Wave forcing effects on coastal aquifer groundwater levels: Observations from the Dead Sea

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Abstract

Wave forcing on coastal aquifers affects groundwater level by elevating the mean water surface at the shoreline (wave set-up), and by infiltration during wave uprush above the mean water level (wave run-up). In this study we distinguish between the two mechanisms by field measurements. Measurements are conducted at the alluvial coastal aquifer of the Dead Sea, Israel, where tides are negligible, by means of georeferenced video camera and pressure transducers installed in observation wells, on the lake’s floor, and buried near the shoreline. Results show that during wave events (i) shoreline set-up is ~50% of maximum wave height, (ii) groundwater level rise decays landward within a few meters from the shoreline, and is not recognized at 90 m landward, (iii) the time lag of groundwater level rise due to wave forcing is not detectable within 8 m away from the shoreline (<5 minutes), but it increases linearly with distance landward and reach ~60 minutes at a distance of 75 meters away from the shoreline, (iv) 4 m offshore the mean lake level is reduced by ~10% of the maximum wave height, compared to the mean lake level further offshore, (v) groundwater level rise is not dependent on wave event duration, and (vi) mean lake level rise effect on groundwater level reaches farther into the aquifer (90m) than wave forcing effect, due to different geometries of the lake’s mean water surface near the shoreline. The main outcome of this study is the distinction between wave set-up and run-up, indicating that only shoreline set-up is responsible for the groundwater level increase during wave events at the Dead Sea.
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1. Introduction

1.1. Wave forcing effects on groundwater

1.1.1. The role of wave forcing in coastal aquifers

The effects of sea level variations on groundwater levels have been studied at different time scales, such as the time scales of tides (hours) and waves (seconds to minutes). Oceanic wave forcing on coastal aquifers has been recognized to have a major role on groundwater levels in the adjacent coastal aquifer and on water and solutes fluxes dynamics near the coast (Cartwright et al., 2004; Li et al., 2000). It is now widely recognized that submarine groundwater discharge (SGD) provides a significant transport pathway for chemicals entering from the continents to the marine environment (e.g., Johannes 1980; Burnett et al. 2003; Brovelli et al. 2007). The groundwater discharge at the shoreline is composed of freshwater and recirculated seawater. The major driving forces for the submarine groundwater discharge include: Topography driven freshwater that also circulates sea water along the fresh-saline interface, tidal forces, and wave forces. The effects of the topography and tidal forcing have been vastly reported, suggesting seasonality in the chemical loading of coastal waters and upper saline plume formation in the intertidal region, respectively (Michael et al., 2005; Robinson et al., 2007). However, field observations of the role of wave forcing are scarce because it is difficult to separate between the effects of waves and tidal forcing. In some field studies, wave amplitudes were measured at far distances from the shoreline, from hundreds of meters (Turner et al., 1997) to tens of kilometers (Rotzoll and El-Kadi 2008) and their actual influence on the shoreline location (wave set-up) was not directly measured but only estimated.
1.1.2. Basic terms in coastal hydraulics

At the meeting zone between the sea and a coastal unconfined aquifer, one needs to decide the location of the shoreline, which is critical for understanding hydrological behavior at the beach. The aim of this section is to briefly describe the basic terms regarding water levels at the beach area and the processes affecting the location of the shoreline and the water table at the beach.

In order to describe sea level surface over time we can divide it into three water surfaces or levels as written below and shown in Figure 1:

**Mean sea level (MSL):** The mean sea level is the measured sea level averaged over a time period long enough to cancel out the effects of long period fluctuation such as weather systems and tides.

**Still water level (SWL):** The still water level is the measured sea level that is averaged over a time period long enough to cancel out the effects of surface waves (waves induced by wind shear), but short enough to show the tidal effect on sea level.

**Mean water surface (MWS):** The mean water surface is the measured sea level that is averaged over a time period short enough (rule of thumb: 100 cycles of the longest wave period expected) to describe only the effect of water waves on sea level.

By definition, the SWL is located above or below the MSL, depending on the tidal stage and both of them are horizontal surfaces. The MWS is somewhat different since it could be a non-horizontal surface during wave activity and tends to be higher than the SWL when approaching the shore, and lower than the SWL at the breaking point of the wave (Figure 1). The shape of the MWS during wave activity was prompted by the hurricane that hit the east coast of the United States in 1938 (Guza and Thornton, 1981).
storm peak the mean shoreline water elevation at an area exposed to waves (Narragansett pier), was approximately 1 meter higher than the mean shoreline water elevation at nearby calmer water (Newport). Longuet-Higgins and Stewart (1964) showed theoretically, that deep water waves breaking on a sloping a beach, result a cross-shore gradient in radiation stress, that rises the MWS towards the shoreline. This was also demonstrated in laboratory experiments done by Bowen et al. (1968). The deviation of the MWS from the SWL was termed wave set-up which is the vertical distance between these two surfaces. At the wave breaking point, wave set-up value is negative, and sometimes referred to as wave set-down. Since the MWS includes the effect of tides and waves, the intersection between the MWS and the beach face describes the location of the shoreline accurately along time.

Figure 1: schematic section of the surf and swash zone, containing mean sea level (MSL) in black; still water level (SWL) in blue; instantaneous water surface in red; mean water surface (MWS) and groundwater table in green. WL represents instantaneous water line and SL represents the shoreline, defined as the intersection between MWS and the beach surface (modified from Nielsen, 1999).
When looking at water waves’ behavior, it is common to divide the sea into three main zones: deep water, surf zone and swash zone. For this discussion we assume that in the deep water, waves do not break and therefore the MWS equals SWL (wave set-up equals zero). A wave that approaches the shore, and enters shallow waters starts to break according to the ratio between wave height and water depth. The zone between the wave breaking point and the shoreline is termed the surf zone. In the surf zone, MWS deviate from SWL and wave set-up is different than zero. Wave set-up is smallest (negative) at the wave breaking point and largest (positive) at the shoreline, according to changes in the radiation stress. The swash zone is the most landward part of the surf zone on which waves rush upward and downward. The maximum and minimum height that the water line reaches up on the beach during one wave is termed wave run-up and wave run down, respectively, and is measured as the vertical distance from SWL. Most importantly, the swash zone is not continuously covered by water and it is therefore inappropriate to calculate wave set-up at this zone according to the wave set-up theory (Bowen et al., 1968; Gourlay, 1992; Nielsen, 1988). The shoreline is located somewhere in the swash zone and there is yet no firm physical theory that can describe its exact location. Nielsen (1988) suggested that the location of the shoreline is the intersection between beach water table and the beach face, while others suggested the measured location of the mean water line on the beach face as the shoreline (Gourlay, 1992; Guza and Thornton, 1981; Holman and Sallenger, 1985; Stockdon et al., 2006). In this study we use the definition of the average waterline as the shoreline, because of the problematic issue of water table measurement in the swash zone. Another important feature of the swash zone is that wave run up flood the beach face landward from the average water line/shoreline, and lead to sea water infiltration into the beach sediment. This infiltration is also depended on sediment properties and beach slope.
In conclusion, water waves in the sea may affect the beach water table through two main mechanisms: (1) wave set-up, which determines the location of the shoreline according to the shape of the MWS; and (2) wave run-up, which contributes sea water into the aquifer by infiltration at the area landward from the shoreline.

1.1.3. Previous work

The influence of wave forcing on coastal groundwater has been studied mostly in the laboratory (Boufadel et al., 2007; Gourlay, 1992; Kang et al., 1994; Longuet-Higgins, 1983; Turner et al., 2016) and by using numerical models (Bakhtyar et al., 2013, 2012; Li and Barry, 2000; Robinson et al., 2014; Xin et al., 2015, 2010) while field observations are scarce.

1.1.3.1. Field studies

Simultaneous field measurements of waves and groundwater levels show that waves in the ocean elevate groundwater levels (Kang et al., 1994; Nielsen, 1999, 1988; Turner, 1998; Turner et al., 1997). Rotzoll and El-Kadi (2008) detected the wave signal at observation boreholes as far away as 5 km from the coast, in case of extremely high waves (in Hawaii). The actual wave set-up at the shoreline is hard to measure, because of non-continuous presence of water at the swash zone. It was measured in the field by two ways: as the average water line (Guza and Thornton, 1981; Holman and Sallenger, 1985) and as the intersection of the mean water table/run up limit with the beach face, (Hanslow and Nielsen, 1993). The latter study derived an empirical equation for wave set-up on natural beaches, but since wave set-up was referred to groundwater table, the measurements where influenced significantly by tidal stage. Turner et al. (1997) pointed out that site specific monitoring is the only practical method for quantifying groundwater overheight at the coast because of complications inherent due to water
table-tide decoupling. Malott et al. (2016), showed that wave forcing in Lake Huron (in Canada), where tide forcing is negligible, can lead to a rapid temporal variation in the chemical conditions of the shallow part of a sandy nearshore aquifer, thereby showing that wave forcing has influence on the fate of reactive pollutants discharging through coastal aquifers.

1.1.3.2. Laboratory studies

The wave forcing effects on groundwater were studied also through laboratory experiments. Longuet-Higgins (1983) showed experimentally that waves induce an onshore upward setup in the phase-averaged sea level that drives a circulation of water within a porous beach. Boufadel et al. (2007) reported laboratory tracers’ studies in an 8 m wave tank and addressed not only the issue of groundwater rise due to waves, but also solute transport, which relates to the exchange fluxes. Turner et al. (2016) studied the effect of different wave conditions on sandy coastal barriers in a field scale laboratory flume (250 m long, 5 m wide and 7 m deep) and found that the sand barrier flow dynamics are dominated by the beach face water exchange due to wave run-up, regardless to the direction of background groundwater hydraulic gradient.

1.1.3.3. Numerical models studies

Through numerical simulations, Li and Barry (2000) found that wave run-up also contributes to the seawater circulation. Their study further showed that the circulation was affected by the beach groundwater table elevation relative to the MSL. Xin et al. (2010); Bakhtyar et al. (2012); Bakhtyar et al. (2013) and Robinson et al. (2014) numerically simulated the effects of wave setup acting on a sloping beach and showed that an extensive upper saline plume (USP) is created with increased seawater recirculation across the aquifer-ocean interface (Figure 2). Xin et al. (2015) further
showed that the interaction of waves and tides causes weaker hysteresis wave effect on SGD and suggested that a tidally influenced coastal aquifer is less sensitive to wave conditions. Therefore, it is preferable to study wave forcing on groundwater in locations with no tides as in the Dead Sea.

**Figure 2**: Conceptual diagram of the water levels, groundwater flows, and salinity distribution in a nearshore aquifer exposed to waves. The still water level (SWL), instantaneous water surface (thick solid line), and wave setup profiles (phase-averaged water surface) for the steady-state condition (pre-storm and post-storm, dashed line) and storm peak (dash-dotted line) are shown. The upper saline plume (USP) formed by wave-driven recirculation (WDR) is shown in addition to the potential expansion of the USP in response to the storm. The saltwater wedge formed by density-driven recirculation (DDR) is also shown. (modified after Robinson et al., 2014)

**1.1.4. Objectives**

This study has two main objectives: to quantify the contributions of wave set-up and wave run-up to nearshore groundwater level rise, and to study the effect of wave forcing on spatial and temporal groundwater level variations at coastal aquifers, using field observations from the coastal area of the Dead Sea, Israel, where tides are negligible.
we consider the shoreline (shoreline set-up) as the height of the mean water line above still water surface (Gourlay, 1992; Guza and Thornton, 1981; Holman and Sallenger, 1985; Stockdon et al., 2006), and differentiate between the mean water level of the sea and the groundwater level of the aquifer. This study is the first to report simultaneous measurements of nearshore wave height, wave setup and groundwater level. Furthermore, it is the first study that quantifies wave forcing on groundwater levels in a natural beach without tidal influence.

1.2. The Dead Sea and its coastal aquifer

1.2.1. The Dead Sea and adjacent Groundwater systems

The Dead Sea (Figure 3 A, C) is a terminal lake situated in the deepest part of the Dead Sea Rift System. The Dead Sea groundwater system consists of two main aquifers (Figure 4): The Upper Cretaceous Judea Group Aquifer and the Quaternary alluvial coastal aquifer (Arad and Michaeli, 1967; Yechieli et al., 1995). The present study is focused on the upper phreatic part of the alluvial coastal aquifer (thickness of ~50 meters), which consists mainly of clastic sediments, such as gravel and sand. Below the phreatic unit, there are several deeper sub-aquifers which are separated by lacustrine clay (e.g. clays, aragonite, gypsum and salts). This aquifer is bounded by normal faults (at 1-3 kilometers from the Dead Sea shoreline), which set Cretaceous carbonate rocks of the Judea Group against Quaternary alluvial and lacustrine sediments. The recharge of the aquifer is mainly through lateral flow from the Judea Group aquifer, which is replenished in the highlands 10-30 km to the west and by flash floods. Direct recharge is negligible because of the arid climate and high evaporation in the Dead Sea region.
Figure 3: A) Satellite image (Terra/MODIS) of the Dead sea. The study area is marked with red rectangle. B) Landview of the shoreline in the study area. C) Study area location at the dead sea coast of Israel. D) Study area map: Boreholes (green circles), sea level sensors (red circles) and the buried water level sensor locations near Ein Gedi, Israel.
Figure 4: Schematic cross section of the Dead Sea groundwater system showing the relation between the different aquifers and the Dead Sea (after Yechieli et al., 1995).

The Dead Sea salinity and density are 340 g/L and 1.24 kg/L, respectively (Anati, 1997; Arnon et al., 2016; Gertman and Hecht, 2002), and its viscosity is 3.6 mPa·sec at 20 °C (Weisbrod et al., 2016). The extremely high density of the Dead Sea induces a very shallow interface between the fresh groundwater and lake’s brine. Although the Dead Sea system is extremely dynamic, the Ghyben-Herzberg approximation for static water was found to be relevant as the slope of the interface is 10 times shallower than the slope normally expected near oceans, due to greater density contrast (Yechieli, 2000).
The Dead Sea level drops by 1m/yr due to its negative water balance, which is translated to a water deficit of $\sim 700 \times 10^6$ m$^3$. This drop is a result of the considerable exploitation and diversion of water upstream from the Sea of Galilee, the Jordan and the Yarmouk rivers, and due to the Dead Sea brine pumping of the Israeli and Jordanian Dead Sea Industries (Lensky et al., 2005; Yechieli et al., 1998). The annual water volume that enters the Dead Sea today is $265\,–\,335 \times 10^6$ m$^3$ (Lensky et al., 2005).

The Dead Sea and the adjoining groundwater system are hydraulically interconnected as expressed in a relatively fast (a few days) groundwater level response to level changes of the Dead Sea (Yechieli et al., 1995). The drop in groundwater level decreases with increasing distance from the shoreline inland (Yechieli et al., 2009). Preliminary measurements imply that in some locations near the shoreline, the groundwater level drop is similar to that of the Dead Sea, while reaching a dynamic steady state (Kiro et al., 2008), whereas in other locations the drop is significantly smaller than that of the Dead Sea (Yechieli et al., 1995).

The fresh-saline water mixing zone is also responding to the drop of the sea level and the eastward shift of the Dead Sea shore line, resulting in the fast flushing of the Dead Sea coastal aquifer. Hydrological modeling of the Dead Sea system shows that the controlling factors of the groundwater response in the different parts of the aquifer are the hydraulic conductivity and the distance from the sea (Kiro et al., 2008; Yechieli et al., 2010).

1.2.2. The Dead Sea as a field site for the studying wave forcing effects on aquifers

The upper gravel sub-aquifer of the alluvial coastal aquifer adjacent to the Dead Sea provides a good case study for quantifying the effect of wave forcing on groundwater
level because of the lack of tides and the diurnal occurrence of wave events. At the Dead Sea, there is typically a late afternoon wind that causes wave events which last approximately 12 hours with amplitude changes of tens of cm (Arnon et al., 2014; Bodzin et al., 2014; Hect and Gertman, 2003; Nehorai et al., 2013). There are also no significant tides and therefore the wave forcing effect on the groundwater of the high permeability gravel coastal sub-aquifer is more distinguishable. Unlike freshwater lakes, the Dead Sea water is quite unique in its high salinity and density, generating a shallow slope of the fresh-saline water mixing zone and allows studies through shallow observation wells. The study of wave effects on groundwater near saline lakes such as the Dead Sea could be applied on other coastal aquifers near oceans. The amplitude of water level decline at Dead Sea is about 1 meter per year. This is an opportunity to compare its influence on groundwater levels, compared to the influence of wave forcing.
2. Materials and methods

2.1. Measuring instruments and field setup

The measuring setup (Figure 5) was designed to quantify the temporal and spatial variations of the water level in the aquifer and in the lake to distinguish between the influence of wave set-up and wave run-up on the groundwater level rise. Water level was measured with pressure transducers that were installed in four observation wells (5 cm diameter) fully perforated that reach 10-17 meters below the water table (see Appendix 1), one buried pressure transducer in the swash zone and two pressure transducers on the lake's floor. The pressure transducers used in this study were: i) Campbell Scientific model CS451, resolution <1 mm, accuracy 0.5 cm, data storage in an external CR200X data logger, and ii) Solinst Levelogger model 3001 (a standalone pressure transducer with a built in data logger limited to 40,000 pressure readings) with a resolution of 1-3 mm, accuracies: 0.5-1 cm. All pressure recordings were compensated for barometric pressure, corrected for water density and calibrated with a manual water level measurement. The water density was considered constant for each measuring point before during and after each wave event. The observation wells showed increasing density with depth and the correction there was done according to an average density, calculated from the pressure transducer depth, pressure reading and water table depth. The calculated average density could be used for short term groundwater level measurements (few days), where groundwater level variations are small (few centimeters), and the water density gradients with depth are assumed to be stable. Water density of between 1.06 and 1.24 g/cm³ was measured in the study area, and pressure transducers that were installed for more than one week, were sealed in durable plastic bags filled with fresh water, as a corrosion protection (For further information on
density profiles and density correction see Appendix 3). Wave activity at the shoreline was tracked by a GoPro 3+ video camera shooting at rate of 25 frames per second (fps), and a geo-referenced measuring bar in the video camera frame was used to obtain referenced level data (Figure 5A). All level measurements were georeferenced by means of GPS RTK, a leveling instrument and Israel’s water authority Dead Sea water level data to give absolute elevations.

2.2. Field campaigns

Three field campaigns (Table 1) were performed, each one concentrating on a different aspect of the research objectives: (i) the 1 day campaign was designed to quantify wave setup profile during a wave event including the average waterline as the maximum wave set-up (shoreline set-up), using 0.5 Hz frequency water pressure measurements and video camera processing, (ii) the 5 days campaign was designed to quantify the influence of wave set-up and wave run-up mechanisms on groundwater level, and (iii) the 2.5 months campaign was design to compare the wave forcing effects to the long term water level change in the lake. The division into three different campaigns was done due to technical limitations of the spatial and temporal resolution of the measuring system in the field.

Table 1: field campaigns dates and instrumentation in this study

<table>
<thead>
<tr>
<th>campaign</th>
<th>dates</th>
<th>groundwater instrumentation</th>
<th>swash zone instrumentation</th>
<th>Lake instrumentation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 day</td>
<td>6-7-/10/2014</td>
<td>pressure transducer (2 sec interval)</td>
<td>video camera</td>
<td>pressure transducer (2 sec)</td>
</tr>
<tr>
<td>5 days</td>
<td>20-25/3/2014</td>
<td>pressure transducer (10 sec and 5 min interval)</td>
<td>buried pressure transducer (10 sec interval)</td>
<td>pressure transducer (10 sec interval)</td>
</tr>
<tr>
<td>2.5 months</td>
<td>15/2/2014-28/4/2014</td>
<td>plastic bag protected pressure transducer (5 min interval)</td>
<td></td>
<td>pressure transducer (5 sec interval)</td>
</tr>
</tbody>
</table>
Figure 5: Instrumental setup in two field campaigns. (A) One day campaign: a GoPro camera records the location of the waterline at the swash zone, two pressure transducers located on the lake floor and one pressure transducer installed in a shallow borehole at the beach face (not flooded by waves); (B) Five days campaign: A pressure transducer located on the lake floor, a buried pressure transducer located in the swash zone (flooded by waves), and four pressure transducers installed in observation wells at different distances landward from the shoreline.
One day campaign - wave setup

The 1 day campaign included high frequency level measurements by a video camera and pressure transducers. Waterline elevation measurements were done using a GoPro video camera recording the waterline along a measuring bar attached to ground, perpendicular to the shoreline (Figure 5A, Figure 6). The videos were taken at a rate of 25 frames per second (fps) along 1-10 minutes sessions during different stages of a wave event: high waves, medium waves and still water. These 3 stages represent the common pattern of a wave event in the Dead Sea that lasts only a part of the day. The video processing included identifying the location of the waterline as the highest position of water along the lower side of the measuring bar, assuming that measuring bar angle represents the average angle of the cross shore profile at all stages. Tracker software was used to digitize the location of the waterline on the measuring bar and the product of this procedure is a time series of the waterline location. Maximum wave setup (mean waterline) was determined as the mean value of waterline records measured at each stage of the wave event. The video method was used in order to measure maximum wave set up (mean waterline) because pressure transducer water level measurements require a continuous submersion of the sensor in water, which is only partly attained at the swash zone.

The procedures used for converting water line measurements from distance on the measuring bar \( l \) into elevation above mean sea level \( z \) and horizontal distance from still water shoreline \( x \), are shown in Figure 7. During the still water phase of the lake, mean waterline elevation and distance on the measuring bar are defined as \( z_0 \) and \( l_0 \) respectively (measured during the one day campaign on the 6/10/2014).
Figure 6: A view of the waterline measuring system. The Gopro video camera (in the red circle) is installed on an iron stake with its lens aimed at the measuring bar that is fixed to the beach surface.
In each video measurement, the Gopro camera was set at a different position, in order to achieve a good view of the waterline location on the measuring bar, and thereby required a different calibration between the movie view and the measuring bar. The calibration was done by comparing the real length of the measuring bar at 10 cm increments with the x coordinate of the movie view (Figure 8). Appendix 5 contains Tracker software view of axis and calibration points (Appendix 5, Figure 38), calibration curves (Figure 39), and calibration data (Table 3A) of all the video measurements done in this study. In order to find waterline elevation, measuring bar was also geo-referenced using a leveling instrument. The geo-referencing result was verified during still water stage of the lake by comparing it with the Israel Water Authority continuous (every 30 minutes; Givati, personal communication, 2014 and 2016) Dead Sea Level measurement. By using Tracker software, Waterline location was marked every 3 frames (0.125 sec) at the bottom of the measuring bar, and then converted into real distance on the measuring bar according to the calibration curve. Waterline elevation was then calculated according to the geo-referencing and the angle of the measuring bar (10°) as shown in Figure 7.
Figure 7: A schematic diagram of video measuring bar on the beach face. Still water shoreline location ($l_0$, red circle) is marked on the measuring bar (blue line). Time varying waterline location on the measuring bar, elevation above mean sea level and horizontal distance from still water shoreline are represented by $l_t$, $z_t$ and $x_t$ respectively. Equations for calculating shoreline elevation ($z_t$) and water line horizontal distance from still water shoreline ($x_t$) are presented below the diagram. The values of $\alpha$ and still water shoreline coordinates ($l_0, z_0, x_0$) are presented below the equations.

\[
\Delta l = -(l_t - l_0) \\
z_t = z_0 + \Delta l \cdot \sin \alpha \\
x_t = x_0 - \Delta l \cdot \cos \alpha
\]

$l_0 = 2.931 \text{ m}$, $x_0 = 0 \text{ m}$, $z_0 = -428.632 \text{ m}$, $\alpha = 10 \pm 1^\circ$
Figure 8: Calibration between video image and measuring bar during high waves stage. (A) Snapshot from Tracker video processing that reveals the measuring bar on the beach face. Ten cm intervals are marked on the measuring bar as blue diamonds and magenta line represents the picture tilted X axis (B) Calibration curve (black line) between measuring bar length and x coordinate of the pixels in the picture (measurements in blue diamonds).

Lake level, maximum wave height ($H_{max}$) and groundwater level where measured by pressure transducers according to the configuration in Figure 5A. Two pressure transducers were installed on the lake floor at distances of 4 and 12 meters seaward from the shoreline and one pressure transducer was installed in a shallow borehole, 4 meters landward from the shoreline. Pressure measurements were recorded every 2 seconds, which is sufficient in order to calculate the maximum wave height over a 10 minutes time window. Water level measurements of the lake were averaged over time.
windows of ten minutes (300 readings) to obtain the mean water level (Figure 9A) at each measuring point. Maximum wave height time series were calculated as the difference between maximum and minimum water level measurement of each time window (Figure 9B).

Figure 9: One day of high resolution level measurements in Dead Sea near Arugot Wadi aluvial fan, see location in Figure 3 and Figure 5A. (A) Water level measurements (red dots, pressure transducer location is ~1.5m deep, ~12 m offshore) along with mean (black line), minimum, and maximum (dashed black lines) of 10 minute time moving windows (300 readings). All results are smoothed by an additional moving average procedure; (B) Lakes maximum wave height \[ H_{\text{max}} = (\text{maximum level} - \text{minimum level}) \text{ in each 10 minutes time window} \].
**Five days campaign – wave setup vs. run up mechanisms**

A longer campaign of five days was performed with one pressure transducer installed at the lake floor (12 meters from the shoreline), one buried pressure transducer under the swash zone (~1.5 meters landward from the shoreline and ~7 cm below the water table during the still water phase of the lake. The pressure transducer depth below the surface was ~40 cm), and four pressure transducers installed in observation wells located at different distances from the shoreline (Figure 5B). The time interval of water level measurement at the observation wells was every five minutes. The time interval for the water level measurements in the Dead Sea and in the swash zone (buried pressure transducer) was every 10 seconds. Mean, minimum, and maximum lake level were calculated by a moving window of 120 readings (20 minutes).

**Two and a half months campaign – wave forcing and lake level effects on the aquifer**

The longest campaign included water level measurements in all four observation wells and at the lake floor. The time interval of water level measurements in the observation wells was every 5 minutes and at the lake floor every 5 seconds.
3. Results

3.1. The diurnal cycle of maximum wave height, wave setup and wave set down (One day campaign)

The results of the one day campaign are divided into two main parts, the waterline video measurement and the pressure transducers water level measurement.

The purpose of the waterline video measurement was to determine the shoreline set-up (maximum wave set-up) which is defined here as the mean waterline elevation. According to the wave conditions at the study area, it was sufficient to use movie frames in increments of 0.125 sec, though during the high wave’s phase of the lake, a significant layer of foam partly covering the lake water and beach face, prevented a continuous clear identification of the waterline position. At the presence of the foam, waterline position was clear mostly during wave up rush (run-up) on the beach face while impossible to be detected on the wave down rush. An example of water line measurements from the high waves’ phase of the lake is shown in Figure 10 where the waterline is marked, only when possible, on the measuring bar in each snapshot, to create the waterline time series. In order to calculate the mean waterline position, a linear interpolation was applied (Figure 11), since it is the simplest form of predicting the waterline position during time of foam covering the beach face. In cases where waterline interpolation was used, run-up and run-down average was different from the interpolated waterline average (Figure 12) showing the asymmetry between times of waves up rush and down rush. Movies of 1 to 10 minutes taken during the one day campaign where analyzed to give the shoreline set-up of the different phases of the wave event. During the still water phase of the lake, a short time of ~1 minute was
Figure 10: Snapshots from Gopro video camera during high wave’s phase of a wave event. Gopro camera was started on 7.10.2014 8:12 (GMT+2). Red arrows point to water line position on the measuring bar. The question marks represent uncertainty in waterline location determination. The water line elevation time series at bottom of the figure is composed of measurements (black circles) and linear interpolation data (red X’s) when measurement uncertainty is too high. The number labels on the time series refer to the snapshots.
needed for waterline position analysis to have a small measurement error, while at least 5 minutes were needed for the same analysis and measurement error during high waves phase.

The comparison between waterline time series of the different phases of the wave event (Figure 13) summarizes the behavior of the Dead Sea shoreline due to waves. In this case, waterline elevation is rising by ~20 cm from still water phase to high waves’ phase. The error increase of the waterline measurement seen as waves get higher is mostly due to the measurement error of the measuring bar angle ($\alpha = 10 \pm 1^\circ$). Appendix 5 contains a more detailed view of the beach face (Figure 37), waterline position on the measuring bar (Figure 40) and results (Table 3B) of all the video measurements done in this study.

Figure 11: video measurement and interpolation from the high waves' phase of the Dead Sea on 7/10/14. A 30 seconds section of the waterline measurement (black circles) and linear interpolated waterline time series (red marks and line).
Figure 12: Video measurement and interpolation from the high waves' phase of the Dead Sea on 7/10/14. Mean wave run-up (RU), run-down (RD) and shoreline (SL) are represented by black circles on the right and horizontal black lines. The Mean shoreline elevation is calculated from interpolated waterline data. Error bars refer to the standard error of the mean and the measurement error of measuring bar angle.

Figure 13: Waterline elevation time series of three different phases of a wave event: still water (blue), mid waves (black) and high waves (red). Time is shifted for comparing the series. The average waterline (shoreline set-up) of each phase is marked with horizontal lines.
Water level measurements with high resolution and high frequency of the one day campaign, using pressure transducers, show a diurnal cycle of wave amplitude, with ~7 hours of still water (15:00-22:00 local time) followed by a period dominated by $H_{\text{max}}$ (maximum wave height) of ~40 cm (Figure 14B). The mean lake level is stable during the measurement period, and is not affected by tides and change in $H_{\text{max}}$ (Figure 14A). However the mean water level measured closer to the shoreline at shallow water show set-down - a decrease in mean water level of 10% (4 cm) of $H_{\text{max}}$, whereas the mean water level measured at the shoreline show set-up (Figure 14C). Wave set-up at the shoreline (shoreline set-up) that was measured by means of video camera correlates linearly with $H_{\text{max}}$ (Figure 15). The shoreline set-up magnitude is $54 \pm 5\%$ of the $H_{\text{max}}$ at this location of the Dead Sea. This relation will be used later to evaluate wave setup when only $H_{\text{max}}$ measurements are available in the study area. The geometry of the mean lake and groundwater level changes during the daily wave events from the geometry that exists during the still water phase (Figure 16). At the shoreline wave set-up reaches its maximum value whereas at a short distance into the lake (4 m), wave set-up is lower than the still water level (set-down). Further offshore (12 m), the mean level is not affected by wave action. Groundwater level rises by ~5% of the $H_{\text{max}}$ at the observation well located 4 m onshore, during the high waves’ stage of the wave event.
Figure 14: High resolution level measurements in N. Arugot aluvial fan, see location in Figure 3-D and Figure 5-a. (A) Lake level measurements (red dots, pressure transducer located ~12 m offshore) along with mean (black line), minimum, and maximum (dashed black lines) lake levels of 10 minutes time windows (300 measurements). All results are smoothed by an additional moving average procedure; (B) Maximum Wave height \( H_{\text{max}} = (\text{maximum} - \text{minimum}) \) at the lake; and (C) lake, waterline and groundwater elevations: lake’s deep water mean elevation (12 m offshore, black line) and maximum elevation (dashed black line); lake’s shallow water mean elevation (4 m offshore, blue line); beach face borehole groundwater elevation (4 m onshore, magenta line); mean waterline elevation from the GoPro camera measurement (green diamonds with calculated error as vertical bars).
Figure 15: Correlation between maximum wave height ($H_{\text{max}}$) measured by a pressure transducer and shoreline wave setup (average water line) measured by the GoPro camera.

Figure 16: Mean water elevation profiles during the daed sea high waves phase (red) and still water phase (blue). The grey line marks the beach face topography (profiles are vertically exaggerated).
3.2. The contribution of wave setup and wave run up to groundwater level rise (Five days campaign)

During the five days campaign, a buried pressure transducer recorded the groundwater level changes within the swash zone (green line in Figure 17 and Figure 18). These measurements represent the actual groundwater level changes 1.4 m landward from the shoreline. Wave setup (magenta line in Figure 17 and Figure 18 A) is calculated from $H_{\text{max}}$ using shoreline wave setup = $0.54$ times $H_{\text{max}}$ (Figure 15). The correlation in Figure 18 B shows that the calculated shoreline set-up is 1.1 times the measured groundwater level at the shoreline. This observation implies that shoreline set-up alone is enough to generate the observed groundwater level rise, and no groundwater level rise is attributed to wave run-up mechanism that can increase groundwater level through water infiltration occurring landward from the average waterline (Figure 18 C). This means that the wave setup is the only mechanism responsible for the groundwater level increase during wave events at the study area. The increase of the water level due to waves decays from the shoreline landward, and cannot be recognized 90m from the shoreline (Figure 17 and Figure 19). There is a hydraulic gradient inversion very near the shoreline whereby at times with no waves the gradient is from land towards the sea and opposite gradient during high waves (Figure 19). During a wave event, groundwater elevation at the buried pressure transducer is higher than at observation well EG-27 which is located 8m inland from the shoreline.
Figure 17: Mean, maximum, and minimum of the lake water level (black), along with groundwater levels of four observation wells, and one pressure transducer buried under the swashzone, 1.4 meters inland from the shoreline (green). Shoreline setup (magenta) is calculated from the maximum wave height using the correlation found in Figure 15. The distances of the boreholes from the shoreline are: EG-27 (blue) – 8m; EG-25 (grey) – 30m; EG-28 (brown) – 75m; EG-22 (purple) – 90m. The datum for the water level was shifted to emphasize the different scale water level change in each observation point.
Figure 18: (A) Time series of the calculated shoreline set-up elevation (magenta line) and $h_{\text{wash}}$ - groundwater elevation measured by the buried pressure transducer (BPD) 1.4 m landward from the shoreline, under the swash zone (green line). (B) Correlation between $h_{\text{wash}}$ and shoreline set-up elevation ($R^2 = 0.94$). (C) Schematic diagram of mean water surface (MWS) and groundwater level during a wave event (dashed blue line) vs. lake’s still water level (SWL) and groundwater level during the lake’s still water phase (blue line). Green arrow represent groundwater level rise at the swash zone and magenta arrow represents shoreline elevation rise (shoreline set-up). RL is the wave run-up limit.
Figure 19: Groundwater elevations and Dead Sea maximum wave height ($H_{\text{max}}$) during the five day campaign. Observation well name and its distance from the shoreline are given above each curve. Black arrows mark times of inversed hydraulic gradients between the buried pressure transducer (BPD) under the swash zone and EG 27 observation well, located 1.4 m and 8 m respectively away from the shoreline.

Wave forcing signal (set-up and run-up) amplitude attenuation and time lag in the aquifer were measured relative to swash zone groundwater level (buried transducer located 1.4 m landward from the shoreline) time series (Figure 20). The amplitude attenuation in the aquifer (Figure 20 A) is decreasing very rapidly at relatively short distances from the shoreline (1.4 m – 8 m), and more moderately at distances farther away (8 m - 75 m). Time lag of wave forcing signal in the aquifer is increasing linearly with distance away from the shoreline (Figure 20 B)
Figure 20: Amplitude attenuation and time lag distributions of four wave events, during the five days campaign, measured relative to the swash zone (buried pressure transducer, 1.4 m from the shoreline) at 3 observation wells located at different distances from the shoreline (EG27, 8m; EG25, 30m; EG28, 75m). (A) Groundwater level rise amplitude $\Delta h$, is measured in each wave event as the difference between peak groundwater level and the pre-event groundwater level. $\Delta h^*$ is the amplitude of groundwater level change normalized to the amplitude of swash zone groundwater level change ($\Delta h^* = \Delta h(x) / \Delta h(1.4m)$). (B) Time lag of each wave event is determined as the time difference between the time of peak groundwater level in each well, and the time of peak groundwater level measured at the swash zone.
3.3. Wave forcing and lake level effects on the groundwater level (2.5 month campaign)

The long term level decline of the Dead Sea affects the groundwater level in all observation wells that were measured in this study (Figure 21). Groundwater levels decrease at the same rate as the Dead Sea level (4.5 mm/day during the second part of the 2.5 months campaign - Figure 21). Similar behavior was found at an observation well farther from the Dead Sea shoreline, up to distance of ~700 meters (Yechieli et al., 2010), in the Darga alluvial fan (12 km north from the location of this study). However, the daily wave events decay with distance from the shoreline and cannot be recognized 90m from the shoreline (Figure 21). The long term measurement (2.5 months) shows that there is almost a daily wave event. Detailed observation during a specific wave event of the 21-22/04/2014, shows small time lag (0 to 60 minutes) between the $H_{\text{max}}$ and the maximum groundwater elevation (Figure 22). In this specific case time lag is not detectable until a distance of 8 m away from the shoreline, but it increases to 15 and 60 minutes at distances of 30 and 75 meters respectively away from the shoreline.

As opposed to the wave event of the 21-22 of April 2014 where lake’s $H_{\text{max}}$ time series shows a sharpen peak (Figure 22), the duration of the 4-5 of April 2014 wave event is much longer (~10 hours) with a quite stable $H_{\text{max}}$ (Figure 23). Here, Wave forcing effect on the aquifer leads to a quasi-steady water level, at different distances from the shoreline, which follows the pattern of $H_{\text{max}}$ time series. This finding implies that in the gravel coastal aquifer of the Dead Sea, the amplitude of groundwater level rise due to wave forcing is fast (< 1 hour).
Figure 21: Maximum wave height ($H_{\text{max}}$), Dead Sea mean level, and groundwater elevations with linear trend lines at four observation wells with different distances from the shoreline. Observation well name and its distance from the shoreline are given above each curve.
Figure 22: Maximum wave height ($H_{max}$), Dead Sea mean level, and groundwater elevations at four boreholes with different distances from the shoreline. Borehole name and its distance from the shoreline are given above each curve. Time lag is measured according to the vertical lines that indicates the peak amplitude of the wave and water level signal.
Figure 23: Long duration wave event. Maximum wave Height ($H_{max}$), Dead Sea mean level, and groundwater elevations at four boreholes with different distances from the shoreline. Observation well name and its distance from the shoreline are given above each curve. All measurements are represented by black crosses and the time series moving average is represented by black lines.
During winter, in the first part of the 2.5 month campaign, two flash floods events in Arugot Wadi were visible as groundwater level rise (Figure 24). The floods effect on groundwater level was characterized by increasing amplitudes with distance from the shoreline, due to the lake level influence on the aquifer boundary (for a more detailed view see Appendix 4). The first flash flood, occurring from 16/2/2014 led to a short period (~2 days) groundwater level rise with higher amplitudes at observation wells located 75 m and 90 m away from the shoreline and was not accompanied by any Dead Sea water level rise. The second flash flood, occurring from 12/3/2014 was actually a combination of four flooding fronts that overlapped each other and lasted for approximately 6 days. This series of flooding was part of a major storm event in the Dead Sea basin as it was accompanied by a ~10 cm rise of the lake’s level. According to the short recovery time (~2 days) of groundwater level after the first small flash flood, it is reasonable to claim that lake’s level rise is responsible for most of the apparent groundwater level rise in all observation wells during and after the second large flash flood. This means that the effect of lake level rise on groundwater level reach farther away from the shoreline (at least 90 m), than does the effect of wave forcing.
Figure 24: Maximum wave Height ($H_{\text{max}}$), Dead Sea mean level, and groundwater elevations at four observation wells with different distances from the shoreline during flash floods, lake level rise and wave events. Observation well name and its distance from the shoreline are given above each curve. Magenta arrows represent mean sea level and far most observation well water level rise ($dH$). Two flash flood events are visible: a small flash flood accompanied by a stable lake level (left rectangle) and a group of flash floods accompanied by a ~10 cm lake level rise (right rectangle).
4. Discussion

Distinguishing between wave forcing and tide forcing on coastal aquifers is challenging because both act with similar time scales. In a lake such as the Dead Sea, tide forcing is negligible and a clear diurnal cyclic wave activity is present. The detailed measurements of both waves in the lake (at very close distance from the shoreline) and water levels in observation wells, together with the exact measurements of the variations in the lake's mean water surface (wave setup), facilitate the study of wave forcing effect on groundwater levels. The combination of both factors (no tide and detailed measurements at the shoreline) allows the distinction between wave setup and wave run-up mechanisms in the Dead Sea system, which is very difficult in coastal aquifers that are subjected to large tide signal.

Field observations during the one day and five days campaigns show that the entire groundwater level rise due to wave forcing is caused by wave set-up mechanism (Figure 18 A, B). An explanation for this might be inferred from (Figure 26), where outcrop of the beach sediment exhibits a gravel layer covering a fine sand layer. This change in lithology leads to anisotropic hydraulic conductivity near the shoreline and should cause higher water flow in the horizontal direction, along the gravel layer, after each wave that run up landward from the mean waterline. Such anisotropic hydraulic conductivity will prevent a significant contribution of the run-up infiltration mechanism to groundwater level rise.

The response of the groundwater level near the shoreline (8-30 m) to the wave forcing, as wave set-up at the shoreline, is almost immediate (time lag <5min) indicating a very high hydraulic diffusivity of the gravel coastal aquifer. At the observation well located 75 m landward from the shoreline, the amplitude of groundwater level response to wave
forcing decreases, and the time lag increases (~1 hour). The high hydraulic conductivity was inferred from lake level effects on groundwater level during this study’s 2.5 month campaign and also from the long term annual respond of groundwater in the Dead Sea coastal gravel aquifer (Kiro et al., 2008). The transmissivity of the aquifer was estimated in an interference pumping test in Arugot Wadi to be 1500 m²/day (Wollman et al., 2003), yielding a hydraulic conductivity value of 30-100 m/day (Yechieli et al., 2010), depending on the specific thickness of the aquifer which is not known. Groundwater level response to wave forcing at different distances from the shoreline was used by Rotzoll and El-Kadi (2008) to estimate a coastal unconfined aquifer hydraulic parameters in Hawaii, where ocean wave forcing effect on distant observation wells was clean of any ocean tidal forcing signal. Rotzoll and El-Kadi (2008) assumed that wave set-up signal could be used in the same manner of tide signal to find hydraulic diffusivity using the signal time lag and amplitude attenuation at different distances from the shoreline, and assuming water flow in one dimension between the ocean and the aquifer. In this study, field measurements suggest that mean water level geometry at the shoreline (Figure 25A) creates two main opposite water flow directions, during a wave event, thereby making the influence of wave forcing signal on groundwater to be different than tide signal influence (Figure 25B). With wave forcing, the highest water table is created on the shoreline, making it complex to model its relationship with groundwater levels in the aquifer. Calculating the hydraulic diffusivity of coastal aquifers using wave forcing (wave set-up and run-up) signal should give lower values than using tide signal, because a major part of the additional water coming from the ocean, are circulated back through the aquifer due to hydraulic gradient from the shoreline seawards (Figure 25A).
Although the long term decline of the Dead Sea level is visible in all observation wells that were monitored in the study area, and reach even farther distances (~500m; Yechieli et al. 2010), the effect of the daily wave events on groundwater level decays with distance from the shoreline and cannot be recognized at distance of ~90m from the shoreline. The decay could be explained by the wave set-up being constricted only to the shoreline and not throughout the entire region of the lake (Figure 25). Actually, the wave events do not change the mean water level of the lake; only in immediate vicinity of the shoreline (less than 12 meters from the still water shoreline in this case study) as wave set-down and wave set-up. Thus, most of the excess pressure created by wave set-up at the shoreline is released at the location of set-down and increases the wave forcing signal amplitude attenuation. This study shows that the signal of high water level at the shoreline, due to wave forcing, has different amplitude attenuation in the coastal aquifer than the attenuation of mean lake level signal, which is, in this case the Dead Sea continuous level decline and its occasional level rise (Figure 21 and Figure 24 respectively). In this specific aquifer, the effect of wave forcing on groundwater level is less dependent on wave event duration than on the maximum wave height (Figure 23).

During winter time, the observed 10 cm lake level rise, caused by major flash floods (March 2014), affected all observation wells in the study area by raising the groundwater level (Figure 24). This finding was mostly significant in observation well EG-22, located 90 m away from the shoreline, which showed no response to wave forcing throughout this study. The groundwater level rise caused by the above mentioned lake level rise during winter, allows a comparison between wave forcing level rise and mean lake level rise mechanisms as shown in Figure 25, and resembles the comparison between wave forcing and tide forcing water level change discussed before.
The obtained data in this study at the Dead Sea indicate that there is an upper saline plume (USP) created by a wave driven recirculation (Figure 2) of saline water from the lake, which occurs generally within about 10 meters from the shoreline. The specific distance could reach more than 8 m and less than 30 m in the example given here (Figure 19), and varies depending on wave amplitude. Similar results were found in previous works, based on modeling and laboratory experiments, showing that circulation could be extending to distances of 20 meters depending on hydraulic conductivity and wave amplitude (Bakhtyar et al., 2013, 2012; Li and Barry, 2000; Longuet-Higgins, 1983; Robinson et al., 2014; Xin et al., 2010). Although groundwater salinity fluctuation during wave events was not measured in this study, we suggest that the wave driven water circulation should lead to the formation of an upper saline plume (Figure 2), and increase the water exchange between the Lake and the aquifer as have been shown numerically by previous studies (Robinson et al., 2014; Xin et al., 2010). Xin et al. (2010) found that this mechanism could be responsible to 60% of the total submarine groundwater discharge.
Figure 25: Schematic diagram of two lake processes (not to scale), wave event (A) vs. regional lake level rise (B). Groundwater level and lake’s mean water surface (MWS) are represented by dashed blue lines. Still water level (SWL) is the lake water level in the absence of wind and waves. Green arrows represent the direction of groundwater level change and Blue arrows represent water flow due to water level geometry in each process. RL is the waves’ run-up limit.
Figure 26: piezometer installation 4 meter landward from the shoreline on the one day campaign. The red line marks the contact between the gravel and the sand layers. Groundwater is exposed at the bottom of the borehole.
5. Summary and Conclusions

The Dead Sea resembles oceanic coastal environments by the presence of a fresh-saline water mixing zone, on the one hand, and show negligible tide oscillations on the other hand. Thus, it can be used as a field study for examining wave forcing effect on groundwater in coastal aquifers without any tidal interference.

In this study, we distinguish between the contribution of wave set-up and run-up to groundwater levels, for the first time in a field study. We find that for the case of the Dead Sea gravel coastal aquifer, the groundwater level increases during wave events only due to wave set-up mechanism, compared to the groundwater levels during still water level phase of the lake. Wave run-up infiltration mechanism does not significantly contribute to groundwater level rise in this case study, probably due to anisotropic hydraulic conductivity near the shoreline. This conclusion could be achieved due to the fact that the Dead Sea tidal forcing is negligible and it has a strong diurnal wave action, and due to accurate and detailed water level measurements that were conducted on both groundwater and lake surfaces.

The influence distance of the wave forcing effect on groundwater level is smaller than the influence distance of mean lake level rise, because of the change in mean water surface geometry near the shoreline caused by braking waves. Field measurements, conducted in this study, by a video camera and pressure transducers, detected this change in mean water surface during wave events, as a rise at the shoreline (wave set-up) and as a fall further offshore (wave set-down).

Wave events generate groundwater level rise in gravel coastal aquifer that reaches tens of meters landward from the shoreline. The time lag of this effect is very small (<5 minutes) nearshore and it increases linearly with distance landwards from shoreline.
Wave events in the Dead Sea generate saline water recirculation at the near-shore zone of the adjacent coastal gravel aquifer. This is a result of the changing hydraulic gradient direction near the shoreline, caused by wave set-up.

This study results demonstrate the ability of coastal aquifers near saline lakes, such as the Dead Sea, to serve as field sites for studying wave forcing effects on coastal groundwater. It calls for further field investigations on the wave forcing effect on the groundwater salinity distribution and on the water and solutes fluxes in the Dead Sea coastal aquifer.
6. Bibliography


Appendix 1 – observation wells log and location of observation points in the study area

EG - 22
Coordinates (ITM) 237955.3/596485.8
Elevation (ASL): -405.824 m

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Drilling date: 20/10/2007

Figure 27: Drilling log of observation well EG – 22 (modified after Yechieli and Shalev, 2008).
Figure 28: Drilling log of observation well EG – 23 (modified after Yechieli et al., 2013, GSI technical report, in preparation).
Figure 29: Drilling log of observation well EG – 25 (modified after Yechieli et al., 2013, GSI technical report, in preparation).
**EG - 27**
Coordinates (ITM) 238054.2/596670.8
Elevation (ASL): -421.53 m

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<tr>
<td>Gravel with some sand (conglomerate)</td>
<td><img src="image" alt="Gravel with some sand (conglomerate)" /></td>
</tr>
<tr>
<td>Gravel with some sand (conglomerate) and halite crystals</td>
<td><img src="image" alt="Gravel with some sand (conglomerate) and halite crystals" /></td>
</tr>
</tbody>
</table>

Drilling date: 1/12/2015

**Figure 30:** Drilling log of observation well EG – 27 (modified after Yechieli et al., 2016, GSI technical report, in preparation).
**Figure 31: Drilling log of observation well EG – 28** (modified after Yechieli et al., 2016, GSI technical report, in preparation).
Table 2: Observation points at Ein Gedi study area. Location and elevation were measured using GPS-RTK (Topcon GR3) and optical level (Leica Jogger 30).

<table>
<thead>
<tr>
<th>Location name</th>
<th>North coordinate</th>
<th>East coordinate</th>
<th>Reference point elevation ASL (m)</th>
<th>Well depth (m)</th>
<th>Depth to water level on 21/7/2014 (m)</th>
<th>Distance from shoreline (m)</th>
<th>Distance from wadi bank (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EG-22</td>
<td>596485.77</td>
<td>237955.30</td>
<td>-405.82</td>
<td>30.0</td>
<td>21.31</td>
<td>90</td>
<td>80</td>
</tr>
<tr>
<td>EG-23</td>
<td>596607.28</td>
<td>238049.92</td>
<td>-420.03</td>
<td>20.0</td>
<td>8.09</td>
<td>20</td>
<td>55</td>
</tr>
<tr>
<td>EG-25</td>
<td>596637.69</td>
<td>238041.89</td>
<td>-421.53</td>
<td>10.0</td>
<td>6.57</td>
<td>30</td>
<td>26</td>
</tr>
<tr>
<td>EG-27</td>
<td>596670.77</td>
<td>238054.17</td>
<td>-425.09</td>
<td>20.0</td>
<td>3.14</td>
<td>8</td>
<td>5</td>
</tr>
<tr>
<td>EG-28</td>
<td>596634.05</td>
<td>237995.82</td>
<td>-422.92</td>
<td>20.0</td>
<td>4.65</td>
<td>75</td>
<td>5</td>
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<tr>
<td>Beach face PZ</td>
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<td>238057.60</td>
<td>-427.67</td>
<td>1.0</td>
<td></td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Measuring bar western end</td>
<td>596682.06</td>
<td>238058.57</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Measuring bar eastern end</td>
<td>596683.35</td>
<td>238062.22</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Appendix 2 – groundwater temperature, density and conductivity profiles at observation well EG-23

Temperature profiles of the coastal gravel aquifer measured at observation well EG-23, located 20 meters from the shoreline are affected by groundwater level changes. A continuous measurement (every 15 minutes) of the temperature profile and groundwater level done on 27/2/2013 – 16/5/2013 using a thermistor chain (30 instruments of Onset - Hobo Water Temperature Pro V2 Data loggers, in increments moving between 0.25 to 1 meter) and a pressure transducer (Solinst – Levelogger Junior Model 3001) showed correlation between frequencies of groundwater level fluctuations and Saline-water temperature profile fluctuations (Figure 32). The temperature profile of the fresh-saline water mixing zone is not affected by the water level fluctuations. This effect was measured only at observation well EG-23 and needs to be further investigated in order to rule out the existence of artificial well effects on the results. The relatively stable temperature of the fresh-saline mixing zone could be an indicator for continuous monitoring of its location, which is impossible to measure using water electric conductivity profiles, due to extremely high salinity of the Dead Sea water.
Figure 32: Density, Water temperature and water level measurements at observation well EG-23. A) Water density profile measured at EG-23 on 27/2/2013. WL (blue line) represents the water level (all depths in the figure are measured from the well reference point), MZ is the fresh-saline water mixing zone and SW is Saline water zone (Dead Sea Salinity). B) Contour map of air and water temperature time series at EG-23. C) Water level measurement at EG-23. Black arrows mark part of the daily scale level fluctuation that show approximately the same fluctuation frequency of the saline-water temperature profile.
Appendix 3 – density evaluation and groundwater level correction

High salinity Water level measurements in the Dead Sea and in its coastal aquifer, using pressure transducers, require a correction for water density. It assumed that water level rise in the well will also raise the fresh-saline water interface. It means that measurements of rising pressure in an observation well, could be a combination of two factors: water level rise and water column density increase above the pressure transducer (P=ρ*g*h, P is pressure, ρ is water density, g is gravity acceleration constant and h is water column height). Pressure transducer Water level measurement must be referenced to manual water level measurements in order to know the exact change in water level and density. This could be problematic when high accuracy and high frequency measurements are needed. In order to get better water level measurements one needs to know the density structure of the well.

Groundwater density profile along with temperature and electrical conductivity was measured in the study area at observation well EG-23 on the 27/2/2013 and on the 16/5/2013 (Figure 33). Water Density measurement results show that the thickness of the fresh-saline mixing zone is ~ 3 meters. This conclusion can’t be simply inferred from the electric conductivity profile of EG-23 (Figure 33, middle plot). Density gradient in the fresh-saline mixing zone is ~0.05g/cm³ per m (Figure 33). Thus it is possible to assume In the study area, that groundwater level small fluctuations (few cm) does not significantly change the total water column density above the pressure transducers, installed between 2-3 meter below groundwater table.

Although groundwater density profile in the study area fluctuates along time due recharge and due to the continuous Dead Sea level decline, it was assumed to be constant on the hourly and daily scale.
Figure 33: water Density, electric conductivity (E.C.) and Temperature profiles measured at observation well EG-23, located 20 meters from the shoreline. The parameters were measured on 27/2/2013 (black plots) and on 16/5/2013 (red plots). Water electric conductivity and Temperature were measured in-situ while water density was measured outside of the observation well, after pumping water samples using peristaltic pump.
Appendix 4 – Arugot wadi flash floods effect on the coastal gravel aquifer groundwater level

Flash floods at Arugot wadi occurred on February and March 2014. Discharge of these floods was measured by the Israel water authority (Galili, personal communication, 2014). During flooding groundwater level, lake level and wave height were measured in the study area (Figure 34), and the results are shown in Figure 35 and Figure 36.

Flash floods in Arugot wadi raise groundwater level in the Dead Sea gravel coastal aquifer as a function of distance from the wadi bank and the distance from the shoreline. Groundwater level at observation wells close to the shoreline (8, 20 and 30 meters) rise less by the floods than far away observation wells (75 and 90 meters). Also, groundwater level rise is higher as the distance to the wadi is closer.

Figure 34: Study area map. Observation wells are represented by green circles and lake level meter is represented by red circle. Blue lines represent the Dead Sea shoreline and Wadi Arugot banks.
Figure 35: The effect of the February 2014 flood in Arugot wadi on groundwater levels. Water discharge at Arugot wadi (upper plot, data taken from Israel water authority; Galili, personal communication, 2014), Maximum wave height ($H_{\text{max}}$, magenta line), Dead Sea mean level (black line), and groundwater elevations at five observation wells at different distances from the shoreline and from Arugot wadi-bank are presented. Observation well name and distance from the shoreline (first distance value) and from Arugot wadi-bank (second distance value) are given next to each curve.
Figure 36: The effect of the March 2014 flood in Arugot wadi on groundwater levels. Water discharge at Arugot wadi (upper plot, data taken from Israel water authority; Galili, personal communication, 2014), Maximum wave height ($H_{max}$, magenta line), Dead Sea mean level (black line), and groundwater elevations at five observation wells at different distances from the shoreline and from Arugot wadi-bank are presented. Observation well name and distance from the shoreline (first distance value) and from Arugot wadi-bank (second distance value) are given next to each curve.
Appendix 5 – waterline measurement calibration and results

Figure 37: Snapshots of Ein Gedi beach face from Gopro camera during each movie taken on the one day campaign (6-7/10/2014). Start time and file name of each video are specified on the snapshots.
Figure 38: Snapshots of measuring bar calibration from Gopro camera during each movie taken on the one day campaign (6-7/10/2014). Axes of the tracker software view are represented by magenta lines. Calibration points on the measuring bar are represented by cyan diamonds. Start time and file name of each video are specified on the snapshots.
Figure 39: calibration curves of the for each Gopro camera video. The x axis represents Tracker software x axis as shown in Figure 38. The y axis represents the length of the measuring bar. The file name of the video and the curve equation are written next to each curve.
Figure 40: Waterline locations on measuring bar. Snapshots of Ein Gedi beach face from Gopro camera during each movie taken on the one day campaign (6-7/10/2014). Start time and file name of each video are specified on the snapshots. Shoreline location measurements of each movie are represented by red diamonds.
Table 3: Summary of calibration parameters for waterline video measurements, i.e. measuring bar start value of calibration and Tracker software view axes origin and tilt angle. B) Summary of waterline measurement results.

### A) Summary of Calibration Parameters

<table>
<thead>
<tr>
<th>Data Description</th>
<th>Measuring Bar Calibration Data</th>
<th>Video Image Origin</th>
<th>Calibration Start Point on Measuring Bar (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Local Time (gmt-2)</td>
<td>Results Interpolated</td>
<td>Wave Event Stage</td>
<td>File Name</td>
</tr>
<tr>
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<td>GOPR2124</td>
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<tr>
<td>06/10/14 15:23</td>
<td>yes</td>
<td>low waves</td>
<td>GOPR2124</td>
</tr>
<tr>
<td>06/10/14 17:06</td>
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<td>still water</td>
<td>GOPR2242</td>
</tr>
<tr>
<td>06/10/14 17:30</td>
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<td>medium waves</td>
<td>GOPR2279</td>
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<tr>
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<td>GOPR2384</td>
</tr>
<tr>
<td>07/10/14 08:12</td>
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<td>high waves</td>
<td>GOPR3260</td>
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</tbody>
</table>

### B) Summary of Waterline Measurement Results

<table>
<thead>
<tr>
<th>Data Description</th>
<th>Results (relative to measuring bar)</th>
<th>Results (georeferenced)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Local Time (gmt-2)</td>
<td>Results Interpolated</td>
<td>Wave Event Stage</td>
</tr>
<tr>
<td>06/10/14 12:15</td>
<td>yes</td>
<td>medium waves</td>
</tr>
<tr>
<td>06/10/14 15:25</td>
<td>yes</td>
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</tr>
<tr>
<td>06/10/14 17:06</td>
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</tr>
<tr>
<td>06/10/14 22:43</td>
<td>yes</td>
<td>medium waves</td>
</tr>
<tr>
<td>07/10/14 08:12</td>
<td>yes</td>
<td>high waves</td>
</tr>
</tbody>
</table>
המכון הגיאולוגי
משרד התשתיות האמונית
האנרגיה והמים

השעת ארועי גלים בים על מפלס מי התהום באקווייפר חוף:
תצפית מי המלח

הלל צבקי

עבדה זו ה♧ossaך כחיבור לקיבול תואר "מוסמך" באוניברסיטת בן גוריון בנגב.

העבודה נשענת בהדרכות של:

ד"ר אייל שלף, המכון הגיאולוגי, ירושלים.
ד"ר נדב لنסקי, המכון הגיאולוגי, ירושלים.
פרופ. נעים וייסברוד, מכון עקרבר לחקר המים, אוניברסיטת בן גוריון בנגב.

יושלמה, איר טshe',מאי 2017

GSI/09/2017