Quantitative assessment of glacial fluctuations in the level of Lake Lisan, Dead Sea rift

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A quantitative understanding of climatic variations in the Levant during the last glacial cycle is needed to support archaeologists in assessing the drivers behind hominin migrations and cultural developments in this key region at the intersection between Africa and Europe. It will also foster a better understanding of the region’s natural variability as context to projections of modern climate change. Detailed documentation of variations in the level of Lake Lisan – the lake that occupied the Dead Sea rift during the last glacial cycle – provides crucial climatic information for this region. Existing reconstructions suggest that Lake Lisan highstands during cold intervals of the last glacial cycle represent relatively humid conditions in the region, but these interpretations have remained predominantly qualitative. Here, I evaluate realistic ranges of the key climatological parameters that controlled lake level, based on the observed timing and amplitudes of lake-level variability. I infer that a mean precipitation rate over the wider catchment area of about 500 mm y⁻¹, as proposed in the literature, would be consistent with observed lake levels if there was a concomitant 15–50% increase in wind speed during cold glacial stadials. This lends quantitative support to previous inferences of a notable increase in the intensity of Mediterranean (winter) storms during glacial periods, which tracked eastward into the Levant. In contrast to highstands during ‘regular’ stadials, lake level dropped during Heinrich Events. I demonstrate that this likely indicates a further intensification of the winds during those times.

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1. Introduction

The Levant is located at the intersection of Africa and Europe, and is one of a few prime routes for hominin migrations and cultural exchanges between these continents (e.g., Fernandes et al., 2006; Abbate and Sagri, 2012; and references therein). Archaeologists require a detailed and well-dated understanding of climatic variations in the region in order to assess the potential role of climate in driving or facilitating hominin migrations and cultural developments.

Implications for hominins are not the only reason for striving toward a better understanding of Levantine climate variability. Parts of the region are highly sensitive to potential aridification under global climate change (e.g., Watson et al., 1997), and future projections require a sound understanding of any underlying natural variability. Similarly, intensive anthropogenic water use (e.g., diversion of the Jordan River) and proposed engineering projects such as a Red Sea–Dead Sea canal are presenting unprecedented challenges to the environment, assessment of which may be helped by a better understanding of the region’s natural variability (for evaluations of these contemporaneous issues, see for example: Salhotra et al., 1985; Stanhill, 1994; Alpert et al., 1997; Asmar and Ergenzinger, 2002a,b,c,d; 2003; Lensky et al., 2005). The present paper focuses on natural variability during the last glacial cycle (specifically, the last 120,000 years). A recent overview (Frumkin et al., 2011), which extends on previous syntheses (e.g., Enzel et al., 2003, 2008), offers a comprehensive overview of climatic changes in the Levant based on varied evidence. This included changes in the level, sedimentology, and chemical properties of Lake Lisan (Fig. 1), cave speleothem deposits, and loess deposits in the region. In particular, reconstructions indicate a similarity between Lake Lisan level fluctuations and Northern Hemisphere climate ‘cycles’ as recorded in Greenland ice-core records (among many others: Bartov et al., 2002; Migowski et al., 2006; Stein and Goldstein, 2006; Waldmann et al., 2007, 2010; Lisker et al., 2009; Torfstein et al., 2009, 2013a; and Stein, Goldstein, and Enzel, pers. comm. at the Ein Gedi fieldtrip, 2012). Early work
had suggested that cold episodes in Greenland correlated with arid conditions in the Levant on decadal to centennial timescales (e.g., Prasad et al., 2004). However, the detailed studies listed above comprehensively demonstrate that cold episodes were instead associated with Lake Lisan highstands on multi-centennial to millennial timescales, reaching up to 200 m or more above the modern Dead Sea level. Intervals corresponding to Heinrich events appear to have deviated from that pattern, with dry conditions and relative lowstands in Lake Lisan (although still 60–120 m above the present-day Dead Sea level; Bartov et al., 2003; Torfstein et al., 2013a).

The previous work generally argued that Lake Lisan highstands represent climatically wet episodes. To date, however, such interpretations have remained largely phenomenological, did not really account for variability in the (large) evaporative loss term, and were expressed in qualitative or semi-quantitative terms only. The present study builds on this previous work and on modelling of recent changes in the region (Salhotra et al., 1985; Stanhill, 1994; Alpert et al., 1997; Asmar and Ergenzinger, 2002a,b,c,d; 2003; Lensky et al., 2005), to advance a quantitatively coherent understanding of relationships between the various observed aspects of regional climate variability over the last glacial cycle. A basic model is developed in which lake level depends on changes in evaporation from the variable Lake Lisan surface area, and on mean net precipitation over the wider catchment area. I then evaluate — for observed timings and amplitudes of lake-level variations — realistic precipitation ranges over the wider catchment, given values (with uncertainties) of evaporation fluctuations that are based on realistic ranges of variability in key climatological parameters. The solutions provide internally consistent suites of potential variability in the regional climate (precipitation, evaporation, wind speed) and lake level over the last glacial cycle. The model is not exhaustive and lacks details that are included in the aforementioned studies of recent changes, which is deliberate to make it applicable to the long timescales and large uncertainties (shortage of direct measurements) associated with geological studies. Despite these simplifications, the results provide indicative, internally consistent, first-order estimates of regional environmental variability.

2. Method

As with any lake, fluctuations in the level of Lake Lisan (H) reflect the integration over time of mean net precipitation (Ψ) over the lake’s wider catchment area (Acatch), and evaporation (E) from the lake’s surface area (A). The lake’s surface area varies with its level; as lake level rises, the lake occupies more of the rift valley, and its surface area increases (Fig. 1). This variation of surface area with lake level is important, because the rate of evaporation, which affects the lake surface area, is high in this region. Currently, the rate of evaporation from fresh water (often referred to as ‘pan evaporation’) is of the order of ~3 m y⁻¹ in this region (Alpert et al., 1997). Due to the influences of high salinity (around S = 276), however, the rate of evaporation from the sea itself is near to ~1.3 m y⁻¹ (note, these are pre-1951 values that pre-date extensive anthropogenic diversion of the Jordan River (Salhotra et al., 1985; Stanhill, 1994; Alpert et al., 1997). In contrast to A, the size of Acatch (which receives the precipitation that feeds the lake) is determined by the larger-scale regional topographic configuration, which is assumed to have remained constant throughout the last glacial cycle (Fig. 1).

Today, Acatch is roughly 40 times larger than the Dead Sea surface area (e.g., Waldmann et al., 2010; Frumkin et al., 2011) (Fig. 1). To maintain the Dead Sea at steady state, water gain from mean net precipitation (Ψmod) over the catchment must balance the evaporative water loss (~1.3 m y⁻¹) over the Dead Sea surface area (ADSsurf). Rapid responses of the Dead Sea level to modern droughts demonstrate that this steady state response is not greatly delayed by groundwater inertia (Enzel et al., 2003), which agrees with experiments that reveal high groundwater flow rates (Magal et al., 2010). Hence, Ψmod over Acatch in this steady state scenario needs to be of the order of 1.3/40 m per year; i.e., 33 mm y⁻¹. Using the approximation given in Frumkin et al. (2011) for modern mean gross precipitation (Pmod), which is Pmod = Ψmod/0.3, Pmod is estimated at 110 mm y⁻¹, where the difference between Pmod and Ψmod is due to evaporation and evapotranspiration. The value of 110 mm y⁻¹ determined here for mean annual precipitation over the lake’s catchment area compares well with the rainfall distribution compilations of Enzel et al. (2003, 2008), Waldmann et al. (2010) and Frumkin et al. (2011) (Fig. 1).

Evaporation from a freshwater body predominantly depends on surface-water and air temperature (and, therefore, on the lake-air temperature difference), relative humidity, and wind stress. Here I adapt the calculations of Rohling (1999) to the Dead Sea/Lake Lisan configuration. In that formulation, evaporation from low-salinity waters (in m y⁻¹) is given by:

$$E_{\text{low-salt}} = -\rho_a L C V (q_h - q_r) 1.26 \times 10^{-2}$$  \hspace{1cm} (1)

where $\rho_a$ is the air density at mean lake-level pressure (set to 1012 mbar); $L = (2500.83 - 2.34T_f) \times 10^3$ is the latent heat of vapoourisation in J kg⁻¹ at surface-water temperature $T_s$ in °C (Abbott and Tabony, 1985); $C = 1.15 \times 10^{-4}$ is an exchange
coefficient (Garrett et al., 1993; Wells, 1995); $V$ is wind speed in m s$^{-1}$; $r$ is relative humidity as a fraction between 0 and 1; $q_s$ is the saturation vapour pressure at surface-water temperature $T_a$ and $q_a$ is the saturation vapour pressure at air temperature $T_a$ (measured at 10 m above the lake surface). Values for $q$ are determined after Abbott and Tabony (1985) and Wells (1986) (for details, see Rohling, 1999).

For the Dead Sea/Lake Lisan situation, the influence of high salinity needs to be accounted for. It reduces the rate of evaporation relative to that determined for low-salinity water according to a factor $a = E(E_{\text{Lows}})$ (Salhotra et al., 1985; Saltzblatt, 1994), which depends on density but can be sufficiently approximated for the problem studied here based on the dominating salinity dependence. Comparison of the values for $a$ with salinity ($S$) in the experiments of Salhotra et al. (1985) gives a reasonable linear relationship of $a = -0.0015 S - 0.82$; $N = 8; 48 < S < 278$.

Fluctuations in the level of Lake Lisan, with amplitudes of change of order of 200 m, appear to have followed Northern Hemisphere climate variations (e.g., Bartov et al., 2003; Waldmann et al., 2010; Frumkin et al., 2011; ). At the millennial scale, those variations are portrayed by the air-temperature dominated Dansgaard-Oeschger (D-O) cycles in Greenland ice-core $\delta^{18}O$ (Dansgaard, 1964; and I.G. Grootes et al., 1995). Sigean similarity to the D-O cycles has also been reported for a number of Mediterranean sea surface temperature records (e.g., Cacho et al., 1999, 2001, 2002; Sprovieri et al., 2003; Martrat et al., 2004, 2007; Frigola et al., 2008).

Although the eastern Mediterranean climate is not likely to have exactly followed the D-O cycles, a suitable first-order experiment can be performed in which the lake-level model is driven by prescribed variations in air temperature $T_a$ and lake surface-water temperature $T_s$ that are assumed to be in direct proportion with D-O fluctuations in the NGRIP ice-core record (North Greenland Ice Core Project Members, 2004), using $T_a = T_{a\text{mod}} + f\Delta T_a$ and $T_s = T_{s\text{mod}} + f\Delta T_s$. Here $\Delta T_a$ and $\Delta T_s$ are single values that represent the total glacial-interglacial amplitudes of change in air and surface-water temperature imposed in each scenario, respectively, while $T_{a\text{mod}}$ and $T_{s\text{mod}}$ are each scenario’s modern value. The parameter $f$ is a non-dimensional scaling coefficient that imposes fluctuations in proportion with the D-O cycles: $f = (\Delta NG_{\text{mod}} - \Delta NG_{\text{Grange}}) / \Delta NG_{\text{Grange}}$ where $\Delta NG$ is the NGRIP $\delta^{18}O$ at each time step, $\Delta NG_{\text{mod}}$ is the modern value (top of the record) and $\Delta NG_{\text{Grange}}$ is the total range between the minimum glacial value and $\Delta NG_{\text{mod}}$. Changes in mean net precipitation over the wider catchment area ($P$, in m y$^{-1}$) are similarly considered, using $P = (1 + fP)P_{\text{mod}}$. Here $P_{\text{mod}}$ is the modern value for each scenario, as determined by dividing the calculated $E_{\text{mod}}$ for each scenario (i.e., $E_{\text{Lows}(\text{mod})}$ from Equation (1), times $a_{\text{mod}}$) by the ratio between the wider catchment and modern Dead Sea surface areas (i.e., $P_{\text{mod}} = E_{\text{mod}}/40$). The term $fP$ represents the total glacial-interglacial amplitude of precipitation change in fractional form; for example, $fP = 0.4$ would mean that maximum glacial $P$ was 40% greater than $P_{\text{mod}}$.

Evaporation acts over the surface area of the lake. Changes in lake surface area are calculated as a function of lake level following a simple approximation, namely that $\varepsilon = \lambda / \lambda_{\text{DSmod}} = 1.25 + 0.025H$. Here the constant approximates the area of the Dead Sea before major anthropogenic diversion of the Jordan River (relative to that of the modern Dead Sea area) at which time the Dead Sea level stood above the threshold that separates the southern Dead Sea from the main basin (Statham, 1994). $A$ is the surface area of past Lake Lisan, $\lambda_{\text{DSmod}}$ is that of the modern Dead Sea, and $H$ is the height of the past Lake Lisan level in m relative to the southern Dead Sea threshold at about 10 m above the current level. The applied relationship follows a geometric ratio within the rift topography as estimated from topographic sections of Wdowinski and Zilberman (1997). Salinities are determined by volumetric dilution of the present-day Dead Sea salinity (about 330), using lake-level dependent volume changes based on a triangular basin morphology. With a present-day mean Dead Sea depth of 118 m, relative changes in volume ($\Delta V$) are approximated by $\eta = \lambda / \lambda_{\text{DSmod}} = 1 + 0.5H/118$. The salinity effect on evaporation is determined using the relationship for $\alpha$ described above. Although the geographic approximations used here may certainly be refined using digital elevation models and hydro-isostatic modelling, the scale of such refinements remains secondary to the large uncertainties in the climatic factors that underlie lake-level changes in the geological past. Therefore, simple geographic approximations suffice for this study’s objective of a first quantitative exploration of the Lake Lisan variations through time.

Past values of $E$ are calculated by multiplying $a$ (a function of $S$) with $E_{\text{Lows}}$ from Equation (1), which uses the temporally varying climate parameters described above. These results are then used to determine bi-decadal (20-year) steady states between $E, \Psi$, and the evolving $H, S$, and $A$, relative to a constant value for $A_{\text{catch}}$. Solutions are determined by iteratively optimising $F_{\text{P}}$ so that the maximum (centennial-scale averaged) lake-level within the last glacial cycle reaches $H = 200$ m, in approximation of observations. Finally, the difference term $\Delta P$ (in m y$^{-1}$) $\Delta P = \Psi - \Psi_{\text{mod}}$ represents the change in mean net precipitation over the last glacial cycle. It comprises a component of change in gross precipitation ($\Delta P_G$), as well as a component of change in evaporation over the catchment region ($\Delta E_r$ where it must be noted that this depends on the area-weighted mean of $E_{\text{Lows}}$ over the catchment region outside Lake Lisan, and $aE_{\text{Lows}}$ over Lake Lisan); $\Delta P = \Delta P_G + \Delta E_r$. Hence, the change in precipitation can be evaluated using $\Delta E_r = \Delta P_G - \Delta E_r$. Besides $\Delta P$ and $\Delta E_r$ relative to the present, the solutions are expressed in absolute values of $P$ and $E$ by adding each scenario’s $\Delta P$ and $\Delta E_r$ values to each scenario’s $P_{\text{mod}}$ and $E_{\text{mod}}$ (about 110 mm y$^{-1}$ and 1.3 m y$^{-1}$, respectively). The entire sequence of calculations is summarised in a flowchart (Fig. 2), and a summary of key parameters is given in Appendix I.

Experiments were conducted over a range of scenarios. Values for $\Delta E_r$ are considered between $-1$ and $2.5$ m s$^{-1}$ (the range within which realistic solutions exist in which lake level reaches $+200$ m). The inter-related terms $\Delta P$ and $\Delta E_r$, are varied together, within realistic constraints to their ratio ($\Delta E_r / \Delta P$). Given that land-temperature changes are on average about 1.5 $>$ larger than marine temperature changes (Branocnct et al., 2007, 2012), that Lake Lisan has less thermal inertia than the open sea, and that lake and overlying air temperatures ultimately are strongly related, the range considered is $1.0 \leq \Delta E_r / \Delta P \leq 1.5$. Estimates for $\Delta E_r$ are obtained from regional proxy data. Soreq Cave speleothem data suggest that $\Delta E_r$ was roughly between $-8$ and $-10$ °C in the Levant (Matthews et al., 2000; McGarry et al., 2004; Afek et al., 2008). Conversely, a $\Delta E_r$ range of $-2$ to $+6$ °C with an extreme of $-9$ °C can be estimated from glacial coolings of 1 to 4 °C (Hayes et al., 2005) and 3 to 4 °C with a maximum of 5 to 6 °C (Arz et al., 2003) in nearby eastern Mediterranean and northern Red Sea surface waters, respectively (allowing that land temperature changes on average are about 1.5 $>$ larger than marine temperature changes). Combined, these results suggest a range of potential $\Delta E_r$ values of $-2$ to $-10$ °C, with most likely values in the region of $-6$ to $-9$ °C. Hence, the present study uses $\Delta E_r = -7.5$ °C for the main assessment, and subsequently considers the sensitivity of the solution to a range between $-6$ and $-9$ °C. Finally, values for $r$ differ strongly between studies, but were assessed and organised by the Dead Sea Research Team (2011) as: “the relative humidity over the Dead Sea varies between 25% to 65%; however, these extreme values are infrequent, and 95% of the relative humidity is between 33% and 52%.” Hence, the present study uses $r = 0.4$ for the main assessment, and
1. Set $T_{r, \Delta T_0/\Delta T}$ and $r$ as discussed in text.

2. Calculate $E_{\text{mod}}$ (eq. 1).

3. Initialise using $H = 0$; $S = 330$; $\varepsilon = 1$; $\eta = 1$; $\alpha = 0.505$. Use $\Psi_i = (1 + f) F_{\text{mod}}^i$ (see text). Set a seed-value for $F_{\text{mod}}$ (e.g., 4). Run model 7000 y through step 4 to stabilise (generally needs only 1000 y).

4. For ‘real’ conditions after initialisation, iterate through entire record 10 times ($\geq 3$ is enough): $\Delta H_{\text{net}} = E_{\text{net}}(t) - \Delta H_{\text{mod}} - H_{\text{mod}} - H_{\text{h}}$.

5. Repeat steps 3 & 4 $N$ times, changing $F_{\text{mod}}$ each time so $F_{\text{mod}} = F_{\text{mod}}^{i-1} + 200 (H_{\text{max}} - H_{\text{mod}} - 1)$, to optimise the solution toward $H_{\text{mod}} = +200$ m (generally $N = 4$ is sufficient).

6. Find solutions for $t = i$, the time at which the maximum highstand of the glacial cycle ($H_{\text{max}}$) is achieved:* $\Delta E_{i}^{(f)} = E_{i} - E_{\text{mod}}$ $\Delta E_{\text{lowS}}(i) = E_{\text{lowS}}(i) - E_{\text{lowS}}(\text{mod})$ $\Psi_i = \Psi_{i} - \Psi_{\text{mod}}$ $\Delta P_{i}^{(f)} = \Delta P_{i} + \Psi_{\text{mod}} - (\varepsilon_2 \Delta E_{i}^{(f)} + 40/\varepsilon_2) \Delta E_{\text{lowS}}(i) / 40$ $P_{i} = \Delta P_{i} + P_{\text{mod}}$

* [My solutions use a fixed value so $i = 74100$ y BP].

**Fig. 2.** Sequence of procedures and calculations followed in the model for Lake Lisan level reconstructions, as presented in this study.

Subsequently considers the sensitivity of the solution to a range of $r$ between 0.3 and 0.5.

Propagation of uncertainties assumes that total ranges for the various variables represent the equivalent of ±2 standard deviations (spanning about 95% of possible values), and chooses all parameters in such a way that extreme scenarios are formulated regardless of whether covariations between parameters might limit the range of possibilities in reality. The solutions then represent the ranges of precipitation change that would be coherent with evaporation assessments based on the controlling climate parameters, and lake-level change that would be coherent with reconstructions from the Lake Lisan sedimentary record. Finally, the field of solutions is compared with evidence from other reconstructions, to provide more insight can be obtained from stable oxygen isotope records from the Lake Lisan sedimentary record, although little additional proxy support can be found to independently constrain the range of possible values and their uncertainties. Instead, I here accept that the current model output does not give a perfect fit with the observations, while noting that there is sufficient agreement in the larger-scale patterns to warrant further assessment of the propagation of uncertainties and sensitivities through the model to identify the major parameter ranges within which reasonable solutions exist.

The main assessment outlined above concerns a narrow set of values for the climatic control parameters. Therefore, the maximum

between –0.2 and 0.2 m y$^{-1}$. Salinities vary between 50 during lake-level highstands at times of cold glacial conditions and 350 during the peak warmth of the early to middle Holocene. There is a clear relationship between lake-level fluctuations and the imposed climatic fluctuations, which seems almost instantaneous/in phase on the scale plotted in Fig. 3. A magnified view of the variability between 60,000 and 45,000 years ago (Fig. 4) reveals that, more precisely, the strong hydrological fluxes caused by the imposed climatic variations lead to an in-phase response of the modelled rates of lake-level change, and that the resultant lake-level record is almost a century delayed relative to the climatic variations.

Lithological observations of change (e.g., Bartov et al., 2003, and its latest version as compiled in Torfstein et al., 2013a) suggest a more smoothed response of lake level than the model (blue line in Fig. 3), as well as a pronounced response to Heinrich Events, which will be explored later. There are several potential explanations for this difference between the model-inferred and observed responses. First it should be noted that the lithological record in the Holocene denotes very rapid variability. Hence, it is not likely that the lack of reported short-term variability in parts of the glacial cycle would have arisen because the system could not respond quick enough. In other words, the strong Holocene variability argues against the existence of long-term smoothing processes in the development of Lake Lisan’s sedimentary record, which might have obscured more rapid lake-level variations. Moreover, a new interpretation of the pre-Holocene part of the record (pink line in Fig. 3, also after Torfstein et al., 2013a) begins to reveal much more fast, large-amplitude variability as well. Such developments in the observational record suggest more evidence about pre-Holocene variability may still remain to be obtained from continued detailed field studies in this intriguing region. However, it is (at least) equally likely that there is a flaw in the model’s simplification that all crucial regional climate conditions (wind speed, temperature, and net precipitation) exactly covaried with the Greenland oxygen isotope record. Finally, there are intervals with signal amplitude similarity but slight temporal offsets between the observations (pink line) and the present Greenland-based simulation (black line) in Fig. 3, which may imply that some of the discrepancies simply reflect chronological uncertainties between the Lake Lisan record and the Greenland ice-core record. All of these suggestions represent interesting avenues for further development, for example using variation of climatic control parameters according to a regional climate model, and incorporating a model for sedimentation tied to the Lake Lisan level model. Unfortunately, little additional insight can be obtained from stable oxygen isotope records from Lake Lisan, because these predominantly reflect changes in the composition of eastern Mediterranean source waters (Kolodny et al., 2005). The latter in turn depend on global sea-level changes and superimposed surface-water dilution events due to African monsoon discharge (Grant et al., 2012).

The fine-scale offsets between model results and observations might be ‘tuned’ away by inverting the calculations (i.e., determining climate variations as a function of the exactly observed lake-level record), but this becomes somewhat circular before additional proxy support can be found to independently constrain the range of possible values and their uncertainties. Instead, I here accept that the current model output does not give a perfect fit with the observations, while noting that there is sufficient agreement in the larger-scale patterns to warrant further assessment of the propagation of uncertainties and sensitivities through the model to identify the major parameter ranges within which reasonable solutions exist.
amplitude changes in the main assessment are evaluated as a function of the aforementioned ranges for control parameters $D_{T_a}$ and $D_{V}$ (Fig. 5). The panels in Fig. 5 are related according to $E = E_{\text{mod}} + E_{\Delta}$ and $P = P_{\text{mod}} + P_{\Delta}$. The first observation from Fig. 5 is that the slopes of the contours clearly identify high sensitivity of the solutions to wind speed (through its impact on evaporation), which thus appears to be the dominant environmental control parameter involved in temporal changes in the regional

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**Fig. 3.** Results for the model’s main assessment outlined in the text. From bottom to top: the three key input records $T_a$, $T_s$, and $V$; and the reconstructed rates of lake-level change, salinities, and lake-levels. The light blue line in the upper panel represents lake-level variations inferred from observations, as compiled from previous work by; see references therein). The pink line with more highly resolved variability represents a new interpretation by Torfstein et al. (2013a), which draws upon the lake-level reconstructions of Torfstein et al. (2013b). Blue and pink HE codes indicate lake level drops that were associated with Heinrich Events in the source publications. All model-based lines have been smoothed with a rectangular 200-y window, for clarity. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

**Fig. 4.** As Fig. 3, but without salinity and zoomed in on the interval from 60,000 to 45,000 years ago. No smoothing has been applied.
hydrological budget. The second observation is that there are two ‘special’ zones in the graphs of Fig. 5. First is the blacked-out, top-right corner, which marks a field where evaporation rates are so high that there are no solutions for which lake level achieves a maximum of +200 m (to understand this, note that changing evaporation rates also affect net precipitation over the catchment). Second is the hatched bottom field in the graphs, where calculated rates of mean precipitation are negative (see panel d). Although mathematically possible (‘negative precipitation’ simply being an additional evaporation term), this condition is unrealistic in the real world; it would require lake water to be advected upslope away from the lake into the wider catchment in order to there sustain the ‘negative precipitation’/evaporative loss.

There are further arguments against the plausibility of the hatched bottom field. Stable oxygen isotope calculations suggest that glacial winds may have strengthened over the Red Sea (Rohling, 1994) and were constant or slightly stronger over the eastern Mediterranean, relative to the present (Rohling, 1999). Enzel et al. (2003, 2008) and Frumkin et al. (2011) also suggested that any precipitation increase over the Levant during the last glacial would have been related to enhanced intensity/frequency of Cyprus Lows, which would lead to increased mean wind speeds.

Fig. 5. Dependence of the solution for $H_{\text{max}} = +200$ m from the main assessment to realistic ranges for the critical control parameters $\Delta T_a / \Delta T_s$ and $\Delta V$, as outlined in the text. Blacked-out and grey hatched zones are explained in the ‘results’ section. All panels show contour values in m yr$^{-1}$. Bold line indicates no change relative to the present.

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**Fig. 5.** Dependence of the solution for $H_{\text{max}} = +200$ m from the main assessment to realistic ranges for the critical control parameters $\Delta T_a / \Delta T_s$ and $\Delta V$, as outlined in the text. Blacked-out and grey hatched zones are explained in the ‘results’ section. All panels show contour values in m yr$^{-1}$. Bold line indicates no change relative to the present.
Dashed lines connect probability maxima and 95% confidence limits. Uncertainty in \( \Delta T_a \), of the implications of uncertainty in the ratio \( \Delta T_a/\Delta T_b \), for the inferred \( \Delta V \) values. Uncertainty in \( \Delta T_a/\Delta T_b \) is captured along the X-axis using a normal distribution with 95% probability within the bounds of 1.0–1.5 (peak of the probability curve is set to an arbitrary level for clearest illustration). Projection of the \( \Delta T_a/\Delta T_b \) probability distribution onto the model solution gives the probability distribution for \( \Delta V \) along the Y-axis, which becomes clearly skewed with an extended tail to high values. curved lines connect probability maxima and 95% confidence limits. Illustration of additional uncertainty in \( \Delta V \) due to 2e (95%) equivalent uncertainties in \( r (\Delta T_a) \) and \( \Delta V \). Root mean square addition of uncertainties from \( \Delta T_a \) gives \( \Delta V = 1.10 (95\%) \times 0.55 \text{ m s}^{-1} \) (95% probability).
assess the implications of this estimate, the 500 mm y\(^{-1}\) contour from Fig. 5d is singled out in Fig. 7a. Next, a normal distribution is formulated for \(\Delta V / \Delta T\) using a mean of 1.25 and a 2\(\sigma\) (95\% probability) window of \(\pm 0.25\), based on the range arguments presented before. This normal distribution is projected onto the \(P = 500\) mm y\(^{-1}\) line to derive the corresponding estimates for \(\Delta V\), which gives \(\Delta V = 1.10 \times 0.73 / 0.42\) m s\(^{-1}\) (95\% probability).

In addition to the uncertainties already accounted for, the inferred probabilities for glacial \(\Delta V\) during peak lake-level highstands also depend on uncertainties in \(r\) and \(\Delta T\). In other words, realistic ranges of \(r\) and \(\Delta T\) need to be accounted for; as discussed above, these amount to \(r = 0.4 \pm 0.1\) and \(\Delta T = -7.5 \pm 1.5\)°C (2\(\sigma\) equivalent). These ranges produce extra uncertainty in the inferred values of \(\Delta V\) that amounts to \(\pm 0.36\) m s\(^{-1}\) (Fig. 7b). Root mean square addition gives a final estimate for glacial \(\Delta V\), namely \(\Delta V = 1.10 \times 0.81 / 0.55\) m s\(^{-1}\) (95\% probability). The observations of Lake Lisan levels up to 200 m above the Dead Sea level, along with \(P = 500\) mm y\(^{-1}\), are therefore found to be quantitatively consistent if glacial wind speeds were significantly (beyond 95\% uncertainty) increased relative to their present-day mean value of about 4 m s\(^{-1}\).

Finally, the model provides insight into the sharp drops in Lake Lisan level that have been reported for the coldest millennial-scale events during the last glacial cycle that coincided with North Atlantic Heinrich Events (Bartov et al., 2003; Torfstein et al., 2013a,b). These contrast sharply with the ‘normal’ pattern of high lake levels during cold periods. Fig. 5d illustrates that, close to (but not inside) the blacked-out zone, a given period of time with relatively high wind speeds and/or high \(\Delta V / \Delta T\) ratios may be marked by high lake levels, as long as precipitation rates were high. However, if (predominantly) windspeed were to increase further, then the solution would shift into the blacked-out zone. In that zone, evaporation rates are so large that lake levels drop sharply, regardless of high precipitation rates. To evaluate the potential impacts, the main assessment’s inferred record of \(V\) has been adapted by manually replacing its values of about 4.6 m s\(^{-1}\) in the interval of Heinrich Event 1 (taken as 17,450–16,250 years ago to match the timing of the observed lake-level drop) by values of 7.5 m s\(^{-1}\) (i.e., \(\Delta V(HE1) = 3.5\) m s\(^{-1}\)) and subsequently calculated lake levels are compared with the main assessment (Fig. 8). Although this simply imposed, large change represents an artificial experiment, it clearly highlights exceptional wind speeds as a key candidate process for explaining the observed lake-level drops during Heinrich Events.

4. Conclusions

Although the full intricacies of D–O scale variability in Lake Lisan level are not yet reproduced with the model as is, it already demonstrates that increased windiness (and thus evaporation) must have been an important component of environmental changes around Lake Lisan if realistic bounds on precipitation increase are considered from the compiled data. For glacial rainfall reaching 500 mm y\(^{-1}\), a value inferred from the literature (Enzel et al., 2008; Frumkin et al., 2011), the estimated increase in mean wind speed is 0.6–1.9 m s\(^{-1}\) (95\% probability). Wind-speed increase over the region is consistent with the development of glacial windblown dust (loess) deposits along the northern Negev Desert margin (Enzel et al., 2008).

The results support a notable increase in (winter) storminess that caused both precipitation and increased winds, as was qualitatively suggested before from geological and geomorphological evidence (Enzel et al., 2003, 2008; Frumkin et al., 2011). The present study therefore provides quantitative support for their proposed mechanism of a stronger west-east “funnelling of Mediterranean cyclones into the central Levant, replacing the moister and semi-arid Mediterranean zone with a wetter temperate climate.” This mechanism also is consistent with evidence of enhanced atmospheric instability over the central Mediterranean during glacial times, with intensified formation of Mediterranean depressions due to cold polar/continental air outbursts through gaps in the mountain ranges to the north of the basin (Kuhlemann et al., 2008).

During Heinrich Events, which were marked by regional cooling, the level of Lake Lisan dropped sharply (Bartov et al., 2003; Torfstein et al., 2013a,b), in contrast with the ‘normal’ association between cold conditions and high lake levels during the last glacial cycle. The present study identifies exceptional wind speeds as a key candidate process for explaining these drops at times of Heinrich Events. Distinct increases in wind speed at times of Heinrich Events agree with other proxy evidence from the Mediterranean and Red Sea region (e.g., Rohling et al., 1998; Moreno et al., 2002; Larrasoña et al., 2003; Enzel et al., 2008; Roberts et al., 2011).

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Appendix I. Key parameters used in the calculations.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
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<tbody>
<tr>
<td>(A)</td>
<td>area of Lake Lisan</td>
</tr>
<tr>
<td>(A_{\text{catch}})</td>
<td>area of catchment basin</td>
</tr>
</tbody>
</table>
References
changes in the western Mediterranean over the past 250,000 years. Science 306, 1762–1765.


