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Elchanan Zucker, Zohar Gvirtzman, Didier Granjeon, Daniel Garcia-Castellanos, Yehouda Enzel. The accretion of the Levant continental shelf alongside the Nile Delta by immense margin-parallel sediment transport. *Marine and Petroleum Geology*, 2021, 126, pp.104876. 10.1016/j.marpetgeo.2020.104876 . hal-03150787

**HAL Id: hal-03150787**

**<https://ifp.hal.science/hal-03150787v1>**

Submitted on 24 Feb 2021

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# 1 The accretion of the Levant continental shelf alongside the Nile 2 Delta by immense margin-parallel sediment transport

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## 11 12 **ABSTRACT**

13 Following the termination of the Messinian salinity crisis ~5.3 million years ago, massive  
14 sedimentation in the Eastern Mediterranean Sea formed the huge Nile Delta. Alongside delta  
15 propagation, a continental shelf was accreted along the Levant margin. For several decades it was  
16 assumed that these two sedimentary structures were closely connected. Levant shelf deposits are  
17 composed of Nile-derived sediments and present-day measurements show that sand is currently  
18 being transported alongshore from the Nile Delta to offshore Israel. This study reexamines the  
19 existing paradigm about sediment transport and shelf-delta connection. We show that longshore  
20 sand transport is just a small part of a much larger process termed here margin-parallel sediment  
21 transport (MPST). Sand is transported in a nearshore shallow-water belt where marine currents are  
22 highly energetic. At the same time, shale is transported at greater depths over the deeper shelf and

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23 the continental slope where marine currents are weaker. To model the accretion of the Levant shelf  
24 alongside the Nile Delta we use a 3D, diffusion-based, stratigraphic modeling tool (DionisosFlow)  
25 with a new module representing MPST. Our results show that margin-parallel transport of silt and  
26 clay in the deeper waters accounts for the bulk of deposition offshore Israel and is responsible for  
27 the development of the Levant shelf. Moreover, though MPST has begun coevally with delta  
28 formation, massive accretion of the Levant shelf was delayed by 2-3 My. Initially, a continental  
29 shelf formed offshore Sinai, then offshore Israel, and most recently along the Lebanon coast. Our  
30 model also demonstrates the significant lithological differences observed between sedimentation  
31 in front of the Nile River mouth and along adjacent continental margin. High energy down slope  
32 sediment transport carries sand, silt, and clay, whereas margin-parallel currents are relatively weak  
33 and carry mainly silt and clay. One exception within the margin-parallel system is the highly  
34 energetic nearshore current that transports sand. Thus, we point out, MPST is an efficient separator  
35 between shale and sand.

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38 **INTRODUCTION**

38 The Nile Delta and its associated sedimentary cone are the largest in the Mediterranean Sea and  
39 one of the largest worldwide (Fig. 1). Five and a half million years ago, during the Messinian  
40 salinity crisis (Hsü et al., 1973; Ryan and Cita, 1978), the Mediterranean sea-level dropped and an  
41 approximately 1.5-km deep canyon (EoNile) incised into the Egyptian continental margin of the  
42 Mediterranean (Barber, 1981; Said, 1981; Fig. 2A). After this crisis, at the beginning of the  
43 Pliocene (5.3 Ma), when normal marine conditions resumed, sediments quickly filled this EoNile  
44 canyon and prograded from the canyon mouth into the open sea (Said, 1981; Sestini, 1989; Stanley  
45 and Warne, 1998). Today, after 5.3 My of progradation, the modern Nile Delta and its deep-sea

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fan and sedimentary cone spread ~200 km Northward from the ancient canyon mouth and ~400 km in an east-west direction (Fig. 1). The thickness of the Pliocene-Quaternary (PQ) sediments building the Nile Delta and cone reaches ~4000 m in its depocenter offshore Egypt (Fig. 2B). Alongside this PQ delta development, a continental shelf was accreted in the southeastern Mediterranean, offshore Sinai and Israel (termed here the Levant shelf, Fig. 1). The shelf deposits offshore Israel are composed of Nile-derived sediments. It was hypothesized that the accretions of both these features are coeval (Buchbinder and Gvirtzman, 1976; Emery et al., 1960; Gvirtzman, 1970; Nachmias, 1969; Shukri, 1950; Shukri and Philip, 1960). Using sedimentological and seismic data, numerous studies support that sediment transport is in a coast-parallel direction (Buchbinder et al., 1993; Carmel et al., 1985; Davis et al., 2012; El Din and Mahar, 1997; Emery et al., 1960; Emery and Bentor, 1960; Goldsmith and Golik, 1980; Golik et al., 1999; Golik, 2002, 1997, 1993; Klein et al., 2007; Manohar, 1981; Perlin and Kit, 1999; Schattner et al., 2015; Schattner and Lazar, 2016; Shoshany et al., 1996; Stanley, 1989; Zaghloul et al., 1982; Zviely et al., 2007, 2006). In addition, present day measurements and estimations of sediment fluxes indicate that effective longshore transport delivers large amounts of sand mainly northwards along the Nile-Levant littoral cell (e.g., Rosentraub and Brenner, 2007; Zviely et al., 2007 and references within). These observations and summaries have focused most of the research for several decades on the near-shore sand transport (Fig. 1, Black arrows). However, as new observations for the relative young age (mainly Pleistocene) of the Levant continental shelf become available (Bar, 2009; Bar et al., 2013; Ben-Zeev and Gvirtzman, 2019; Elfassi et al., 2019), we question this existing view. If the Levant shelf coevolved alongside the Nile cone, why did its accretion start much later? The answer to this question presents major implications for understanding sediment transport along continental margins, in general, and

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69 particularly to the sedimentary source of the Levant shelf. Since current sediments moving over  
70 the Levant shelf do derive from the Nile Delta, what are the main controlling processes that  
71 transport sediments along the continental margin and what types of sediments are transported by  
72 these processes?

73 In what follows we develop, first the concept, and then the quantification of margin-parallel  
74 sediment transport (MPST) not limited to sand grain size and not limited to the nearshore belt.  
75 Earlier evidence regarding a mid-shelf mud-dominated belt shaped by oceanographic processes,  
76 were documented in the Gulf of Lions at the mouth of the Rhone River (Rabineau et al., 2005; and  
77 references therein) and suggested for the Levant margins by Schattner et al. (2015). Here, we  
78 further develop this concept exploring the hypothesis that deeper and distal currents first conveyed  
79 large amounts of clays and silts along the ancient continental margin and only then, after generating  
80 this clay and silt foundation, the shelf was established on top with sand transport, as we see it  
81 today. I.e., we hypothesize that only after filling the quite large accommodation space with the  
82 finer grains and generating the shallow water environment, sand could have started to accumulate  
83 as we observe today on the present-day shelf.

84 Here, we test this hypothesis using the stratigraphic simulation software DionisosFlow (Granjeon,  
85 1997, 2014; Granjeon and Joseph, 1999). DionisosFlow is a state-of-the-art, 3-D stratigraphic  
86 modeling tool that produces synthetic stratigraphy according to prescribed subsidence history, sea-  
87 level variations, sediment supply, water discharge, and sediment distribution by diffusion. The  
88 problem is that to date, DionisosFlow has no module to simulate margin-parallel transport at  
89 shallow (alongshore) or deep (off shelf) waters. Therefore, we added to it a new module that may  
90 answer some of the above questions. The new module includes a simple advection law in which  
91 sediment flux is defined by the user, and deposition rate is calculated from this flux accounting for

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4 92 mass conservation. This volumetric exercise is constrained by ages and thicknesses of units. The  
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6 93 sediment flux maps obtained by trial and error, indicate how much sediments pass through each  
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9 94 point in the basin at various modelled time steps.

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11 95 To account for lithological variations across the margin we also develop a depth-dependent  
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13 96 function for the sand/shale ratio for the deposited sediments and the resulting stratigraphic section  
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16 97 is calibrated by well data. Our results in terms of sediment flux and sand/shale ratio can be the  
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19 98 basis for future physically based simulations that can better examine how surficial, intermediate,  
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21 99 and deep marine currents transport sand and shale along various prescribed morphologies of  
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23 100 continental margins.

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## 27 28 102 **GEOLOGICAL BACKGROUND**

### 29 103 Sediment transport along the Nile littoral cell

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33 104 The Nile littoral cell is one of the longest in the world with shorelines extending along ~700 km  
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36 105 from the Nile Delta to northern Israel (Inman and Brush, 1973; Inman and Jenkins, 1984). The  
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38 106 source of this littoral cell is in the Nile River, it passes along the shallow (up to 30 m deep)  
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41 107 nearshore water belt of the Levant, with current northernmost terminal deposition at the Haifa Bay  
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43 108 (Buchbinder et al., 1993; Carmel et al., 1985; and the other references quoted in the above  
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46 109 introduction). Some sediments that are transported northward turn west, fall off the shelf and then  
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48 110 deposited along the continental slope of the Levant (Ben-Gai et al., 2005; Schattner et al., 2015;  
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51 111 Schattner and Lazar, 2016; see data about Units 3-4 below). In addition to the nearshore transport,  
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53 112 Woodward et al. (2007) showed that present-day Nile sediments extend across the Levant and  
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55 113 Herodotus basins as far as Rhodes, Crete and north of Cyprus.

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114 Longshore transport usually refers to the net movement of coastal and nearshore sand parallel to  
115 the shore (Seymour, 2005). This transport is either by currents carrying sand in suspension or by  
116 waves on the seabed (i.e., bedload transport). It is easier to measure concentration of suspended  
117 sediment, but quite difficult to measure the bedload component. Quantitative estimates of  
118 longshore transport, therefore, are based commonly on inferences from the suspended sediment  
119 concentrations indicating that longshore transport is at a maximum under the breaking waves and  
120 in the swash zone on the beach face (Seymour, 2005).

121 The annual longshore flux of sand from the Nile Delta (Fig. 1) eastward is estimated at about  $10^6$   
122  $\text{m}^3 \text{yr}^{-1}$  (Inman et al., 1993, 1976; Inman and Jenkins, 1984). Offshore Sinai, this flux slightly  
123 decreases to  $\sim 8.5 \times 10^5 \text{m}^3 \text{yr}^{-1}$  and  $\sim 7 \times 10^5 \text{m}^3 \text{yr}^{-1}$  off Port Said and off El Arish, respectively  
124 (Inman et al., 1976). This flux is estimated at  $4.5 \times 10^5 \text{m}^3 \text{yr}^{-1}$  and  $\sim 10^5 \text{m}^3 \text{yr}^{-1}$  at Ashkelon and  
125 Herzlia (Fig. 1), respectively (Perlin and Kit, 1999). Only a small portion ( $\sim 8.5 \times 10^4 \text{m}^3 \text{yr}^{-1}$ ) of the  
126 initial sand flux reaches the currently considered final sink at Haifa Bay (Zviely et al., 2007).

### 127 Different Hypotheses

128 An important point was raised by Schattner et al. (2015) who argued that longshore transport is  
129 limited to the shallow (<30 m depth) and narrow (350-400 m) near-shore belt and cannot explain  
130 the deposition at greater depths. In addition, we stress that longshore transport only moves sand  
131 and silt, whereas the PQ shelf is mainly composed of shales (Yafo Formation) with only its topmost  
132 part composed of sand (Kurkar Group, e.g., Gvirtzman and Buchbinder, 1978).

133 Based on bathymetric, seismic, and oceanographic data, Schattner et al. (2015) suggest that since  
134 the Pleistocene-Holocene transition ( $\sim 11 \text{ka}$ ) fast water currents, termed the Levant Jet System  
135 (LJS; Fig. 1, red arrows), mobilizes sediments along the continental margin not only nearshore,  
136 but also in intermediate and deep waters (between 0 and 400 m water depths).

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137 Timing of the Levant shelf accretion

138 A general PQ age was assigned for the continental shelf offshore Israel (e.g., Ben-Avraham, 1978;  
139 Ben-Gai et al., 2005; Gvirtzman and Buchbinder, 1978; Tibor et al., 1992). However, recent  
140 reexaminations of the biostratigraphy led Bar et al. (2009) to propose that most of the shelf was  
141 accreted later and that is composed primarily of Pleistocene sediments. This relatively young age  
142 is supported by Elfassi et al. (2019) and Ben-Zeev and Gvirtzman (2020) who demonstrate that  
143 after the Pliocene the pattern of sedimentation in the Levant Basin changed drastically; from  
144 thicker accumulation in the deep basin to thicker accumulation along the continental margin.  
145 Particularly interesting and crucial is the observation that the preferred deposition along the margin  
146 was delayed by 2-3 My after the commencement of the Nile Delta formation offshore Egypt (Ben  
147 Zeev and Gvirtzman, 2020). This, once more, raises the question why did margin-parallel sediment  
148 transport and shelf accretion begun so late? Why were sediments not transported along the Sinai-  
149 Israel margin already during their initial supply in the Pliocene?  
150 Hypothetically, this can be answered by proposing that external forcing such as climate or ocean  
151 circulation could have triggered margin-parallel marine currents only in the Pleistocene.  
152 Alternatively, we hypothesize here, sediment transport along continental margins is a strong depth-  
153 dependent process that becomes more effective under certain bathymetry conditions. In the case  
154 of the Nile, we hypothesize that only after 2-3 My of delta accretion, the continental margin  
155 offshore Sinai could have become an effective conduit for sediment transport and large amounts  
156 of sediments started arriving to offshore Israel.

157 **METHODS**

158 **Diffusion modeling in sediment transport**



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159 Since the early work of Culling (1960) and Young et al. (1973) and with the lack of specific  
160 information on the actual terms in physical sedimentology processes, diffusion laws have been  
161 used in geomorphology and geology to represent large-scale, spatially-averaged transport of  
162 sediment by creep, overland flow, and channel flow (e.g., Begin et al., 1981; Paola et al., 1992).  
163 Diffusive sediment transport states that the transport capacity of a water flow is proportional to the  
164 local basin slope and water discharge. It has been used in modeling sediment transport along  
165 alluvial fans, rivers and floodplains (Begin et al., 1981; Ben Moshe et al., 2008; Coulthard, 1999;  
166 Murray and Paola, 1994; Parker et al., 1998), mountains, foreland basins (Jordan and Flemings,  
167 1991; Tucker and Slingerland, 1994) and passive margins (Rabineau et al., 2005).

168 Since the 1980s, when seabed morphology was shown to resemble lowland landscapes (e.g.,  
169 Adams et al., 1998; Galloway, 1998; Mitchell and Huthnance, 2007; Schlager and Adams, 2001),  
170 laws of landscape evolution were extended to simulating progradation of deltas (Kenyon and  
171 Turcotte, 1985) and continental shelves (Granjeon, 1997; e.g., Jordan and Flemings, 1991;  
172 Kaufman et al., 1991; Mitchell and Huthnance, 2008; Rivenæs, 1997).

173 DionisosFlow is a 3D stratigraphic forward modeling tool aiming to simulate geometry and facies  
174 of sedimentary units on a scale of several tens to hundreds of kilometers and over geological  
175 intervals of  $10^3$ - $10^8$  years (Granjeon, 2009; Granjeon and Joseph, 1999; Granjeon and Wolf, 2007);  
176 these are the scales of the problem at hand. At these spatial and temporal scales and the needed  
177 resolution, unfortunately, there is practically no way to apply physically based modeling. Using a  
178 diffusive approach for sediment transport (Gerber et al., 2009; Kaufman et al., 1991; Lai and  
179 Capart, 2007; Mitchell and Huthnance, 2008; Schlager and Adams, 2001; Spinewine et al., 2011;  
180 Steckler et al., 1999; Van Heijst et al., 2001), DionisosFlow averages the cumulative mass  
181 movement in time steps that combine many individual flows. In this way, DionisosFlow, for

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182 example, ignores individual turbidites, but considers their cumulative mass movement. Similarly,  
183 and important for our study, it does not account for the discontinuity between fluvial and submarine  
184 systems in cases that sediments temporarily settle down on the delta front or inner shelf and only  
185 later are transported downslope by gravity-driven processes. For simplicity, it assumes that flows  
186 continue seaward either directly by hyperpycnal flow or by slope failures.

187 DionisosFlow formulates two large-scale sedimentary processes: slow creep and fast transport  
188 driven by water discharge (Tucker and Slingerland, 1994; Willgoose et al., 1991). Creep is  
189 simulated by a linear, slope-driven diffusion equation (i.e., transport is proportional to gradient).  
190 Water-driven flow is simulated by a non-linear water- and slope-driven diffusion equation.  
191 Combining the driving equations of these transport laws (Granjeon, 2014; Granjeon and Joseph,  
192 1999) leads to the sediment transport equation:

$$(1) \quad Q_S = -(K_S + K_W Q_W^n S^{m-1}) \vec{\nabla} h$$

194 Where  $Q_S$  is sediment flux [ $\text{km}^2/\text{yr}$ ];  $h$  [m] is topographic elevation;  $K_S$  and  $K_W$  [ $\text{km}^2/\text{yr}$ ] are the  
195 diffusion coefficient, respectively for slow creep transport and water flow process;  $Q_W$  is  
196 dimensionless local water discharge at the cell;  $S$  is local gradient of basin slope;  $n$  and  $m$  are  
197 constants, usually ranging between 1 and 2. Transport equation and mass conservation are applied  
198 to each grain size, leading to definition of local sedimentation or erosion. Sedimentation occurs  
199 where transport capacity decreases (by decreasing gradient or slower water flow) and erosion  
200 occurs where transport capacity increases.

201 The capability of DionisosFlow to simulate the production of large-scale systems such as deltas,  
202 fluvial plains, shelf plains, incised valleys, and deep water systems was tested and validated in  
203 several studies including, for example: (1) the XES 02 flume experiments designed to reproduce

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204 fluvial plain and incised valleys (Granjeon, 2014); (2) Synthetic case studies built to investigate  
205 the impact of millennial-scale climatic and eustatic cycles on the growth of a delta and the  
206 associated fluvial channel belts and incised valleys (Csato et al., 2014; Rabineau et al., 2006,  
207 2005); (3) On modern delta, shelf, and deep-water turbiditic systems such as the Niger Delta  
208 (Laurent et al., 2020) and the Orinoco deep-sea fans (Deville et al., 2015).

209 **Simulating margin-parallel sediment transport**

210 The margin-parallel transport module developed here begins with a sediment flux map describing  
211 the intensity [km<sup>2</sup>/My] of the flux transported through a given location in a coast-parallel,  
212 dynamically updated direction, which evolves with time. The flux obeys a simple user-defined  
213 advection law that accounts for mass conservation, i.e., when the flux decreases, sediments depart  
214 from the suspension plume and deposit. The rate of deposition, SED [km/My] is:

$$(2) \quad \begin{cases} \vec{q}_{mp} = Q_{mp}(x, y) \cdot \vec{n} \\ SED(x, y) = -div(\vec{q}_{mp}) \end{cases}$$

216 With  $q_{mp}$  the margin-parallel sediment flux [km<sup>2</sup>/My],  $Q_{mp}$  the user-defined intensity of this flux,  
217  $n$  a unit vector parallel to the closest shoreline [dimensionless], and SED [km/My] the  
218 sedimentation rate induced by this margin-parallel sediment transport.

219 The sand/shale ratio within the deposited sediment is computed from a given depth-dependent  
220 function, derived from observations. In addition to this deposition rule, we use the classical  
221 diffusion equations to transport sediments as soon as they reach the sea floor. By the end of this  
222 computation, we obtain a stratigraphic layer whose thickness is compared to observations. By trial-  
223 and-error (i.e., sensitivity over a range), the flux map is corrected, and simulations rerun until they  
224 satisfy borehole and geophysical observations.

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226 **Data**

227 Basin structure

228 The stratigraphic PQ sequence simulated here is bounded by the present-day bathymetry at its top  
229 and the base Pliocene surface at its bottom. Bathymetry (Fig. 1) is taken from ETOPO1 (Amante  
230 and Eakins, 2009). The base Pliocene structure map (Fig. 2A) is after Zucker et al., (2019). The  
231 PQ isopach (Fig. 2B), which is the subtraction of Fig. 2A from Fig. 1, defines a major depocenter  
232 under the Nile Delta and cone (reaching a thickness of 4 km) and a secondary and connected, local  
233 depocenter under the Israeli coastal plain and shelf (reaching a thickness of 2 km). Resolving these  
234 two depocenters - a delta and an adjacent continental shelf - are the targets of this simulation.

235 Paleo-bathymetry

236 For the paleo-bathymetry at 5.3 Ma (Fig. 3A), which is the surface on which we begin to model  
237 deposition, we used the base Pliocene map corrected for the load of the overlying sediments (PQ  
238 section). For this flexural backstripping procedure, we assumed an average density of  $2000 \text{ kg/m}^3$   
239 for unloaded sediments, adopting a pure elastic thin plate approach to isostasy and an elastic  
240 thickness of 30 km, a relatively large value consistent with the old tectonothermal age of the  
241 northeast African margin. For the decompaction of the underlying sediments, we considered an  
242 approximately 10-km thick pre-Pliocene section. The decompaction results are very insensitive to  
243 the depth of the base of this layer because only the upper ~2 km retains a significant porosity.  
244 Details about this pseudo-3D (planform) flexural backstripping procedure are described in Garcia-  
245 Castellanos et al. (2003) and in Amadori et al. (2018). To correct for post-Miocene tectonics, the  
246 sediment-unloaded map was smoothed in the area of the Rosetta and Tamsah faults (Fig. 2A),

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247 which were active in the PQ (Aal et al., 2000). In addition, the northern Levant Basin that subsided  
248 towards the Cyprus Arc during the Plio-Quaternary (Ben-Avraham, 1978; Kempler, 1998;  
249 Robertson, 1998; Schattner, 2010) was uplifted by ~200 m. The uncertainty of post Miocene  
250 tectonics is discussed below.

### 251 Subsidence

252 Subsidence (Fig. 3B) is calculated in the model by taking the difference between the  
253 paleobathymetry at 5.3 Ma (Fig. 3A) and the present-day base Pliocene surface (Fig. 2A), the  
254 surface which should be obtained at the end of simulation. Average subsidence rate for the entire  
255 simulation is taken as the total calculated subsidence divided by 5.3 My. This assures that at the  
256 end of the simulation, the modeled and observed base Pliocene surfaces will coincide to allow  
257 comparing the modeled and observed bathymetries (Fig. 3C).

### 258 Sediment supply

259 A first order approximation for sediment supply is derived from the total volume of the PQ isopach  
260 (Fig. 2B) divided by 5 million years (i.e., 70,330 km<sup>3</sup>/My). From this number we subtract biogenic  
261 carbonates produced in the water column. Well samples and logs from the Israeli coastal plain  
262 indicate that the Plio-Quaternary Yafo Formation contains ~10% calcite in the form of micro- and  
263 nano-fossils (Gvirtzman, 1970). Considering that 10% of the total PQ volume did not arrive from  
264 the clastic feeders, we initially applied 90% of the total deposited volume. After a set of trials  
265 constrained by the sediment thickness maps, we applied an average value of 64,417 km<sup>3</sup>/My  
266 (~91.5%) to represent sediment supply by the Nile River for the entire duration of the model. It is  
267 interesting to note that this average value obtained from sediment volume is on the same order of  
268 magnitude as the sediment yield presently measured at the mouth of the Nile River (60,000

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269 km<sup>3</sup>/My, Syvitski and Milliman, 2007). Sand/shale ratio for Nile sediments was taken as 20% and  
270 carbonate production was taken as 5,910 km<sup>3</sup>/My.

271 Margin-parallel transport

272 Regarding present-day measurements of longshore transport (references above) it should be noted  
273 that they refer only to the sand fraction (1000 km<sup>3</sup>/My from Nile Delta to Sinai; 850 km<sup>3</sup>/My at  
274 Port Said, 700 km<sup>3</sup>/My at El Arish; 450 km<sup>3</sup>/My at Ashkelon; 100 km<sup>3</sup>/My at Herzlia; and 85  
275 km<sup>3</sup>/My arrive to Haifa Bay). The big unknown in our simulation is the total amount of sediments  
276 (sand to clay) moving along the continental margin nearshore, on and off the shelf.

277 A lower limit for the total sediment flux along the Levant continental margin is obtained from the  
278 average deposition rate along the Israeli margin since the early Pliocene. Considering a rectangular  
279 200 km (Rafah to Southern Lebanon) by 20 km (approximately the mean width of shelf) by 2 km  
280 (thickness of PQ section offshore Tel Aviv), the total volume of the Israeli shelf is about 8,000  
281 km<sup>3</sup>. Dividing this by 5.3 million years, yields ~1500 km<sup>3</sup>/My. This set a minimal constraint for  
282 the flux that entered the Israeli margin from offshore Sinai, because the actual flux includes  
283 sediments that continued northward to offshore Lebanon. No siliclastic sediments can enter the  
284 shelf from the onshore carbonate platform of the Levant. It is interesting to note that this estimated  
285 minimum value of total sediment flux triples the sand flux (~500 km<sup>3</sup>/My) currently entering to  
286 Israeli margins at Rafah (the southern boundary of the Gaza strip between El Arish and Ashkelon).  
287 Below, the unknown flux of sediments along the Levant continental margin is modeled by trial-  
288 and-error.

289 Water discharge

290 Present day water discharge of the Nile River is 2830 m<sup>3</sup>/s (Milliman and Syvitski, 1992). The  
291 paleo-discharge of the ancient Nile River can be further, but only roughly, constrained by

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292 paleogeographic (drainage area) and paleoclimatic (rain) considerations and compared with  
293 modern rivers worldwide that have similar drainage areas (Dai and Trenberth, 2002; Milliman and  
294 Syvitski, 1992; Syvitski and Milliman, 2007). Although we are aware of the impact of Quaternary  
295 climatic changes on the Nile discharge, for the lack of data and for the sake of simplicity, we took  
296 the present-day value as a constant throughout the entire duration of the model.

297  
Diffusion coefficients

299 Efficiency of sediment transport depends on difficult-to-quantify factors such as grain size,  
300 lithology, roundness, water depth, water discharge, vegetation, river type and more. In practice,  
301 coefficients used to model various basins do not always express the same factors and their values  
302 vary greatly from case to case. One reason for this large range is that efficiency of diffusion can  
303 be controlled either by the diffusion coefficient or by the amount of water pushing the sediments  
304 (Equation 1). Accordingly, Gvirtzman et al. (2014) and Hawie et al. (2015) showed that by  
305 constraining water discharge, the range of possible diffusion coefficients is narrowed. Recent  
306 stratigraphic forward models run at a basin scale, using diffusion laws to simulate slow creep and  
307 water flow processes, and calibrated on seismic and well log data suggested the following range  
308 of values (Table 1):  $K_s$  (slow creep) is about 0.001 to 0.05 km<sup>2</sup>/ky for sand, and 0.001 and 0.1  
309 km<sup>2</sup>/ky for shale; while  $K_w$  (water flow process) is about 50 to 2,000 km<sup>2</sup>/ky for sand in terrestrial  
310 environments, 100 to 4,000 km<sup>2</sup>/ky for shale in terrestrial environments, 0.01 to 20 km<sup>2</sup>/ky for  
311 sand in marine environments, and 0.063 to 100 km<sup>2</sup>/ky for shale in marine environments (Csato et  
312 al., 2014, 2013; Gvirtzman et al., 2014; Harris et al., 2016; Hawie et al., 2018, 2017; Zhang et al.,

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313 2019). Here we take the present-day water discharge of the Nile as a starting point and identify the  
314 best-fit diffusion coefficients of shale and sand inside these ranges by trial and error.

315 Sand/shale ratio

316 Present-day measurements of sediment composition at the seabed offshore Israel (Almogi-Labin  
317 et al., 2012) indicate the presence of three coast-parallel belts (Fig. 4A): (a) An almost pure, well-  
318 sorted sand belt from the shoreline to a water depth of 35 m. (b) A silt-dominated belt from water  
319 depths of 35-100 m in which the amount of silt gradually grows from a few percent at a depth of  
320 35 m to 85% at depths of 60-100 m. The rest is shale; this shale represents only a few percent of  
321 the total sediment volume near the coast and 10-15% at depths of 60-100 m. (c) A shale dominated  
322 belt at depths >100 m, where sand is negligible. For modeling we simplified these observations to  
323 only two components (sand and shale) with a depth-dependent ratio shown in Fig. 4B.

324  
325 **Modeling strategy**

326 The first objective of this modeling study is to obtain the diffusion parameters that explain the Nile  
327 Delta formation with the given values of water discharge and sediment supply. For this, we first  
328 use DionisosFlow without the added new module of MPST. Considering that the base of the  
329 simulated PQ package is forced to coincide with the observed base Pliocene surface (see above  
330 constraints for subsidence), testing the model is carried out by comparing the modeled and  
331 observed bathymetries. If bathymetry fits, sediment thickness also fits. This stage of the modeling  
332 is expected to provide a rough fit between the model and observations at the delta area, but not at  
333 the Levant margin.



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334 The second objective is to simulate the formation of the Levant shelf. For this, we apply all model  
335 parameters realized in the first stage (and particularly the diffusion coefficients) and rerun  
336 DionisosFlow; this time the runs include the upgraded module for margin-parallel transport. By  
337 numerous trial-and-error runs, we identify the sediment flux maps that yield the better fit between  
338 model and observation. For calibration of the lithostratigraphy, we use a published  
339 lithostratigraphic section that distinguishes the sandy Kurkar Group from the shales of the Yafo  
340 Formation (data from Gvirtzman and Buchbinder, 1978). For chronostratigraphic calibration, we  
341 use the subdivision of the PQ section offshore Israel identified for four dated seismic units (Ben-  
342 Zeev and Gvirtzman, 2019; Elfassi et al., 2019).

**RESULTS**

**Stage 1: basic model without the margin-parallel sediment transport (MPST)**

346 Fig. 5 presents the simulated bathymetry from 5.3 Ma to the present, using the basic model without  
347 the MPST. The parameters for sediment distribution that generate this model are listed in Table 1.  
348 These results demonstrate that the model successfully produces the Nile Delta, but essentially fails  
349 to produce the Levant shelf. The general fit between model and present-day observations in  
350 bathymetry and thickness (Fig. 6A) is evident in section AA' crossing the Nile Delta. Moreover,  
351 comparison to a published geological section in that area from Kellner et al. (2009) indicates that  
352 even the subdivision between Pliocene and Quaternary is modeled quite well (Fig. 6C). On the  
353 other hand, offshore Israel there is only a poor fit between model and observations (Fig. 7A, B).  
354 This discrepancy in the east-west cross sections is primarily because the model produces by far  
355 too little deposition offshore Israel (Fig. 8A). The map presenting the discrepancy in thickness  
356 between model and observations (Fig. 8B) further indicates that the sediment, unaccounted for at

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357 the Levant margin (~2 km, blue colors), is compensated by excess sediments in the Herodotus  
358 Basin (200-900 m, red); i.e., the sediments were delivered there. This discrepancy strengthens our  
359 hypothesis that without a large margin-parallel sediment supply, it is impossible to simulate the  
360 formation of the Levant continental shelf.

361 Except for this discrepancy, which is related to the lack of margin-parallel transport, the generally  
362 good agreement between model and observations offshore the Nile Delta (e.g., Fig. 6) indicates  
363 that (1) sediment distribution in the model functions quite well, and (2) average values of  $Q_w$ ,  $Q_s$   
364 for the entire simulated interval are suitable approximations. We therefore take model parameters  
365 obtained in the first modeling stage and continue to the next stage.

366

**Stage 2: Adding the margin-parallel sediment transport (MPST) to the model**

368 Fig. 9 shows the simulated bathymetry from 5.3 Ma to the present, using the upgraded model,  
369 which includes the margin-parallel sediment transport (MPST) module. Unlike the basic model  
370 (Fig. 5), here with the addition of this module, a shallow continental shelf is produced (orange  
371 color) offshore Israel with nearly a 2-km-thick section deposited there (Fig. 10A). Throughout the  
372 studied area, the model produces a thickness within <200 m from the observed thickness (Fig.  
373 10B); this is quite a satisfactory result as this value is within <10% of the total observed thickness.  
374 Two modeled chronostratigraphic cross sections offshore Caesarea and Ashdod demonstrate the  
375 good fit with observations (Fig. 11). Furthermore, Two lithological sections at the same location  
376 (Fig. 12) indicate that sand is accumulated only in the upper part of the section. This is also  
377 consistent with observations as the age of the Kurkar Group varies laterally, from Pliocene-  
378 Pleistocene inland to only Pleistocene at the coastline (Fig. 12C, Gvirtzman and Buchbinder,  
379 1978). According to our model, farther west at the shelf area, the age of the sandy Kurkar Group

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380 is limited to a few hundred thousands of years and to only several tens of thousands of years near  
381 the shelf edge (Fig. 12). The noticeable sand intercalations into the shaly section represent intervals  
382 of eustatically low sea levels (Haq et al., 1987) used as part of the model input.

383 The successful simulation of sand deposition and its accretion in the shallow environment is the  
384 outcome of the depth-dependent sand/shale ratio (Fig. 4) that was fed to the model. This depth  
385 dependency is also expressed in Fig. 13 and indicates that more sand was deposited offshore Israel  
386 during the latter part of stratigraphic sequence accretion (1.8-0 Ma), when a shallow shelf was  
387 already developed.

388

389 **Rates, accretion, and shelf buildup**

390 To summarize our results, we compare modeled and observed sedimentation rates in the Levant  
391 Basin over the four time intervals of 5.3-2.6 Ma, 2.6-1.8 Ma, 1.8-0.35 Ma, and 0.35-0 Ma (i.e.,  
392 interval durations of 2.7, 0.8, 1.45, and 0.35 My), which were defined as traceable  
393 chronostratigraphic zones by Elfassi et al. (2019) based on seismic- and bio-stratigraphy. These  
394 four stratigraphic units are the basis for the four thickness maps prepared by Ben Zeev and  
395 Gvirtzman (2019) and are the basis for the rates we use in the comparison with the modeling. The  
396 division of the thickness of the four units corresponds to the above interval durations.

397 Deposition rates calculated for the basic model (without MPST), indicate that deposition was not  
398 sufficient to form a continental shelf in the Levant (Fig. 14). In the upgraded model that includes  
399 MPST a continental shelf is gradually built; this buildup is observed in the model to occur first,  
400 offshore Sinai, then offshore Israel, and most recently offshore Lebanon. The calculated rates of  
401 accretion are consistent with observations but present some exceptions. Unit 2 (Gelasian) in the  
402 actual stratigraphy presents an increase in sedimentation rate both in the southwest, near the Nile

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Delta, and along the Levant margin (Fig. 14). This could have been an expression of a temporal increase of the Nile's sediment discharge, which we do not model. Furthermore, the Israel Slump Complex that provided ~1,000 km<sup>3</sup> of sediments from the Levant slope to the basin (Frey Martinez et al., 2005) in a short episode, limited to Unit 2 (Elfassi et al., 2019) and is not modelled either. In addition, our model ignores other local sediment contribution that may have reached the basin from the east (e.g., the Yafo sand member, Druckman et al., 1995; Frey-Martinez et al., 2007; Frey Martinez et al., 2005; Gardosh et al., 2009).

The importance of the MPST is demonstrated by comparing (Fig. 15) sedimentation rates in the deep basin and at the continental margin (locations appear as white dots in Fig. 14). In the basic model, without MPST, the difference between these two localities is minor and the rate of sedimentation is relatively low in both. This is because the model depends almost entirely on slow and gradual diffusion of sediments from the Nile Delta through and into the deep basin. On the other hand, with the addition of the MPST, increasingly more sediments arrive and deposit along the continental margin of the Levant and as a result less sediments reach and deposit in the deep basin.

### **Margin-parallel sediment flux**

Sediment flux maps obtained by trial-and-error are shown in Fig. 16A-B. In these maps each pixel represents a flux (km<sup>2</sup>My<sup>-1</sup>) in a coast-parallel direction (approximately to the northeast). Integrating the pixels along an east-to-west transect across the continental margin, provides the total northward sediment flux (km<sup>3</sup>My<sup>-1</sup>) through that transect line (Fig. 16C-D).

During the first modeled interval (5.3-1.8 Ma), calculated sediment flux offshore Rafah was ~2500 km<sup>3</sup>My<sup>-1</sup> (Fig. 16D, blue line). Most of these sediments were trapped and deposited offshore Israel

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426 while very little volume continued north of Haifa. This has changed during the 1.8-0 Ma interval  
427 with flux offshore Rafah reaching  $\sim 5,300 \text{ km}^3\text{My}^{-1}$  and the flux offshore Haifa increased to  $\sim 3000$   
428  $\text{km}^3\text{My}^{-1}$  (Fig. 16D, red line). Subtracting the modeled flux north of Haifa from the flux offshore  
429 Rafah indicates that about  $2300 \text{ km}^3\text{My}^{-1}$  of sediments were trapped in the shelf south of Haifa;  
430 this is not very different from the volume deposited during the first interval ( $2500 \text{ km}^3\text{My}^{-1}$ ).  
431 The modelled total sediment fluxes are difficult to compare to present-day observations that  
432 measure only the sand flux, but they do produce satisfactory results for deposition and the quite  
433 correct thickness. Sand fluxes are easier to compare (Fig. 16C). They show that during the early  
434 interval less than  $100 \text{ km}^3\text{My}^{-1}$  had entered the Israeli margin from the south (Rafah) and these  
435 were trapped almost immediately (Fig. 16C, blue line). During the 1.8-0 Ma interval nearly  $350$   
436  $\text{km}^3\text{My}^{-1}$  of sand entered from offshore Rafah, most of it trapped before reaching Ashdod, and  
437 about  $40 \text{ km}^3\text{My}^{-1}$  of sand arrived at Haifa (Fig. 16C, red line). These results are slightly lower  
438 than the present-day observations and may be improved by a sensitivity analysis that will include  
439 changes in the sand/shale ratio considered in the model input. In general, approximately 6% of the  
440 modeled sediment flux offshore Israel is sand, concentrated in near coastal areas.  
441 The two flux maps (Figs. 16A, 16B) indicate a northward shift of transport and alongshore  
442 deposition propagation between the first and second sequences. Initially sediments were primarily  
443 trapped offshore Sinai and only then offshore Israel. Moreover, during the first interval sediment  
444 transport offshore Israel occurred mainly at deep waters (i.e., west of the first isobath after isobath  
445 0 m - isobaths 200 m in Fig. 16A), whereas during the second interval transport mostly occurred  
446 on the shelf (i.e., east of isobaths 200 m depth in Fig. 16B).

447  
448 **DISCUSSION**

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449 **Continental shelves and river deltas**

450 Development of continental shelves adjacent to river deltas is quite common, but the dependency  
451 of the shelf evolution on sediment transport from the nearby delta is not fully resolved. Here we  
452 study the accretion of the Levant shelf alongside the Nile Delta, exploring coevality and the  
453 dependency of the Levant shelf on sediment transport from the Nile Delta. First, we show that  
454 longshore sand transport is only a small fraction of a much larger process, termed here  
455 MPST. In the Levant shelf, sand accumulation in the past 5 My comprises only about 6% of  
456 the total volume of sediments that arrived from the Nile Delta. Furthermore, without the  
457 large transport of silt and clay in the deeper waters, deposition offshore Israel probably  
458 would have been negligible, and the Levant shelf never would have developed (Fig. 14).

459 Our observations and modelling results indicate that MPST has begun coevally with delta  
460 formation, but at first, developed quite slowly. Initially, a continental shelf formed offshore  
461 Sinai, then offshore Israel, and most recently along the Lebanon coast, where it is probably  
462 still in its initial buildup stage. Throughout this time, longshore sand transport and  
463 deposition has taken place near the coastline, but only recently (sometime in the last  
464 hundreds or tens of thousands of years), when the shelf became shallow enough, sand  
465 transport and deposition expanded westward. We hypothesize that this westward  
466 expansion was more pronounced during eustatic sea-level drops.

467 Our results emphasize the importance of deep-water, margin-parallel currents for sediment  
468 transport (e.g., Rebesco, 2018, 2005; Rebesco et al., 2014; Rebesco and Camerlenghi, 2008;  
469 Schattner and Lazar, 2016 and references within), which is studied less than longshore sand  
470 transport (see introduction) and is still poorly understood. The argument that sand transport along  
471 the Levant margin occurs within a wide depth range (0-400 m), was first proposed by Schattner et

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al. (2015), arguing that sediments are transported by the Levant jet system (LJS), which is a fast water current related to the present day thermohaline circulation in the Eastern Mediterranean. According to Schattner et al. (2015) this jet system initiated during the Pleistocene-Holocene transition (~11 ka), intensifying after the last glacial maximum (LGM). Our results show that MPST started coevally with the Nile Delta formation (5.3 Ma) and intensified at the beginning of the Pleistocene (2.6 Ma). We do not have any knowledge of the water circulation during the Pliocene and Pleistocene. However, based on our knowledge of the depositional history, we infer the paleo-sediment fluxes. These fluxes indicate margin-parallel transport much earlier than the Holocene, even at depths greater than 400 m. At the nearest shore environments, where they are dependent on eustatic sea levels, it is possible that the last emplacement of the jet occurred following the last glacial. If the current interglacial is a clue to the past, this could have been a cyclic occurrence during the Quaternary with early interglacials experiencing an increase in Nile River discharge and sediment supply to its delta accompanied by pronounced LJS that intensified the MPST. Such a cyclic process may have facilitated the construction of the Levant continental shelf.

Obviously, our model is not driven by physical processes. However, it presents a first step in transforming available data about sediment deposition in and on shelves (thickness and lithology) into sediment fluxes. This is essential for future physically based simulations that can examine how surficial, intermediate, and deep marine currents transport sediments along various morphologies of continental margins.

**Margin-parallel vs. margin-perpendicular sediment transport**

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494 In the Levant margin sand is forming a relatively thin sequence at the top of the shelf,  
495 reaching ~180 m at the coastline and wedging out landwards. Sand grains comprise 90-  
496 100% of this sequence (i.e., the Kurkar Group; Fig. 12). Conversely, sand is spread  
497 throughout the entire thickness (>3.5 km) of the PQ section of the Nile Delta, ; from its base  
498 to its top and from its eastern to western edges, comprising a maximum of 50% (Fig. 6B). We  
499 propose that this profound difference in sand content is an expression of the available energy  
500 for transport. Sediment flow from the river mouth straight down to the deep basin is  
501 stronger in comparison with margin-parallel flow that keeps a nearly constant elevation.  
502 Accordingly, offshore the Nile River mouth, sand grains were transported straight  
503 downslope over large distances since the early Pliocene. On the other hand, the much weaker  
504 margin-parallel currents transported only fine grains that had built the shaly Levant shelf.  
505 One exception within the margin-parallel system is the highly energetic nearshore current  
506 that transported sand continuously since the Pliocene. This, however, is limited, here and  
507 probably elsewhere, to a very narrow zone and very shallow water depth. Thus, the  
508 combination of near shore energetic currents and deeper weaker currents produce an  
509 efficient differentiation between shale and sand deposits, although they source at the same  
510 river mouth. Such differentiation characterizes margin-parallel sub-horizontal sediment  
511 transport in contrast with the margin-perpendicular steep-gradient transport.

512 An additional basin-wide expression of the margin-parallel vs. the common margin-  
513 perpendicular transport is deltas growing thicker towards the open sea more than they grow  
514 wider along the adjacent continental margins. The Nile Delta started developing immediately  
515 following the recovery from the Messinian salinity crisis (early Pliocene) and then it had  
516 propagated 150-200 km from the river mouth northward within about 1 My (Fig. 9). At the



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517 same time the Levant coast started receiving small amounts of sand (Pleshet Formation,  
518 Gvirtzman and Buchbinder, 1969). The shelf, which has existed earlier (probably was  
519 initially forming in the Miocene) presumably was very narrow (Buchbinder and Zilberman,  
520 1997; Gvirtzman and Buchbinder, 1969). Only 2-3 My after the beginning of delta formation  
521 a notable shelf began developing in the Levant margin (Figs. 9, 11).

522

**523 Implications for shelf exposure during sea level drops**

524 The youngest stage (tens or hundreds of thousands of years) of delta development is the  
525 extension of sand deposition 10-20 km westward during eustatic lowstands (Fig. 13). These  
526 results are consistent with the proposal of Amit et al. (2011) that only around 180-200 ky,  
527 the Nile Delta has been accreted to levels that its surface could have been exposed during  
528 eustatic low stands. Only then, the sand surface and deposits of the delta and the northern  
529 Sinai shelf could have been subaerially exposed and initiate their eolian transport under the  
530 windy glacial climate in northern Sinai. This sand transport has generated the northern Sinai  
531 dune fields and abraded the sand grains into coarse silts that comprise the loess of the Negev  
532 (Amit et al., 2011; Crouvi et al., 2008). We hypothesize that some of the shelf sand was also  
533 exposed and was involved in the accretion of the Eastern Mediterranean coastal plain (Harel  
534 et al., 2017).

535

**536 CONCLUSIONS**

- 537 1. Longshore sand transport is a part of a much broader process termed here margin-  
538 parallel sediment transport (MPST). Sand is transported in a nearshore shallow  
539 water belt where marine currents are highly energetic. Shale transport occurs at

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540 greater depths over the deeper shelf and the continental slope where marine  
541 currents are weaker.

2. Without the large transport of silt and clay in the deeper waters, deposition offshore  
542 Israel probably would have been negligible, and the Levant shelf never would have  
543 developed.  
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3. The delay in the formation of the Levant continental shelf relative to the Nile Delta is  
545 not a result of external forcing. MPST has begun coevally with delta formation but  
546 developed slower. Initially, a continental shelf formed offshore Sinai, then offshore  
547 Israel, and most recently along the Lebanese coast.  
548

4. Throughout this time, longshore sand transport and deposition has taken place very  
549 close to the coastline, but only recently (last hundreds of thousands or tens of  
550 thousands of years), when the shelf became shallow enough, sand transport and  
551 deposition expanded 10-20 km westwards. This implies subaerial exposure of sandy  
552 landscape during recent eustatically low stands.  
553

5. Down slope sediment transport into deep basins is relatively energetic and carries  
554 sand, silt, and clay, whereas margin-parallel currents are relatively weak and carry  
555 mainly silt and clay. One exception within the margin-parallel system is the highly  
556 energetic nearshore current that transport sand. Thus, MPST is an efficient  
557 differentiator between shale and sand.  
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560 **ACKNOWLEDGMENTS**

561 This study was supported by the Israeli Ministry of Energy and by the Mediterranean  
562 Research Center of Israel (MERC I). The article is further based upon work of COST Action

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4 563 “Uncovering the Mediterranean salt giant” (MEDSALT) supported by the European  
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7 564 Cooperation in Science and Technology.

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14 889 **FIGURE CAPTIONS**

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17 890 Figure 1: Topography and bathymetry map of the Eastern Mediterranean. Her = Herzliya.

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19 891 Ash = Ashkelon. ESM = Eratosthenes Seamount. Black circles mark oil and gas wells:

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22 892 1=Mira, 2=Hannah 01, 3=Sara, 4=Hof Ashdod 01, 5=Yam 02, 6=Yinnon, 7=Maanit 01,

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24 893 8=Caesarea. Black line w/triangles mark the Hellenic–Cyprus subduction. Dashed brown

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27 894 lines = Locations of cross sections. Black arrows – longshore sand transport (after Inman

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29 895 and Jenkins, 1984). Red arrows – LJS (after Schattner et al., 2015)

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31  
32 896 Figure 2: (A) Structural map of the base of the Pliocene. Black lines are faults after Aal et al.

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34 897 (2000). (B) A map showing the thickness of the Plio-Quaternary sedimentary sequence.

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37 898 Present-day coastline (gray line) is shown for reference. Figure after Zucker et al (2019).

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39 899 Figure 3: (A) Reconstructed bathymetry for 5.3 Ma (obtained by flexural unloading of the

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41 900 PQ sediment from the base Pliocene map, see text). (B) Subsidence since the time of the

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44 901 base of the Pliocene sequence (5.3 Ma) until today. (C) 2D Illustration of the relationships

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47 902 between the data presented in Figs. 1-3.

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49 903 Figure 4: (A) Sediment grain size as documented on the present-day seabed offshore Israel

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51 904 (after Almogi-Labin et al., 2012); the data indicate the coast-parallel distribution of the

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54 905 sand, silt, and clay belts. (B) Depth dependent sand/shale ratio parameters introduced in

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56 906 the model. Generated following Almogi-Labin et al. (2012).

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907 Figure 5: Maps illustrating the evolution of the bathymetry through time as modeled by  
908 DionisosFlow using the basic sediment transport model without the MPST (see text). Red  
909 lines show locations of cross sections. (A) Same as Fig. 3A.

910 Figure 6: (A) Model results using a basic sediment transport that show a good fit with a  
911 cross section (B-B', see 6D for location) after Kellner et al. (2009) across the Nile Delta. (B)  
912 Cross section A-A' (see D for location) of the Nile Delta sand [%] that builds the delta at the  
913 end of the model that includes MPST (see text). (C) Interpreted seismic composite line  
914 across the Nile Delta (Kellner et al., 2009). (D) Location map of cross sections B-B', and A-  
915 A'. Dashed black line in cross section A-A', which shows present-day bathymetry. Color bar  
916 is the same for Figs. 6A and 6C.

917 Figure 7: (A and B) Results of chronostratigraphic model using basic sediment transport  
918 that indicate a bad fit to the observed present-day bathymetry reference (black dotted line)  
919 across the Levant continental margin. Black vertical lines mark locations of oil and gas  
920 wells with their names written above (see wells and cross section locations in Fig. 1).

921 Figure 8: (A) Total sediment thickness (from 5.3-0.0 Ma) as simulated without MPST (see  
922 text). (B) The difference in sequence thickness between simulated (A) and observed.

923 Figure 9: Bathymetry evolution through time as modeled by DionisosFlow implementing  
924 MPST. Note the difference between this model, which produces the Levant shelf, and the  
925 basic model that does not (Fig. 5). Red lines show locations of cross sections.

926 Figure 10: (A) Total sediment thickness, from 5.3 to 0.0 Ma generated by simulation  
927 including MPST (see text). (B) Difference between simulated (A) and observed sediment  
928 thickness. Note the relatively small difference here in comparison with the basic, without  
929 MPST, model (Fig. 8b).

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4 930 Figure 11: (A and C) Chrono-stratigraphy of sections across the Levant margin derived  
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7 931 from the model after adding the MPST. These sections show a good fit to observed sections  
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9 932 based on seismic data (B and D). Black vertical lines in cross sections mark locations of oil  
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12 933 and gas wells. Well names and cross section locations are in Fig. 1.  
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14 934 Figure 12: Simulated MPST cross sections (A and B) showing the resulted Kurkar (Mainly  
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17 935 sand) and Yafo Formations (mainly shale); black rectangle in B shows the location of the  
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19 936 enlarged cross section (C) (after Gvirtzman and Buchbinder, 1978). The sand fingers into  
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22 937 the shaly section of Yafo Formation is the result of including the eustatic curve of Haq et al.  
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24 938 (1987) in the model. (D) Location map. Color bar is the same for cross sections A and B.  
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27 939 Figure 13: The total sand thickness as simulated for the lower sequence (5.3-2.6 Ma) and  
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29 940 for the upper sequence (2.6 -0.0 Ma), using the model without and with the MPST laws. (A)  
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32 941 Lower sequence, using the basic model. (B) Lower sequence, using the MPST model. (C)  
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34 942 Upper sequence, using the basic model. (D) Upper sequence, using the MPST model.  
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37 943 Figure 14: Sedimentation rates generated by both modeled scenarios in comparison with  
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39 944 the observed sedimentation, divided into the four chronostratigraphic units determined by  
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42 945 Ben Zeev and Gvirtzman (2020). White dots mark the locations points taken, one in the  
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44 946 deep basin and a second at the continental margin, for calculating sedimentation rates  
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47 947 measured in figure 15.  
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49 948 Figure 15: Comparing sedimentation rates for the models without and MPST (red and gray,  
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52 949 respectively) and the observed (black) sedimentation rate, divided in to two locations:  
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54 950 Deep basin (solid line) and continental margin (dashed line). Locations in figure 14.  
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57 951 Figure 16: Sediment flux maps obtained by trial and error for lower (A) and upper (B)  
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59 952 sequences printed over isobaths representing 3 Ma and 1 Ma, respectively. Each pixel in the  
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953 maps represents the flux ( $\text{km}^2\text{My}^{-1}$ ) in a coast-parallel direction. Total sediment (D) and sand  
954 (C) fluxes crossing margin-perpendicular transects during deposition of lower and upper  
955 sequences. Curve representing the observed sand flux is based on references listed in the  
956 introduction.

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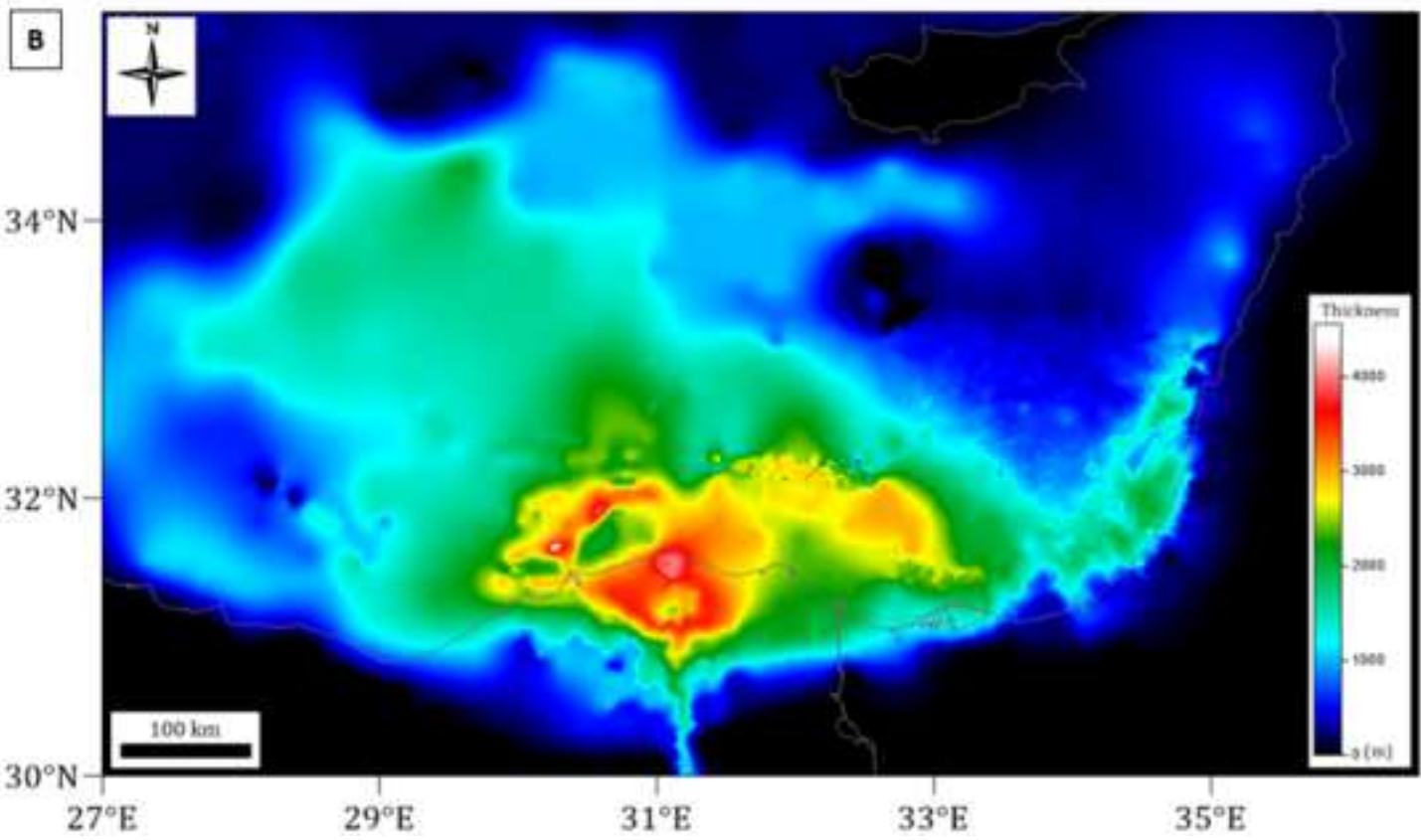
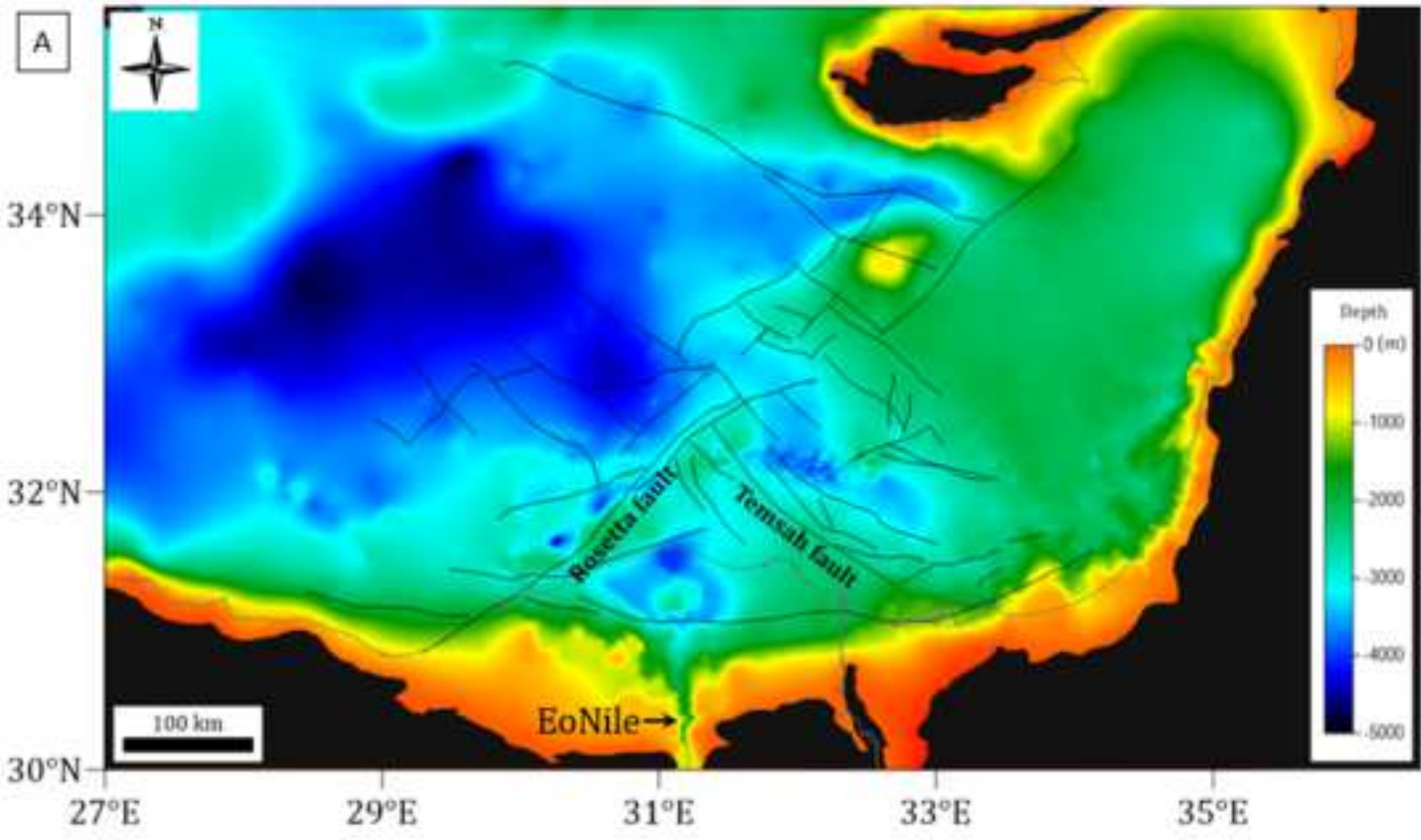
958 Table 1: Model parameters from the literature and this study.

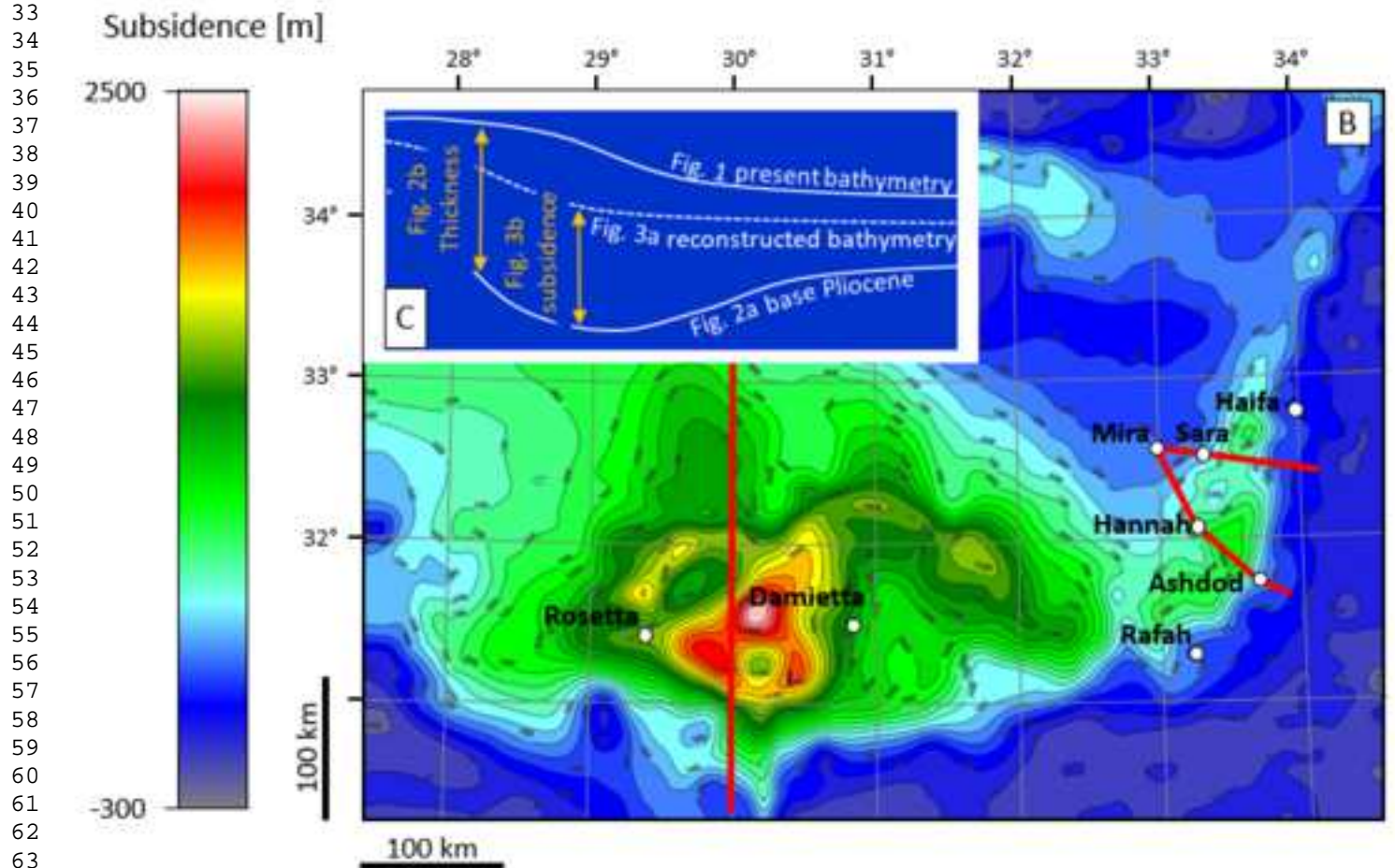
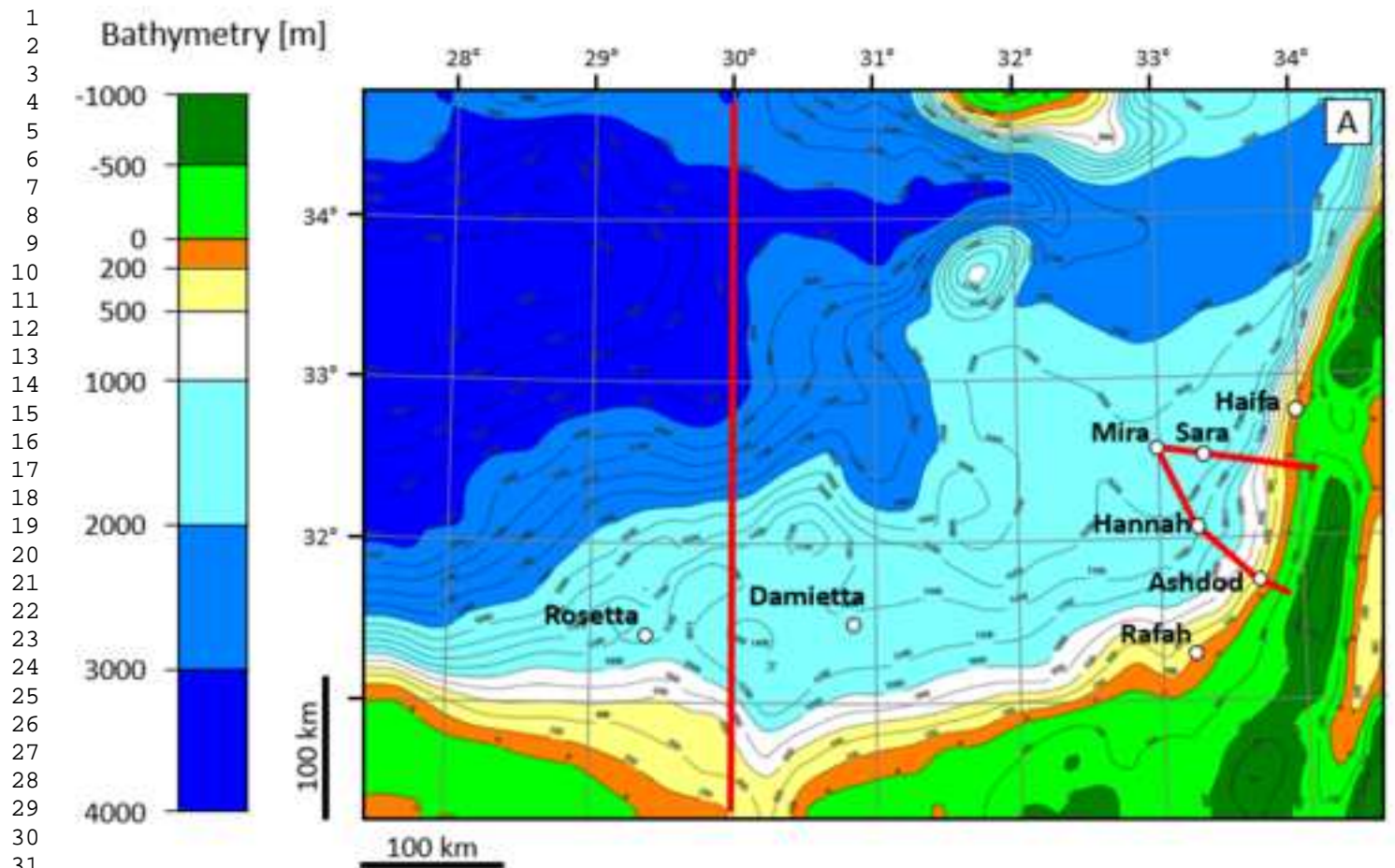
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Figure 2

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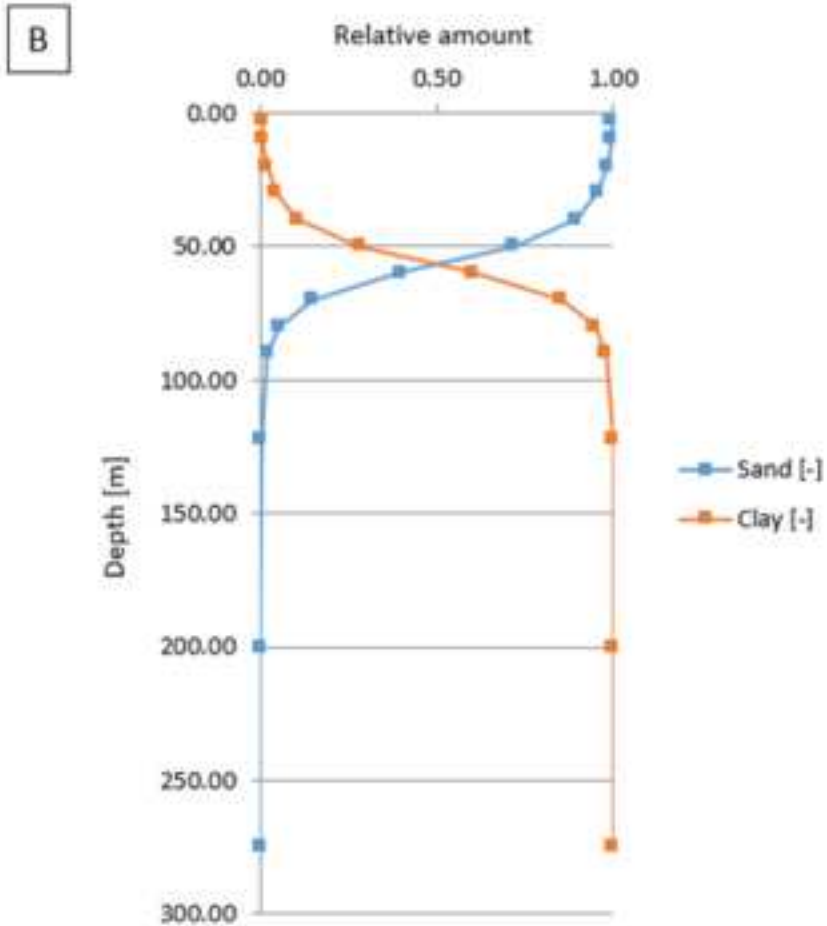
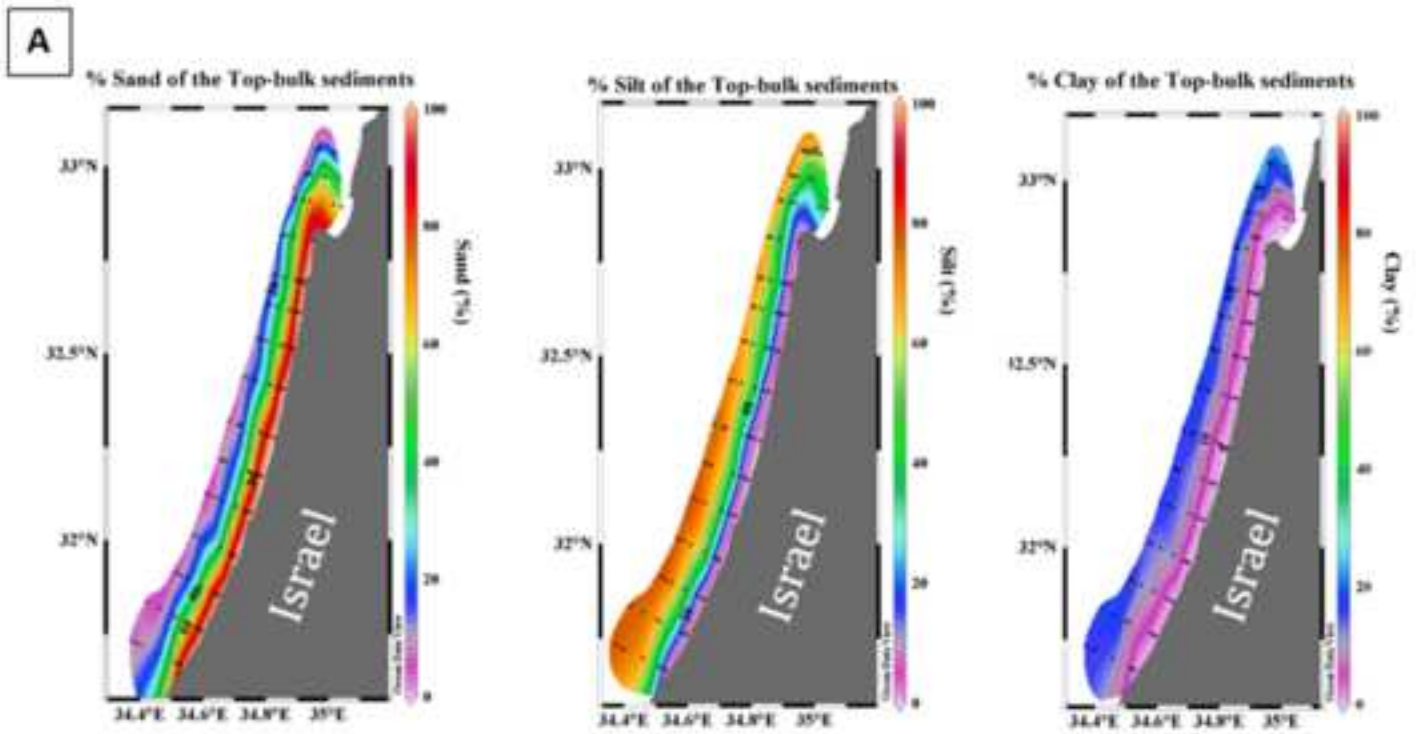
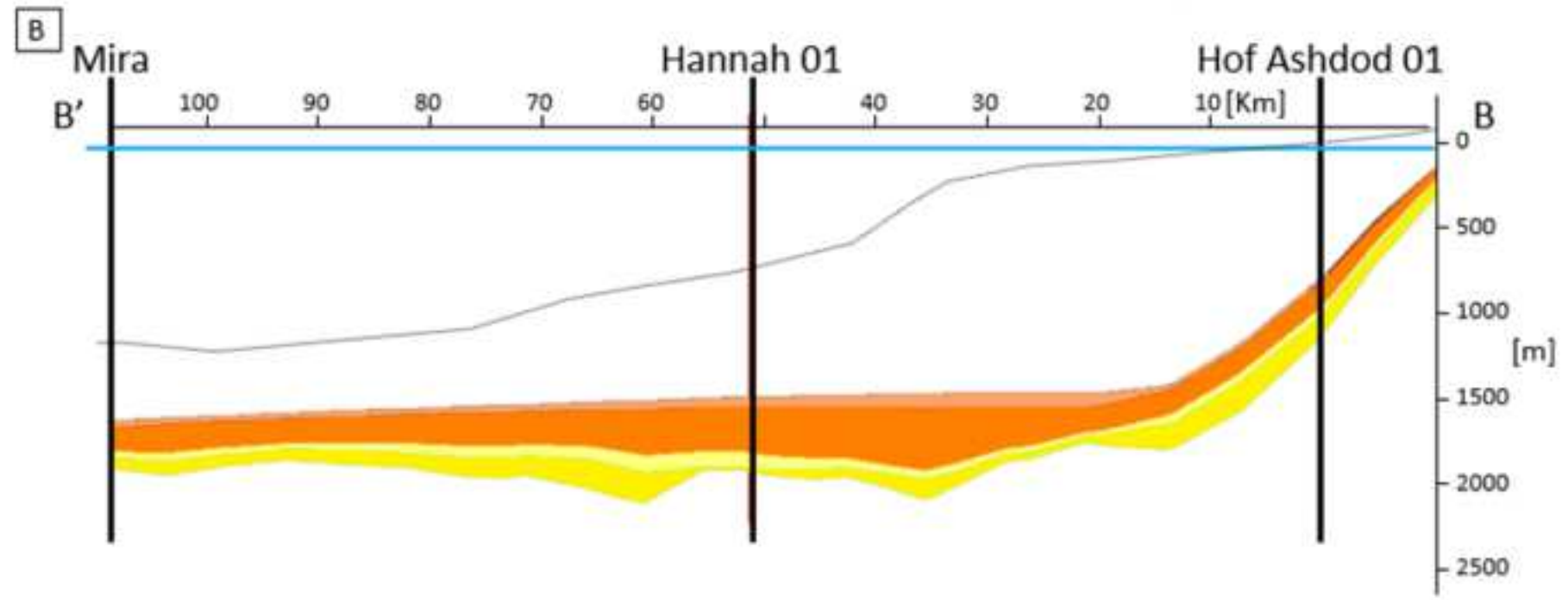
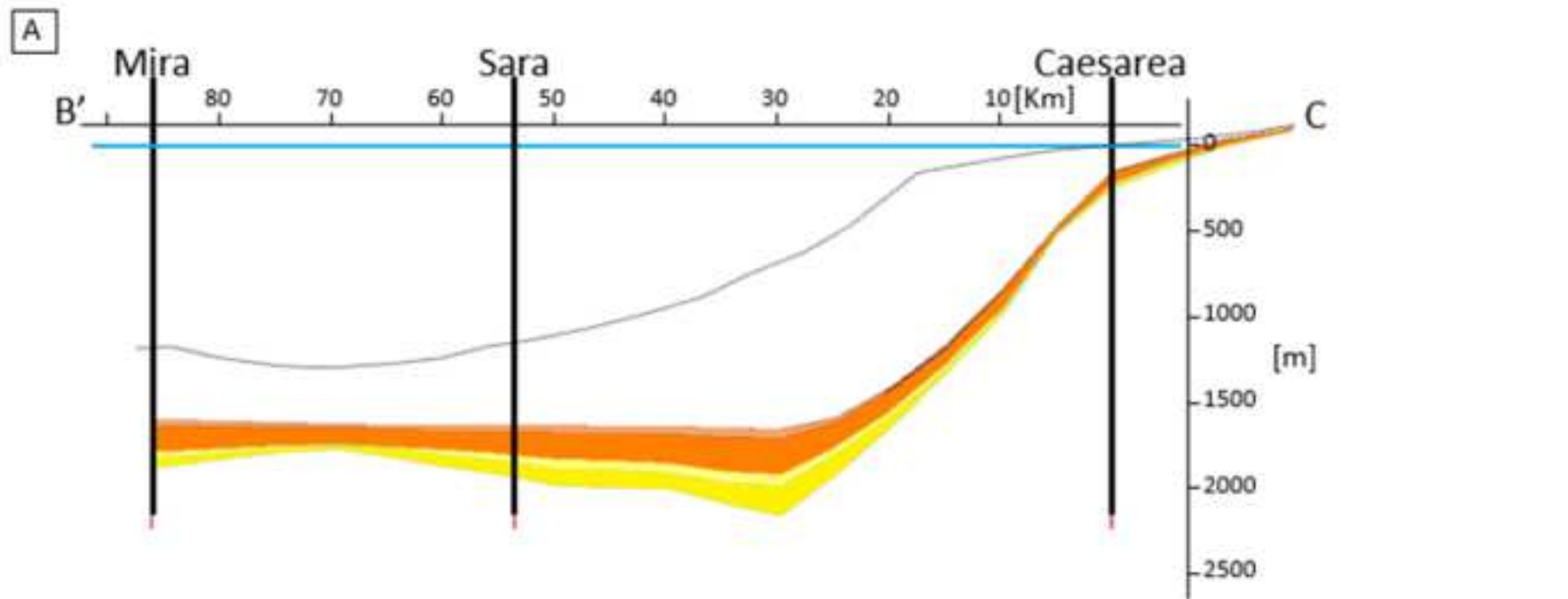


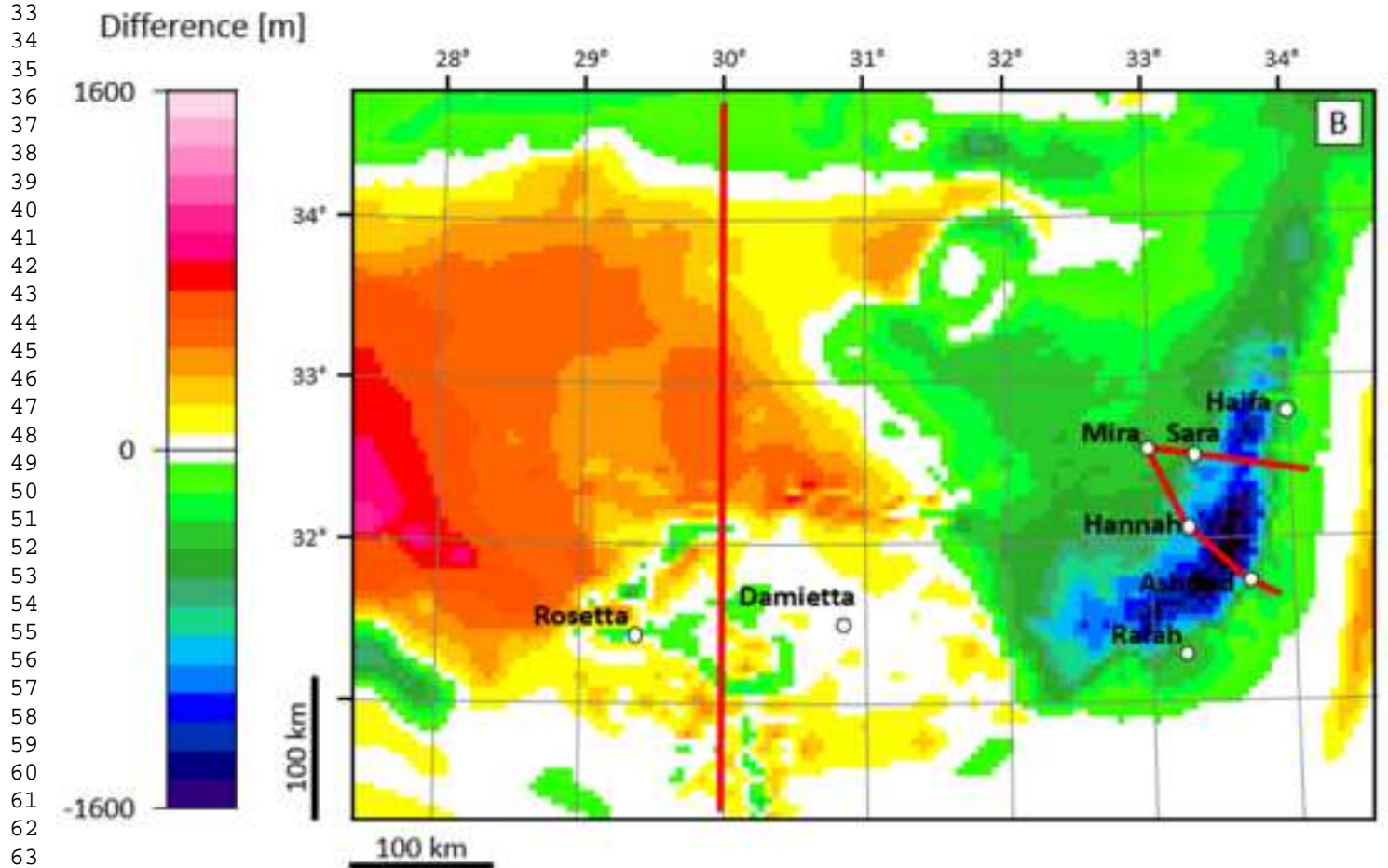
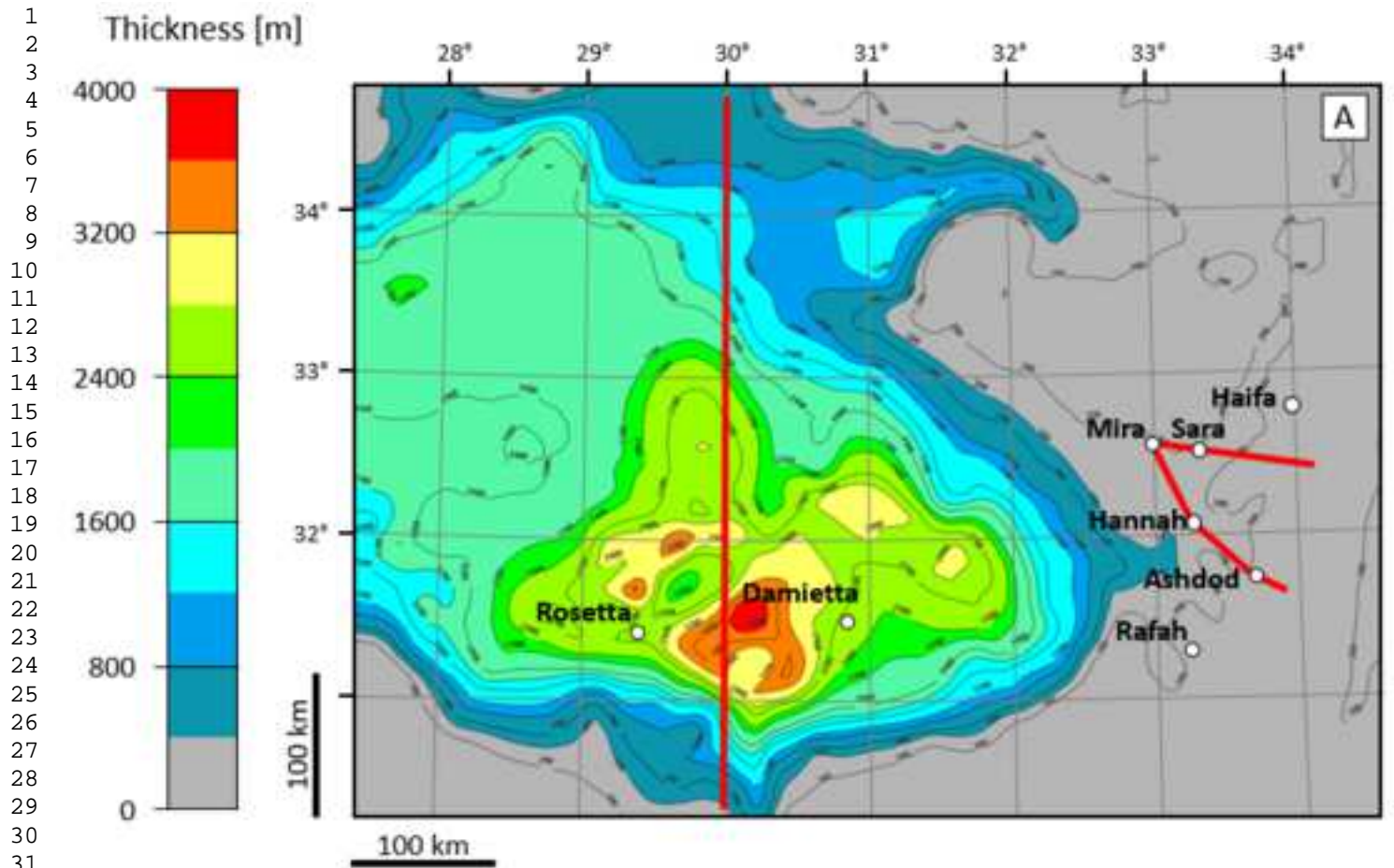
Figure 7

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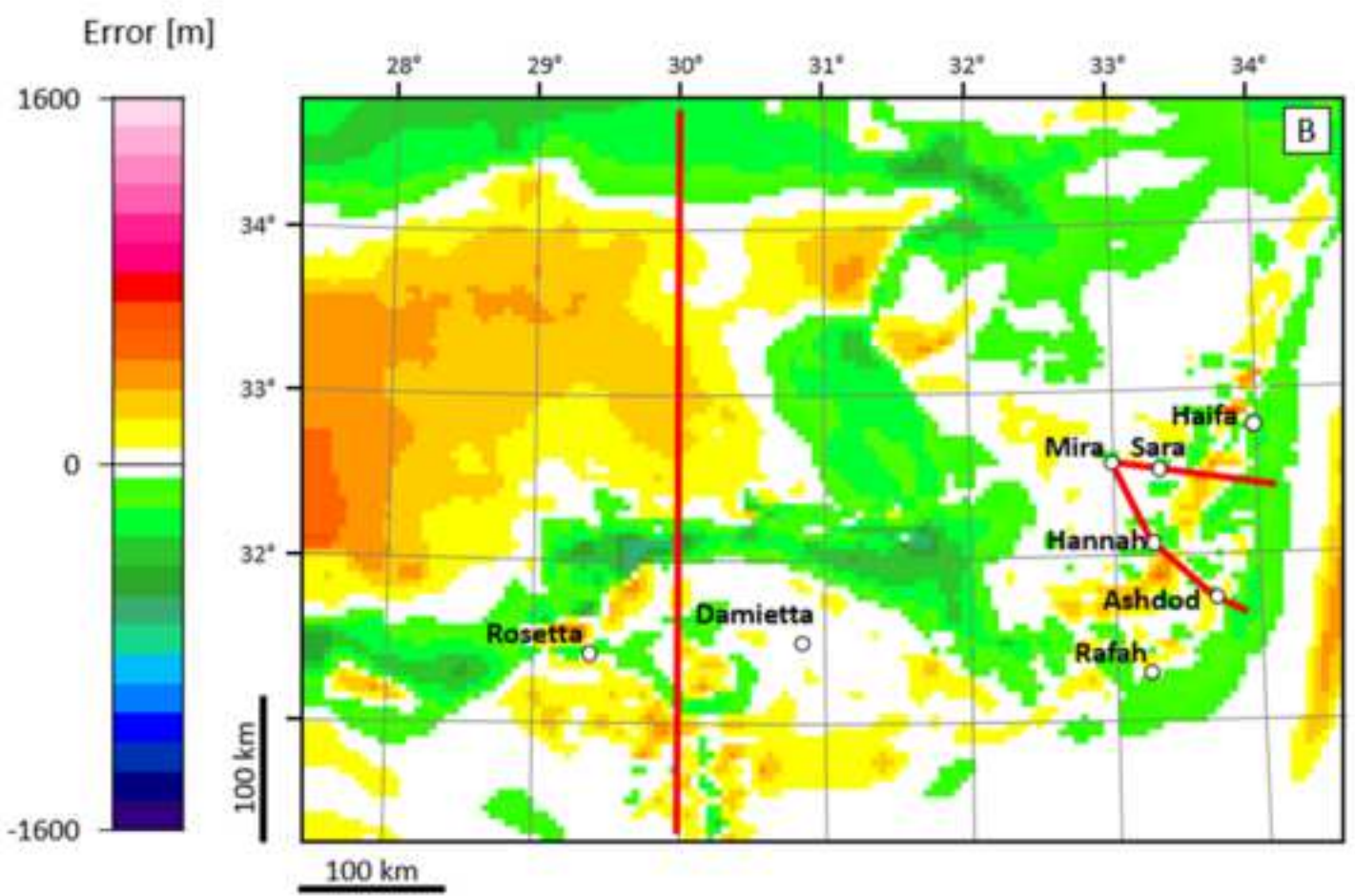
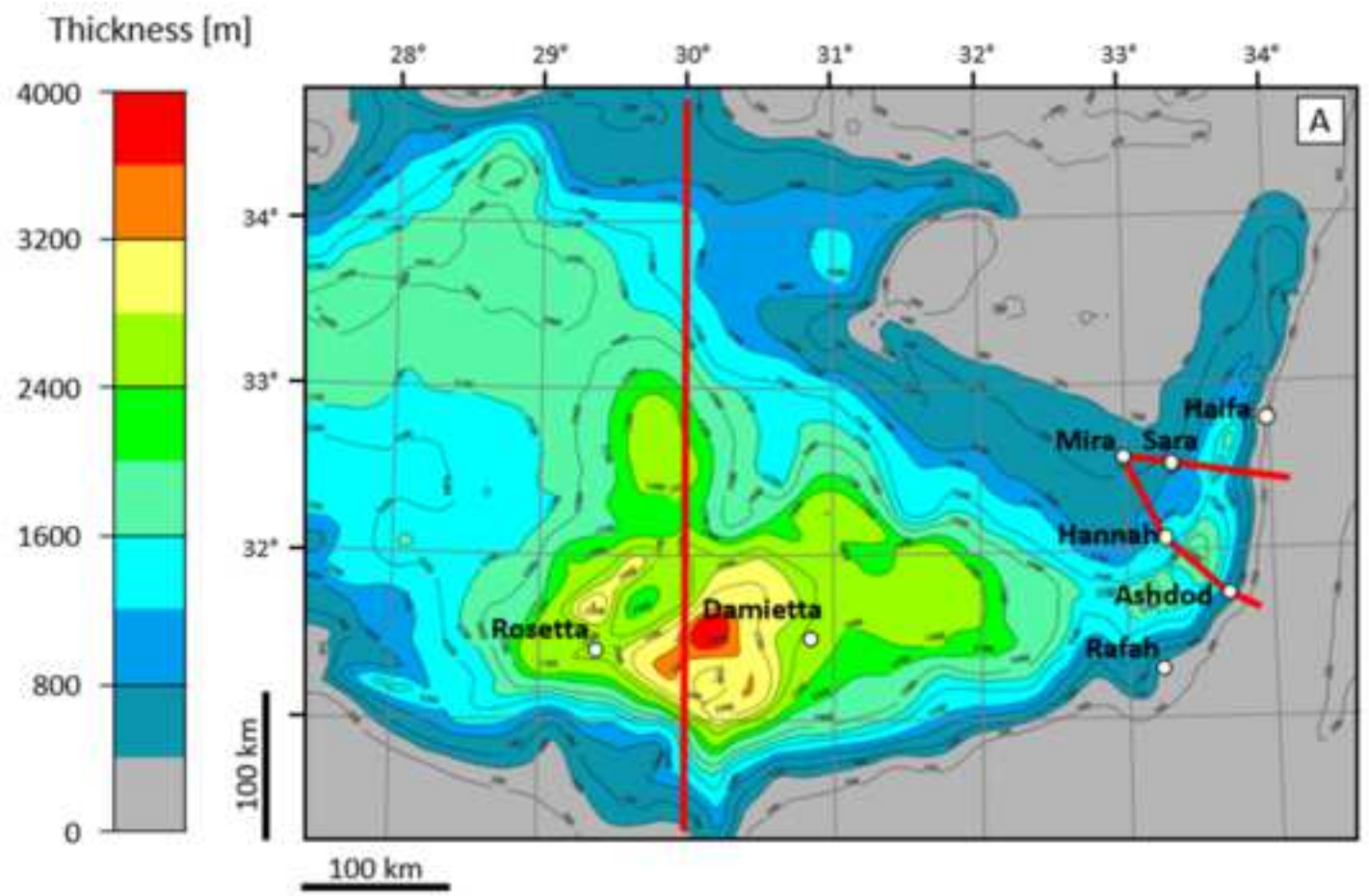


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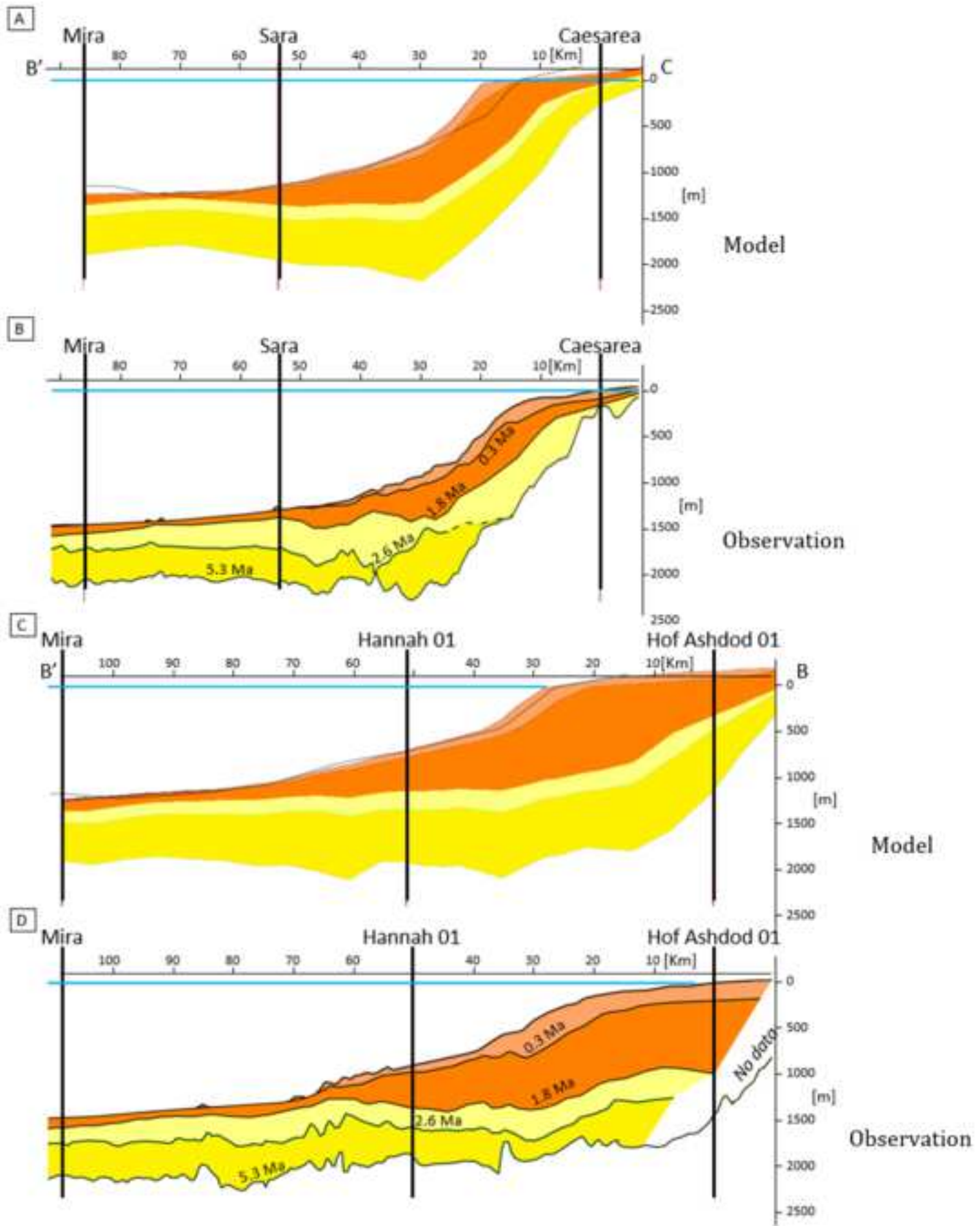




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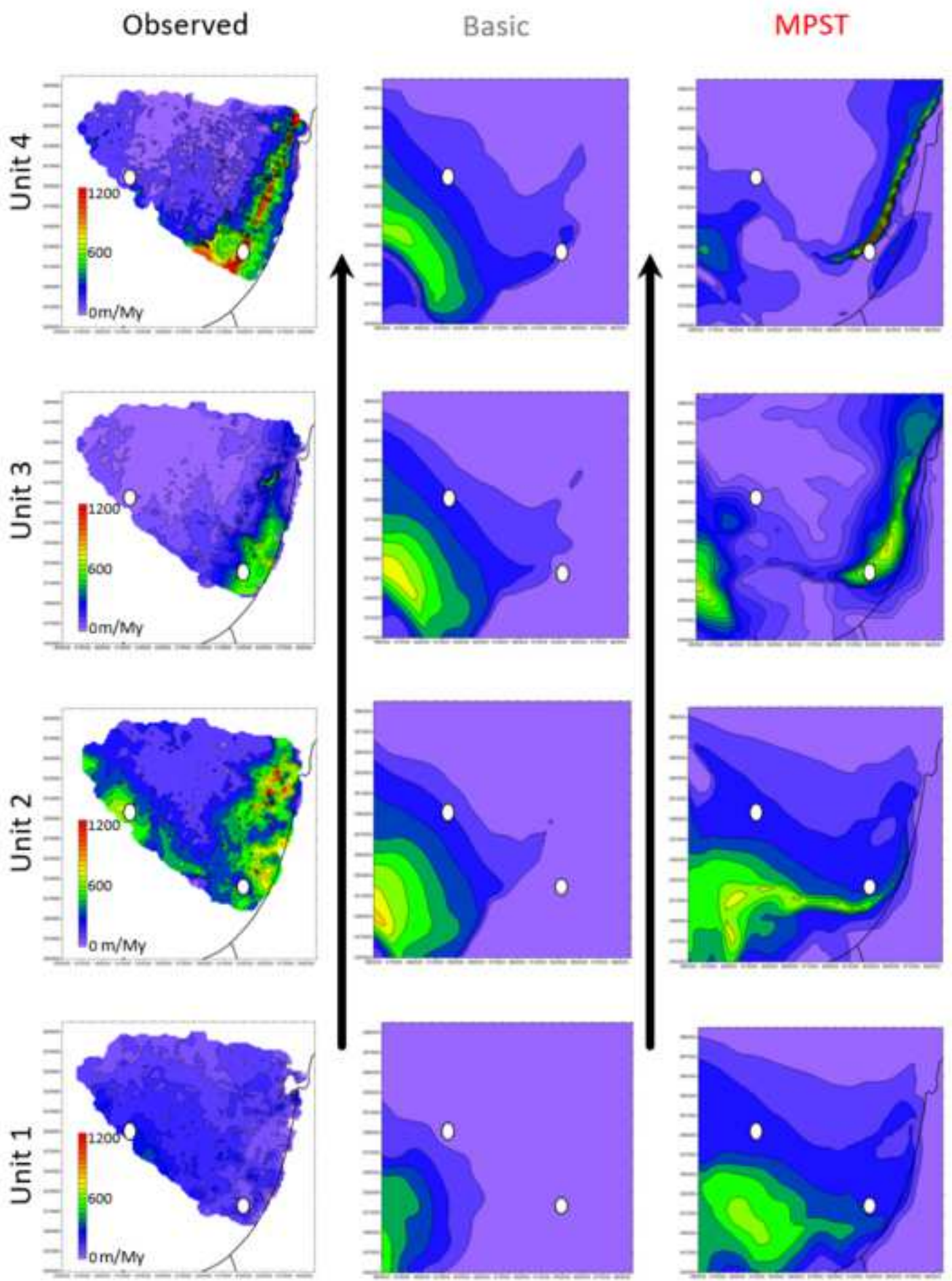
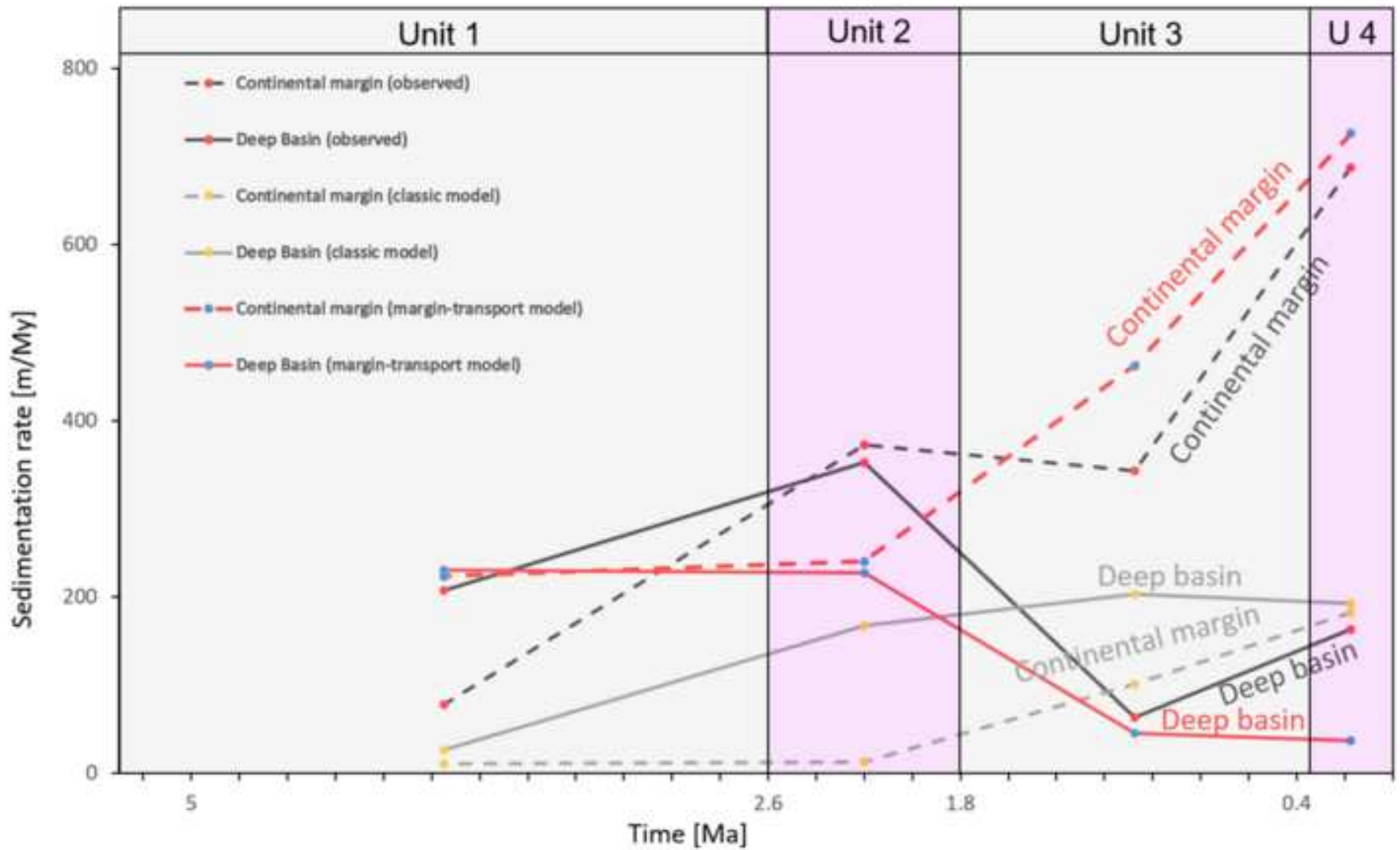
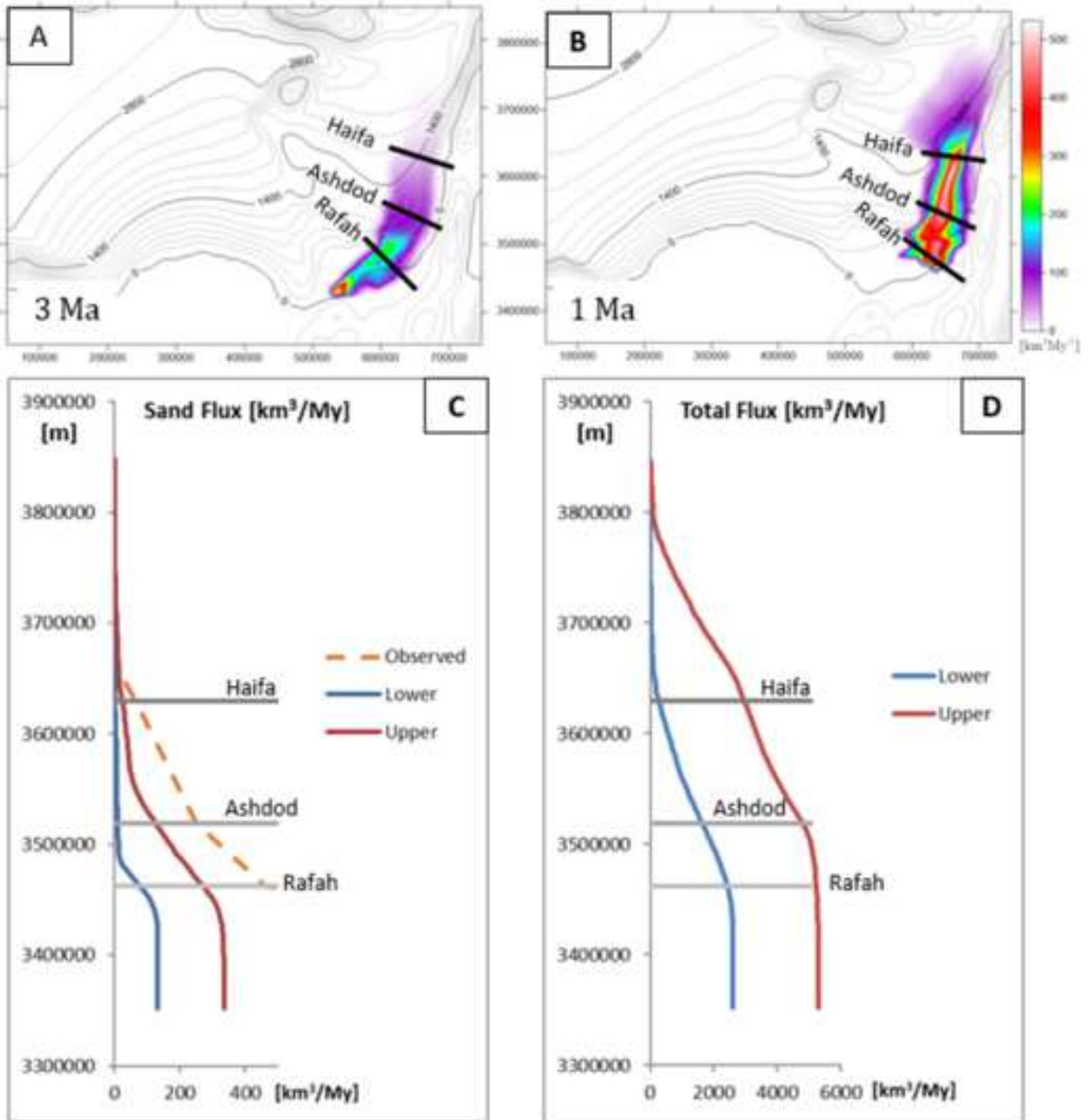


Figure 15



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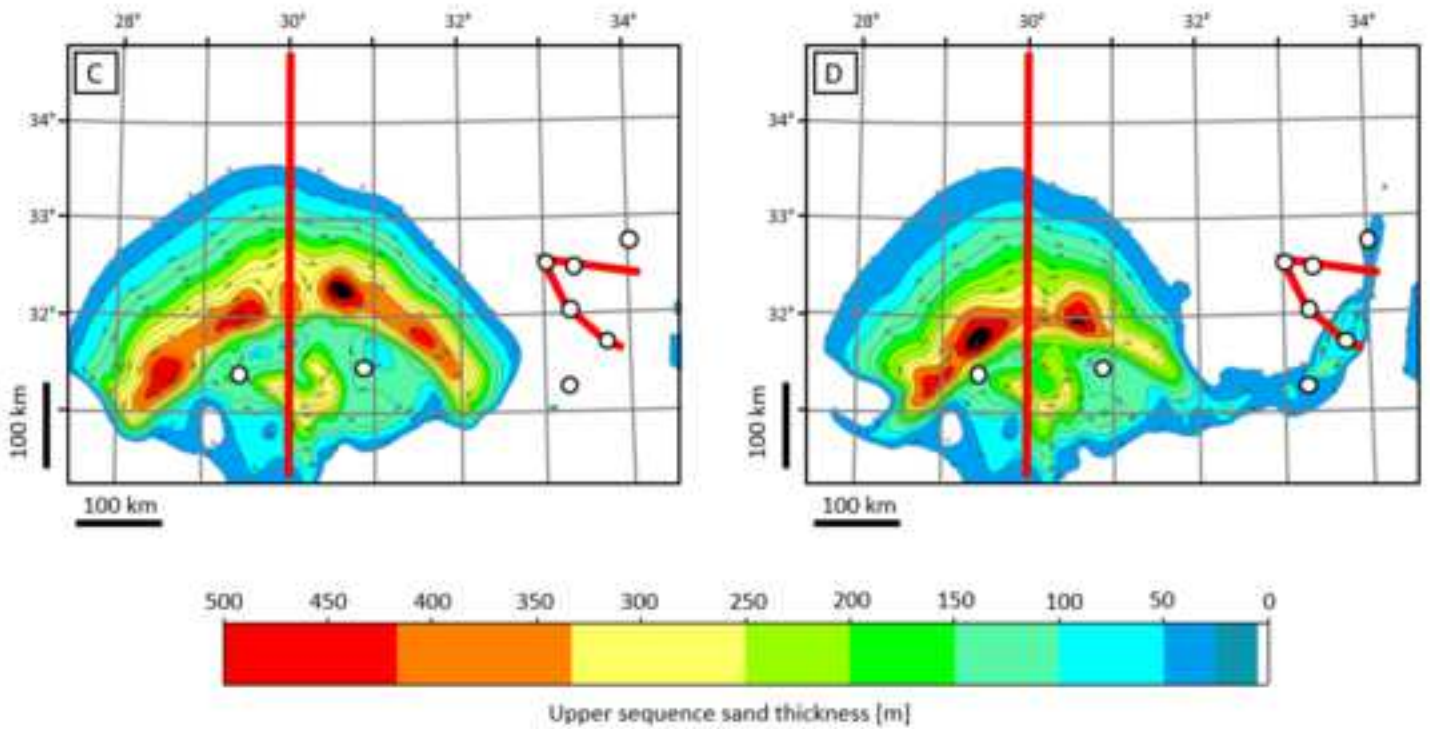
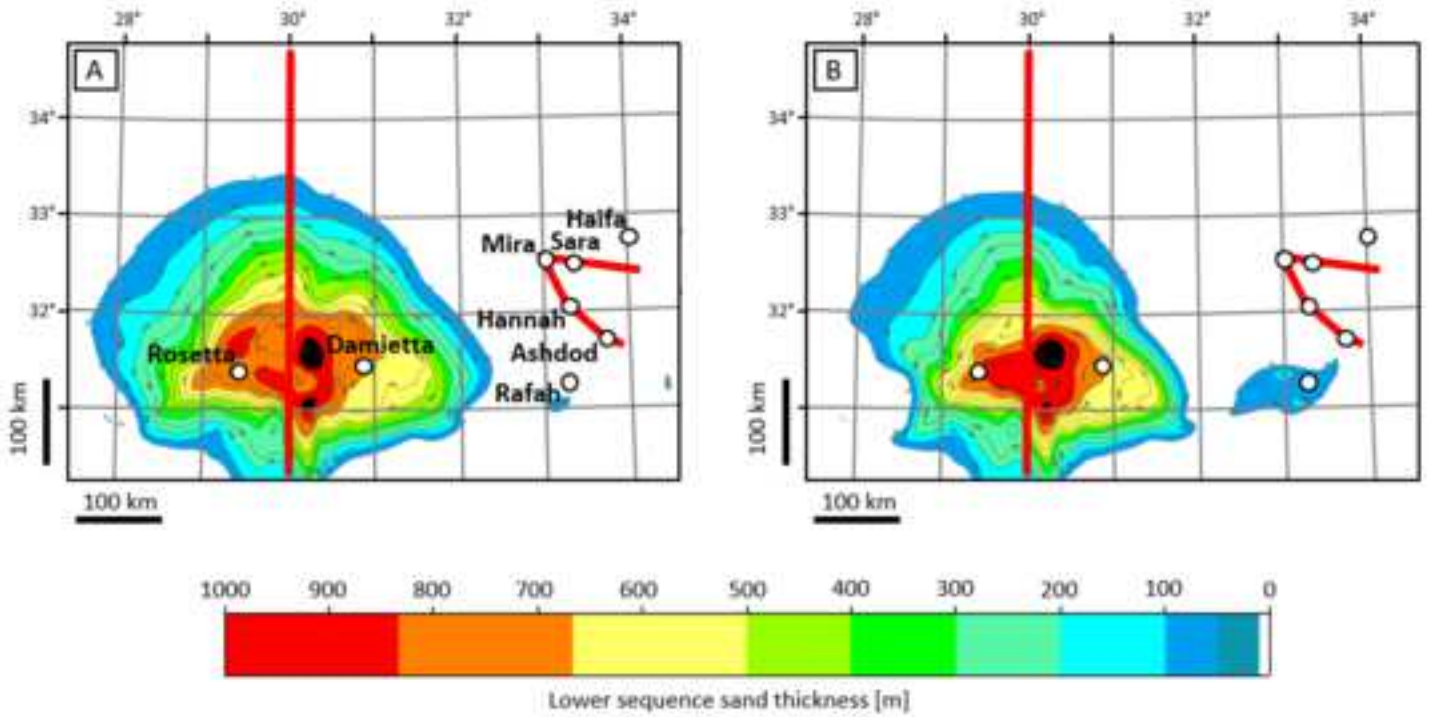
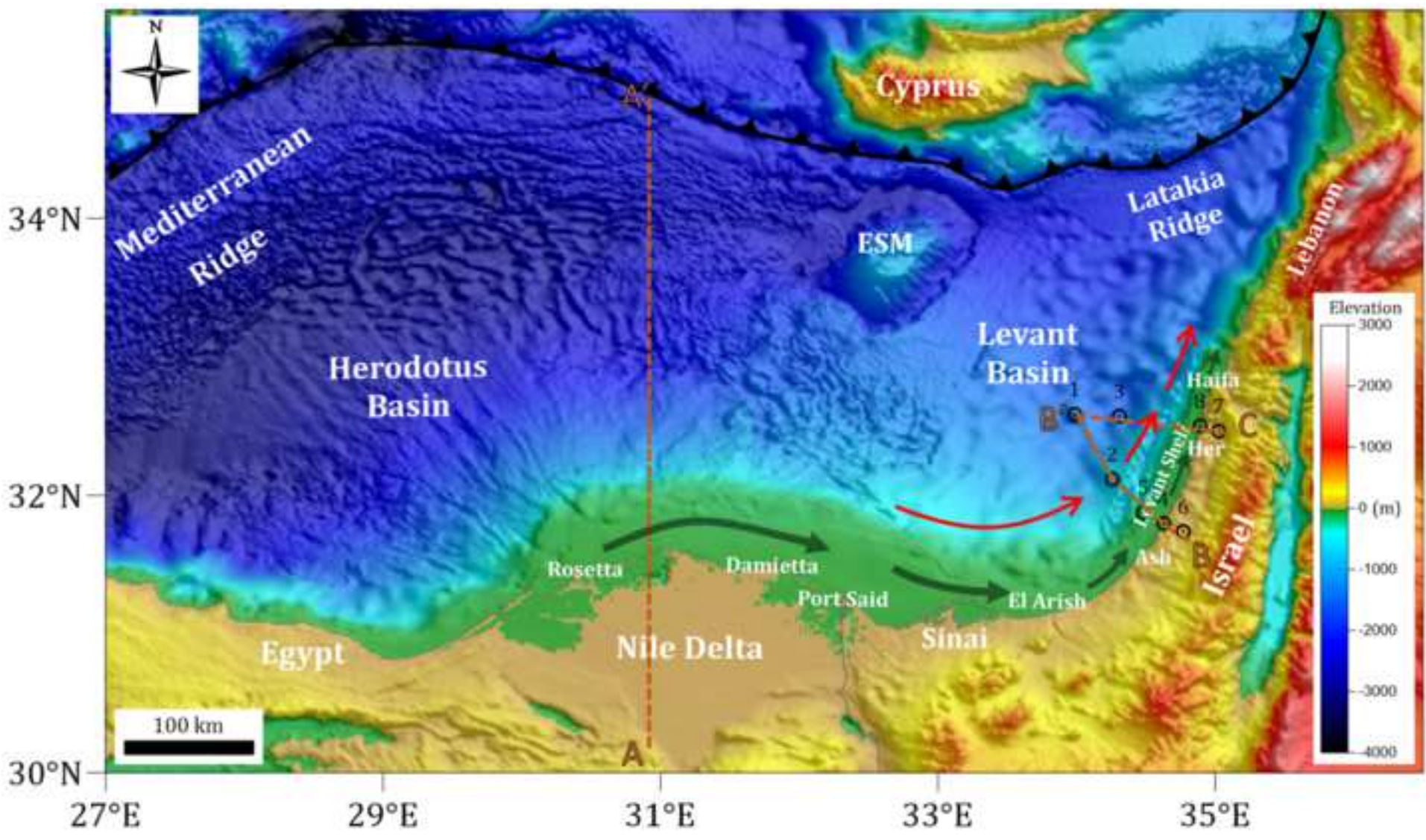
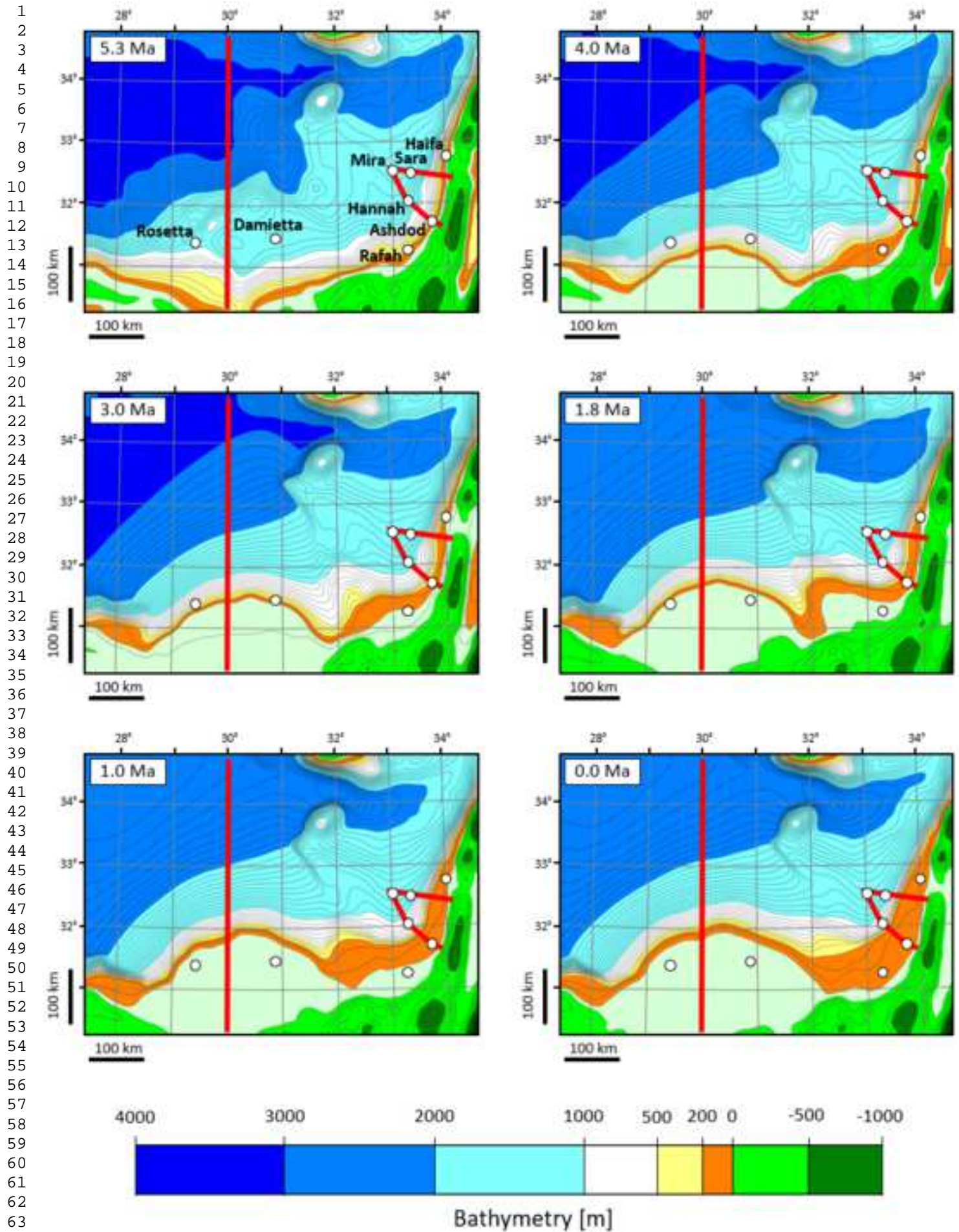




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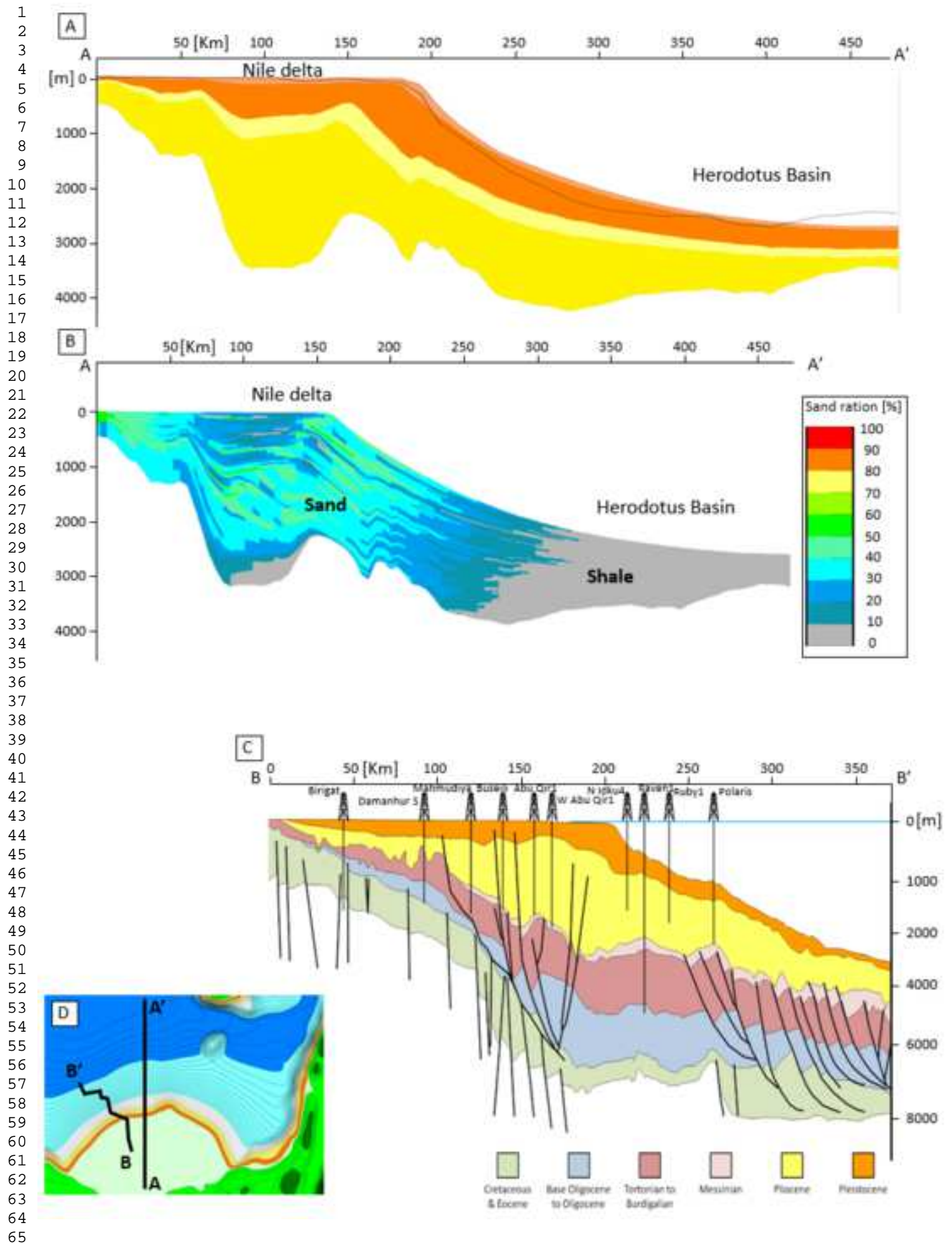
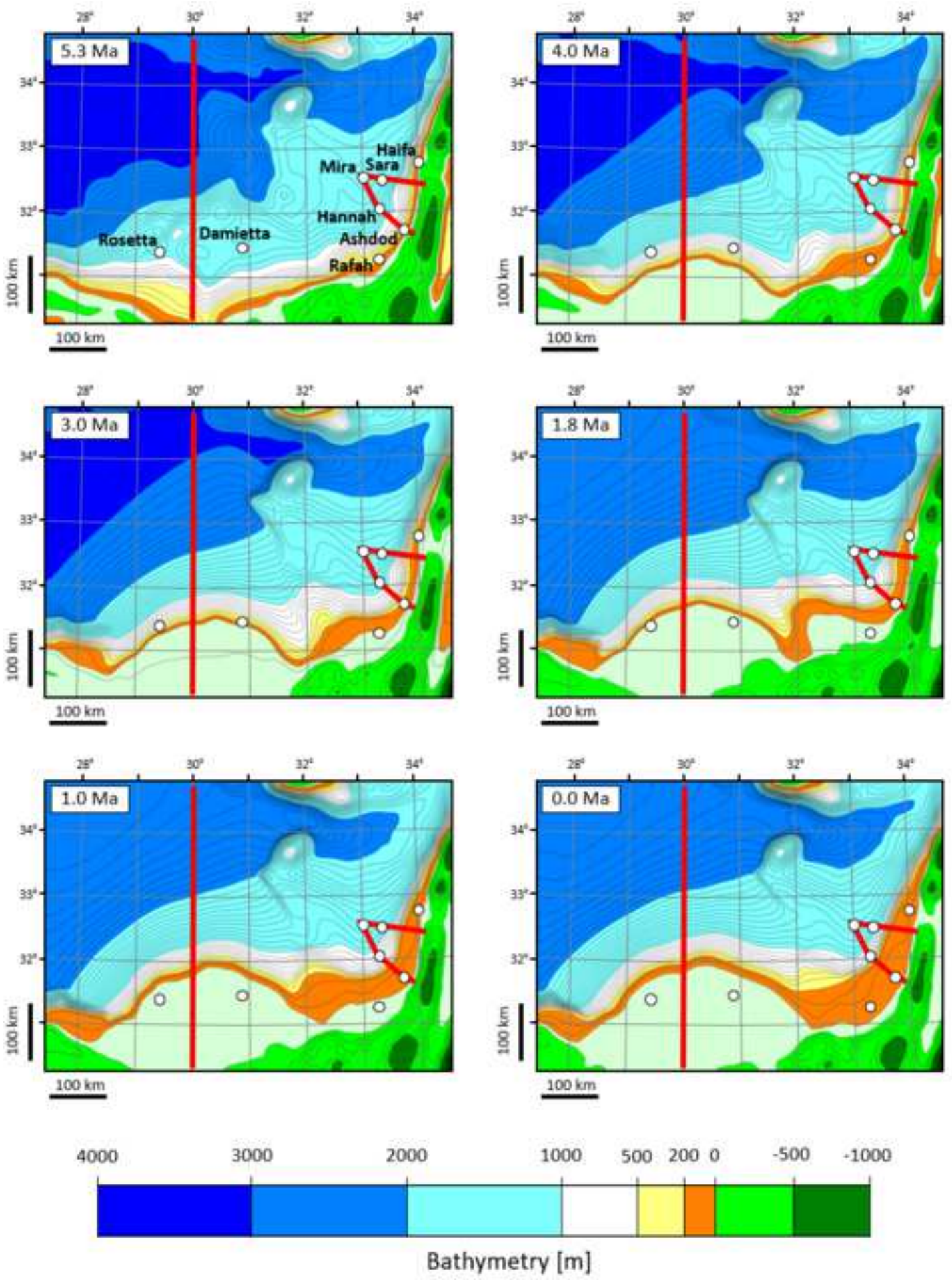


Figure 9

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## Diffusion Coefficient [km<sup>2</sup>/ky]

### Sand

		K gravity terrestrial [km <sup>2</sup> /ky]	K gravity marine [km <sup>2</sup> /ky]	K water terrestrial [km <sup>2</sup> /ky]	K water marine [km <sup>2</sup> /ky]
Csato et al., 2014	Delta	0.001	0.001	250	5
Zhang et al., 2018	Delta	0.05	0.005	50	0.01
Csato et al., 2013	Margin	0.001	0.001	2000	0.4
Harris et al., 2016	Margin	0.001	0.001	250	1
Gvirtzman et al., 2014	Turbidites	0.001	0.001	Not Defined	2
Hawie et al., 2017	Turbidites	0.001	0.001	Not Defined	0.3
Hawie et al., 2018	Turbidites	0.001	0.007	Not Defined	20
<b>range</b>		<b>0.001 - 0.05</b>	<b>0.001 - 0.005</b>	<b>50 - 2000</b>	<b>0.01 - 20</b>
<b>Levant (this study)</b>	<b>Delta and Margin</b>	<b>0.001</b>	<b>0.001</b>	<b>100</b>	<b>0.5</b>

### Shale

		K gravity terrestrial [km <sup>2</sup> /ky]	K gravity marine [km <sup>2</sup> /ky]	K water terrestrial [km <sup>2</sup> /ky]	K water marine [km <sup>2</sup> /ky]
Csato et al., 2014	Delta	0.001	0.001	1000	5
Zhang et al., 2018	Delta	0.1	0.05	100	0.1
Csato et al., 2013	Margin	0.001	0.001	4000	10
Harris et al., 2016	Margin	0.001	0.001	1000	5
Gvirtzman et al., 2014	Turbidites	0.001	0.001	Not Defined	10
Hawie et al., 2017	Turbidites	0.001	0.001	Not Defined	0.063
Hawie et al., 2018	Turbidites	0.001	0.001	Not Defined	100
<b>range</b>		<b>0.001 - 0.1</b>	<b>0.001 - 0.05</b>	<b>100 - 4000</b>	<b>0.063 - 100</b>
<b>Levant (this study)</b>	<b>Delta and Margin</b>	<b>0.01</b>	<b>0.01</b>	<b>500</b>	<b>2.5</b>