Magnetic Properties of Late Holocene Dead Sea Sediments as a Monitor of Regional Hydroclimate

Y. Ebert1, R. Shaar1, E. J. Levy2,3, X. Zhao4, A. P. Roberts4, and M. Stein1,3

1The Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel, 2Department of Geological & Environmental Sciences, Ben-Gurion University of the Negev, Beer-Sheva, Israel, 3Geological Survey of Israel, Jerusalem, Israel, 4Research School of Earth Sciences, Australian National University, Canberra, ACT, Australia

Abstract Diagenetic processes in anoxic sedimentary environments influence sediment magnetic properties mainly through dissolution of detrital magnetite and precipitation of authigenic greigite. Recently exposed late Holocene Dead Sea sediments provide an opportunity to study the processes governing greigite formation and preservation, and their relation to different hydrological settings. Magnetic data and pore-fluid compositions were obtained from three Holocene sections along a N-S transect on the western Dead Sea shore: Og, Ein-Feshkha (EF), and Ein-Gedi. The northern sections are closer to the major freshwater source to the Dead Sea-the Jordan River. Detrital titanomagnetite is present at all sections, but greigite is the dominant magnetic phase at Og and EF. Bulk rock magnetic data vary between and within the sections by over 3 orders of magnitude, where higher values indicate higher greigite concentrations. At the three sites, pore fluids have similar or lower salinity than the modern and Holocene Dead Sea brine, with variable and dissolved iron (Fe2+) and sulfate (SO4^2-). Magnetic property changes are reflected by iron and/or sulfate microbial reduction that controlled sedimentary greigite formation. We propose that the N-S greigite decrease suggests that anoxic microbial activity was controlled by labile organic matter and/or reactive iron brought by, or formed as a result of, freshwater influx from the Jordan River. Hence, greigite concentration changes depended on past freshwater input to the hypersaline lake and proximity to the freshwater source. The apparent relationship between hydrological conditions and magnetic properties provides a new method to trace past hydrological changes in the Dead Sea.

1. Introduction

Microbial sulfate reduction (MSR) occurs commonly in reducing redox environment (Canfield & Berner, 1987; Morse & Cornwell, 1987; Wilkin & Barnes, 1997) where it can affect significantly the magnetic properties of sediments by causing titanomagnetite dissolution and by triggering authigenic greigite (Fe3S4) precipitation (Roberts & Weaver, 2005; Roberts et al., 2011; Ron et al., 2006; Rowan & Roberts, 2006; Rowan et al., 2009; Sagnotti, 2018). MSR is sensitive to environmental variabilities of parameters such as availability of sulfate, reactive iron, organic matter, and salinity. In this paper, we characterize these environmental variables which dictate greigite formation in the late Holocene Dead Sea lake. High sedimentation rates (millimeters per year) and slow sulfate reduction activity enable a detailed investigation of the processes that govern greigite formation and their sensitivity to environmental changes. We combine geochemical and magnetic data from several exposed late Holocene Dead Sea sediment sections to explore the effects of hydrological conditions on diagenetic microbial processes and greigite formation.

1.1. Limnological and Hydrological Setting

The Dead Sea is a terminal hypersaline lake that filled the Dead Sea Basin during the Holocene (Bookman et al., 2004; Migowski et al., 2006; Stein, 2001, 2014). The modern lake (total dissolved solids [TDS] of ~341 g/L) comprises a calcium chloride brine whose formation can be traced to ingestion of the Sedom lagoon into the basin in the late Miocene (Starinsky, 1974; Zilberman et al., 2017). In the past, the Dead Sea lake level was controlled by freshwater supply variations from the lake watershed which resulted in strong oxic stratification (Stein et al., 1997; Torfstein et al., 2008). Holocene lake level fluctuated between ~430 and ~370 m below mean sea level (bmsl) (Kushnir & Stein, 2019; Migowski et al., 2006). Short-lived isotopes indicate that the lake was stratified during the past few hundred years, while the lake overturned during...
the winter of 1979 due to decreasing lake level and increasing salinity (Stiller & Chung, 1984). The modern lake is regressing rapidly (by > 1 m/year), mainly due to human activity; the current lake surface is at 435 m bmsl. The calcium chloride brine that filled the Holocene Dead Sea (and previous Quaternary lakes) is poor in bicarbonate and sulfate and, thus, deposition of mineral phases such as aragonite and gypsum requires addition of sulfate and bicarbonate to the lake. In the modern Dead Sea, these ions are supplied by freshwaters and springs that discharge to the Dead Sea (Torfstein et al., 2008). Thus, both primary (aragonite and gypsum) and detrital sediments that comprise Dead Sea lacustrine formations contain important information on hydroclimate conditions in the lake watershed and global climate that dictate the synoptic conditions that control rain and dust transportation (e.g., Stein, 2014 and references there).

1.2. Sulfate and Iron Reduction in the Dead Sea

The high salinity and high concentration of Mg$^{2+} > 2$ M and Ca$^{2+} > 0.5$ M makes the Dead Sea a harsh environment for microorganisms (Oren, 2001). Nevertheless, sulfate-reducing bacteria (SRB) activity has been detected in the Dead Sea water column, springs, and sediments from sulfur isotopic analyses of sulfide and sulfate (Bishop et al., 2013; Gavrieli et al., 2001; Häusler et al., 2014; Ionescu et al., 2012; Neev & Emery, 1967; Nissenbaum & Kaplan, 1976). Thomas and Ariztegui (2019) analyzed fluid inclusions from within halite and suggested that both archaea and bacteria are involved in the sulfur cycle in the Dead Sea. Overall, the high salinity and poor organic matter content limit MSR activity in the Dead Sea (Häusler et al., 2014; Ionescu et al., 2012; Thomas & Ariztegui, 2019; Thomas et al., 2016).

Little is known about iron reduction in the Dead Sea. Nishri and Stiller (1984) showed that iron reduction occurs in the water column and in sediment pore fluids. They studied the distribution and dissolution of iron and showed that 4,000 tonnes of allochthonous iron enters the Dead Sea via the Jordan River. Sedimentary iron sulfide formation – mackinawite (Fe$^{2+}$S$_2$), greigite (Fe$^{2+}$Fe$^{3+}$S$_4$), and pyrite (Fe$^{2+}$S$_2$) – depends on the availability of iron and sulfides. When iron reduction outcompetes and overcomes sulfate reduction, dissolved sulfide is removed rapidly to form and preserve greigite (e.g., Kao et al., 2004).

1.3. Magnetic Properties of Dead Sea Sediments

Dead Sea sediments contain two dominant magnetic minerals: titanomagnetite and greigite (Ebert et al., 2018; Frank et al., 2007a, 2007b; Ron et al., 2006). Titanomagnetite is transported to the lake by fluvial and alluvial systems (e.g., Jordan River), while greigite is a diageneric product of MSR activity. Frank et al. (2007a, 2007b) investigated four Holocene sediment cores collected along the western Dead Sea shoreline and found large depth variations of magnetic susceptibility, isothermal remanent magnetization (IRM), anhysteretic remanent magnetization (ARM), and the ratio between the low-field IRM and susceptibility ($\chi$). They interpreted these variations as reflecting magnetic mineralogy changes, where greigite-rich sediments have much higher $\chi$ and IRM/$\chi$ ratio values. However, the controlling mechanism of such variations has remained unclear.

In this study, we aim to reveal the mechanism that controls greigite formation in Dead Sea sediments. Additionally, we assess the effects of greigite on sediment magnetic properties and its environmental implications. We investigate three Holocene sedimentary outcrops, which are exposed freshly in new gullies that were entrenched in the lake floor during the (man-made) retreat of the modern lake (Figure 1). From each outcrop, we assemble a detailed composite depth profile of magnetic parameters and pore-fluid geochemistry, where intervals with sharp magnetic parameter changes were further characterized using electron microscopy and first-order reversal curve (FORC; Pike et al., 1999) measurements.

2. Methods

2.1. Sampling Sites

Sampling sites (marked in Figure 1) are located in deep gullies that expose sediments of the Holocene Ze’elim Formation. This unit comprises fine silty detrital sediments (the Dead Sea “muds” or the ld facies [laminat-
ed detritus] described by Haliva-Cohen et al. [2012]) whose major minerals are quartz and calcite along with sequences of laminated aragonite and silty detritus (the aad facies, described by Machlus et al. [2000]). Although the outer surfaces of gully walls appear altered and oxidized, digging less than a few centimeters into the walls reveals dark sediments that contain anoxic hypersaline pore fluid. Images of representative outcrop sediments are presented in Figure 2. The sites selected for this study are from north to south: Nahal Og (Og) (31.7400°E; 35.49029°N), Ein-Feshkha Nature Reserve (EF) (31.70839°E; 35.45524°N), and Ein-Gedi Spa (EG) (31.41917°E; 35.38486°N). The first two sites are located close to freshwater springs (e.g., the EF spring system, which is currently the main freshwater supply to the Dead Sea after damming of the Jordan River in 1964). EG Spa is located close to saline springs that discharge calcium chloride brine into the lake. The studied sedimentary sequences span a similar time interval of \( \sim 2,800 - \sim 1,200 \) years before present (BP). The chronology of each site is based on new radiocarbon ages (Table S1 and Figure S1) on terrestrial organic debris that are integrated with the existing data (EF [Kagan et al., 2011], EG [Migowski et al., 2006]). The Og section was deposited between \( \sim 2,400 \) and \( \sim 1,600 \) years BP; the EF section was deposited between \( \sim 2,800 \) and \( \sim 1,200 \) years BP; and the EG section was deposited between \( \sim 3,000 \) and \( \sim 1,350 \) years BP. At each outcrop, we dug a vertical cross section and collected sediments in nonmagnetic plastic boxes (23 mm \( \times \) 23 mm \( \times \) 19 mm; Figure 2c) for magnetic analyses at \( \sim 25 \)-mm sampling intervals. The thickness of the intervals sampled at the investigated sections is 3.15, 4.6, and 2.6 m for Og, EF, and EG, respectively.

### 2.2. Magnetic Measurements

The magnetic measurement routine used in this study was as follows: ARM acquisition with a 0.1 mT direct current bias field and 100 mT AF field along the z-axis of the sample; AF demagnetization of the ARM in 11 steps from 5 to 110 mT; and acquisition of an IRM in a 1,500-mT field along the sample z-axis. In addition, all samples were weighed and the mass-normalized susceptibility was measured at 200 A/m at 976 Hz. ARM demagnetization was done using a 2G Enterprises superconducting rock magnetometer (SRM) 750 with in-line two-axis AF demagnetizer coils or using a 2G Enterprises RAPID SRM system with in-line two-axis coil demagnetizer and z-axis ARM coil; the ARM was acquired and measured using the RAPID system; IRM was imparted using an ASC pulse magnetizer and

**Figure 1.** Location map of investigated outcrops. (a) Digital elevation model map of the Dead Sea area. Red circles indicate sampling site locations along the Dead Sea. (b–d) Images of outcrops from north to south and estimated age spans with thicknesses of the studied sediment intervals: (b) Nahal Og, (c) EF, and (d) EG. EF, Ein-Feshkha; EG, Ein-Gedi; OG, Nahal Og.

**Figure 2.** Representative images of sedimentary facies exposed in the studied outcrop sequences (scale unit is centimeters). (a) Og, (b) EF, and (c) EG. Illustration of (a and b) the ld facies (laminated detritus), and (c) the aad facies (alternating aragonite and silty detritus laminae). EF, Ein Feshkha; EG, Ein Gedi; OG, Nahal Og.
was measured using an AGICO JR-6A dual speed spinner magnetometer at the Geological Survey of Israel (GSI), Jerusalem. Low-field magnetic susceptibility was measured using an AGICO MFK-1 Kappabridge system. Magnetic analyses were made at the Paleomagnetic Laboratory at the Institute of Earth Sciences, Hebrew University of Jerusalem. FORCs for selected samples were measured at the Black Mountain Paleomagnetic Laboratory, Australian National University (ANU), using a Princeton Measurements Corporation MicroMag™ vibrating sample magnetometer. FORC data were analyzed using xFORC (Zhao et al., 2017).

2.3. Electron Microscopy

Magnetic mineral extracts were isolated from selected intervals representing different magnetic behaviors from all three studied sections. Magnetic minerals were extracted from a mixture of alcohol and the entire sediment content of a sampling box, using a handheld rare-earth magnet within a plastic probe (Nowaczyk, 2011). The extracts were embedded in epoxy and were polished with a series of abrasives down to 1 μm. Mineral identification and composition were performed using two electron probe microanalysis (EPMA) instruments: (1) a JEOL JXA-8530F Plus at ANU and (2) a JEOL 8230 at the Institute of Earth Sciences, Hebrew University of Jerusalem. The EPMA instruments are equipped with an energy-dispersive X-ray spectrometer (EDS) and beam conditions were set to 15 keV for EDS analyses. Data were processed with a PRZ correction procedure, and all phases were analyzed using hematite and pyrite standards. Additional images were taken using a Quanta 200 environmental scanning electron microscope at the Harvey M. Krueger Family Center for Nanoscience and Nanotechnology, Hebrew University of Jerusalem.

2.4. Pore-Fluid Compositions

Sediment samples for pore-fluid chemical analyses were taken from each of the three sections in two phases. During the first sampling at EF, 25 samples were taken from a 2.5-m exposed section at 10-cm intervals. Around 50 cm of the exposed surface of the section was removed to ensure the removal of oxidized sediment. Using end-cut 20-ml syringes, sediment was added to 50-ml plastic falcon tubes flushed with Ar gas to prevent oxidation. During the second sampling, four sediment samples were taken from each outcrop at EF, Og, and EG. Here, relatively large quantities of undisturbed sediment were taken in nylon bags, which were analyzed for pore fluids at the GSI and Ben-Gurion University of the Negev. Pore fluid was extracted using both a centrifuge method (for SO4²⁻ and Fe²⁺ analyses) and a Carver© hydraulic press method (for major ions, salinity, and solution density). For the centrifuge method, pore fluids were extracted from sediments in 50-ml falcon tubes purged with Ar gas placed in a centrifuge at 9,000 rpm for 15 min. Pore-fluid solutions were subsequently purged again to remove any dissolved H2S and were then filtered using a 0.45-μm syringe filter. Solution density was measured using an Anton Paar DMA 35 density meter at the GSI. Major cation concentrations (Na⁺, K⁺, Ca²⁺, Mg²⁺, Sr²⁺) were analyzed using inductively coupled plasma atomic emission spectroscopy at the GSI. Bromide (Br⁻) and lithium (Li⁺) chloride concentrations were analyzed by inductively coupled plasma mass spectrometry (ICP-MS) at the GSI. Combined chloride (Cl⁻) and bromide (Br⁻) were measured by titration. Br⁻ values were measured using ICP-MS, and Cl⁻ concentrations could then be derived from titration values. Analyses are within a charge-balance error of <2%. SO₄²⁻ concentrations were measured using a Metrohm© compact ion chromatography flex at BGU. Fe²⁺ was measured on centrifuged samples by immediately adding pore fluid to 15-ml falcon tubes treated with ascorbic acid and ferrozine. Samples were then diluted with double-distilled water and Fe²⁺ was measured via spectrophotometry. TDS in pore-fluid solutions were calculated by summing cation and anion concentrations.

3. Results

3.1. Magnetic Data

Samples from the northernmost Og section yielded the highest χ values (around 10⁻⁶ m³/kg), whereas those from the southern EG section have the lowest values (10⁻⁷–10⁻⁸ m³/kg). Some parts of the EF section have high χ (10⁻⁶ m³/kg) values, similar to Og, whereas others have intermediate values (10⁻⁷ m³/kg) that are more similar to the higher values of the EG section (Figure 3). Variations in all three parameters (χ, ARM,
and IRM) are similar for each section, which indicates that \( \chi \) changes reflect ferrimagnetic mineralogy changes and are not due to paramagnetic contributions. Therefore, the magnetic parameters are affected by the concentration and grain size of two ferrimagnetic minerals that co-occur in the sediment: titanomagnetite and greigite (Frank et al., 2007b; Ron et al., 2006) (see Section 3.2). However, titanomagnetite and greigite have many similar magnetic properties (Roberts, 1995; Roberts et al., 2011), which makes it important to distinguish between the two minerals. It has been suggested that high IRM/\( \chi \) values are an indicator of greigite (Roberts, 1995; Sagnotti & Winkler, 1999; Snowball, 1991; Snowball & Thompson, 1990). Sediments from Og and part of the EF section are characterized by higher IRM/\( \chi \) values than the EG section (Figures 3j–3l). Lower IRM/\( \chi \) values in the EF section are similar to typical values in the EG section and correspond to intervals with intermediate \( \chi \) values. Taken together, different magnetic properties are revealed in Figure 3: (1) sediment intervals have high \( \chi \), IRM, ARM, and IRM/\( \chi \) values at the Og and EF sections (170–320 cm); (2) sediments have intermediate magnetizations at the EG (0–130 and 200–260 cm) and EF sections (0–170 and 320–460 cm); and (3) sediments have low magnetizations at the EG section (130–200 cm). In the following, we investigate the origin of these differences using electron microscopy and FORC diagrams. The analyses were made on sediments from specific elevations in the sections that represent the different magnetic property groups, as marked by stars in the upper panel of Figure 3.

Figure 3. Depth profiles of magnetic data: susceptibility (\( \chi \)), ARM, IRM, and IRM/\( \chi \) for each outcrop. Green, red, and blue denote Og, EF, and EG, respectively. Yellow stars mark the locations of samples that were studied in more detail, as shown in Figures 4–6. Trends of changes in \( \chi \), ARM, and IRM are similar at each outcrop. ARM, anhysteretic remanent magnetization; EF, Ein-Feshkha; EG, Ein-Gedi; IRM, isothermal remanent magnetization; Og, Nahal Og.
3.2. Electron Microscopy

Titanomagnetite and greigite are the only magnetic minerals identified in backscattered electron images. EDS analyses of the iron sulfide minerals indicate the presence of three dominant minerals: pyrite (FeS₂), greigite (Fe₃S₄), and mackinawite (FeₓS, x = 0–0.11). In general, darker mud is richer in iron sulfide aggregates compared to the gray mud, and its dominant ferrimagnetic mineral is greigite. The Og and EF sediments are richer in iron sulfide aggregates (Figures 4 and 5). Aggregates of greigite and a titanomagnetite grain from the Og section are shown in Figures 4a and 4b (sample locations are marked by stars in the upper panel of Figure 3a). Images for two samples from the EF section with different magnetizations are shown in Figures 5a–5c. An aggregate with a mixture of greigite and pyrite with intermediate magnetizations is shown in Figures 5a and 5b. The aggregates in Figure 5c comprise greigite with higher magnetizations (stars, upper panel of Figure 3b). Sediments from the EG section have lower iron sulfide contents compared to the other two outcrops, and the dominant magnetic mineral is titanomagnetite. Iron sulfides in the EG section are illustrated in Figure 6, including pyrite from a sample with weak magnetization (Figure 6a), and aggregates of greigite and titanomagnetite from samples with low and moderate magnetizations, respectively (Figures 6b and 6c).

Comparisons of electron microscope observations and mineral magnetic data (Figures 3–6) reveal a correlation with iron sulfide mineralogy. Samples with high magnetizations are dominated by abundant greigite aggregates (Figures 4a, 4b, and 5c). Samples with intermediate magnetizations are associated with a mixture of pyrite and greigite (Figures 5a and 5b) or only a few greigite aggregates (Figure 6c); samples with low magnetizations have fewer iron sulfide aggregates that consist mainly of pyrite (Figures 6a and 6b).
sedimentation rates affect the thicknesses of detrital and aragonitic parts of laminae, which could contribute to magnetic mineral concentration changes via variable aragonite dilution. Bias due to aragonite layers can at most affect the mass-normalized magnetization by a factor of $\sim 2$. Magnetic mineral concentration changes of orders of magnitude are observed between Og/EF and EG (Figure 3; upper three panels), so laminae thickness is not a critical driver of magnetic mineral concentration changes.

### 3.3. FORC Diagrams

FORC diagrams for samples imaged by electron microscopy are shown in the lower panels of Figures 4–6. The shapes of the FORC distributions are consistent with electron microscopy observations and magnetic data. Samples with higher magnetizations from the Og and EF sections, which are rich in greigite aggregates, are characterized by closed concentric contours with wide vertical spreading that are indicative of interacting greigite particles (Roberts et al., 2006, 2011, 2014). The contours are centered about $B_c$ values that range between $\sim 25$ mT (Figure 4c) and $\sim 40$ mT (Figure 5e), which is lower than for previously reported FORC diagrams for interacting single-domain greigite (e.g., Duan et al., 2017; Kelder et al., 2018; Roberts et al., 2006, 2011; Sagnotti et al., 2010). The peaks of the FORC distributions are located below the $B_i = 0$ line, as previously observed in greigite-bearing sediments (Rowan & Roberts, 2006). Samples from the EG section, which have lower iron sulfide contents, have two components (Figures 6d–6f): (1) vortex state to multidomain behavior (e.g., Pike et al., 2001; Roberts et al., 2017), associated with large titanomagnetite, as indicated by a divergent distribution along the $B_i$ axis with low $B_c$ values; and (2) narrow concentric contours around $B_c = 10$ mT, with lower coercivities than in Figures 4 and 5, which is indicative of the presence of additional greigite aggregates.
3.4. Pore-Fluid Chemistry

The distribution of major and some minor ions (Na⁺, K⁺, Ca²⁺, Sr²⁺, Li⁺, Cl⁻, Br⁻) in pore fluids for 35 samples from all sections varies slightly along the depth profiles, except for a few outliers (Figure S2). TDS profiles (Figure 7a), which represent salinity, have mostly uniform values with an average of 334 ± 45 g/L for all sections. TDS values are similar or lower than the range for Holocene pore fluids in the deep International Continental Scientific Drilling Program (ICDP) Dead Sea core (Levy et al., 2017, 2018). Mg²⁺ (Figure 7b) has an average value of 1.75 ± 0.33 mmol/L for all sections. Mg²⁺ is regarded as a conservative ion because pore fluids are undersaturated with respect to carnallite (KMgCl₃·6[H₂O]), the dominant evaporite that can mineralize Mg²⁺ in evaporated Dead Sea brines, and Mg/Br is mostly uniform (Levy et al., 2017). Given the low permeability typical of these sediments and the high dynamic viscosity of hypersaline solutions, which prevent advection and diffusion, the pore fluid may be a remnant of brine that has remained in situ since sediment deposition, similar to pore fluids from the deep ICDP Dead Sea core (e.g., Levy et al., 2017). The pore-fluid compositions are indicative of insignificant hydrological modification since recent sediment exposure (e.g., post-exposure evaporation or mixing with fresh groundwater), except for one sample from EG (25 cm), which has geochemical values similar to the Ein-Qedem (EQ) brine (marked in Figure 7 and Figure S2) (Weber et al., 2018), which suggests that the lower 20 cm of EG was washed by EQ brine. The main pore-fluid chemistry difference is in the concentration of dissolved Fe²⁺ and SO₄²⁻. EF has the highest Fe²⁺ concentration and EG has the lowest; however, EG has the highest SO₄²⁻ concentration. Comparisons between Fe²⁺ and SO₄²⁻ concentrations and χ in the studied sections are shown in Figures 8 and 9. Both Fe²⁺ and SO₄²⁻ concentrations in the pore fluids appear to track χ changes. These correlations are discussed in the following sections.
4. Discussion

4.1. Greigite Formation in the Dead Sea

We seek here to understand the processes that affect the magnetic properties of late Holocene Dead Sea sediments. Magnetic analyses and electron microscope observations indicate that the magnetic properties are controlled by the sedimentary greigite content. There is a clear N-S greigite concentration trend in the Holocene Dead Sea sediments (Figure 3). A major difference between these sites is their proximity to the main freshwater source to the hypersaline Dead Sea, the Jordan River, which raises the question of whether there is a connection between greigite formation and proximity to the primary freshwater source.

Several iron sulfides were identified in the studied sediments, which are produced during pyritization: mackinawite (Fe\(_{2+}\)S\(_{2-}\)), greigite (Fe\(_{2+}\)Fe\(_{3+}\)S\(_{2-}\)_4), and pyrite (Fe\(_{2+}\)S\(_{2-}\)). These minerals form as products of two reactants that form during early diagenetic anoxic microbial processes: (1) reactive iron (Fe\(_{2+}\))-which is a product of microbial iron (Fe\(_{3+}\)) reduction and (2) sulfide (S\(_{2-}\))-which is a product of MSR. Bishop et al. (2013) and Thomas et al. (2016) suggested that the transformation from monosulfide to pyrite in Dead Sea sediments requires addition of S\(_{0}\). For microbial iron and sulfate reduction to occur, an adequate supply of labile organic matter, iron (Fe\(_{3+}\)), and sulfate are required. In the modern Dead Sea, there is a large sulfate supply via freshwater runoff, mostly from the Jordan River (Figure 1), but also via spring discharge (Torfstein et al., 2008). The Jordan River is also an important iron source to the Dead Sea (Nishri & Stiller, 1984), which can be assumed to have remained active throughout the Holocene because pore fluids from the EF and Og sites have relatively large sulfate and dissolved iron (Fe\(_{2+}\)) concentrations (Figures 8 and 9). This suggests that sulfate and iron were supplied to the Dead Sea throughout the studied time interval.

Dissolved iron (Fe\(_{2+}\)), which is a product of iron reduction, indicates that this process occurred at the studied sites (Figure 7d). The pore fluids are rich in dissolved iron, which suggests that iron reduction outcompetes sulfate reduction and that iron sulfides will form promptly. Availability of dissolved Fe\(_{2+}\) favors greigite formation (Kao et al., 2004; Picard et al., 2018). The Og and EF sites (~5 km apart) are located near the Jordan River inlet to the Dead Sea, and are, therefore, expected to have higher reactive Fe\(_{2+}\) concentrations that promote higher greigite contents (Figure 8). However, the outcrops have variable dissolved Fe\(_{2+}\) concentrations and the 0–170 cm and 320–460 cm intervals at the EF section have lower \(\chi\) values than the 170–320 cm interval (Figure 8b). Dissolved Fe\(_{2+}\) variability at each site and compared to other sites may have variable causes; for example, variable redox conditions, different reactive Fe\(_{2+}\) supplies at different sites, or site-specific iron sinks (e.g., pyrite mineralization).

![Figure 7](image-url)
Similar to dissolved iron, sulfate appears to have been in ample supply (Figure 7c), and correlation with $\chi$ (Figure 9) may suggest a relationship between MSR and greigite formation. High dissolved Fe$^{2+}$ in pore fluids and abundant greigite imply that iron reduction is relatively high compared to sulfate reduction. However, precipitation/dissolution of calcium sulfate minerals can also result in decreased/increased sulfate concentrations, respectively (Levy et al., 2019). Although there are limitations in determining the factors that control dissolved Fe$^{2+}$ and sulfate distribution, there appears to be a similarity between changes in dissolved Fe$^{2+}$ concentrations (Figure 8), sulfate (Figure 9), and $\chi$. Together, this evidence suggests a correlation between greigite formation and anaerobic-microbial activity (iron and sulfate reduction).

Assuming a connection between microbial activity and greigite formation, data from the northern sites (EF and OG) suggest greater greigite formation and predominant iron reduction compared to the south (EG). High $\chi$ occurs when iron reduction is relatively high compared to sulfate reduction (thus favoring greigite relative to pyrite). This raises the question of whether the distance from the Jordan River played a role in determining the degree of microbial activity and greigite formation. The Jordan River is the major freshwater source to the Dead Sea, so salinity differences between sites might have favored microbial activity in the north rather than the south. However, pore-fluid TDS in Figure 7a, which is representative of salinity, has similar values at all three studied sections (all are hypersaline). Furthermore, salinity and conservative ion concentrations at all sites are comparable to those of pore fluids from Holocene ICDP Dead Sea core sediments (Levy et al., 2017, 2019), which suggests that the studied sites were probably closer to the hypolimnion (deep water) than to the fresher epilimnion (surface water) when the sediments were deposited.

Microbial activity might have been controlled by labile organic matter availability as shown by Häusler et al. (2014) and Thomas et al. (2016). Organic matter in the terminal Dead Sea sediments arrives from two sources: (1) allochthonous organic matter brought mainly via the Jordan River and (2) autochthonous organic material formed in the Dead Sea water-column by microorganisms, such as algal *Dunaliella* (Oren, 2010; Oren & Shilo, 1985). The amount of allochthonous organic material may decrease as a function of distance from the Jordan River. Additionally, autochthonous organic material in this hypersaline environment is related to freshwater influx via the Jordan River. Algae blooms under meromictic conditions, when a fresher surface layer forms due to sufficient freshwater input to the lake (Oren, 2010). Algal blooms were observed during the last significant formation of a fresher surface layer during the rainy winters of 1980 and 1992 (Oren, 1995; Oren & Shilo, 1982). Similar conditions to those of the winters of 1980 and 1992 prevailed occasionally in the lake during the late Holocene leading to algal blooms that would have supplied organic matter to the sediment.

To summarize, we suggest that allochthonous and/or autochthonous organic matter and/or iron supply may have been controlling factor(s) on microbial activity and subsequent greigite formation within sediments along the Holocene Dead Sea margin. All three factors are related to freshwater supply from the Jordan River. Therefore, we conclude that the $\chi$ of fresh sediment could be a useful site-specific proxy for Jordan River influx to the Paleo-Dead Sea.
Late Holocene Dead Sea sediments have χ values that fall within three ranges: high, intermediate, and low. We suggest that χ can be used to indicate different hydrological and environmental conditions, which rely on Jordan River freshwater inputs. Sediments with high χ values reflect higher freshwater inputs to the lake, intermediate χ values imply lower inputs, and lower χ values indicate almost no input. In addition, outcrop locations relative to the Jordan River inlet likely dictate χ changes because the freshwater influence from the Jordan River decreases to the south. Therefore, χ values at sites proximal to the Jordan River inlet (Og and EF) are sensitive to freshwater input changes, while χ values at EG are almost insensitive. Therefore, χ changes at northern sites (Figures 3a and 3b) reflect freshwater input changes from the Jordan River, and the χ profile provides a proxy for freshwater volume entering the lake. We correlate lake-level changes with the EF χ profile because it spans a longer time period.

The relationship between the EF χ record that is interpreted as a monitor of freshwater supply to the Dead Sea and other regional hydroclimate recorders is examined in Figure 10 in relation to Dead Sea lake levels for the past 3.5 Kyr (based on Bookman et al., 2004; Kushnir & Stein, 2019; Migowski et al., 2006). We note that high χ values correspond to lake-level rises above the sill at ∼402 m bmsl that separates the northern (deeper) and southern (shallow) Dead Sea basins (yellow line). During most of the Holocene, the southern Dead Sea basin behaved like a large evaporation space (similar to the modern situation) that buffered lake-level rise. Thus, significant freshwater inputs are required to fill the southern basin to allow it to rise above the sill (Bookman et al., 2004). Higher χ values are recorded during these periods (∼2,100–1,700 and ∼1,500–1,300 years BP; Figure 10b) of enhanced freshwater supply to the lake.

Temporal planktonic foraminiferal δ¹⁸O variations (Globigerinoides ruber) in an eastern Mediterranean deep-sea core (Schilman et al., 2001) and in a Soreq Cave speleothem (Schilman et al., 2002) are shown in Figure 10c. Foraminiferal δ¹⁸O values represent the composition, salinity, and temperature of eastern Mediterranean seawater, and mainly have a narrow range of values (δ¹⁸O = 0.2‰ ± 0.2‰), with pronounced negative peaks at 3.2 and 1.3 ka BP. These negative peaks were attributed by Schilman et al. (2001) to enhanced Nile River freshwater ingressions into the eastern Mediterranean, which reflect enhanced monsoon rains in the Ethiopian Highland Blue Nile source area. Enhanced Nile River inflows to the Mediterranean are most marked during sapropel events (e.g., sapropels S5 at ∼128–121 ka and S1 at ∼10⁶ ka [Rohling et al., 2002; Rossignol-Strick & Paterne, 1999]). The speleothem δ¹⁸O pattern resembles that of eastern Mediterranean seawater (e.g., Grant et al., 2012; Kolodny et al., 2005). Small offsets in peak locations are.
5. Conclusions

We present magnetic property and pore-fluid chemistry results for late Holocene Dead Sea lacustrine sediments. Sediments were sampled from three sections exposed on the retreating shores of the modern Dead Sea along a north-south transect: Nahal Og and EF Nature Reserve on the northwestern side of the lake and EG Spa to the south. Magnetic property variations (e.g., magnetic susceptibility, $\chi$) of more than 3 orders of magnitude mainly reflect variable sedimentary greigite contents. Greigite precipitation is controlled by iron and sulfate microbial reduction within the sediment, which varied along the N-S transect as a function of magnitude mainly reflect variable sedimentary greigite contents. Greigite precipitation is controlled by iron and sulfate microbial reduction within the sediment, which varied along the N-S transect as a function of magnitude mainly reflect variable sedimentary greigite contents. 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of distance from the Jordan River, and by the availability of labile organic matter and/or iron influx. Thus, magnetic property variations indicate freshwater influx changes to the lake. χ is extremely sensitive to freshwater availability in this depositional environment and provides a new tool to reconstruct decadal regional hydroclimate regime fluctuations. Relatively easily measured χ can, thus, be used as site-specific tracers of Dead Sea catchment hydrological conditions. The dependence of χ on the local hydrological regime provides a cautionary note for using χ for stratigraphic correlation, which should be assessed carefully when magnetic mineral assemblages are dominated by authigenic minerals such as greigite.

Data Availability Statement

Data are available at http://doi.org/10.17632/ztpkpbm43x.1

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