Non-hydrostatic effects in the Dead Sea

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Abstract

The Dead Sea is the saltiest and lowest terminal lake in the world. For at least three decades, the Dead Sea’s water level has been dropping by more than 1 m per year, due to excessive use of the water that previously flowed into it. The Dead Sea constitutes a unique environment and is important from economic, environmental, and touristic points of view. The winter deep convection of the Dead Sea and its deep and narrow basin suggest that non-hydrostatic effects may significantly affect its circulation. Despite these factors, the expected non-hydrostatic effects on the circulation of the Dead Sea have not been investigated. Here we perform high resolution (100 m) ocean general circulation model (the MITgcm) simulations of the Dead Sea and show that the non-hydrostatic results are very different from the hydrostatic ones. Specifically, we show that the winter non-hydrostatic simulations resulted in a layer of dense water overlaying slightly lighter water during the several last hours of the night; this convection process involved plumes of heavier sinking water and the entrainment of the plumes. We also studied the effect of the wind stress’s diurnal variability and found it to be important, especially during the summer when the wind’s variability drastically increased the surface kinetic energy; however, it did not alter the depth density profile. The results presented here may be important for the Dead Sea’s potash industry and for the planned Red Sea-Dead Sea canal that aims to stop and, possibly, to increase the level of the Dead Sea using the Red Sea’s water.

Keywords: Dead Sea, Non-hydrostatic, circulation

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1. Introduction

The Dead Sea is the saltiest and lowest natural water body in the world (Hall, 1978). It is an elongated (on the south-north axis) terminal lake located between Israel and Jordan, along the Great Syrian-African Rift. The Dead Sea surface is currently more than 430 m below mean sea level, and its deepest point is \( \sim 730 \) m below mean sea level, such that its maximal depth is currently \( \sim 300 \) m. The current dimensions of the Dead Sea are approximately 20 km \( \times \) 50 km. The Dead Sea is located in an extremely arid region with an average precipitation rate of \( \sim 5 \) cm yr\(^{-1}\). Its main water source is the Jordan River, whose current runoff is equivalent to \( \sim 10 \) cm yr\(^{-1}\) (Lensky et al., 2005).

The salinity of the Dead Sea’s water is about eight times larger than standard ocean water salinity (Steinhorn, 1985, 1991; Gertman and Hecht, 2002), reaching values of \( \sim 280 \) ppt. This high salinity is due to the fact that the Dead Sea is a closed basin without outlets, so that natural water loss from the Dead Sea only occurs via evaporation or human usage (such as water pumping for the potash industry). The current evaporation rate exceeds the precipitation and river runoff rates, such that the Dead Sea’s salinity increases with time, and the sea surface level decreases with time by more than 1 m per year (Hecht et al., 1992; Yechieli et al., 1998; Gertman and Hecht, 2002). This also leads to what is known as salt precipitation in the deep water, with salt accumulation at the bottom of the Dead Sea occurring at a rate of \( \sim 10 \) cm yr\(^{-1}\) (Lensky et al., 2005).

For the last several centuries until the mid-1970s, the Dead Sea was stratified throughout the year, in a state known as meromictic stratification (Neev and Emery, 1967; Anati, 1997) in which the water column is not mixed throughout the year. During winters, the surface water became colder, but since it was relatively fresh, it was not dense enough to sink to the bottom of the lake. In the 1970s, the Dead Sea experienced a drastic change in its circulation, as the meromictic stratification was replaced by total vertical mixing during the winter (Steinhorn and Gat, 1983; Steinhorn, 1985; Anati et al., 1987; Gertman and Hecht, 2002; Steinhorn et al., 1979), thus switching to its current holomictic state in which water layers are mixed at least once a year.

Increasing human use of Lake Kinneret waters (the Sea of Galilee, located...
about 100 km north of the Dead Sea, along the African-Syrian Rift) has decreased
the Jordan River runoff to the Dead Sea, making its surface water saltier, and
thus reducing the salinity gradient between the surface water and the deep water.
Moreover, water from the Dead Sea’s main northern part is pumped to the sea’s
former southern part that now contains evaporation ponds used for the potash
industry. During cold winter nights, the surface water becomes colder, and since
the surface layer has become saltier over the years, the surface water is now denser
than the deep water, leading to deep vertical mixing. Since the 1970s, except
for in years with exceptionally high precipitation (such as the winter of 1991-
1992), the Dead Sea’s water column experiences total mixing every winter. This
process of vertical mixing by the sinking of cold surface water is associated with
non-hydrostatic dynamics.

The Dead Sea is currently of local and global interest, due both to the economic
importance of the Dead Sea Works (the factories that deal with the potash industry
of the Dead Sea) and to tourism associated with the Dead Sea, as well as to the
planned Red Sea-Dead Sea conduit. This conduit (Beyth, 2007; The World Bank,
2013) aims to transport water from the Gulf of Eilat/Aqaba to the Dead Sea, in
order to stop the drop in the Dead Sea’s level (and possibly restoring it to its higher
1970s level), and to use the ∼430 m elevation difference between the Gulf of Eilat
and the Dead Sea as an energy source for water desalination. The ocean water
imported through the planned conduit is expected to have a significant impact
particularly, the imported water may affect the overall circulation, the surface
water conditions, the water chemistry and the ecological system surrounding the
Dead Sea, and may have severe implications for the Dead Sea’s environment and
shores. Thus, understanding the dynamics of the Dead Sea and predicting the
effects of the Red Sea-Dead Sea conduit are of major importance from both local
and global perspectives.

The Dead Sea’s circulation has been studied using both simple models (Vadasz
et al., 1983; Asmar and Ergenzinger, 2002, 2003) and more complex general circu-
lation models (GCMs) (Ezer, 1984; Sirkes, 1986), and has been simulated mainly
using 1D and 2D models. Modeling the impact of the high density of the Dead Sea
water on the circulation show that the near-surface wind-driven currents in the
Dead Sea are \(\sim 15\%\) weaker than currents in standard seawater under the action of the same wind (Huss et al., 1986). These simulations were not long enough to study the dynamics incurred by the annual cycle and were too coarse to study the non-hydrostatic effects of the circulation. The Dead Sea’s wind-driven seiches were studied by Sirkes (1986) using a seiche model and a simple oceanic CGM, under the assumptions of homogeneous temperature and salinity fields. To our knowledge, the Dead Sea’s circulation has not been studied using a non-hydrostatic GCM, even though non-hydrostatic effects are expected to be significant in this circulation, due to winter mixing (deep convection) of the water column and due to the deep and narrow shape of Dead Sea; see Eq. (2) of Marshall et al. (1997a). The goal of the present study is to investigate whether and how non-hydrostatic effects impact the Dead Sea’s properties and circulation. Using a high resolution oceanic GCM, the MITgcm (MITgcm-group, 2010), we show that the non-hydrostatic dynamics resulted in a Dead Sea circulation that is much different than the hydrostatic one. This suggest that non-hydrostatic effects must be taken into account in order to more accurately simulate the Dead Sea’s circulation. Accurate simulations of the lake’s circulation and salinity distribution are important for optimal management of the potash industry and for the planned Red-Sea-Dead-Sea canal.

Previous studies had used the MITgcm (and other GCMs) to compare between hydrostatic and non-hydrostatic simulations to study, for example, non-hydrostatic effects on the development of surface ocean fronts (Haine and Williams, 2002), entrainment processes of water flowing down a slope (Legg et al., 2006), the Strait of Gibraltar exchange flow (Sannino et al., 2014), East Greenland dense water flow (Magaldi and Haine, 2015), and coastal north Adriatic Sea vertical processes (McKiver et al., 2016).

Convection processes are dominant during the winter, at which heavy water (resulted from surface nocturnal cooling) sinks to the bottom. Convective processes are naturally simulated through the (full) vertical momentum equation without the need of special convection parameterization. Water plumes and entrainment of water are associated with convective processes. However, under the hydrostatic approximation the vertical momentum equation is reduced to the hydrostatic balance and there is a need to parameterize convection processes. There are several ways to handle convection under the hydrostatic approximation like (MITgcm-
group, 2010): (i) convective adjustment, (ii) much larger vertical diffusion when
the water column becomes stable (e.g., the implicit vertical diffusion scheme of
MITgcm), and the (iii) K-profile-parameterization (KPP) (Large et al., 1994) in
which the vertical diffusion coefficient is determined according to the stability
and vertical density structure of the water column; basically, also for the KPP,
the vertical diffusion coefficient is drastically increased when the water column
destabilized. Still, increased vertical diffusion cannot reproduce the complicated
convection process and the water plumes and entrainment associated with it.

A key question is when and at what (horizontal) scale one may expect to observe
non-hydrostatic effects. When the grid is too coarse, non-hydrostatic effects are
weak and poorly represented by the vertical momentum equation; in this case
there is a need to use convection parameterizations. According to Marshall et al.
(1997b) and Marshall and Schott (1999) non-hydrostatic effects (such as plumes)
are expected when the horizontal scale is less than 1 km. To ensure that we
indeed resolve non-hydrostatic effects in the Dead Sea we used a much finer grid
of 100 m. We note that simulation of the Dead Sea using the non-hydrostatic
configuration, without convection parameterization and using 400 m and 200 m
horizontal resolutions also yielded non-hydrostatic effects.

In this paper, we first describe the model, setups, and the experiments (Sec.
2). We then present the results for annual cycle forcing (Sec. 3), for the winter
diurnal cycle (Sec. 4), and for summer and winter diurnal cycles under fixed and
diurnally varying winds (Sec. 5). We then summarize and conclude the paper
(Sec. 6).

2. Model Setup

2.1. The MITgcm

The Massachusetts Institute of Technology general circulation model (MIT-
gcm) (MITgcm-group, 2010) is a widely used, open source GCM, capable of sim-
ulating the dynamics of both the atmosphere and the ocean, separately or in a
coupled mode (Marshall et al., 1997a; Adcroft et al., 2002). It is a z-coordinate
model, and its main properties are: (i) hydrostatic and full non-hydrostatic (and
non-Boussinesq) capabilities, (ii) a finite volume numerical model with partial
cells, and (iii) adjoint modeling. In particular, the support for non-hydrostatic
dynamics allows the model to be used to study both large-scale and global dy-
namics, and small-scale processes. It also has many parameterizations to account
for various oceanic processes. In the current study we use the hydrostatic and
full non-hydrostatic options of MITgcm together with the partial cells option, to
account for the steep bathymetry of the Dead Sea.

2.2. Setup

To simulate the Dead Sea’s circulation, we used a cascade of three computa-
tional grids, with horizontal grid resolutions of 400 m, 200 m and 100 m, as detailed
in Table 1. We used 15 levels in the vertical direction with variable resolution as
follows (top to bottom): 1, 1.5, 2, 3, 4, 5, 7, 10, 15, 20, 30, 40, 50, 55, 61 m;
the spacing between the vertical levels was chosen to allow a better simulation of
dynamics near the surface, which are affected by temperature and salinity forcing
and by the wind stress. The 400 m simulation was initialized with the water at rest
and with uniform temperature and salinity fields matching the surface conditions,
roughly mimicking winter conditions. This simulation was run under annual cycle
(monthly mean) forcing of surface temperature and salinity and constant northerly
winds: it was run for twenty years to a quasi repeating annual cycle, representing
the Dead Sea’s circulation under annual cycle forcing.

The state of the 400 m simulation after twenty years was used as the initial
condition for the 200 m simulation (using spatial interpolation). This simulation
was run with the same annual cycle and constant winds as the 400 m simulation
and for ten years, or until reaching a steady state, representing a higher resolution
annual cycle circulation. The (quasi) steady states of the 200 m simulations were
then used as initial conditions for the 100 m simulations, which were run with (i)
annual cycle (monthly mean) forcing for more than three years, and with a diurnal
cycle in temperature, constant salinity matching the winter and summer seasons,
and both (ii) constant and (iii) varying winds; see Table 1.

The described process of cascading from the 400 m to the 100 m simulation was
repeated using both hydrostatic and non-hydrostatic modes, to compare between
the two. In all hydrostatic simulations, the K-profile-parameterization (KPP)
(Large et al., 1994) scheme was used; we have also used an alternative (simpler)
vertical diffusion scheme, the implicit vertical diffusion scheme of MITgcm (using different vertical level spacing) and obtained similar results; no convective parameterization were used in the non-hydrostatic simulations.

Under regular stable conditions, the vertical diffusion coefficient of temperature and salinity was chosen to be $10^{-5} \text{ m}^2 \text{s}^{-1}$, which is a typical value for open ocean eddy parameterized diffusion coefficient (Wunsch and Ferrari, 2004). The horizontal diffusion coefficient was chosen according to the horizontal grid resolution and was ranging from $4 \text{ m}^2 \text{s}^{-1}$ for the 400 m horizontal resolution runs to $1 \text{ m}^2 \text{s}^{-1}$ for the 100 m horizontal resolution runs. The vertical viscosity coefficient was chosen to be $5 \times 10^{-4} \text{ m}^2 \text{s}^{-1}$ and the horizontal viscosity coefficient varied according to the horizontal grid resolution, from $4 \text{ m}^2 \text{s}^{-1}$ for the 400 m horizontal resolution runs to $1 \text{ m}^2 \text{s}^{-1}$ for the 100 m horizontal resolution runs.

The numerical simulations involve the solution of 2D and 3D elliptic equations, at every time step—these are solved iteratively until reaching a maximum number of iterations or until the residual value drops below a target value. In all simulations, the 2D solver target residual was $10^{-13}$, and in the non-hydrostatic simulations, the 3D solver target residual was $10^{-9}$. Both solvers were allowed enough iterations to reach the target residual. The time step used was adjusted to the horizontal resolution, and is listed in Table 1; in all runs the time step was small enough to satisfy the (advection based) CFL condition and the Courant number was usually much smaller than 1. Realistic, high resolution (100 m) Dead Sea bathymetry (Fig. 1) was used, based on the 25 m resolution data of Hall (1978); we have used a coarser bathymetry resolution to fit the resolution of the different experiments.

In the simulations presented below we have used the nonlinear free surface option of MITgcm. In addition, we used the nonlinear flux limiter advection scheme (advection scheme 33) of MITgcm.

Standard equations of state are invalid for the Dead Sea due to its extremely high salinity. Instead, we used a linear equation of state, based on Steinhorn (1991) and Ezer (1984):

$$\rho = \rho_0 [1 - \alpha (T - T_0) + \beta (S - S_0)]$$

where $\rho_0 = 1237 \text{ kg m}^{-3}$, $\alpha = 3.5 \cdot 10^{-4} \text{ C}^{-1}$, $\beta = 9.5 \cdot 10^{-4}$, $T_0 = 23 \text{ °C}$, and
$S_0 = 278$ ppt. We note that the thermal and haline coefficients, $\alpha, \beta$, are different than the ones used in Steinhorn (1991) and Ezer (1984), to match the saltier water (corresponding to the simulations) of the Dead Sea where under (future) conditions the value of these parameters may be different.

In all simulations, surface forcing took the form of temperature and salinity restoring, and prescribed wind stress. We use restoring boundary conditions instead of surface fluxes as the former can be estimated more easily from observations. We note that the use of heat and freshwater surface fluxes often yields more realistic mixed layers (e.g., Ezer, 2000), although the use of such boundary conditions is problematic in the Dead Sea as the freshwater flux is significantly negative, disabling the convergence to a quasi-steady state. The salinity and temperature forcing formulations used was a sine wave, given by:

$$T(t) = \frac{T_{\text{min}} + T_{\text{max}}}{2} - \frac{T_{\text{max}} - T_{\text{min}}}{2} \cos \left(2\pi \frac{t}{t_{\text{cycle}}} \right)$$  \hspace{1cm} (2)

$$S(t) = \frac{S_{\text{min}} + S_{\text{max}}}{2} - \frac{S_{\text{max}} - S_{\text{min}}}{2} \cos \left(2\pi \frac{t}{t_{\text{cycle}}} \right)$$  \hspace{1cm} (3)

where $t_{\text{cycle}}$ is one year in the annual cycle simulations, and one day in the diurnal cycle simulations. The values of $T_{\text{min}}, T_{\text{max}}, S_{\text{min}}, S_{\text{max}}$ are based on measurements performed by the IOLR (Israel Oceanographic and Limnological Research) (IOLR, ISAMAR, 2009), and are listed in Table 1. The restoring times of the forcing salinity, $S(t)$, and temperature, $T(t)$, were 2.4 days for surface salinity and 0.8 days for surface temperature (see Tziperman et al., 1994); using 5 time larger restoring times yielded, as expected, smaller and unrealistic seasonal variations. The influx of fresh water from the Jordan river and winter flash floods are ignored here and indirectly considered through the relaxation to fresher sea surface during the winter.

The wind stress used in the constant wind simulations matches a northerly wind of $4.5 \text{ m s}^{-1}$, which is a typical wind for the Dead Sea area; the 2007-2011 mean wind speed at Israel Meteorological Service’s Ein Gedi station (located in the middle of the Dead Sea’s western coast) is $\sim 4.8 \text{ m s}^{-1}$. Based on Gill (1982),
we applied the following conversion from wind velocity, $\mathbf{v}_{\text{air}}$, to wind stress, $\mathbf{\tau}$:

$$\mathbf{\tau} = \rho_{\text{air}} C_D \mathbf{v}_{\text{air}} |\mathbf{v}_{\text{air}}|,$$

(4)

where $C_D = 10^{-3}$ when $|\mathbf{v}_{\text{air}}| < 6.2 \text{ m s}^{-1}$ and $C_D = 10^{-3} (0.61 + 0.063 \cdot |\mathbf{v}_{\text{air}}|)$ when $|\mathbf{v}_{\text{air}}| \geq 6.2 \text{ m s}^{-1}$. While Eq. (4) is standard in the world ocean it may be less accurate in the Dead Sea due to its unusual density, viscosity and waves (Hecht et al., 1997). In the varying wind simulations, data from the Ein Gedi station for the years 2006 – 2010 was used to calculate the climatological mean diurnal cycle of the wind stress for both winter (January) and summer (July). The resultant diurnal cycle of the wind stress is presented in Fig. 2.

3. Results of Annual Cycle Forcing

As described in Sec. 2.2 and Table 1, the annual cycle of the Dead Sea was simulated using a horizontal grid resolution of 400 m, starting from uniform initial conditions (resembling winter conditions) and run for twenty years to a steady annual cycle; the results of this simulation were used as the initial conditions for 200 m simulations. These were run for an additional ten years (or quasi steady state) and were used as initial conditions for the 100 m simulations. The 100 m resolution run were first run for more than three years under annual cycle (monthly mean) forcing and constant winds (see above). Fig. 3 depicts the sea surface temperature (SST) and velocity field snapshots from the 100 m simulation at the end of the runs, in summer and winter, for both the hydrostatic and non-hydrostatic simulations. As this figure shows, the hydrostatic and non-hydrostatic results are similar, to some degree, during the summer while they are much more different during the winter. We will show below that the difference between the winter non-hydrostatic and hydrostatic simulations is much larger when considering the diurnal cycle—this difference is attributed to convection processes that are more accurately simulated when using non-hydrostatic configuration.

Fig. 4 shows the density, salinity, and temperature fields of the 100 m non-hydrostatic run (annual cycle, monthly mean, forcing), averaged over the entire lake in the horizontal directions, presented in the depth-time plane. The results of the hydrostatic simulations are similar to the non-hydrostatic ones, suggesting
that the non-hydrostatic effects affect mainly the surface layers and mainly the
model’s velocities. The difference between the non-hydrostatic and hydrostatic
simulations become much more pronounced under winter, diurnal cycle conditions;
see below. The profiles are uniform below a certain mixing layer. The depth of the
mixed layer varies between 20 m in summer to the entire water column (complete
mixing of the water column) during winter. The surface layer becomes deeper as
the surface water becomes colder and thus denser until at some point during the
winter, the mixed layer disappears and total mixing occurs. This state persists for
approximately two months in the winter, before the mixing layer forms again as the
surface water becomes warmer and lighter. The results described above resembles
observations from the Dead Sea reported in IOLR web-page, and in other studies
(e.g., Anati, 1997; Gertman and Hecht, 2002; Gertman, 2012); see Fig. 5 for a
comparison between observations and model’s results.

A useful measure of the flow is the stream function, $\psi$. We calculated the
stream function (that reflects the overturning circulation) in the Y-Z plane
\[
\psi(y, z) = \int_{-H}^{z} V(y, z') \, dz',
\]
where $V = \int v \, dx$ (i.e., integration over the entire width of the lake). The stream
function, $\psi$, indicates the net northward volume flux through the X-Z plane located
at $y$ and stretching from the bottom up to $z$.

The overturning circulation in the Dead Sea peaks at February and then drasti-
cally relaxed during March. In Fig. 6 we present the stream function from the 100
m simulations, of both the hydrostatic and non-hydrostatic runs during these two
months. The February stream function basically shows 1-2 circulation cells with a
counterclockwise flow; the non-hydrostatic circulation is different than the hydro-
static one were the former circulation extend over the entire lake. The southward
flow near the surface corresponds to the action of the northerly winds. The March
results show the counterclockwise cell compressed to the top 20 m near the surface,
with a clockwise circulation cells occupying the deep water; the abyssal lake cir-

ulation is counterclockwise. Also here the non-hydrostatic results are similar, to
some degree, to the hydrostatic ones; we will show below that the non-hydrostatic
dynamics is very different from the hydrostatic dynamics, when considering the
lakes surface and when using the winter conditions diurnal cycle. Note that as seen in Fig. 4, in February, the entire water column is mixed, while in March, the mixing layer is forming, occupying approximately the top 20 m near the surface. The transport during February is about six times larger than during March. Following the above, we conclude that the differences between the hydrostatic and non-hydrostatic simulations for the annual cycle (monthly mean) 100 m resolution case are moderate, either in the Dead Sea’s circulation or in its properties (i.e., temperature, salinity, and hence density). Therefore, we proceeded to study possible non-hydrostatic effects when the diurnal cycle is taken into account.

4. Results of Winter Diurnal Cycle Forcing

As described in Sec. 2.2, the final states of the 200 m simulations in both winter and summer were used as input for the 100 m simulations with a diurnal cycle in winter and summer, and with both constant and time varying wind stress. In this section, we focus on the diurnal cycle in winter, with constant winds, and compare the hydrostatic results to the non-hydrostatic ones. We expected to see a marked difference due to the presence of significant vertical convection: during cold winter nights, the surface water becomes colder and denser than the deep water, which results in convective instability, leading to non-hydrostatic vertical convection. A spatial resolution of 100 m (Marshall et al., 1997b) should suffice to simulate most of the convection events in the Dead Sea.

The diurnal cycle simulations were run for thirty days, and we present the results of the last five days of the simulations. Thirty days proved to be enough to switch from the 200 m resolution to the 100 m resolution and from the annual cycle to the diurnal cycle. We note that in contrast to the annual cycle simulations with the seasonal forcing, it did not make sense to run the simulation until it reached a steady state, as a perpetual January or a perpetual July until a steady state is reached is not representative of actual circulations (as opposed to a perpetual annual cycle until reaching steady state), which are (seasonally) transient by their nature.

The horizontal mean (i.e., mean over the x and y directions) versus depth and time of the density, salinity, and temperature of the hydrostatic and non-hydrostatic runs are presented in Fig. 7. Only the top 14 m of the water column is
shown, to allow better visualization of the difference between the non-hydrostatic
and hydrostatic simulations; below this depth, the water properties are much more
uniform. Both figures clearly present the effect of the diurnal cycle on the top water
layer near the surface, which becomes colder and denser during the nights, and
warmer and lighter during the days.

While both the hydrostatic and the non-hydrostatic simulations presented the
effect of the diurnal cycle, there was a substantial difference between them. In the
hydrostatic results (top three panels of Fig. 7), the surface water got colder and
denser during the night, but it was only slightly denser than the water below the
surface, since the surface water very fast mixed once it became denser than the
water below it. In the non-hydrostatic run (bottom three panels of Fig. 7), the
surface water did become colder and denser than the water below it during the
night, but this state persisted until the morning. Moreover, the nocturnal SST of
the non-hydrostatic run was much colder (by $\sim 0.5^\circ\text{C}$) than that of the hydrostatic
run as a result of its slower sinking. However, neither the hydrostatic nor the
non-hydrostatic simulations presented in Fig. 7 shows complete mixing during
the night as we would expect, but this is a result of the horizontal averaging over
the entire lake when, in fact, the density (depth) profile varies from one location
to another. The difference between the non-hydrostatic and hydrostatic results
may be attributed to water plumes and entrainment processes that are present in
the non-hydrostatic simulations while absence in the hydrostatic simulations. As
we shall show below, the actual density profiles do present total mixing during
the night. [We have estimated the Monin-Obukhov scale (e.g., Shay and Gregg,
1986) of the non-hydrostatic run during the winter time and found it to increase
from slightly more than 1 m at midnight to tens of meters toward the morning.
Thus, the model’s vertical resolution resolves the Monin-Obukhov scale, especially
during deep convection that occurs a few hours after midnight.]

The left two panels of Fig. 8 show the SST and velocity field snapshots at 4 am,
for both the hydrostatic and non-hydrostatic simulations—the hydrostatic results
are significantly different from the non-hydrostatic ones. The surface water was
colder in the non-hydrostatic simulation, as already indicated in Fig. 7, and the
spatial features of both the temperature and velocity fields were different from the
corresponding hydrostatic fields; these differences are attributed to non-hydrostatic
convection occurring at this time of night.

As suggested by Fig. 7, the most apparent difference between the hydrostatic
and non-hydrostatic simulations was the different dynamics of the surface water
density relative to the deep water density. For this reason, we studied the density
difference between the surface (upper layer) and deep water, taken as the mean
density below 100 m. The right two panels of Fig. 8 show a snapshot of this
density difference at 4 am (approximately the time of minimum SST), of both the
hydrostatic and non-hydrostatic simulations, with the contour of $\Delta \rho = 0$ overlayed.
This figure demonstrates a significant difference between the hydrostatic and non-
hydrostatic simulations. As seen, in the non-hydrostatic simulation, the lake’s
surface water was denser than the deep water by about 0.22 kg m$^{-3}$, equivalent to
a relative density difference of $\Delta \rho / \rho_0 \sim 1.8 \times 10^{-4}$. For the hydrostatic case, the
surface water was also almost everywhere heavier than the deep water (except for
a very small region along the western coast of Dead Sea). However, the density
difference was less than 0.009 kg m$^{-3}$, i.e., a relative difference of $\Delta \rho / \rho_0 < 7 \times 10^{-6}$;
i.e., the density difference in the non-hydrostatic run was about 25 times larger
than that of the hydrostatic run. Thus, the non-hydrostatic simulation showed a
much unstable state of heavy water above light water compare to the hydrostatic
simulation.

To visualize the vertical convection process, we show in Fig. 9, a cross-section
of the density field along the X-Z plain at $Y = 30$ km, from the 4 am snapshot,
for both the hydrostatic and non-hydrostatic simulations. This figure also shows
dramatic differences between the hydrostatic and non-hydrostatic results. In the
non-hydrostatic case, the surface water was denser than the water below it (due to
atmospheric cooling), and it sank in plumes that are typical for vertical convection
(e.g., Marshall et al., 1997b). This is very different from the hydrostatic results,
in which no such plumes existed.

We next show the density profile at a selected point in the middle of the lake (at
$X = 10$ km, $Y = 30$ km) versus depth and time (Fig. 10), with a vertical line that
indicates the time of the snapshot presented in Figs. 8, 9. Also here, there were
dramatic differences between the hydrostatic and non-hydrostatic simulations. As
before, in the non-hydrostatic simulation, the surface water became slightly heavier
than the deep water (a relative density difference of $\approx 1.8 \times 10^{-4}$), unlike the
much smaller difference for the hydrostatic simulation. In the non-hydrostatic simulation, the surface water started to become denser than the deep water around midnight and became denser with time due to surface nocturnal (night) cooling; as a result, the mixed layer deepened, and after several hours (around 7 am), it sank to the bottom, resulting in total mixing of the water column. Later in the day, the surface water became lighter (as it is heated), resulting in a shallow (∼2 m) surface layer of warmer and lighter water, persisting until the next night. In the hydrostatic simulation, the behavior during the day was similar, but during the night, it was very different: the sinking occurred as soon as the surface water became heavier than the water beneath it. In this case, total vertical mixing started around midnight and persisted until a little before midday.

To present this difference more clearly, we plotted $\Delta \rho = \rho_{\text{surface}} - \rho_{\text{deep}}$ versus time, at the point located at $X = 10$ km, $Y = 30$ km (Fig. 11), the same location used in Fig. 10. The major difference between the hydrostatic and non-hydrostatic simulations occurred during the night. In the non-hydrostatic simulation, $\Delta \rho$ demonstrated a diurnal oscillation between negative values during the days and positive values during the night, roughly with the same magnitude up and down. In the hydrostatic simulation, on the other hand, $\Delta \rho$ demonstrated a cut-off near $\Delta \rho = 0$, with much smaller positive (unstable) density difference. We have also examined the vertical velocity at the same location as above (at depth of 1m, not shown) and found much larger variability for the non-hydrostatic run (approximately between $-10^{-3}$ to $10^{-3}$ m/s) in comparison with hydrostatic run (approximately between $-10^{-4}$ m/s to $10^{-4}$ m/s).

Based on the above (Figs. 9, 10) it is possible to roughly estimate the temporal and spatial scales associated with convective processes in the Dead Sea. From Fig. 9, the estimated spatial scale of the convection process is a few hundreds meters. From Fig. 10, the estimated temporal scale of the plumes is a few hours. Consequently and from Fig. 10, the frequency of the convective occurrences is 2-3 events per night at a given point.

The major discrepancy between the hydrostatic and non-hydrostatic runs, presented in Figs. 7–11, can be explained by the different ways that the vertical convection was simulated in the two modes. In the hydrostatic simulations, we used the KPP scheme which caused any unstable vertical density gradients to dif-
fuse and disappear rapidly. Essentially, this means that the hydrostatic simulation allows the surface water to become only slightly denser than the water below the surface, which led to the cut-off that appears in Fig. 11. This is the reason that in the snapshot presented in the two right panels of Fig. 8, the hydrostatic simulation has a very small value almost over all the surface of the Dead Sea. However, in the non-hydrostatic simulation, the physical process of cold and dense surface water sinking by vertical convection was fully simulated, resulting in plumes such as those presented in Fig. 9. This means that the non-hydrostatic simulation did sustain the unstable state of heavy water above lighter water for several hours of surface water cooling during the night during which plumes of downward and upward moving water occurred. This situation was probably related to the fact that the SST continued to drop during the night, creating increasingly heavier water that maintained the unstable state of heavier water above lighter water throughout the night. A situation similar to the non-hydrostatic results presented above has been observed in the Gulf of Elat (see Fig. 2 of Biton et al., 2008) where dense water was observed to overlay lighter water. The fact that such a seemingly unstable situation has been observed in a nearby environment supports the conclusion that the non-hydrostatic simulation provided a better description of the actual physical process, which as explained above, allows the seemingly unstable situation to exist. Such conditions may be important for the potash industry, to optimize their production. A similar situation was observed in the Dead Sea measurements performed by the IOLR (on their online near-real-time monitoring web-page) e.g., during 12-14 of March 2014 and 26-28 Feb. 2015 where cold surface water overlaid warmer water for more than 4 hours; see Fig. 12.

The difference in the way the hydrostatic and non-hydrostatic modes simulated the vertical mixing has a significant impact mainly on the upper circulation of the Dead Sea. Fig. 13 presents vertical profiles of density, salinity and temperature, horizontally averaged over the entire lake and time averaged over the last five days of the simulation (days 25–30). As seen, in the hydrostatic results, the deep water had a lower temperature, slightly lower salinity, and, overall, a higher density than the deep water in the non-hydrostatic simulation. [Note that all the differences are small in magnitude.] This is explained by the fact that the diffusive mixing was stronger than the actual, convective mixing, which made the deep water respond
more strongly to surface conditions in the hydrostatic simulation than in the non-
hydrostatic one. Thus, in winter, when the overall lake becomes colder and fresher,
these changes affected the deep water more in the hydrostatic simulation than in
the non-hydrostatic simulation.

5. Results of the Summer and Winter Diurnal Cycle Under Spatially
Fixed Diurnally Varying Winds

After studying the hydrostatic and non-hydrostatic effects under the action of
temporally constant winds on the Dead Sea’s dynamics and properties, we then
focused on more realistic non-hydrostatic configuration and studied the effect of
wind (temporal) variability on the sea’s dynamics; the winds in the simulations
are spatially uniform. As described in Sec. 2.2, the time varying winds were based
on the climatological hourly means of observed winds in the Israel Meteorological
Service’s Ein Gedi station (presented in Fig. 2).

The surface snapshots of temperature and velocity field spanning the diurnal
cycle at six-hour intervals are shown in Fig. 14. In particular, this figure shows that
the velocity field of the two runs differed greatly, as one would expect, as the wind
stress significantly affected the surface current field. Similar difference between
constant and time varying wind stress is also found in the mean fields of density,
salinity, temperature, and velocity (not shown). The complex temperature field
at the end of the winter night (6 am) is associated with the convection processes.
In both runs, during summer, the northern part of the lake was colder than the
southern part. This difference between the northern and southern parts of the
lake is partially related to the northerly winds that upwell deep colder water in
the northern part of the basin.

The mean density and square of the current speed of the surface water, for
the non-hydrostatic winter simulations with both constant and varying winds, are
shown in the upper two rows of Fig. 15. The mean current speed squared represents
the kinetic energy of the water layer near the surface, which is mainly driven by
the winds. As shown, during the winter, the simulations with constant winds had
higher mean surface kinetic energy compared to the time varying winds. This
difference may be related to the relatively uniform wind stress during the winter
(Fig. 2) and to time varying smaller mean wind stress (∼0.009 N m$^{-2}$) compare to the constant wind stress case (∼0.024 N m$^{-2}$). The differences in the surface density were very small.

In addition to the above, it is noticeable that there is a connection between the surface water density and the surface current kinetic energy. During the day, when the surface water was lighter than the water below the surface, the water column was stratified, and, in practice, the wind forced a thin surface layer of water, which resulted in higher velocities of the surface water. During the late night and early morning, mixing occurred, and the mixing acted to decrease the velocity field, as the wind forced a deeper layer of water.

We now present the results of the simulations of the diurnal cycle in the summer. The conclusions of the comparison between the hydrostatic and non-hydrostatic cases in the winter diurnal cycle are also valid for the summer simulations, although the differences are less dramatic in the summer. Still, the non-hydrostatic mode is more suitable for simulating the circulation, as vertical (night) convection is a significant process even in the summer. Thus, we present the results from the non-hydrostatic simulations with a diurnal cycle in summer, for both constant and time varying winds. As seen in Fig. 2, the observed winds in the summer presented a high peak of westerly wind during the evening, centered around 7 pm; probably, this peak in wind activity was associated with the Mediterranean Sea breeze, which reaches the Dead Sea only during the late afternoon hours. We expected this peak of strong westerly wind to affect the circulation.

Snapshots of surface temperature and velocity field spanning the diurnal cycle of the summer (July) simulations are shown in bottom two rows of Fig. 14. This figure as well as mean surface density, salinity, temperature, and velocity (not shown) indicate that in the summer, the winds had a great impact on the surface conditions, where the temperature and velocity fields of the two cases were very different.

The two lower panels of Fig. 15 present the mean density and current speed squared (representing the kinetic energy) of the surface water, from the non-hydrostatic summer simulations. The current speed squared curves show a very dramatic difference between the time varying and constant wind simulations, which is due to the large difference between the varying wind stress (see Fig. 2) and the
constant northerly winds used in the latter simulation. The mean wind stress during July is \( \sim 0.022 \text{ N m}^{-2} \), close to the winds stress of the constant wind runs (\( \sim 0.024 \text{ N m}^{-2} \)). The kinetic energy of the time varying wind was much higher than the one under the action of constant winds, where the latter was almost constant with time. However, in spite of the dramatic difference in the kinetic energy of the surface water, the density profiles show very similar behavior, with only a slight offset between the constant wind and the varying wind simulations. The variability of mean surface kinetic energy in response to constant wind stress is greater during winter in comparison to the summer. This is probably due to the convection processes that are present in the winter while absent during the summer.

6. Summary and Conclusions

The Dead Sea is a lake with the saltiest natural water in the world, with a salinity that is about eight times greater than standard ocean water (Gertman and Hecht, 2002). During winter, the Dead Sea’s circulation is significantly affected by the sinking of cold surface water, leading to a total mixing of the water column. This process is linked to non-hydrostatic processes. The Dead Sea is currently of local and global interest, as the planned Red Sea-Dead Sea canal (Beyth, 2007) may significantly alter the circulation in the Dead Sea, which may, in turn, cause negative economic impacts on the Dead Sea Works and Dead Sea tourism. Despite the clear importance of the Dead Sea and despite the expected non-hydrostatic effects on the sea’s circulation and properties, no studies have investigated the Dead Sea using non-hydrostatic equations. In addition, studies of the effect of diurnal wind variability under non-hydrostatic effects are also lacking. In this paper, we aimed to begin filling these gaps.

We presented numerical (MITgcm) simulations of the Dead Sea’s circulation, using both the hydrostatic and the non-hydrostatic modes. The results demonstrated that for simulations with a spatial resolution of 100 m when using seasonal forcing (monthly means) for temperature and salinity and constant winds, the hydrostatic and non-hydrostatic modes exhibit some degree of similarity. In simulations with a spatial resolution of 100 m, when using diurnal cycle forcing, the
The diurnal variability of the wind had some effect on the Dead Sea’s circulation during winter in comparison with the constant wind case—the latter resulted in a slightly smaller diurnal variability of the surface kinetic energy but with a higher
mean kinetic energy. This is probably related to the larger mean wind stress of the
constant winds compare to the time varying ones. The situation is very different
during the summer where the diurnal wind variability resulted in a much higher
surface kinetic energy compared to the constant wind case. In the case of constant
winds, the surface kinetic energy was almost constant during the day. The density
depth profiles of the diurnally variable and constant winds were similar, both
during the summer and winter. We note that the daily mixing pattern can change
dramatically from day to day as due to the variability in the wind pattern and
intensity (Essen et al., 1995; Hecht et al., 1997); in addition, the seasonal cycle
can change significantly from year to year, mainly as a result of the annual fresh
water supply (Anati, 1997).

To properly simulate non-hydrostatic effects the lateral resolution should be fine
enough, otherwise numerical dispersion may be confused with real non-hydrostatic
effects (Vitousek and Fringer, 2011). To safely study non-hydrostatic effect the
leptic ratio (the ratio between the horizontal resolution and the depth of the mixed
layer) should be small enough (Vitousek and Fringer, 2011). The reported differ-
ences between the hydrostatic and non-hydrostatic simulations occur during the
winter, when the mixed layer span the entire water depth, resulting in leptic ratio
that is \( \sim \frac{1}{3} \), indicating that the reported non-hydrostatic effects are not due to
numerical dispersion. In addition, the lack of similarity between the hydrostatic
and non-hydrostatic results is another indication that the (fine resolution) non-
hydrostatic simulations are not due numerical artifacts, as when the leptic ratio is
large, hydrostatic and non-hydrostatic simulations yield similar results (Vitousek
and Fringer, 2011). Still, during the summer, the reported differences may be less
reliable as the mixed layer is much shallower then. Yet, the summer is not ex-
pected to have significant non-hydrostatic effect due to the lack of deep convection
processes.

While we present here high resolution non-hydrostatic simulations of the Dead
Sea, we keep in mind that the results only crudely estimate the sea’s state. There
are several reasons for that:

1. The simplified forcing we have used. We have used restoring boundary con-
ditions for the surface temperature and salinity instead of using heat and
freshwater fluxes. In addition, we have used spatially uniform forcing in-
stead of spatially variable forcing fields. During the winter, the Dead Sea experiences intensive pulses of freshwater from the Jordan River and the wadis surrounding it. These can affect the Dead Sea’s surface properties and circulation. The wind stress field was also spatially uniform. The seasonal and diurnal cycle was also simplistic (sinus function); note that the diurnal wind stress was more realistic.

2. The inability to handle the decline in the Dead Sea level. This problem is related to forcing that we have used that lead to a quasi steady state, in contrast to the reality where the salinity and density increase with time due to the decline in Dead Sea level. Thus, even for more realistic forcing the model could fit the observations only for one specific year while for the next year the observations would indicate denser and saltier water while the model will not.

3. The simple choice of a linear equation of state. The equation of state of the Dead Sea water is far more complicated and nonlinear (due to, for example, the saturation of the different components of the salt at different salinity levels).

The above can affect the Dead Sea’s surface properties and circulation and may underlie the discrepancy between the model’s salinity and density and observations. Moreover, double diffusion processes that expectedly occur in the Dead Sea are ignored here—these processes are expected since the surface water of the Dead Sea is saltier and warmer than its deep water during most of the year. In addition to the above, we did not study the (barotropic and baroclinic) waves in the lake including basin scale oscillations, trapped Kelvin waves, and short internal waves, although these are of course present in the lake (e.g., Sirkes, 1986, 1987; Hecht et al., 1997) and can be different when using hydrostatic or non-hydrostatic configurations. However, we believe that our results shed important light on the implications of non-hydrostatic effects on the dynamics and properties of the Dead Sea.

7. Acknowledgment

We thank Erick Fredj and Hezi Gildor for helpful discussions. The model setups and results described in this paper are available upon request from the authors.
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Table 1: Details of the Dead Sea simulations. VW stands for varying winds. Only the 100 m simulations are shown in the present study. All simulations were run using the non-hydrostatic and hydrostatic configurations.

<table>
<thead>
<tr>
<th>Name</th>
<th>Res. (m)</th>
<th>Duration</th>
<th>Time Step (s)</th>
<th>Temp. (°C)</th>
<th>Salinity (ppt)</th>
<th>Winds (m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Annual 400 m</td>
<td>400</td>
<td>20 yrs.</td>
<td>120</td>
<td>22 - 35</td>
<td>278 - 282</td>
<td>4.5, Northerly</td>
</tr>
<tr>
<td>Annual 200 m</td>
<td>200</td>
<td>10 yrs.</td>
<td>120</td>
<td>22 - 35</td>
<td>278 - 282</td>
<td>4.5, Northerly</td>
</tr>
<tr>
<td>Annual 100 m</td>
<td>100</td>
<td>&gt; 3 yrs.</td>
<td>60</td>
<td>22 - 35</td>
<td>278 - 282</td>
<td>4.5, Northerly</td>
</tr>
<tr>
<td>Diurnal Winter</td>
<td>100</td>
<td>30 days</td>
<td>60</td>
<td>18 - 26</td>
<td>278</td>
<td>4.5, Northerly</td>
</tr>
<tr>
<td>Diurnal Winter VW</td>
<td>100</td>
<td>30 days</td>
<td>60</td>
<td>18 - 26</td>
<td>278</td>
<td>Observed</td>
</tr>
<tr>
<td>Diurnal Summer</td>
<td>100</td>
<td>30 days</td>
<td>60</td>
<td>30 - 40</td>
<td>282</td>
<td>4.5, Northerly</td>
</tr>
<tr>
<td>Diurnal Summer VW</td>
<td>100</td>
<td>30 days</td>
<td>60</td>
<td>30 - 40</td>
<td>282</td>
<td>Observed</td>
</tr>
</tbody>
</table>

Figure 1: Dead Sea bathymetry (based on Hall, 1978), corresponds to surface water level of -427 meter below mean sea level (year 2013). The horizontal dashed line indicates the section location shown in Fig. 9 and the white circle in the middle of figure indicates the location of the profiles shown in Figs. 10, 11.
Figure 2: (a) Winter (January) and (b) summer (July) climatological diurnal cycles of the wind-stress components, used in the diurnal cycle simulations with varying winds, with a 100 m resolution. The data is based on hourly mean wind data of the Israel Meteorological Service’s Ein Gedi station. Note the different scale vector for the winter and summer.
Figure 3: SST maps (color) and surface horizontal velocity field (vectors), for hydrostatic (left panels) and non-hydrostatic (right panels) simulations of the annual cycle with 100 m resolution, in both summer (July, upper panels) and winter (January, lower panels).
Figure 4: Density (upper panel), salinity (middle panel), and temperature (lower panel) averaged in the horizontal directions and presented in the depth-time plane, from the non-hydrostatic simulation of the annual cycle with a 100 m resolution. The results of the hydrostatic simulation are similar.
Figure 5: Salinity (a) and temperature (b) profiles measured by the IOLR at the deepest point of the Dead Sea (see https://isramar.ocean.org.il/isramar2009/DeadSea/DS_Database.aspx). (c) The density as calculated using Eq. (1). The simulations profiles (indicated by the dashed lines) at $x = 12$ km and $y = 30$ km, which is approximately the location of the Dead Sea hydrographic station EG320, are in general agreement with these observations, although the model’s salinity and density are systematically smaller than the observed profiles. Note that the original salinity is given in units of quasi-salinity (Anati, 1999; Gertman and Hecht, 2002) and were converted here to “ppt”, to allow easier comparison with the simulations. Courtesy to Isaac Gertman from the IOLR for giving the permission to present this data.
Figure 6: Stream function (in mSv, 1 mSv is equivalent to $10^3$ m$^3$ s$^{-1}$) in the Y-Z plane from hydrostatic and non-hydrostatic simulations of annual cycle with 100 m resolution. Upper left panel: February, hydrostatic. Upper right panel: February, non-hydrostatic. Lower left panel: March, hydrostatic. Lower right panel: March, non-hydrostatic. The Y origin is taken at the southern most tip of the lake. The figure depicts the mean four Feb. and March months.
Figure 7: Horizontal (x and y) mean of density (top panels), salinity (middle panels), and temperature (bottom panels) as a function of depth and time, of the hydrostatic (upper 3 panels) and non-hydrostatic (lower 3 panels) simulations of the diurnal cycle with a 100 m resolution during winter (January) under the action of constant winds.
Figure 8: Left two panels: Surface current velocity (vectors) superimposed on the SST maps (color), for the hydrostatic (left panel) and the non-hydrostatic (2nd panel) simulations of the diurnal cycle with a 100 m resolution, in winter (January) from a snapshot at 4 am. Right two panels: Density difference between the surface water and the deep water density (defined as the mean density of water below a depth of 100 m) in the hydrostatic (3rd panel) and the non-hydrostatic (right panel) simulations of the diurnal cycle with a 100 m resolution, in winter (January) from a snapshot at 4 am. The zero density difference is indicated by the black contour line. The positive areas in the hydrostatic case are all below 0.009 kg m\(^{-3}\).
Figure 9: Density slice (4 am snapshot) in the X-Z plane ($Y = 30$ km, indicated by the horizontal dashed line in Fig. 1), of the hydrostatic (upper panel) and the non-hydrostatic (lower panel) runs of the diurnal cycle with a 100 m resolution, in winter (January).
Figure 10: Density profile versus time in the middle of the lake ($X = 10$ km, $Y = 30$ km, indicated by the white circle in Fig. 1), for the hydrostatic (upper panel) and the non-hydrostatic (lower panel) simulations of the diurnal cycle with a 100 m resolution, in winter (January). The vertical line labeled 4 am indicates the time of the snapshots presented in Figs. 8–9.
Figure 11: Density difference between the surface water and the deep water at the middle of the lake ($X = 10$ km, $Y = 30$ km, indicated by the white circle in Fig. 1) versus time, for the hydrostatic (black) and the non-hydrostatic (red) runs of the diurnal cycle with a 100 m resolution, in winter (January). The density difference is defined as the difference between the upper layer density and the mean water density below 100 m.

Figure 12: Several days long time series of water temperature at several depths when the starting time is (a) March 11, 2014 and (b) February 26, 2015. These time series were recorded automatically by the IOLR at the Dead Sea platform that is located somewhere in the middle of the lake (see http://isramar.ocean.org.il/isramar2009/DeadSea/position.aspx). Note the colder surface water that overlaid warmer water for several hours. Yet, river runoff and precipitation at that time may also contributed to this situation by making the surface water fresher. Courtesy to Isaac Gertman from the IOLR for giving the permission to present this data.
Figure 13: Vertical profiles of density (left panel), salinity (middle panel), and temperature (right panel), averaged over the horizontal directions and over time (last five days of the simulation) for the hydrostatic and the non-hydrostatic simulations of the diurnal cycle with a 100 m resolution, during winter (January).
Figure 14: Temperature and velocity field snapshots, spanning the diurnal cycle, of the non-hydrostatic simulations of the diurnal cycle with constant (1st and 3rd rows) and time varying (2nd and 4th rows) winds with a 100 m resolution, in winter (January, 1st and 2nd rows) and summer (July, 3rd and 4th rows). The vectors indicate the surface current field.
Figure 15: Current speed squared (1st and 3rd panels) and density (2nd and 4th panels), averaged over the surface of the entire lake, as a function of time, of the non-hydrostatic simulations of the diurnal cycle with constant (black) and varying (red) winds with a 100 m resolution, in winter (January, upper two panels) and in summer (July, lower two panels).