Mechanical modeling and InSAR measurements of Mount Sedom uplift, Dead Sea basin: Implications for effective viscosity of rock salt

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We present a mechanical model for the growth of an emerging salt diapir in a tectonically active basin. The analytical model is applied to and serves to constrain the effective viscosity of rock salt and strain rates during diapirism of the wall-shaped Mount Sedom rock salt diapir, Dead Sea basin. The model is based on one-dimensional flow of Newtonian viscous fluid (salt) in a vertical channel that has been driven by the load of the overburden and affected by shear along the channel walls. Because the Poiseuille (channel) flow profile is parabolic and the Couette (shear) flow profile is linear, a one-dimensional model provides three sets of predicted profiles: topography, uplift rate, and shear strain. The present topography of Mount Sedom represents the shape of the Sedom diapir, and hence the effective viscosity of rock salt can be constrained by a model that best fits the present topography of the mountain. The resulting Sedom rock salt viscosity is determined to be between 2 and $3 \times 10^{18}$ Pa s, and the associated strain rate is between 5 and $6 \times 10^{-13}$ s$^{-1}$. Geological structures indicate strain rates of $9 \times 10^{-13}$ s$^{-1}$ and $3 \times 10^{-14}$ s$^{-1}$ during the Holocene emerging stage and at the Plio-Pleistocene pre-emergent stage of the Sedom diapir, respectively. The uplift history of Mount Sedom predicted by the model and the current topography are compared to Interferometric Synthetic Aperture Radar (InSAR) measurements of salt uplift. The maximum uplift rates of Mount Sedom are 8.3 and 5.5 mm/yr for its northern and southern parts, respectively. The InSAR uplift profiles resemble topographic profiles obtained along the same traverses, implying that the uplift history during the last 14,000 years is stable. Steep uplift gradients observed by InSAR along the western margin of the diapir are higher than predicted by modeling of Newtonian viscous flow. This could imply that flow of power law viscous fluid may be more suitable than that of Newtonian viscous fluid for the Sedom rock salt at high strain rates above $8 \times 10^{-13}$ s$^{-1}$.

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1. Introduction

Salt flow commonly forms discordant penetrating structures known as diapirs. These structures have attracted considerable attention over the last few decades due to their importance in controlling the location of natural hydrocarbon traps. Salt diapirs possess physical properties that allow storage of fuel or waste in engineering cavities. Hence constraining the properties of rock salt mass and revealing the strain rates associated with their growth are of great interest both for academic and economic reasons.
[1] The uplift rates of salt diapirs vary considerably between different diapirs, and between stages in the emplacement history of a particular diapir [Jackson and Talbot, 1986]. Generally, diapirs that breach the surface rise faster than those still at the subsurface, mainly because of the effect of the overburden resistance [e.g., Weinberg, 1993]. For example, typical pre-emergent diapirs in the US Gulf Coast and Germany show uplift rates of 0.2 mm/yr and 0.1–0.5 mm/yr, respectively [Jackson and Talbot, 1986; Jartiz, 1987]. On the other hand, the inferred uplift rates of the exposed diapirs of Iran are higher than 10 mm/yr and may reach rates of meters a year [e.g., Talbot et al., 2000; Talbot and Aftabi, 2004]. Likewise, a wide range of strain rates has been obtained for natural rock salt flow between 10^{-15} and 10^{-13} s^{-1} for pre-emergent diapirs, and higher strain rates between 10^{-11} and 10^{-9} s^{-1} for emerging salt diapirs [Talbot and Jackson, 1987]. However, documenting strain rate variations during the emplacement history of a single diapir has not been presented so far.

[4] Modeling of salt diapirism has traditionally been carried out numerically simulating a Rayleigh-Taylor type gravitational instability of fluid layers with suitable density contrasts [e.g., Ramberg, 1981]. These models have been extended to include the effects of topography [Poliakov and Podladchikov, 1992], erosion and sedimentation [Poliakov et al., 1993a, 1993b], tectonic regime [Daudre and Cloetingh, 1994], and rheologically layered structure [Ismail-Zadeh et al., 2002]. The models assume certain viscosity of the rock salt, usually at the lower range, between 10^{17} and 10^{18} Pa s, and simplify the overburden as a viscous layer. Other models treat the overburden as a plastic layer [e.g., Schultz-Ela et al., 1993; Gemmer et al., 2004]. While the model results promote our understanding of the factors controlling diapir shapes and growth, these types of models are mainly applicable to the pre-emergent stage of diapir growth. Other works emphasize the importance of regional tectonics and faulting in the sedimentary overburden rather than gravitational instability [e.g., Jackson and Vendeville, 1994]. It seems that each of those factors can be important and contribute to diapirism in various proportions depending on the geologic setting.

[5] One of the major factors controlling the rate of diapir growth is the effective viscosity of rock salt. This property depends on variables such as water (brine) content, strain rate, temperature and crystal size [e.g., Urai et al., 1986]. It has been shown that at high strain rates, dislocation creep is the dominant flow mechanism of dry coarse-grained (crystal size > 10 mm) salt [Handin et al., 1986; Urai et al., 1986]. On the other hand, solution-transfer diffusive creep dominates at low strain rates for fine-grained (crystal size < 5 mm) salt containing traces of brine [Urai et al., 1986; van Keken et al., 1993]. While dislocation creep is usually described by a power law stress strain rate relation [e.g., Weertman, 1978; Ranalli, 1995], the diffusive creep corresponds to linear Newtonian flow with viscosity depending only on temperature and grain size [van Keken et al., 1993]. A wide range of rock salt viscosities (10^{17} to 10^{20} Pa s) has been suggested depending upon which creep law is implemented for rock salt in numerical models of diapirism [van Keken et al., 1993]. Weinberg [1993] calculated the effective viscosity of a salt mass that lifts large inclusions of dense rocks, and showed that values between 10^{15} Pa s and 10^{18} Pa s are suitable for most diapirs rising at rates higher than 0.5 mm/yr. Talbot et al. [2000] used the dimensions and velocities of the salt mass at Kuh-e-Jahani, Iran, to tune a simple numerical model and constrain the effective viscosity of rock salt to between 10^{16} and 10^{17} Pa s for rate of rise at 2–3 m/yr. It is desirable to further constrain the range of rock salt viscosity values with independent methods that rely on accurate field data and mechanical modeling.

[6] In this study, we construct an analytical model for the growth of an emerging salt diapir in a tectonically active basin. Using accurate measurements of uplift, we apply this model to analyze the growth of Mount Sedom rock salt diapir, Dead Sea basin. The creep behavior of Sedom rock salt is examined at a field-related scale. On the basis of the present topography of the diapir and the time of its emergence, we constrain the effective viscosity of the Sedom rock salt and infer the strain rates associated with its flow. We compare the obtained results with strain measurements as well as with recent observations of Interferometric Synthetic Aperture Radar (InSAR), a technique that has recently become widespread for measurements of subtle displacements of the ground surface [e.g., Gabriel et al., 1989; Massonnet and Feigl, 1998; Baer et al., 2002a, and references therein]. Finally, the Holocene uplift history of the diapir is assessed and compared with the Pleistocene history in order to examine the strain rate variations during the
transition from subsurface to emerging stage of diapir growth.

2. Geologic Setting

Mount Sedom is a north-south trending ridge about 10 km long and 2 km wide, located in the southwestern part of the Dead Sea basin (Figure 1). The mountain is the surface expression of a salt diapir that emerged in the Pleistocene [Zak, 1967]. The diapir attracted some attention in the past as a possible hydrocarbon storage facility in its excavated salt caverns [D’Appolonia Consulting Engineers, 1979, and references therein]. Recommendations for further work that have not been implemented included laboratory estimation of the geotechnical parameters and creep behavior of the Sedom rock salt. The Dead Sea basin is a continental depression located within the rift valley that stretches along the Dead Sea Fault (Transform). It is widely agreed that the basin is a rhomb-shaped pull-apart graben that was formed due to the left-lateral displacement along the segmented Dead Sea Fault [e.g., Quennell, 1958; Freund et al., 1970; Garfunkel, 1981]. The basin is bounded by a series of oblique-normal faults. The total

Figure 1. (a) Location map showing the regional setting of the Dead Sea Transform (DST) and Mount Sedom. Salt bodies along the Dead Sea rift are marked by solid circles [Weinberger et al., 2006, and references therein]. Arrows indicate the ingression path of the Mediterranean Sea during the Neogene. (b) Shaded relief image of Mount Sedom and its vicinity. The western border fault (partly dotted) forms an escarpment and separates outcrops of Cretaceous rocks to the west of Ami’az Plain from the Quaternary fill of the Dead Sea basin. Solid squares mark the locations of three drill holes in the vicinity of Mount Sedom. Location of seismic profile RV7003 (Figure 2) is marked.
throw of pre-basin rocks in its center is estimated from gravity and seismic data to be in the range of 8.5–10 km [Zak, 1967; ten Brink et al., 1993; Ben-Avraham, 1997; Al-Zoubi and ten Brink, 2001]. Temperature measurements in deep oil wells drilled in the Dead Sea basin show low geothermal gradients of ~20°/km [Kashai and Croker, 1987].

[s] Neogene massive evaporite beds and several types of evaporite bodies have been found in the Dead Sea rift, between the Sea of Galilee and the Dead Sea (Figure 1) [Weinberger et al., 2006, and references therein]. They are overlain by a thick Plio-Pleistocene overburden comprising mainly sedimentary fill of fluvo-lacustrine origin. The geographic extent of these evaporite bodies delineates a narrow, elongated gulf within the tectonically subsiding Dead Sea rift, into which seawater penetrated through the Yizre’el valley during the Late Miocene to Pliocene (Figure 1) [Avnimelech, 1937; Zak, 1967; Zak and Bentor, 1972; Shaliv, 1991]. Of this series of salt bodies the Sedom diapir is the only one exposed. The source of its salt layer is located in the deepest part of the southern Dead Sea basin, now buried beneath sediments of a maximum thickness of ~5,500 m [Al-Zoubi and ten Brink, 2001].

[s] Seismic profiles in the Dead Sea basin indicate that many, if not all, salt bodies in the basin are located above or near deep-seated faults [Neev and Hall, 1979; ten Brink and Ben-Avraham, 1989, Gardosh et al., 1997]. The piercing Sedom diapir is found adjacent to the north-south striking Sedom fault, a major tectonic line in the western margin of the basin. The diapir has an antiformal structure, the thin (<500 m) western limb of which either remained buried below the Ami’az Plain [Zak, 1967] or is exposed [Frumkin, 1996c; Weinberger et al., 1997]. The piercing salt diapir tilted the overlying sequence about 70° eastward along the eastern flank of Mount Sedom, and about 40° westward along its western flank. Seismic sections in the Dead Sea basin show low geothermal gradients of ~20°/km [Kashai and Croker, 1987; Gardosh et al., 1997]. The evaporite series, mainly highly tilted bedded rock salt 1,500–2,000 m thick, builds most of Mount Sedom and is known as the Sedom Formation [Zak, 1967]. It pierced the overlying sediments, reaching an elevation of ~160–200 m below mean sea level (msl). Mount Sedom, which comprises the diapir and a veneer of Late Quaternary sediments (including up to a 40 m thick caprock) rises about 90 m above the Ami’az Plain and 250 m above the Dead Sea (Figure 1). The age of the Sedom Formation is most likely Pliocene, on the basis of stratigraphic [Zak, 1967] and palynological [Horowitz, 1987] evidence. The formation is composed of sequences of well-bedded fine-grained (<5 mm) rock salt [Zak, 1967, p. 33] alternating with thin clastic units. The piercing salt diapir tilted the overlying sequence about 70° eastward along the eastern flank of Mount Sedom, and about 40° westward along its western flank. This sequence consists of marl, chalk, sandstone and conglomerate that overlie the Sedom Formation, ~400 m of which are exposed on the eastern flank of Mount Sedom and are known as the Amora Formation [Zak, 1967]. The upper ~40 m consists of authigenic aragonite and gypsum layers alternating with silty and sand layers forming the Lisan Formation [Zak, 1967]. The age of the Lisan Formation, as established through U-series, is ~70,000–14,000 years B.P. [Kaufman, 1971; Schramm et al., 2000; Haase-Schramm et al., 2004]; on top of Mount Sedom only the younger section (<43,000 years B.P.) is exposed [Weinberger and Bar-Matthews, 2004]. Parts of the rock salt of the Sedom diapir were dissolved by meteoric groundwater and by lake water to which it was exposed during the late Pleistocene [Vroman, 1951; Zak, 1967]. Less soluble components of the Sedom Formation, such as anhydrite, gypsum, marl, dolomite, and clastic material formed a ~40 m thick caprock over the entire Sedom diapir. A flat dissolution surface termed “salt mirror” (salt table) separates the rock salt from the overlying caprock. The salt mirror is thought to have been “fossilized” when a new dissolution base developed at a lower level, when the southern basin of the Dead Sea desiccated [Zak, 1967]. This event took place between 14,000 and 11,000 years B.P., when lake level dropped sharply to 400 m below msl [Neev and Emery, 1967; Yechiel et al., 1993; Frumkin, 1996b].

[10] A 900 m thick evaporite series overlain by 3,700 m thick fluvo-lacustrine series were penetrated in Sedom Deep-1 drill hole southeast of Mount Sedom, and ~1,300 m thick evaporite series overlain by 1,900 m thick fluvo-lacustrine series were penetrated in Ami’az East-1 drill hole west of Mount Sedom [Kashai and Croker, 1987; Gardosh et al., 1997]. The evaporite series, mainly highly tilted bedded rock salt 1,500–2,000 m thick, builds most of Mount Sedom and is known as the Sedom Formation [Zak, 1967]. It pierced the overlying sediments, reaching an elevation of ~160–200 m below mean sea level (msl). Mount Sedom, which comprises the diapir and a veneer of Late Quaternary sediments (including up to a 40 m thick caprock) rises about 90 m above the Ami’az Plain and 250 m above the Dead Sea (Figure 1). The age of the Sedom Formation is most likely Pliocene, on the basis of stratigraphic [Zak, 1967] and palynological [Horowitz, 1987] evidence. The formation is composed of sequences of well-bedded fine-grained (<5 mm) rock salt [Zak, 1967, p. 33] alternating with thin clastic units. The piercing salt diapir tilted the overlying sequence about 70° eastward along the eastern flank of Mount Sedom, and about 40° westward along its western flank. This sequence consists of marl, chalk, sandstone and conglomerate that overlie the Sedom Formation, ~400 m of which are exposed on the eastern flank of Mount Sedom and are known as the Amora Formation [Zak, 1967]. The upper ~40 m consists of authigenic aragonite and gypsum layers alternating with silty and sand layers forming the Lisan Formation [Zak, 1967]. The age of the Lisan Formation, as established through U-series, is ~70,000–14,000 years B.P. [Kaufman, 1971; Schramm et al., 2000; Haase-Schramm et al., 2004]; on top of Mount Sedom only the younger section (<43,000 years B.P.) is exposed [Weinberger and Bar-Matthews, 2004]. Parts of the rock salt of the Sedom diapir were dissolved by meteoric groundwater and by lake water to which it was exposed during the late Pleistocene [Vroman, 1951; Zak, 1967]. Less soluble components of the Sedom Formation, such as anhydrite, gypsum, marl, dolomite, and clastic material formed a ~40 m thick caprock over the entire Sedom diapir. A flat dissolution surface termed “salt mirror” (salt table) separates the rock salt from the overlying caprock. The salt mirror is thought to have been “fossilized” when a new dissolution base developed at a lower level, when the southern basin of the Dead Sea desiccated [Zak, 1967]. This event took place between 14,000 and 11,000 years B.P., when lake level dropped sharply to 400 m below msl [Neev and Emery, 1967; Yechiel et al., 1993; Frumkin, 1996b].

[11] Milestones in the rise of the Sedom diapir were summarized by Weinberger et al. [2006] from various types of evidence, including angular and
erosional unconformities, thickness variations, chemistry and isotope variations of aragonite laminae in the Lisan Formation, and cave deposits. It seems that the Sedom rock salt began to migrate laterally and upward at about 2.2 Ma (Figure 2). It emerged at \( \frac{1}{2} \times 300,000 \) years, when its upper part penetrated into a lake that filled the Dead Sea depression and the exposed rock salt was dissolved. The diapir breached the surface at ca. 70,000 to 43,000 years B.P., and between 43,000 and 14,000 years B.P. it was inundated by Lake Lisan. At 14,000 years B.P., when Lake Lisan level declined sharply, the Sedom diapir reemerged and started forming the present topography. The Holocene rate of rise of the Sedom diapir has been estimated by several approaches to vary between 5 and 11 mm/yr [Weinberger et al., 2006, and references therein].

3. InSAR Observations

[12] InSAR measurements were previously applied to the Mount Sedom uplift by Baer et al. [2002b], Pe’erí et al. [2004], and Weinberger et al. [2006]. In this study we reanalyzed some of our previous data using the ROI-PAC software [Rosen et al., 2004]. SAR data used for this study were collected by the European Space Agency Remote Sensing satellites ERS-1 and ERS-2. Interferograms were made for periods of 3–89 months between June 1992 and January 2001 (Table 1). The uplift of the diapir is directly related to the phase changes as observed in the interferogram. This phase is measured modulo \( 2\pi \), in fringe cycles representing half a wavelength (28 mm) of range change in the satellite to ground line of sight. For analysis of the diapir geometry and uplift rate we convert these cycles to a continuous signal (unwrap the phase), using a procedure built in the ROI-PAC software that adds the correct integer number of phase cycles to each measurement point. Only Interferograms with higher coherence within the interval 1993–2001 were phase-unwrapped. The pattern of uplift in Mount Sedom is best shown by the most coherent unwrapped interferogram for the period 10/1997 – 01/2001 (Figure 3). A series of oblique three-dimensional unwrapped interferograms showing the growth of Sedom diapir from 1993 to 2001 is presented in Figure 4.

[13] The geometry of the rising diapir is examined by uplift profiles along ten W-E sections and one N-S section for the period 10/1997 – 01/2001 (Figure 5). The stable Ami‘az Plain provides a...
reference surface. In most profiles the uplift patterns resemble the current topography. To exclude the likelihood that these similarities are due to topographic contribution or atmospheric layering, we examined phase changes along the steep topographic slopes of the Dead Sea basin away from the Sedom diapir. We see no phase-topography correlation, and thus we rule out the possibility of topographic or atmospheric artifacts in the Mount Sedom interferograms. Most W-E profiles show a pronounced asymmetry, with steep gradients and a relatively flat top on the western side of the diapir, and milder gradients on the eastern side (Figure 5). The steep gradients coincide with previously mapped piercement faults [Zak, 1967] or their along-strike continuations. Mild uplift gradients are generally associated with zones where bedding plane slippage has been observed or suggested [Zak and Freund, 1980]. The maximum uplift rates for Mount Sedom rock salt diapir are 8.3 and 5.5 mm/yr for its northern and southern parts, respectively (Figures 3 and 5).

A notable narrow ridge is observed in almost all the uplift profiles, encircling the eastern margins of the diapir (Figures 3, 4, and 5). This feature coincides with the bottom of the steep eastern wall of the mountain and does not show any surface evidence for differential uplift. We interpret this feature as follows: The satellite look direction is from ESE to WNW (descending track), and the incidence angle is 23° from the vertical. Along the eastern wall of Mount Sedom the slope is steeper than the line perpendicular to the direction of the radar pulse. The higher parts of the rising eastern wall are thus imaged before some of the area that lies at the bottom of the wall (which is closer in range to the satellite) forming a layover SAR effect. The interferogram resulting from this effect duplicates the uplift signal and shows it both on the diapir and at the overlaid area in front of the mountain. The narrow ridge is thus interpreted as a SAR artifact.

### 4. Strain and Strain Rate Measurements

Six types of deformation features were described and measured at the Sedom Quarry

![Figure 3. Unwrapped interferogram for the period 10/1997 to 1/2001 showing the amount of uplift in Mount Sedom. The location of uplift and topography profiles and fault map of Mount Sedom [Zak and Freund, 1980] are annotated.](image)
(Figure 1b), on the eastern side of Mount Sedom, and their strain was calculated by Zak and Freund [1980]. They all occur within a vertical rock salt sequence about 200 m thick, containing minor anhydrite, dolomite and clay interbeds. Three of these features (stretched desiccation polygons, rotated boudins, and flattened anhydrite aggregates) were formed at great depth. The other three features (folded dolomite beds, folded rock salt beds, and vertical displacement of the salt mirror along bedding planes) were formed at shallow depth and are therefore younger than the features of the first group. Zak and Freund [1980] concluded that the features of both groups indicate that the uplift of Mount Sedom was achieved by vertical shear. The sequence at the eastern margin of the diapir underwent a finite shear strain of at least 2.1, the sum of 1.7 ductile shear recorded by features of the first group, and 0.4 slip of brittle shear recorded by the salt mirror [Zak and Freund, 1980]. The shear is smaller westward, and over the whole 1-km-wide backbone of Mount Sedom the displacements add only another 50 m before reaching the western margin of the mountain with its reciprocal slip displacements [Zak, 1967]. Because the salt mirror was fossilized about 14,000 to 11,000 years ago, the strain recorded by the displacement of this surface represents the Holocene finite strain of the diapir.

[16] The ductile shear strain of 1.7 was achieved within a period of upward salt migration at depth in the last 2.2 Ma (Figure 2). These values suggest a minimum shear strain rate of $\sim 3 \times 10^{-14}$ s$^{-1}$ for
Figure 5
the Plio-Pleistocene pre-emergent stage of Sedom diapir. The brittle shear strain of 0.4 achieved during the last 14,000 years is associated with an average strain rate of \( \sim 1 \times 10^{-12} \text{ s}^{-1} \). The latter strain rate is higher by more than an order of magnitude relative to the average strain rate calculated for the pre-emergent stage of the diapir.

[17] The current shear strain rate is calculated from the InSAR results. The shear strain rate is the slope of the uplift rate over a certain interval of distance. For the 200 m interval from the easternmost margin of the diapir westward along profiles h, i, j, which cross the southern part of the mountain near the Sedom Quarry, the present calculated shear strain rate is between \( 5 \times 10^{-13} \) and \( 6 \times 10^{-13} \text{ s}^{-1} \).

5. Mechanical Modeling of the Sedom Diapir Growth

5.1. Model Formulation

[18] On the basis of the geology of Mount Sedom and its vicinity, we formulate a simple mechanical model for the Holocene rise of the Sedom diapir. Its basic assumptions and their justification are as follows: (1) The wall-shape of the Sedom diapir [e.g., Zak, 1967; Gardosh et al., 1997], which resulted from its association with N-S striking oblique-normal faults [Weinberger et al., 2006], allows one-dimensional approximation of salt flow in a vertical channel. (2) Salt flow within the channel is driven by the load of the overburden and affected by the differential subsidence of the diapir’s flanks due to displacement along the downthrown (basinward) normal faults. (3) The thickness of the relict salt source layer at the southern Dead Sea basin is now about 900 m in large areas in the southern Dead Sea basin [Al-Zoubi and ten Brink, 2001], and hence the available amount of salt at the channel entrance is practically infinite for the relatively short time interval that is considered here (\(< 20,000\) years). (4) Over this time interval, the source salt layer is loaded by a constant pressure \( P_r \), that is applied by the overlying sedimentary column. The additional pressure exerted by the veneer of sediments accumulated during the last 14,000 years in the Deep Block (Figure 2), which is less than 50 m [Neev and Emery, 1995] is negligible with respect to that exerted by the thick Plio-Pleistocene overburden (\( \sim 4,000\) m). (5) The present topography of Mount Sedom is the result of salt emergence during the last 14,000 years B.P. since the recession of Lake Lisan [Zak, 1967; Frumkin, 1994; Weinberger et al., 2006]. We assume that at the time of emergence, most of the host-rock roof had already been displaced upward and outward and the thickness of the veneer of late Quaternary sediments at the top of Mount Sedom is less than 40 m [Zak, 1967]. The corresponding pressure at the channel vent is less than 0.5 MPa \(< P_r \) and hence is negligible. (6) The geological reference surface above which the salt has emerged is the top of the Lisan Formation at the Am‘az Plain, which is currently stable (Figure 3) and has been most likely stable during the last 14,000 years. East of the Am‘az Plain, the position of this reference surface has been controlled initially by the bottom slope of Lake Lisan and it changed with time due to the uplift of Mount Sedom and subsidence of the southern Dead Sea basin along the stepped faults. (7) The overburden is about 1,700 m thick west of Mount Sedom at the Intermediate Block (Figure 2) and more than 3,700 m thick east of Mount Sedom at the Deep Block. The latter provides a buoyancy force that is twice as high as that at the Intermediate Block. A paleomagnetic reconstruction [Weinberger et al., 1997] and a structural mismatch of strata in caves [Frumkin, 1996c] indicate that more than three quarters of the exposed section of the diapir is built of salt rising from the Deep Block. We ignore the migration of salt from the salt source layer in the Intermediate Block eastward into the adjacent diapir.

[19] Salt creep is commonly approximated by flow of either Newtonian or power law viscous fluid [e.g., Weinberg, 1993; Poliakov et al., 1993a, van Keken et al., 1993; Ismail-Zadeh et al., 2002]. We assume that diffusive creep at low strain rates of \( 10^{-12}\text{ to }10^{-13} \text{ s}^{-1} \), deformed the fine-grained Sedom rock salt. Hence we adopt a Newtonian rheology with a constant effective salt viscosity \( \eta \) and construct an analytical model of the Sedom diapir emergence. We consider a pressure-driven salt flow against the gravity through a vertical channel of length \( h \) and a constant half-width, \( w \). Conveniently, the origin of the coordinate system \((0,0)\) is located along the centerline of the channel and is fixed to the
The pressure at the top of the horizontal salt source layer in the Deep Block, $P_f$, is

$$P_f = \int_0^H \rho_{\text{sediment}}(z) \cdot g \cdot dz,$$

(1a)

where $g$ is the acceleration due to gravity, and $\rho_{\text{sediment}}(z)$ is the density of the sedimentary overburden that varies with depth and is calculated by integrating the density variations of the Sedom salt $\rho_{\text{sedom}}$ measured in Sedom Deep-1 drill hole over depth $H$. As mentioned above, $P_f$ is assumed to remain constant over the time interval considered here. The pressure at the entrance to the channel located at $(0,0)$, $P_{in}$, is not exactly equal to $P_f$. This is due to continuous subsidence of the Deep Block (see text). (a) Initial conditions at $t = 0$. The pressure in the channel vent is zero. (b) Conditions during diapir rise ($t > 0$). $v_t$ is the tectonic (subsidence) rate which accounts for the relative offset between Ami’az Plain and the Dead Sea base level. $S(x, t)$ is the evolving surface topography of the diapir above the base level. The model neglects discreet vertical bedding-plane faults shown in the uppermost part of the diapir.

middle of the line connecting the western and eastern walls of the channel (Figure 6). For comparison with the geological data (topography, InSAR data) presented in Figures 7 and 8, the origin of the coordinate system was shifted up to the level of the Ami’az Plain.

[20] The pressure at the top of the horizontal salt source layer in the Deep Block, $P_f$, is

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where $g$ is the acceleration due to gravity, and $\rho_{\text{sediment}}(z)$ is the density of the sedimentary overburden that varies with depth and is calculated by integrating the density variations of the Sedom salt $\rho_{\text{sedom}}$ measured in Sedom Deep-1 drill hole over depth $H$. As mentioned above, $P_f$ is assumed to remain constant over the time interval considered here. The pressure at the entrance to the channel located at $(0,0)$, $P_{in}$, is not exactly equal to $P_f$. This is due to continuous subsidence of the Deep Block, leading to increase of the depth difference $\delta h(t)$ between the location of the channel entrance $(0,0)$ and the top of the salt source layer (Figure 6). The difference in pressure $\delta P(t)$ is approximated by the weight of the salt layer with a thickness of $\delta h(t)$. In the following section, we show that $\delta h(t) \ll H$ and, consequently, $\delta P(t)$ could be estimated by a constant value $\delta P$ calculated by averaging $\delta P(t)$ for $t = 0$ and for $t = 14,000$ years.
Finally, the pressure at the entrance of the channel, $P_{in}$, is calculated by subtracting the pressure exerted by a column of salt in the channel $P_s = \rho_{sedom} \cdot g \cdot H$ and $\delta \rho$ from $P_f$:

$$P_{in} = P_f - P_s - \delta P = \int_0^H \rho_{sedom}(z) \cdot g \cdot dz - \rho_{sedom} \cdot g \cdot H - \delta P.$$  

[21] The pressure at the channel vent, $P_{out}(t)$, changes over time due to the pressure build-up exerted by the extruded salt above the reference surface. The volume of the extruded salt is calculated by integrating the topography above the reference surface. Hence $P_{out}(t)$ has the form

$$P_{out}(t) = \frac{1}{2W} \int_{-W}^{W} k_{sedom} \cdot g \cdot S(x, t) dx,$$

where $S(x, t)$ is the evolving surface topography of the diapir above the base level, and the constant $k < 1$ accounts for the effective reduction in salt-mass density due to existence of a perforated karst system above the base level. $k$ is strongly dependent on the annual precipitation in the area and the corresponding amount of dissolved salt from Mount Sedom.

[22] The differential pressure, $P_{in} - P_{out}(t)$, drives salt expulsion into the diapir. The velocity profile $v(x, t)$ results from the superposition of the parabolic profile of the Poiseuille (channel) flow and

Figure 7. East-west topographic profiles b and c across the northern part of Mount Sedom and h, i, and j across the southern part of Mount Sedom drawn from 25-m-grid DTM [Hall, 1997]. For location of the profiles, see Figure 3. For $t = 14,000$ years, best fitting model result for profiles h, i, and j (heavy black line) yields $\eta = 2.3 \times 10^{18}$ Pa s. For the same viscosity, the best fitting model result for profiles b and c (heavy black line) yields $t = 9,800$ years. Dashed line represents the uplift for $\eta = 2.3 \times 10^{18}$ Pa s and $t = 14,000$ years. See text for details.

Figure 8. East-west uplift-rate profiles b and c across the northern part of Mount Sedom and h, i, and j across the southern part of Mount Sedom based on InSAR data and the present model. Heavy line shows the model prediction for $t = 14,000$ years and $\eta = 2.3 \times 10^{18}$ Pa s. Dashed line shows the model prediction for the northern part with $t = 9,800$ years and $\eta = 2.3 \times 10^{18}$ Pa s.
the linear profile of the Couette (shear) flow [e.g., Turcotte and Schubert, 1982, pp. 232–234]:

\[ v(x, t) = \frac{P_{in} - P_{out}(t)}{2\eta H} \left( W^2 - x^2 \right) - v_i \frac{x}{W}, \]  

(3)

where \( 2v_i \) is the tectonic (subsidence) rate (Figure 6). The volumetric flux of salt per unit area, \( q(t) \), is

\[ q(t) = \frac{1}{2W} \int_{-W}^{W} v(x, t) \cdot dx. \]  

(4)

Substituting equation (3) into equation (4) and integrating along the channel width provides

\[ q(t) = \frac{P_{in} - P_{out}(t)}{3\eta H} \cdot W^2. \]  

(5)

Equating the volumetric flux per unit area to the pressure build-up during salt extrusion \((dP_{out}(t)/dt)\) gives

\[ \frac{dP_{out}(t)}{dt} = \frac{P_{in} - P_{out}(t)}{3\eta H} W^2 \cdot k \cdot \rho_{sedom} \cdot g. \]  

(6)

The solution for equation (6), with the initial condition \( P_{out}(0) = 0 \), is

\[ P_{out}(t) = P_{in} \left[ 1 - \exp \left( -\frac{t}{\tau} \right) \right]. \]  

(7)

where the characteristic timescale \( \tau \) has the following form:

\[ \tau = \frac{3\eta H}{W^2 \cdot k \cdot \rho_{sedom} \cdot g}. \]  

(8)

The solution for \( P_{out}(t) \) (equation (7)) shows an exponential buildup from 0 to \( P_{in} \) with the rate depending on channel geometry and salt properties. Respectively, the driving pressure, \( P_{in} - P_{out}(t) \), decays exponentially by a factor “\( e \)” for \( t = \tau \) (“half-life time”).

[23] Substituting equation (7) into equation (3) provides an expression for the velocity profile:

\[ v(x, t) = \frac{(W^2 - x^2)}{2\eta H} P_{in} \exp \left( -\frac{t}{\tau} \right) - v_i \frac{x}{W}. \]  

(9)

[24] To obtain an expression for the evolving surface topography of the emerging diapir \( S(x, t) \), we account for the sediment erosion:

\[ \frac{\partial S}{\partial t} = v(x, t) - \hat{n}(t), \]  

(10)

where \( \hat{n}(t) \) is the erosion rate. Substituting equation (9) into equation (10) after integrating, we obtain

\[ S(x, t) = S_0(x) + \left( \frac{W^2 - x^2}{2\eta H} P_{in} \cdot \tau \left[ 1 - \exp \left( -\frac{t}{\tau} \right) \right] \right) - v_i \frac{x}{W} t - \hat{n}(t), \]  

(11)

where the first term \( S_0(x) \) accounts for the initial slope between the reference surface and the subsiding base level (Figure 6); the second and the third terms account for the Poiseuille and Couette flow components, respectively. The last term in equation (11), \( \hat{n}(t) \), is the result of integrating \( \hat{n}(t) \) in equation (10), and accounts for the effect of erosion above the reference surface.

[25] To obtain an expression for the shear strain rate \( \dot{\varepsilon}(x, t) \) profile, we calculate the first derivative of the velocity (equation (9)) by \( x \):

\[ \dot{\varepsilon}(x, t) = -\frac{W^2 - x^2}{\eta H} P_{in} \cdot \exp \left( -\frac{t}{\tau} \right) - v_i \frac{x}{W}. \]  

(12)

The shear strain profile is the first derivative of the surface topography (equation (11)) by \( x \), ignoring the strain accumulated prior to the breaching of the diapir at \( t = 0 \).

\[ \varepsilon(x, t) = -\frac{W^2 - x^2}{\eta H} P_{in} \cdot \tau \left( 1 - \exp \left( -\frac{t}{\tau} \right) \right) - v_i \frac{x}{W} t. \]  

(13)

[26] The Poiseuille component of the velocity and strain rate profiles (equations (9) and (12), respectively) exponentially decays with the characteristic timescale \( \tau \), while this component in the surface topography and strain profiles (equations (11) and (13), respectively) approaches asymptotic values.

5.2. Model Input

[27] The input for the model includes the diapir (channel) geometry, the driving pressure, the time of diapir Holocene emergence, the slope of the top Lisan Formation above the diapir prior to current emergence, the tectonic (subsidence) rate of the southern Dead Sea basin compared to the Ami’az Plain, and the erosion rate. Table 1 summarizes the estimated values and their sources. Several points on the estimated values are elaborated below.

5.2.1. Diapir Geometry

[28] The width of the exposed sections of the diapir provides a first order approximation for the channel width, \( 2W \). Combined with data from seismic profiles (Figure 2), a value of 2,000 m is assigned to the channel width. The channel length \( H \) of
3,700 m is based on the interpretation of the seismic profile [Gardosh et al., 1997], and the thickness of the overburden close to the diapir, which is taken from the Sedom Deep-1 drill hole.

5.2.2. Subsidence (Tectonic) Rate

[29] Palynological evidence for the Holocene rate of accumulation in the southern Dead Sea basin indicates a considerable subsidence of about 100 m during this period [Horowitz, 1989, p. 66]. This implies a very high rate of tectonic subsidence of at least 9 mm/yr. Holocene deposits were penetrated by several drill holes [Neev and Emery, 1967] and their accumulation rate is about 3 mm/yr [Neev and Emery, 1995]. For a wider range of time intervals, the reported subsidence at the Dead Sea basin is between 1.5 and 5.5 mm/yr [Begin and Zilberman, 1997, and references therein]. A small fraction of these values (<1 mm/yr) may be attributed to compaction and subsidence due to thinning of the source salt layer during withdrawal of salt to the Sedom and Lisan diapirs [Al-Zoubi and ten Brink, 2001]. The inferred displacement of the top of Lisan Formation along the Amazyahu fault based on the present escarpment topography is at least ~70 m, indicating a subsidence rate of 5 mm/yr. We assign this value for the average subsidence (tectonic) rate at the southern Dead Sea basin over the last 14,000 years.

5.2.3. Initial Dip of the Lisan Sediments (Lake Bottom Slope)

[30] The top of the Lisan Formation serves as a reference for the diapir uplift since ~14,000 years B.P. This reference has an elevation of about 265 m below msl at Ami’az Plain, and 165 m below msl at the top of Mount Sedom. East of Mount Sedom, the top of the Lisan Formation is covered by Holocene deposits. These deposits were penetrated by several drill holes [Neev and Emery, 1967] and the top of the Lisan Formation was recognized at a depth of 415–420 m bsl [Neev and Emery, 1995]. The 150 m elevation difference between the two sides of Mount Sedom is attributed to the combined effect of sedimentation on an initial slope of the lake bottom and basin subsidence during the Holocene. The above estimated subsidence rate of 5 mm/yr at the southern Dead Sea basin over the last 14,000 years indicates that 70 m of the total 150 m can be attributed to subsidence, and the remaining 80 m to initial dip of the Lisan sediments. The latter value indicates a lake bottom slope of ~4% (80 m over the ~2,000 m width of Mount Sedom). This value is comparable to slopes of the top of the Lisan Formation elsewhere within the Dead Sea basin. This surface was preserved in the southern Dead Sea basin and is only mildly disturbed by Holocene tectonics within the block [Sneh et al., 1998]. The calculated slopes of this surface along W-E and N-S traverses are 5% and 1%, respectively (see data of Bartov et al. [2002] and Druckman et al. [1987]).

5.2.4. Driving Pressure

[31] $P_f$ is calculated by integrating the density profile measured in Sedom-Deep-1 drill hole [Rybakov et al., 1995] over a depth of $H = 3,700$ m. This calculation may underestimate $P_f$ due to possible escape of brines from the pores of the sedimentary fill before measurements. To evaluate this effect, we calculate the density profile of sedimentary column with similar composition following the compaction curves of Baldwin and Butler [1985], assuming that all pores are filled with modern Dead Sea brine of density 1,235 kg/m³ [Salhotra et al., 1985; Gavrieli, 1997]. The similarity between the calculated density profile and the density log measured directly from Sedom Deep-1 drill hole suggests that $P_f$ is adequately calculated from these data.

[32] $P_{in}$ is calculated by equation (1b). The density of Sedom rock salt does not vary significantly with depth and a constant value of $\rho_{sedom} = 2,115$ kg/m³ is assigned. This value is slightly higher than that of pure halite to account for the denser minor components of Sedom Formation. According to the lake-bottom slope and tectonic (subsidence) rate, $h_b$ and the corresponding $\delta P$ are calculated (Figure 6). Values of $\delta P = 0.2$ MPa for $t = 0$ and $\delta P = 1.6$ MPa for $t = 14,000$ years lead to an average value of $\delta P = 0.9$ MPa, resulting in a constant $P_{in} = 7.3$ MPa.

5.2.5. Time of Diapir Current Emergence

[33] The time of the model onset is $t_{onset} = 14,000$ years B.P. At this time, Lake Lisan level declined sharply and the salt diapir started its current phase of emergence. This time is consistent with the exposure age of karst landscape for the southern part of Mount Sedom [Frumkin, 1996b] and with the age assigned to the salt mirror. For the northern part of the mountain the onset time may lag by 3,000 years [Frumkin, 1996b].

5.2.6. Erosion Rate and the Constant k

[34] The current climate in the Sedom area is hyper-arid with an average rainfall of 50 mm/yr.
Table 2. Values of Measurements and Model Output Variables

<table>
<thead>
<tr>
<th>Variable</th>
<th>Value</th>
<th>Time Interval, years</th>
<th>Stage</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Effective viscosity</td>
<td>$2.0 \times 10^{18}$ Pa s</td>
<td>14,000</td>
<td>emergence</td>
<td>Topography and modeling</td>
</tr>
<tr>
<td>Shear strain</td>
<td>1.7</td>
<td>2,200,000</td>
<td>pre-emergence</td>
<td>rotated boudins</td>
</tr>
<tr>
<td>Shear strain</td>
<td>0.4</td>
<td>14,000</td>
<td>emergence</td>
<td>displacement of salt mirror</td>
</tr>
<tr>
<td>Shear strain</td>
<td>0.27 – 0.33</td>
<td>14,000</td>
<td>emergence</td>
<td>modeling</td>
</tr>
<tr>
<td>Average strain rate</td>
<td>$3.0 \times 10^{-14}$ s$^{-1}$</td>
<td>2,200,000</td>
<td>pre-emergence</td>
<td>rotated boudins</td>
</tr>
<tr>
<td>Average strain rate</td>
<td>$9.0 \times 10^{-13}$ s$^{-1}$</td>
<td>14,000</td>
<td>emergence</td>
<td>displacement of salt mirror</td>
</tr>
<tr>
<td>Strain rate</td>
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<td>3.25</td>
<td>emergence</td>
<td>InSAR data</td>
</tr>
<tr>
<td>Strain rate</td>
<td>$6.0 - 7.5 \times 10^{-13}$ s$^{-1}$</td>
<td>14,000</td>
<td>emergence</td>
<td>modeling</td>
</tr>
</tbody>
</table>

*Strain and strain rate are measured and calculated for a 200 m interval from the easternmost margin of the diapir westward.

For this climate, Gerson [1972] estimated an erosion rate of 0.2 mm/yr, and we assume a similar erosion rate over the last 14,000 years.

Frumkin [1994] calculated a minimum solute discharge at Mount Sedom of $15 \times 10^6$ kg/yr that is attributed mainly to rock salt dissolution along karstic conduits. Assuming that this annual discharge has been constant over the Holocene, a first order approximation of the removed weight from the mountain is up to 5% of its total weight above the base level. Hence $k$ that accounts for the effective reduction in salt-mass density due to existence of the perforated karst system is 0.95.

6. Results

The model output includes three sets of data that can be obtained for any time after the current phase of emergence at $t_{onset} = 14,000$ years B.P. (Table 2). The first data set is based on equation (11) and it provides the predicted topography profile $S(x, t)$ of the Sedom diapir. The second data set is based on equation (9) and it provides the uplift rate profile $v(x, t)$. The third data set is based on equation (13) and it provides the shear strain profile across the channel $\varepsilon(x, t)$. We first estimate the effective viscosity of the Sedom rock salt by standard curve fitting $S(x, t)$ to the present topography. We then use the estimated value of the effective viscosity to calculate uplift rate profiles $v(x, t)$ and compare them to uplift rate profiles based on available InSAR data. Finally, the calculated shear strain $\varepsilon(x, t)$ is compared to independent field measurements of shear strain at the eastern margins of the Sedom diapir. To calculate the topography, uplift rate and shear strain, the variables constrained above (see section 5.2 and Table 1) are inserted into equations (9)–(13).

6.1. Effective Viscosity of the Sedom Rock Salt

The model output data sets depend on the effective viscosity $\eta$ of the Sedom rock salt, which was not previously determined. Because the topography of Mount Sedom represents the shape of the Sedom diapir [Zak and Bentor, 1972], we obtain $\eta$ by searching for the best fit between the calculated topography $S(x, t)$ for $t = 14,000$ years and the present topography of Mount Sedom using the least squares method. Hence, for all of the following fits, the elevations at the western and eastern margins of the channel are 265 and 415 m below msl, respectively, according to the present elevation of top Lisan Formation at these localities. The search for the best fit is done simultaneously along several E-W topographic traverses across the mountain drawn from 25-m-grid DTM [Hall, 1997]. Because more data is available from the southern part of the mountain (i.e., exposed sections, strain measurements, seismic data) and its structure is more adequate for modeling (i.e., sheared “card-deck like” vertical beds, Figure 6b), we first fit the modeled topography to this part of the mountain. Figure 7 shows three representative topographic traverses for the southern part of Mount Sedom (profiles h, i, j; see Figure 3 for profile locations), and the model best fit predicted topography to these traverses for $t = 14,000$ years (assuming $t_{onset} = 14,000$ years B.P.). The best fit with a standard deviation of $2\sigma = 25$ m is obtained for a salt viscosity of $2.3 \times 10^{18}$ Pa s. The predicted diapir surface profile is asymmetric with respect to the channel centerline, as indicated by the
westward offset of the highest point (170 below msl).

[38] Assuming that Sedon rock salt viscosity is everywhere similar along the mountain, and recalling that ages of the most ancient salt cave indicate that the northern part of Mount Sedon might lag by about 3,000 years after the emergence of the southern part of Mount Sedon [Frumkin, 1996b], we insert the obtained viscosity (\(\eta = 2.3 \times 10^{18} \text{ Pa s}\)) and search for the best fit \(t_{\text{onset}}\). The model best fit with \(2\sigma = 30\) m to two representative northern traverses b, c yields \(t_{\text{onset}} = 9,800\) years B.P. for the emergence of the northern part of Mount Sedon. This fit suggests a time lag of 4,200 years after the emergence of the southern mountain (Figure 7), in a satisfactory agreement with the independently estimated time lag. The peak elevation predicted by the model for the northern part of Mount Sedon is somewhat lower than the actual one. Moreover, while the model satisfactorily reproduces the profile of the present topography from the centerline of the mountain eastward, it fails to reproduce the sharp topographic gradients from the centerline westward. If the model topography is calculated using both \(t_{\text{onset}} = 14,000\) years B.P. and \(\eta = 2.3 \times 10^{18} \text{ Pa s}\), it reproduces better the peak elevation of the northern Mount Sedon than that obtained for \(t_{\text{onset}} = 9,800\) years B.P., but shows increasing mismatch from the centerline of the mountain eastward with \(2\sigma = 65\) m (Figure 7).

6.2. Predicted and Observed Uplift Rate

[39] We calculate uplift rate profiles for the northern and the southern parts of the mountain by inserting the value obtained above for the effective viscosity (equation (9)). The calculations for the northern part were done both for \(t_{\text{onset}} = 9,800\) years B.P. and \(t_{\text{onset}} = 14,000\) years B.P. The model predicted uplift rate profiles are calculated independently of the uplift rate profiles obtained from the InSAR data.

[40] The model uplift rates along the centerline of Mount Sedon are between 6.0 and 7.0 mm/yr (Figure 8). These rates are within the range of previously reported uplift rates based on geological considerations, and are in agreement with an average uplift rate obtained by previous and the present analysis of InSAR data [Baer et al., 2002b; Pe'eri et al., 2004]. A more detailed assessment of the model output is obtained by comparing the InSAR uplift rate profiles with that predicted by the model. The origin of the coordinate system of the analytical model is located along the channel centerline (Figure 6), whereas the InSAR data provide the uplift rates relative to the stable surface along the Ami’az Plain. Hence, to compare these two data sets, we subtract a constant \(v_t\) from the calculated uplift rate profiles \(v(x, t)\) in equation (9), and set zero uplift rate at the western edge of the mountain.

[41] Figure 8 shows representative InSAR profiles at the southern and northern part of the mountain, respectively, and the model-predicted uplift rates for those traverses. In general, the model profiles are in fairly good agreement with those of the InSAR data. Two drawbacks should be mentioned: (1) InSAR uplift rates at the northwestern margin of the diapir are 2–3 mm/yr higher than the predicted rates, and the asymmetry of the observed profiles is more pronounced than that predicted by the model (Figure 8); (2) InSAR uplift rates along the eastern margin of the diapir are occasionally less regular and usually higher than the predicted uplift rates.

6.3. Predicted and Observed Strain and Strain Rate

[42] Direct field measurements of shear strain are restricted to the eastern margin of the diapir. Zak and Freund [1980] measured brittle shear strain of 0.4 over a distance of 200 m from the outer eastern margin of Mount Sedon westward. Our model predicts shear strain along the same interval of 0.27–0.33, which is in reasonable agreement with the strain measurements. Consequently, the model-predicted shear strain rate at that interval is between \(5.6 \times 10^{-13}\) and \(6.8 \times 10^{-13} \text{ s}^{-1}\), which is comparable with the somewhat higher strain rate of \(9 \times 10^{-13} \text{ s}^{-1}\) based on Zak and Freund’s [1980] measurements. It should be noted that the latter estimate of strain rate averages over a time period of about 14,000 years, whereas the predicted strain rate is calculated 14,000 years after onset of the topography buildup. In that sense, the model-predicted strain rate is better compared to the shear strain rate of \(6 \times 10^{-13} \text{ s}^{-1}\) calculated from InSAR data for the eastern margin of the diapir.

7. Discussion

7.1. Strain Rate Variations During Uplift

[43] A wide range of strain rates has been obtained for natural rock salt flow based on studies of either pre-emergent or emergent diapirs. At the transition from pre-emergent to salt extrusion strain rates were estimated between \(10^{-14}\) and \(10^{-12} \text{ s}^{-1}\), but
Strain rate variations during the growth of Sedom diapir. Estimation of strain rates during the subsurface pre-emergent stage of the diapir is based on strain measurements of rotated boudins [after Zak and Freund, 1980]. Estimation of strain rates during the Holocene and currently emerging stage of the diapir is based on strain measurements of salt mirror offsets and InSAR data, respectively. The mechanical modeling results are for (1) the easternmost margin of the diapir and (2) 200 m from the easternmost margin of the diapir westward. Initiation time of upward salt flow and salt extrusion is based on interpretation of seismic profiles and field observations [Weinberger et al., 2006]. Strain rate variations due to possible emergence episodes before 14,000 years are not well constrained and therefore are not shown here.

![Graph showing strain rate variations](image)

**Figure 9.** Strain rate variations during the growth of Sedom diapir. Estimation of strain rates during the subsurface pre-emergent stage of the diapir is based on strain measurements of rotated boudins [after Zak and Freund, 1980]. Estimation of strain rates during the Holocene and currently emerging stage of the diapir is based on strain measurements of salt mirror offsets and InSAR data, respectively. The mechanical modeling results are for (1) the easternmost margin of the diapir and (2) 200 m from the easternmost margin of the diapir westward. Initiation time of upward salt flow and salt extrusion is based on interpretation of seismic profiles and field observations [Weinberger et al., 2006]. Strain rate variations due to possible emergence episodes before 14,000 years are not well constrained and therefore are not shown here.

Examples of field-based strain rates are rare [Talbot and Jackson, 1987]. In the present study, we have been able to document strain rate variations during the emplacement history of a single diapir (Figure 9). At the Plio-Pleistocene pre-emergent stage of the Sedom diapir, the Sedom rock salt flowed at a strain rate of $3 \times 10^{-14}$ s$^{-1}$, which is within the upper range of strain rates reported for other pre-emergent diapirs [Talbot and Jackson, 1987; Weinberg, 1993]. This strain rate is one order of magnitude lower than the strain rate of $5 \times 10^{-13}$–$9 \times 10^{-13}$ s$^{-1}$ obtained for the Holocene emerging stage of Sedom diapir. The Holocene increase in strain rate occurred after an episode of active diapirism, during which the Sedom diapir pierced the overburden and reached the surface. At that stage, the model predicts flow under the highest strain rate over the entire history of the diapir growth. At later stages of passive diapirism and salt extrusion, the strain rate decays due to reduction in the driving pressure (see equation (12)). This scenario is consistent with InSAR observations (Figure 9), which indicate that the current strain rate is similar to the model prediction and is lower than the average strain rate calculated for the Holocene based on displacement of the salt mirror.

### 7.2. Effective Viscosity of the Sedom Rock Salt

[44] The viscosity of $2.3 \times 10^{18}$ Pa s reported above has been obtained using a constant value of diapir width and overburden pressure for both parts of the mountain. While other variables used in our model are well constrained by observations and subsurface data (Table 1), the diapir width and overburden pressure are associated with errors that may change the value of the effective viscosity reported here. The diapir width may vary with depth as implied by traverse seismic lines thought the southern part of the mountain. Hence we varied the diapir width by ±10% and searched for the effective viscosity related to this range of widths. The estimated viscosity range falls between $1.8 \times 10^{18}$ and $3.0 \times 10^{18}$ Pa s. The present analysis neglects the possible increase in the overburden pressure due to continuous sedimentation during the last 14,000 years, inserting an error in the estimation of $\delta P$. Accounting for pressure variations associated with a relatively high sedimentation rate of 3 mm/yr during this time interval (i.e., about 1 MPa direct additional pressure to $P_{in}$ and maximum error in $\delta P$, related to 0.7 MPa additional pressure to $P_{in}$), we get variations of effective viscosity between $2.1 \times 10^{18}$ and $2.9 \times 10^{18}$ Pa s. Thus the errors in the diapir geometry and the driving pressure do not significantly change the estimated effective viscosity of the Sedom rock salt for strain rate between $5 \times 10^{-13}$ and $6 \times 10^{-13}$ s$^{-1}$, which is between 2 and $3 \times 10^{18}$ Pa s.

[45] van Keken et al. [1993] discussed two creep mechanisms of salt: transfer-diffusive and dislocation creep. They suggested that below a strain rate of $10^{-12}$ s$^{-1}$ the transfer-diffusive mechanism dominates the viscosity of fine-grained (crystal size < 5 mm) salt and the salt viscosity is independent of the strain rate. Hence it is likely that the fine-grained Sedom rock salt has been dominantly deformed by transfer-diffusive creep, and behaves as Newtonian fluid during emergence. This allows us to construct an analytical model describing viscosity-dependent topography evolution and poses constraints on the viscosities of the Sedom rock salt by fitting the model results to the present topography of the diapir. Comparison of our value to that derived numerically for fine-grained salt at room temperature [see van Keken et al., 1993, Figure 2] shows a remarkable agreement and
strengthens the applicability of the transfer-diffusive creep. Weinberg [1993] showed that salt with a viscosity of \(10^{17}\) Pa s should rise faster than 100 mm/yr, and that of \(10^{18}\) Pa s rises faster than 10 mm/yr to lift large inclusions of dense rock. The Sedom rock salt of \(2 \times 10^{18}\) Pa s, which rises faster than \(\sim 5\) mm/yr is comparable to the values calculated by Weinberg [1993], and may be suitable for most diapirs rising by a rate of several mm/yr.

### 7.3. Long-Term Versus Short-Term Uplift History

[46] The topography of Mount Sedom represents the averaged uplift history of the mountain since it breached the surface 14,000 years ago [Weinberger et al., 2006]. Likewise, we applied the mechanical model to predict this history of the mountain. On the other hand, the current InSAR observations represent the short-term uplift history of the mountain during the last decade. The InSAR uplift profiles are generally similar to topographic profiles obtained along the same traverses in Mount Sedom. The central and southern InSAR and topographic profiles are relatively more symmetrical and show a better similarity than the northern profiles (see below). These profiles are also in good agreement with synthetic profiles generated by the mechanical model. This similarity implies that the uplift history during the last 14,000 years is quite stable. This is confirmed by the characteristic timescale of growth \(\tau\) (i.e., “half-life” time, equation (8)), which is between 55,000 and 68,000 years, and is 4–5 times greater than \(t = 14,000\) years. This is comparable to the 42,000 years for extrusive lifetime estimated for Qum Kuh diapir in central Iran [Talbot and Aftabi, 2004].

[47] The similarity between the current topography and uplift patterns indicates a relatively low contribution of surface processes such as erosion, dissolution, or slope deposition. In some exceptional cases the uplift and topographic profiles show an apparent lateral shift (e.g., profiles e and h, Figure 5) indicating that the current uplift is different than the long-term time-averaged uplift presented by the topography. The higher uplift rate in the northern part of the mountain where the current elevation is lower could be explained by the suggestion that the emergence of the northern part of Mount Sedom lags by 3,000–4,000 years after the emergence of the southern part (see Frumkin [1996b] and the model predictions above).

[48] Several InSAR profiles (i.e., a, c, k, Figure 4) indicate that the country rocks east of the eastern margins of the diapir are currently subsiding. These observations are in agreement with the Pleistocene geological indication that the southern Dead Sea basin is subsiding at a rate of several mm/yr. Hence the shape of the diapir is also dictated by basin subsidence along its eastern margins.

### 7.4. Limitations on Predictions Based on the Model

[49] The present model provides a predictive tool for estimating the effective viscosity of Sedom rock salt and strain rate variations during the growth of the diapir based on the topography of Mount Sedom. The model initiation time is well-defined and is correlated with the beginning of salt breaching the surface and topography buildup 14,000 years ago. However, it is less clear how long after the present time the model prediction is still valid. Three drawbacks of the model are considered: (1) The rigid-wall assumption applied in the model poses constraints on the duration of the predicted time interval. The relatively simple wall-like shape of the Sedom diapir allows one-dimensional analysis of salt flow in a vertical rigid-wall channel. Because the basin is considered to be a pull-apart, the effects of deformation perpendicular to the transform are commonly neglected. Even if we hypothetically assume that the amount of W-E extension rate across the Dead Sea basin is as high as 5 mm/yr, the maximum possible extension of diapir walls over the last 14,000 years is smaller than 3.5%. This extension is less significant than the uncertainty related to the diapir width and can be neglected for this time interval. However, modeling the future growth of Sedom diapir is limited by the possibility that a large error in the diapir width will be accumulated over time. The extension creates accommodation space, which must be filled by lateral flow of salt that would otherwise contribute to vertical rise of the diapir. It seems likely that the rigid-wall assumption holds if the N-S trending normal faults in the western margin of the basin would be active for less than 20,000 years (i.e., error on diapir width < 100 m).

(2) The future growth of Sedom diapir is based on the fact that the modeled source salt layer is loaded by a constant pressure. Indeed, the additional pressure of sediments accumulated over the last 14,000 years is negligible compared to that exerted by the thick Plio-Pleistocene overburden. However, over time the error on the driving pressure will increase and may be significant for \(t > 40,000\) years.

(3) The model accounts for constant erosion that may well represent the geomorphological processes
that affected the diapir during the initial stages of its growth. In later stages of growth, it is likely that gravitational instability accompanied by sliding and rockfalls from the diapir head will be more significant than during the initial stages of growth. The present model cannot account for gravitational instability of the diapir during later stages of growth. We, therefore, limit our model to predict the topography of Mount Sedom to a time interval no longer than 20,000 years after salt breached the surface, that is, for less than an additional 6,000 years into the future.

[50] Along the western margin of the Sedom diapir the InSAR uplift profiles are somewhat steeper than the topographic profiles, and are significantly steeper than the synthetic profiles generated by the mechanical model. This discrepancy means that the current uplift and strain rates along the western margin are quite higher than predicted by the model. This can be explained by invoking several mechanisms. Irregularities and tectonically related changes in the channel geometry (Figure 2) might lead to deviation from the idealized constant-width geometry adopted in the analytical model. Local narrowing of the channel width should increase uplift rates and cause deviation from the Poiseuille flow. Distributed displacements along the stepped normal faults might pump excess salt at up-thrown blocks near the western margin of the diapir. This margin is also prone to influx of excess salt from the salt source layer in the Intermediate Block into the growing diapir (Figure 2). From a rheological point of view, the relatively high strain rates at the western margin of Mount Sedom could imply that the Bingham (e.g., power law) rheology is more suitable than the Newtonian rheology for this side of the diapir. The transition between these two behaviors seems to be associated with strain rates around \( 8 \times 10^{-13} \text{ s}^{-1} \), based on InSAR data from the western margin of Mount Sedom (Figure 8), in agreement with independent results of van Keken et al. [1993]. Support for this suggestion is the asymmetry in most uplift profiles with steep gradients and a relatively flat top on the western side of the diapir, and milder gradients on the eastern side (Figure 5). The above geometrical and rheological variations cannot be accounted for by the present analytical model and require numerical simulations, which are beyond the scope of this study.

8. Conclusions

[51] We present a simple mechanical model for the growth of an emerging salt diapir in a tectonically active basin. The model is based on one-dimensional flow of Newtonian viscous fluid (salt) in a vertical channel that has been driven by the load of the overburden and affected by the differential subsidence of the diapir’s flanks due to displacement along the marginal normal faults. It provides three sets of predicted profiles: topography, uplift rate, and shear strain. The model serves to constrain the effective salt viscosity and strain rates during the currently emerging stage of the Sedom diapir in the Dead Sea basin. The resulting Sedom rock salt viscosity is determined to be between \( 2–3 \times 10^{18} \text{ Pa s} \), and the associated strain rate is between \( 5–6 \times 10^{-13} \text{ s}^{-1} \). Geological structures indicate a strain rate of \( 3 \times 10^{-12} \text{ s}^{-1} \) during the Plio-Pleistocene pre-emergent stage of the Sedom diapir, which is one order of magnitude lower than that associated with the currently emerging stage of the diapir growth. The uplift history of the mountain predicted by the model and the current topography are compared to InSAR measurements of salt uplift. The maximum uplift rate from the InSAR data of Mount Sedom is 8.3 mm/yr. The InSAR uplift profiles resemble topographic profiles obtained along the same traverses. This similarity implies that the uplift history during the last 14,000 years has been stable.

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