Relationships between lake-level changes and water and salt budgets in the Dead Sea during extreme aridities in the Eastern Mediterranean

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Abstract

Thick halite intervals recovered by the Dead Sea Deep Drilling Project cores show evidence for severely arid climatic conditions in the eastern Mediterranean during the last three interglacials. In particular, the core interval corresponding to the peak of the last interglacial (Marine Isotope Stage 5e or MIS 5e) contains ~30 m of salt over 85 m of core length, making this the driest known period in that region during the late Quaternary. This study reconstructs Dead Sea lake levels during the salt deposition intervals, based on water and salt budgets derived from the Dead Sea brine composition and amount of salt in the core. Modern water and salt budgets indicate that halite precipitates only during declining lake levels, while the amount of

dissolved Na^+ and Cl^- accumulates during wetter intervals. Based on the compositions of Dead Sea brines from pore waters and halite fluid inclusions, we estimate that ~12-16 cm of halite precipitated per meter of lake-level drop. During periods of halite precipitation, the Mg^{2+} concentration increases and the Na^+/Cl^- ratio decreases in the lake. Our calculations indicate major lake-level drops of ~170 m from lake levels of 320 and 310 m below sea level (mbsl) down to lake levels of ~490 and ~480 mbsl, during MIS 5e and the Holocene, respectively. These lake levels are much lower than typical interglacial lake levels of around 400 mbsl. These lake-level drops occurred as a result of major decreases in average fresh water runoff, to ~40% of the modern value (pre-1964, before major fresh water diversions), reflecting severe droughts during which annual precipitation in Jerusalem was lower than 350 mm/y, compared to ~600 mm/y today. Nevertheless, even during salt intervals, the changes in halite facies and the occurrence of alternating periods of halite and detritus in the Dead Sea core stratigraphy reflect fluctuations between drier and wetter conditions around our estimated average. The halite intervals include periods that are richer and poorer in halite, indicating (based on the sedimentation rate) that severe dry conditions with water availability as low as ~20% of present day, continued for periods of decades to centuries, and fluctuated with wetter conditions that spanned centuries to millennia when water availability was ~50-100% of present day. These conclusions have potential implications for the coming decades, as climate models predict greater aridity in the region.

1 Introduction

In many regions of the world droughts are a major concern, affecting agriculture, industry and everyday life. In drought sensitive regions, such as the Middle East, water scarcity impacts political stability. Climate models and observations show a drying trend in the lands around the

47 Mediterranean, reflecting both natural variability and increased anthropogenic greenhouse gas
48 concentrations and predict up to 20% decreases in water availability by the end of the 21st
49 century (Kelley et al., 2012; Lelieveld et al., 2012; Mariotti et al., 2008; Seager et al., 2014).
50 While aridification of the Mediterranean region might be explained by poleward expansion of
51 the Hadley circulation and associated widening of the subtropical dry zone (Frierson et al.,
52 2007), the mechanism and strength of the Mediterranean climate response to increasing
53 greenhouse gases is not well-understood (e.g. Seager et al., 2014). In the Eastern Mediterranean
54 region, low lake levels in the Dead Sea have been shown to reflect increases in drought
55 frequency that are also associated with higher annual temperatures and northward migrations of
56 storm tracks (Enzel et al., 2003).

57 The climate during the Holocene and MIS 5e varied both spatially and temporally in the
58 Mediterranean, Europe, North Africa and the Levant (Bar-Matthews, 2014; Damnati, 2000;
59 deMenocal et al., 2000; Harrison et al., 1993; Kushnir and Stein, 2010; Mayewski et al., 2004).
60 The Levant region, comprising Israel, Jordan, Palestine, and Syria, experienced much drier
61 conditions than today, as indicated by halite precipitation in the Dead Sea (Kiro et al., 2016;
62 Neugebauer et al., 2014; Torfstein et al., 2015). Yet, during the last interglacial peak (MIS 5e,
63 ~130-115 ka), which was characterized by stronger insolation cycles, higher average global
64 temperatures, higher sea levels and smaller continental ice sheets than the Holocene (Dutton and
65 Lambeck, 2012; e.g. Felis et al., 2004; Govin et al., 2015; Kukla et al., 2002; Otto-Bliesner et al.,
66 2013; Rohling et al., 2002), paleoclimate records from the eastern Mediterranean document
67 extended intervals of relatively wet conditions, especially during the peak insolation interval of
68 MIS 5e at around ~125 ka BP (Bar-Matthews, 2014; Bar-Matthews et al., 2003; Torfstein et al.,
69 2015; Vaks et al., 2007). This wet interval was associated with the tropical northern hemisphere

summer insolation peak, which forced intense African summer monsoon rainfall that resulted in increased Nile flow, and the formation of a major sapropel (S5) in the Eastern Mediterranean (Rohling et al., 2002; Rossignol-Strick, 1985; Torfstein et al., 2015; Vaks et al., 2007; Waldmann et al., 2010). Nevertheless, during the past 140 kyr in the Levant, the main source of precipitation moisture has primarily been the Mediterranean Sea (Kolodny et al., 2005; McGarry et al., 2004; Torfstein et al., 2015). Sorting out the hydrological variability in the Levant during the last interglacial and its relation to orbital and global climate variations remains a challenge.

The Dead Sea watershed straddles both the Mediterranean climate zone and the Sahara-Arabian desert belt, reflecting the interplay between these two climate belts. Considering that the effects of anthropogenic warming on the semi-arid and Mediterranean climate is a concern for the future, it is important to know what has occurred in the past, and sedimentary archives provide a means to learn about the severity, duration and frequency of past droughts. The ICDP Dead Sea Deep Drilling Project (DSDDP) recovered such a record from the deep floor of the Dead Sea (Figure 1) during 2010-11 (Neugebauer et al., 2014; Stein et al., 2011). The drill site was 297 m below the lake surface, and cores were recovered to a depth of 456 m below the lake floor (mblf), covering ~200 ka (Figure 2 and Torfstein et al., 2015), thus including the Holocene and MIS 5e.

The age model used in this study is constructed mainly after Torfstein et al. (2015), based on U-series dating of aragonite, the lithology of the core, and correlation of aragonite $\delta^{18}\text{O}$ in the DSDDP core with $\delta^{18}\text{O}$ in the Soreq cave (Bar-Matthews et al., 2003), the marine benthic $\delta^{18}\text{O}$ LR04 stack (Lisiecki and Raymo, 2005) and Mediterranean core records (e.g. Wang et al., 2010). For the last glacial maximum to the present-day, the age model is based on the ^{14}C data of Kitagawa et al. (2016). Torfstein et al. (2015) discussed the general paleohydrology of the

region, based on the lithology of the core, and showed that the last interglacial was characterized by a high degree of climate variability, with a dry interval marked by halite precipitation during the MIS 6/5 transition, followed by wetter conditions during the peak of MIS 5e without halite precipitation, in turn followed by an extreme dry interval characterized by a thick accumulation of halite.

Here we develop a salt and water budget based on halite fluid inclusions and pore water chemical compositions, together with halite thicknesses from the DSDDP cores. Based on these considerations, we calculate the lake-level changes during the driest intervals in this region and estimate the discharge of water and the amount of precipitation. Our results show that the Levant has experienced severe dry conditions lasting up to a few thousand years during both the Holocene and MIS 5e.

2 Geological and hydrological settings

During the late Quaternary, the Dead Sea basin, the lowest elevation on the continents (the deepest Dead Sea lake floor is >720 mbsl, while the 2016 lake level is at 431 mbsl), was occupied by a succession of lakes that varied in their compositions and limnological configurations as a result of the changing hydro-climatic conditions in the watershed (Neev and Emery, 1995; Stein, 2001). The modern Dead Sea brine originated from seawater that penetrated the Dead Sea Valley during the late Neogene, forming the Sedom Lagoon (Zak, 1967). The evaporated seawater interacted with Cretaceous limestone surrounding the Valley and produced the Ca-chloride brine that deposited >2000 m of salt in the subsiding Dead Sea basin (Starinsky, 1974; Stein et al., 2000; Zak, 1967). After the Sedom Lagoon became disconnected from the open sea at ~3 Ma (Matmon et al., 2014; Torfstein et al., 2009), the water-body occupying the Dead Sea Basin became a terminal lake, that was continuously recharged by freshwater from

runoff and by saline springs (García-Veigas et al., 2009; Stein et al., 1997; Torfstein et al., 2008) precipitating mainly aragonite, gypsum and halite (e.g. Stein, 2014). The lakes received water, dissolved salts and detritus via runoff from the Jordan and Arava Rivers, smaller rivers, and springs (Figure 1). The minerals aragonite, gypsum, and halite precipitated, depending on the amount of freshwater inflow and how it mixed with the Dead Sea brine (Katz and Starinsky, 2008; Stein et al., 1997; Torfstein et al., 2008).

Over the late Quaternary, Dead Sea water levels have fluctuated with glacial cycles and has been higher during glacials and lower during interglacials (Bartov et al., 2003, 2002; Stein et al., 2010; Torfstein et al., 2013; Waldmann et al., 2009). For example, over much of the last ice age the lake level fluctuated between ~300 and ~250 mbsl, and it rose to ~170 mbsl during the last glacial maximum high-stand (~27-24 ka), when it expanded over the present-day Jordan River Valley to north of the Sea of Galilee (Bartov et al., 2002; Hazan et al., 2005; Torfstein et al., 2013). These fluctuations are reflected in the sediment stratigraphy (Figure 2). The sediments deposited during glacials are mainly primary aragonite and silty-detritus layers of a few mm in thickness that form annual pairs of the '*aad*' facies (Machlus et al., 2000; Prasad et al., 2004) and reflect significant runoff and addition of bicarbonate to the lake. During interglacials, the lake levels are lower, typically at ~400 mbsl (e.g. the late Pleistocene Lake Samra and the Holocene Dead Sea), and sediments deposited are mostly sequences of silty-detritus (mainly calcite and quartz grains that comprise the *laminated detritus* or '*ld*' facies (Haliva-Cohen et al., 2012). Gypsum-rich sequences were deposited during both interglacials and glacials (Kiro et al., 2016; Migowski et al., 2006, 2004; Neugebauer et al., 2015). They indicate dry intervals associated with abrupt lake-level drops, and reflect CaSO₄ saturation following build-up of sulfate in the Ca-chloride brine (Stein et al., 1997; Torfstein et al., 2008, 2005). Halite is deposited only during

interglacials and reflects extreme dryness (Figure 2).

Halite thus indicates high evaporation. It has not been recovered from the marginal terraces, with the exception of an uplifted and tilted late Quaternary halite sequence on the eastern flank of the Mount Sedom salt-diapir (Figure 1) dated by U-Th to ~400 ka (possibly MIS 11, Torfstein et al., 2009). With the exception of this halite layer, and modern and Holocene-age deposits along the present-day (human-driven) retreating shore of the Dead Sea, halite deposits over the late Quaternary has been recovered only by drilling. A thick sequence of halite (~10-20 m) was recovered by drilling to the base of the Holocene on the Dead Sea margin (Migowski et al., 2004; Yechieli et al., 1993), which is dated by ^{14}C at 11-10 ka BP (Stein et al., 2010). Layers of salt have also been recovered by drilling into the Holocene section in the southern basin of the Dead Sea, which was dry during various time intervals (Neev and Emery, 1995).

The present-day Dead Sea is one of the most saline lakes in the world, with a salt content of ~340 g/L (e.g. Katz and Starinsky, 2008), nearly an order of magnitude higher than seawater. Lake-level fluctuations in the Dead Sea are a function of water discharge into the lake, which is directly related to the regional precipitation (Enzel et al., 2003), because precipitation in the watershed varies more than evaporation from the lake surface (e.g. Lensky et al., 2005; Stanhill, 1994). The volume of evaporating water decreases with declining lake level, due to the decrease in the surface area and lower vapor pressures associated with increases in salinity (Krumgalz et al., 2000; Yechieli et al., 1998), while the volume evaporated increases with increased brine temperature and wind speed (Lensky et al., 2005; Rohling, 2013). The modern Dead Sea evaporation rate, based on energy and mass balances, is 1.05-1.30 m/y (Lensky et al., 2005; Stanhill, 1994; Yechieli et al., 1998), whereas during intervals of greater wetness and lower surface salinity rates may be as high as 1.6 m/y (Stanhill, 1994). During the last glacial period

the evaporation rate has been estimated to be considerably higher, between 1.9 and 2.6 m/y (Rohling, 2013). Evaporation rates are expected to decrease in the future due to increasing salinity, for example, down to ~0.7 m/y when the salinity reaches ~460 g/L (Yeichieli et al., 1998).

Since the 1960s the freshwater input has decreased significantly due to diversion of the Jordan River and large-scale utilization of freshwater from the Dead Sea watershed. The total amount of freshwater entering the Dead Sea today is only ~300 million m³/y, compared with ~1600-2000 million m³/y in the 1950's (Lensky et al., 2005). Together with enhanced evaporation of the brine by potash salt production by Israel and Jordan, the lake level is currently declining at a rate of more than ~1 m/y. In this study, all estimates of water and salt budgets are relative to the natural (i.e., pre-1964) characteristics of the Dead Sea.

3 Methods

Our model is based on the chemical compositions of major elements in the primary fluid inclusions in the halite, and in pore fluids (Levy et al., 2016). Analytical details are in Appendix A, and data are listed in Appendix B. Calculations of brine evolution and halite accumulation rates used the USGS PHREEQC geochemical computer program (Parkurst and Appelo, 2013), using the Pitzer database for calculating ion activity coefficients and water activity in high ionic strength solution according to the Pitzer equations (Pitzer, 1973) in 25°C. Appendix C lists input and output parameters.

4 Halite distribution in the Dead Sea core

The halite intervals in the DSDDP core occur only during interglacials (MIS 1,5,7, Figure 2, and Table 1), when the lake levels were relatively low (Torfstein et al., 2015). The thickest halite

interval occurs during the MIS 6/5 deglaciation and MIS 5e, between ~135 and 116 ka, which contains ~30 m of recovered halite over ~85 m of drilling (~330-235 m depth in the core). Most of the recovered halite (24.5 m over 45 m of drilling at 280-235 m depth in the core) precipitated during the waning stages of MIS 5e between ~122-116 ka. The peak northern hemisphere insolation interval at ~128-122 ka (Figure 2) is characterized by a ~23 m thick halite-free detritus layer (303-280 m depth in the core), indicating a wetter period (Torfstein et al., 2015). The Holocene halite (89.25-0 m depth in core) is distributed more evenly over time compared with the MIS 5e halite and contains 33.8 m of halite over 89.25 m of drilling.

5 Water and solute budgets

In Section 5.1 we first outline the basis for our conclusion that halite precipitation reflects lake-level drops and negative water budgets in the lake. In Section 5.2 we describe in the chemical evolution and precipitation of halite during lake-level drops. Section 5.3 outlines our conceptual model of two stages of halite precipitation, based on the data from the DSDDP core and the changes in the bathymetry of the lake. In Section 5.4 we outline our quantification of the lake-level changes based on the trends of Mg^{2+} and Na^+ in the core, taking into account initial and final concentrations and the thickness of halite. Section 5.5 deals with the fluctuations between wet and dry periods based on the alternating halite and detritus.

5.1 Na, Cl and Mg budgets

The Na^+ and Cl^- budgets in a saline water body are dictated by their removal due to halite precipitation, and contributions from other sources such as fresh water flows and brine discharge. When the Na^+/Cl^- molar ratio in a water body is <1 (the case for Dead Sea brine and for marine and most lacustrine environments), the Na^+/Cl^- ratio decreases during halite precipitation (Figure

3), and the amount of halite that can precipitate is dictated mainly by the amount of Na. However, dissolution of halite in undersaturated brine adds equal molar amounts of Na and Cl to the brine, increasing the Na^+/Cl^- ratio. At the same time, Mg^{2+} is nearly conservative, as long as Mg-salts are undersaturated, and therefore its concentration decreases during lake-level rise and increases during lake-level drop (Levy et al., 2016). On a long time scale (longer than the 200 kyr represented in the DSDDP core), the Mg^{2+} concentration (hereafter $[\text{Mg}^{2+}]$) in the terminal lake should increase due to prolonged input with freshwater (and no precipitation of Mg-salts). However, on the shorter time scales such as the past 200 kyr, lake-level changes have much larger impact on $[\text{Mg}^{2+}]$ than the input flux. For example, in the present-day Dead Sea, for a lake-level decline of 20 cm in a year, the amount of Mg^{+2} within the evaporated volume is $\sim 10^{13}$ g, and this amount is added to the residual brine, while the Mg^{+2} flux is nearly two orders of magnitude lower at $\sim 2 \cdot 10^{11}$ g/y (Table 2). Thus, to a first approximation, the $[\text{Mg}^{2+}]$ is related directly to lake volume (Figure 4). We note that $[\text{Mg}^{2+}]$ in pore waters reflects the bottom water body during times of a stratified water column. During halite intervals, when the water column of the lake is generally well-mixed, the $[\text{Mg}^{2+}]$ -volume relationship is direct. During wetter intervals the relationship is affected by water column stratification, and there is probably a delay in the response of $[\text{Mg}^{2+}]$ at the bottom of the lake to the volume change.

Here we estimate the water discharge into the Dead Sea, applying salt and water mass balances. Table 2 summarizes the compositions of various water sources in the Dead Sea watershed. Among these sources the potential water sources contributing Na^+ and Cl^- into the modern Dead Sea (prior to the diversion of the Jordan River during the 1960s) are: brine springs, the Jordan River north of Sea of Galilee and near the Dead Sea, and contributions from Sedom Formation halite.

These values are used to calculate the *potential accumulation rate of halite* on the lake's floor resulting from the annual input flux from these sources (bottom row in Table 2). The mass balance of water and salt translates into halite thickness by the equation:

$$H_R = \frac{\sum_i F_i [Na]_i \cdot \mu}{\rho \cdot S} \quad (1)$$

where H_R [m/y] is the potential accumulation rate of halite, $[Na]_i$ [mol·L⁻¹] is the concentration of Na in a water source i with a flux of F_i [m³/y], μ =58.433 g·mol⁻¹ and ρ =2170 g·L⁻¹ are the molecular weight and density of halite, and S [m²] (=600-700·10⁶ m²) is the Dead Sea surface area. The halite in the Dead Sea core is compacted, with no pores (Kiro et al., 2016), and thus its density is taken as the density of halite. Substituting the values from Table 2 into Eq. 1 yields a potential halite accumulation rate of ~1 mm/y from the pre-1964 input, with brine seepage and the Jordan River as the major Na⁺ and Cl⁻ sources.

The estimated potential halite accumulation rate from the annual input (~1 mm/y) is an order of magnitude lower than the rate of halite deposition of 1-2 cm/y estimated in the DSDDP core (Neugebauer et al., 2014; Torfstein et al., 2015). *Thus, the halite that precipitated in the Dead Sea basin during the last 220 kyr must have originated from evaporation of the brine itself, which in turns means that the halite layers signify declining lake volume.*

However, there is a limit to the amount of halite that can be precipitated by simple evaporation of the Dead Sea brine. For example, a total halite thickness of ~11 m would precipitate by evaporation of the total volume of the modern Dead Sea (Krumgalz et al., 2002). The total amount of halite in the DSDDP core during the last 220 kyr is ~79 m (Table 1). This requires addition of significant salt to the lake. As an example, if we evaporate the water volume

of $V \sim 190 \cdot 10^9 \text{ m}^3$, corresponding to a lake level of $\sim 350 \text{ mbsl}$ when halite starts to precipitate (further discussed below), and if we consider a starting brine composition with a very high potential to precipitate large quantities of halite, such as an $[\text{Na}] = 4.1 \text{ mol} \cdot \text{L}^{-1}$ (Figure 3) along with a high Na^+/Cl^- ratio, then we would still need at least twice as much Na^+ in order to precipitate the observed 79 m of halite ($T = \frac{[\text{Na}]V\mu}{\rho S} \sim 32 \text{ m}$ where $T[\text{m}]$ is the thickness of halite, $\mu = 58.433 \text{ g} \cdot \text{mol}^{-1}$ is the halite molecular weight, $\rho = 2.17 \cdot 10^6 \text{ g} \cdot \text{m}^{-3}$ is the halite density and $S = 650 \cdot 10^6 \text{ m}^2$ is the surface area of the lake). Thus, a significant additional supply of Na^+ and Cl^- into the lake is required to precipitate the observed thickness of halite in the DSDDP core. The identification and quantification of the additional source is discussed below.

The compositions of Dead Sea brines, as shown in halite fluid inclusions and pore waters, lie on an evaporation curve of the fluid inclusions with the least $[\text{Mg}^{2+}]$ (Figure 5). This means that the chemical composition in the brine during the late Quaternary changed mainly due to precipitation and dissolution of evaporite minerals. During the last glacial period (and possibly also MIS 6), dissolution of the Mt. Sedom salt-diapir (Weinberger et al., 2007; Zak, 1967) was a potential source of dissolved salts into the lake (Figure 1).

The dissolved Na^+ and Cl^- accumulate in the lake when its level and freshwater runoff are high, particularly during glacials, and are removed through halite precipitation when the lake-level drops during interglacials. Thus, during lake-level drops and halite precipitation, any increases in the element concentrations due to fresh water input are negligible compared with decreases due to precipitation of salts as a result of evaporation, for example, Na removal due to a halite precipitation rate of 20 cm/y is $\sim 10^{13} \text{ g/y}$, while $\sim 10^{10} \text{ g/y}$ is added by input from the Jordan River.

5.2 Brine evolution and halite accumulation during evaporation

During lake-level drops and halite precipitation, $[\text{Mg}^{2+}]$ increases with the decrease in the lake volume, while $[\text{Na}^+]$ and Na^+/Cl^- decrease, and $[\text{Cl}^-]$ increases, due to halite precipitation (Figure 3a). The main constraint on halite precipitation is $[\text{Na}^+]$ and therefore the halite accumulation rate decreases as $[\text{Na}^+]$ decreases in the brine.

The thickness of halite that accumulates per meter of lake-level drop, assuming a brine saturated with halite, can be calculated according to

$$D = \frac{dH}{dW} \cdot \frac{\mu}{\rho} \quad (2)$$

where D [m-halite/m-lake-level-drop] is the halite accumulation rate, H [mol] is the amount of precipitated halite, and W [L] is the amount of water loss by evaporation, dH/dW is the rate of halite precipitation per amount of water evaporated, $\mu=58.433 \text{ g}\cdot\text{mol}^{-1}$ is the molecular weight of halite, and $\rho=2170 \text{ g}\cdot\text{L}^{-1}$ is the halite density.

Because the Na^+/Cl^- of the brine varies with salt precipitation and addition (Figure 5), the Na^+/Cl^- ratio of pore waters and fluid inclusions in halite layers in the core indicates the rate that halite is accumulating (D [m-halite/m-lake-level-drop]) during that time interval (Figure 3). In the DSDDP core, the range of Na^+/Cl^- ratios is 0.2-0.6 (Figure 5) indicating that the accumulation rate of halite did not vary much, between ~12.5 and 16 cm per meter of lake-level drop (Figure 3).

5.3 Conceptual model of halite precipitation and lake levels

During halite precipitation, $[\text{Mg}^{2+}]$ increases and the Na^+/Cl^- ratio decreases (Figure 3). Because Mg^{2+} behaves (to a first order) as a conservative ion (that is, its total amount in the lake

remains constant), and its concentration is directly related to the lake volume, for any initial lake volume V_i [m³], the final volume of the lake, V_f [m³], is described by:

$$V_f = \frac{[Mg]_i}{[Mg]_f} \cdot V_i \quad (3)$$

where $[Mg]$ [mol·L⁻¹] is the Mg²⁺ concentration and subscripts i and f indicate initial and final, respectively. This relationship between Mg²⁺ and lake volume is directly related to the initial lake surface S_i [m²] and level L_i [m] via the hypsometric curve of the modern Dead Sea (Figure 4). According to Eq. 3, the increase in Mg²⁺ from 0.7 to 1.8 mol·L⁻¹ during MIS 5e, and from 0.6 to 1.7 mol·L⁻¹ during the Holocene (Figure 6, the $[Mg^{2+}]$ units in the figure are in molality) is equivalent to lake volume decreases of ~40% during both time periods (Table 3). Figure 4 shows that the increase of the Mg²⁺ during MIS 5e corresponds to a lake-level change from ~320 to 490 mbsl, (where the initial $[Mg^{2+}]$ and lake level are those at the beginning of MIS 5e).

Comparison of the amount of halite precipitated with $[Mg^{2+}]$ in the DSDDP core reveals that there are two stages (illustrated by the MIS 5e data, Figure 7). During Stage 1 there is a large increase in Mg²⁺ with little halite (from 0.7 to 1.6 mol·L⁻¹ over a 4.5 m thick halite layer), while during Stage 2, a lot of halite precipitates with little increase in Mg²⁺ (from 1.6 to 1.8 mol·L⁻¹ over a 28.9 m thick halite layer). In stark contrast to the 28.9 m of halite observed during Stage 2, the calculated thickness of halite expected from the changes in Mg²⁺ and Na⁺ is only ~4 m. This indicates that the decreases in lake volume were not the sole factor determining the amount of halite that precipitated in the middle of the lake (details are in Section 5.4, and relationships between halite precipitation, Na⁺ and Mg²⁺ are described by Eqs. 4, 5).

The two stages of halite precipitation (Figure 7) are related directly to the lake's hypsometric curve (Figure 4), which shows an inflection point at ~400 mbsl, reflecting important changes in

the geometry of the basin, whereby the slope of the margins is much steeper below that elevation. As a result, at lake levels higher than 400 mbsl, for each meter of lake-level change there are much larger changes in volume and surface area, than at lake levels below 400 mbsl.

The basin geometry has important impacts on halite deposition at the lake margins versus the middle (Figure 8). During Stage 1, when lake levels are above 400 mbsl, thicker layers of halite are deposited at the lake margins than in the center. For example, along the margins of the Holocene Dead Sea there is a ~15 m thick halite layer that precipitated between ~10.7 and 10.2 ka (Stein et al., 2010), while only ~7.7 m are found in the deep DSDDP core. However, halite is generally absent in the marginal sections of the Dead Sea, because the marginal salt is re-dissolved and added back to the lake brine, to be re-precipitated in the lake center. As a result, for each meter of lake-level drop during Stage 2 there is more halite precipitated in the center of the basin. Therefore, the halite recovered from the DSDDP core, at the depocenter of the lake, comprises most of the halite that deposited in the entire basin. These relationships, that is, the changes in chemical composition and the total thickness of halite, can be used to calculate changes in lake levels. In Section 5.4 below we outline our method to quantify lake-level changes based on the initial and final concentrations of Mg^{2+} and Na^+ assuming that all the halite is re-precipitated in the lake center.

5.4 Reconstructing lake-level changes and average fresh water runoff

The lake-level changes are calculated using the halite thicknesses, the Mg^{2+} and Na^+ concentrations in the lake brine (Table 3), and the lake's hypsometric curve (Figure 4), through a halite interval in the DSDDP core, such as MIS 5e or the Holocene (Table 1). The use of the hypsometric curve yields a single solution for initial and final lake levels for each of the halite intervals (Figures 9, 10).

The model calculates the initial and final lake levels as described in a flow chart (Figure 9). Given an initial volume V_i [m³], the final lake surface area S_f [m²], lake level L_f [m], and volume V_f [m³] are all related by Eq. 3 and the Dead Sea hypsometric curve. The amount of halite H [mol] that is precipitated can be calculated from the change in $[Na^+]$ (Table 3) by

$$H = \left([Na]_i \frac{[Mg]_f}{[Mg]_i} - [Na]_f \right) V_f \quad (4)$$

where $[Na]$ [mol·m⁻³] is the Na⁺ concentration. Assuming an evenly distributed halite layer precipitates over the final lake surface (Section 5.3), the thickness of halite T [m] is a function of the surface area, and can be calculated by:

$$T = \frac{H}{\rho \cdot S_f} \cdot \mu = \frac{\left([Na]_i \frac{[Mg]_f}{[Mg]_i} - [Na]_f \right) V_f}{\rho \cdot S_f} \cdot \mu \quad (5)$$

for any given initial volume V_i [m³], where V_f [m³] is calculated by Eq. 3. S_f [m²] is the final surface area corresponding to V_f according to the hypsometric curve (Figure 4), and $\mu=58.433$ g·mol⁻¹ and $\rho=2.17 \cdot 10^6$ g·m⁻³ are the molecular weight and density of halite, respectively. Uncertainties regarding these assumptions are discussed in Appendix D. Thus *the thickness of halite deposited can be calculated for any given initial lake-level elevation* (Figures 9, 10).

Lake levels versus halite thickness calculated for the Holocene and MIS 5e periods are plotted in Figure 10, which shows this relationship for a *range* of initial lake-level elevations. The calculation indicates that in order to precipitate the amount of halite observed in the DSDDP core, the lake level dropped by ~170m during both the Holocene and MIS 5e, from 308 mbsl and 320 mbsl at the end of the MIS 2 and MIS 6 glacials, to lake levels of 475 and 490 mbsl, respectively (Table 4). We note that the thickness of Holocene halite takes into account 7.5 m of non-dissolved marginal halite (Section 5.3 and Table 3), while this is not the case for the MIS 5e

estimate because there is no evidence for non-dissolved marginal halite from that time in the DSDDP cores from the near-shore sites (Neugebauer et al., 2015).

In order to reach these low lake levels, the average water discharge into the lake would have to be very low. Plotting the discharge to the lake as a function of the lake-level (Figure 11a) as calculated from the evaporation rate of 1.1-1.3 m/y (Lensky et al., 2005; Stanhill, 1994) and the hypsometric curve (Figure 4) indicates that during the interglacials, the annual discharge was just ~40% of the modern (pre-1964) value.

The modern multi-year averaged precipitation over the Dead Sea watershed region has been correlated with the instrumental rainfall record in Jerusalem for the years 1870-1960, and with other stations in the Levant region (Enzel et al., 2003). This is the longest existing southern Levant rainfall record. Using the modern Jerusalem rainfall record, a discharge of 40% of the modern (pre-1964) value into the Dead Sea ($\sim 600 \cdot 10^6 \text{ m}^3/\text{y}$) during the Holocene and MIS 5e salt intervals implies that the average rainfall in Jerusalem decreased to ~350 mm/y, which is ~60% of the modern average rainfall of $605 \pm 195 \text{ mm/y}$ (average and standard deviation of precipitation in Jerusalem between 1865-1995; Israel Meteorological Service, Table 4, Figure 11b). On decadal time scales, the Jerusalem record varies in concert with less continuous precipitation records from Beirut and Damascus in the north to Beer Sheva in the south (Enzel et al., 2003; Kushnir and Stein, 2010). Therefore we assume that the changes in Jerusalem are representative of the whole region. Uncertainties of these estimates are discussed in Appendix D.

The Levant region sporadically experiences years with low precipitation and droughts, and the annual rainfall varies significantly from year to year. According to the instrumental rain record, clusters of years with only ~300 mm/y in Jerusalem occur occasionally and are considered very dry intervals. According to modern observations, between 1864 and 1995

rainfall lower than 400 mm/y occurred on average every 8 years. Two nearly decade-long, historical dry intervals centered on 1930 and 1960 (Kushnir and Stein, 2010). Based on the modern instrumental data, lake-level declines were correlated to time intervals when 50% of the years experienced less than 450 mm/y (Enzel et al., 2003). Based on the modern data, Enzel et al. (2003) concluded that low lake levels during the past 4000 years reflect long periods when the annual rainfall fell frequently below the modern average. Using the modern comparison between lake levels and regional annual precipitation to infer the past, our results thus indicate that during the interglacial salt intervals, average precipitation was extremely low, with high frequency of years with severe droughts during which annual precipitation was much lower than 350 mm/y.

5.5 Fluctuations between wetter and drier intervals

Despite the overall extremely arid conditions that prevailed in the Eastern Mediterranean during MIS 5e and early Holocene, the salt intervals in the DSDDP core show alternations between layers of halite and detritus (Figure 2) indicating fluctuations between drier and wetter intervals. This means that the dry intervals were punctuated with periods of wetness and increasing lake levels even though overall the lake level was declining. Within the core, different halite facies also reflect changes between dryer and wetter conditions, often showing seasonal changes represented by different types of crystals (Kiro et al., 2016). The halite layers contain two main crystal types, large crystals that precipitate at the lake bottom, and small cumulate crystals that precipitate at the lake surface. The large crystals reflect wetter conditions than the small cumulate crystals, and they are associated with more detritus and show effects of mild dissolution (Kiro et al., 2016). Frequent alternations between these two crystal types are seasonal, with the large “bottom-growth” crystals forming during winter and the small cumulates during summer (Kiro et al., 2016). The seasonal halite layers range in thickness between ~0.5-2

400 cm implying sedimentation rates between ~1-4 cm/year, while a typical sedimentation rate is ~2
401 cm/y.

402 The average lake-level drop during the whole MIS 5e halite interval (between 328 and 234 m
403 in the DSDDP core) is ~1 cm/y, taking into account detritus-rich layers. However, during halite
404 rich intervals rate of lake-level drop was much higher. Based on estimates of the halite
405 sedimentation rates, and 12-16 cm of halite precipitated per meter of lake-level drop (Figure 3),
406 we conclude that the lake-level drop rate during halite-rich intervals was between 6 and 33 cm/y
407 ($L_r = S_r / D$ where L_r [cm/y] is the lake-level drop rate, S_r [m/y] is the sedimentation rate and D
408 [cm/m] is the thickness of halite precipitated per meter lake-level drop). Based on this rate of
409 lake-level drop, an evaporation rate of 1.1 m/y, and a surface area of 560 km², corresponding to
410 a lake level of 450 mbsl (Figure 4), we calculate that the most arid intervals, reflected by the
411 thick cumulate layers of halite, are characterized by an annual discharge of only ~430 million
412 m³/y ((1.1-0.33)·560×10⁶), which is ~20% of the modern (pre-1964) value. A caveat to this
413 calculation is that the distribution of halite might not be homogeneous beneath the lake surface.
414 As discussed in Section 5.3, at least during the Stage 1 of halite precipitation (reflected by rapid
415 changes in [Mg²⁺] in the Dead Sea brine, Figure 7), the halite layers are thicker on the lake
416 margins compared with the center. In this case, the estimated discharge would be even lower.

417 The halite thicknesses indicate that the most arid intervals lasted for decades to centuries: in
418 between >1 m intervals of halite-free detritus, the average thickness of halite layers is ~1.7±1.4
419 m (1SD). These extreme arid conditions fluctuated with milder conditions, as reflected by the
420 detritus intervals that indicate increases in the lake level lasting up to a few thousand years.
421 During the peak of MIS 5e, a 23 meter thick detritus layer indicates a wet period that lasted
422 several thousands of years (~128-122 ka, at ~303-280 m depth in the DSDDP core, Torfstein et

al., 2015). This was coeval with a major sapropel event in the Eastern Mediterranean (the ‘S5’ event; e.g. Cane et al., 2002; Rohling et al., 2002; Rossignol-Strick, 1985), when deep water in the Mediterranean became anoxic as a result of a major increase in the Nile flux. During this period in the Dead Sea, $[Mg^{2+}]$ decreased from ~ 1.68 to ~ 1.27 mol·kgH₂O⁻¹ (Figure 6). This indicates that the lake level increased from ~ 470 mbsl to at least 411 mbsl (Figures 4, 7; the lake level at the end of Stage 1 of halite precipitation is calculated according to Eq. 5) due to increased discharge (to >760 million m³/y, $>50\%$ of modern), as estimated from the lake-level rise or surface area and evaporation rate (Figure 11a).

In summary, the characteristics of the halite-rich and detritus-rich intervals during the interglacials can be used to estimate the water discharge and duration of these cycles. Given the detritus and halite deposition rates and thicknesses, together with the calculated discharge to the lake, it follows that the climate fluctuated between extremely arid periods (with water availability $\sim 20\%$ of present day) spanning decades to centuries (15-300 y), and wetter periods (water availability 50-100% of present day) spanning centuries to millennia (90-6000 y).

6 Levant climatic context

Our knowledge of the Levant regional climate during MIS 5e (Bar-Matthews, 2014; Develle et al., 2011; Lazar and Stein, 2011; Petit-Maire et al., 2010; Torfstein et al., 2015; Vaks et al., 2006; Waldmann et al., 2010) is not extensive, and tends to emphasize high precipitation (Bar-Matthews, 2014). For example, $\delta^{18}O$ and $\delta^{13}C$ in speleothem records (e.g. Soreq cave, Bar-Matthews, 2014), have been interpreted as indicating higher amounts of precipitation in the region compared to present-day (Bar-Matthews et al., 2003). Torfstein et al. (2015) highlighted the climatic variability during this time interval based on the DSDDP core record, which indicates lake-level decline during the MIS 6/5 deglaciation between ~ 135 -128 ka, a wet period

during the MIS 5e peak insolation interval between ~128-122 ka as indicated by mud layers and an absence of halite, followed extreme aridity to ~115 ka. The wetter interval in the core coincides with the ages of the speleothems (Vaks et al., 2006) and travertine deposition (Waldmann et al., 2010) in the southern Negev desert as well as ages of diagenesis by freshwater of coral reefs in the more generally hyperarid Gulf of Aqaba (Lazar and Stein, 2011; Yehudai et al., 2016). Our record and our lake-level estimates however, clearly show that the whole MIS 5e interval was *on average* much drier than today.

As an alternative to high amounts of precipitation in the region between 128-122 ka, moisture originating in a tropical source (Torfstein et al., 2015; Vaks et al., 2006), or an increase in rainfall intensity per event, could also explain the increased wetness indicated by the DSDDP record and the $\delta^{18}\text{O}$ in Soreq cave speleothems. In this case, the increased precipitation could reflect changes in the temporal and spatial distribution of rainfall, and fluctuations between relatively wetter conditions similar to the present-day and more extreme arid conditions.

During MIS 5e the summer insolation peak was higher than during the Holocene (Berger and Loutre, 1991). Moreover, the difference between summer and winter insolation was greater, indicating greater seasonal contrast. This is seen in temperature proxies ($\delta^{18}\text{O}$, Sr/Ca) of Red Sea corals, which show increased seasonal contrast at the peak of the last interglacial (Felis et al., 2004). Moreover, according to climate models, winters during this time interval (characterized by very low insolation in the northern hemisphere) were warmer over Europe and colder over the south and central Levant compared to pre-industrial climate (Felis et al., 2004; Otto-Bliesner et al., 2013). The colder winters in the Levant and warmer winters over Europe have been associated with wetter winters in the southern Levant (Felis and Rimbu, 2010; Kushnir and Stein, 2010) .

469 While there has been significant Holocene climate variability in the Levant, it was not as
470 variable in time and space as during MIS 5e. For example, Holocene $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in
471 speleothem records from Soreq cave do not vary as much as during MIS 5e (Bar-Matthews et al.,
472 2003), and there is no evidence of increased precipitation at the edges of the desert belt, as
473 represented by the MIS 5e speleothems in the Negev (Vaks et al., 2006). Nevertheless, the
474 presence of Holocene salt in the DSDDP core, and our reconstructed water-budget, as well as
475 significant hiatuses in the sedimentary records of the marginal terraces of the Dead Sea
476 (Migowski et al., 2006) indicate that the early Holocene period was on average drier than today.
477 Both lake levels (Migowski et al., 2006) and the DSDDP core halite sequences (Figure 12) show
478 that the climate varied between conditions slightly wetter than today and extremely dry (when
479 lake level was low and halite precipitated).

480 Several studies have discussed changes in spatial pattern of the climate variability in the
481 Levant and Middle East on interannual to multidecadal time-scales, and refer in particular to
482 variations in precipitation (Cook et al., 2016; Cullen and deMenocal, 2000; Feliks et al., 2010;
483 Felis et al., 2000; Felis and Rimbu, 2010; Kushnir and Stein, 2010). The major pattern of
484 precipitation variability in the modern record is associated with the North Atlantic
485 Oscillation/Arctic Oscillation (NAO/AO), the two paradigms of a prominent mode of Northern
486 Hemisphere, low-frequency circulation variability (Hurrell et al., 2001; Wallace, 2000). The
487 NAO/AO is particularly strong in winter (December-March) when it represents changes in the
488 intensity and position of the permanent pressure centers in the North Atlantic, together with
489 meridional swings in the Atlantic jet stream and storm tracks that significantly impact the
490 Mediterranean Basin. A positive phase of the NAO, with a deep Icelandic Low and strong Azores
491 High, is associated with a colder and wetter than normal climate in the Levant (Givati and

Rosenfeld, 2013; Kushnir and Stein, 2010) and the southeastern Mediterranean (Egypt and Libya; Felis and Rimbu, 2010; Givati and Rosenfeld, 2013), while in the countries of the Mediterranean northern rim, over Turkey and south Europe, the climate is drier than normal (Cullen and deMenocal, 2000; Felis and Rimbu, 2010). The DSDDP cores record the southern Levant climate.

Comparison of the Levant climate to higher and lower latitudes can be used to infer the changes in the meridional variability during the Holocene and MIS 5e. More paleoclimatic records, and with higher resolution, are available during the Holocene compared with MIS 5e. During MIS 5e, pollen records from Europe indicate a warm and wet climate (e.g. Brauer et al., 2007; Kukla et al., 2002). The southern Sahara desert (Osborne et al., 2008; Smith et al., 2007; Szabo et al., 1995) and Arabia (Rosenberg et al., 2013) climate records also show wetter episodes particularly during the peak summer insolation interval of MIS 5e, when the increase in precipitation is associated with the intensification of the African summer monsoon and the northward contraction of the desert boundary (Battisti et al., 2014; Herold and Lohmann, 2009). However, while the wet episodes in the desert belt during MIS 5e coincide with the relatively wetter interval in the Levant and temporary rises in lake levels, it was not so during the Holocene African Humid Period (deMenocal et al., 2000), when the Dead Sea lake levels were low and halite precipitated (Figure 12). At that time, most of Europe was relatively wet, but lakes in southern Europe were in phase with the Dead Sea levels (Harrison et al., 1993). The higher summer insolation peak and the greater difference between summer and winter insolation during MIS 5e compared to the Holocene (Berger and Loutre, 1991) may explain the different response of Levant climate in these two time intervals.

7 Summary

This study uses brine compositions measured in fluid inclusions in salts and pore waters, and the thicknesses of salt layers (recovered from coring of the deepest basin of the Dead Sea by the Dead Sea Deep Drilling Program) to calculate water and salt budgets during the last interglacial MIS 5e and the Holocene, and infer the hydro-climatic conditions.

The calculations show that the climate was significantly drier than the present-day during MIS 5e and much of the Holocene. Lake levels decreased from ~308 and ~320 mbsl to ~475 and ~490 mbsl during the early Holocene and MIS 5e, respectively, resulting in substantial halite deposition, with 33.4 and 33.8 meters during MIS 5e and the Holocene, respectively). The lake level decreases were associated with significant reductions in average fresh water inputs to the lake, to ~30% of the present-day discharge in both the Holocene and MIS 5e and lasting over thousands of years. Nevertheless, during these major dry intervals, halite deposition alternated with halite-free detritus deposition, indicating a persistent pattern of intermittent flooding events when the discharge was >50% of the modern amount. These water discharge estimates and lake levels indicate that during the interglacial salt intervals the average precipitation in the Eastern Mediterranean was extremely low, with high frequency of years experiencing severe droughts, and with precipitation in Jerusalem <350 mm/y, compared to ~600 mm/y today. We use the halite and detritus sedimentation rates and the calculated lake-level changes to estimate the water discharge and duration in cycles within the interglacials. The halite-rich-detritus-rich cycles reflect fluctuations between extreme dry periods, with water availability as low as 20% of the present day spanning decades to several centuries (15-300 y), and wetter periods with water availability of ~50-100% of the present day spanning centuries to millennia (90-6000 y).

The DSDDP salt record documents intervals during the last interglacial and the early

Holocene characterized by much more severely arid conditions than those predicted by global climate models (e.g. the forecast of a ~20% decrease in water availability around the Mediterranean due to warming by the end of the 21st century, Mariotti et al., 2008). Because what happened in the past may happen again, this result may have implications for future water resources in the densely populated East Mediterranean-Levant.

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553   *Figure 1. Location map showing the main water sources into the Dead Sea, and the location of*
554   *the DSDDP deep core.*

Figure 2. Stratigraphy of the Dead Sea core. On the left summer (red) and winter (blue) insolation curves for 30° north, and the ages of halite layers (in grey).

Figure 3. Changes in chemical compositions and the halite accumulation rate (Eq. 2) during brine evaporation. (a) vs. Na⁺/Cl⁻ and (b) [Mg²⁺]. Initial molalities are [Mg²⁺]=0.53, [Na⁺]=4.9 and [Cl⁻]=6.44 (more details are in Appendix C). During halite precipitation, in the brine [Mg²⁺] increases conservatively, the Na⁺/Cl⁻ ratio decreases, and the amount of halite per meter of lake-level drop decreases because of the decrease in [Na⁺]. The blue field marks the, range of compositions of the Dead Sea brines.

Figure 4. The Dead Sea hypsometric curve (Hall, 1997) and an example of the variation in $[\text{Mg}^{2+}]$ vs. lake volume (discussed in Section 5.3). Due to the topography and bathymetry of the Dead Sea basin, the surface area and volume show significant decreases when lake level decreases from 100 to 400 mbsl, after which the volume and surface area decrease becomes very small. This behavior affects the amount of halite precipitating during lake-level drops (when $[\text{Mg}^{2+}]$ increases) as shown in Fig. 3. The $[\text{Mg}^{2+}]$ vs. volume plot is calculated according to the relationship between the initial $[\text{Mg}^{2+}]$ ($0.66 \text{ mol}\cdot\text{L}^{-1}$) and lake level (320 mbsl) at the onset of halite precipitation during MIS 5e.

Figure 5. Na^+/Cl^- molar ratio vs. Mg^{2+} molal concentration in pore waters, halite fluid inclusions and several water bodies (García-Veigas et al., 2009; Zilberman-Kron, 2008) in the Dead Sea basin. The evaporation curves of seawater (black line) and the fluid inclusion with the least $[\text{Mg}^{2+}]$ (red line) are shown. The Na^+/Cl^- of the evaporated brine decreases due to halite precipitation while $[\text{Mg}^{2+}]$ increases. The DSDDP core pore water and halite fluid inclusions all lie along a single the evaporation curve, which is different from seawater.

Figure 6. Mg^{2+} molar concentrations and Na^+/Cl^- molar ratio in pore waters (from Levy et al., 2017) and in halite fluid inclusions (Appendix B) vs depth and age in the DSDDP deep core. Halite layers are marked in grey. During glacials, $[\text{Mg}^{2+}]$ decreases and Na^+/Cl^- increases, while during interglacials, when halite precipitates, $[\text{Mg}^{2+}]$ increases and Na^+/Cl^- decreases. The core depths here are on a linear scale.

Figure 7. Mg^{2+} molal concentrations in pore waters and halite fluid inclusions in the MIS 5e interval. Halite layers are grey lines. The blue bands mark the general trend of $[\text{Mg}^{2+}]$ during precipitation of halite and the two stages (Section 5.3). During Stage 1 $[\text{Mg}^{2+}]$ increases sharply while a small thickness of halite accumulates, while during Stage 2 $[\text{Mg}^{2+}]$ increases a small amount while a large thickness of halite accumulates. The smaller increase of $[\text{Mg}^{2+}]$ during Stage 2 is probably due to halite dissolution at the lake's margins and its re-precipitation in the deep basin (halite focusing, shown in Fig. 8).

Figure 8. Conceptual model describing halite precipitation and halite focusing during MIS 5e lake-level drop. Stage 1: At lake levels higher than ~400 mbsl, halite precipitates both at the margins and in the middle of the lake. Stage 2: Halite focusing. Due to the basin topography, at low lake levels the marginal halite dissolves and re-precipitates at the center of the lake. The increase in $[\text{Mg}^{2+}]$ per cm of salt precipitated is much higher during Stage 1 than during Stage 2 (Fig. 7). Dissolution of marginal halite by fresh water decreases $[\text{Mg}^{2+}]$ and increases the Na^+/Cl^- ratio, leading to the deposition of thick layers of halite with small changes in the Mg^{2+} .

Figure 9. A flow chart of the model used to calculate halite thicknesses as a function of lake levels and changes in $[\text{Na}^+]$ and $[\text{Mg}^{2+}]$.

561

Figure 10. Initial (blue) and final (red) lake levels vs halite thickness, based on changes in $[Mg^{2+}]$ and $[Na^+]$ (Section 5.4, Table 3) during MIS 5e (solid) and the Holocene (dashed). The estimated halite thicknesses during the Holocene and MIS 5e are ~41 and ~33 meters, respectively, corresponding to initial and final lake levels of 308 and 475 mbsl during the Holocene and 320 and 490 mbsl during MIS 5e, respectively.

(a)

(b)

Figure 11. Relationships between lake levels or rainfall in Jerusalem and runoff to the Dead Sea. (a) The calculated total discharge to the Dead Sea vs lake level (based on an evaporation rate of 1.1-1.3 m/y (Lensky et al., 2005; Stanhill, 1994) and a surface area based on the Dead Sea hypsometric curve (Hall, 1997). (b) Jordan River discharge (~70% of total discharge) vs rainfall in Jerusalem between 1928 and 1965 (data is from the Israel Meteorological Service and the Israel Water Authority).

Figure 12. Halite layers (in grey) and lake levels during the Holocene. The lake-level curve is from Kushnir and Stein (2010) and Stein et al., (2010), and uses their ages. The halite layers are from the DSDDP stratigraphy, the DSDDP age model for this interval is derived from Kitagawa et al.(2016). There may be some small discrepancies between low lake levels and halite layers due to inconsistencies in the age models. The green area marks the peak of the African Humid Period (deMenocal et al., 2000).

Tables

Table 1. The main halite intervals in the Dead Sea core.

Interval depth (m)	Known halite thickness (m)	Estimated halite thickness (m)	Known detritus thickness (m)	Estimated detritus thickness (m)	Age (ka)		Calculated total lake-level drop (m)
390.36-456.7	4.5	8.1	56.1	58.2	188-220	MIS 7	58
328-233.97	29.5	33.4	56.4	60.6	116-134	MIS 5e	239
<i>279.09-233.97*</i>	<i>24.5</i>	<i>28.4</i>	<i>15.6</i>	<i>16.7</i>	<i>116-121</i>	<i>end of MIS 5e</i>	<i>203</i>
199.21-205.63	3.5	3.7	2.4	2.8	90-93	MIS 5b	26
89.25-0	30.5	33.8	47.9	55.5	0-11.4	MIS 2/1, MIS 1	141

* The italics gives the entire thick halite interval of the end of MIS 5e

Table 2. Water and major element sources into the modern Dead Sea

	Units	Brine	Jordan River (Arik Bridge)	Jordan River (Dead Sea)	Dissolution of the Sedom formation
Discharge	m ³ /y	10 ⁷	1.2·10 ⁹	1.2·10 ⁹	
Na ⁺ concentration	g/m ³	25950	12.2	253	
Cl ⁻ concentration	g/m ³	122500	18.3	474	
Mg ²⁺ concentration	g/m ³	20008	9.7	71	
SO ₄ ²⁻ concentration	g/m ³	754	21.3	174.5	
Na ⁺ flux	g/y	2.6·10 ¹¹	1.5·10 ¹⁰	3·10 ¹¹	
Cl ⁻ flux	g/y	1.2·10 ¹²	2.2·10 ¹⁰	5.7·10 ¹¹	
Mg ²⁺ flux	g/y	2·10 ¹¹	1.2·10 ¹⁰	8.6·10 ¹⁰	
SO ₄ ²⁻ flux	g/y	7.5·10 ⁹	2.6·10 ¹⁰	2.1·10 ¹¹	
Na ⁺ /Cl ⁻ molar ratio		0.32	1	0.83	
Flux in terms of potential accumulation rate of halite	mm/y	0.5	0.03	0.6	~0.2*

Data from Bentor (1961); Stein et al.(1997); Salameh and El-Naser (1999); The hydrological service of Israel; Dead Sea surface area is estimated as 600-700 km². Potential contribution of the Sedom Formation was calculated according to the rate of uplift (6.2 mm/y, Frumkin, 1996) and multiply by the diapir area (~14 km²) divided by the Dead Sea area.

*Total halite dissolved during the rise of the diapir was estimated to be 200 m (Zak, 1967). Multiplying this thickness by the Mt. Sedom to Dead Sea area ratio (~14 km²/~600 km², respectively), yields a halite thickness of just ~5 m halite within the Dead Sea.

Table 3. Initial and final concentrations of Na⁺ and Mg²⁺ during the Holocene and MIS 5e.

The second stage of halite precipitation is characterized by slower change in the chemistry and more halite in the lake center.

	Initial Mg ²⁺	Final Mg ²⁺	Initial Na ⁺	Final Na ⁺	change in volume	Halite thickness
	mol·L ⁻¹				%	meters
Holocene	0.623	1.667	3.75	1.49	37	41.3 ¹
MIS 5e	0.662	1.824	3.09	1.09	36	33.4
Stage 2						
Holocene	1.489	1.667	2.07	1.49	94	16.4
MIS 5e	1.607	1.824	1.71	1.09	88	28.9

¹ Estimated by the thickness of halite in the core during the Holocene (33.8 m) plus estimated marginal halite thickness of 15 m over surface area of 300 km² distributed over a lake basin of 600 km² (=7.5 m).

Table 4. Modeled initial and final lake levels, discharge to the lake, and rainfall during the Holocene and MIS 5e

	Initial lake level (m)	Final lake level (m)	Average discharge (million m ³ /y)	% of modern discharge	Average precipitation in Jerusalem (mm/y)	% of average modern precipitation
Holocene	-308	-475	580-690	30-50	300-400	50-67
MIS 5e	-320	-490	560-670	30-50	300-400	50-67

References

- Ayora, C., Garcia-Veigas, J., Pueyo, J.-J., 1994. X-ray microanalysis of fluid inclusions and its application to the geochemical modeling of evaporite basins. *Geochim. Cosmochim. Acta* 58, 43–55. doi:10.1016/0016-7037(94)90444-8
- Bar-Matthews, M., 2014. History of Water in the Middle East and North Africa, in: *Treatise on Geochemistry*. Elsevier, pp. 109–128. doi:10.1016/B978-0-08-095975-7.01210-9
- Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., Hawkesworth, C.J., 2003. Sea – land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. *Geochim. Cosmochim. Acta* 67, 3181–3199.
- Bartov, Y., Goldstein, S.L., Stein, M., Enzel, Y., 2003. Catastrophic arid episodes in the Eastern Mediterranean linked with the North Atlantic Heinrich events. *Geology* 31, 439–442. doi:10.1130/0091-7613(2003)031<0439:CAEITE>2.0.CO;2
- Bartov, Y., Stein, M., Enzel, Y., Agnon, A., Reches, Z., 2002. Lake Levels and Sequence Stratigraphy of Lake Lisan, the Late Pleistocene Precursor of the Dead Sea. *Quat. Res.* 57, 9–21. doi:10.1006/qres.2001.2284
- Battisti, D.S., Ding, Q., Roe, G.H., 2014. Coherent pan-Asian climatic and isotopic response to orbital forcing of tropical insolation. *J. Geophys. Res. Atmos.* 119, 11997–12020. doi:10.1002/2014JD021960.Received
- Bentor, Y.K., 1961. Some geochemical aspects of the Dead Sea and the question of its age. *Geochim. Cosmochim. Acta* 25, 239–260.
- Berger, A., Loutre, M.F., 1991. Insolation values for the climate of the last 10 million years.

589 Quat. Sci. Rev. 10, 297–317. doi:10.1016/0277-3791(91)90033-Q

590 Bookman (Ken-Tor), R., Enzel, Y., Agnon, A., Stein, M., 2004. Late Holocene lake levels of the
591 Dead Sea. Geol. Soc. Am. Bull. 116, 555–571. doi:10.1130/B25286.1

592 Brauer, A., Allen, J.R.M., Mingram, J., Dulski, P., Wulf, S., Huntley, B., 2007. Evidence for last
593 interglacial chronology and environmental change from Southern Europe. Proc. Natl. Acad.
594 Sci. U. S. A. 104, 450–455. doi:10.1073/pnas.0603321104

595 Cane, T., Rohling, E.J., Kemp, A.E.S., Cooke, S., Pearce, R.B., 2002. High-resolution
596 stratigraphic framework for Mediterranean sapropel S5: Defining temporal relationships
597 between records of Eemian climate variability. Palaeogeogr. Palaeoclimatol. Palaeoecol.
598 183, 87–101. doi:10.1016/S0031-0182(01)00461-8

599 Cook, B.I., Anchukaitis, K.J., Touchan, R., Meko, D.M., Cook, E.R., 2016. Spatiotemporal
600 drought variability in the Mediterranean over the last 900 years. J. Geophys. Res. Atmos.
601 121, 2060–2074. doi:10.1002/2015JD023929.Received

602 Cullen, H.M., deMenocal, P.B., 2000. North Atlantic influence on Tigris-Euphrates streamflow.
603 Int. J. Climatol. 20, 853–863.

604 Damnati, B., 2000. Holocene lake records in the Northern Hemisphere of Africa. J. African Earth
605 Sci. 31, 253–262. doi:10.1016/S0899-5362(00)00089-0

606 deMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Sarnthein, M., Baker, L., Yarusinsky, M.,
607 2000. Abrupt onset and termination of the African Humid Period: Quat. Sci. Rev. 19, 347–
608 361. doi:10.1016/S0277-3791(99)00081-5

609 Develle, A.-L., Gasse, F., Vidal, L., Williamson, D., Demory, F., Van Campo, E., Ghaleb, B.,
610 Thouveny, N., 2011. A 250ka sedimentary record from a small karstic lake in the Northern

611   Levant (Yammoûneh, Lebanon). *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 305, 10–27.
 612   doi:10.1016/j.palaeo.2011.02.008

613   Dutton, A., Lambeck, K., 2012. Ice Volume and Sea Level During the Last Interglacial. *Science*
 614   (80-.). 337, 216–219.

615   Enzel, Y., Bookman (Ken-Tor), R., Sharon, D., Gvirtzman, H., Dayab, U., Ziv, B., Stein, M.,
 616   2003. Late Holocene climates of the Near East deduced from Dead Sea level variations and
 617   modern regional winter rainfall. *Quat. Res.* 60, 263–273. doi:10.1016/j.yqres.2003.07.011

618   Feliks, Y., Ghil, M., Robertson, A.W., 2010. Oscillatory Climate Modes in the Eastern
 619   Mediterranean and Their Synchronization with the North Atlantic Oscillation. *J. Clim.* 23,
 620   4060–4079. doi:10.1175/2010JCLI3181.1

621   Felis, T., Lohmann, G., Kuhnert, H., Lorenz, S.J., Scholz, D., Pätzold, J., Al-Rousan, S.A., Al-
 622   Moghrabi, S.M., 2004. Increased seasonality in Middle East temperatures during the last
 623   interglacial period. *Nature* 429, 164–168. doi:10.1038/nature02546

624   Felis, T., Pätzold, J., Loya, Y., Fine, M., Nawar, A.H., Wefer, G., 2000. A coral oxygen isotope
 625   record from the northern Red Sea documenting NAO, ENSO, and North Pacific
 626   teleconnections on Middle East climate variability since the year 1750. *Paleoceanography*
 627   15, 679–694. doi:10.1029/1999PA000477

628   Felis, T., Rimbu, N., 2010. Mediterranean climate variability documented in oxygen isotope
 629   records from northern Red Sea corals — A review. *Glob. Planet. Change* 71, 232–241.
 630   doi:10.1016/j.gloplacha.2009.10.006

631   Frierson, D.M.W., Lu, J., Chen, G., 2007. Width of the Hadley cell in simple and comprehensive
 632   general circulation models. *Geophys. Res. Lett.* 34, 1–5. doi:10.1029/2007GL031115

633 Frumkin, A., 1996. Uplift rate relative to base-levels of a salt diapir (Dead Sea Basin Israel) as
634 indicated by cave levels. *Geol. Soc. London, Spec. Publ.* 100, 41–47.

635 García-Veigas, J., Rosell, L., Zak, I., Playà, E., Ayora, C., Starinsky, A., 2009. Evidence of
636 potash salt formation in the Pliocene Sedom Lagoon (Dead Sea Rift, Israel). *Chem. Geol.*
637 265, 499–511. doi:10.1016/j.chemgeo.2009.05.013

638 Garcia-Veigas, J., 1993. *Geoquímica de inclusiones fluidas en formaciones salinas, Microanálisis*
639 *Cryo-SEM-EDS*. Universitat de Barcelona.

640 Givati, A., Rosenfeld, D., 2013. The Arctic Oscillation , climate change and the effects on
641 precipitation in Israel. *Atmos. Res.* 132–133, 114–124. doi:10.1016/j.atmosres.2013.05.001

642 Govin, A., Capron, E., Tzedakis, P.C., Verheyden, S., Ghaleb, B., Hillaire-Marcel, C., St-Onge,
643 G., Stoner, J.S., Bassinot, F., Bazin, L., Blunier, T., Combourieu-Nebout, N., El Ouahabi,
644 A., Genty, D., Gersonde, R., Jimenez-Amat, P., Landais, A., Martrat, B., Masson-Delmotte,
645 V., Parrenin, F., Seidenkrantz, M.-S., Veres, D., Waelbroeck, C., Zahn, R., 2015. Sequence
646 of events from the onset to the demise of the Last Interglacial: Evaluating strengths and
647 limitations of chronologies used in climatic archives. *Quat. Sci. Rev.* 129, 1–36.
648 doi:10.1016/j.quascirev.2015.09.018

649 Haliva-Cohen, A., Stein, M., Goldstein, S.L., Sandler, A., Starinsky, A., 2012. Sources and
650 transport routes of fine detritus material to the Late Quaternary Dead Sea basin. *Quat. Sci.*
651 *Rev.* 50, 55–70. doi:10.1016/j.quascirev.2012.06.014

652 Hall, J.K., 1997. Topography and bathymetry of the Dead Sea Depression, in: Niemi, T.M., Ben-
653 Avraham, Z., Gat, J.R. (Eds.), *The Dead Sea -- the Lake and Its Settings*. Oxford
654 *Monographs on Geology and Geophysics* No. 36, pp. 11–21.

655 Harrison, S.P., Prentice, I.C., Guiot, J., 1993. Climatic controls on Holocene lake-level changes
 656 in Europe. *Clim. Dyn.* 8, 189–200. doi:10.1007/BF00207965

657 Hazan, N., Stein, M., Agnon, A., Marco, S., Nadel, D., Negendank, J.F.W., Schwab, M.J., Neev,
 658 D., 2005. The late Quaternary limnological history of Lake Kinneret (Sea of Galilee), Israel.
 659 *Quat. Res.* 63, 60–77. doi:10.1016/j.yqres.2004.09.004

660 Herold, M., Lohmann, G., 2009. Eemian tropical and subtropical African moisture transport: an
 661 isotope modelling study. *Clim. Dyn.* 33, 1075–1088. doi:10.1007/s00382-008-0515-2

662 Hurrell, J.W., Kushnir, Y., Visbeck, M., 2001. The North Atlantic Oscillation. *Science* (80-.).
 663 291.

664 Katz, A., Starinsky, A., 2008. Geochemical History of the Dead Sea. *Aquat. Geochemistry* 15,
 665 159–194. doi:10.1007/s10498-008-9045-0

666 Kelley, C., Ting, M., Seager, R., Kushnir, Y., 2012. Mediterranean precipitation climatology,
 667 seasonal cycle, and trend as simulated by CMIP5. *Geophys. Res. Lett.* 39, L21703.
 668 doi:10.1029/2012GL053416

669 Kiro, Y., Goldstein, S.L., Lazar, B., Stein, M., 2016. Environmental implications of salt facies in
 670 the Dead Sea. *Geol. Soc. Am. Bull.* 128, 824–841. doi:10.1130/B31357.1

671 Kitagawa, H., Stein, M., Goldstein, S.L., Nakamura, T., Lazar, B., DSDDP Scientific Party,
 672 2016. Radiocarbon chronology of the DSDDP core at the deepest floor of the Dead Sea.
 673 *Radiocarbon* 1–12. doi:10.1017/RDC.2016.120

674 Klein, C., Flohn, A., 1987. Contribution to the knowledge in the fluctuations of the Dead Sea
 675 level. *Theor. Appl. Climatol.* 38, 151–156.

676 Kolodny, Y., Stein, M., Machlus, M., 2005. Sea-rain-lake relation in the Last Glacial East

677 Mediterranean revealed by $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ in Lake Lisan aragonites. *Geochim. Cosmochim.*
678 *Acta* 69, 4045–4060. doi:10.1016/j.gca.2004.11.022

679 Krumgalz, B.S., Hecht, A., Starinsky, A., Katz, A., 2000. Thermodynamic constraints on Dead
680 Sea evaporation: can the Dead Sea dry up? *Chem. Geol.* 165, 1–11.

681 Krumgalz, B.S., Magdal, E., Starinsky, A., 2002. The evolution of a chloride sedimentary
682 sequence-simulated evaporation of the Dead Sea. *Isr. J. Earth Sci.* 51, 253–267.
683 doi:10.1560/EL8J-PVU9-EH88-M083

684 Kukla, G.J., Bender, M.L., de Beaulieu, J.-L., Bond, G., Broecker, W.S., Cleveringa, P., Gavin,
685 J.E., Herbert, T.D., Imbrie, J., Jouzel, J., Keigwin, L.D., Knudsen, K.-L., McManus, J.F.,
686 Merkt, J., Muhs, D.R., Müller, H., Poore, R.Z., Porter, S.C., Seret, G., Shackleton, N.J.,
687 Turner, C., Tzedakis, P.C., Winograd, I.J., 2002. Last Interglacial Climates. *Quat. Res.* 58,
688 2–13. doi:10.1006/qres.2001.2316

689 Kushnir, Y., Stein, M., 2010. North Atlantic influence on 19th–20th century rainfall in the Dead
690 Sea watershed, teleconnections with the Sahel, and implication for Holocene climate
691 fluctuations. *Quat. Sci. Rev.* 29, 3843–3860. doi:10.1016/j.quascirev.2010.09.004

692 Lazar, B., Stein, M., 2011. Freshwater on the route of hominids out of Africa revealed by U-Th
693 in Red Sea corals. *Geology* 39, 1067–1070. doi:10.1130/G32257.1

694 Lelieveld, J., Hadjinicolaou, P., Kostopoulou, E., Chenoweth, J., El Maayar, M.,
695 Giannakopoulos, C., Hannides, C., Lange, M. a., Tanarhte, M., Tyrlis, E., Xoplaki, E., 2012.
696 Climate change and impacts in the Eastern Mediterranean and the Middle East. *Clim.*
697 *Change* 114, 667–687. doi:10.1007/s10584-012-0418-4

698 Lensky, N.G., Dvorkin, Y., Lyakhovsky, V., Gertman, I., Gavrieli, I., 2005. Water, salt, and

699 energy balances of the Dead Sea. *Water Resour. Res.* 41, W12418.
700 doi:10.1029/2005WR004084

701 Levy, E.J., Stein, M., Lazar, B., Yechieli, Y., Gavrieli, I., Sivan, O., 2016. Pore fluids in Dead Sea
702 sediment core reveal linear response of lake chemistry to global climate changes. *Geology*
703 IN PRESS. doi:10.1130/G38685.1

704 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally distributed
705 benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20, 1–17. doi:10.1029/2004PA001071

706 Lowenstein, T.K., Dolginko, L.A.C., García-Veigas, J., 2016. Influence of magmatic-
707 hydrothermal activity on brine evolution in closed basins: Searles Lake, California. *Geol.*
708 *Soc. Am. Bull.* B31398.1. doi:10.1130/B31398.1

709 Machlus, M., Enzel, Y., Goldstein, S.L., Marco, S., Stein, M., 2000. Reconstructing low levels of
710 Lake Lisan by correlating fan-delta and lacustrine deposits. *Quat. Int.* 73–74, 137–144.
711 doi:10.1016/S1040-6182(00)00070-7

712 Mariotti, A., Zeng, N., Yoon, J.-H., Artale, V., Navarra, A., Alpert, P., Li, L.Z.X., 2008.
713 Mediterranean water cycle changes: transition to drier 21st century conditions in
714 observations and CMIP3 simulations. *Environ. Res. Lett.* 3, 44001. doi:10.1088/1748-
715 9326/3/4/044001

716 Matmon, A., Fink, D., Davis, M., Niedermann, S., Rood, D., Frumkin, A., 2014. Unraveling rift
717 margin evolution and escarpment development ages along the Dead Sea fault using
718 cosmogenic burial ages. *Quat. Res.* doi:10.1016/j.yqres.2014.04.008

719 Mayewski, P. a., Rohling, E.E., Curt Stager, J., Karlén, W., Maasch, K. a., David Meeker, L.,
720 Meyerson, E. a., Gasse, F., van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G.,

721 Rack, F., Staubwasser, M., Schneider, R.R., Steig, E.J., 2004. Holocene climate variability.
 722 Quat. Res. 62, 243–255. doi:10.1016/j.yqres.2004.07.001

723 McGarry, S., Bar-Matthews, M., Matthews, A., Vaks, A., Schilman, B., Ayalon, A., 2004.
 724 Constraints on hydrological and paleotemperature variations in the Eastern Mediterranean
 725 region in the last 140ka given by the δD values of speleothem fluid inclusions. Quat. Sci.
 726 Rev. 23, 919–934. doi:10.1016/j.quascirev.2003.06.020

727 Migowski, C., Agnon, A., Bookman, R., Negendank, J.F., Stein, M., 2004. Recurrence pattern of
 728 Holocene earthquakes along the Dead Sea transform revealed by varve-counting and
 729 radiocarbon dating of lacustrine sediments. Earth Planet. Sci. Lett. 222, 301–314.
 730 doi:10.1016/j.epsl.2004.02.015

731 Migowski, C., Stein, M., Prasad, S., Negendank, J.F.W., Agnon, A., 2006. Holocene climate
 732 variability and cultural evolution in the Near East from the Dead Sea sedimentary record.
 733 Quat. Res. 66, 421–431. doi:10.1016/j.yqres.2006.06.010

734 Neev, D., Emery, K., 1995. The Destruction of Sodom, Gomorrah, and Jericho. Geological,
 735 Climatological, and Archaeological Background. Oxford University Press, New York.

736 Neugebauer, I., Brauer, A., Schwab, M.J., Waldmann, N.D., Enzel, Y., Kitagawa, H., Torfstein,
 737 A., Frank, U., Dulski, P., Agnon, A., Ariztegui, D., Ben-Avraham, Z., Goldstein, S.L., Stein,
 738 M., 2014. Lithology of the long sediment record recovered by the ICDP Dead Sea Deep
 739 Drilling Project (DSDDP). Quat. Sci. Rev. 102, 149–165.
 740 doi:10.1016/j.quascirev.2014.08.013

741 Neugebauer, I., Schwab, M.J., Waldmann, N.D., Tjallingii, R., Frank, U., Hadzhiivanova, E.,
 742 Naumann, R., Taha, N., Agnon, A., Enzel, Y., Brauer, A., 2015. Hydroclimatic variability in

743 the Levant during the early last glacial (∼ 117–75 ka) derived from micro-facies
744 analyses of deep Dead Sea sediments. *Clim. Past Discuss.* 11, 3625–3663. doi:10.5194/cpd-
745 11-3625-2015

746 Osborne, A.H., Vance, D., Rohling, E.J., Barton, N., Rogerson, M., Fello, N., 2008. A humid
747 corridor across the Sahara for the migration of early modern humans out of Africa 120,000
748 years ago. *Pnas* 105, 16444–16447. doi:10.1073/pnas.0804472105

749 Otto-Bliesner, B.L., Rosenbloom, N., Stone, E.J., Mckay, N.P., Lunt, D.J., Brady, E.C.,
750 Overpeck, J.T., 2013. How warm was the last interglacial ? New model – data comparisons.
751 *Philos. Trans. R. Soc. A Math. Phys. Eng. Sci.* 371, 1–20.

752 Parkurst, D.L., Appelo, C.A.J., 2013. Description of Input and Examples for PHREEQC Version
753 3—A Computer Program for Speciation, Batch-Reaction, One-Dimensional Transport, and
754 Inverse Geochemical Calculations, in: *U.S. Geological Survey Techniques and Methods*,
755 Book 6. p. 497.

756 Petit-Maire, N., Carbonel, P., Reyss, J.L., Sanlaville, P., Abed, A., Bourrouilh, R., Fontugne, M.,
757 Yasin, S., 2010. A vast Eemian palaeolake in Southern Jordan (29 ° N). *Glob. Planet.*
758 *Change* 72, 368–373. doi:10.1016/j.gloplacha.2010.01.012

759 Pitzer, K.S., 1973. Thermodytriarnics of Electrolytes. I. Theoretical Basis and General Equations.
760 *J. Phys. Chem.* 77, 268–277. doi:10.1021/j100621a026

761 Prasad, S., Vos, H., Negendank, J.F.W., Waldmann, N., Goldstein, S.L., Stein, M., 2004.
762 Evidence from Lake Lisan of solar influence on decadal- to centennial-scale climate
763 variability during marine oxygen isotope stage 2. *Geology* 32, 581. doi:10.1130/G20553.1

764 Rohling, E.J., 2013. Quantitative assessment of glacial fluctuations in the level of Lake Lisan,

765 Dead Sea rift. *Quat. Sci. Rev.* 70, 63–72. doi:10.1016/j.quascirev.2013.03.013

766 Rohling, E.J., Cane, T.R., Cooke, S., Sprovieri, M., Bouloubassi, I., Emeis, K.C., Schiebel, R.,
767 Kroon, D., Jorissen, F.J., Lorré, a., Kemp, a. E.S., 2002. African monsoon variability
768 during the previous interglacial maximum. *Earth Planet. Sci. Lett.* 202, 61–75.
769 doi:10.1016/S0012-821X(02)00775-6

770 Rosenberg, T.M., Preusser, F., Risberg, J., Pliikk, A., Kadi, K.A., Matter, A., Fleitmann, D., 2013.
771 Middle and Late Pleistocene humid periods recorded in palaeolake deposits of the Nafud
772 desert, Saudi Arabia. *Quat. Sci. Rev.* 70, 109–123. doi:10.1016/j.quascirev.2013.03.017

773 Rossignol-Strick, M., 1985. Mediterranean Quaternary sapropels, an immediate response of the
774 African monsoon to variation of insolation. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 49,
775 237–263. doi:10.1016/0031-0182(85)90056-2

776 Salameh, E., El-Naser, H., 1999. Does the actual drop in Dead Sea level reflect the development
777 of water sources within its drainage basin? *Acta Hydrochim. Hydrobiol.* 27, 5–11.

778 Scheff, J., Frierson, D.M.W., 2014. Scaling Potential Evapotranspiration with Greenhouse
779 Warming. *J. Clim.* 27, 1539–1558. doi:10.1175/JCLI-D-13-00233.1

780 Seager, R., Liu, H., Henderson, N., Simpson, I., Kelley, C., Shaw, T., Kushnir, Y., Ting, M.,
781 2014. Causes of Increasing Aridification of the Mediterranean Region in Response to Rising
782 Greenhouse Gases*. *J. Clim.* 27, 4655–4676. doi:10.1175/JCLI-D-13-00446.1

783 Smith, J.R., Hawkins, A.L., Asmerom, Y., Polyak, V., Giegengack, R., 2007. New age constraints
784 on the Middle Stone Age occupations of Kharga Oasis, Western Desert, Egypt. *J. Hum.*
785 *Evol.* 52, 690–701. doi:10.1016/j.jhevol.2007.01.004

786 Stanhill, G., 1994. Changes in the rate of evaporation from the Dead Sea. *Int. J. Climatol.* 14,

787 465–471.

788 Starinsky, A., 1974. Interaction between calcium chloride brines and sedimentary rocks. The
789 Hebrew University, Jerusalem.

790 Stein, M., 2001. The sedimentary and geochemical record of Neogene-Quaternary water bodies
791 in the Dead Sea basin-inferences for the regional paleoclimatic history. *J. Paleolimnol.* 21,
792 271–282.

793 Stein, M., 2014. The evolution of Neogene-Quaternary water-bodies in the Dead Sea rift valley,
794 in: Garfunkel, Z., Ben-Avraham, Z., Kagan, E. (Eds.), *Dead Sea Transform Fault System:*
795 *Reviews.* Springer, pp. 279–316. doi:10.1007/978-94-017-8872-4

796 Stein, M., Ben-Avraham, Z., Goldstein, S.L., 2011. Dead Sea Deep Cores: A Window A Library
797 of Georeferenced Into Past Climate and Seismicity. *Eos (Washington. DC).* 92, 453–454.
798 doi:10.1111/j.2006.0906-7590.04596.x.Xiao

799 Stein, M., Starinsky, A., Katz, A., Goldstein, S.L., Machlus, M., Schram, A., 1997. Strontium
800 isotopic, chemical, and sedimentological evidence for the evolution of Lake Lisan and the
801 dead Sea. *Geochim. Cosmochim. Acta* 61, 3975–3992.

802 Stein, M., Starinsky, a., Agnon, a., Katz, a., Raab, M., Spiro, B., Zak, I., 2000. The impact of
803 brine-rock interaction during marine evaporite formation on the isotopic Sr record in the
804 oceans: evidence from Mt. Sedom, Israel. *Geochim. Cosmochim. Acta* 64, 2039–2053.
805 doi:10.1016/S0016-7037(00)00370-7

806 Stein, M., Torfstein, A., Gavrieli, I., Yechieli, Y., 2010. Abrupt aridities and salt deposition in the
807 post-glacial Dead Sea and their North Atlantic connection. *Quat. Sci. Rev.* 29, 567–575.
808 doi:10.1016/j.quascirev.2009.10.015

809 Szabo, B.J., Haynes, C.V., Maxwell, T. a., 1995. Ages of Quaternary pluvial episodes determined
 810 by uranium-series and radiocarbon dating of lacustrine deposits of Eastern Sahara.
 811 *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 113, 227–242. doi:10.1016/0031-0182(95)00052-
 812 N

813 Torfstein, A., Gavrieli, I., Katz, A., Kolodny, Y., Stein, M., 2008. Gypsum as a monitor of the
 814 paleo-limnological-hydrological conditions in Lake Lisan and the Dead Sea. *Geochim.*
 815 *Cosmochim. Acta* 72, 2491–2509.

816 Torfstein, A., Gavrieli, I., Stein, M., 2005. The sources and evolution of sulfur in the hypersaline
 817 Lake Lisan (paleo-Dead Sea). *Earth Planet. Sci. Lett.* 236, 61–77.
 818 doi:10.1016/j.epsl.2005.04.026

819 Torfstein, A., Goldstein, S.L., Kagan, E.J., Stein, M., 2013. Integrated multi-site U-Th
 820 chronology of the 1st glacial Lake Lisan. *Geochim. Cosmochim. Acta* 104, 210–231.

821 Torfstein, A., Goldstein, S.L., Kushnir, Y., Enzel, Y., Haug, G., Stein, M., 2015. Dead Sea
 822 drawdown and monsoonal impacts in the Levant during the last interglacial. *Earth Planet.*
 823 *Sci. Lett.* 412, 235–244.

824 Torfstein, A., Haase-Schramm, A., Waldmann, N., Kolodny, Y., Stein, M., 2009. U-series and
 825 oxygen isotope chronology of the mid-Pleistocene Lake Amora (Dead Sea basin). *Geochim.*
 826 *Cosmochim. Acta* 73, 2603–2630. doi:10.1016/j.gca.2009.02.010

827 Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Halicz, L., Frumkin, A., 2007. Desert
 828 speleothems reveal climatic window for African exodus of early modern humans. *Geology*
 829 35, 831–834. doi:10.1130/G23794A.1

830 Vaks, a., Bar-Matthews, M., Ayalon, a., Matthews, a., Frumkin, a., Dayan, U., Halicz, L.,

831 Almogi-Labin, a., Schilman, B., 2006. Paleoclimate and location of the border between
 832 Mediterranean climate region and the Saharo–Arabian Desert as revealed by speleothems
 833 from the northern Negev Desert, Israel. *Earth Planet. Sci. Lett.* 249, 384–399.
 834 doi:10.1016/j.epsl.2006.07.009

835 Waldmann, N., Stein, M., Ariztegui, D., Starinsky, a., 2009. Stratigraphy, depositional
 836 environments and level reconstruction of the last interglacial Lake Samra in the Dead Sea
 837 basin. *Quat. Res.* 72, 1–15. doi:10.1016/j.yqres.2009.03.005

838 Waldmann, N., Torfstein, a., Stein, M., 2010. Northward intrusions of low- and mid-latitude
 839 storms across the Saharo-Arabian belt during past interglacials. *Geology* 38, 567–570.
 840 doi:10.1130/G30654.1

841 Wallace, J.M., 2000. North atlantic oscillation annular mode: Two paradigms-one phenomenon.
 842 *Q. J. R. Meteorol. Soc.* 126, 791–805. doi:10.1002/qj.49712656402

843 Wang, P., Tian, J., Lourens, L.J., 2010. Obscuring of long eccentricity cyclicity in Pleistocene
 844 oceanic carbon isotope records. *Earth Planet. Sci. Lett.* 290, 319–330.
 845 doi:10.1016/j.epsl.2009.12.028

846 Weinberger, R., Bar-Matthews, M., Levi, T., Begin, Z.B., 2007. Late-Pleistocene rise of the
 847 Sedom diapir on the backdrop of water-level fluctuations of Lake Lisan, Dead Sea basin.
 848 *Quat. Int.* 175, 53–61. doi:10.1016/j.quaint.2007.03.007

849 Yechieli, Y., Gavrieli, I., Berkowitz, B., Ronen, D., 1998. Will the Dead Sea die? *Geology* 26,
 850 755–758.

851 Yechieli, Y., Magaritz, M., Levy, Y., Weber, U., Kafri, U., Woelfli, W., Bonani, G., 1993. Late
 852 Quaternary Geological History of the Dead Sea Area, Israel. *Quat. Res.* doi:DOI:

853 10.1006/qres.1993.1007

854 Yehudai, M., Lazar, B., Bar, N., Kiro, Y., Agnon, A., Shaked, Y., Stein, M., 2016. U-Th dating of
855 calcitic corals from the Gulf of Aqaba. *Geochim. Cosmochim. Acta* In press.

856 Zak, I., 1967. *The Geology of Mount Sedom*. The Hebrew University, Jerusalem.

857 Zilberman-Kron, T., 2008. *The origin and geochemical evolution of brines in the sinkholes along*
858 *the Dead Sea coast*. The Hebrew University.

859

860