JGR Solid Earth

RESEARCH ARTICLE

10.1029/2019JB018963

Key Points:

- We recorded the July 2018 seismic swarm that occurred below the Sea of Galilee using a local temporary seismic network deployed around the lake
- We relocated more than 600 events and computed a new 1-D local velocity model for the area; the swarm occurred in depth range of 6–13 km
- Depth and focal mechanisms (normal and strike-slip) point toward a natural fluid-driven activity associated with the Dead Sea Fault and the Kinarot-Beit-Shean pull-apart basin

Supporting Information:

- Supporting Information S1

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Citation:

Haddad, A., Alcanie, M., Zahradník, J., Lazar, M., Antunes, V., Gasperini, L., et al. (2020). Tectonics of the Dead Sea fault driving the July 2018 seismic swarm in the Sea of Galilee (Lake Kinneret), Israel. *Journal of Geophysical Research: Solid Earth*, 125, e2019JB018963. <https://doi.org/10.1029/2019JB018963>

Received 3 NOV 2019

Accepted 21 AUG 2020

Accepted article online 24 AUG 2020

Tectonics of the Dead Sea Fault Driving the July 2018 Seismic Swarm in the Sea of Galilee (Lake Kinneret), Israel

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Abstract Northern Israel was struck during July 2018 by a M_L 4.4 earthquake followed by a seismic sequence that lasted about 30 days. This seismic sequence occurred in the center of a temporary seismic network deployed around the Sea of Galilee (Lake Kinneret). The network was installed to investigate the regional kinematics of the Dead Sea Fault, which is a major transform fault running N-S for more than 1,000 km. The data allowed us to develop a local velocity model for the Sea of Galilee. We relocated more than 600 earthquakes and calculated 27 focal mechanisms pointing out a complex kinematic setting, possibly controlled by fluids at depth. The seismic sequence developed along a NNW-striking direction and it is bounded to the east by the N-striking Dead Sea fault. Hypocenter depths range between 6 and 13 km. Directions of the principal stress tensors suggest a transtensional deformation, in agreement with the overall kinematics of the region.

We analyze and discuss our data set to investigate mechanisms that potentially triggered the observed seismic swarm, including exacerbated ground water pumping proposed by previous authors. We suggest that the seismic sequence is driven by the dissipation of the elastic load that accumulated in this region.

1. Introduction

The Dead Sea Fault (DSF) is a ~1,000 km long continental transform structure (Figure 1a) that produced at least 14 historical earthquakes with magnitudes ≥ 6.0 since 551 AD (Galli, 1999; Zililidis et al., 2015). This major plate boundary caused a moderate- to low level of seismicity during the last eight centuries (Meghraoui et al., 2003; Wetzler & Kurzon, 2016). However, large-magnitude earthquakes occurred in the region over the past 4,000 years, affecting cities and resulting in a high number of casualties (Ben-Menahem, 1991). Geophysical studies conducted along the DSF over the last decades, including an ICDP experiment, focused mostly on the Dead Sea (cf. Marco et al., 1996; Neugebauer et al., 2014), which has been interpreted as a pull-apart basin forming at a major left-lateral overstep/bending of the DSF (Garfunkel, 1981).

To the North, the DSF intersects the Sea of Galilee (SoG), or Lake Kinneret (Figure 1b), a tectonic depression marked by a complex deformation pattern (Figure 1c). This SoG segment of the DSF is characterized by low seismic activity with no major earthquakes ($M_S \geq 5.0$) detected since 1900 (Heidbach & Ben-Avraham, 2007). A complete overview of the seismic activity recorded around the SoG is proposed by Navon (2011) who analyzed a 25-year-long record of local events. Navon (2011) integrated this data set with 9 months of observations using a local network composed of four temporary stations and the Israel National Seismic Network (INSN).

The SoG was struck by a M_L 4.4 earthquake in July 2018. This was followed by a sequence that lasted around 30 days. Wetzler et al. (2019) discussed a possible anthropogenic cause for the 2018 earthquake sequence. The authors observed that from the beginning of the instrumental record in 1985 until 2013, the seismicity around the SoG was limited to sporadic events, while more recently (2013–2018) the NW part of the SoG was affected by two main seismic swarms of magnitude $M_w > 3$ along NNW-SSE strands. It should be noted

Figure 1. Geodynamic setting of the investigated region at different scales. The Dead Sea transform fault connects the Red Sea rifting system in the south and the east Anatolian Fault in the North. The investigated area is offset by NW and N striking faults. FB: Fault Belt. BKFS: Bet-Kerem Fault System. DSF: Dead Sea Fault. JF: Jordan Fault. KBSB: Kinarot-Beit-Shean Basin. MMLI and UJAP (in b) represent the BroadBand stations used for the moment tensor inversion not present on Figure 2. The yellow star marks the location of the 2018 earthquake swarm. Geological setting is modified after Garfunkel (1981), Hurwitz et al. (2002), and Matmon et al. (2010).

that the INSN in this region is sparse (<https://www.fdsn.org/networks/detail/IS/>) and velocity models are constrained at the regional scale only (Aldersons et al., 2003; Gitterman et al., 2002). The lack of an ad hoc velocity model for the SoG region makes the inversion of the focal mechanism for the M_L 4.4 2018 earthquake challenging. Wetzler et al. (2019) suggest that the seismic sequence may have been atypical for this region describing a swarm-like character. The authors pointed out that the inversion problems may have been caused by, verbatim: “a local masking of the regional stress field by a local and more dominant stress field.” The surroundings of the SoG are affected by continuous groundwater extraction, which has accelerated since 2010 and resulted in a total water level decrease of ~50 m at the time of the 2018 earthquake swarm. Taking into account these observations and the anomalously shallow hypocentral depth of 2 km, Wetzler et al. (2019) proposed that the SoG earthquake swarm was caused by exacerbate groundwater pumping.

The July 2018 earthquake sequence struck in the center of a temporary seismic network (Figure 2) that we deployed around the SoG from September 2017 to August 2018 in the framework of an international project to study local seismicity (Lazar et al., 2018). The data coverage was ideal to study the 2018 seismic sequence and provides a useful tool to obtain further details on the nature of this seismic event. The goal of this work is to understand the mechanisms that promoted the July 2018 seismic sequence using the seismological information derived from our temporary network deployed in the region.

Figure 2. Network distribution and coverage. Blue hexagons show the position of the permanent seismic network (INSN) and green triangles the temporary network deployed in July 2018. The earthquakes shown in red are the ones used for the velocity model inversion with *Velost* (see Figure 3).

2. Geological Setting

The SoG is a shallow freshwater lake with a subellipsoidal shape that developed along the DSF (Ben-Avraham et al., 1990). The N-S and E-W axes are about 21 and 12 km long, respectively. The maximum depth of the lake is presently 45 m while the water surface is located 210 m below mean sea level (Lazar et al., 2019; Sade et al., 2009). The structural and stratigraphic settings of the SoG result from a complex geological evolution, mainly driven by deformations along the DSF (Eppelbaum et al., 2007; Hurwitz et al., 2002).

The DSF is a mature continental transform fault that developed starting from the early Miocene (Matmon et al., 2003). This major plate boundary separates the Sinai microplate that is part of the African plate, from the Arabian plate (Figure 1a). The DSF shows a sinistral displacement and connects the divergent motion of the Red Sea to the south with the left-lateral east Anatolian Fault to the north and the convergent motion at the Mediterranean Ridge accretionary complex (Figure 1a). The DSF accommodated ~ 100 km of plate motion since the Miocene (Marco & Klinger, 2014) with an average slip rate of 4.9 ± 1.4 mm/yr and a locking depth of about 10–12 km (Hamiel et al., 2016; Le Beon et al., 2008). In the central part of the DSF extending south of the Palmyride Fault Belt (Figure 1a), three main pull-apart basins formed as a result of the left-stepping behavior of the transform fault (Garfunkel, 1981). These topographic depressions are bounded by strike slip and normal faults creating half-grabens and tilted block morphologies (Matmon et al., 2003). The SoG is located in the northern part of the Kinarot-Beit-Shean basin (Figure 1b) which developed between two left-stepping left-lateral strike-slip faults joining the Korazim Plateau to the north and the Kinarot valley to the south (Figure 1c). According to gravimetric measurements, the Kinarot-Beit-Shean basin is composed of two subbasins filled with Neogene to Quaternary sediments and igneous rocks units (Eppelbaum et al., 2004; Sneh & Weinberger, 2003). To the north of this basin, another prominent feature is the Bet-Kerem Fault System (BKFS) spanning between the Lower and the Upper Galilee. This system is

Figure 3. Velocity model, V_p/V_s values, b values, and error calculations. (a) Preliminary velocity models (Gitterman et al., 2002; Aldersons et al., 2003, *High-Vel* and *Low-Vel*) and the associated 150 best fit solutions obtained using *Velest*. (b and c) V_p/V_s ratio and b value computed using *GISMO* package (Thompson & Reyes, 2018) (Red squares: inverse cumulative number of events against magnitude; white triangles: number of events for each magnitude). (d) Depth and error comparison for the main events using the velocity models presented in panel (a) and computed with *HypoInverse*. (e) Table showing the average RMS and location errors associated with each velocity model.

characterized by E-W shallow normal faulting extending from the Mediterranean Sea to the SoG (Figure 1) (Matmon et al., 2010). The SoG is suggested to be affected by a higher heat flux compared to the surrounding region (Shalev et al., 2013). This may be due to the occurrence of magmatic activity during the Eocene to Miocene, as suggested by the magmatic formations of the Golan to the east of the SoG (Meiler et al., 2011). The Korazim Plateau is composed of an alternation of sedimentary and magmatic units from the same period (Heimann & Ron, 1993). No signs of magmatic activity have been recognized in the south. Several geophysical studies have been conducted around the SoG to characterize the upper crust of this region. Gravimetric studies suggest that the SoG is composed of two units (Eppelbaum et al., 2004) with the northern sector being more affected by the kinematics of the DSF. A dense-spaced grid of seismic reflection profiles has been collected in the SoG (e.g., Gasperini et al., 2020; Hurwitz et al., 2002; Lazar et al., 2019; Reshef et al., 2007; Tibor et al., 2004). Analysis of such profiles suggest the presence of two different sets of active faults oriented N-S and NNW-SSE. Geophysical and geological data are consistent with the SoG being considered as a pull-apart basin with complex extensional behavior on its northern side and a clear eastern boundary composed of a steep block displaced by the DSF (Eppelbaum et al., 2004; Hurwitz et al., 2002).

3. Methods

3.1. Seismic Network

From September 2017 to July 2018, we maintained a temporary seismic network deployed around the SoG (Figures 1 and 2). The network was composed of 12 three-component short-period Lennartz LE-3Dlite 1 Hz

sensors equipped with Omnirecs Data Cube digitizers powered by local electrical source or solar panels. The temporary network was complemented with six INSN broadband stations (for the type of sensors refer to the INSN website) (Figure 2). The network operated uninterruptedly for 10 months except a few gaps. One station did not operate throughout the experiment.

To process the data set we first applied a STA/LTA routine to the seismic sequence, which resulted in 789 detections. We manually picked 14,263 *P* and 10,862 *S* wave phases corresponding to more than 660 events. We used the *automag* routine implemented in Seisan for local magnitude (M_L) estimation (Havskov & Ottemoller, 2008). In this initial step, we used the constants of Hutton and Boore (1987). We obtained 306 microearthquakes ($M_L < 2$) and four events with $M_L > 4$. Based on the Wadati method (Wadati & Oki, 1933), a V_p/V_s ratio of 1.75 was calculated (Figure 3b). We estimated a M_c (magnitude of completeness) of 1.9 using the *GISMO* toolbox and the maximum curvature method (Thompson & Reyes, 2018). The *b* value calculated for this experiment is 0.853 (Figure 3c).

3.2. Earthquake Relocation

To improve our locations we constrained multiple 1-D velocity models using *Veles*t (Kissling et al., 1994) and based on the initial regional velocity model of Israel proposed by Aldersons et al. (2003) and Gitterman et al. (2002). First, we selected a subset of 385 events with reliable and clear pickings, a local magnitude $M_L > 2$, and low RMS and location errors. A Moho depth of 30 km was assumed according to Koulakov and Sobolev (2006). We ran *Veles*t using the input velocity models of both Aldersons et al. (2003) and Gitterman et al. (2002) as well as for two other velocity models with low and high starting values (see Figure 3a). The models converged to the final 1-D minimum model as shown in Figure 3, where we plot the 150 best-fit solutions from all inversions performed with *Veles*t. We then selected the model with the lowest uncertainties (calculated with *HypoInverse* and *Veles*t, see Figures 3d and 3e). This model, calculated for the local SoG area, was derived from the low-velocity model and presents high accuracy for our network in terms of location errors and RMS if compared with the regional velocity models of Aldersons et al. (2003) and Gitterman et al. (2002).

Meiler et al. (2011) suggested that shallow magmatic formations cropping out to the east of the SoG may cause a velocity inversion in the 1-D velocity model. They propose that velocities between 0 and 500 m depth may be higher than those between 0.5 and 2.5 km depth. However, the thickness of these shallow magmatic formations is not evenly distributed in the region (Meiler et al., 2011). For this reason, and in agreement with Wetzler et al. (2019), we decided not to include this laterally discontinuous thin and shallow layer in the proposed velocity model as it is not representative for the entire region.

To compare the obtained depth and location of our events with another method of calculation, we used the *NonLinLoc* package (Lomax et al., 2000, 2014; Lomax, 2005) that calculates probabilistic and nonlinear earthquake locations in a 3-D environment and computes the associated confidence ellipsoids. *NonLinLoc* calculates the optimal hypocenters based on a 3-D grid that we interpolated with the previously estimated velocity model. All events were processed using both *P* and *S* wave pickings. Weights were assigned to the pickings using the *PS-Picker* package (Baillard et al., 2014) based on the signal-to-noise ratio for both *P* and *S* waves. The resulting locations were calculated with the local velocity model previously constrained and the Gitterman et al. (2002) regional velocity model in order to evaluate how they affect epicenter locations (Figure S1 in the supporting information). In order to assess the associated uncertainties, we also determined the confidence ellipsoids computed with *NonLinLoc* for the largest events ($M_L > 3$) (Figure S1).

We relocated the seismic events of our catalog with *HypoDD* (Waldhauser & Ellsworth, 2000) to highlight possible fault structures. The double-difference method implemented in *HypoDD* reduces residuals between observed and theoretical traveltimes at each station for earthquake pairs (Waldhauser & Ellsworth, 2000). The conjugate gradient method (LSQR) was also applied using both waveform cross-correlation and traveltimes (Paige & Saunders, 1982; Waldhauser & Ellsworth, 2000). This method resulted in 659 relocated events among the 666 events previously located. The comparison of the epicenters before and after the relocation is shown in Figure 4.

3.3. Moment Tensor Determination

We used the *FPFit* program (Reasenber & Oppenheimer, 1985) integrated in SEISAN to assess focal mechanisms using *P* wave first motion polarities and depth computed with *HypoInverse* with the new velocity

Figure 4. Seismicity relocation. Panel (a) presents seismicity located with *HypoInverse* and the ad hoc velocity model presented in Figure 3. Panel (b) shows the relocation after *HypoDD*. Note that the depth sections are not to scale compared to the map.

model (Figure 3). We also derived azimuth and plunge for the principal stress directions of the whole cluster using the *StressInverse* package (Vavryčuk, 2014) and display the ternary diagram using Frohlich (1992) on Figure S3.

To better constrain and verify the initial focal mechanism solutions we also worked with *ISOLA*, which allowed us to estimate the moment tensor of two of the largest events (i.e., the $M_L > 4.0$ occurring on 4 July 2018 at 1:50 and 19:45) (Sokos & Zahradník, 2008, 2013; Zahradník & Sokos, 2018). We applied wave-form inversion using both short-period stations of the temporary network and the broadband stations from the INSJ.

4. Results

4.1. Seismic Sequence Distribution

The July 2018 SoG seismic sequence struck in the center of our temporary network granting us an ideal coverage (Figure 2). The earthquakes located using the velocity model constrained with *Velast* resulted in relatively low hypocentral uncertainties (i.e., average errors of latitude, longitude, and depth computed with *HypoInverse* of 1.6, 2.3, and 2.2 km, respectively) and an average RMS of 0.35 s (Figures 3 and S2). Based on this model we calculated the stations' corrections ranging from -0.2 to 0.9 s for the *P* waves and -0.28 to 1.5 s for the *S* waves for our network. We selected the model with the lowest average depth error (final minimum 1-D velocity model) compared to all other available (and calculated) velocity models (Figures 3d and 3e). The effects and the importance that the velocity model may have on the final earthquake location should be kept in mind. Our network was ideally distributed around the SoG to capture microseismic activity of the DSF in this zone. Therefore, the velocity model calculated herein is applicable only for this region. When working with stations further away, the use of a global velocity model (i.e., Gitterman et al., 2002 velocity model for the Wetzler et al., 2019 study or Crust1.0 regional velocity model Laske et al., 2013) is required.

Figure 4 shows a comparison between the location of the seismic events before and after processing with *HypoDD*. The whole seismic sequence appears to be confined between latitudes 35°33' and 35°36' and longitudes 32°49' and 32°54' and the resulting cluster is about 10 km long, 6 km wide, and 7 km thick. The cross sections indicate that *HypoDD* clustered the seismic sequence by setting a defined upper boundary at about 6 km depth. The *HypoDD* results (Figure 4b) show a narrower structure compared to the distribution of the events before relocation (Figure 4a) where earthquakes are gathered toward the north-east of the sequence into patches of seismicity. No clear earthquake lineaments denoting fault structures were observed.

Figure 5. Evolution of the seismic swarm from 4 to 25 July 2018 with HypoDD relocations. The cross sections include the event comprised within the dashed rectangles shown on the map. Events are colored based on their order of appearance. The blue star corresponds to the largest event that occurred at the very beginning of the swarm.

Figure 6. Correlation matrix and energy released. (a) We used the *GISMO* package (Thompson & Reyes, 2018) to calculate the correlation matrix of the events with M_L higher than 3 for the station ANG (see Table S1). Inputs for both the vertical and horizontal axes of the matrix are 42 events with M_L higher than 3 sorted by time. (b) Number (orange) and cumulative number (blue) of earthquakes per hour (Time in horizontal axis is not evenly distributed).

Figure 7. Fault planes of the focal mechanisms, locations and stress tensor orientations of 27 events recorded during the seismic sequence. The focal mechanisms were calculated using *FPFit* and manual polarity pickings. Earthquakes are shown with their *HypoInverse* location. The average errors in Dip, Rake, and Fit are 6°, 5°, and 12°, respectively. Overall the sequence shows a strike-slip kinematics, in agreement with the focal mechanism of the largest event. The bottom right panel shows the distribution of the principal stress axes for all the events calculated with the *StressInverse* package (Vavryčuk, 2014). The subhorizontal σ_3 suggests minor extensional component and the subvertical σ_2 lateral kinematics. Details on polarity pickings for each focal mechanism can be found on Figures S4 to S7.

HypoDD has the advantage to minimize residuals and to locate an event relative to another; however, it is also greatly affected by the velocity model and lateral variations (Waldhauser & Ellsworth, 2000). However, the depth obtained when using three different location methods (*HypoInverse*, *HypoDD*, and *NonLinLoc*) gives consistent results (Figures 4 and S1).

Little seismicity was recorded before 4 July 2018 after which a sequence of more than 650 events that lasted until the end of July 2019 followed. The relocated seismicity is distributed in the northern part of the SoG between 6 and 13 km depth with just a few events above and below these limits (Figure 4). Figure 4b shows the presence of an eastern boundary in the distribution of earthquakes, which appears poorly confined in all other directions. The largest magnitudes were observed between depths of about 10 and 12 km (Figure 4b). The INSN location are shown in the supporting information (Figures S11 and Tables S3–S5). They counted 92 events spanning from 0 to 8 km depth (<https://seis.gii.co.il/>).

From a preliminary analysis, it is not possible to recognize any spatial cluster of earthquakes while the temporal evolution shows a concentration toward S-W at the end of the sequence (yellow-orange events in Figure 5).

Figure 8. Moment tensor of the largest events of the SoG July 2018 seismic swarm calculated with *ISOLA*. Six broadband stations and local Velest model are used.

A temporal analysis of the seismic sequence (Figure 5) suggests that it is possible to qualitatively recognize two distinct families of events. This is confirmed by the correlation matrix of Figure 6.

The waveform correlation matrix in Figure 6a was computed using the *GISMO* toolbox (Thompson & Reyes, 2018) and calculates the similarity matrix of the events with $M_L > 3$. We observe similarity for two groups of waveforms around 4 July and the 8–9 July with correlations spanning from 70% to 90% within each group, the latter corresponding to the events recorded in the southeastern part of the distribution (Figure 5). The two groups have a very low correlation with each other. The sequence analysis in time is shown on Figure 6b. We notice that seismicity occurred mostly during the first five days of the sequence in a N-S direction (Figure 5), including all the major events.

4.2. Inversion of Focal Mechanisms

Polarity picking and FPFit with three allowed errors resulted in the 27 focal mechanisms presented in Figures 7 and S3. The resulting focal mechanisms show, for the vast majority, the strike slip and mixed

faulting (strike slip with normal or thrust). Waveform inversion of the 4 July 01:50 event using local short-period stations at frequencies 0.2–0.5 Hz was successful just for three stations and indicated shallow normal faulting (Figure S10). These three stations are displaying small station correction errors (~ 0.08 s).

Six regional broadband stations were well fitted with the local Velest model between 0.05 and 0.09 Hz. The inversion pointed to a shallow normal faulting, or a deeper source with a strike-slip component (Figure 8), in accordance with Wetzler et al. (2019). This result was also confirmed using a regional sample of the Crust1.0 velocity model (Laske et al., 2013), both for the 01:50 and 19:45 events of 4 July, shown in Figure S8.

Since location of the 01:50 and 19:45 events pointed toward greater depths where polarities indicate a strike-slip solution, we show in Figure S9 the broadband waveform fit in local Velest model for the source depth of 9 km. Major features of the six stations are still explained by the synthetics.

Earthquakes larger than $M_L \geq 4.0$ occurred from 4 to 8 July 2018 with the exception of the $M_L 4.1$ earthquake that occurred on 22 July (Figure 6b). The average depth is 9.11 km for the whole swarm.

5. Discussion

The 2018 SoG seismic sequence occurred in a region that has not experienced large-magnitude (i.e., $M_w > 5$) earthquakes during the last century. The DSF is a well-developed long crustal-scale continental transform fault zone. According to the empirical relations proposed by Wells and Coppersmith (1994) and the historical seismicity studied by Hamiel et al. (2009), the DSF should be capable of generating large-magnitude seismic events. The July 2018 seismic swarm lasted about 1 month and was characterized by four seismic events with M_L larger than 4. The swarm character of the seismic sequence was first suggested by Wetzler et al. (2019) and our analysis supports this suggestion. Swarms are mostly induced in areas prone to fluid movement within the crust (e.g., volcanic environments Roman & Cashman, 2006, hydrothermal systems Horton, 2012, and other active geothermal areas Maeda et al., 2010). Inguaggiato et al. (2016) proposed that the region of the SoG may be pervaded with mantle fluids upwelling through the deep-reaching DSF. The widespread existence of fluids lubricating the fault plane may explain both the moderate magnitude of the earthquakes occurring along this segment of the DSF and the swarm character of the seismic sequence described here. The positive anomaly of the heat flux shown by Shalev et al. (2013) for this region further supports a fluid-rich upper crust.

The cross section running E-W in Figure 4b shows that the seismic swarm is bounded on the east. This limit may correspond to the Eastern Marginal fault and the Jordan Fault (both included in the DSF system and described by Hurwitz et al., 2002) or the Almagor fault dipping toward the east (see Figure 9) but could also be associated with the lithological heterogeneities of this zone. The fault dynamics of the Almagor fault was proposed to be left-lateral based on seismic data (Hurwitz et al., 2002). However, reprocessing and reinterpretation suggest a normal behavior (Reznikov et al., 2004).

Considering the focal mechanisms resulting from the waveform inversion (Figures 8 and S8–S10) and polarities (Figure 7), we investigated the possibility of an overlooked physical effect causing the shallow normal fault. The waveform inversion was performed using deviatoric mode (i.e., neglecting volume changes) and shows significant variations of the double-couple percentage (DC%) with depth. Therefore, to avoid possible bias of the depth and focal mechanism caused by trade-off with a non-DC source component, we also calculated DC-constrained moment tensors (DC = 100%), assuming pure shear events. This inversion also resulted in normal faulting at shallow depth (i.e., less than 5 km). When enabling a nonzero volume (isotropic) source component, full moment tensors indicated that shallow (i.e., less than 5 km) normal faulting and deeper (i.e., less than 10 km) strike-slip faulting fit the waveforms almost equally well. The best fitting full moment tensor models featured a positive volume component (20–60%). However, the deviatoric inversion is better conditioned than the full MT inversion and presents more stable results with the velocity models used. In other words, the 20–60% volume change is not sufficiently constrained by the available waveform data, and, intentionally, is not presented here. As such, we can only speculate that greater faulting depth (suggested by hypocenter locations) was masked by discarding volume change in our source model. Possible source depth bias in deviatoric inversions has been theoretically proven by Kířová et al. (2016).

We constantly observe differences between the centroid depth (obtained with ISOLA) and the hypocenter depth observed in the seismic catalog here presented. We relocated the events with three different location methods (NonLinLoc, Hypoinverse, and HypoDD) and obtained consistent depth values regardless of the

Figure 9. 3-D representation of the seismic sequence (HypoDD relocations) and geological setting of the Sea of Galilee region. Isotherms were extracted from Shalev et al. (2013). Events are color coded based on their order of appearance. The focal mechanism corresponds with FPFit solution of the M_L 4.4 event. WMF: Western Margin Fault. AF: Almagor Fault. DSTF: Dead Sea Transform Fault. JF: Jordan Fault.

location method (pointing out a depth location of 9 km). The difference between the centroid depth and the hypocenter depth might be a result of the stations selected for moment tensor inversion (i.e., distant stations). When testing the location of the largest events ($M_L > 4$) using our local velocity model and selecting the permanent stations only (at epicentral distances of >40 km), we obtain shallower depths (3 km). However, when introducing the seismic stations from the temporary network (at epicentral distances of 4–16 km), the depth location increases. Husen et al. (2004) states that if the distance of the closest station is less than 1.5 times the focal depth, then the depth location is reliable. For this reason, and because the closest station is at a distance of 4 km from the seismic swarm, we suggest that the depth location at 9 km depth is the most likely solution. We additionally obtained a shallow centroid depth when calculating moment tensors using three of the closest stations of the temporary network. However, the calculation of focal mechanisms and moment tensor using waveform inversion techniques in a complex tectonic region is challenging when using higher frequencies. As the local stations are short period, higher frequencies had to be used. At higher frequencies, the waveform inversion is more sensitive to lateral variations of velocity and possible site effects that cannot be represented by the 1-D velocity model (Antunes et al., 2020). Besides, the velocity model used here was computed based on the P waves, but complete seismograms are used for waveform inversion. This difference can also bring some anomalous time shifts and therefore, affect the centroid depth of the solution. Finally, Antunes et al. (2020) showed that sometimes the selection of stations affect the centroid depth, especially when using higher frequencies.

Overall, while the calculation of the swarm dynamic is challenging the results of both moment tensor inversion and polarity picking point toward a mix of normal and strike-slip deformation. Therefore, we suggest a global transtensional dynamic associated with the sequence described herein.

Overall, the seismic sequence is elongated N-S and the upper boundary of the relocated events is about 6 km depth. From Figures 5 and 9 it is possible to notice two different clusters of events. The waveform correlation (Figure 6) supports this observation suggesting that seismic events may originate from two different fault systems interacting under the SoG.

The seismic sequence that started in July 2018 in the SoG invoked particular attention because of the suggested anthropogenic activity that may have caused the swarm. Wetzler et al. (2019) proposed that the 2013 and 2018 SoG seismic swarms were caused by exacerbated water pumping. This may have unloaded the hydrostatic column acting on the faults beneath the SoG promoting slip at shallow depths.

Wetzler et al. (2019) calculated that groundwater pumping (which is operated at 1 km depth and 10 to 16 km to the west of the SoG swarm) may have affected pore pressure distribution at depth with a horizontal stress of 0.1 MPa at 2 km depth causing the 2013 and 2018 earthquakes. This value is within the range of the commonly accepted triggering thresholds (e.g., Bonini, 2007; Lupi et al., 2013; Manga & Brodsky, 2006; Saar & Manga, 2003) and can promote the reactivation of geological systems in a near-critical state. Wetzler et al. (2019) argue that this horizontal stress may have caused high dilation in the impermeable lacustrine SoG sediments ultimately triggering the sequence.

However, this hypothesis requires further scrutiny as our new data challenge the anthropic triggering scenario. These data analysis reveals that the hypocenters of the seismic sequence are most likely located in the middle crust (with the Moho at 30 km depth Koulakov & Sobolev, 2006). The seismic swarm is articulated at a depth of $9 \text{ km} \pm 1 \text{ km}$ where the influence of anthropogenic activity is probably less prominent.

As mentioned previously, the accuracy of focal depth is mostly linked to the network geometry that provides a low gap between stations ($\leq 180^\circ$), the high number of *P* and *S* wave phase picking and the dedicated velocity model with station corrections (Havskov et al., 2012). For this reason, and based on the various location methods presented in the present study alongside the converging velocity model calculation, we speculate that the deeper locations (i.e., articulated around 9 km depth) are more reliable than the shallower depths (shown by location with the permanent network and waveform inversion). However, we cannot discard the possibility of a shallower seismic swarm (shown by waveform inversion) as the uncertainties created by the strong lithological and structural 3-D heterogeneity and anisotropies in the area require deeper analysis.

Finally, the perturbed stress regime first pointed out by Wetzler et al. (2019) is compatible with a fluid-rich environment, where swarm sequences may take place. We agree on the fluid-driven nature of the July 2018 seismic swarm and argue that elevated pore pressures may reactivate secondary fault planes masking the kinematics of the largest events.

6. Conclusions

Between September 2017 and July 2018 a temporary network of 12 seismometers was deployed to investigate the seismicity in the Sea of Galilee. The network was complemented with the data from the Israel National Seismic Network. During the survey period, a 30-day-long long seismic swarm took place at the center of the network and was characterized by magnitudes up to $M_L 4.4$ and an average epicentral depth of 9 km. We relocated the entire seismic sequence with an ad hoc velocity model (that we derived) that constrained the events between 6 and 13 km depth. We suggest that the July 2018 swarm is a fluid-driven seismic sequence triggered by the active structures of the Dead Sea Fault, upon which the transtensional basin of Kinarot-Beit-Shean developed (the Sea of Galilee is part of it). Our deep earthquake location results question the hypothesis that exacerbated groundwater pumping may have triggered the earthquake sequence. Further geophysical analysis are needed to investigate the mechanisms responsible for the July 2018 Sea of Galilee swarm. We point out the importance of future broadband local deployment for accurate earthquake location and focal mechanisms in this region characterized by heterogeneous deposits and complex structural patterns.

Data Availability Statement

The data recorded by the temporary stations is freely available online (at <https://doi.org/10.5281/zenodo.3524185>).

Acknowledgments

We thank Efthimios Sokos, Vladimir Plicka, Naama Sarid, Guy Lang, Victor Haddad, Christian Baillard, and all the colleagues involved in the fieldwork of this study. Permanent data of the INSN network belong to the Geological Survey of Israel. Figures 2, 4, 5, and 7 were created using the Generic Mapping Tool software (Wessel & Smith, 1991). Code for the ternary diagram written by M. Halló was used (<https://geo.mff.cuni.cz/hallo/>). J. Zahradnik was supported from the Czech Science Foundation Grant GACR-18-06716J. A. Mazzini acknowledges the support from the Research Council of Norway through its Centres of Excellence funding scheme, (project 223272-CEED) and the HOTMUD (project-288299). M. Lupi is a SCCER-SoE professor supported by KTI funding.

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