Semi-automated bathymetric spectral decomposition delineates the impact of mass wasting on the morphological evolution of the continental slope, offshore Israel

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Abstract
Understanding continental-slope morphological evolution is essential for predicting basin deposition. However, separating the imprints and chronology of different seafloor shaping processes is difficult. This study explores the utility of bathymetric spectral decomposition for separating and characterizing the variety of interleaved seafloor imprints of mass wasting, and clarifying their role in the morphological evolution of the southeastern Mediterranean Sea passive-margin slope. Bathymetric spectral decomposition, integrated with interpretation of seismic profiles, highlights the long-term shape of the slope and separates the observed mass transport elements into several genetic groups: (1) a series of ~25 km wide, now-buried slide scars and lobes; (2) slope-parallel bathymetric scarps representing shallow faults; (3) slope-perpendicular, open slope slide scars; (4) bathymetric roughness representing debris lobes; (5) slope-confined gullies. Our results provide a multi-scale view of the interplay between sediment transport, mass transport and shallow faulting in the evolution of the slope morphology. The base of the slope and focused disturbances are controlled by ~1 km deep salt retreat, and mimic the Messinian base of slope. The top of the open-slope is delimited by faults, accommodating internal collapse of the margin. The now-buried slides were slope-confined and presumably cohesive, and mostly nucleated along the upper-slope faults. Sediment accumulations, infilling the now-buried scars, generated more recent open-slope slides. These latter slides transported ~10 km³ of sediments, depositing a significant fraction (~3 m in average) of the sediments along the base of the studied slope during the past < 50 ka. South to north decrease in the volume of the open-slope slides highlight their role in counterbalancing the northwards diminishing sediment supply and helping to maintain a long-term steady-state bathymetric profile. The latest phase slope-confined gullies were presumably created by channelling of bottom currents into slide-scar depressions, possibly establishing incipient canyon headword erosion.

KEYWORDS
landslide volume, Levant Basin, mass transport complexes, morphometric analyses, semi-automated mapping, slope confined gullies, submarine slide, thin skin faulting
INTRODUCTION

Characterization of the continental slope physiography, and its changes in time and space, is important for understanding and modeling the development of depositional systems on the slope and adjacent basin, and for predicting potential reservoir settings (e.g., Prather et al., 2017; Ross et al., 2006; Ward et al., 2014; Hampton et al., 1996). The slope undergoes morphological modifications and readjustment through time by the interplay of constructional processes, causing net progradation or aggradation, and destructional processes, causing net retrogression or erosion (Brothers et al., 2013; D. C. Mosher et al., 2017; Prather et al., 2017; Ross et al., 2006; ten Brink et al., 2009). These studies usually consider the large-scale slope morphology primarily as a regional perspective, by inspecting submarine slides over substantial expanses of continental margins (Chaytor et al., 2004;を中心とする；十面人ほか，2009). These studies highlight specific, usually exceptionally large, submarine slide scars (Gee et al., 2006). While others take a regional perspective, by inspecting submarine slides over substantial expanses of continental margins (Chaytor et al., 2004;肖特別ほか，2006; ten Brink et al., 2006; ten Brink et al., 2009). These studies usually consider the large-scale slope morphology primarily as pre-conditioning factor for sliding. While studies on the morphological evolution of the margin usually take a generalized view of submarine mass transport features. This study examines in detail the mass wasting features over a broad section of the southeastern Mediterranean Sea continental margin, and their impact on the evolution of the slope morphology at large.

1.1 Submarine landslides

The term submarine slide is used here to refer to a range of mass transport features, initiated as movements of coherent masses of sediment bounded by distinct failure planes (Masson et al., 2006; Mulder et al., 1996). These include features that develop brittle as well as plastic deformation and flow (i.e., slumps and debris flows). Several indicators characterize the morphological imprints of submarine slides (Gee et al., 2006; Hampton et al., 1996): the slide translates material above a detachment from a general extensional source area at the ‘head’ of the slide; through a transitional sheared domain; to the depositional ‘toe’. Statistical analyses of submarine slides around margins of the Northern Atlantic (Hüllnerbach et al., 2004), U.S.
Atlantic margin (McAdoo et al., 2000) and Mediterranean (Camerlenghi, Urgeles, & Fantoni, 2010); show a general agreement on the distribution of submarine slide morphological parameters. The headscarps are usually located near the shelf break in canyon-based slides. In contrast, for open-slope slides they are located at the mid-slope, approximately at its maximal dip (e.g., Lastras, Canals, Urgeles, Hughes-Clarke, & Acosta, 2004). Slides range widely in the area and volumetric extent of their debris (usually reaching \( \sim 10^5 \) km\(^2\) and \( \sim 2.5 \times 10^3 \) km\(^3\), respectively) (e.g., Chaytor et al., 2009). The height of the headscarps of slides generally extends up to hundreds of meters over a width of several kilometers, while the run-out distances of the slide debris reach mostly several to tens of kilometers (e.g., McAdoo et al., 2000). Albeit, in exceptional cases submarine slides may have significantly larger dimensions (as summarized by Shanmugam (2010)). In general, submarine slides occur as a combined result of long-term preconditioning of the sedimentary pile and a transient triggering event (e.g., Masson et al., 2006). Slides may occur in a single failure, or multiple and complex failures (ten Brink, Barkan, Andrews, & Chaytor, 2009; ten Brink et al., 2014). In many cases localized repetition of slides may be associated with sediments trapping within the scars of previous slides (e.g., Frey Martinez, Cartwright, & Hall, 2005; Masson et al., 2006; Solheim, Bryn, Sejrup, Mienert, & Berg, 2005).

1.2 | Geological setting

The Israeli passive margin forms the eastern boundary of the southern Levant Basin in the southeastern Mediterranean Sea (Figure 1). It is situated between the Anatolian, African and Arabian plates and in the vicinity of the Dead Sea Fault, in a seismically active region (e.g., Salamon, Hofstetter, Garfunkel, & Ron, 1996). This margin was formed in early
Mesozoic times by the break-up of the northern edge of Gondwanaland but attained its present appearance since the Neogene (e.g., Garfunkel, 1984). Restriction of Atlantic connectivity during the Messinian Salinity Crisis resulted in the deposition of a thick evaporitic sequence within the Mediterranean basin (CIESM, 2008; Melilijson et al., 2018; Roveri et al., 2014). This layer reaches a thickness of ~2 km in the central part of the Levant Basin and pinches out upslope towards the southeastern margin, extending farther landwards within Oligo-Miocene canyons (Gardosh & Druckman, 2006; Gradmann, Hübscher, Ben-Avraham, Gajewski, & Netzeband, 2005). The top of this sequence is generally imaged as a pronounced high amplitude seismic reflection, the M reflection (Ryan, 1978). A 1–2 km thick northwestward prograding Pliocene to Recent sedimentary wedge overlies the Messinian evaporites constituting the eastern flank of the Nile littoral cell (Schattner & Lazar, 2016; Segev et al., 2006; Tibor, Ben-Avraham, Steckler, & Fligelman, 1992).

Updated Northward prevailing currents and waves dictate a net longshore transport of sediments from the Nile Delta northwards along the Israeli continental shelf and slope (Brenner, 2003; Schattner, Gurevich, Kanari, & Lazar, 2015; Schattner & Lazar, 2016; Stanley & Warne, 1998). Thus, the Pliocene-Quaternary interval is primarily composed of Nile-derived fine-grained marine siliciclastics, which include clay-rich marls, sandstones and claystone's (Buchbinder & Zilberman, 1997; Frey-Martínez, Cartwright, & James, 2006). Sediment composition varies along the Israeli continental margin in correlation with its distance from the Nile. The dominating Nile derived clastic composition is gradually and relatively depleted to the north with increasing distance from the source and dilution by local supply of skeletal carbonates (Almagor & Schilman, 1995). During the Pleistocene to Recent, terrestrial and marine conditions alternated on the shelf resulting in interlayered deposition of calcareous sandstones with various degrees of cementation, shaly to silty red sandstone, marine and continental clays, conglomerates and sand dunes (Gvirtzman, Zilberman, & Folkman, 2008).

The Holocene transgression resulted in deposition of up to a few tens of meters of silty-clay on the middle to outer continental shelf (Neev, Bakler, & Emery, 1987). Estimated late Pleistocene to present sediment accumulation rates over the centre of the Israeli slope range between ~0.15 to 1.3 m/kyr (Schilman, Bar-Matthews, Almogi-Labin, & Luz, 2001). At the base of the slope sedimentation rates on the order of ~0.2 m/kyr have been estimated (Almogi-Labin et al., 2009; Castañeda et al., 2010; Hamann et al., 2008).

The Pliocene to Quaternary sediments in the southeastern Levant Basin are deformed, as indicated by growth and blind faults, tilted beds and folds (Almagor, 1984; Almagor & Garfunkel, 1979; Baudon & Cartwright, 2008; Ben-Avraham, 1978; Garfunkel, 1984; Gvirtzman et al., 2015; Neev & Ben-Avraham, 1977). These features are believed to represent, at least in part, shallow deformation that resulted from loading and mobilization of the Messinian evaporites since the Pliocene (Almagor & Garfunkel, 1979; Gvirtzman, Reshef, Buch-Leviatan, & Ben-Avraham, 2013). The shelf and slope offshore central Israel is characterized by a series of extensional faults, generally believed to be localized above and detached into the pinch-out zones of Messinian evaporites (Almagor, 1984; Cartwright & Jackson, 2008; Gradmann et al., 2005; Mart & Ryan, 2007; Netzeband, Hübscher, & Gajewski, 2006). However, Safadi, Melilijson, and Makovsky (2017) argued that at least some of the deformation observed below the margin was detached from the Messinian evaporites. Several areas of focused evaporite-rooted deformation form rotational slump structures, most notably the Palmahim Disturbance and Dor Disturbance in the south and north of the study area, respectively (Almagor, 1984; Almagor & Garfunkel, 1979; Garfunkel, 1984).

An abundance of small to medium (~10–50 km²) Quaternary submarine slides were mapped on the modern seafloor (e.g., Almagor & Garfunkel, 1979; Almagor & Wiseman, 1977; Katz et al., 2015), and additional larger slides were mapped in the subsurface by 3D seismic imaging (e.g., Frey Martinez et al., 2005). Correspondingly, coarse-grained sediments and clasts of reworked shelf material are found in sedimentary cores from the mid-to-lower open slope in the south, while laminated turbiditic material is found in cores from mid-slope canyons (Almagor & Schilman, 1995). Hübscher, Betzler, and Reiche (2016) attributed periodic mass wasting to forestet oversteepening due to late base level fall or early base level rise. Recent and possible future activity of sliding on the Israeli slope is suggested by multiple historical accounts of locally generated tsunamis (Salamon, 2010), and the association of the seafloor morphology with proximity to seismogenic zones (e.g., Katz et al., 2015).
2 | METHODS

2.1 | Bathymetric analysis

This study is based on a composite ‘best available’ bathymetric digital elevation model (DEM) of the continental slope offshore Israel, integrating a multibeam sonar (50 m resolution) dataset (Hall et al., 2015) with bathymetric grids (12.5 m resolution) extracted from commercial 3D seismic datasets (Figure S1; Table S2) (after Gvirtzman et al., 2015). As a result, the grid sizes and orientations vary according to the datasets and are not uniform (Figures 1, S1; Table S2). Seafloor picks that were obtained from time-migrated seismic data (Table S2) were scaled to depth using a water sound velocity of 1,520 m/s (e.g., Hall et al., 2015).

Analysis of the composite bathymetric DEM was carried out in two stages. In the first stage, the main recognizable slide headscars, lateral margins and toe domains were manually interpreted to provide preliminary information about the magnitude, direction and distribution of slide scars along the investigated slope. This information was used to construct the spectral analysis scheme and provide the context and controls for the spectral decomposition results. The second analysis stage utilized 2D spectral decomposition of the DEM to its main spatial components for morphometrical investigation of the morphological imprints of mass waste and associated features along the slope. Previous examples of the use of spectral decomposition in morphometric analysis include studies of de-glaciated debris slopes (Thornes, 1972), aeolian dunes (Cazenave, Dix, Lambkin, & McNeill, 2013), wave rippled seafloors (Cazenave et al., 2013; Lefebvre & Lyons, 2011; Lyons, Fox, Hasiotis, & Pouliquen, 2002) and rhythmic river bedforms (van Dijk, Lindenbergh, & Egberts, 2008; Lisimenka & Rudowski, 2013). The lowest frequency component of the bathymetry was obtained by smoothing the bathymetric DEM with a ~12 × ~12 km filter utilizing the Garcia (2010, 2011) procedure. This component and its calculated gradient were used to study the general regional changes along the continental slope (Figure 2a). The bathymetric DEM was then detrended by subtracting this smooth regional component, tapered at its edges to avoid filtering artefacts and transformed to the 2D wave numbers (kx, ky) spectral domain (upper left corner of each of the panels in Figure 2). Spectral domain 2D tapered muting was used to divide the de-trended DEM into three spectral slices; each slice represents the inverse of a different scale range of geomorphological features: 100 to 50 m (~10^3 m), 1,500 to 400 m (~10^2 m), and 10 km to 1,500 m (~10^4 m) (Figure 2). An additional spectral slice was obtained from the ~10^3 m spectral component through an anisotropic filter in the slope strike direction, in order to extract slope parallel perturbations that reflect fault scarps and other morphological steps. A corresponding slope perpendicular ~10^3 spectral component was obtained by subtracting the slope parallel DEM from the full ~10^3 spectral component. Each spectral slice (component) was translated back to the spatial domain and was separately draped on a shaded relief map of the original bathymetric DEM (Figures 1, 2). Note that the ~10^2 to ~10^4 m bathymetric spectral components measure local (at the specified lateral scale) deviations of the bathymetric grid from the broader trend. These deviations will henceforth be termed ‘anomalies’ in accordance with signal-processing terminology. Interpretation of the anomalies observed in terms of the features they represent was carried out by their correlation with features observed in the full bathymetry and seismic sections.

2.2 | Feature digitization and statistical analysis

The ~10^3 m bathymetric spectral component anomalies were digitized to create a morphological database of the mapped slide scars along the slope (Table S2). To estimate the excavated volume of the slides we removed from the original bathymetric DEM all the areas that overlies the scar polygon and created a bathymetric DEM with data holes corresponding to all slide scar locations. This DEM was then re-interpolated to create a new DEM, in which all the slide scars are filled and smoothed to approximate the morphology of the slope before the formation of the slide scars. Slide volumes were then calculated by subtracting the original DEM from the filled and smoothed DEM. Note that these volume calculations provide minimum estimates, as no account is taken of post-failure sediment deposition within the scars. Positive anomalies of the ~10^2 m component, which represent debris blocks and morphological edges, were digitized to line-shape-objects. A ‘roughness density’ was estimated by the aerial density of these line-shape-objects (see the Supporting information for details), and used to semi-quantify the surface roughness of slide debris (e.g. as done by Micallef, Masson, Berndt, & Stow, 2007).

2.3 | Interpretation of 2D high-resolution seismic profiles

Three seismic datasets were interpreted in order to identify the submarine mass transport and deformation features: (1) TGS—2D time migrated sections of multichannel seismic reflection profiles acquired and processed by TGS-NOPEC in 2001; (2) METEOR—2D high resolution (~4 KHz) time sections acquired with a “PARASOUND” parametric sub-bottom profiler during R/V Meteor 2002 cruise M52/2 (Hübscher et al., 2003); (3) Yam Hadera—a 3D depth migrated seismic volume acquired by PGS in 2011. Note that datasets 1 and 2 were interpreted in the two-way-time
(TWT) domain, while dataset 3 was interpreted in the depth domain. Scaling of the TWT data to depth was approximated using constant seismic velocities of 1,520 m/s for the water (Hall et al., 2015), 1,600 for the shallow (<100 m below the seafloor) and 2000 m/s for the deeper (100–1000 m below the seafloor) subsurface (Berndt et al., 2012; Garziglia, Migeon, Ducassou, Loncke, & Mascele, 2008; Sagy, Gvirtzman, Reshef, & Makovsky, 2015). The different seismic datasets and bathymetric attributes were integrated and co-interpreted using Paradigm Epos software multi-surveys Project desktop.

Identification of mass transport complex (MTC) related features within seismic profiles follow the criteria established in previous studies (Bull et al., 2008; Embley, 1976, 1980; Evans, King, Kenyon, Brett, & Wallis, 1996; Frey Martinez et al., 2005; Hampton et al., 1996; Lee et al., 2002; Normark & Gutmacher, 1988; Omeru & Cartwright, 2015). By these criteria, we interpreted lens-shaped chaotic, discontinuous or profoundly disrupted seismic facies as mass transport deposits. Basal shear surfaces (BSS) were recognized as continuous unconformities below the deposit lobes, sub-parallel to the underlying sedimentary layers, which may ramp up and down several layers. The headwall domains were recognized as listric concave upward continuations of the BSS, cutting upslope younger sedimentary layers.

A more detailed description of the methodology is in the supplementary information.

3 | RESULTS

The combined bathymetric dataset reveals numerous mass transport imprints along the continental slope of Israel (Figures 1,2). The sizes and morphologies of these imprints are variable, and many of them appear morphologically complex, interleaved and discontinuous. However, the different spectral components of the bathymetry, extracted through our spectral decomposition procedure, reveal several basic underlying patterns. These, in turn, are interpreted by us below to reflect the activity of several distinct MTC systems, their controlling mechanisms and their role in shaping the continental margin. Four spectral components were identified:

1. The regional component > 12 km.
2. The ~10^4 m (1.5–1.0 km) component.
3. The ~10^3 m (400–1500 m) component.
4. The ~10^2 m (50–100 m) component.

3.1 | The regional > ~10^4 m (>12 km) spectral component

Based on the regional spectral component of bathymetry and its gradient (Figure 2a) we divide the continental slope of southern to north-central Israel into four morphological segments (Table 1). Note that slope segments 1 and 3 are two approximately aligned open slope segments, which are similar in their morphological parameters (mainly slope and width). In contrast, slope segments 2 and 4 incorporate the Palmahim and Dor Disturbances respectively, which are carved more eastwardly into the continental shelf.

3.2 | The ~10^4 m (1.5–10 km) spectral component

This spectral component of bathymetry is mainly composed of nine upper slope bathymetric depressions coupled with lower slope bulges; together each couple spans the downslope (~20 km in average) width of the continental slope (Figure 2b).

| Slope segment number (ordered from south to north) | 1 | 2 | 3 | 4 |
|-----------------------------------------------|---|---|---|---|
| Map location                                 | offshore Ashqelon—Palmahim Disturbance | Palmahim Disturbance—offshore Netanya | Palmahim Disturbance—offshore Netanya | Offshore Netanya—Offshore Dor |
| Latitudes range                              | 31°40′N–31°59′N | 31°59′N–32°10′N | 32°10′N–32°24′N | 32°24′N–32°42′N |
| Slope strike azimuth                         | 27° | 21° | 20° | 29° |
| Water depths                                 | 200–900 m | 100–1100 m | 200–1100 m | 200–1100 m |
| Slope width                                  | 17 km | 50 km | 21 km | 19–13 km (from south to north) |
| Slope area                                   | ~520 km² | ~677 km² | ~427 km² | ~630 km² |
| Slope average regional dip                   | 3.5° | 1°–2° | 3.5° | 3.5°–4.5° (from south to north) |
The northernmost depression and coupled bulge are associated with central elements of the Dor Disturbance. The top of this arcuate depression correlates with a normal growth fault, which is rooted in the Messinian layer (Figure 3). This coupled bulge is the bathymetric expression of a fault ramp structure identified by Gradmann et al. (2005). The rest of the 10^4 m component coupled depression-bulge features are denoted C1 to C8 from north to south, respectively (Figure 2b). Interpretation of seismic sections reveal that C1 through C7 lower slope bulges (Table 2; Figures 4–7), characterized by disrupted and chaotic seismic facies underlain by well-defined basal detachments, represent 30 to 40 ms-thick (~24–32 m) lenticular MTD lobes. The detachments continue upslope from the lobes and outline buried slide scars. The headwalls of these scars are buried beneath 30–50 ms (~24–40 m) of layered sediments at the edge of the current continental shelf; and are generally aligned with growth faults, rooted within Messinian evaporites or chaotic MTD units within the post-Messinian sedimentary stack (Figures 4,6,7). A thicker (up to ~160 ms/128 m) truncated divergent sigmoidal sedimentary infill, deposited within the main scars depression, is located between the headwall and lobe area of each of these MTDs (Figure 4).

All of the C1 to C7 MTD lobes pinch out beneath the current base of the slope and are nowhere observed extending downwards into the basin. Although continuous tracing of the sedimentary layering along the slope was inhibited by sparse seismic data coverage, the seismic character and correlative position of the basal surfaces of the C1 to C7 slides suggest that they track a common sedimentary horizon (Figures 4–9 and S1–S11). This common horizon appears, therefore, to have constituted a regional detachment surface. Seismic sections traversing the MTD lobes along the slope suggest that they were deposited successively from north to south: the southern part of the C1 lobe overlies the northern part of the C2 lobe (Figure 5a,b); the edges of the C5 lobe overlies the edges of the C4 and C7 lobes; and the C6 lobe overlaps a part of the C7 lobe, with its northern edge overlying also the C5 lobe (Figure 5c). Also, the C1 to C7 depositional lobes display a general decrease from south to north in their dip-directed dimensions (Figure 2b).

The southernmost C8 10^3 m scale lobe is directly associated with the seafloor expression of a slide complex, named here ‘the Goliath Slide Complex’. The chaotic seismic facies lobe associated with the Goliath is up to 60 ms (~48 m) thick (the largest of the previously described lobes), 33 km long from east to west and 10 km wide from south to east (Figures 2b, 5d, 7, 8). It originates upslope at a ~80 m high headscarp that incises the seafloor. Seismic sections crossing the Goliath Slide Complex suggest possible segmentation of its MTD by an internal reflection into two or more interleaved lobes (Figure 5d).

3.3 The ~10^3 m (400–1500 m) spectral component

Two types of elongate anomalies are delineated by the 10^3 m spectral component. These types are separated by anisotropic filtering (Figures 2c,d, 8a). Elongate bathymetric scarps and steps that extend approximately parallel to the strike of the slope are the bathymetric expressions of shallow faulting.
Generally, elongated bathymetric scars that extend in the slope dip direction represent a multitude of submarine slide scars etched into the seafloor.

### 3.3.1 Slope strike parallel bathymetric anomalies

A dense set of slope parallel bathymetric anomalies extend discontinuously along segment 4 of the slope, representing a set of pronounced bathymetric steps (Figures 2d, 8a). These < 50 m-tall bathymetric steps are aligned with growth faults, rooted predominantly in the Messinian evaporites ~1.2 km below the seafloor (Figure 10). In the northern part of segment 4, the steps are ~10 km long and are ~1 km apart. In the southern part of segment 4 the steps become more branched (only ~700 m apart) and relatively short (~5 km long). Farther south, in segment 3 of the slope, the mid-slope strike-parallel anomalies discontinue and the strike-parallel anomalies branch to two separate bands, aligned with the top and bottom of the continental slope. The upper slope band of strike-parallel anomalies extends across segments 3 and 1 of the slope and is aligned with surface projections of growth faults that are rooted in chaotic MTD units within the post-Messianian sedimentary stack (Figures 4, 6, 7, 10). The headwalls of the ~10⁴ m scale scars, C1 and C2 in segment 3 and C4 to C8 in segment 1, are aligned with the same band of faults. These faults are highlighted by the upper slope band of slope-parallel anomalies (Figure 8a). Along-slope segments 3 and 4, the lower band of strike-parallel anomalies bounds the bottom of the slope and outlines a ~5 km-wide (from east to west) belt of seafloor incised grabens (Figure 4). The grabens are delimited by a set of normal faults that are rooted in the pinch-out zone of the Messinian evaporites unit. They accommodate offsets of ~50 m at the seafloor. These grabens also outline the downslope termination of the C1 and C2 lobes. In the Palmahim Disturbance (segment 2 of the slope) the strike-parallel anomalies converge to the Disturbance's top bounding faults, and define the extensional domain (Garfunkel, Arad, & Almagor, 1979). These faults create < 50 m-high bathymetric steps, highlighting the top of the Disturbance (Figure 2d). An additional set of strike parallel anomalies marks the distal end of the Disturbance. To the south, in segment 1 of the slope, strike parallel anomalies mark the surface outcrop of salt-rooted strike-slip faults to the west of the base of the slope (e.g., Gvirtzman et al., 2015). Additional strike parallel anomalies result from north trending elongate depressions that traverse the chaotic deposit of the Goliath Slide Complex. These depressions border a graben, which incises the C8 and P8 chaotic seismic facies (Figure 7) and is represented at the seafloor by a 2 km-wide (from east to west) ~20 m-deep depression. Thus, in segment 1 some strike parallel anomalies correlate with the lower slope, but do not bound the base of the slope as in segment 3.

An additional band of low amplitude strike-parallel anomalies stretches along a major part of the outer shelf, to the east of the upper slope bathymetric steps, which highlights a set of sediment waves (Figure 2d) previously described by Schattner et al. (2015).

### 3.3.2 Dip parallel bathymetric anomalies

In total, 106 dip-parallel anomalies, representing open slope slide scars were identified and digitized within the survey area (Table S2). Overall these escarpments are distributed between water depths of 100 and 1,000 m (Figures 8, 9), but are predominantly clustered at a water depth of 400 m. In segment 3 of the slope and in the Palmahim Disturbance (segment 2 of the slope), an additional cluster of escarpments appears along the lower slope. In segment 4, headscars are distributed throughout the slope (Figure 9). The majority of the dip parallel headscars incise the scars of the C1 to C7 ~10⁴ m-scale anomalies (Figures 2, 8). These features truncate the divergent sigmoid sedimentary fill that accumulated within these scars (e.g., Figure 4). An additional cluster of dip-parallel scars incises the depositional lobes of the C1 and C2 features between water depths of 700–950 m (Figures 8, 9). Differently, in segment 4 of the slope and on top of the Palmahim Disturbance the dip parallel escarpments incise...
the multiple bathymetric fault steps (Figures 2, 3, 8). In particular, scars SL1 and SL2 (Figure 2b) are elongate chains of depressions, that start at the continental shelf edge (water depths of ~200 m) and incise the bathymetric steps down to

the base of the slope, where they converge to a wider depression (water depth 800 m).

The areal extent of slide scars in the study area ranges from a minimum of 0.47 km$^2$ to a maximum of 31.3 km$^2$.

**FIGURE 4**  (a) A 2D time migrated seismic section of the TGS 2D survey across slope segment 3 (for location see Figure 1b). Vertical exaggeration is × 10. The imaged lenticular depositional lobe of chaotic seismic facies, bounded by the green and yellow lines, correlates with the C2 10$^4$ m scale bathymetric bulge (Figure 2b). This lobe terminates within the confines of the slope; itself bounded downslope by a seafloor incising graben rooted at the Messinian pinch out (black lines). The basal surface of the lobe (yellow) is traced upslope to a buried headwall, adjacent to upper slope faults. A 10$^3$ m scale open slope slide seafloor scar incise the sigmoid deposits burying the C2 scar, and its deposited debris covers the C2 deposition lobe. (b) The same seismic profile, flattened to the seafloor horizon, emphasizes the sedimentary accumulation within the C2 scar (labeled)

**FIGURE 5**  Time migrated seismic sections of the TGS 2D survey approximately in the strike direction of the slope (for location see Figure 1b). Vertical exaggerations (VE) are noted in each section. Lenticular depositional lobes of chaotic seismic facies correlate with the location of ~10$^4$ m bulges delineating their field relations: (a) C1 (blue and orange) and C2 (green and yellow); (b) C1 (blue and orange) and C2 (green and yellow); (c) C4-C7 (pink and light blue, blue orange and green) (d) C5 (pink and blue), C8 (pink and yellow), with possible segmentation within the C8 lobe (white). The 10$^3$ m scale slide scars (black brackets) incise the 10$^4$ m scale bulges
**FIGURE 6** A time migrated seismic section of the TGS 2D survey across the center of slope segment 1 (for location see Figure 1b). Vertical exaggeration is ×10. A lenticular depositional lobe of chaotic seismic facies (bounded by the green and red horizons) correlate with the C7 104 m scale bathymetric bulge. The basal surface of the lobe (red) is traced upslope to a buried headwall, associated with normal faults (black lines). An overlying 103 m scale open slope slide scar incise the deposits accumulating in the C7 scar, and its debris covers the C7 deposit.

**FIGURE 7** A time migrated seismic section of the TGS 2D survey across the Goliath slide, in the southern part of slope segment 1 (for location see Figure 1b). The C8 chaotic depositional lobe represents a part of the Goliath slide complex (red and yellow), which overlies an older P8 depositional lobe (blue and green). The two lobes are vertically offset by a fault rooted into the Messinian interval, forming a graben (black).
Their total area affected by landslide scars along the studied ~100 km-long slope is 609 km². The removed volume calculated for a single scar ranges between a minimum of 0.001 km³ to a maximum of 0.64 km³, with a total 8.2 km³ of sediment removed by all slides together. Several attributes, representing the intensity of sliding that formed the mapped slide scars, correlate with the scar’s south-north position along the slope (Figures 8, 9). The number of the slide scars increases from south to north, reaching a maximum in region 4 (Figure 9c). In contrast, the area of each scar (Figure 9c), the scar depths (Figure 9a) and the cumulative volume of transported material (excluding the Palmahim Disturbance) decrease from south to north with a pseudo-linear trend. Three interleaved scars at the southern edge of the studied area, east of the C8 lobe, constitute the Goliath Slide Complex headscarp region (Figures 7, 8). The two southernmost of these adjacent scars extend together over an area of ~8 km² and are up to ~80 m deep, starting at a water depth of ~400 m. The third northern scar displays a ~15 km long, ~600 to 1,500 m wide and up to ~45 m deep channel-like morphology, starting at a water depth of ~650 m.

3.4 | The ~10² m (50–100 m) spectral component

The fidelity of this component is variable, changing spatially due to differences in resolution (12.5–50 m grid spacing) of the different datasets incorporated in this study (Figure 1a). This component derives two types of bathymetric information. The first type maps the sharpest gradients present on the continental slope, concordant with the headscarp, lateral margins and bathymetric steps of the mapped 10³ m scale scars. The second type maps small-scale bathymetric perturbations, representing surface roughness along the continental slope. Variations in the surface roughness delineate landslide debris lobes and lineaments of smooth morphology (e.g., Migeon et al., 2014; Rovere, Gamberti, Mercorella, & Leidi, 2013) (Figures 11–13). In segment 4 of the slope, the internal areas of the 10³ m dip-parallel scars display relatively low roughness, corresponding to smoother surfaces than the surrounding seafloor (Figure 11). In this segment of the slope, high roughness is observed in the mid to lower slope (water depths 700–1000 m) areas outside of the dip-parallel scars.
These rough surfaces also extend downslope into the basin (water depths 1000–1300 m) marking debris lobes and elongate debris deposits. In particular, a 10 km by 8 km wide fan-shaped debris lobe extends to water depths of ~1,300 m, ~16 km basinward of the bottom of the continental slope (Figure 11).

In segment 3 of the slope, the 10^3 m component delineates rough debris lobes extending downslope from mapped 10^3 m scars (Figure 11). These lobes start in the mid-slope (water depth of ~700 m) and extend into the basin, overprinting the mapped slope-edge grabens. A mid-slope scar (Esc-1 in Figure 11b) incises through one of these lobes, suggesting that it is formed by more recent sliding. In segment 1 of the slope, two large debris lobes (Lobe-1 and Lobe-2 in Figure 13) are 20 km wide (south to north) by 14 km long (east to west) and 10 km wide by 30 km long, respectively. Lobe-2 converges upslope into the Goliath slide complex scars and constitutes their depositional lobes system. Density analysis of the 10^2 m component within Lobe-1 (Figure 13b) delimits different smaller lobes that compose this large-scale lobe. Each delineated smaller lobe correlates with a dip-parallel scar mapped upslope of it, thus suggesting that Lobe-1 is an interleaved stack of several lobes. This analysis also delineates several areas varying in their roughness within Lobe-2. However, the separation of the different constituent lobes is not as clear as in Lobe-1. Lobe-1 appears to be truncated by the outline of Lobe-2, suggesting that Lobe-2 was the latest lobe deposited in this area. Moreover, the overall outline of Lobe-2 is surrounded by a smooth “halo”, presumably created by settling of suspended sediments from the sliding event at the distal-most edge of the depositional lobe. This halo overprints the southern outline of Lobe-1.

Elongate, generally east-west trending lineaments of particularly smooth seafloor are observed down the slope across segments 3 and 4 (Figure 11). These lineaments start at the edge of the continental shelf (water depth of ~200 m), diverge down the slope along the 10^3 m dip-parallel scars and seem to terminate at the bottom of the slope (water depth of

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**FIGURE 9** The distribution of the 10^3 m scale slide scars in the four slope segments (marked) along and across the continental slope. (a) A scatter plot of the headscarp water depths versus the south to north distance along the slope. (b) A histogram of the number of slide scars versus the headscarp water depth, colour-coded and numbered for each slope segment. (c) The scatter of slide scar areas (blue dots) and the cumulative number of scars (histogram) versus the south to north distance along the slope. (d) A histogram of the cumulative mobilized volume estimated from the identified slide scar, showing general south to north linear decrease. See the text for discussion.
Where these lineaments cut through the dip-parallel scars and morphologic steps, they created discontinuities that locally decrease their sharp gradients. High-resolution seismic profiles that cross these lineaments (Figure 12) reveal that they are ~50 m wide and ~10 m deep sharply incised gullies. Their bases are locally flat or V-shaped. A few additional gullies are observed in the profiles that are not mapped by our bathymetric analysis. This result is probably due to the lack of sufficient bathymetric resolution in the eastern part of the region. No similar lineaments are found in the southern part of the studied area.

DISCUSSION

The spatial components delineated by our bathymetric spectral decomposition are characterized by consistent sets of patterns, while the regional component constitutes the base level for these patterns. Based on the correlation with seismic profiles we find that the extracted patterns correspond with distinct sets of mass transport features. Moreover, we find that these sets occupy distinct stratigraphic levels, conveying the relative timing of their formation. In the following discussion, we examine the characteristics and controls of the regional base level and mass transport feature sets, and their significance for the morphological evolution of the slope.

4.1 | Controls of the continental slope morphology

4.1.1 | The broad scale morphology of the slope

The extraction of the regional (>~10^4 m) spectral component corresponds to removal of the morphological imprints of local slope sedimentary processes from the slope bathymetry. We therefore postulate that this component approximates the long-term steady-state morphology of the slope, as implied by the slope readjustment model (Ross et al., 1994). Corroboration of this hypothesis requires a comparison of the regional bathymetric component with an independent estimate of the long-term steady-state bathymetric profile. Such long-term profiles are estimated by averaging the shapes of mid-Pleistocene to present stratigraphic sequences boundaries, as imaged on seismic profiles crossing the same continental slope offshore Israel (Hübscher et al., 2016) (see the Supporting information for details). Each of the sequence boundaries represents a palaeo-bathymetric profile, and thus averaging their shapes should provide an estimate of the long-term bathymetric profile. Three of the seismic sections interpreted by Hübscher et al. (2016) cross the continental slope within our study area, in the true dip direction, and outside of the deformed Palmahim and Dor Disturbance slope segments.
The long-term averaged profiles match our regional (>~10^4 m) component profiles, extracted along these seismic sections published by Hübscher et al., (2016) (Figure 14). These matches affirm that the bathymetric spectral decomposition of the regional (>~10^4 m) component approximates the long-term steady-state profile of the slope.

Segments 1 and 3 of the continental slope, together stretching over a total distance of 60 km of the margin, are fundamentally indistinct in their regional (>~10^4 m) component morphologies, while being distinctly different from slope segments 2 and 4 (Figures 2a, 14; Table 1). Both segments, separated by the Palmahim Disturbance (segment 2), have a similar average gradient of 3.5°, matching the same long-term gradient estimated from seismic) sections (Figure 14). The gradual shelf breaks of these segments are aligned at the same (~200 m) water depth, albeit with a 7° anti-clockwise rotation of their strike orientation in accordance with the coastline. The southward thinning, implied by the smaller width of segment 1 (Figures 1a, 2a) corresponds with shoaling of the base of the slope. This base of the slope is constrained on the west by the southward widening Nile fan (e.g. Gvirtzman et al., 2015 their ‘Sinai slope turbidite channels domain’; Segev et al., 2006). These results contrast with previous studies discussing the large scale morphology of the continental margin offshore southern to central Israel (e.g., Emery & Bentor, 1960; Nir, 1984; Stanley & Warne, 1998). These latter studies pointed out the predominance of the Nile sedimentary source in shaping a gradually northward narrowing and steepening continental margin. Emery and Bentor (1960) also argued that the continental shelf edge deepens southwards in response to the load of the Nile delta and fan. However, our regional (>~10^4 m) component analysis reveals that the continental slope does not obey a northward narrowing and steepening trend, nor does the shelf edge seem to deepen southwards. The northern segment 4 of the slope indeed shows that the shelf edge shoals and the slope narrows over a distance of ~35 km. However, we argue that these changes mark the local influence of the Dor Disturbance, whereby...
the slope steepens, rotates 9° clockwise and cuts into the continental shelf towards the centre of the Disturbance.

Our analysis suggests therefore that over a spatially large (~10 km) extent coupled with the duration of sedimentary cycles (e.g. as described by Hübscher et al., 2016), and away from focused salt controlled deformation (i.e., slope regions 1 and 3), the slope in the study area preserves a general steady-state bathymetric profile. These results undermine the supposed effects of differential sediments supply as a function of distance from the Nile in shaping the long-term bathymetric profile of the slope. Alternative mechanism are needed to explain the observed morphology. The following discussion will demonstrate the utility of bathymetric spectral decomposition to delineate the interplay of internal slope collapse and surficial mass transport processes of different scales that distributed the accumulated sedimentary load across the slope and into the basin over time. These processes jointly counteract the effects of differential sediment supply and accumulation over time and regain the steady state shelf to basin bathymetric profile of the continental slope (Pyles et al., 2011; Ross et al., 1994, 1995).

4.1.2 Faulting controls the bounds of the slope

The set of elongate grabens north of the Palmahim Disturbance, as highlighted by the 10^3 m scale strike-parallel anomalies (Figure 8), were formed by a system of conjugate normal faults. These faults are rooted in the westward pinch-out of Messinian evaporites, and deform the overlying Pliocene-Quaternary sediments. Together they mark an extensional belt outlining the base of the continental slope. Garfunkel (1984), Gradmann et al. (2005), Cartwright and Jackson (2008), Katz et al. (2015) and others observed this extensional belt and related its development to the basin-ward retreat of the edge of the Messinian evaporites. Recent activity of these faults is evidenced by their pronounced bathymetric expressions, as was also argued by Katz et al. (2015). We suggest that the position of the base of the continental slope north of the Palmahim Disturbance is continuously controlled primarily by the basinward retreat of Messinian evaporites, mimicking
the base of the Messinian age slope (as also depicted by Cartwright & Jackson, 2008).

The 10^3 m strike-parallel bathymetric anomalies highlight a set of approximately coast-parallel normal faults, marking the transitional edge of the continental shelf at the top of the continental slope. In some places, these faults deform and even truncate the seafloor, indicative of recent activity. In segment 2 and the northern part of segment 4, namely the Palmahim and Dor Disturbances, respectively, the faults are rooted in the Messinian M reflection (Figures 3, 10) and are controlled by the evaporite's retreat, as already shown by Almagor and Garfunkel (1979). In contrast, in segments 1 and 3 and the southern part of segment 4 of the slope, the shelf edge faults are rooted within the upper Pliocene-Quaternary sediments, and generally, do not reach the M reflection (Figure 4). Safadi et al. (2017) illustrated the continuous post-failure deformation of buried MTDs within the area define here as segment-3, and suggested that these MTDs accommodate a long-term internal collapse of this section of the continental margin. The faults delineated along the entire length of the

![FIGURE 13](image)

**FIGURE 13** The seafloor surface roughness map of slope segment 1 derived from the 102 m (wavelengths between 50 and 100 m) spectral component. (a) The roughness elements are colour-coded between 0 (smooth) and 3 (roughest). Two debris lobe complexes are delineated with dashed black lines and labelled, while the 10^3 m scale slide scar polygons are outlined in red. (b) A colour-coded roughness-density version of the same map. Red dashed lines delineate several interleaved lobes composing Lobe 1 and Lobe 2, while thin black lines outline the 10^3 m scale slide scar polygons. A dashed black line outlines the combined eastern and southern boundaries of the higher (12 m) resolution Ashqelon and Southern Israel datasets (Table S1, Figure 1a)

![FIGURE 14](image)

**FIGURE 14** A comparison between measured bathymetry (red), and the regional (>10^4 m) spectral component (blue) along representative sections (a–e) across the slope segments defined in this study (Figure 1; Table 1). The location map (f) depicts the bathymetric gradient map (Figure 1a, displayed in gray scale) overlaid by the outlines and numbers of the slope segments (white) and the layout and labelling of the representative sections (a–e). Sections a, b, and d (marked orange in f) are Hübscher et al., (2016) sections b, c, and d (their Figures 2, 3, 4, 5), respectively, along which we estimated (see the Supporting information for details) the long-term average slope profiles (green). Note that this profile is hard to decipher because of the good fit with the regional component (blue). The long-term slope gradient estimated as the average of these three long-term profiles is marked as a reference in all profiles (grey line). The vertical exaggeration (VE as note on each section) is uniform in a, b, d and e, but is increased in c. The long-term profiles and gradient, estimated from Hübscher and et al., (2016) fit well with sections a, b, and d, representing slope segments 1 and 3, and clearly do not match sections b and c, representing the Palmahim and Dor Disturbances, segments 2 and 4, respectively. This figure demonstrates that the regional (>~10^4 m) component approximates the long-term bathymetric profile, and the morphological equivalence of segments 1 and 3
shelf edge of segments 1 and 3, expand the range of Safadi et al.’s (2017) observations and support the prevalence of their proposed internal deformation mechanism. We suggest that, away from the Disturbances, internal deformation of the continental margin controls the upper bound of the continental slope at the edge of the shelf.

4.1.3 | The Palmahim and Dor Disturbances

The locations of the Palmahim and Dor Disturbances, as mapped by Almagor and Garfunkel (1979) and revised by Garfunkel (1984), correspond to regional-scale bathymetric anomalies (Figure 2a). Both features shift the shelf edge landward, and eastward, to water depths < 100 m. Our analysis finds fundamental differences between the two disturbances. The Palmahim Disturbance is a confined feature that moderates the general gradient of the continental slope (to ~2°), and extends it westward to a width of ~50 km across a ~17 km long strip. This observation is in agreement with the Garfunkel et al. (1979) description of the Palmahim Disturbance as a rotational slide, bounded between two slope-perpendicular translational fault systems, that glided over Messinian evaporites deposited within a pre-Messinian erosive channel (Ashdod Channel).

Unlike the Palmahim Disturbance, the Dor Disturbance appears to be the focus of a regional deformation zone. It modified the continental slope at least over a width of ~35 km to the south of the Disturbance’s center, as previously mapped by Garfunkel (1984). Towards the centre of the Dor Disturbance, the general gradient of the slope gradually increases (to ~4.5°), and the slope width narrows to 13 km. The convergence of the extensional fault belts coincides with the narrowing of the slope, and replacement of the upper slope fault system with faults rooted in the Messinian unconformity. The < 50 m-tall morphologic steps that etch the continental slope in segment 4 of the slope are the bathymetric expressions of numerous growth faults that bound landward tilted blocks. In addition, shoaling of the base of the slope at the centre of the Dor Disturbance to a water depth of ~600 m coincides with the large (>30 km across) radial seafloor bulge to the west of it. Tilt of the blocks and the bulge are rooted at the Messinian unconformity, as formerly described by Garfunkel (1984) and Gradmann et al. (2005).

4.2 | Mass wasting features on the continental slope and their local controls

Three morphologically distinct types of submarine mass transport features, that occupy three distinct stratigraphic positions, were mapped based on their predominance in the 10^4, 10^3 and 10^2 m spectral components of our bathymetric analysis.

4.2.1 | Buried cohesive slope-wide slides

The ~10^4 m spectral component primarily highlights a few buried open slope slides (represented by the C1 to C7 features; Figure 2b) that imprinted a convex upper slope and concave lower slope profile. The headscars of these slides are approximately 10 km wide, cut the seaward part of underlying sigmoidal clinoforms, and construct the outer shelf and slope in the study area (e.g., Figure 4). Based on the analysis of high-resolution reflection seismic profiles perpendicular to the Israeli continental slope (Figures S13, 14) Hübscher et al. (2016) argued that foreset oversteepening facilitated mass wasting. In line with their interpretation, we find that the headscars of the ~10^4 m-scale buried slides are aligned with surface impingements of the faults at the head of the slope. This alignment suggests that faulting promoted the failure of these slides, and possibly movements on these faults were even the trigger factor. The confinement of the depositional lobes of the ~10^4 m-scale slides to the lower slope constrains run-out to relatively short distances; on the order of their slope parallel dimensions. Moreover, their depositional lobes do not extend out of the lateral bounds of their head scars. Taken together these observations imply that sliding was relatively cohesive, and possibly that the sedimentary load was relatively small. Interpretation of the seismic profiles suggests that the basal shear surfaces of the ~10^4 m-scale slides align along a common stratigraphic horizon. Lastras (2004) argued that a common basal shear surface shared by four landslides in the Eivissa Channel in the western Mediterranean Sea indicates that similar and possibly common, preconditioning factors and trigger mechanisms initiated collapse of sediments. They also suggested that gas accumulation within this basal stratigraphic layer promoted sliding. Preconditioning related to focused fluid overpressure within a specific layer (e.g., as suggested by Dugan & Flemings, 2000) and headscarf faulting, may have promoted the failure of relatively cohesive sediments, which otherwise might have been less likely to fail.

4.2.2 | Disintegrative open slope slides

Overlying and etched into the stratigraphic level of the 10^4 m spectral component slides are 106 open slope slide scars, as represented and mapped by the negative dip-parallel 10^3 m spectral component anomalies (Figures 2c, 8). Based on manual investigation of the Hall et al. (2015) multibeam bathymetry (Table S1), Katz et al. (2015) reported 447 primary, secondary and tertiary slide scars over the same study area. Their mapped scars generally correlate with the scars mapped in this study with the 10^3 m spectral component, which lumps the different scar levels into the same spectral anomalies.
While this study identified fewer slide scars than that of Katz et al. (2015), it does result in a larger distribution of scars in segments 2 and 3 of the slope. These differences suggest that the spectral decomposition methodology is more robust but lower in resolution than manual picking of the data. Marked by steep (up to 35°) gradients and up to 80 m-high relief, the headscars of the 10^3 m component slides are mostly confined to a water depth of ~400 m (Figures 8, 9). A second modality of headscars is concentrated in the lower slope, primarily from the Palmahim Disturbance and northwards through slope segments 2 to 4 (Figures 8, 9). This broader distribution observed in slope segment 3 (Figure 9), is coincident with failures both on the buried headscarp regions (the prominent peak in Figure 9b) and depositional lobes of the 10^2 m spectral component slides. The headscars of the 10^3 m component slides are notably etched into local sediment depocenters and do not seem to be directly associated with faults (Figure 4). We suggest that these slides are controlled primarily by the overload and collapse of sediments accumulated within the scars of previous slides, primarily the ~10^4 m scale scars. Downslope of these scars, the slope is characterized by several-meters-high seafloor roughness, observed as scattered anomalies in the 10^2 m spectral component. It is interpreted that this roughness element represents the lobes and debris trails of the open slope slides. The extensive spread of these debris lobes and their relatively long run out distances (up to 16 km beyond the base of the continental slope) suggests that the open slope scars and associated debris are formed by disintegnative sliding of relatively unconsolidated sediments. In slope segment 4 these open slope slides are significantly smaller (<5 km^2), and generally span the width of one bathymetric step. Also, the 10^4 m spectral component slides are missing in this region. This fact suggests that the faults that controlled bathymetric steps of the Dor Disturbance constrained the development of longer and larger slides and acted as controls for the locations of the 10^3 m slides.

Within the group of the 10^3 m disintegnative slides, the southernmost scar (i.e., the scar of the Goliath Slide Complex), stands out (Figures 7, 8, 13). It exhibits the sharpest gradients (>30°) and the most extreme bathymetric step (~80 m-high), and its debris lobe has the roughest preserved morphology observed throughout our analysis (Figure 13). Thus, the Goliath scar appears to have been less affected by post-slide smoothing processes. Such processes may include subsequent covering by sedimentation (as described by McAdoo et al., 2000; and Vanneste, Sultan, Garziglia, Forsberg, & L’Heureux, 2014); or alternatively erosion (as described by Hampton et al., 1996). We speculate that this scar is one of the most recent open-slope slide tracks within the bounds of our high quality (12.5 m grid size) data. Its depositional lobe is the largest (large enough to be detected as a 10^4 m component) of the mapped debris lobes. Density analysis of the debris roughness (Figure 13), seismic sections cutting through the debris lobe (Figures 5d, 7) and the morphology of this features headscarp (Figure 13), all suggest that this feature was created by several events. The separation of this same scar to represent several sliding events was previously suggested by Frey Martinez et al. (2005; their Figure 7) and Bull et al. (2008; their Figure 4a). Existing data does not allow the determination of whether these sliding events occurred coevally or separately.

### 4.2.3 Slope confined gullies

Between the Palmahim and Dor Disturbances (segments 3 to 4 of the slope), the continental slope is imprinted by a set of ~50 m-wide and ~10-m deep slope-confined gullies. These appear as smooth streaks on our 10^2 m spectral component (Figures 11, 12). They are observed threading through scars and deposition lobes, and diverging downslope in a manner reminiscent of subaerial fluvial systems. Along their tracks, these gullies overprint the rest of the previously described slope-shaping features creating relatively moderate gradients, where they cut through open slope scars and bathymetric steps; and relatively smooth morphology, where they pierce through debris lobes. Similar gullies were described by Migeon et al. (2014) as sediment pathways eroded by downslope bottom flows. In a number of studies, slope confined gullies were associated with cascading dense shelf water, diverted from alongshore currents (Covault, Normark, Romans, & Graham, 2007; Dalla Valle et al., 2013; Micallef & Mountjoy, 2011; Trincardi, Verdicchio, & Misereocchi, 2007). Dalla Valle et al. (2013) also suggested that topographic lows formed by landslide scars may control the location for the initiation of slope confined gullies. Porębski and Steel (2003) suggested that gullied slopes are associated with diminished sediment flux from the shelf to continental slopes, Micallef and Mountjoy (2011), Porębski and Steel (2003) and Pratson and Coakley (1996) suggested that they evolve into canyons through headward erosion. Rosenthal and Brenner (2007) established the presence of longshore currents and showed evidence that these currents evolved into downslope cascading flows. We speculate that the gullies observed on the continental slope were formed by recent erosive downslope cascading flows, related to sediment transport across the continental shelf (e.g., as discussed by Puig, Palanques, Orange, Lastras, & Canals, 2008; Wright, Friedrichs, Kim, & Scully, 2001), or along the edge of the shelf (e.g. as discussed by Schattner & Lazar, 2016), and onto the slope. These gullies may represent an incipient stage in the formation of submarine canyons on the continental slope in the presumably relatively sediment-deprived northern half of our study area.
4.3 | The evolution of slumping affecting the current bathymetry

Stratigraphic relationships and differences in spectral characteristics between the various scars and depositional lobes allow inferences to be made on the relative sequence of mass transport events on the continental slope. The generally smaller disintegrative open slope slumps (delineated mostly on the $10^3$ m spectral component) consistently appear to overlie the slope-wide cohesive slides (delineated mostly in the $10^4$ m spectral component) and cut their scars and lobes. The two groups of slides are often separated by relatively thin (mostly $< 70$ m thick) conformable sediment sections, accumulated after the failure of the larger cohesive slides. Moreover, the scars of the cohesive slides served as sediment traps, accumulating a sedimentary load that was subsequently transported by the disintegrative slides that incise the upper slope (at water depths around $\sim 400$ m, Figures 4, 8). The additional level of lower slope ($\sim 600$–$800$ m depth) slides in segment 3 of the slope are eroded into the stacked C1 and C2 depositional lobes (Figure 8). The two groups of cohesive and disintegrative slides appear therefore to have formed at distinct time-periods, separated by a considerable phase of sediment deposition. Moreover, the different mechanical behaviour between the buried ($10^4$ m component) and seafloor ($10^3$ m component) slides imply that they mobilized different sedimentary loads. As discussed above, the older slope spanning slides were presumably cohesive and mobilized relatively firm sediments in contrast with weakly consolidated sediments that are associated with the later disintegrative slides. Alternatively, the differences in the runout distances of the two generations of slides represent different sedimentary loads. In either case, the nature and rates of accumulation of the sediments on the continental slope of Israel are controlled by variations in sediment supply from the Nile (Box et al., 2011), which in turn is controlled by climatic changes (Ducassou et al., 2009; Hübscher et al., 2016). Thus, we speculate that two different sliding phases occurred in different climatic periods. Moreover, sediment trapping and accumulations related to the earlier cohesive slides appear to have served as a pre-conditioning factor for the disintegrated slide scars, as previously suggested by Solheim, et al. (2005) for the Storegga slide.

The significant differences observed in areas covered by high-resolution ($12.5$ m grid size) bathymetric data, in scar gradients and depositional lobe roughness (Figures 11, 13) imply that the bathymetric expressions of the different slides have gone through different amounts of healing. These differences may be the result of post-collapse smoothing processes, and therefore possibly associated with the failure ages. We note, however, that such high-resolution ($\sim 10^2$ m) correlation of the ‘freshness’ or ‘smoothness’ of the seafloor surface with its age is commonly misleading. For example, surveying of the 1929 Grand Banks slide area, eastern Scotian slope, showed that the latest scars are sometimes the smooth ones (Mosher & Piper, 2007), while on western Scotian slope, a rough surface tracked the relatively deeply ($\sim 30$ m) buried Barrington MTD (Mosher & Campbell, 2010). Moreover, at least in one location (Esc-1 in Figure 11), it is clear that a lower-slope slide scar overprints the debris trail of an upper-slope slide scar, giving evidence of two separate sliding events. The smooth seafloor imprint of the slope-confined gullies (Figure 11) resembles the appearance of the ‘fresh’ scars imaged within channels in the 1929 Grand Banks slide area (Mosher & Piper, 2007). These gullies overprint the bathymetric expressions of a large number of the disintegrative slide imprints, and presumably represent one of the most recent post-failure processes. The standard resolution seismic sections do not show any evidence of decipherable ($< 10$ m thick) layering above any of the disintegrative slide scars. However, some of these slides may be overlain by up to a few meters of sediments, which may partly account for the observed smoothing of their bathymetric expressions. Thus, our results do not constrain the relative ages of the slides, modifying the current seafloor. Yet, the observations are suggestive of the disintegrative sliding in the southern to central Israel offshore being an on-going process, possibly at a diminishing rate.

Taken together our results show a correlation between the spatial scales of the bathymetric imprints, delineated by our spectral decomposition results, and the relative timing of their formation. The $\sim 10^2$ m component represents the most recent imprints of the gullies. The $\sim 10^3$ and $\sim 10^4$ m components represent, respectively, older imprints of submarine slides, sediment accumulation in the scars of the earlier slides and offsets of co-active faults. The regional ($\sim 10^5$ m) component approximates the long-term steady state shape of the slope. We note that the change in the dimensions of the slides represented by the $\sim 10^3$ and $\sim 10^4$ m components may have been affected by differences in sediment and supply characteristics (as discussed above), and therefore may be incidental to some degree. Moreover, there are some overlaps between different generations of imprints within a single spectral component, such as the slide debris and gullies overlapping in the $\sim 10^2$ m component. Finally, although appealing, our suggested age relations still need to be examined by age dating of the different mass transport imprints and associated deposits. We therefore suggest that when utilized within a broader framework of geophysical and geological analysis methodologies, bathymetric spectral decomposition can yield first order information on the characteristics and evolution of mass transport on continental margins.
4.4 | Slope failure and sediment transport along and across the continental margin

Our observations highlight sediment transport and accumulation along the continental margin as a primary control for slope failure and the corresponding role of mass failure in sediment transport across the continental slope. Clear south to north diminution of the number of submarine landslides is notable in the mid-range spectral components of the bathymetry. The $10^4$ m scale depositional lobes of cohesive buried slides (C7 to C1 bulges) display a general decrease in run-out distance from south to north (Figures 2b, 8). The $10^3$ m scale open slope scars show a pronounced several-fold decrease in their scar areas and cumulative volume (Figure 9). Considering the similar bathymetric profile and sediment characteristics in slope segments 1 and 3, the observed diminution in landslide numbers and volumes along and between these segments is consistent with the argued decrease in sediment supply with northward distance from the Nile. Also, the seemingly adverse effect of a northward increase in slide population (Figure 9) is consistent with the localization of sediment accumulations as a combined result of decreased sediment supply and the Dor Disturbance related bathymetric complexity. The correlation of presumed sediment supply with mass failure, while preserving a fixed general bathymetric profile, implies that the failure is controlled by, and counteracts, the impact of differential sediment-supply and accumulation along the slope.

The observed diminution in landslide abundance is also accompanied by a south to north change in the style of mass transport. The southern parts of the slope are characterized by a few broad $10^3$ m - scale slide scars with widely spread depositional lobes (Figure 6), with their scars limited to the middle slope (between 500 to 700 m water depth; Figures 8, 9). In contrast, narrow slide scars with channelized debris lobes (Figure 8) are distributed across the entire slope in the north (Figures 8, 9). The latter frequently evolve into a downslope complex series of interconnected slide scars and debris channels, as shown by Katz et al. (2015). The $10^2$ m scale slope-confined gullies appear northward of the Palmahim Disturbance. The changes in scar morphology appear to reflect a gradual, ongoing transition from open slope sliding in the south to the canyon-incised continental slope to the north of our study area. This transition is possibly the result of reduced sediment supply towards a sediment-starved slope in the north, presumably accentuated by the local steepening of the slope and incision of the shelf by the Dor Disturbance.

Schattner and Lazar (2016) proposed that bottom currents serve as the main bypass mechanism for sediment transport from the Nile source, along the Israeli continental shelf to the basin in the north. Our results highlight the preponderance of submarine landslides in sediment transport across the margin and into the basin. The translated debris derived from the open slope slides, defined here by the $10^2$ m scale roughness component (Figures 11, 13), appears to reach ~16 km basinwards from the base of the slope. Approximated by a polygon stretching from the mid-continental slope (~600 m depth) down to ~16 km basinward from the bottom of the slope, the deposition of the $10^3$ slides is estimated to be spread over an area of approximately $\sim 2,700 \text{ km}^2$. Dividing the total computed slide scars volume of 8.17 km$^3$ (representing the total sediment volume transported by all open slope disintegrative slides) by the area of this polygon, yields an average debris thickness of ~3 m. Katz et al. (2015) constrained the disintegrative open slope slides offshore Israel to have occurred within the last 50 ka and possibly within 17 ka, depending on the estimated sediment accumulation rate on the continental slope. A similar estimate is implied by Hübscher et al. (2016), with the suggestion that mass wasting occurred predominantly when sea level was low. These estimates imply that the slide deposits have contributed a significant fraction (>17% and possibly ~50%) of the ~0.2 to 0.35 m/ka (Almogi-Labin et al., 2009; Box et al., 2011; Castañeda et al., 2010) sedimentation rate just below the base of the slope during the same time frame.

5 | CONCLUSIONS

Spatial bathymetric spectral decomposition is shown to be an efficient tool in delineating and separating the imprints of mass transport systems that shaped the present-day seafloor morphology of the southeastern Mediterranean continental slope. Integrated with interpretation of seismic sections and morphometric statistical analysis constrains the processes and sequence of events in the morphological evolution of this slope.

- We found that the regional ($>10^4$ m) spectral component represents the long-term steady-state shape of the continental slope.
- Three types of recent mass transport processes have sequentially shaped the current bathymetry of the continental slope on a sub-regional scale:
  1. Broad scale ($\sim 10^4$ m spectral component) relatively cohesive slides originated at shelf edge faults and deposited the sediments within narrowly confined lobes at the base of the slope. Although now buried, these slides created upper slope morphologic depressions coupled with lower slope confined bulges, imprinting an inverse S shape to the slope.
  2. Disintegrative open slope slides ($\sim 10^3$ m spectral component) formed through the collapse of sediments accumulated in upper slope sediment traps, formed within the scars of the earlier cohesive slides and their depositional lobes. The sediments transported by the disintegrative slides created rough texture debris lobes and
3. Slope-confined gullies (~10² m spectral component) incise the northern half of the continental slope, formed presumably by cascading gravity flows. These latest phase gullies may mark the beginning of headward erosion and the creation of new canyons.

- Both the Dor and the Palmahim disturbances are the result of faulting rooted in the Messinian evaporites pinch-out. However, there is a fundamental morphological difference between the Palmahim and Dor Disturbances. The Palmahim Disturbance is a clearly confined morphological feature. In contrast, the Dor Disturbance is a broad and diffuse feature whose fault systems cut through the continental slope and reach the outer shelf forming ~50 m high bathymetric steps. The primary mechanism of mass transport related to the Dor Disturbance is long-term continuous mobilization of faulted blocks, while catastrophic collapses probably play only a minor role in its formation.

- The base of the slope north of the Palmahim Disturbance is controlled by faults rooted at the pinch-out of the presumably retreating Messinian evaporitic unit. The top of the slope is controlled by faults that are mostly rooted within the Pliocene to Quaternary sedimentary wedge. The latter records an internal collapse of the outer shelf to slope region.

- The cumulative volume of sediments mobilized by the recent disintegrative slides is sufficient for the deposition of a ~3 m-thick layer on the lower slope and proximal basin. Thus, a significant fraction of the sediments deposited near the base of the slope since the late Pleistocene was presumably transported by these slides.

- In lieu of expected northward reduction in sediment accumulation and away from focused disturbances, a generally fixed long-term steady-state shelf-to-basin bathymetric profile of the continental margin is maintained. The clear north to south increase in submarine slide volumes appears to counteract the impact of suggested southward increasing sediment supply from the Nile. The northward diminution of both processes possibly marks a transition from open slope sliding to canyon incision.

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DATA AVAILABILITY STATEMENT

All bathymetric and commercial seismic data are archived in the State of Israel Ministry of Energy national data archives, with restricted availability by permission of the State’s Petroleum Commissioner Office. The relevant Parasound data are available upon reasonable request from Christian Hübscher, University of Hamburg, email: christian.huebscher@uni-hamburg.de.

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**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section.

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