

DIPARTIMENTO DI SCIENZE
SCIENZE DELLA TERRA

XXIX

Late Cenozoic uplift, active tectonics and natural hazards of the Central

Anatolian Plateau southern margin, Turkey

Nazik Öğretmen

Prof. Domenico Cosentino

Prof. Claudio Faccenna



DOCTORAL DISSERTATION

LATE CENOZOIC UPLIFT, ACTIVE TECTONICS AND NATURAL HAZARDS OF THE CENTRAL ANATOLIAN PLATEAU SOUTHERN MARGIN, TURKEY

PhD Student: Nazik Öğretmen

Advisor: Prof. Domenico Cosentino

Co-Advisor: Prof.ssa Elsa Gliozzi

Table of Contents

GENERAL INTRODUCTION.....	4
1. LONG-TERM UPLIFT AND PLEISTOCENE PALAEOGEOGRAPHY OF THE CENTRAL ANATOLIAN PLATEAU SOUTHERN MARGIN.....	11
1.1 Introduction.....	13
1.2 Geology of the Central Anatolian Plateau (CAP) southern margin.....	14
1.3 Post-Miocene evolution of the southern flank of the CAP.....	17
1.4 Material and methods.....	18
1.5 Stratigraphy of the sampled sections.....	22
1.6 Results.....	25
1.6.1 Gülnar East section.....	25
1.6.2 Tol section.....	33
1.7 Discussions.....	36
1.7.1 Age model.....	36
1.7.2 Palaeoecology of the Pleistocene marine environment at the CAP southern margin	40
1.8 Plio-Pleistocene palaeogeography at the CAP southern margin, Pleistocene uplift rate, and mechanisms for the uplift.....	44
1.9 Conclusions.....	48
Acknowledgements.....	50
2. COASTAL EVIDENCE FOR SHORT-TERM UPLIFT AND ACTIVE TECTONICS AT THE CAP SOUTHERN MARGIN.....	52
2.1 Introduction.....	54
2.2 Coastal geomorphic markers and archaeological infrastructures useful to evaluate sea-level changes.....	58
2.2.1 Marine notches.....	58
2.2.2 Beachrocks.....	60
2.2.3 Algal rims	61
2.2.4 Alluvial deposits.....	62
2.2.5 Archaeological infrastructures.....	63
2.3 Material and methods.....	63

2.4 Discussions and conclusions.....	69
Acknowledgements.....	71
 3. NATURAL HAZARDS ALONG THE COASTS OF SOUTHERN FLANK OF THE CENTRAL ANATOLIAN PLATEAU.....	 75
3.1 Introduction.....	76
3.2 Natural disasters in the Eastern Mediterranean as recorded in the archives.....	78
3.3 Storm vs. tsunami boulders.....	84
3.4 Different boulder transportation modes.....	87
3.5 Methodology.....	89
3.5.1 Identification and measurements of out-of-place boulders.....	91
3.5.2 Classification of the out-of-place boulders (storm vs tsunami).....	96
3.6 Results.....	96
3.6.1 Narlıkuyu	96
3.6.2 Yeşilovacık.....	98
3.6.3 Aydıncık.....	100
3.7 Discussion.....	104
3.8 Conclusion.....	104
Acknowledgements.....	105
GENERAL CONCLUSIONS.....	106
REFERENCES.....	112
SUPPLEMENTARY MATERIAL.....	128



GENERAL INTRODUCTION

Orogenic plateaus are topographic features with high mean surface elevation that are mainly formed as a result of mountain building processes along the boundaries of the plateau that are generated by plate tectonics (Strecker et al., 2007; Allen et al., 2015). They form in collision, extrusion, and subduction plate-boundary settings and so are associated with a variety of geodynamic processes. Orogenic plateaus are special environments where the tectonics (e.g. collision, subduction) influence climate dynamics (Strecker et al., 2007; Ehlers and Poulsen, 2009; Ehlers, 2013). As a result, they are characterized by intensive erosional processes along the margins, which are also affected by relatively high precipitation rate, whereas the internal part of the plateaus are characterized by arid climate and less intensive erosional processes (Molnar et al., 1993; Allmendinger et al., 1997; Montgomery et al., 2001).

Orogenic plateaus are very useful places to make geological research considering their both long- and short-term evolution in terms of geodynamic processes (e.g., collision, subduction, etc.), climate changes (e.g. humidity vs aridity), and natural hazards (e.g., earthquakes, landslides, tsunamis, volcanic eruptions, etc.). From well investigated regions such as Tibetan Plateau and Andean Plateau, it has been revealed that they show very distinctive features such as steep margins (e.g., Himalaya Mts. for the Tibetan Plateau; Andes Mts. for the Andean Plateau), arid to semi-arid climates (e.g., Taklamakan Desert in the Tibetan Plateau; Atacama Desert in the Andean Plateau), and low-relief topography in the central parts of the orogenic plateaus in contrast to their steep margins.

The Central Anatolian Plateau (CAP) is an example for collisional-type plateaus, considering its geodynamical position in relation with the colliding Africa, Eurasia and Arabian plates. Anatolia has been continuously built for millions of years by several geodynamic processes such as: continental collision (Şengör & Yılmaz, 1981; Bozkurt & Mittweide, 2001; Jolivet & Faccenna, 2000; Keskin, 2003; Okay, 2008; van Hunen & Allen, 2011; McQuarrie & Van

Hinsbergen, 2013), plateau uplift (Şengör, 2003; Keskin, 2007; Göğüş & Pysklywec, 2008; Şengör et al., 2008; Bartol et al., 2012; Cosentino et al., 2012; Schildgen et al., 2012, 2012a; Bartol & Govers, 2014; Radeff et al., 2015;), oceanic subduction (Şengör & Yılmaz, 1981; Robertson & Dixon, 1984; Dercourt et al., 1986; Stampfli, 2000; Okay et al., 2001; Parlak et al., 2002; Parlak & Robertson, 2004; Robertson et al., 2004a, 2004b) and escape tectonics (Dewey and Şengör, 1979; Şengör et al., 1985; Burke and Şengör, 1986; Barka and Reilinger, 1997; Jaffey and Robertson, 2004; Faccenna et al., 2006). All those geodynamic processes gave rise to a convenient smaller-scale natural laboratory in the eastern Mediterranean, suitable to study the main steps of orogenic plateaus development.

Therefore, the small CAP shows the same features of the larger orogenic plateaus in the world, such as the Tibetan and the Andean plateaus. It is characterized by steep margins, both at the northern and at the southern margins. The northern margin of the CAP goes together with the Pontide Mts., whereas the basement of the southern margin is made by the central Tauride tectonic units. The CAP shows: (1) arid to semi-arid environments in the “ova” province (Şengör et al., 1985), (2) high precipitation along the northern and southern flanks of the plateau, which increase their landslide hazard (e.g., northern flank, Duman et al., 2005; Erener and Düzgün, 2013; southern flank, Sivrikaya et al., 2008), (3) high seismicity at the plateau margins (e.g., North Anatolian Fault, East Anatolian Fault, Ecemiş Fault, etc., Kalafat et al., 2009), and (4) subdued interior topography (Schildgen et al., 2014).

The geodynamic mechanism responsible for the development of orogenic plateaus and the active tectonic boundary at its margins is still matter of debate. Different geodynamical processes for growing orogenic plateaus have been suggested by different authors, such as: (1) crustal shortening and thickening (Avouac, 2003; Hatzfeld and Molnar, 2010), (2) magmatic underplating and lower-crustal flow (Royden et al., 1994), (3) mantle delamination (Kay and Kay, 1993), and (4) surface processes (e.g., erosion, Avouac, 2003; Hatzfeld and

Molnar, 2010). For the development of the CAP, four hypotheses have been suggested by the researchers, which are: (1) Eratosthenes Seamount subduction/collision (Robertson and Shipboard Scientific Party, 1996), (2) lithospheric slab break-off and tearing (Duretz et al., 2011), (3) lithospheric delamination and upper mantle convection (Kay and Kay, 1993; Göğüş and Pysklywec, 2008), or (4) multi-phased uplift including all these hypotheses (Schildgen et al., 2012; 2014). Matter of debate are also the uplift rates that affected Anatolia during the different phases of plateau growth, and the differential uplift between Cyprus and Anatolia, if any. More open questions include differences during the different phases of the plateau uplift and possible changes in the uplift rates through time. Finally, is the uplift of the CAP still going on?

This PhD thesis has been developed within the Initial Training Network on the “Anatolian pLateau climatE and Tectonic hazards” (Marie Curie ITN-ALeRT), which is focused on the tectonic and climatic boundary conditions along the densely populated margins of the Central Anatolian Plateau (Turkey), and associated natural hazards. The main active processes of the study area are: (1) subsiding Adana-Cilicia Basin versus uplifting CAP southern margin (Aksu et al., 2005; Burton-Ferguson et al., 2005; Ben-Avraham et al., 2006; Cosentino et al., 2012; Radeff et al., 2015; 2016); (2) subducting plate beneath the Cyprus Arc (Kelling et al., 2005; Aksu et al., 2009); (3) seismicity and movements along the active Ececi and Kozan Fault systems (Barka & Reilinger, 1997; Akhun et al., 2008; Aksu et al., 2014a, 2014b); and (4) movements along the active Dead Sea Fault Zone (Salamon et al., 1996; Akyuz et al., 2006). The tectonic activity on those different features makes the Eastern Mediterranean a complex region to be studied.

Within this tectonic framework, in this work we tried (1) to better define the uplift rate of the CAP southern margin, (2) to understand if the area underwent any Holocene uplift and if the uplift is still going on, (3) to recognize any active tectonic features related to the uplift of the

CAP southern margin, (4) to evidence any natural hazards connected with the geodynamics of the CAP southern margin.

At the CAP southern margin, Neogene marine deposits unconformably overlie the basement units of the Central Taurides, up to 2 km of elevation (Cosentino et al., 2012; Cipollari et al., 2012). The age of these marine deposits was classically used to constrain the uplift of the CAP southern margin at the Middle-Late Miocene (Özgül, 1976; Dhont et al., 1999).

More recently, according to the age of the marine deposits cropping out in the Ermenek Basin (Başyayla section), a post-Tortonian age was suggested for this event (Cosentino et al., 2012; Cipollari et al., 2013a). Moreover, the stratigraphical architecture of the marine deposits capping the CAP southern margin, which shows an unconformity surface within the late Neogene marine succession, was recently used for defining a multi-phased uplift of the CAP southern margin, with a second uplift phase in the early Calabrian (ca. 1.6 Ma) (Schildgen et al., 2012a).

One of the aims of this thesis is to refine the starting age for the second uplift phase that affected the CAP southern margin after the late Messinian uplift (Schildgen et al., 2012b; Radeff et al., 2016). To do this, a micropaleontological study on samples collected at around 1000 m a.s.l. on the CAP southern margin (Gülнар and Tol villages) has been carried out on both benthic and planktonic foraminifera, ostracods, as well as calcareous nannofossils.

Coastal geomorphology (i.e., abrasion platforms, trottoirs, and notches), geoarchaeology, and AMS ^{14}C datings have been used to evidence the Holocene behavior, in terms of uplift or subsiding, of the CAP southern margin between Mersin and Aydıncık (southern Turkey). Finally, the occurrence along the coastal area between Narlıkuyu and Aydıncık (southern Turkey) of out-of-place boulders, evidenced for the first time working for this thesis, shows the natural hazards of the study area in connection with the geodynamical processes active at

the CAP southern margin.

Following the main aims of the study, the thesis has been organized in three different chapters, which will be considered for submitting manuscripts to be published in international journals. Part of the contents of the last chapter, which are related to the tsunami hazard of the coastal area of the CAP southern margin, has been already published in Öğretmen, N., Cosentino, D., Gliozzi, E., Cipollari, P., Iadanza, A., & Yıldırım, C. (2015): Tsunami hazard in the Eastern Mediterranean: geological evidence from the Anatolian coastal area (Silifke, southern Turkey), *Natural Hazards*, 79(3), 1569–1589. <http://doi.org/10.1007/s11069-015-1916-2>



CHAPTER 1

LONG-TERM UPLIFT AND PLEISTOCENE PALAEOGEOGRAPHY OF THE CENTRAL ANATOLIAN PLATEAU SOUTHERN MARGIN

Table of Contents

1.1 Introduction.....	12
1.2 Geology of the CAP southern margin.....	14
1.3 Post-Miocene evolution of the southern flank of the CAP.....	17
1.4 Material and methods.....	18
1.5 Stratigraphy of the sampled sections.....	22
1.6 Results.....	25
1.6.1 Gülnar East section.....	25
1.6.2 Tol section.....	33
1.7 Discussions.....	36
1.7.1 Age model.....	36
1.7.2 Palaeoecology of the Pleistocene marine environment at the CAP southern margin.....	40
1.8 Plio-Pleistocene palaeogeography at the CAP southern margin, Pleistocene uplift rate, and mechanisms for the uplift.....	44
1.9 Conclusions.....	48
Acknowledgements.....	50

1.1 Introduction

The orogenic plateaus, with elevation of several kilometers and remarkable relief contrast with the surrounding areas, are first order tectonic and topographic features on the Earth's surface. Their development, resulting from a fast regional rock uplift, caused the major palaeogeographical and paleoclimatic changes occurred on the Earth. According to many authors, the evolution of the Asian monsoons is strictly linked to phases of Himalaya-Tibetan Plateau uplift (Ruddiman and Kutzbach, 1989; Molnar et al., 1993; Zhisheng et al., 2001; Harris, 2006) , whereas the South American and Andean climate changed significantly in response to the Altiplano-Puna Plateau growth (Gregory-Wodzicki, 2000; Hartley, 2003; Strecker et al., 2007; Ehlers and Poulsen, 2009). In the Middle East, as a consequence of the Arabia and Eurasia collision, which started at ca. 20 Ma (e.g., Okay et al., 2010) along the Bitlis-Zagros thrust zone, in the Serravallian the Black Sea-Caspian region became isolated from the global ocean (e.g., van Couvering & Miller, 1971), whereas to the south, in Eastern Anatolia, the end of marine sedimentation between ca. 13 and 11 Ma attests to the final closure of the Mediterranean-Indian Ocean gateway and the start of surface uplift of the Eastern Anatolian Plateau (e.g., Gelati, 1975; Şengör and Kidd, 1979; Şengör et al., 1985; Hüsing et al., 2009).

Assessing the vertical movements of the orogenic plateaus is crucial to quantitatively define some of the active geodynamic processes on the Earth, such as surface response of mantle dynamics, lithospheric loading and unloading, active tectonics, etc. For reconstructing the past vertical movements, high precision estimates have been provided by uplifted, datable marine and terrestrial reference surfaces (Cosentino and Gliozzi, 1988; Merritts and Bull, 1989; Gardner, 1992; Westaway, 1993; Chappell et al., 1996; Abbott et al., 1997; Perg et al., 2001; Zazo et al., 2003; Cosentino et al., 2009; Jordan et al., 2010; Schildgen et al., 2012a). At the southern margin of the Central Anatolian Plateau, Neogene marine deposits

unconformably overlies the basement units of the Central Taurides, up to 2 km of elevation (Şenel, 2002). The age of these marine deposits was classically used to constrain the uplift of the CAP southern margin at the Middle-Late Miocene (Dhont et al., 1999; Özgül, 1976). In fact, in the Neogene marine deposits of the Mut-Ermenek Basin, Bassant et al. (2005) identified the upper part of the marine succession on the east side of the Mut Basin to be Langhian, while Tanar and Gökçen (1990) identified marine deposits as young as Serravallian. According to the age of the marine deposits cropping out in the Ermenek Basin (Başyayla section), a post-Tortonian age was recently suggested for this event (Cosentino et al., 2012; Cipollari et al., 2013). Indeed, the stratigraphy of the subsiding Adana-Cilicia Basin, to the south of the uplifted CAP southern margin, provides evidence of even younger age (end of the Messinian, ca. 5.45 Ma, Cipollari et al., 2012; Faranda et al., 2013; Radeff et al., 2016), for the occurrence in the stratigraphic architecture of the Adana Basin of a clastic formation (Handere Formation, *sensu* Radeff et al., 2017), sourced by the Central Tauride uplifted units, with a maximum thickness of 1.9 km (Radeff et al., 2016b). Moreover, the stratigraphical architecture of the marine deposits capping the CAP southern margin, which shows an unconformity surface within the late Neogene marine succession (Schildgen et al., 2012a; Radeff et al., 2016), was recently used for defining a multi-phased uplift of the CAP southern margin, with a second uplift phase in the early Calabrian (ca. 1.6 Ma) (T. F. Schildgen et al., 2012a).

Field work carried out at the CAP southern margin within this PhD project showed that the Pleistocene marine deposits are more widespread than reported in previously published papers or maps. Indeed, in the area of Gülnar and Tol villages, where the official geological map of Turkey indicates the presence of Middle Miocene deep-water marine marls (e.g., Şenel, 2002), Pleistocene marine deposits extensively crop out at an elevation between 1000 to 1200 m.

The principal aim of this paper is to precisely define the age of the youngest marine deposits capping the Central Taurides (southern Turkey) and, consequently, to better constrain the second uplift phase of the study area, which affected the CAP southern margin after the Messinian salinity crisis of the Mediterranean Basin (Hsü et al., 1973; Ryan, 2009 and references therein). In addition, efforts have been done to reconstruct the late Cenozoic palaeogeography of the CAP southern margin, especially to define changes in palaeodepth through time. Palaeodepth investigations on the youngest Pleistocene marine deposits capping the Central Taurides helped us to reconstruct the sea level position and, consequently, to calculate the long-term uplift rate that affected the CAP southern margin in the late Cenozoic.

1.2 Geology of the CAP southern margin

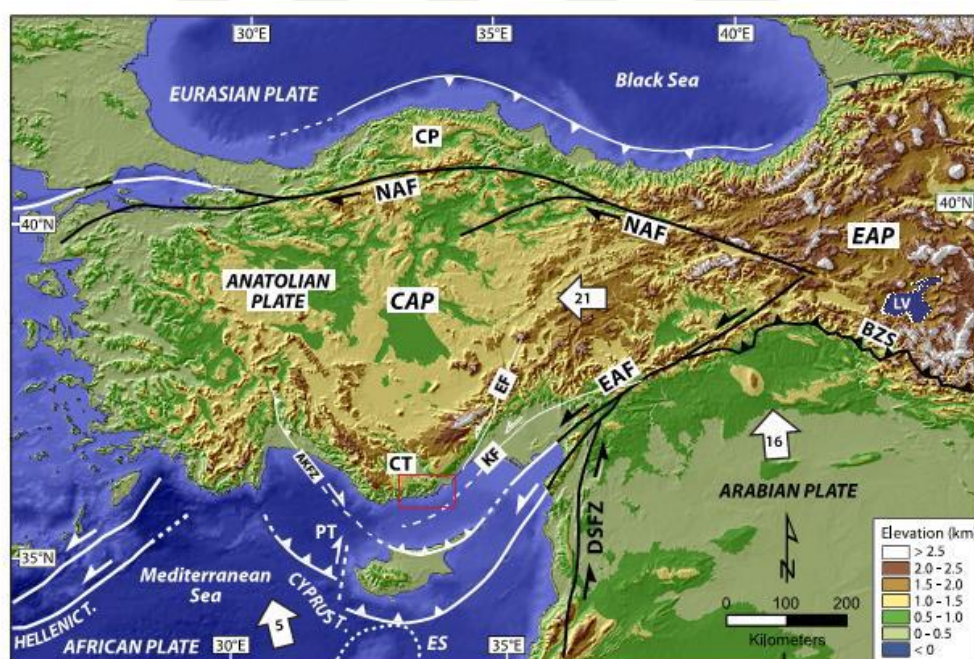


Figure 1.1: Tectonic sketch of the Anatolian microplate showing the location of the study area (red rectangle) rectangle (modified after Schildgen et al. 2014 and Aksu et al. 2014). CAP: Central Anatolian Plateau; EAP: Eastern Anatolian Plateau; CP: Central Pontides; CT: Central Taurides; BZS: Bitlis–Zagros suture; NAF: North Anatolian Fault; EAF: East Anatolian Fault; EF: Ecemiş Fault KF: Kozan Fault; AKFZ: Aksu-Kyrenia Fault Zone; DSFZ: Dead Sea Fault Zone; ES: Eratosthenes Seamount; PT: Paphos Transform; LV: Lake Van. White arrows with numbers inside represent plate movement (mm/yr) with respect to a stable Eurasia (Reilinger et al., 2006).

The Central Taurides constitute the southern Anatolian part of the Alpine-Himalayan orogenic complex, which is characterized by Cambrian to Tertiary stratigraphic units (Özgül, 1976;

Şengör & Yılmaz, 1981). The Tauride belt consists of Neoproterozoic basement that is covered by Paleozoic-Mesozoic platform sediments (Göncüoğlu, 2010). Those units underwent shortening and overthrusting in the late Cretaceous-Eocene time span while the closure of Mesozoic Neotethys Ocean was going on and Tauride carbonate platforms and Central Anatolian Crystalline Complex were colliding (Şengör & Yılmaz, 1981).

The region is composed of 2 autochthonous units (e.g., Geyikdağı and Beydağları) and several allochthonous units (e.g., Bolkardağı, Aladağ, Alanya, Bozkır, and Antalya) (Özgül, 1976, 1984). Those units crop out in the following order, from north to south: Bozkır Unit, Bolkardağı Unit, Aladağ Unit, Geyikdağı Unit, Antalya Unit, and Alanya Unit, (see Fig 1 in Özgül 1976; Özgül 1984; Göncüoğlu 2010). Tectonically, the southern part of the orogen is represented by the Alanya Unit that thrust over the Antalya Unit; whereas, in the northern part, the Bozkır Unit overthrust both the Aladağ and the Bolkardağ units (Özgül, 1984). The rocks of all those continental and oceanic tectonostratigraphic units, which are bounded to the east by the Eceiş Fault Zone, and to the west by the Antalya Basin, were formed during the late Paleozoic-middle Eocene. The orogeny of the Tauride Mountains occurred in the middle Eocene, though the regional uplift of the Anatolian microplate continued through the late Miocene up to the present with ~2 km of uplift at its southern margin. The post-orogenic uplift phase was responsible for the development of mainly extensional intramontane basins such as the Mut Basin (Tanar & Gökçen, 1990; Bassant, 1999; Şafak et al., 2005; Göncüoğlu, 2010). In those intramontane basins, lake and/or fluvial deposits together with shallow-marine limestones were sedimented. From a tectonic point of view, the post-orogenic intramontane deposits at the CAP southern margin are not deformed by compressional tectonics as an evidence of tectonic quiescence (Tanar & Gökçen, 1990; Eriş et al., 2005; Şafak et al., 2005; Cosentino et al., 2012).

From the late Miocene, the tectonic regime of the Anatolian microplate is mainly a

consequence of the collision between Eurasia and Arabian plate, resulting in a westward-tectonic escape of the microplate (Şengör et al., 1985) and in the uplift of the Anatolian plateau.

Synchronously, starting from the middle Miocene (Serravallian), the rising of the Anatolian plateau was accompanied by other events, such as (1) the disconnection between the Indo-Pacific realm and Paratethys, due to the collision between African and European continents, (2) the closure of the Betic and Rifan straits by orogenic processes that brought to the closure of the Atlantic - Mediterranean gateway (Hsü et al., 1977; Kenneth et al., 1978; Krijgsman et al., 1999). As the Atlantic - Mediterranean connection was blocked in the late Messinian, the ocean water inflow to the Mediterranean Basin was firstly extremely reduced and then definitely stopped. As a consequence, the evaporation of the Mediterranean water body led to a deep desiccation of the Mediterranean Basin, which caused the “Messinian Salinity Crisis” (MSC, Hsü et al., 1973). The late Messinian desiccation event of the Mediterranean Basin led to a regional drawdown of the base level, which induced a deep erosion of the basin margins with the development of the Messinian Erosional Surface (MES) all around the Mediterranean Basin. The MES has been recently recognized also within the Messinian deposits of the Adana Basin (southern Turkey) (Cipollari et al., 2013; Faranda et al., 2013; Radeff et al., 2016; 2017), testifying that the MSC deeply impacted the evolution of the Eastern Mediterranean, as far as the Adana-Cilicia and Levantine basins. Deep Messinian incisions (e.g., Messinian canyons) and subsequent Pliocene refilling characterize the Messinian/Zanclean transition at the CAP southern margin (Cipollari et al., 2013; Radeff et al., under review), as well as in the rest of the Mediterranean Basin (e.g., southern Cyprus basins, Rouchy et al., 2001; Italy, Grossi and Gennari, 2008; Cosentino et al., 2013; Spain, Guerra-Merchán et al., 2010).

The youngest marine deposits of the study area (in Gülnar East and Tol districts) pertain

to the Plio-Pleistocene marine transgression of the Mediterranean Sea onto the southern margin of Anatolia, as a consequence of the Zanclean reflooding event (e.g., Cipollari et al., 2012 with references therein). In the literature, sediments pertaining to this sedimentary cycle have been previously signaled within the Mut Basin, along the Goksu River valley (e.g., Yıldız et al., 2003; Schildgen et al., 2012; Radeff et al., under review). In the Gülnar district, sediments pertaining to this sedimentary cycle unconformably onlap the horizontal (undeformed) Miocene limestones and marls of the Mut and Köselerli formations, from ~1 km up to 1.5 km above the present sea level. In the area, the Neogene units were unconformably deposited on the basement rocks of the Geyikdağı Unit (Fig. 1.2).

1.3 Post-Miocene evolution of the southern flank of the CAP

The post-Miocene evolution of the CAP southern margin has not been studied with high resolution data yet. In the previous studies of the last decade, it was reported that in the Eastern Mediterranean, as well as in the rest of the Mediterranean realm, the MSC ended with the Zanclean flooding event, a Pliocene episode also related to a shift to wetter climates in the Mediterranean realm (Griffin, 2002; Cosentino et al., 2010; Cipollari et al., 2012). In the Mut Basin, Pliocene marine sediments started to deposit following this climate shift (Kelling et al., 2005; Cosentino et al., 2012; Schildgen et al., 2012b). Based on micropaleontological data, previous authors reported Pliocene deposits in Mut and Adana basins (e.g., Şafak, 1997; Nazik, 2004) even though some of them are stratigraphically still controversial (Yıldız et al., 2003). However, the complete stratigraphy of the region is not completely covered yet. Recently, Radeff et al. (under review) suggested to include the Plio-Pleistocene deposits of the Mut Basin in a stratigraphic unit called Sarıkavak Formation (Fig. 1.2).

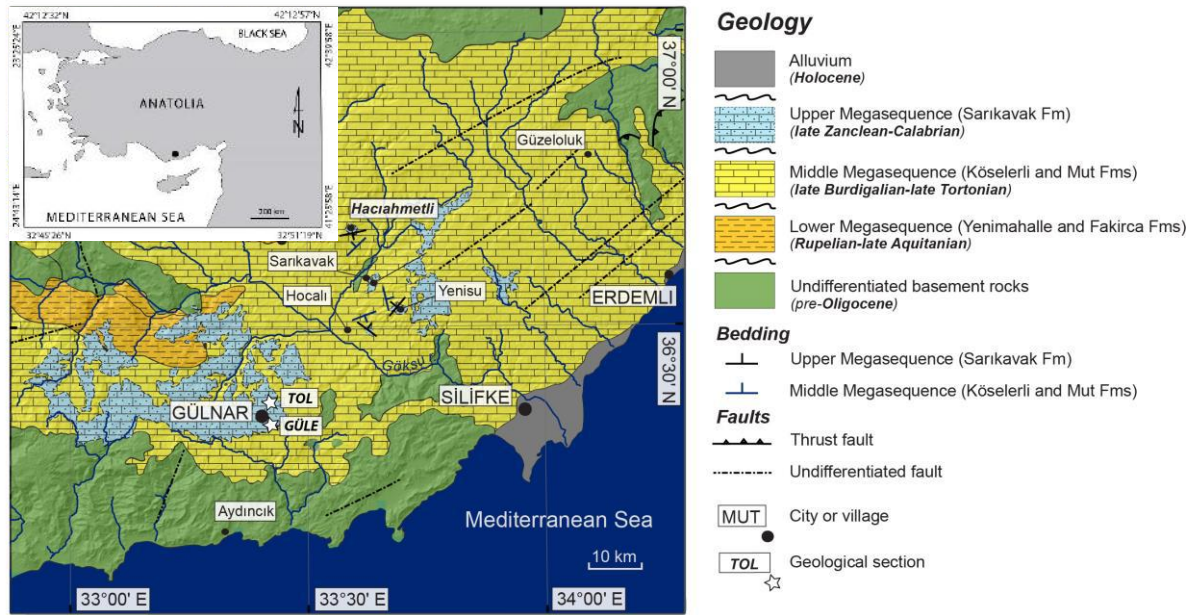


Figure 1.2: Map showing the location of Plio-Pleistocene deposits in the study area (modified after Radeff et al, under review).

Aksu et al. (2014), using the offshore seismic data of the Adana-Cilicia Basin evidenced the occurrence of Plio-Quaternary extensional tectonics at the southeastern margin of the CAP. According to this study, an ENE-WSW extensional fault system (the “Kozan Fault System”) is affecting the marine deposits of the CAP southern margin, cutting through upper Quaternary and Holocene deposits.

According to some authors (Robertson et al., 1996; Schattner, 2010; Schildgen et al., 2014b), the collision of the Eratosthenes Seamount with the Cyprus Arc would have played a role in the Plio-Pleistocene geomorphological evolution of the CAP southern margin.

1.4 Material and methods

In the Gülnar district, deposits pertaining to the Sarikavak Formation unconformably lie onto middle Miocene shallow-water limestones (Mut Formation). Both formations show horizontal bedding in the whole area, showing no major deformation in comparison with the pre-Miocene tectono-sedimentary units. Only a few high-angle normal faults cutting the

Sarıkavak Formation have been observed in the sampled sites. In the study area, the boundary between Mut and Sarıkavak formations is usually marked with lag deposits, which lie on top of an erosional surface cut within the middle Miocene limestones of the Mut Formation. The erosional surface at the base of the Sarıkavak Formation is well exposed in the Gülnar area at around 1000 m a.s.l.

In the study area, some randomly distributed samples were collected from the Sarıkavak Formation, together with three continuously sampled long sections: Gülnar East, Tol, and Tol1 sections (Fig. 1.2). The 27 m-thick Gülnar East section and the 18.5 m-thick Tol section were sampled approximately every 50 cm and, in total, 54 and 26 samples, respectively, were collected and analyzed for fossil content. From the Tol1 section, 71.50 m-thick, 68 samples were collected approximately every 1 m. Unfortunately, the fossil content and the biostratigraphical results of Tol1 section are not completed yet. Thus, in the “Results” section of this PhD thesis only the biostratigraphical and palaeoecological results from Gülnar East and Tol sections will be presented.

For both Gülnar East and Tol sections (Fig. 1.3), as well as for the randomly distributed samples, micropaleontological analyses were carried out on benthic and planktonic foraminifers, ostracods, and calcareous nannofossils.

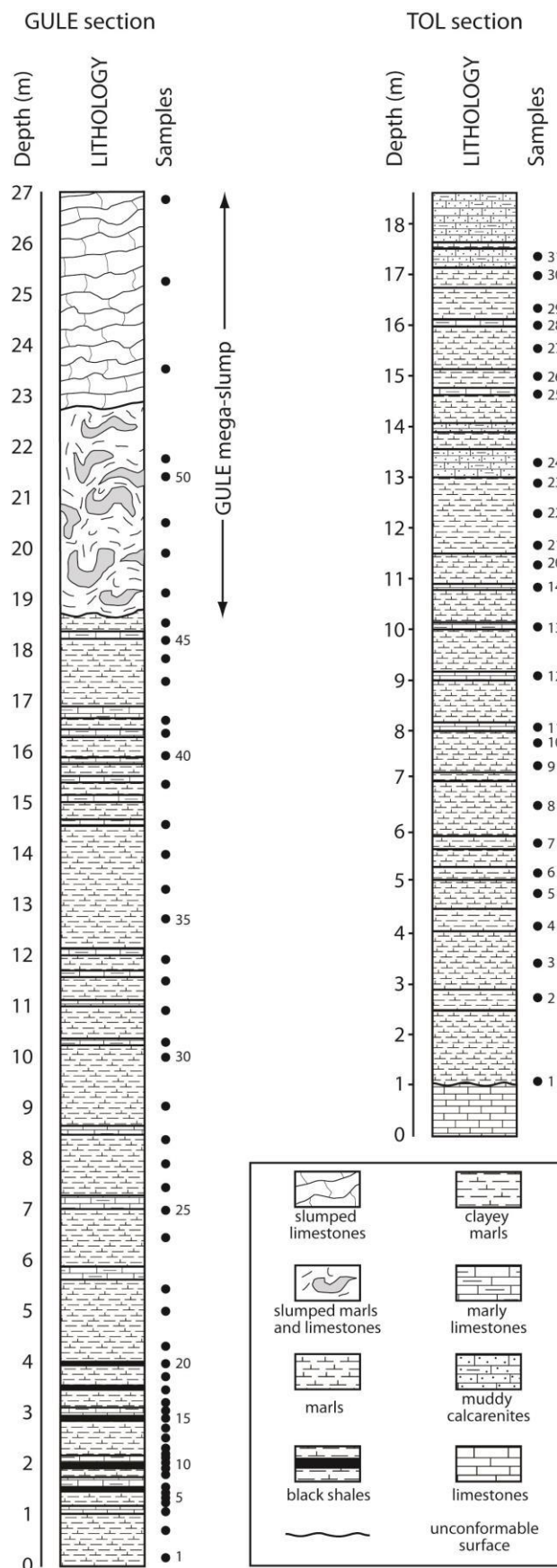


Figure 1.3: Stratigraphic logs of the GULE and TOL sections. The stratigraphic position of the collected samples for micropaleontologic analyses is also shown. See the main text for a complete description of the study sections.

For the microfaunistic studies (both foraminifers and ostracods), all samples (except for the uppermost Gülnar East samples 52, 53 and 54) were disaggregated in a H₂O₂ 5% solution for 24-48 h, washed through 63 µm- and 125 µm-mesh sieves, and dried in oven at 40°C. Quantitative and qualitative benthic and planktonic foraminifer and ostracod analyses were carried out by counting, when possible, up to 300 specimens from the dry residue fraction >125 µm. Frequency matrices were obtained normalizing the picked foraminifers to 1 g of sieved sample and the picked ostracods to 10 g of sieved sample. Samples GULE 52, 53 and 54, which are from the cemented calcarenite at the top-section, were analyzed on thin sections.

The taxonomic identification of planktonic foraminifera was based on Parker (1962), Kennett and Srinivasan (1983), and Iaccarino and Premoli Silva (2007). Benthic foraminifer classification was based on AGIP (1982) and Meriç et al. (2014).

Biostratigraphic considerations followed the planktonic foraminifer biozones of Iaccarino and Premoli Silva (2007) and the Global Time Scale 2012 edited by Anthonissen and Ogg (2012). Paleoclimate analysis on the studied samples followed Amore et al. (2004) and Stefanelli and Capotondi (2013). Paleodepth calculations were done using Hohenegger (2005) methodology and using Fisher α -Shannon (H) diversity calculations on PAST software paleo-depositional environment identification were applied (Hammer et al., 2001; Murray, 2006). Oxygen index study following Kaiho (1994) on benthic foraminifera was applied to highlight the sapropel layers of the study sections.

For the calcareous nannofossil analyses, smear slides were prepared from raw samples using standard techniques (Bown and Young, 1998). Optical adhesive (Norland #61) was used as a mounting medium for all of the smear slides, which were exsiccated under ultraviolet light. The preservation degree and total abundance of the assemblages were determined. The estimates of the individual species abundances were given following letter

codes: R= rare (1 specimen per 51-200 fields of view); F= few (1 specimen per 11-50 fields of view); C= common (1 specimen per 2-10 fields of view); A= abundant (1-10 specimens per field of view). The biostratigraphic scheme used in this paper is referred to Raffi et al. (2006). The taxonomic concepts of gephyrocapsids vary according to different authors (Flores et al., 1999; Raffi, 2002). Here, the Rio et al. (1990) criterion, based on biometric features, for distinguishing gephyrocapsids has been adopted, that is: gephyrocapsids $<3.5\ \mu\text{m}$ in size were labeled "small Gephyrocapsa", gephyrocapsids $>4\ \mu\text{m}$ and $<5.5\ \mu\text{m}$ in size "medium Gephyrocapsa", and gephyrocapsids $>5.5\ \mu\text{m}$ in size "large Gephyrocapsa".

The results from both the biostratigraphic analysis and the palaeoecological investigation have been used to better constrain the age of the youngest marine deposits capping the CAP southern margin, to define the palaeodepth of deposition, and, consequently, to calculate the uplift rate of the study area. In addition, the sedimentation rate for the Pleistocene succession has been calculated and compared with the results of Vrica and Singa sections (Hilgen, 1991; Lourens et al., 1998; Cita et al., 2008), and ODP Site 967 (Lourens et al., 1998).

1.5 Stratigraphy of the sampled sections

The Gülnar East (GULE) section (base of the section: $36^{\circ} 20' 8.26''$ N; $33^{\circ} 25' 51.44''$ E) crops out along a new road cut to the east of the Gülnar village, on the road to Silifke, with a bedding attitude of $N322^{\circ}, 10^{\circ}$ (Fig. 1.4). The section, 27 m-thick, is mainly characterized by fine-grained deposits (marly clays and clayey marls) in the lower and middle part, whereas in the uppermost part of the section an 8.5 m-thick horizon shows clear evidence of slumping. The slumped horizon is made of 3.5 m of clayey marls, containing laterally-discontinuous-and-folded carbonate layers, and 5 m of carbonate deposits at the top of the section (top of the section: $36^{\circ} 20' 10.9''$ N; $33^{\circ} 25' 51.9''$ E).



Figure 1.4: Panoramic view of the Gülnar East section. The lowermost part of the section is characterized by the occurrence of black shales referable to sapropel layers. At the top of the section, carbonate layers are parts of a slumped horizon.

In the lowermost part of the GULE section, five dm-thick black horizons, rich in organic matter (black shales/ sapropel layers), have been distinguished from the quasi-homogeneous light-gray color that characterizes the rest of the section. From field observations, the fossil content of the GULE section consists of rich planktonic foraminifera assemblages, containing large *Orbulina universa* throughout the section (Fig. 1.5), rare thin shells of mollusk (mainly gastropods), shark teeth, and plant remains in the middle/upper part of the section. Both lithology and fossil content from the GULE section point to a deep-water marine environment.

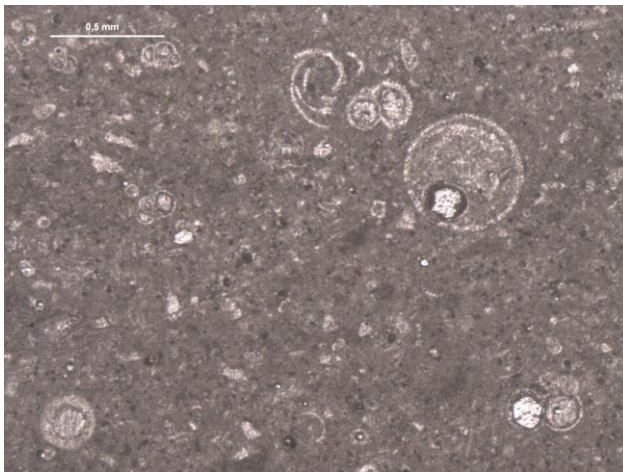


Figure 1.5: Microphotograph of a thin section from sample GULE 47. The fossil assemblage is mainly characterized by planktonic forams, with large *Orbulina universa*, radiolarians and sponge spicules, which points to a pelagic environment.

On the contrary, the carbonate layers within the slumped horizon (mega-slump), especially the topmost carbonates of the section, contain both benthic (among the others *Amphistegina* sp.) and planktonic forams, *Ditrupa* sp., and coralline red algae (*Lithothamnium*

spp.) (Fig. 1.6). Thus, the fossil assemblage within the topmost part of the GULE section, which it is worth to note they come from a slumped horizon, point to a shallower marine environment referable to a distal carbonate ramp, no more than a few tens of meters depth (40-50 m) (outer ramp *sensu* Civitelli and Brandano, 2005). The mega-slump of the GULE section, which could indicate instability in the sedimentary basin possibly due to tectonics (i.e., tectonic uplift of the plateau margin), characterizes all the youngest marine deposits that crop out east of the Gülnar village, and is well visible along the cut of the road to Silifke (Fig. 1.7).

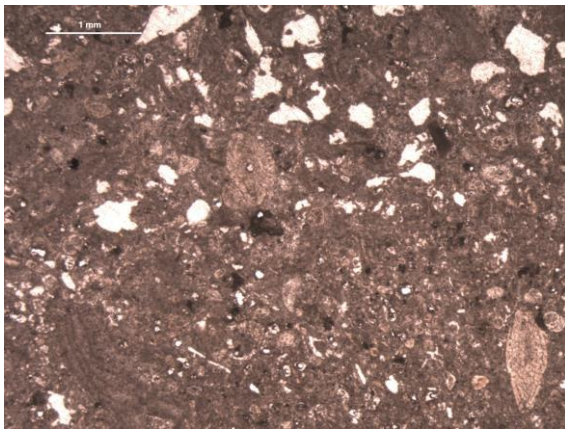


Figure 1.6: Microphotograph of a thin section from sample GULE 52. The fossil assemblage is mainly characterized by planktonic forams, with globigerinids and large *Orbulina universa*, benthic forams (*Amphistegina* sp.), and coralline red algae (*Lithothamnium* sp.), which points to a shallow marine environment (outer carbonate ramp, *sensu* Civitelli and Brandano, 2005)



Figure 1.7: Mega-slump structure affecting the youngest marine deposits cropping out throughout the area to the east of the Gülnar village. The picture is from the road to Silifke.

The Tol (TOL) section (base of the section: 36° 23' 05.9" N; 33° 25' 25.9" E; 1087 m elev.) crops out at the base of the eastern flank of the Kepezçalı T. (1229 m), ca. 600 meters NE of the Tol village. The TOL section starts on top of an erosional surface that cuts Middle Miocene shallow-water limestones of the Mut Formation. The section, 18.5 m-thick, which shows some carbonate layers in the upper part, is mainly characterized by clayey marls and

marly clays, with a few dark horizons of bioturbated marly clays rich in planktonic foraminifers and plant remains (Fig. 1.8). The bedding attitude of the limestones at the base of the TOL section is N 198°, 10°. From field observations, the fossil content of the TOL section consists of rich planktonic foraminifera assemblages, containing large *Orbulina universa*, benthic foraminifers, and thin shells of mollusks (mainly bivalves). As for GULE section, both lithology and fossil content from the lower-middle part of the TOL section point to a deep-water marine environment. The topmost part of the TOL section show carbonate layers, which are mainly biocalcarenites from shallower-water environments (top of the section: 36° 23' 07.9" N; 33° 25' 22.9" E; 1090 m elev.).



Figure 1.8: Panoramic view of the TOL section, which crops out at the base of the eastern flank of the Kepezçali T. (1229 m).

1.6 Results

1.6.1 Gülnar East section

Planktonic and benthic foraminifers, ostracods and calcareous nannofossil analyses were performed on 51 samples from Gülnar East section, with the aim to constrain the age of the succession and reconstruct the paleoenvironmental conditions of the sedimentary basin. Microfossils were recovered throughout almost all the sedimentary succession from 0.3 m up to 21.9 m (only few samples were barren), with fluctuating abundances (Fig. 1.9) and different state of preservation, from poorly preserved (incrusted, abraded or strongly recrystallized), to well preserved. All samples seems to be affected by detrital dilution and

reworking. For example, among nannofossils the co-presence of *Sphenolithus abies*, whose highest occurrence is referred to Pliocene (3,7 Ma, Lourens et al., 1996), with Pleistocene species such as medium *Gephyrocapsa* points out to the presence of huge reworking, as well as the presence of *Helicosphaera sellii* (its highest occurrence in eastern Mediterranean is at 1,245 Ma, (Lourens et al., 1996, 1998) and *Calcidiscus macintyreii* (highest occurrence at 1,619 Ma; de Kaenel et al., 1999) in samples above those containing large *Gephyrocapsa*. Finally, also the great abundance of long ranging specimens in most of the samples, can be related to reworking that apparently increases their abundance.

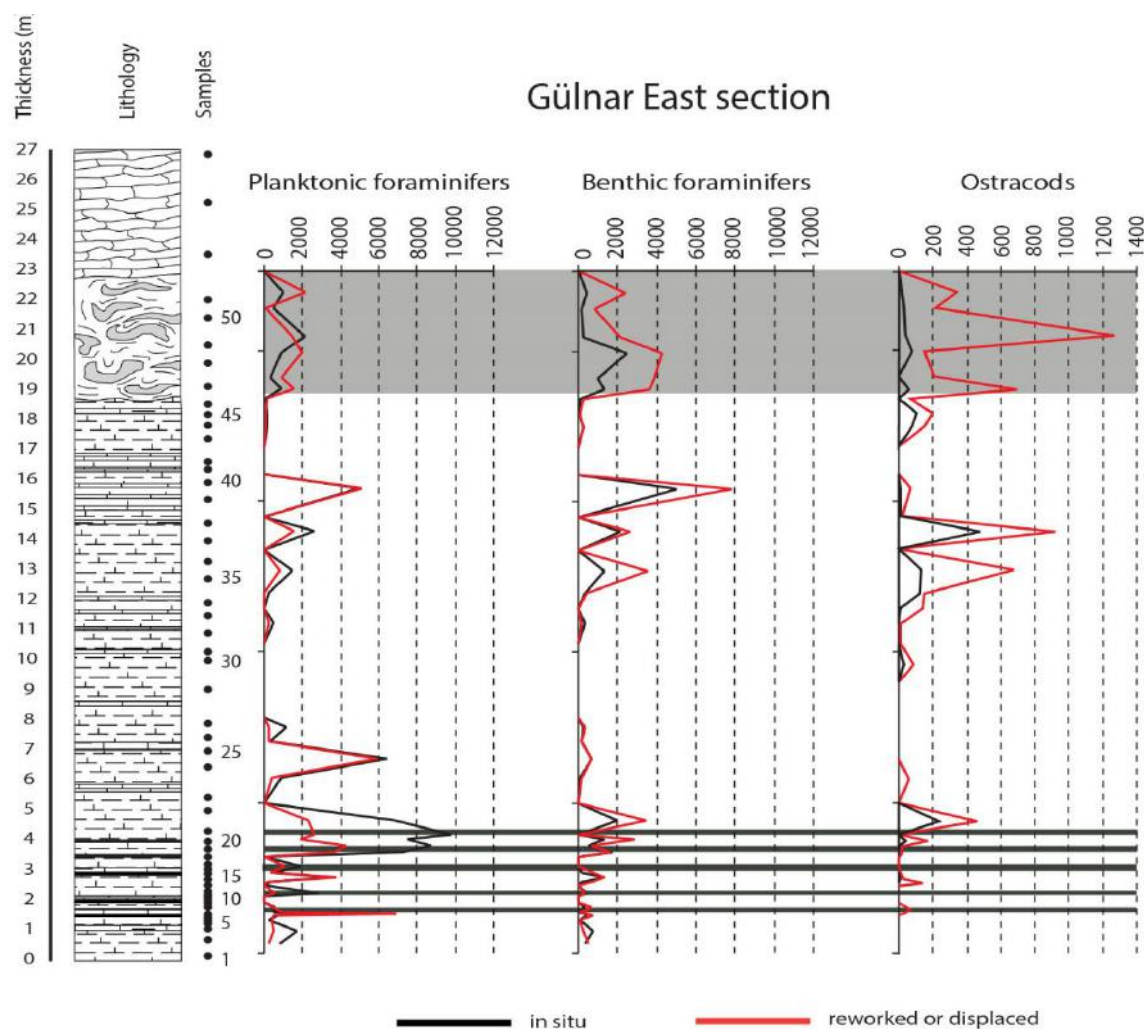


Figure 1.9: Normalized frequencies of planktonic and benthic foraminifers and ostracods from Gülnar East section. The grey lines correspond to the five sapropel layers. Legend: dotted lines=displaced or reworked contingent; black lines=autochthonous contingent. Dark grey lines=sapropel layers.

Also among planktonic foraminifers, Miocene species such as *Globigerinoides bulloideus*, *G. subquadratus*, and *G. altiaperturus*, were found together with Plio-Pleistocene *G. ruber*, *Globorotalia crassaformis*, *Neogloboquadrina pachyderma* (sx), *N. dutertrei* and *Globoturborotalita rubescens* (Tab. 3, supplementary material). Even among benthic microfossils it is possible to distinguish two different contingents. The first one includes poorly preserved specimens referable to infra-circalittoral species (Fig. 1.9; Tabs. 2 supplementary material). Ostracods of this group are referable to species represented mainly by instars and very few adults. Such assemblage is attributable to the allochthonous component of a low energy thanatocoenosis (Boomer et al., 2003). The reworked or displaced status of this contingent is definitely strengthened by the stratigraphic distribution of the species that shows a mixing of species disappeared during late Miocene or appeared during Pliocene. The second contingent is made of rather well or well preserved deep marine dweller species, and in this case, the ostracod remains include both several adults and instars. According to Boomer et al. (2003) this population age structure is referable to the autochthonous component of a moderate energy thanatocoenosis. As a consequence, the first contingent of benthic microfossils includes reworked (or at least displaced) species. Conversely, the second contingent, inhabiting deeper waters, must be considered autochthonous (Fig. 1.9; Tab. 5 supplementary material).

The following detailed results are limited to the autochthonous contingents recognized among calcareous nannofossils, planktonic and benthic foraminifers, and ostracods.

The nannoflora of the entire section (Fig. 1.10) is dominated by the presence of *Umbilicosphaera jafari*, which is considered capable of opportunistic behavior (Wade and Bown, 2006), *Reticulofenestra* spp. (group including all the species of *Reticulofenestra* smaller than 7 micron) and *Dictyococcites* spp. *S. abies* is present in nearly all the samples as well as *Helicosphaera carteri*, *Coccolithus pelagicus* and *Pontosphaera* spp. On the contrary,

marker species, as small *Gephyrocapsa*, medium *Gephyrocapsa* and large *Gephyrocapsa*, are rare and scattered along the section, preventing the precise definition of their Lowest Occurrence (LO). More precisely, medium *Gephyrocapsa* is present since the lowermost sample of Gülnar East section and its last occurrence has been detected at about 3 m from the bottom. Large *Gephyrocapsa* has been detected only in two samples respectively located at 4 and 5 meters from the base of the section. Finally, *C. macintyreii* and *H. sellii*, even if discontinuously, are present in the entire section.

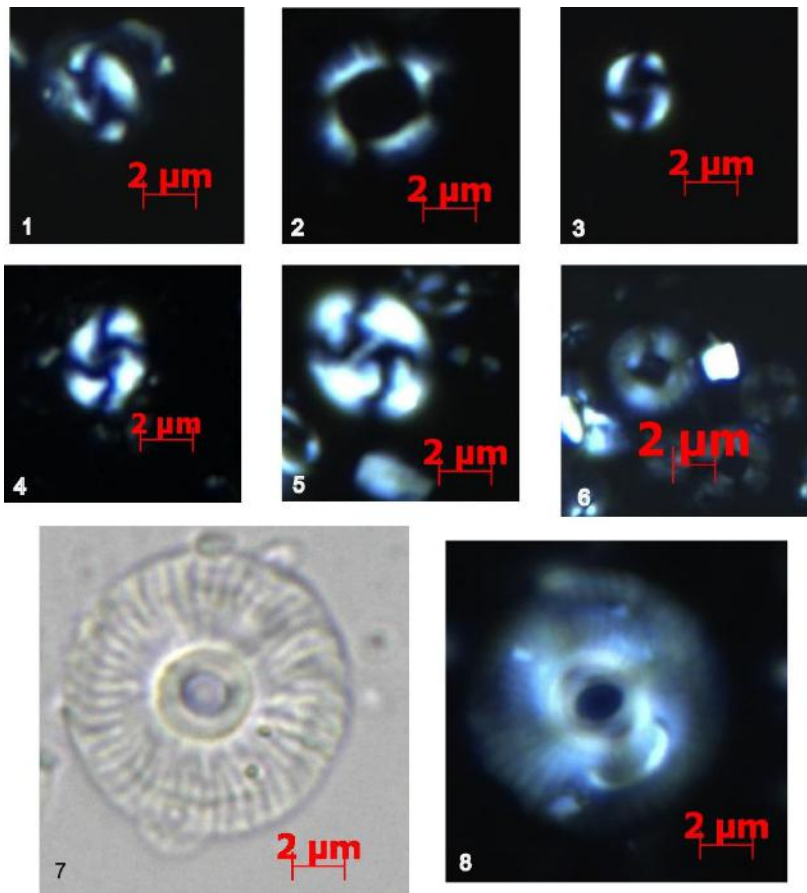
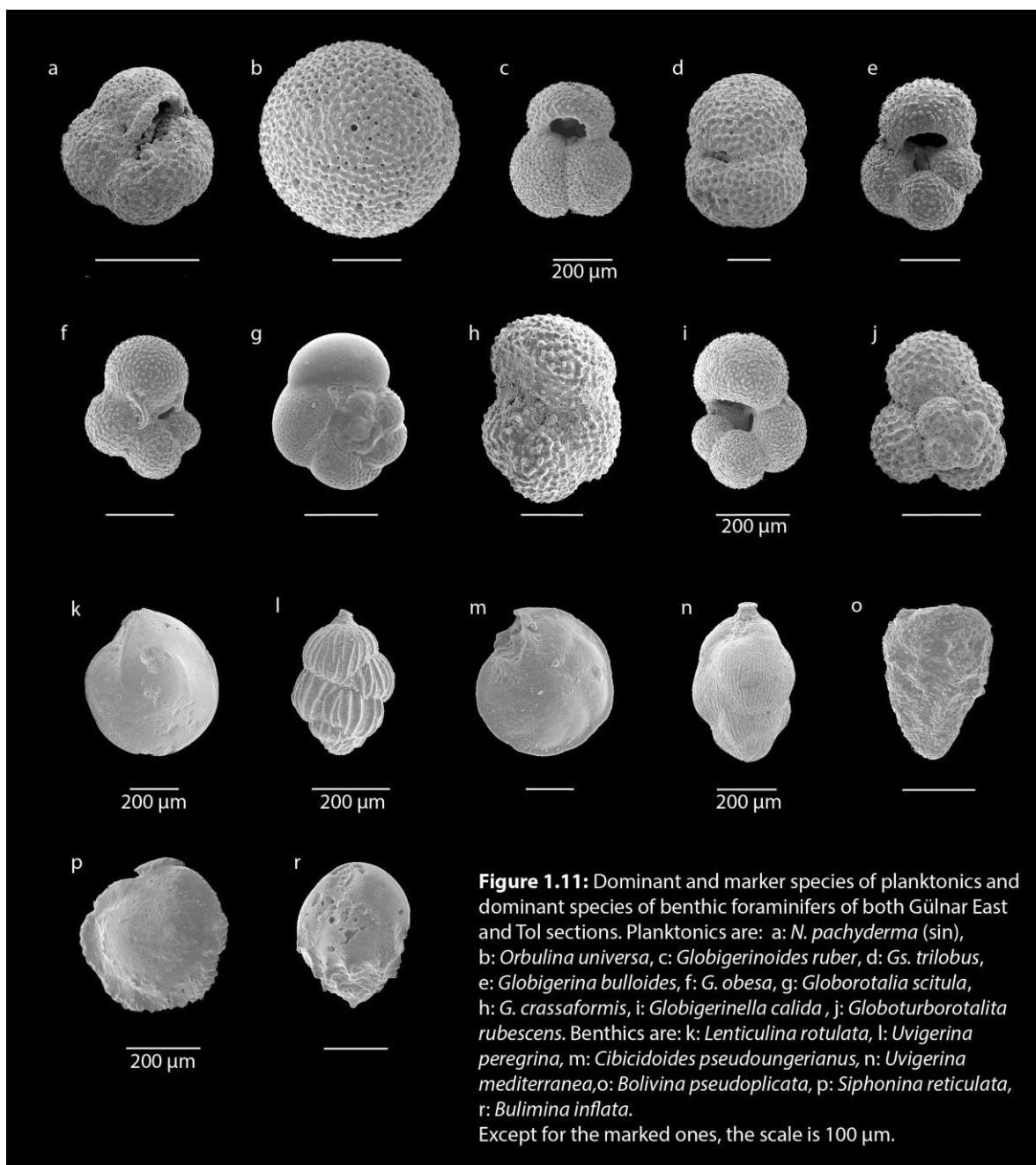


Figure 1.10: Nannoflora of Gülnar East section. 1. *Helicosphaera sellii*; 2. *Pseudoemiliana lacunosa*; 3. small *Gephyrocapsa*; 4. medium *Gephyrocapsa*; 5. large *Gephyrocapsa*; 6. *Umbilicosphaera jafarii*; 7. *Calcidiscus macintyreii* (tr); 8. *Calcidiscus macintyreii* (pol).

Autochthonous planktonic foraminifer assemblages (Figs. 1.11, 1.12) are characterized by the high abundance of *Orbulina universa*, *Globigerinoides trilobus*, *G. ruber*, *Globigerinella obesa*, *G. siphonifera*, *Globorotalia scitula*, *Globigerina bulloides* and *G. falconensis* (Fig.1.11). Other species, such as *Globorotalia crassaformis*, *N. acostaensis*, *N. pachyderma*, *N. dutertrei*, and *Globoturborotalita rubescens*, are present scattered and with lower

abundances. Nevertheless, some of them have a high biostratigraphical importance (Tab. 1, supplementary material). *Neogloboquadrina* spp. (sin) show an irregular pattern with high abundances from 0.3 to 7.5 m, and with scattered distribution from 10.95 to 17.45. *N. pachyderma* (sin) is present from the bottom and shows some peaks between 2.35 and 3.75 m. Afterwards, it disappears and appears again showing two more peaks at 15.40 and 17.45 m (Fig. 1.12). *G. crassaformis* is present with low percentages from 0.30 to 0.75 m, at 1.7, 2.0, 7.05 to 7.50 and 11.95 m.



Autochthonous benthic foraminifer assemblages are present almost in all samples along the section, characterized by relatively high abundances of species typical of deep marine environment such as *Bolivina pseudoplicata*, *Cibicidoides pseudoungerianus*, *Lenticulina rotulata*, *Trifarina bradyi* and *Uvigerina* spp. In the first two sapropel layers and in the fourth one the assemblages are less diversified and mainly made by *Uvigerina* spp. and *Bolivina* spp. that are in concordance with the study of Sen Gupta and Machain-Castillo (1993), while sapropel layers three and five are barren. The abundances of the most significative species are reported in Fig. 1.12 and Fig 1.13 for planktonic and benthic foraminifers, respectively.

The planktonic foraminifer tests are generally more abundant than benthic ones, in some cases making up to 100% of the foraminiferal assemblage, at 2.9 and 3.95 m along the section. In general, the planktonic/total foraminifer ratio (Fig. 1.12) shows an irregular trend, but three different intervals can be detected: 1) a first interval from the bottom to 7.5 m where the P/P+B ratio is always higher than 60%, 2) a second one between 10.95 to 15.4 m in which it drops between 50 and 60%, and, finally, a third one from 17.9 m to the top in which it increases again up to 90%, except for samples at 19.1 and 19.95 m where it reaches its minimum values (around 25%). The P/P+B values further support the deposition of the Gülnar East sediments in deep marine conditions.

The deep marine ostracod contingent is represented by 15 species (2 left in open nomenclature) referable to 11 genera (Fig. 1.14, Tab.5, supplementary material). None of them were recovered in the sapropel layers. Frequencies are not very high (Fig. 1.15), but some species, such as *Bairdoppilata profunda*, *B. producta*, *Cytherella* spp., *Henryhowella sarsii profunda*, and *Parakrithe* spp. are rather well represented in several samples. The ostracod assemblages generally show low diversity (on average 2-6 species, except for sample at 14.0 m that reaches 11 species).

Gülınar East section

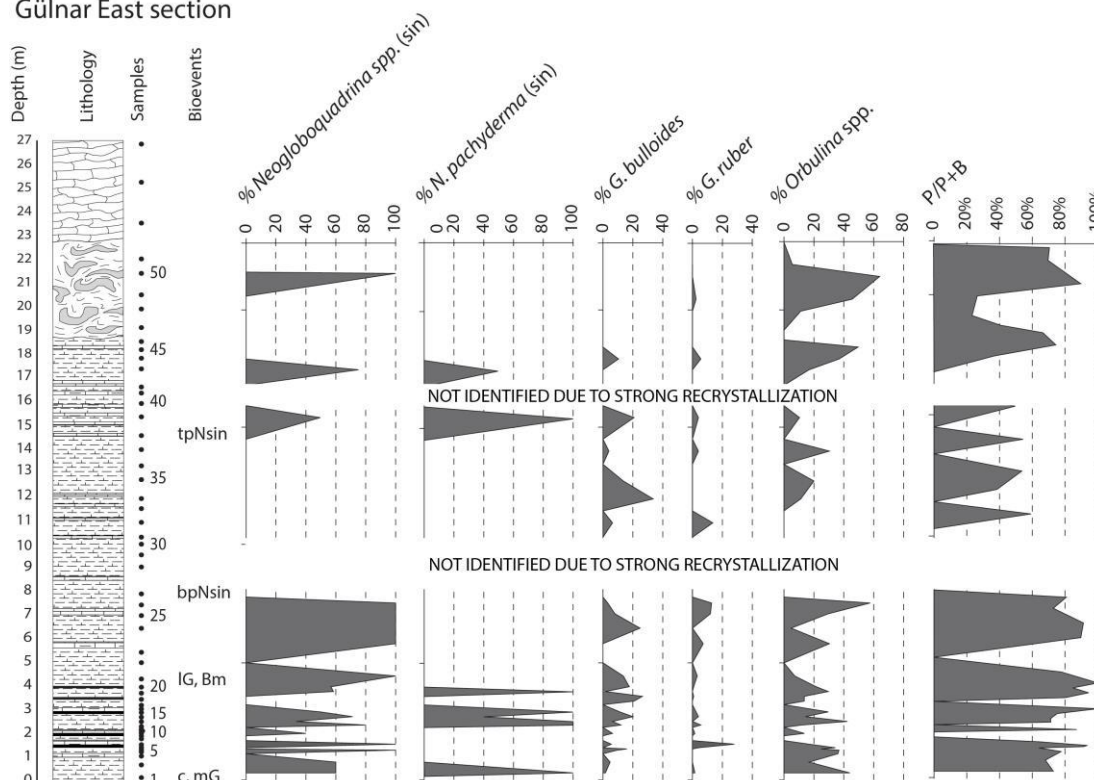


Figure 1.12: Normalized abundance frequencies of the most significant planktonic foraminifera of the Gülınar East section.

Gülınar East section

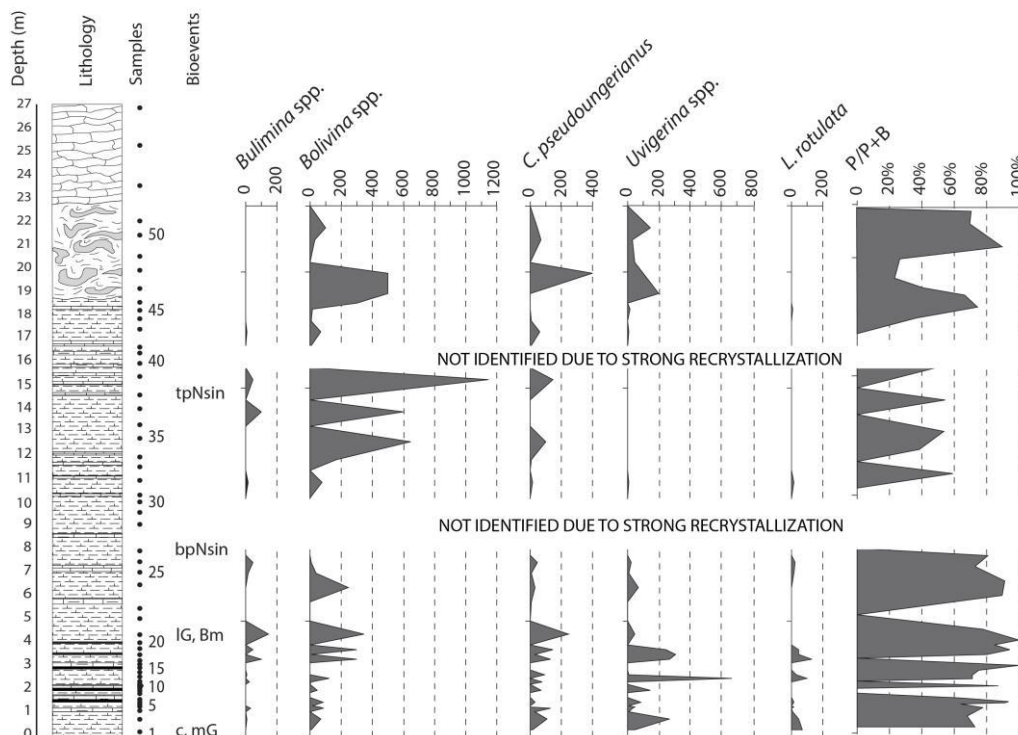


Figure 1.13: Normalized abundance frequencies of the most significant benthic foraminifera of the Gülınar East section.

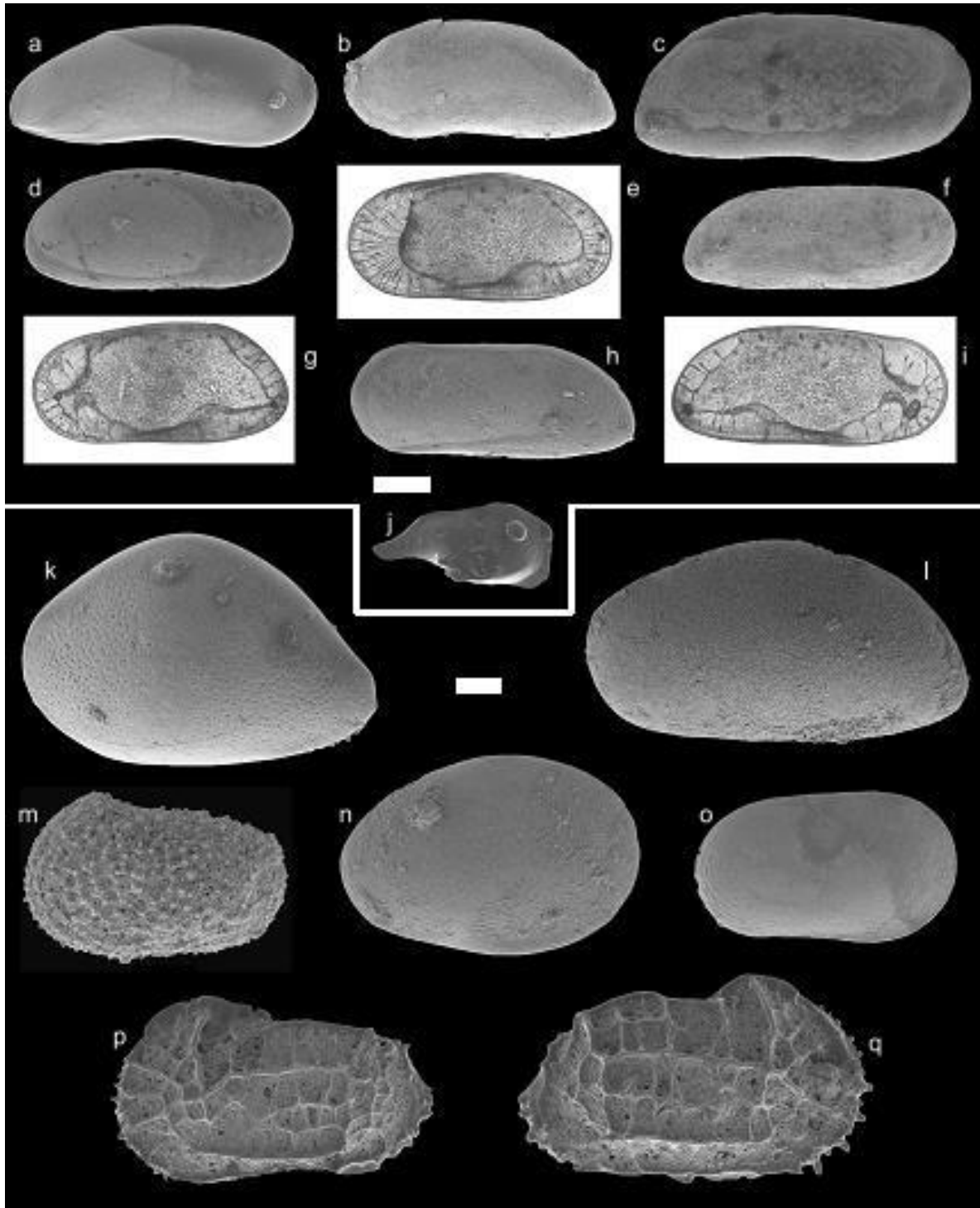


Figure 1.14: SEM and transmitted light pictures of the autochthonous epibathyal ostracod contingent from Gülnar East section. a. *Argilloecia acuminata*, RV, external view, sample 21; b. *Argilloecia fatua*, VS, external view, sample 44; c. *Krithe iniqua*, RV, external view, sample 37; d. *Parakrithe rotundata*, RV, external view, sample 44; e. *Parakrithe rotundata*, LV, external view, sample 48; f. *Parakrithe lamellosa*, RV, external view, sample 37; g. *Parakrithe ariminensis*, LV, external view, sample 48; h. *Parakrithe ariminensis*, LV, external view, sample 44; i. *Parakrithe ariminensis*, RV, external view, sample 48; j. *Paijenborchella iocosa*, RV, external view, sample 35; k. *Bairdoppilata profunda*, LV, external view, sample 44; l. *Bithocypris producta*, LV, external view, sample 51; m. *Henryhowella sarsii profunda*, VS, external view, sample 43; n. *Cytherella russoi*, RV, external view, sample 37; o. *Cytherella robusta*, RV, external view, sample 13; p. *Oblitacythereis mediterranea*, VS, external view, sample 21; q. *Oblitacythereis mediterranea*, VD, external view, sample 21. Legend: VD=right valve; VS=left valve. White bars correspond to 0.1 mm

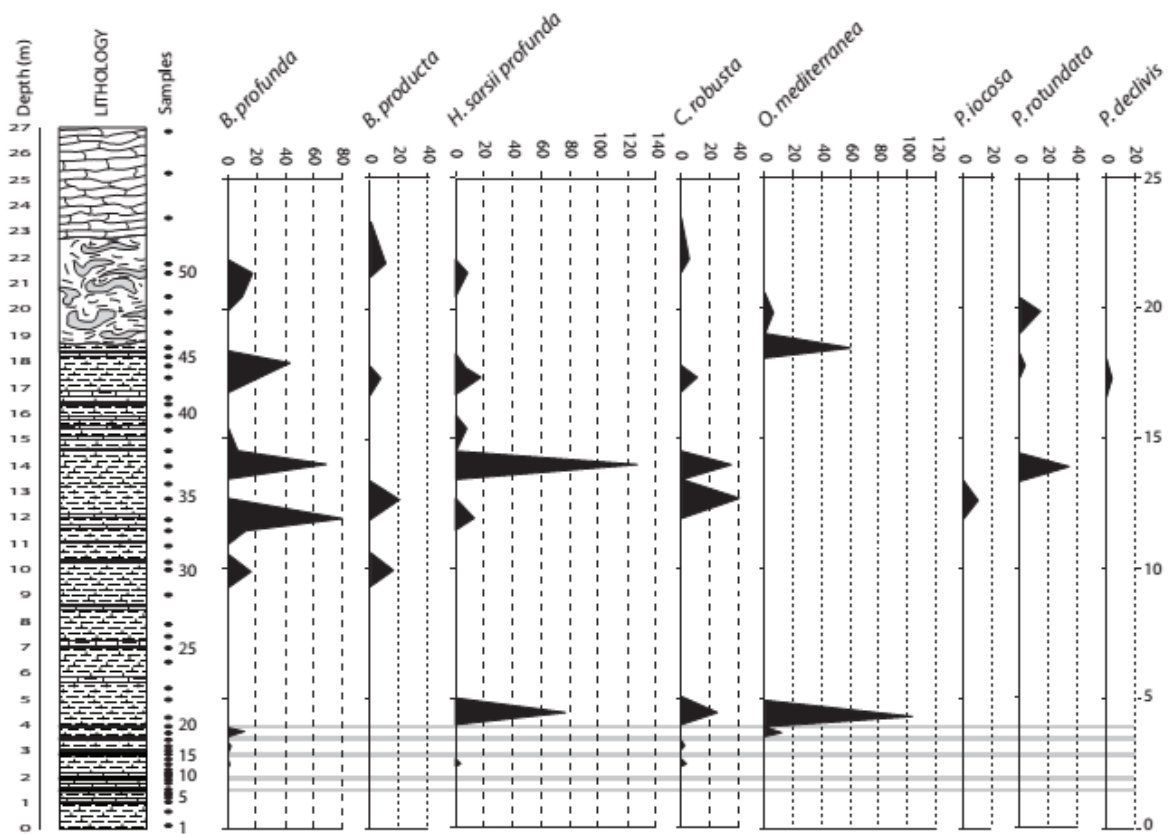


Figure 1.15: Normalized frequencies of selected true epibathyal ostracod species from Gülnar East section. The grey lines correspond to the five sapropel layers, barren of ostracods.

1.6.2 Tol section

Microfaunas and nannofloras analyses were performed on 26 samples from Tol section. They were recovered throughout almost all the sedimentary succession from 1.0 m up to 16.4 m (only few samples were barren), with fluctuating abundances. On the whole the microfossils are rather well to well preserved (only one sample at 4.75 m bore fully re-crystallized specimens) and reworking/displacing is less strong than in Gülnar East section (Fig. 1.16).

The preliminary results on the nannoflora of Tol section indicate Pleistocene ages however, this study is still ongoing to refine the preliminary results.

Even in Tol section the planktonic foraminifer assemblages (see Fig 1.11) are dominated by *Orbulina* spp., *Globigerinoides trilobus*, *G. ruber*, *Globigerinella obesa*, *Globigerina bulloides* and *Globigerinita parkerae* (Fig. 1.17). Other species, such as

Globigerinoides tenellus, *G. elongatus*, *G. conglobatus*, *Globoturborotalita rubescens*, and *Globigerinella calida* are present scattered and with lower abundances. No *Globorotalia crassaformis* and *Neogloboquadrina* spp. are present in the analyzed samples. It is worth to note that in Tol section the specimens referred to *G. calida* corresponds to the original description and illustrations of Parker (1962) and Kennet and Srinivasan (1983).

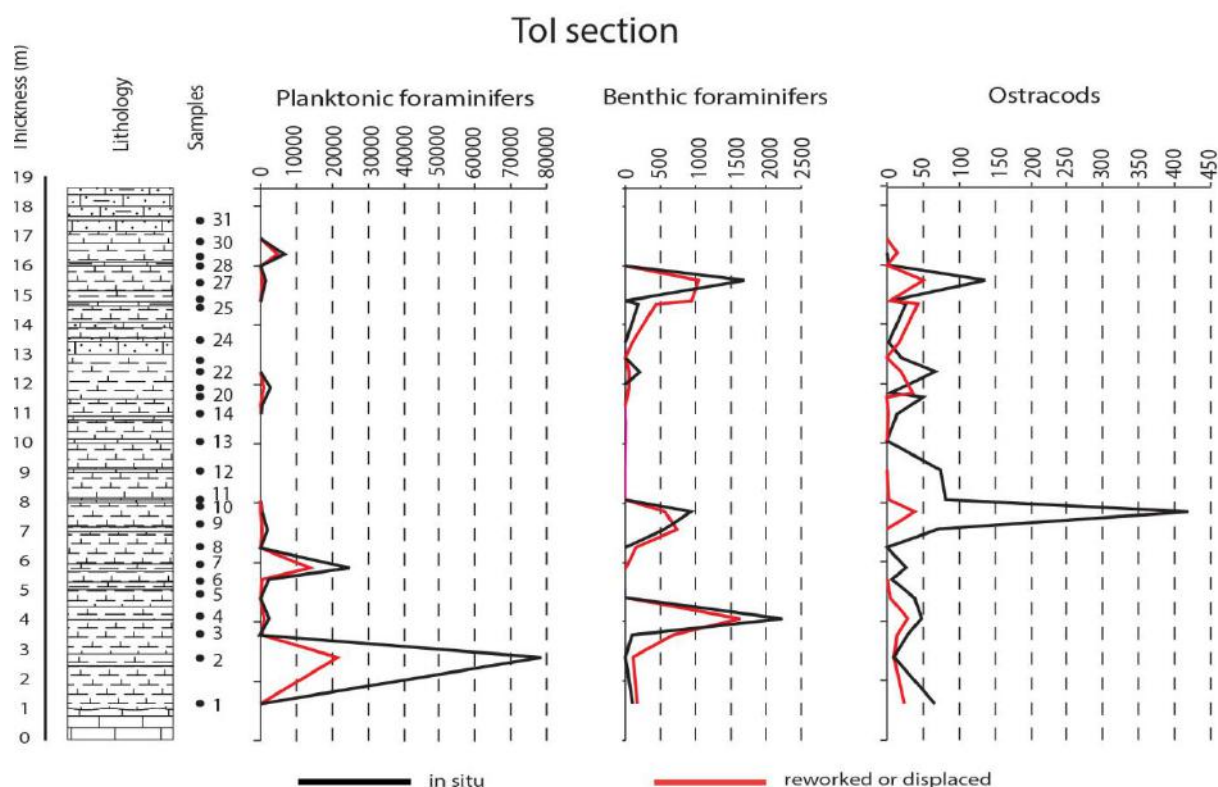


Figure 1.16: Normalized frequencies of planktonic and benthic foraminifers and ostracods from Tol section. Legend: red lines=displaced or reworked contingent; black lines=autochthonous contingent.

Autochthonous benthic foraminifer assemblages are rather similar to those of Gülnar East section, still pointing to the deposition in a deep marine environment (Tab. 4, supplementary material). *Lenticulina* spp. is very abundant, followed by *Siphonina reticulata*, *Cibicidoides pseudoungerianus*, *Bolivina* spp., *Praeglobobulimina pupoides*, *Uvigerina* spp., and *Stilostomella monilis* (Fig. 1.18).

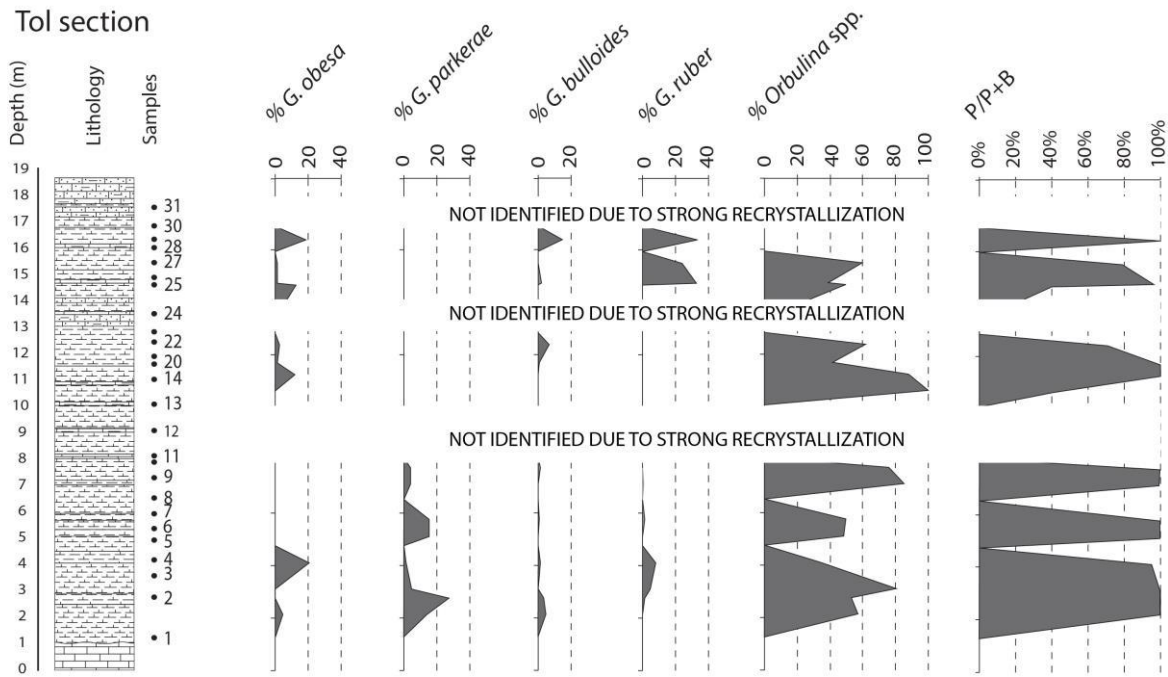


Figure 1.17: Normalized abundance frequencies of the most significant planktonic foraminifera of the Tol section.

The planktonic foraminifer tests are generally more abundant than benthic ones, in most cases making up to 100% of the foraminiferal assemblage, at 2.75, 5.10, 5.75, 11.25, 11.70 and 14.75 m along the section. In general, the planktonic/total foraminifer ratio (Fig. 1.17 and 1.18) shows an irregular trend, but two different intervals can be detected: (1) a first interval at 12.40 and 15.50 m where the P/P+B ratio is between 70-80%, (2) a second one is always more than 90% except previous two levels. The P/P+B values further support the deposition of the Tol sediments in deep marine conditions.

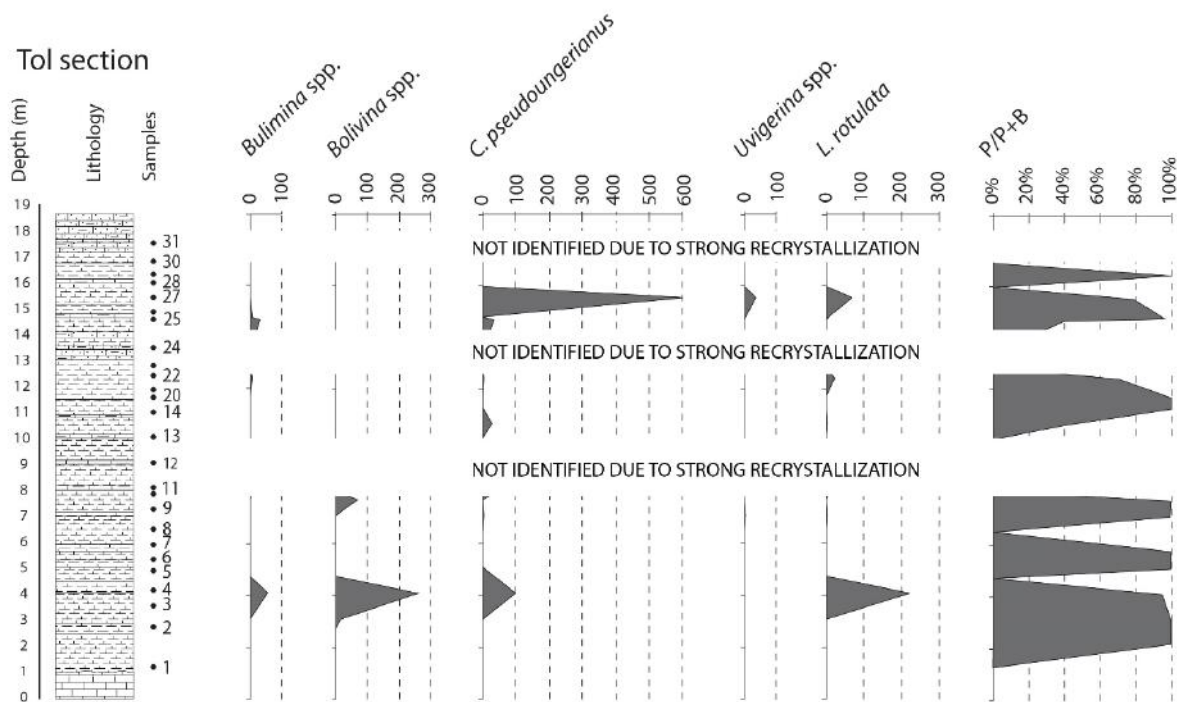


Figure 1.18: Normalized abundance frequencies of the most significant benthic foraminifera of the Tol section.

The generic composition of the ostracod assemblages from Tol section is very similar to that of Gülnar East section, but, among the 13 recognized species, some new occurrences have been detected, such as *Cytherella cercinata*, *Bithocypris bosquetiana*, *Krithe compressa*, *Krithe monosteracensis*, *Parakrithe lamellosa*, and *Macrocypris compressa*. *C. cercinata* and *B. sarsii profunda* are the most abundant species, followed by *B. producta*, *B. bosquetiana*, and *K. monosteracensis*. The ostracod assemblages of Tol section are, on average, more diversified than those of Gülnar East section, displaying 6 to 10 species, with a peak of 13 species at 7.80 m. The figures for TOL ostracod assemblage are under preperation.

1.7 Discussions

1.7.1 Age model

The micropaleontological analyses carried out throughout the Gülnar East and Tol sections highlighted the presence of some marker species among calcareous nannofossils, planktonic and benthic foraminifers. These markers, coupled with the correlation of the sapropel layers

occurring at the base of the section with those of Vrica and Singa sections (Calabria, Southern Italy) and ODP Site 967 (Eastern Mediterranean Sea) (Cita et al., 2008; Emeis & Sakamoto, 1998; Lourens et al., 1998) lead to obtain a well constrained age model. (Tab 1, supplementary material; Fig. 1.19).

The Gülnar East section starts with the presence of medium *Gephyrocapsa* from the first basal sample (0.3 m), providing a maximum age of 1.73 Ma (Raffi, 2002). This age corresponds to the bottom medium *Gephyrocapsa* (bmG) event in the Eastern Mediterranean, recognized for the first time by Shackleton and Crowhurst (1997) and calibrated by Lourens et al., 2004 and Anthonissen and Ogg (2012). The basal age of the section could be further constrained by the relevant occurrence, in the same first sample, of *G. crassaformis* that could be correlated with the first influx of this species in the ODP Site 967 dated at 1.722-1.723 Ma (Lourens et al., 1998). A definite age of around 1.72 Ma for the base of the section was essential to correlate the five sapropel layers recognized in the sedimentary succession of Gülnar East with those occurring in the outcropping Calabrian sections of Vrica and Singa and in the ODP Site 967 (Cita et al., 2008; Emeis & Sakamoto, 1998; Lourens et al., 1998). In particular, Gülnar East sapropels correlate well with sapropels 42-38 (sensu Lourens et al., 1998), which spans from 1.542 to 1.564 Ma. Furthermore, the age of the 5th sapropel is in agreement with the presence of some specimens of large *Gephyrocapsa*, which indicate an age younger than 1.617 Ma (Raffi, 2002).

Going upwards, at 4.40 m in the section it is relevant the presence, for the first time, of a modern morphotype of *B. marginata*, whose first occurrence is dated at 1.543 Ma in the Vrica Section (Cita et al., 2008; Lourens et al., 1996a). This bioevent is followed by the temporary disappearance of *Neogloboquadrina* spp. (sin) between 7.8 to 14.50 m. This event could correlate with the *Neogloboquadrina* spp. (sin) paracme aged between 1.367-1.37 and 1.21-1.22 Ma in the Vrica Section and in the ODP Site 967 (Lourens et al., 1996a; Lourens, et

al., 1996b; Lourens et al., 1998; Lourens et al., 2004; Cita et al., 2008).

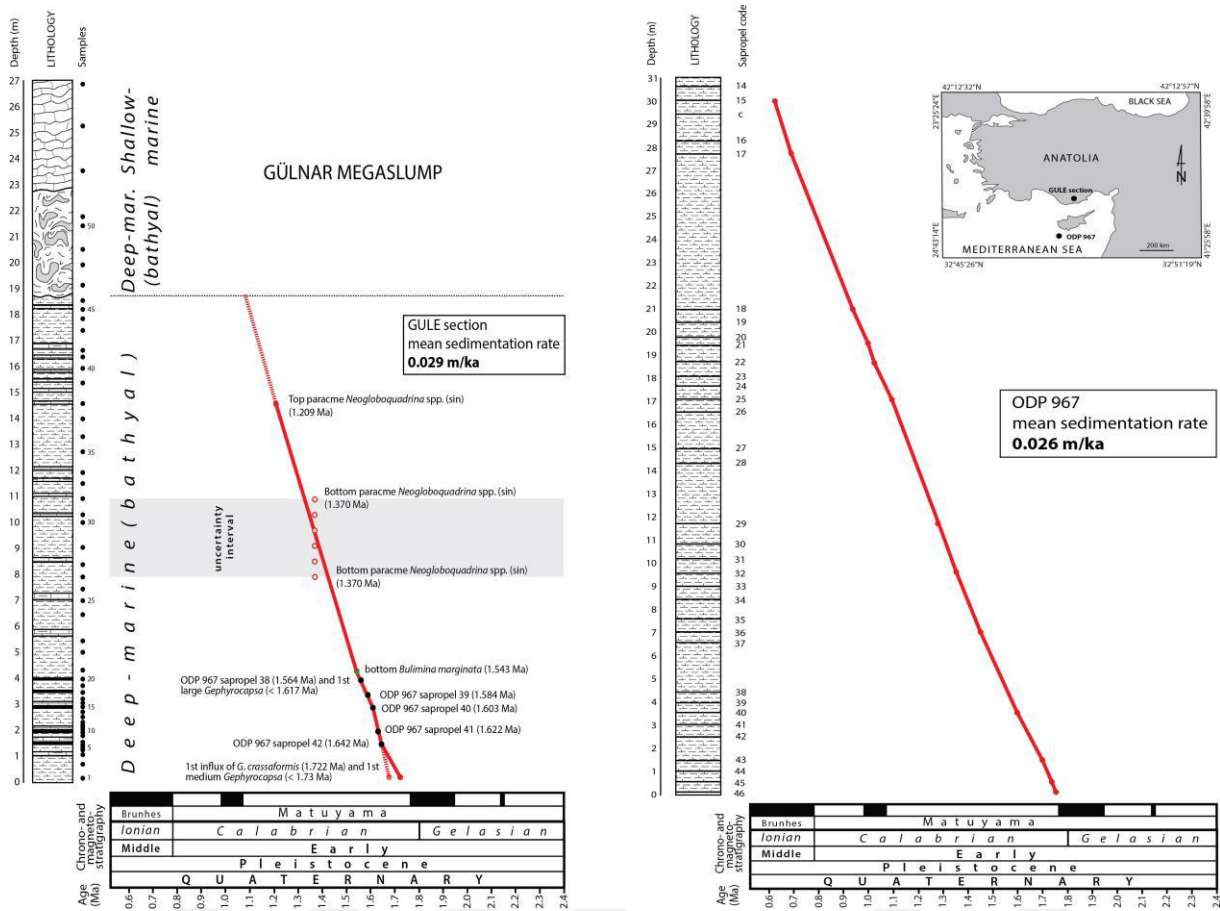


Figure 1.19: Graphics showing the age constraints of Gülnar East section and its comparison with ODP site 967 (ODP site 967 data from Emeis and Sakamoto, 1998).

The definition of a well constrained age model for the lower and middle part of the Gülnar East section led to the construction of the sedimentation rate curve illustrated in Fig. 1.19, which shows a rather constant rate throughout the section, with an average value of 0.029 m/kyr. The mean sedimentation rate of the Gülnar East section is comparable to the one calculated for the ODP Site 967, which is 0.026 m/kyr (Emeis & Sakamoto, 1998) (Fig. 1.19). Using this value, it was possible to calculate the possible age of the top of the sedimentary succession below the mega-slumping at about 1.08 Ma.

In conclusion, based on the occurrence of bio- and lithological markers, the sedimentation of the Gülnar East succession can be constrained between 1.72 (early Calabrian) to 1.08 Ma (late Calabrian).

Conversely, the sedimentation of Tol section starts with the presence of *Globigerinella calida* at 4.30 m and the total absence of neogloboquadrinids. The relevance of *G. calida* as a stratigraphic marker is not clear. Bolli and Premoli Silva (1973), followed by Bolli and Saunders (1985) established, in the Caribbean region, the *G. calida calida* subzone (within the *Globorotalia truncatulinoides truncatulinoides* Zone) aged between 0.14 and 0.08 Ma while, according to Anthonissen and Ogg (2012), the first occurrence of this species in the South Pacific area is considered a reliable marker bioevent aged 0.22 Ma. On the contrary, in the Mediterranean area the appearance of this species seems to occur earlier: according to Yıldız et al. (2003) it is a Calabrian event and, following Iaccarino and Premoli Silva (2007), *G. calida* appears within the MPle 2 Zone (*Globorotalia truncatulinoides excelsa* Interval Zone) slightly after 0.94 and before the Last Common Occurrence of *Neogloboquadrina* spp. (sin) dated 0.61 Ma (Lourens et al., 2004; Iaccarino & Premoli Silva, 2007; Anthonissen & Ogg, 2012). Since in Tol section neogloboquadrinids are not present, we must conclude that the study section should pertain to the MPle 2b Zone (*Globigerinella calida* Interval Zone), with an age for the base of the section younger than 0.61 Ma. Unfortunately, no other biomarkers have been found to better constrain the time span of the entire section and most importantly its uppermost part. However, it is worth to note that no *Globigerinoides ruber rosa* has been found in the Tol section, suggesting the top of the section should be older than 0.33 Ma, which defines its First Occurrence in the Eastern Mediterranean (Lourens et al., 2004; Anthonissen and Ogg, 2012).

Taking into account the obtained age of the Gülnar Est and Tol sections, the stratigraphic range of some ostracod species must be updated and rejuvenated: *Paijenborchella iocosa*, *Parakrithe declivis*, and *Parakrithe ariminensis* must be extended to the Calabrian, while *Bairdoppilata profunda*, *Bythocypris producta*, *Cytherella russoi*, *Cytherella cercinata*, *Parakrithe rotundata*, *Parakrite lamellosa*, *Krithe monosteracensis*, and

Krithe compressa to the Middle Pleistocene, at least for their occurrence in the Eastern Mediterranean.

1.7.2 Palaeoecology of the Pleistocene marine environment at the CAP southern margin

The micropaleontological analyses carried out on benthic microfossils showed that the Gülnar East and Tol sections deposited in a deep marine environment. The low diversity displayed by the foraminifer and ostracod assemblages is comparable with those typical of deep waters (Benson, 1975; Casale, 1982; Van Harten and Droste, 1988; Sgarrella and Moncharmont Zei, 1993; Sciuto, 2003; Spezzaferri and Tamburini, 2007) and in agreement with the lowest diversity recognized in the eastern Mediterranean area in comparison with the western one (Bonaduce, 1983; Van Harten & Droste, 1988; Baldrighi et al., 2014). Tabs. 2 and 4 in the supplementary material report the depth preferences of the recovered species derived from literature. Together with some strongly eurybathic species (foraminifers: *Martinottiella communis*, *Melonis pompilioides*, *Cassidulina laevigata*, *Stilostomella monilis*, *Siphonina reticulata*; ostracods: *Argilloecia* spp., *C. russoi*, *K. iniqua*, *P. ariminensis*), the assemblages include true epibathyal dwellers (from 200-1000 m following Sgarrella and Moncharmont Zei, 1993) among which *Uvigerina* spp., *Lenticulina* spp., *B. profunda*, *B. producta*, *C. robusta*, *H. sarsii profunda*, *O. mediterranea*, *P. iocosa*, *P. declivis*, and *P. rotundata*, or even bathyal species such as *Praeglobobulimina pupoides*. According to Sciuto et al. (2003) the co-occurrence in the Mediterranean of *Bythocypris* and *Krithe* indicates waters deeper than 250 m. *O. mediterranea* is a typical low thermospheric ostracod (Benson, 1977; Bonaduce and Barra, 2002) inhabiting deep waters around 400 m with bottom temperatures >10°C. Seldomly, this species has been recovered also with *Agrenocythere pliocenica* (Seguenza, 1880) in the deeper and colder waters of the psychrosphere (Barra et al., 1998; Aiello et al., 2000). The lack of this latter species in the Gülnar East section and the abundance of *Uvigerina* spp. suggests to infer depositional palaeodepths around 400 m. Moreover, Fisher α -

Shannon (H) diversity analysis shows that both of the sections represent shelf-deep marine environment (Fig 1.20).

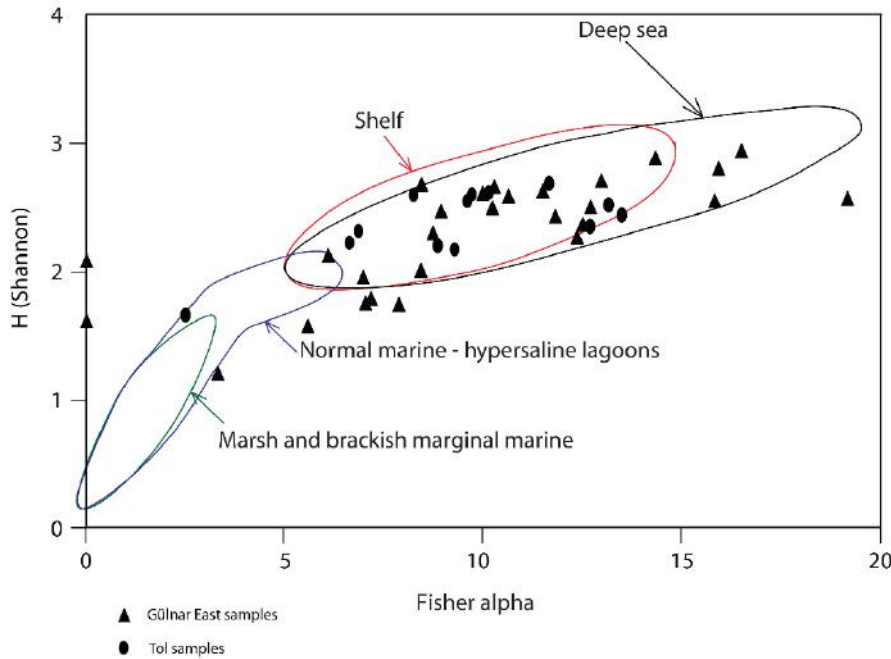


Figure 1.20: Plot showing the palaeo-depositional environments for both the Gülnar East and Tol sections.

More precisely, following Hohenegger (2005) we estimated the paleodepth of the marine environments for both the study sections to be between 400 to 500 m depth (Fig. 1.21), which shows that the qualitative estimation of the paleodepth is in good agreement with the results from the quantitative analysis.

The presence of important fractions of displaced and/or reworked species in each sample (Fig. 1.9 for GULE; Fig. 1.16 for TOL) is not uncommon in the deep marine environments (see Colalongo and Pasini, 1980, at Vrica; Aiello et al., 2000, in the upper part of the Monte S. Nicola section; Sciuto and Rosso, 2008, at Furnari; Tsourou et al., 2012, at Zakynthos), due to the transport from the Pleistocene shallow-water marine deposits cropping out along the margin of the Mut Basin at Hacıhametli, Sarıkavak, and Kargıcak localities (Yıldız et al., 2003), or due to the erosion of the ancient margins of the Mut Basin. At Gülnar East section, the littoral reworked ostracods most probably derive from the erosion of the Late Miocene Köşelerli Formation cropping out in the surrounding areas (Başayla section, Mut Basin, Cosentino et al., 2012 and unpublished data; Saraycık and Sarıalan sections, Antalya

Basin, (Babinot, 2002; Schildgen et al., 2012 and unpublished data).

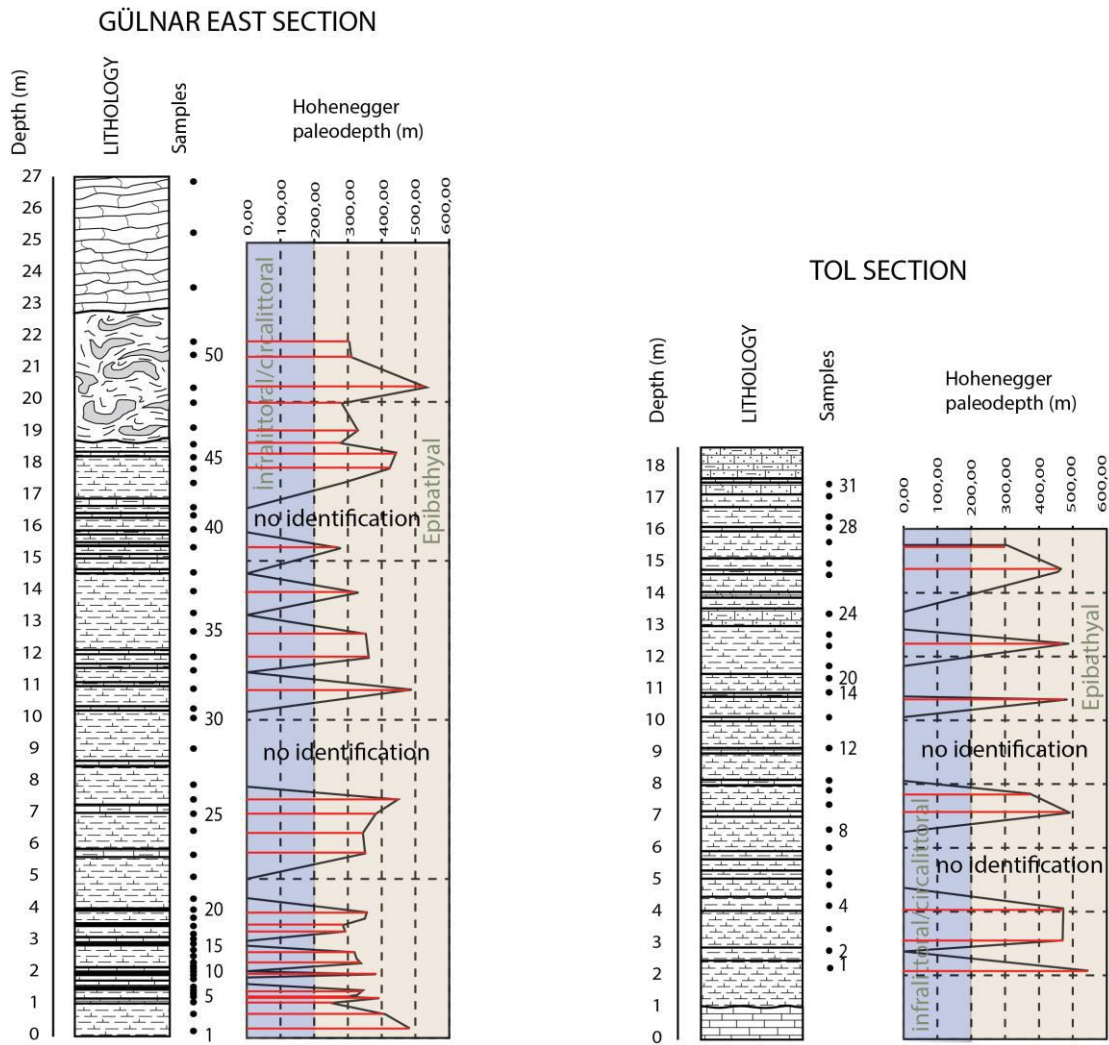


Figure 1.21: Estimated paleodepths for Gülnar East and Tol sections by using the Hoenegger (2005) method. The results show paleodepth oscillations within the epibathyal zone (between 300 to 500 m depth) for both the study sections. Only the red lines represent the depths to consider.

Planktonic foraminifer frequency curves of climate significant species could give some further information on the temperature variations of the Eastern Mediterranean surface waters during the Pleistocene (Fig. 1.22). The high occurrence of *Globigerinoides ruber*, but also *G. trilobus*, *Globigerina bulloides* and *Orbulina universa*, is indicative of warm conditions, whereas high frequencies of *Neogloboquadrina acostaensis* (sin), *N. pachyderma* (sin), as well as *G. scitula* indicates cooler waters (Hemleben et al., 1989; Pujol & Vergnaud Grazzini, 1995; Triantaphyllou et al., 2010) (Fig. 1.12 for GULE; Fig 1.17 for TOL). In particular, *G.*

ruber is a typical species of warm surface water, and blooms during maxima insolation in the spring/summer periods. Neogloboquadrinids are typical species of cool/cold waters, and bloom during minima insolation and autumn/winter periods (Caruso, 2004). At Gülnar East section, Neogloboquadrinids (sin) show an opposite trend to *G. ruber*, indicating an alternation between phases of cooling and warming.

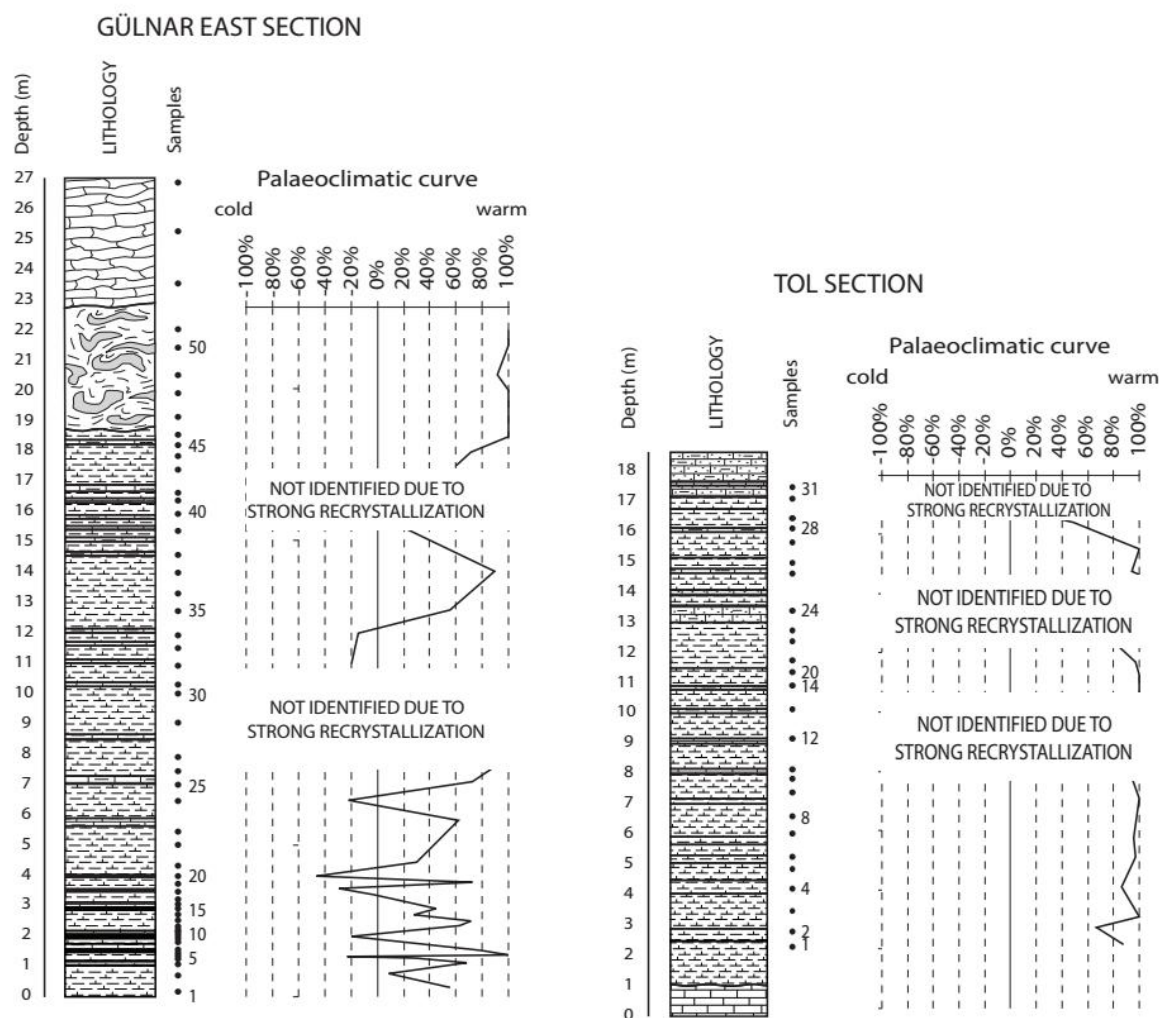


Figure 1.22: Paleoclimatic curves of Gülnar East and Tol sections following Amore et al (2004). It is worth to note that Tol section is younger than Gülnar East section, suggesting a long lasting period of warm surface water from the late Early Pleistocene (late Calabrian) to the Middle Pleistocene (Ionian).

Benthic foraminifera oxygen index gives an additional information on the oxygen conditions during the sedimentation of the study sections. Following the methodology of Kaiho (1994), we plotted a diagram showing oxic-suboxic levels throughout the sections (Table 1.1, Fig 1.23).

Table 1.1: Faunal groups representing oxygen conditions (Kaiho, 1994).

Oxygen condition	Faunal groups
Oxic	<i>Cassidulinoides bradyi</i> , <i>Cibicidoides pseudoungerianus</i> and <i>Cibicidoides wüllerstorfi</i>
Suboxic	<i>Cassidulina</i> spp., <i>Bulimina striata</i> , <i>Dentalina</i> spp., <i>Fissurina</i> spp., <i>Gyroidina</i> spp., <i>Gyroidinoides</i> spp., <i>Lagena</i> spp., <i>Lenticulina</i> spp., <i>Melonis</i> spp., <i>Oridorsalis umbonatus</i> , <i>Pullenia</i> spp., <i>Sphaeroidina bulloides</i> , <i>Uvigerina</i> spp.
Dysoxic	<i>Bolivina alata</i> , <i>Bolivina punctata</i> , <i>Praeglobobulimina pupoides</i>

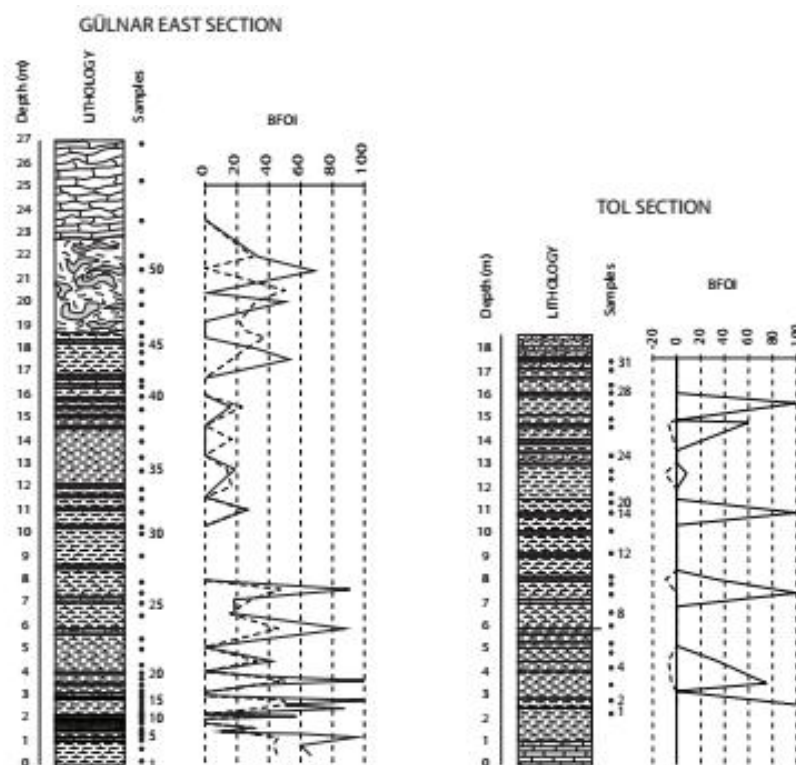


Figure 1.23: Figure showing the peaks related to presence of oxic, suboxic and dysoxic species for both sections. Legend: Dashed line for suboxic-dysoxic condition; solid line for oxic condition.

Accordingly, the relative increase in suboxic-dysoxic species at the bottom part of the Gülnar East section corresponds to the presence of the sapropel layers that were correlated to the sapropel 38-42 identified in ODP 967.

1.8 Plio-Pleistocene palaeogeography at the CAP southern margin, Pleistocene uplift rate, and mechanisms for the uplift

The palaeogeographic reconstruction in Fig. 24B tentatively shows the palaeo-topography during Plio-Pleistocene times. Extended outcrop of Plio-Pleistocene clastic continental

sediments all over the plateau interior (Fig. 24A; Şenel, 2002; Ulu, 2002) testifies that this area was already emerged at this time. Contrariwise, the southern side of the CAP southern margin was almost completely drowned. As previously stated, palaeoecological analyses performed on the benthonic fossil content (benthic foraminifers and ostracods) of both Gülnar East and Tol sections indicate that, in this area, epibathyal condition with palaeodepths of 400-500 m persisted from 1.72 (Early Pleistocene, early Calabrian) up to 1.08 (Early Pleistocene, late Calabrian) (Gulnar East section) and even after 0.61 Ma (early Middle Pleistocene, Ionian) (Tol section).

Summing the 400-500 m palaeo-bathymetric data to the ca. 1090 m a.s.l. present-day elevation of the youngest Pleistocene deposits of the Tol section, the post- early Middle Pleistocene uplift for the CAP southern margin results to be ca. 1500 – 1600 m.

Palaeogeographic reconstruction of Fig. 24B. shows how the inferred Plio-Pleistocene palaeo-coastline was retreated inland and insisted along the areas today lying at ca. 1500 m a.s.l. (Fig. 24B), even though part of the Plio-Pleistocene coastline has been evidently disrupted by the fluvial erosion of the Mut River and its tributaries. The uplift estimate of ca. 1500 - 1600 m increases up to ca. 1550 - 1650 m when taking into account the elevation of the sea level for the time-span comprised between 1.72 and 0.61 Ma, which was ca. 50 m lower than the present sea level (Miller et al., 2005). Nevertheless, even being conservative and using the ca. 1500-1600 m of uplift experienced by the CAP southern margin since 0.61 Ma, fast uplift rates reaching values up to 2.45 - 2.62 mm/yr can be evaluated. The resulted uplift rates for the CAP southern margin, obtained by considering the new time constraints for the youngest marine deposits capping the CAP southern margin, are higher than the uplift rates suggested by different authors for the same area considering both different deposits and time span (Cosentino et al., 2012; Schildgen et al., 2012a; 2012b).

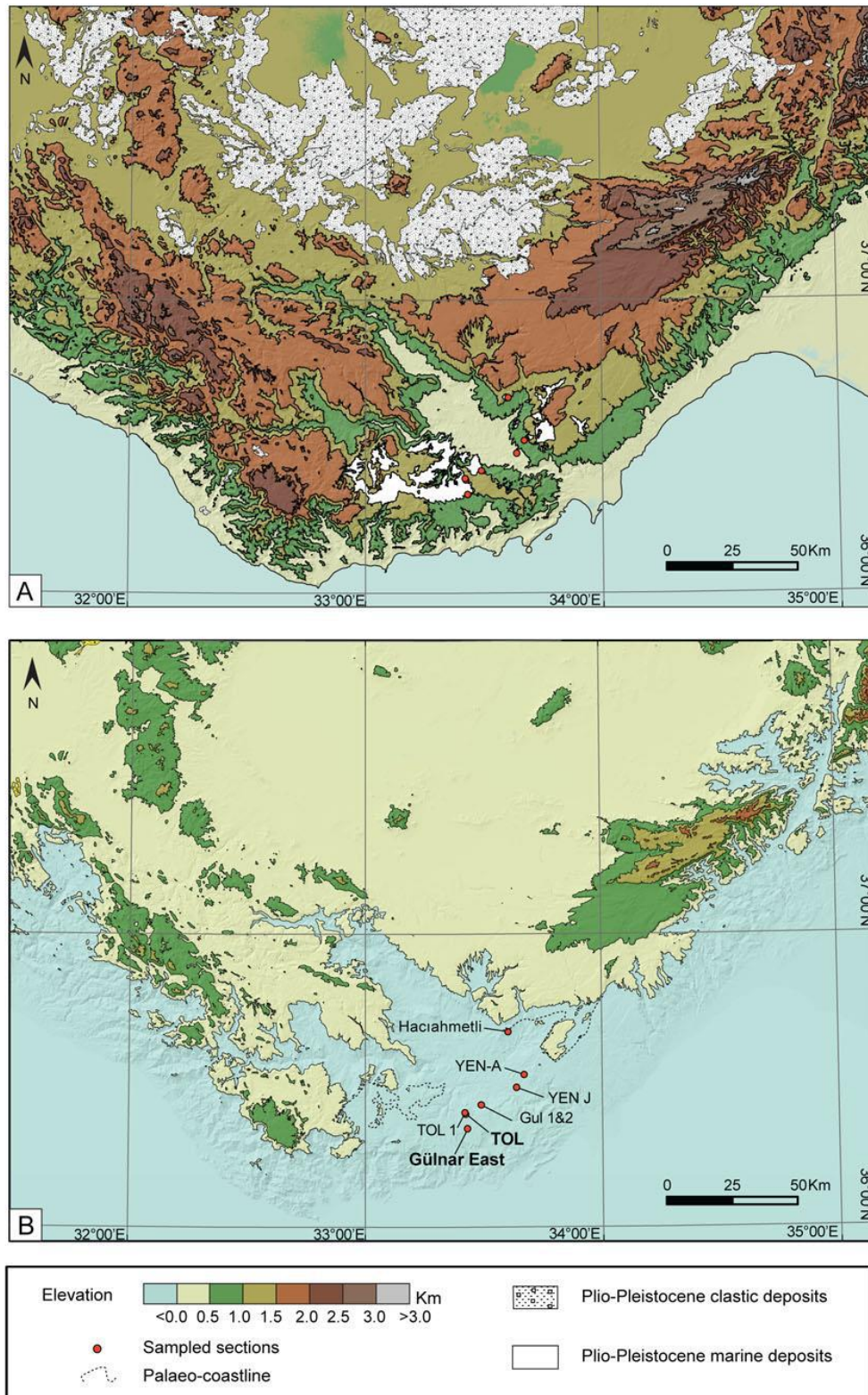


Figure 1.24: [A] Present-day topography of the Central Anatolian Plateau southern margin showing the outcrops of continental (dotted white pattern) and marine (white area) Plio-Pleistocene deposits. [B] Palaeogeography at Plio-Pleistocene times, inferred using palaeo-elevation and palaeo-bathymetric indications coming from both continental and marine Plio-Pleistocene deposits. Red dots show the location of the Plio-Pleistocene marine outcrops. Dotted black line shows the portion of the limit between the Miocene and the Plio-Pleistocene sequences interpreted as the remnants of the Plio-Pleistocene palaeo-coastline.

Cosentino et al. (2012), considering the late Tortonian age (ca. 8 Ma) of the highest marine deposits at the CAP southern margin (ca. 2000 m a.s.l.) calculated an average uplift rate of 0.24-0.25 mm/yr. The sub-horizontal beds of the marine deposits exposed at the CAP southern margin suggested those authors that the post-8 Ma uplift mechanism must be rooted deep in the crust or in the upper mantle. This observation, together with tomographic evidence of slab break-off beneath the Eastern Anatolian plateau (Keskin, 2003; Şengör et al., 2003), led Cosentino et al. (2012) to suggest that at least some of the ca. 2 km of post-8 Ma uplift of the CAP southern margin is compensated by low-density upwelling asthenospheric mantle following slab break-off.

Schildgen et al. (2012), combining uplifted marine deposits, which reveal average uplift rates of 0.25 to 0.37 mm/yr since Late Miocene time and 0.72 to 0.74 mm/yr after Early Pleistocene (early Calabrian), with abandonment ages of Late Pleistocene strath terraces showing average incision rates of 0.52 to 0.67 mm/yr, evidenced a multi-phased uplift of the CAP southern margin. According to Schildgen et al. (2012), the multi-phased uplift and the post-Calabrian increase in the uplift rate at the CAP southern margin can be explained by lithospheric slab break-off and possibly also the arrival of the Eratosthenes Seamount at the collision zone south of Cyprus.

The previous uplift rates for the CAP southern margin were well in the range of uplift rates predicted from numerical modeling of spontaneous slab break-off at different depths (Duretz et al., 2011). A slab break-off at a depth of 300 km produce an uplift rate of 0.2 mm/yr, whereas uplift rates of 0.8 mm/yr are predicted for break-off developed at about 50 km of depth. The post 0.61 Ma uplift rate of the CAP southern margin documented in this paper is even three time faster than the uplift rate predicted by the numerical model of a shallow slab break-off (50 km).

This result call into question the effectiveness of a lithospheric slab break-off as the only mechanisms for triggering the uplift of the CAP southern margin. In addition, a multi-phased uplift of the CAP southern margin with different uplift rates can be better explained with the combination of multiple sources of uplift. In the framework of a dynamic uplift of the Central Anatolian Plateau (Cosentino et al., 2012; Schildgen et al., 2012; 2014), the local mantle flow due to the slab break-off of the subducting Arabian plate should be coupled with an additional force capable to increase the uplift rate of the overriding plate. In the Arabian-Anatolian collision zone, the local increase in mantle flow due to the slab break-off can be coupled with the arrival of hot material through the northward directed mantle flow sourced by the Afar deep mantle upwelling (Faccenna et al., 2013). Numerical as well as analogical models are needed to test if the topographic response of a such complex mantle flow system combines with the multi-phased uplift of the CAP southern margin and its post 0.61 Ma fast uplift rate. However, even though field evidence and models can fit well, the timing suggested in Faccenna et al. (2013) for the Afar mantle upwelling to flow beneath Central Anatolia and even reach the Aegean area (3-5 Ma) does not fit with 1500 m of uplift of the CAP southern margin in the last 600 kyr.

1.9 Conclusions

At the CAP southern margin, in the Gülnar district, the youngest marine deposits capping the Taurides tectonic units at elevations between 1000 to 1500 m a.s.l. have been classically considered as Middle Miocene in age. Actually, this study demonstrates that they astonishingly contain a Pleistocene fossil content, which includes benthic and planktonic foraminifers, and calcareous nannofossils. Combining biomarkers from both Gülnar East and Tol sections, we can conclude that the Mediterranean Plio-Pleistocene transgression reached the CAP southern margin in the Gülnar district in the Early Pleistocene (ca. 1.7 Ma), with the

deposition of Early Pleistocene epibathyal marine deposits onto Middle Miocene shallow-water limestones of the Mut Formation. In addition, given that in the Tol section neogloboquadrinids are not present, the youngest bioevent recognizable in the Pleistocene deposits capping the CAP southern margin is the LCO of *Neogloboquadrina* spp. (sin), which is dated 0.61 Ma. This last biostratigraphical result lead to consider the youngest marine deposits in the Gülnar district as at least early Middle Pleistocene. Qualitative and quantitative analyses on the benthic fauna from both the Gülnar East and Tol sections point to a marine environment from the epibathyal zone (400-500 m of paleodepth), with a palaeo-coastline attested at ca. 1500-1600 m a.s.l.

Considering the elevation of the early Middle Pleistocene palaeo-coastline (1500-1600 m a.s.l.) and the maximum age (0.61 Ma) for the youngest marine deposits at the CAP southern margin, fast uplift rates reaching values up to 2.45 - 2.62 mm/yr can be evaluated. These uplift rates are higher than those suggested by different authors for the same area based on both different deposits and even shorter time span. This result on the uplift rate call into question the effectiveness of a lithospheric slab break-off as the principal mechanisms for triggering the uplift of the CAP southern margin, as recently suggested by some authors (e.g., Cosentino et al., 2012; Schildgen et al., 2012; 2014). Indeed, the maximum uplift rate predicted by numerical model for shallow slab break-off (50 km depth) is only 0.8 mm/yr (Duretz et al., 2011).

In the framework of a dynamic uplift of the Central Anatolian Plateau, a more complex mantle flow system could be responsible for the post 0.61 Ma uplift rates of the CAP southern margin. Even though the timing seems to be slightly different, this complex system can be due to the combination of the local mantle flow, due to the Arabia-Anatolia subduction system, with the arrival of hot material through the mantle flow sourced by the Afar deep mantle upwelling (Faccenna et al., 2013).

In conclusion, the results of this work can be useful to set up numerical and analogical models to test if the topographic response of a such complex mantle flow system combines with the post 0.61 Ma fast uplift rate of the CAP southern margin.

Acknowledgements

This study is part of the Anatolian Plateau Climate and Tectonic Hazards (ALeT) project, funded by the European Union Marie Curie Actions “Initial Training Networks” program (FP7-PEOPLE-2013-ITN)” (grant agreement number 607996).







CHAPTER 2

COASTAL EVIDENCE FOR SHORT-TERM UPLIFT AND ACTIVE TECTONICS AT THE CAP SOUTHERN MARGIN

Table of Contents

2.1 Introduction.....	54
2.2 Coastal geomorphic markers and archaeological infrastructures useful to evaluate sea-level changes.....	58
2.2.1 Marine notches.....	58
2.2.2 Beachrocks.....	60
2.2.3 Algal rims	61
2.2.4 Alluvial deposits.....	62
2.2.5 Archaeological infrastructures.....	63
2.3 Material and methods.....	63
2.4 Discussions and conclusions.....	69
Acknowledgements.....	71

2.1 Introduction

For a better investigation on the coastal geodynamics and sea-level changes, to focus on some specific deposits and geologic structures is essential. Among those, marine notches, trottoir rims, beachrocks, and alluvial deposits at the outlet of the river, where the alluvial plain meets the shoreline, host very important geologic archives.

Along the southern coasts of the Central Anatolian Plateau (CAP), where previous papers (e.g., Cosentino et al., 2012; Schildgen et al., 2012a; 2014; chapter 1 in this thesis) evidenced the highest long-term uplift rate of the CAP, limestone and/breccia rocky shorelines are good environments for notches, beachrocks, and trottoir rims development useful for evidencing the tectonic/climate dynamics of the region even for short time span (mainly late Pleistocene-Holocene). Coastal archaeological sites of southern Anatolia, including Roman harbors and fishponds, are also very useful to evidence vertical movements connected with the geodynamics of a region with so strong long-term uplift rates and to define its short-term uplift rate, if any.

Since eustatic sea-level changes are well known in the Quaternary (Milliman and Emery, 1968; Haq et al., 1988; Schnack and Pirazzoli, 1990; Lambeck and Chappell, 2001) to precisely define the short-term and recent uplift rate of a coastal region, it is crucial to compare the evidence for palaeo-coastlines with the eustatic position of the sea-level in the study area. For what concern the Holocene sea-level changes in the Mediterranean, Lambeck & Purcell (2005) reconstructed the curve of the sea-level rise following the major sea-level drop of the Last Glacial Maximum (LGM). The Lambeck & Purcell (2005) model predict the eustatic position of the sea-level in a reference area by the spatially variable glacio-hydro-isostatic response of the earth and ocean to the growth and decay of ice sheets. Superimposed on these glacio-hydro-isostatic variation of the sea-level are the relative changes of any vertical movement of the land. The most recent study about the Holocene sea-level changes in

the Mediterranean Basin belongs to Antonioli et al (2006), which was conducted in the Central Mediterranean region, and then improved by Antonioli et al. (2007). On the other hand, for what concern the easternmost coastal areas of the Mediterranean, there are local studies that cannot be extended to all Eastern Mediterranean region. In Lebanon, Holocene deposits and sea-level markers point to regional crustal uplift episodes for the last ~6000 years (Morhange et al., 2006). In Cyprus, reduced uplift and eustatic sea-level changes were reported, while some local submergence were observed during the late Pleistocene-Holocene (Poole and Robertson, 1991). In the northern coasts of Israel, based on archaeological data, it has been reported a relative sea-level stability for the last 4000 years (Galili et al., 2005). Along the southeastern coasts of Turkey, in the Hatay region, 2 rapid uplifts of the late Holocene were observed (Pirazzoli et al., 1991). Another detailed study based on benthic foraminifera (Spezzaferri and Tamburini, 2007) shows that even though the main reason for the deepening of the Eratosthenes Seamount in south of Cyprus is the collapse of the seamount, it is stated that the minor shift in the paleobathymetry of the study area could be linked to the global sea level changes from the Miocene/Pliocene boundary to present.

Those differences in the Holocene vertical movements are due to the complex geodynamics of the Eastern Mediterranean, which is characterized by active subduction zones, continental transform faults, continental collision zones, triple junction, etc. (Fig. 2.1). All these tectonic structures that affect the Eastern Mediterranean induce differential vertical movements of the Holocene coastlines (Antonioli et al., 2015).

For the coastal area between Mersin and Anamur (southern Turkey), the Lambeck & Purcell (2005) glacio-hydro-isostatic model predicts the following position of the sea-level in comparison with the present one: -0.5 m at 2 ka; -1.5 m at 4 ka; -4 m at 6 ka. For defining the Holocene vertical movements of the coastal area between Mersin and Anamur, we refer to the Holocene sea-level curve as predicted by the Lambeck & Purcell (2005) model (Fig. 2.2).

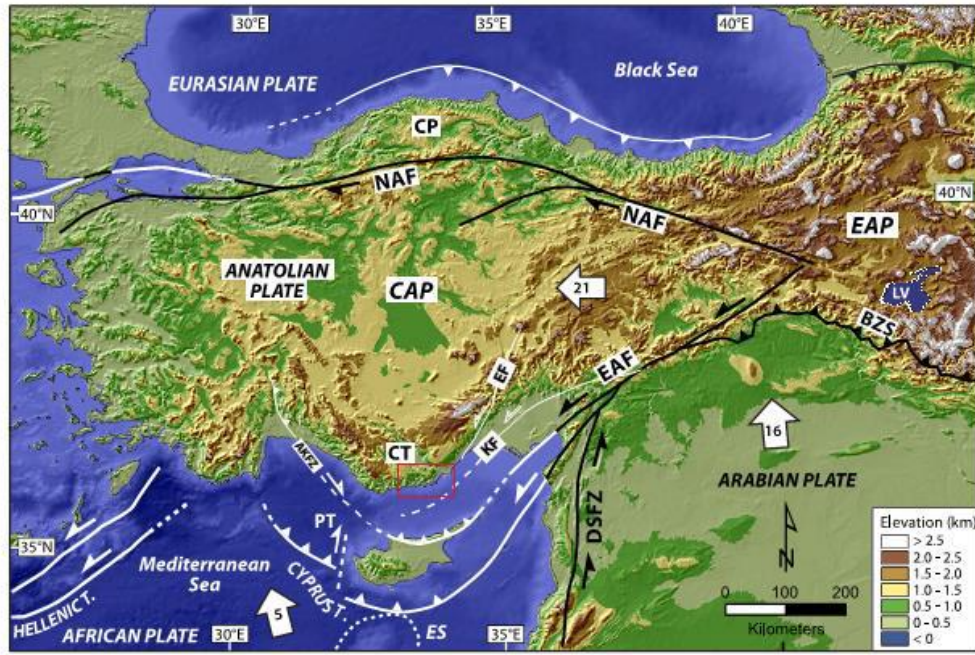


Figure 2.1: Map showing the main tectonic structures in the study area identified in red square (modified after Schildgen et al. 2014 and Aksu et al. 2014). CAP: Central Anatolian Plateau; EAP: Eastern Anatolian Plateau; CP: Central Pontides; CT: Central Taurides; BZS: Bitlis–Zagros suture; NAF: North Anatolian Fault; EAF: East Anatolian Fault; EF: Ecemiş Fault KF: Kozan Fault; AKFZ: Aksu-Kyrenia Fault Zone; DSFZ: Dead Sea FaultZone; ES: Eratosthenes Seamount; PT: Paphos Transform; LV: Lake Van. White arrows with numbers inside represent plate movement (mm/yr) with respect to a stable Eurasia (Reilinger et al., 2006).

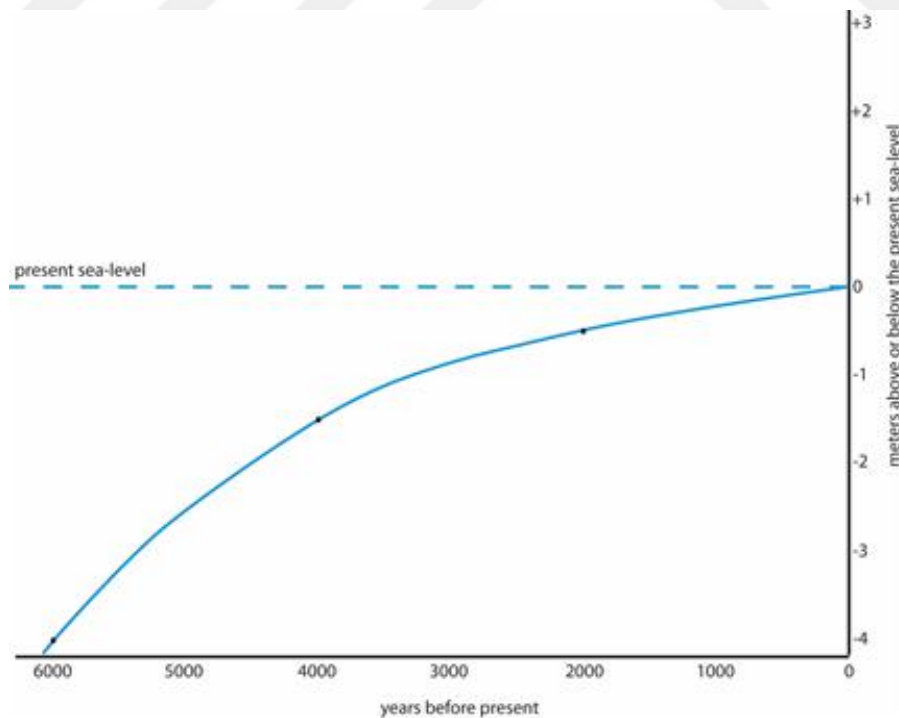


Figure 2.2: Holocene sea-level curve as predicted by the glacio-hydro-isostatic model of Lambeck & Purcell (2005) for the southern coastal area of Turkey between Mersin and Anamur.

In the tectonic complexity of the Eastern Mediterranean, the CAP southern margin is a region that is located in between the Tauride Mts., in the north, and the subsiding Adana-Cilicia Basin in the south, with the colliding Cyprus Arc further south. The CAP southern margin is bounded to the southeast by the seismically active Ecemiş-Kozan Fault Zone (Burton et al., 1984; Aksu et al., 2014a; 2014b), which towards SW reaches the offshore realms of the study area (Figs 2.1). Considering that the study area is mainly characterized by rocky coasts with many coastal archaeological sites, it could potentially be a comprehensive archive of geological data related to the short-term uplift of the CAP southern margin, the active tectonics that possibly affected the study area, and the historical seismicity of the region.

The aim of this study is to define the vertical movements that affected the study area during the Holocene, to evaluate the short-term uplift rate at the CAP southern margin, and, possibly, provide evidence of active tectonics, in an area that, although a really fast long-term uplift rate (ca. 2.5 mm/yr), the Active Fault Map of Turkey (Emre et al., 2012) does not show any active fault in the study area.

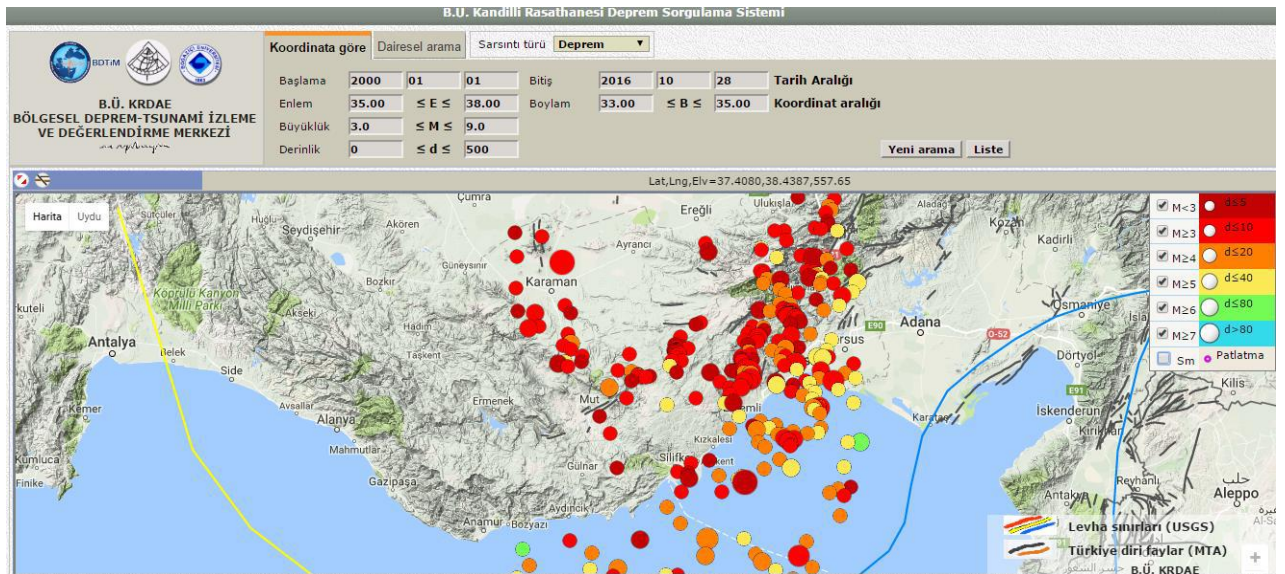


Figure 2.3: Map showing the last 15 years earthquakes ($M \geq 3$) in the study area that are clustered around Ecemiş-Kozan Fault Zone (from B.Ü. Kandilli Observatory and Earthquake Research Institute, <http://www.koeri.boun.edu.tr/sismo/zeqdb/>).

2.2 Coastal geomorphic markers and archaeological infrastructures useful to evaluate sea-level changes

2.2.1 Marine notches

Marine notches are coastal geomorphologic features due to bio-erosion and/or dissolution processes. They are cut into rocky coasts and can be used as indicators of mean sea-levels in areas with a limited tidal range, which in the Mediterranean Basin is generally less than half a meter (Pirazzoli, 1986a; Stewart and Morhange, 2009). Even though it is not certain the process for the development of notches, in general 4 processes are considered: (1) physical, (2) biological, (3) chemical or (4) mechanical processes. However, since all those processes can contribute in the formation of marine notches, a general term “sea corrosion” is being used to cover all those physico-chemical and biological processes that can develop in the intertidal zone of a rocky coast (Carobene, 2015; Guilcher and Pont, 1957; Pirazzoli, 1986a). In this context, the main agents in the morphogenesis of a rocky coast for sea corrosion are considered chemical dissolution and bio-erosion (Pirazzoli, 1986a).

The main principle for giving a marine notch in a rocky coast is that continuous wave movement acting on the rock surface is accompanied by bio-erosion from endolithic boring organisms, such as *Lithophaga* and/or grazers such as *Patella*, which using their abrasive organic fluids dissolve carbonate rocks (Carobene, 2015; Thomas, 1996; Trudgill, 1987). The discrimination for tectonic control or sea-level change regarding the notch formation is mainly based on the shape and relative position of the formed notch. In that sense, if the sea-level is relatively stable, the notch must have U- or V-shaped profile (Pirazzoli, 1986a; Pirazzoli and Evelpidou, 2013). It is assumed that if a U- or V-shaped notch profile formed, it means that the uplift/submergence rate matched the sea-level rise/drop, respectively; on the other hand, ill-formed notches develop when the uplift/submergence rate exceeds the sea-level changes (see figs. 2 and 3 in Boulton and Stewart 2013). Accordingly, it is possible to

detect the relative tectonic movement of a region looking at the shape of a notch, or its position respect to the mean sea-level.

Even if there is no agreement on the formation of marine notches, the common assumption is that the generation of notches is not only related to bio-erosional processes but also to sea-level changes and tectonic movements, when the rate of eustatic sea-level rise matches the tectonic uplift, it results in forming notches on rocky coastlines (Cooper et al., 2007; Antonioli et al., 2015). Following this assumption, in theory, the present-day elevation of the marine notch is a good clue to investigate the uplifted/submerged former shorelines and/or climate dynamics in terms of sea-level changes in tectonically silent areas (Cooper et al., 2007; Boulton & Stewart, 2013; Carobene, 2015).

It is worth to strike out that, recently, some authors claimed in the Mediterranean notches are not forming anymore due to the slow rise of the Mediterranean sea-level (Evelpidou et al., 2012; Pirazzoli & Evelpidou, 2013). In contrast, other scientists documented the existence of present-day notches in several sites around the Mediterranean, such as : Italy, France, Croatia, Montenegro, Greece, Malta, and Spain (Antonioli et al., 2015).

Among the best materials for ^{14}C datings marine notches, it has been recommended to use mollusk shells such as *Lithophaga*, *Balanids*, *Vermetids* or *Serpulids* (Laborel and Laborel-Deguen, 2016). As the best material, in the same study, it has been recommended to use *Dendropoma* because of its depth accuracy ($\pm 5\text{cm}$). Unfortunately, in the study area it was not found, than for AMS ^{14}C datings we used other organisms such as algae trottoir, *Bivonia triquetra*, others mollusk shells, and large mammal teeth.



Figure 2.3: Rocky coast in Narlıkuyu. The present day notch, with a well developed algae trottoir, and an older uplifted marine notch are visible. Some flat surfaces in the picture, cutting the Middle Miocene limestone of the Mut Formation, are referable to ancient wave-cut platforms.

One strong evidence that is attributable to the studies of Holocene elevated marine notches and sea-level changes are their impact on the archaeological sites of the study area and surrounding localities (Pirazzoli, 1986b; Lambeck et al., 2004; Galili et al., 2005; Marriner & Morhange, 2007; Auriemma & Solinas, 2009; Anzidei et al., 2011).

Here, we try to combine our measurements and AMS ^{14}C datings of marine notches with some archaeological remains, such as ancient harbors and/or fishponds (Figs 2.10), to possibly reconstruct the Holocene vertical movements of the study area.

2.2.2 Beachrocks

Beachrocks are carbonate-cemented marine deposits that form in the intertidal zone, between high tide and low tide. They can form in several months up to several hundreds of years in relation to vertical stability of sea-level (Frankel, 1968; Turner, 2005; Desruelles et al., 2009). Beachrocks are mainly common in tropical and subtropical seas (Cooper, 1991; Goodwin, 2009). One of the most recognizable features of beachrocks is their sea-ward dipping (Turner,

2005). In previous studies, it is highlighted that the material to date the beachrock should be taken with attention. In fact, trying to date a beachrock, the maximum age of deposition could be obtained from the constituent organic material, whereas the cement could give the minimum age (Hopley, 1986). For this reason, additional comparable data are required. Even though also for beachrocks there are some doubts regarding the chemical factors in the cementation process, there is a general consensus that they are good sea-level indicators as long as they are correctly evaluated (Hopley, 1986; Kelletat, 2006).

In our study, to date the beachrocks we used mollusk shells collected within the coarser layers of the beachrock. So we interpret our results as indicating the maximum age of the beachrock. In the same study (Hopley, 1986), it is also highly suggested not to use beachrocks as sea-level indicators in places of high tidal range, which is not the case of the Eastern Mediterranean where the tidal range reaches max. 40 cm (Fig 2.4).

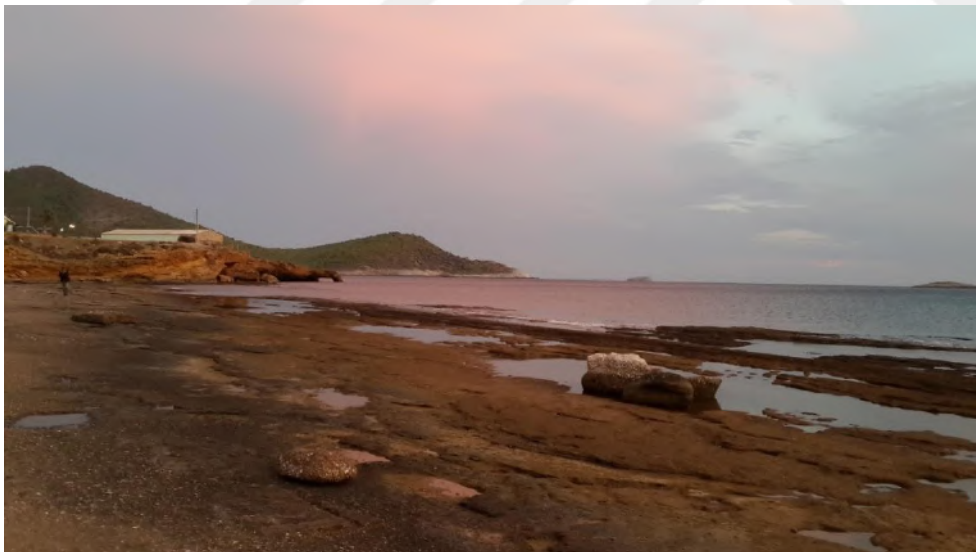


Figure 2.4: Sea-ward dipping beachrocks in Aydıncık locality. The innermost part of the Aydıncık beachrock (the highest part of the deposits) is ~0.2 m above the high tide. The picture has been taken at low tide.

2.2.3 Algal rims

Algal rims or *algal trottoirs* are intertidal landforms that form as a result of bioconstruction activity of algae mainly on rocky coasts (Hopley, 2005; Laborel, 2010). It is stated that the

presence of algal platforms on elevated shorelines are very strong evidence of past sea-levels because of their narrow repartition around mean sea-level and their lateral growth rate (Hopley, 2005; Laborel, 2010). In this study, some algal trottoirs have been analyzed and AMS ^{14}C dated.

2.2.4 Alluvial deposits

Alluvial plains are continental depositional environments of clastics that are sourced by material derived from the erosion of higher elevated regions (e.g. mountain chains). Rivers and streams transport the clastics to lower elevation areas forming plain surfaces characterized by clastic materials. The increase of clastic sediments in the succession of continental sedimentary basins can possibly be related to uplift in the river catchement areas and/or to climatic changes (Bull, 1972). They are good indicators for vertical movements when the alluvial plain reach the outlet of the river and the alluvial deposits laterally pass to the beach deposits. In that case the depositional surface of the alluvial plain can be considered as the base level of the river network, correlatable with the sea-level surface.



Figure 2.5: Uplifted Fossil-bearing (large-mammals remains) alluvial deposits at Yanışlı. The foreground beach gravels are laterally eteropic to the present day alluvial plain, which is 1.43 m below the uplifted alluvial plain.

2.2.5 Archaeological infrastructures

The coastal archaeological sites, with their infrastructures, are among the most reliable area for collecting evidence of vertical movements, sea level changes, seismic activity. They are comprehensive archives for studying the tectonic/climatic stability and/or instability of a region (Auriemma and Solinas, 2009; Marriner and Morhange, 2007). They can potentially give an exact age of each natural event (e.g., sea level changes, earthquakes, landslides, floodings, etc.) recorded in the historical archive of the site. For long-term tectonically stable areas, it is possible to make predictions in terms of sea-level change, or in terms of tectonic oscillations for climatically standstill archaeological areas. The southern coast of Anatolia is really rich in archaeological sites with coastal infrastructures, such as, among the others: Pompeiopolis (Mersin), Lamos (Lamos), Elaiussa-Sebaste (Ayaş), Corycus (Kızkalesi), Anemorium (Anamur); etc.

In our study area, the most common archaeological evidence are Roman fish tanks (fishponds), which were excavated in rocky coasts (Fig. 2.10), Roman pools excavated in rocky coasts to take a bath, and ancient Roman harbors (e.g., Pompeiopolis and Elaiussa-Sebaste). The functional height of the fishponds that were discovered in different places around the Mediterranean were considered at the same level of the mean sea-level of that period (Lambeck et al., 2004; Auriemma & Solinas, 2009). In that case, the topmost elevation of the external wall of the fish tank can indicate the sea-level during the activity of this coastal infrastructure (1st century B.C. and 2nd century A.D., Evelpidou et al., 2012).

2.3 Material and methods

In the area between Mersin and Anamur (southern Turkey), a geomorphological study to recognize the principal coastal landforms related to ancient coastlines has been carried out. The rocky coastal area, mainly characterized by outcrops of Middle-Upper Miocene shallow-

water limestones (Mut Formation) and Mesozoic to Paleozoic rocks of the Taurides tectonic units, well preserve geomorphic elements related to the position of ancient coastlines, such as: marine notches, abrasion platforms, pool rocks, trottoir rims, beachrocks, alluvial plains in the vicinity of the river outlet, etc. When possible, fossil remains, such as mollusk shells, large mammals, and plants, have been collected for dating the deposits by AMS ^{14}C dating technique. Elevation of the sea-level markers have been measured by using a metric rigid roll-up tape measure. All the measurements have been referred to the present mean sea-level. In the study area, the present mean sea-level has been defined as the mean position between high and low tide. In the eastern Mediterranean the tide amplitude is ca. 40 cm (Evelpidou et al., 2012). Usually, in the study area the present day algae trottoir, as well as the major concavity of the marine notch, well approximates the mean sea-level position. All the elevations in this study have been measured with reference to the present day algae trottoir or to the major concavity of the present day marine notch.

In addition, evidence for vertical movements have been collected from some archaeological sites containing coastal infrastructures such as harbours and Roman pool (Elaioussa-Sebaste, the present day Ayaş), and fish tanks (Narlıkuyu).

In the Narlıkuyu coastal area, which is mainly characterized by the outcrop of Middle Miocene shallow-water limestones, several markers of past sea-level, such as marine notches and abrasion platforms with rock-pools, are affected by tectonics (Fig. 2.3). In a small portion of the study area (Narlıkuyu village), 39 fault planes have been measured. Offsets of the sea-level markers (both marine notches and abrasion platforms with rock pools) reach values between a few centimetres to 70 cm. Meter-scale structural analyses carried out along the coastal area of Narlıkuyu revealed the presence of two different fault systems affecting the Holocene sea-level markers.

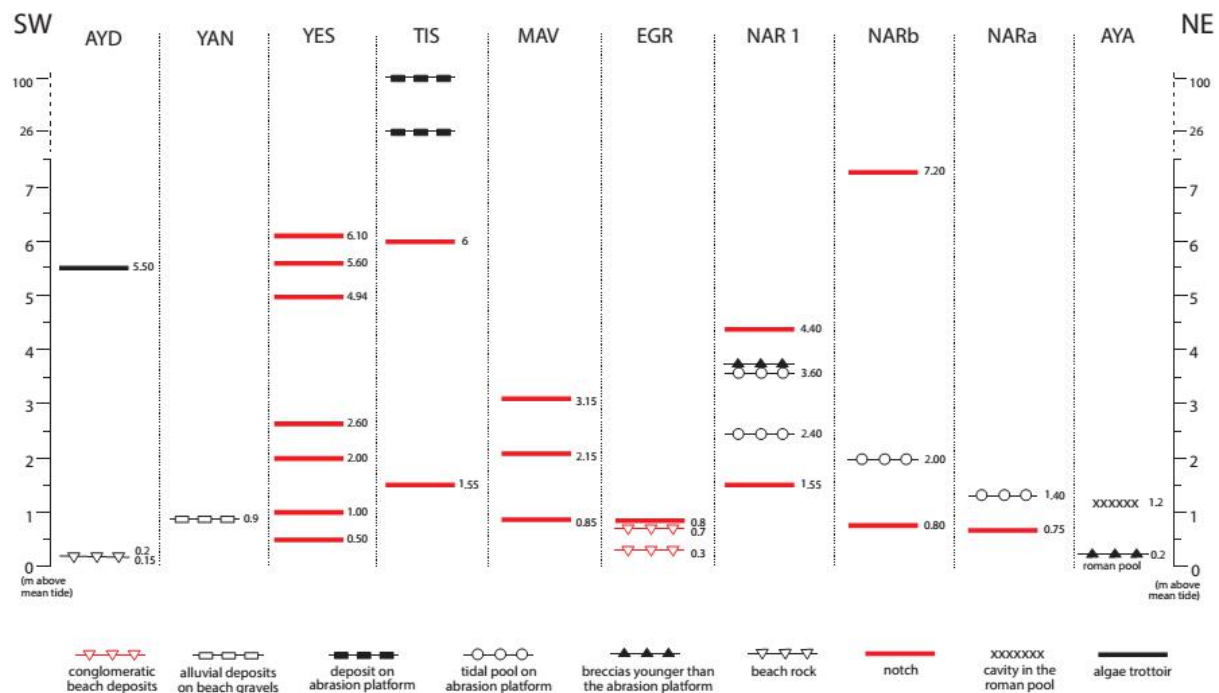


Figure 2.6: SW-NE section showing the altitudes of shoreline structures. AYD: Aydıncık; YAN: Yanışlı; YES: Yeşilovacık; TIS: Tisan; MAV: Mavikent; NAR: Narlıkuyu; AYA: Ayaş.

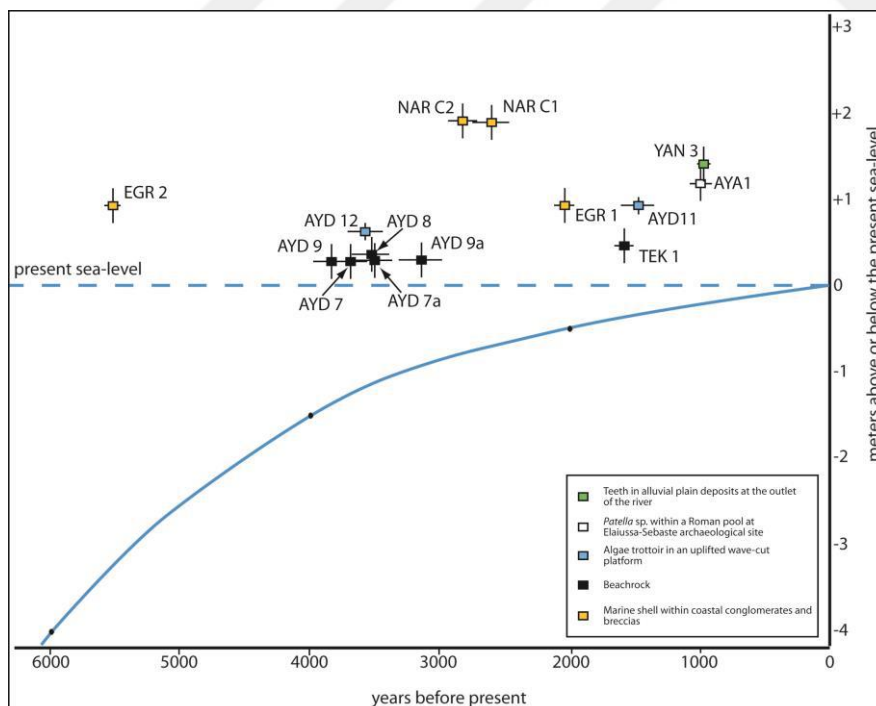


Figure 2.7: Holocene sea-level curve as predicted by the glacio-hydro-isostatic model of Lambeck & Purcell (2005) for the southern coastal area of Turkey between Mersin and Anamur. In the graph, AMS ¹⁴C datings have been plotted considering the elevation of the samples and possible palaeodepth corrections. All the samples plot above the present sea-level, showing Holocene uplift of the study area.



Figure 2.8: Extensional fault plane affecting a Holocene abrasion platform with rock pools and the relative tidal notch, with offset of ca. 70 cm.

The main fault system is represented by conjugate fault planes striking NNE-SSW, which shows normal offsets of the Holocene marine notches and tidal abrasion platforms. In the same measurement station, a secondary fault system shows deformation along WNW-ESE direction. A preliminary interpretation of those fault systems points to consider this latter to be represented by transfer faults kinematically related to extensional tectonics along NNE-SSW faults.

This conclusion is corroborated by the crystal growth of a syn-kinematic fibrous aragonite observed on one of the major WNW-ESE fault plane measured in the area. The c-axes of those syn-kinematic crystals, which is the growth axes of aragonite, is parallel to the strike of the fault plane, suggesting a strike-slip movement along the WNW-ESE fault during the aragonite growth.

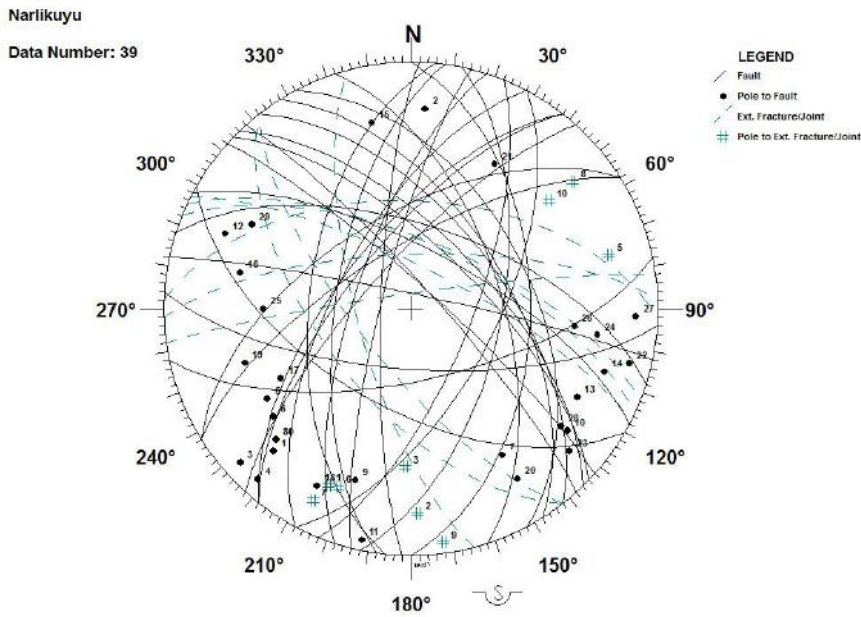


Figure 2.9: Stereoplot (Schmidt net lower hemisphere) of the fault planes and fractures affecting the Holocene sea-level markers at Narlıkuyu (southern Turkey).

Although according to the Active Fault Map of Turkey (Emre et al., 2012) no active faults are affecting the coastal area SW of Adana (Mersin –Silifke region), fault planes cutting Holocene sea-level markers should be indisputably considered as active faults, even capable to cut the ground surface. Active tectonics is well documented also in the offshore of the study area.

Offshore the Silifke area, active tectonics affecting all the Quaternary succession was recognized by Ozel et al. (2007) by interpreting single channel seismic reflection profiles. These authors evidenced a NE-SW major fault system that was considered to be an offshore extension of the Ecemiş Fault Zone. This fault system is mainly characterized by high angle faults, dipping to the SE, which show normal offsets even affecting the youngest seismic reflectors within the Quaternary sequence.

Usually, active and capable fault planes are related to local big earthquakes. Although the southeastern margin of the CAP is seismically active along a NE-SW tectonic lineament (Fig. 2.1 and 2.3), in the last century no major earthquakes ($M > 6.0$), capable to produce ground ruptures, struck the onshore area between Mersin and Anamur. In the last century, only an earthquake with magnitude more than 6.0 ($M = 6.0-6.4$) has been recorded in the

Silifke offshore (Jenkins et al., 2013). However, along the Mersin-Anamur coastal area there are historical evidence for local earthquakes. In the 4th century Corycus, the present day Kizkalesi, was struck by an earthquake that killed at least two people (father and son) as reported in an epitaph from the archaeological site of Corycus (Guidoboni, 1989; Guidoboni et al., 1994). At Elaïoussa-Sebaste, the present day Ayaş, the archaeological evidence shows indisputable signs of violent destruction due to earthquakes in the second half of the 7th century (Equini Schneider, 2008). According to the Digital Atlas of the Roman Empire, the Roman Empire was active in southern Anatolia, including Ayaş (Elaïoussa/Sebaste), Kizkalesi (Corycus), and Narlıkuyu (Arima?/Kime?) until 640 AD (Åhlfeldt, 2015). Roman infrastructures such as: harbor, aqueduct, cistern, theatre, rock-cut tombs, baths, fish tanks, etc., are evidence of the political and economic importance of the Roman province of Cilicia for the Roman empire. Ancient harbor infrastructures and Roman fish tanks, as well as rock-cut pool, are suitable to detect the vertical stability of a coastal area. Along the coastal area of Narlıkuyu, we found a fish tank cut into the local marly limestones, which shows partially eroded walls in their uppermost part. For their function, the Roman fish tank were built in the intertidal zone, to be alternatively in connection with the sea following the tide. For this reason, those Roman coastal infrastructures have been used as good sea-level indicators. According to Evelpidou et al. (2012), the Roman fish tank that is possible to find along the Mediterranean coasts were built in the time span 100 BC – 200 AD.

The Narlıkuyu Roman fish tank, which shows the external walls at an elevation of about 1 m (Fig 2.10), is no more connected with the sea, showing uplift of the Narlıkuyu coastal area in the last two millennia. Based on the predicted sea-level in the eastern Mediterranean by Lambeck & Purcell (2005), the past sea-level in the Narlıkuyu area 2000 years ago was - 0.5 m, indicating a total uplift of the fish tank of ca. 1.5 m. However, uncertainties in calculating the vertical movement of the area are from the partial erosion of the Roman

infrastructure that prevent to measure any precise elevation of the structure correlatable with the sea-level during the activity of the fish tank. In addition, considering the correction for the uncertainties due to the tidal range (± 0.2 m), it is possible to estimate uplift rates between 0.65 to 0.85 mm/yr. Even though the big uncertainties with this estimation, the Narlıkuyu Roman fish tank is a further evidence of the Holocene uplift of the CAP southern margin.



Figure 2.10: Roman fish tank (100 BC – 200 AD) cut into marly limestones along the coastal area of Narlıkuyu. The disconnection with the sea (the tank is partially full of fresh water by the rain) is an evidence of uplift of the Narlıkuyu coastal area.

2.4 Discussions and conclusions

Geological structures studied along the coasts of the CAP southern margin gave important information about the recent tectonic history of the study area. In the literature, it has been emphasized that for detecting uplift rates the marine notches must be interpreted with a great attention since it is not possible to date them directly (Antonioli et al., 2015). In addition, they may not always be the direct indicators of tectonic movements because they may not be clustered around seismogenic zones (Boulton and Stewart, 2013). Keeping this into account, our investigation along the CAP southern margin, which took into account not only marine

notches but also algae trottoirs, beachrocks, alluvial deposits, and archaeological coastal infrastructures, shows indisputable late Holocene uplift (Table 2.2 and Fig.2.7). To make a regional comparison, the general figure of the vertical movements in southern Anatolia that was reported from the beachrocks in southern Turkey and Cyclades from Desruelles et al. (2009) is also supported by our study on the beachrocks of Aydınçık. In general, looking at the different elevation of the marine notches affecting the coastal area of the CAP southern margin, it is possible to recognize three main events of sea-level highstand, responsible for the development of marine notches at around 1 m, 2 m and 5-6 m of elevation (Fig. 2.6). Considering our AMS ^{14}C datings, the marine notch observed at around 1 m a.s.l. along the coastal area of Yeşilovacık (1 m), Mavikent (0.85 m), Eğribük (0.80 m), and Narlıkuyu (0.75 m) should be related to a sea-level highstand occurred at ca. 1000 yr BP. The decimal differences in the heights of the marine notches in the different areas of investigation could be linked to vertical offsets due to the active faults that in the study area clearly cut late Holocene sea-level markers. Using the late Holocene sea-level predicted by Lambeck & Purcell (2005) as sea-level reference for the coastal area of the CAP southern margin, in the last 4 ka the uplift rate of the study area has been ca. 1 mm/yr. This result from our AMS ^{14}C datings is in agreement with the GPS vertical velocity field measured from the GPS station in Silifke, which shows uplift between 0.8 to 1.4 mm/yr (Serpelloni et al., 2013). The late Holocene uplift of the CAP southern margin is coupled with a daily intense seismic activity along a NE-SW tectonic lineament. Although the present seismic activity is mainly due to earthquakes with $M < 3.0$, in the historical archives there are evidence for destructive earthquakes along the coastal area of southern Anatolia (e.g., Corycus, 4th century; Elaioussa-Sebaste, 7th century).

The occurrence of extensional faults that cut through late Holocene sea-level markers is an evidence of active tectonics linked to the local seismicity of the study area, even capable to give destructive earthquakes. For this reason, we propose to add in the Active Fault Map of

Turkey a seismogenic active fault zone, as a fault strand of the Ecemiş Fault in its southwestern prolongation. In the study area, the main fault zone that is known along this active tectonic lineament is the Ecemiş-Kozan Fault (Aksu et al 2014a), which should have its main deformation zone in the offshore domain (Figs. 2.1 and 2.2).

Acknowledgements

This study is part of the Anatolian Plateau Climate and Tectonic Hazards (ALeT) project, funded by the European Union Marie Curie Actions “Initial Training Networks” program (FP7-PEOPLE-2013-ITN)” (grant agreement number 607996). I would like to thank Prof. Dr. Celal Şengör for his support and especially Prof. Dr. Bilal Haq for his valuable scientific contributions.

Table 2.2: Radiocarbon datings (AMS ^{14}C) of samples as sea-level indicators and uplift rates calculated based on predicted past sea-levels (Lambeck & Purcell, 2005)

Sample ID	Coordinates	Elevation (a.s.l.)	Sample material	Further observations	Paleodepth correction	Past sea-level (Lambeck & Purcell, 2005)	Conventional Radiocarbon Age (yrs)	Calibrated Radiocarbon Age (yrs)	max. & min uplift rates (mm/yr)	Locality
AYA 1	N 36 29' 00.1" E 34 10' 41.6"	1.2 m	Patella sp.	breccia level in a Roman pool	± 0.2 m	-0.25 m	1347 $\pm$ 40	BP 1166 to 852	1,07-0.98	Ayaş
AYA 4	N 36°28' 49" E 34°10' 33"	0.5 m	Littorina (living)	upper part of a present day notch	-	-	Present day	-	-	Ayaş
AYD 7	N 36°09'07.91" E 33°20'45.39"	0.35 m	Shell fragment	beachrock	± 0.2 m	-1.35 m	BP 3710 $\pm$ 30 BP	BP 3840 to 3555	0.49-0.39	Aydıncık
AYD 7a	N 36°09'07.91" E 33°20'45.39"	0.35 m	Shell fragment	beachrock	± 0.2 m	-1.25 m	3560 $\pm$ 30 BP	BP 3640 to 3375	0,49-0,41	Aydıncık
AYD 8	N 36°09'07.91" E 33°20'45.39"	0.4 m	Shell fragment	beachrock	± 0.2 m	-1.25 m	3580 $\pm$ 30 BP	BP 3675 to 3390	0,5-0,39	Aydıncık
AYD 9	N 36°09'07.91" E 33°20'45.39"	0.35 m	Shell fragment	beachrock	± 0.2 m	-1.5 m	3830 $\pm$ 30 BP	BP 3985 to 3690	0,51-0,41	Aydıncık
AYD 9a	N 36°09'07.91" E 33°20'45.39"	0.35 m	Shell fragment	beachrock	± 0.2 m	-0.9 m	3260 $\pm$ 30 BP	BP 3320 to 2990	0,43-0,32	Aydıncık
YAN 1	N 36° 09' 11.4" E 33° 29' 35.3 "	1.4 m	bone	Alluvial deposists	± 0.2 m	-0,25 m	1270 $\pm$ 30 BP	BP 1280 to 1175	1,46-1,13	Yanışlı
YAN 2	N 36° 09' 11.4" E 33° 29' 35.3 "	1.4 m	teeth	Alluvial deposists	± 0.2 m	-0,25 m	1190 $\pm$ 30 BP	BP 1180 to 1055	1,57-1,29	Yanışlı
YAN 3	N 36° 09' 11.4" E 33° 29' 35.3 "	1.4 m	teeth	Alluvial deposits	± 0.2 m	-0,25 m	1060 $\pm$ 30 BP	BP 1005 to 930	1,84-1,44	Yanışlı
NAR C1	N 36° 26' 57.0" E 34° 07' 38.7 "	1.9 m	Shell fragment	Marine breccias/conglomerate	± 0.2 m	-0.70 m	2830 $\pm$ 30 BP	BP 2750 to 2480	1,02-0,87	Narlıkuyu
NAR C2	N 36° 26' 57.0" E 34° 07' 38.7 "	1.9 m	Shell fragment	Marine breccias/conglomerate	± 0.2 m	-0.80 m	2990 $\pm$ 30 BP	BP 2940 to 2715	0,99-0,85	Narlıkuyu
EGR 1	N 36° 15' 18.8" E 33° 48' 31.7 "	0.9 m	Shell fragment	Marine conglomerate/beachrock?	± 0.2 m?	-0.5 m	2470 $\pm$ 30 BP	BP 2130 to 1965	0,75-0,56	Eğribük

EGR 2	N 36° 15' 18.8" E 33° 48' 31.7 "	0.9 m	Shell fragment	Marine conglomerate/beac hrock?	± 0.2 m?	-3.25 m	5200 ±30 BP	BP 5575 to 5445	0,78-0,71	Eğribük
AYD 11	N 36° 08' 37.2" E 33° 21' 30.3 "	1 m	Marine carbonate	Algae trottoir	± 0.1 m	-0.35 m	1880 ±30 BP	BP 1610 to 1345	0,90-0,78	Aydıncık
AYD 12	N 36° 08' 37.2" E 33° 21' 30.3 "	0.60 m	Marine carbonate	Algae trottoir	± 0.1 m	-1.20 m	3620 ±30 BP	BP 3715 to 3440	0,51-0,46	Aydıncık
TEK 1	N 36° 08' 06.9" E 33° 07' 50.6 "	0.45 m	Shell fragment	beachrock	± 0.2 m	-0.30 m	2070 ±30 BP	BP 1850 to 1555	0,51-0,30	Tekeli

CHAPTER 3

NATURAL HAZARDS ALONG THE COASTS OF THE SOUTHERN FLANK OF THE CENTRAL ANATOLIAN PLATEAU

Table of Contents

3.1 Introduction.....	77
3.2 Natural disasters in the Eastern Mediterranean as recorded in the archives.....	78
3.3 Storm vs. tsunami boulders.....	84
3.4 Different boulder transportation modes	87
3.5 Methodology.....	89
3.5.1 Identification and measurements of out-of-place boulders.....	91
3.5.2 Classification of the out-of-place boulders (storm vs tsunami).....	96
3.6 Results.....	96
3.6.1 Narlıkuyu.....	96
3.6.2 Yeşilovacık.....	98
3.6.3 Aydıncık.....	100
3.7 Discussion	102
3.8 Conclusion.....	104
Acknowledgements.....	105

3.1 Introduction

The Anatolian plate as a geodynamically very active area in between the collision zone of Africa, Eurasia and Arabian plates, has been built continuously by several geodynamic processes for millions of years such as continental collision (Bozkurt and Mittwede, 2001; Jolivet and Faccenna, 2000; Keskin, 2003; McQuarrie and Van Hinsbergen, 2013; Okay, 2008; Şengör and Yılmaz, 1981; van Hunen and Allen, 2011), plateau uplift (Bartol and Govers, 2014; Bartol et al., 2012; Cosentino et al., 2012; Göğüş and Pysklywec, 2008; Keskin, 2007; Radeff et al., 2015; Schildgen et al., 2012; Schildgen et al., 2012; Şengör, 2003; Şengör et al., 2008), oceanic subduction (Dercourt et al., 1986; Okay et al., 2001; Parlak et al., 2002; Parlak and Robertson, 2004; A. Robertson et al., 2004; Robertson et al., 2004; Robertson and Dixon, 1984; Stampfli, 2000; Şengör and Yılmaz, 1981) and escape tectonics (Dewey and Şengör 1979; Şengör et al 1985; Burke and Şengör 1986; Barka and Reilinger 1997; Jaffey and Robertson 2004; Faccenna et al 2006). In present, the high seismicity of the region is due to the activity of those geodynamic processes in the Eastern Mediterranean (Burton et al 1984; Barka and Reilinger 1997; Meier et al 2007; Jenkins et al 2013; Geological Survey Department of Cyprus 2014). As a result of those geodynamic processes in the region several main tectonic features have been originated such as the Aegean-Cyprean subduction zone, the North Anatolian fault zone, the East Anatolian fault zone and the Dead Sea fault zone, which are the main controlling elements of the seismogenicity of the Eastern Mediterranean (Burton et al., 1984; Jenkins et al., 2013; Salamon et al., 1996). The active geodynamic processes of the region resulting in sloppy submarine morphology, strongly could be the responsible for the natural hazards at the coastline in terms of tsunamis. Based on the historical archives, some studies show that the activity of those tectonic elements is responsible for both earthquakes and tsunamis with some severe damages (Altınok and Ersoy, 2000; Yolsal et al., 2007). Here, we tried to examine the geological

evidence of tsunami hazards in different areas along the southern coastline of the Central Anatolian Plateau linked to the margin uplift-related morphology changes in the submarine realm and importance of coastal geomorphology considering the transportation modes of tsunamites to better understand the trigger mechanisms of the tsunamis along the southern Anatolian Plateau margin (Fig. 3.1).

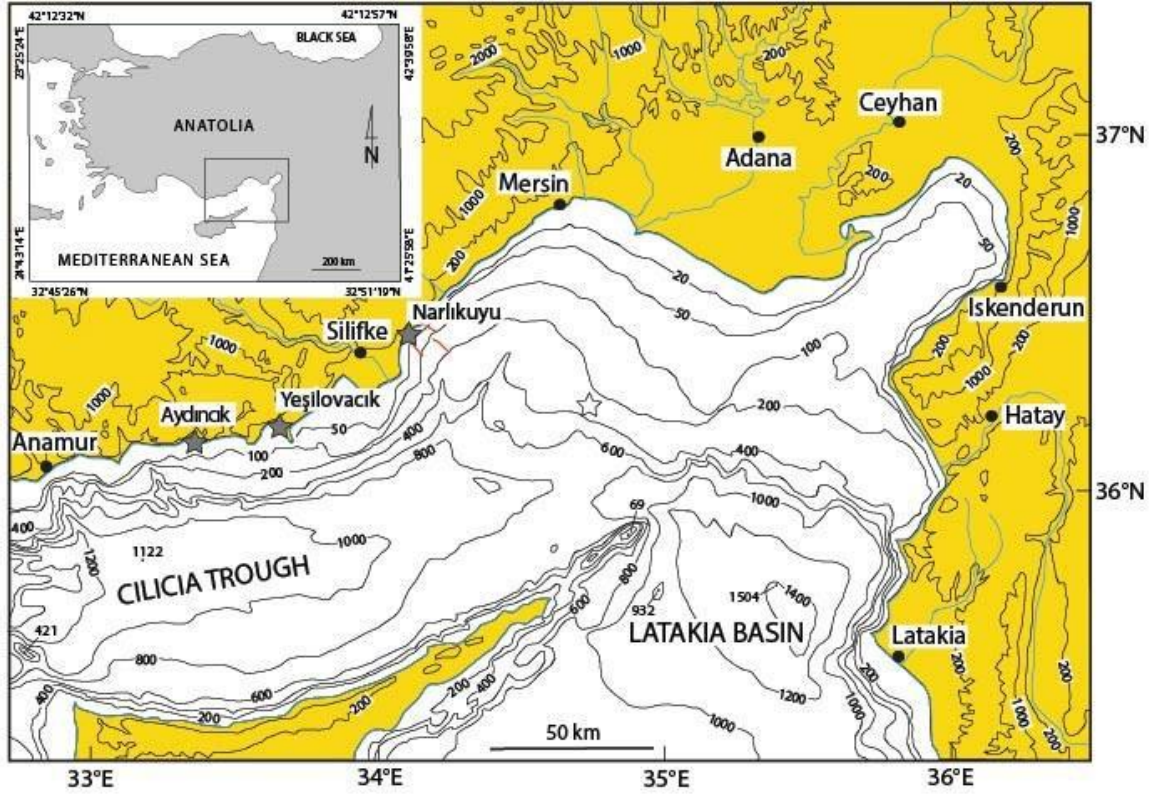


Figure 3.1: Map showing the coastal area of the Levantine Sea between the Cilicia trough and the Latakia Basin. Gray stars show the location of Aydıncık, Yeşilovacık and Narlikuyu villages, where out-of-place boulders discussed in the text were found. Bars offshore Narlikuyu (a, b, c) are the traces of the high-resolution seismic reflection profiles showed in Fig. 3.2. Isobaths and elevation contour lines are in meters. Redrawn and modified from Okyar et al. (2005).

3.2 Natural disasters in the Eastern Mediterranean as recorded in the archives

As reported in historic chronicles or as inferred from geological observations, due to the active tectonic and geodynamic processes in the area, the Eastern Mediterranean region has been affected by destructive and tsunamigenic earthquakes from at least 1410 ± 100 BC (Altınok et al., 2011; Papadopoulos et al., 2014; Yolsal et al., 2007). The closest seismogenic

zones to the southern coast of central Anatolia capable of generating $M > 6.0$ earthquakes include two onshore zones (hence with low tsunamigenic potential), namely the East Anatolian fault [Antakya (ancient Antioch), İskenderun, Misis] and the Ececiş-Kozan fault (Kozan, Ceyhan) (Tables 1, 2), and a third offshore seismogenic zone with higher tsunamigenic potential, namely the subduction zone of the Cyprus arc (Tables 3.3, 3.4) (Noller et al., 2011). However, there are more tsunamigenic zones in the vicinity of the southern margin of Anatolia that can generate tsunamis capable of affecting its coastal area, including the Hellenic arc and the Dead Sea fault. Along the Cyprus and Hellenic arcs, earthquakes mostly occur in the submarine environment; therefore, their co-seismic displacements are expected to generate strong tsunamis capable of propagating far from the epicentral area. In contrast, locally strong tsunamis associated with the onland strike-slip tectonics of the Dead Sea fault system are mainly due to landslides affecting the eastern margin of the Levantine Basin (Papadopoulos et al., 2014; Salamon et al., 2009, 2007). These local tsunamis tend not to propagate for long distances due to the strong sea-wave attenuation of landslide-generated tsunamis (Papadopoulos and Kortekaas, 2003; Yolsal et al., 2007).

Table 3.1: Historical earthquakes in southern Turkey (from Boğaziçi University, Kandilli Observatory and Earthquake Research Institute and Poirier & Taher, 1980).

Date	Epicenter	Mercalli scale	Location
69 BC	36°25'-36°10'	IX	Antakya, Syria
13.12.115	36°25'-36°10'	IX	Antakya
245	36°25'-36°10'	X	Antakya
334	36°25'-36°10'	IX	Antakya, Beirut, Cyprus
14.09.458	36°25'-36°10'	IX	Antakya, Northern Syria
10.09.506	36°25'-36°10'	IX	Antakya, Samandağ
29.05.526	36°25'-36°10'	IX	Antakya, Samandağ
29.11.529	36°25'-36°10'	IX	Antakya
30.09.587	36°25'-36°10'	IX	Antakya (60.000 deaths)
757	36°57'-35°35'	VII	Misis
803	36°57'-35°35'	VIII	Misis
835	36°50'-36°15'	VIII-IX	Antakya
845	36°50'-36°15'	IX	Antakya
08.04.859	36°25'-36°10'	IX	Antakya, Lattakia
867	36°25'-36°10'	IX	Antakya
17.09.1091	36°50'-36°15'	IX	Antakya
10.08.1114	36°50'-35°50'	IX	Ceyhan, Antakya, Maraş (Tsunami)
1268	37°35'-35°80'	IX	Kozan, Ceyhan (60.000 deaths)
30.12.1408	36°50'-36°15'	X-XI	Antakya

13.08.1822	36°40'-36°20'	X	Antakya, İskenderun (20.000 deaths-tsunami)
02.04.1872	36°25'-36°10'	IX	Antakya, Samandağ (1800 deaths)

Table 3.2: Earthquakes of the last century in southern Turkey (from: Kalafat et al 2009 and Boğaziçi University, Kandilli Observatory and Earthquake Research Institute)

Date	Epicenter (N,E)	Magnitude	Location
02.17.1908	37°40'-35°80'	6.0	Işıkli-Kozan, Adana
04.30.1911	36°00'-30°00'	6.1	Antalya Offshore
06.05.1927	36°00'-31°00'	6.2	Antalya Offshore
20.03.1945	37°11'-35°70'	6.0	Ceyhan-Misis (Adana)
30.04.1975	36°19'-30°73'	6.0	Gulf of Antalya
27.06.1998	36°87'-35°78'	6.3	Hasanbeyli-Sarıçam, Adana
28.12.2013	36°00'-31°30'	6.0	Gulf of Antalya

Table 3.3: Historical earthquakes in Cyprus (from: Geological Survey of Cyprus)

Date	Epicenter	Mercalli scale	Location
15 BC	Southwest Pafos	IX	Pafos, Kourion
76 AD	Southeast of Cyprus	IX	Salamis, Pafos, Larnaca
342 AD	Southwest coast	X	Pafos
365 AD	Southwest of Cyprus	VIII	South coasts of Cyprus
394 AD	East of Cyprus	VII	Pafos, Salamis
1144	West of Cyprus	VII	Pafos
1183	Near Pafos	VIII	Pafos
03.05.1222	Southwest of Cyprus	IX, VIII	Pafos, Lemesos
25.04.1491	Cyprus	IX, VIII, VII	Mesaoria Valley, Nicosia, Lemesos
25.04.1567	South of Cyprus	VII	Lemesos
10.12.1741	Near Famagusta	VIII	Famagusta
29.06.1896	190 km South of Cyprus	VIII, VII	Akrotiri, Lemesos

Table 3.4: Earthquakes of the last century in Cyprus (from: Geological Survey of Cyprus).

Date	Epicenter	Magnitude	Location
29.09.1918	35°10'-34°80'	6.3	Lemesos
18.02.1924	34°80'-34°30'	6.0	Larnaca
10.09.1953	34°72'-32°24'	6.0	Pafos
10.09.1953	34°80'-32°78'	6.1	Pafos
09.10.1996	34°53'-32°10'	6.5	SW-Cyprus

Given that the coastal area of the Silifke district is characterized by a narrow and relatively steeply sloping shelf, local tsunamis related to offshore slumps and landslides, possibly triggered by either onland or offshore earthquakes from the Mersin seismic district, cannot be ruled out. Indeed, in the offshore realm of the study area, seismic facies related to gravity mass movements (e.g., slumps and landslides) affecting the Holocene sequence were interpreted from high-resolution seismic reflection profiles (Fig 3.3, Okyar et al 2005). Moreover, on Narlıkuyu shorelines, we have found normal fault that cut Holocene structures such as tidal pools as a clue of the tectonic activity of the region that is correlable with the major offshore faults that were detected by previous researchers (Fig 3.2 and 3.3).

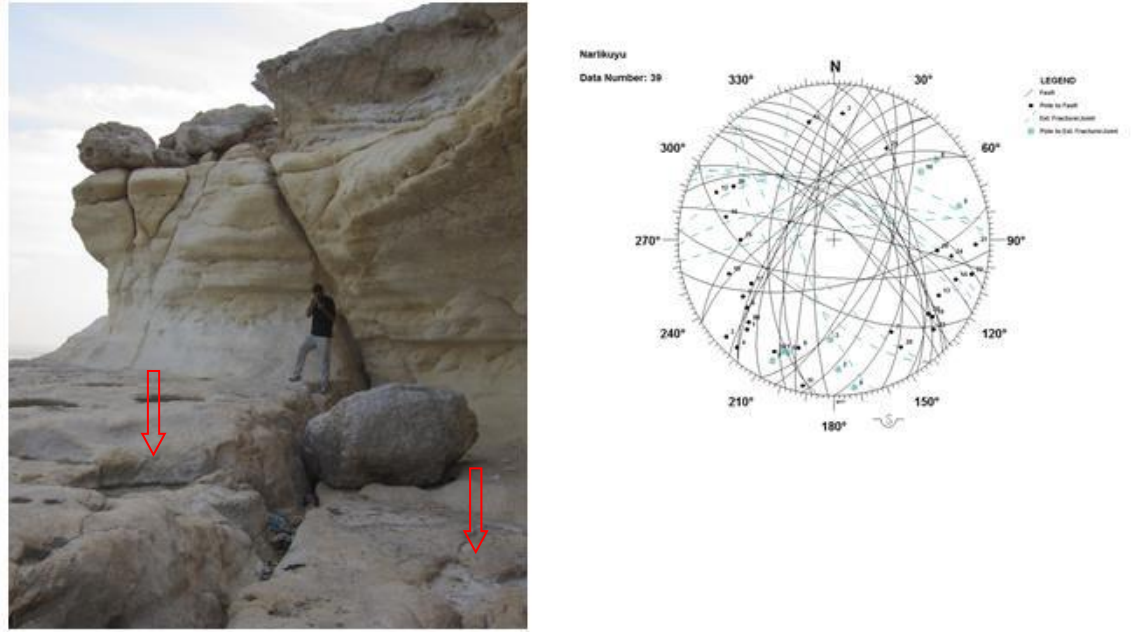


Figure 3.2: Photo showing the NE-SW trending normal fault at Narlıkuyu shoreline with ~40 cm displacement of Holocene tidal pools (shown with red arrows). The figure shows the fault plane and fracture measurements.

In the tsunami catalogs for the Eastern Mediterranean (Altınok et al., 2011; Altınok and Ersoy, 2000; Guidoboni and Comastri, 2005; Papadopoulos et al., 2014; Yolsal et al., 2007) there are only a few records of tsunamis that affected the southern Anatolian coast.

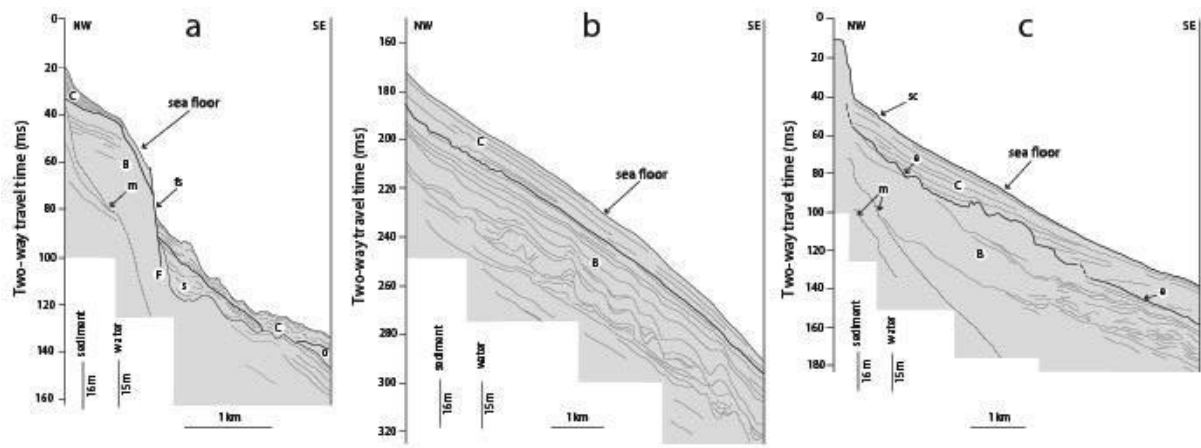


Figure 3.3: Line-drawing of high-resolution seismic profiles acquired offshore Narlikuyu (for their location see Fig. 3.1). Legend B pre-Holocene sequence, C Holocene sequence, F fault; m multiples, sc erosional scarp, fs fault scarp, s slope front fill, o onlap fill, e pre-Holocene erosional surface. It is worth noting that slumped masses affect both pre-Holocene (B) and Holocene (C) sequences. The original seismic profiles are from Okyar et al. (2005).

In our study area, the most ancient records of tsunamis are from 140 and 524–525 BC (Altınok and Ersoy, 2000). Between these early events and the medieval age, there are no records of tsunamis in southern Turkey. During the medieval period (476–1492 AD), a strong tsunami related to an earthquake that happened either on 12 March 1036 or on 11 March 1037 (Guidoboni and Comastri 2005) struck Cilicia and southern Turkey. Subsequent tsunami events recorded in southern Anatolia occurred on 16 November 1403, in 1489, and most recently on 10 September 1953 (Altınok and Ersoy, 2000). Although Altınok and Ersoy (2000) assigned a very low grade of reliability (Rel = 1) to this last event, more details were subsequently provided by Altınok et al. (2011), including parameters of the source area such as: (1) fault type (right-lateral transpressional); (2) fault length (25 ± 5 km); (3) max depth (20 ± 5 km); (4) fault displacement (0.35 ± 0.1 m); (5) max magnitude (6.2 ± 0.1); and a higher reliability grade (Rel = 3–4). However, according to Fokaefs and Papadopoulos (2007), the 10 September 1953 Cyprus earthquake was associated with a small tsunami along the coast of Paphos that caused no damage, suggesting that this event was incapable of producing appreciable run-up along the southern Anatolian coast.

To evaluate tsunami wave propagation in Eastern Mediterranean, Yolsal et al. (2007) simulated the tsunamis generated by two major historical earthquakes that affected Eastern Mediterranean in 1222 AD and 1303 AD. The 11 May 1222 earthquake ($M \sim 7.5$), which is reported as 5 May 1222 in Table 2, is related to the activity of the SW Cyprus arc. Macroseismic observations related to this earthquake have been reported from the coastal area all around the Eastern Mediterranean, including the southern coast of Anatolia (Guidoboni and Comastri 2005). The propagation of the tsunami wave generated by this earthquake, simulated by using nonlinear shallow-water theory (Yolsal et al., 2007), indicates the arrival of water surface perturbations along the southern Anatolian coast 60 min after the earthquake, with water surface elevations of a few centimeters (~ 4 cm) and maximum wave heights of 8 cm (Yolsal et al. 2007).

The simulation of the tsunami generated by 8 August 1303 earthquake ($M \sim 8.0$) shows slightly stronger effects on the southern Anatolian coast (Yolsal et al. 2007). This earthquake occurred along the subduction zone of the Hellenic arc south of Crete. It was one of the largest, best-documented seismic events in the history of the Mediterranean area and generated very destructive tsunami waves that hit all the Eastern Mediterranean coasts (Guidoboni and Comastri 2005). The simulation of the tsunami wave propagation for this earthquake indicates that the water surface perturbation reached the southern Anatolian coast near Aydınçık 82 min after the main shock, with water surface elevation of 8 cm and maximum wave heights of 14.6 cm (Yolsal et al. 2007). Similar water surface perturbation was observed from the time series of the sea level data recorded by DART (Deep-ocean Assessment and Reporting of Tsunamis) buoys during the propagation of the 2009 Samoa tsunami, which was triggered by a $M_w = 8$ earthquake that occurred off the Samoa Islands (Kowalik, 2012). This earthquake generated a tsunami with height of 2.7 m near the coast of American Samoa Islands in Pago Pago (Kowalik, 2012).

3.3 Storm vs. tsunami boulders

According to the literature, the transport and deposition of coastal boulders could involve either tsunami (Mastronuzzi and Sansò 2000; Bryant and Nott 2001; Kelletat and Schellmann 2002; Scheffers 2002, 2004; Whelan and Kelletat 2005; Goff et al. 2006, 2012; Morhange et al. 2006; Bryant and Haslett 2007; Scicchitano et al. 2007; Goto et al. 2007; Scheffers et al. 2008; Bourgeois and MacInnes 2010; Paris et al. 2009; Vella et al. 2011; Papadopoulos et al. 2014) or storm waves and hurricanes (Massel and Done 1993; Fletcher et al. 1995; Williams and Hall 2004; Hall et al. 2006; Spiske et al. 2008). Experimental studies motivated by proposed hydrodynamic equations (e.g., Goto and Imamura 2007) have demonstrated that the initial transport and the mode of a boulder to be transported from the wave (by sliding, rolling, and saltation) vary with the flow velocity, bottom friction, and the shape and weight of the boulder. Nandasena et al. (2011), revising Nott's hydrodynamic equations (Nott 1997; 2003), defined a boulder transport histogram to represent the range of flow velocities necessary for initial transport by sliding, rolling, or saltation of boulders from different case studies (e.g., Australia, SE Italy, Iceland, and Sumatra). The boulder transport histogram is a valuable tool to help reconstruct the flow velocity of past high-energy events such as tsunamis or storm surges, although it does not help to discriminate among different high-energy events. Given that no definitive criteria have been suggested to distinguish tsunami boulders from storm boulders, the genetic origin of such deposits remains the most controversial issue in studying out-of-place boulders along coasts (Noormets et al. 2004; Imamura et al. 2008; Goto et al. 2009). For this reason, the origin of the coastal boulder deposits (storm vs tsunami) is still highly debated (e.g., Dawson et al. 2008).

The term “tsunamite” in the literature is being identified as, sediments not only transported by a tsunami itself but also deposited by a tsunami-induced current (Shiki and Yamazaki, 2008). However, it is very common to confuse storm and tsunami deposits because

of the absence of a definitive distinguishing criteria (Kortekaas and Dawson, 2007; Paris et al., 2011; Weiss, 2012). In general, on-land tsunami sediments are characterized by out-of-size massive sand, silt and fine gravel (Dawson, 1999; Scheffers and Kelletat, 2003; Dawson and Stewart, 2007; Morton et al., 2007; Mastronuzzi, 2010; Goff et al., 2012; Papadopoulos et al., 2014) . On the other hand, it has been stated that on the rocky coasts, small sized sediments may not be preserved due to physical processes (e.g. erosion) related to the coastal geomorphology, instead, tsunami boulders may remain (Dawson, 1994; Dawson and Stewart, 2007; Scheffers, 2008; Bourgeois, 2009; Maouche et al., 2009; Pignatelli et al., 2009).

In the developing study of tsunamite deposition, it is important to clarify several parameters such as the position of boulders relative to source (altitude of boulder accumulation), size, weight, form, distance to the shoreline and accumulation surface for a better distinguish between tsunami and storm boulders (Scheffers, 2008; Pignatelli et al., 2009; Paris et al., 2011). To make an accurate description of the Aydıncık's out-of-place boulders, the features of the outcrop that we considered are listed below:

- a) the boulders display some imbrications with seaward dips;
- b) the majority of boulders have their A-axes aligned parallel to the shore or in one particular direction;
- c) the boulders are scattered on the underlying lithology with distance inland or upslope;
- d) the boulders are deposited in preferential locations;
- e) marine fauna (oysters) attached to boulders above the highest tide.

Those features show quite fit to the Nott's tsunami boulder criteria (1997) that were used to distinguish storm and tsunami boulders in Great Barrier Reef, Australia. Those signatures together with the absence of a local cliff in the coastal area where we found the Aydıncık's

out-of-place boulders allow us to eliminate the possibility to consider such boulders as due to a cliff collapse and to take into account a high energy-wave transportation mode.

Moreover, some observations from recent historical tsunamis, such as the 2004 Indian Ocean tsunami (Goto et al. 2007) and the 1975 Kalapana tsunami (Hawaii, Goff et al. 2006), could help us to discriminate between those different high-energy waves (storm vs tsunami) in interpreting coastal boulder deposits. On Pakarang Cape (Thailand), abundant m-scale boulders appeared on the tidal bench of its western shore just after the 26 December 2004 Indian Ocean tsunami (Goto et al. 2007). According to the authors, no boulders were observed from a satellite image prior to the tsunami, allowing them to indisputably relate the boulders to the 2004 Indian Ocean tsunami. In addition, no boulders were found on land after the tsunami, indicating that the hydraulic force of the tsunami wave rapidly dissipated on reaching the land. It is worth noting that after more than 1 year of observation (September 2005– November 2006), all the boulders along a monitored transect crossing the tidal bench of Pakarang Cape were at the same position, indicating that high waves generated by the annual monsoon did not have sufficient energy to move those boulders from their post-tsunami position (Goto et al., 2007). The 1975 Kalapana tsunami (Hawaii, Goff et al 2006) deposited a fining-inland sequence of basalt boulders and carbonate sand. According to the authors, the fining-inland sequence of high-energy deposits is characteristic of tsunami sediments, in contrast to a more random pattern associated with storm-wave deposits. The same difference was noted at sites where a comparison between tsunami and storm deposits was made on the same coastal area (e.g., Nanayama et al 2000; Goff et al 2004; Tuttle et al 2004). In contrast to these results, no landward fining of boulder sizes was detected after the 26 December 2004 Indian Ocean tsunami at Pakarang Cape (Thailand, Goto et al. 2007) nor in Lhok Nga Bay (northwest Sumatra, Paris et al 2009).

Additional clues to distinguish storm and tsunami boulders are from the Caribbean Sea, where coarse littoral sediments and boulders occurring along the coasts of the Leeward Netherlands Antilles were interpreted as due to Holocene tsunamis (Scheffers, 2004). According to the author, field evidence and radiocarbon dates obtained from vermetids encrusting boulders indicate that storm or hurricane deposition can be excluded. This conclusion is strengthened by the rather limited impacts of heavy surf conditions related to Hurricane Lenny on the southwestern coasts of the Leeward Netherlands Antilles. Hurricane Lenny was a rare storm with peak winds of 250 km/h, which in November 1999 passed close to the islands. The most impressive effect of Lenny along the coasts was the accumulation of at least 10,000 t of coral rubble with a grain size distribution of centimeters to decimeters. No boulders were transported onshore (Scheffers, 2004).

3.4 Different boulder transportation modes

In the literature, there are only a few number of studies regarding the beachrock slab geomorphology and its importance for high-energy wave transportation (Taveuni Island, Fiji, Etienne and Terry 2012; Luda Island, Taiwan, Lau et al 2015). Although given studies are focused on high-energy waves that are produced by typhoons and/or cyclones, for the facts that are stated above regarding the low occurrence possibility of storms in the Eastern Mediterranean, we are not interpreting the presence of out-of-place boulders in Aydıncık district related to storms, but a tsunami event that might be triggered by an earthquake and /or submarine landslide.

Concerning the transportation modes of boulders, although it still remains controversial, there are several field observations, experimental and numerical studies have been applied (Nott, 2003a; Noormets et al., 2004; Imamura et al., 2008; Barbano et al., 2011; Costa et al., 2011; Etienne et al., 2011; Nandasena et al., 2011; Liu et al., 2014). The transport

mechanisms of coastal boulders have been classified in different modes such as sliding, rolling and saltation (Etienne et al., 2011; Nandasena et al., 2011). Even though it is highly controversial matter to interpret the origin of coastal boulders of past events as reported before (e.g. tsunamis, storm waves, and hurricanes), current knowledge is based on the recent tsunami events. For example, after the 26 December 2004 tsunami in Pakarang Cape (Thailand), m-scale scattered boulders have been observed on the shoreline that were 1.3 m^3 in volume in average, with an estimated weight of 2.1 t and an ellipsoidal shape (Goto et al. 2007). The well-rounded and ellipsoidal shapes of the boulders suggest that they were transported by rolling and/or saltation rather than sliding. According to the boulder transport histogram of Nandasena et al. (2011), in comparison with the sliding mode, they required higher flow velocities for their initial transport. For example, a calcareous boulder $1.3 \times 1.1 \times 0.85 \text{ m}$ from a seawall in Sumatra requires a minimum flow velocity of 4.1 m/s to start sliding, whereas it needs a minimum flow velocity of 12 m/s to start moving by saltation (Nandasena et al., 2011). In total three types of boulders have been described based on their initial positions: submerged, subaerial and joint-bounded (Nott, 2003b; Nandasena et al., 2011). Furthermore, it is possible to indicate that since the only upper part of the boulders have been covered with marine mollusk biomarkers, this could be related to that the lower part of the boulders were not exposed to marine fauna because they were joint-bounded. Nandasena et al. (2011) states that if a boulder has a joint-bounded initial position, the initial motion of the block should be saltation/lifting (Fig 2B in Nandasena et al., 2011), however progressive motion should not necessarily be the same as initial motion. The form of the boulders with preserved edges that explained above, leads us to focus on the initial motion of the boulder is likely to be lifting instead of saltation.

3.5 Methodology

To understand the origin and the mechanism of the boulders, we measured their sizes and calculated their weights related to their composition materials that are given in the following subtitle. To identify their mechanism of the initial motion and transportation of the boulders, we have applied the equations that are revised by Nandasena et al. (2011) after Nott (2003b).

a) Boulder transport will be initiated when, in the case of a joint-bounded boulder:

$$F_l \geq F_r \cos \theta + \mu_s F_r \sin \theta$$

$$u^2 \geq \frac{2 \left(\frac{\rho_s}{\rho_w} - 1 \right) g c (\cos \theta + \mu_s \sin \theta)}{C_l}$$

b) Boulder transport will be initiated when, in the case of a subaerial boulder:

$$u^2 \geq \frac{2 \left(\frac{\rho_s}{\rho_w} - 1 \right) g c (\cos \theta + \left(\frac{c}{b} \right) \sin \theta)}{C_d \left(\frac{c^2}{b^2} \right) + C_l}$$

However, we have to remember to consider the type of the initial motion of the boulders. As stated in Nandasena et al. (2011), for joint-bounded boulders sliding or rolling could not be possible because drag and inertia forces are not imposed on the boulder effectively because all the faces of the boulder is covered by adjacent rock faces. For this reason for the joint-bounded boulders the initial motion is required to be lifting/saltation. In that case, the formulation (a) that is given above is used for Aydıncık locality boulders to make a comparison with the boulder transport histogram of Nandasena et al. (2011). For the submerged/subaerial boulders sliding, overturning/rolling or saltation type initial motion should be considered. The formulations that are used are the following:

a) initial motion will be sliding when:

$$u^2 \geq \frac{2 \left(\frac{\rho_s}{\rho_w} - 1 \right) g c (\mu_s \cos \theta + \sin \theta)}{C_d \left(\frac{c}{b} \right) + \mu_s C_l} \quad (1)$$

b) initial motion will be overturning/rolling when:

$$u^2 \geq \frac{2 \left(\frac{\rho_s}{\rho_w} - 1 \right) g c (\cos \theta + \left(\frac{c}{b} \right) \sin \theta)}{C_d \left(\frac{c^2}{b^2} \right) + C_l} \quad (2)$$

c) initial motion will be saltation when:

$$u^2 \geq \frac{2 \left(\frac{\rho_s}{\rho_w} - 1 \right) g c (\cos \theta)}{C_l} \quad (3)$$

where u: flow velocity; ρ_s : density of boulder; ρ_w : density of sea water; g: gravity; c: height of the boulder; b: intermediate axis of the boulder; Θ : bed slope angle; μ_s : coefficient of static friction; C_l : coefficient of lift; C_d : coefficient of drag.

The constant values applied in the formula are given below (Nandasena et al 2011):

$$C_l = 0.178; C_d = 1.95; \rho_w = 1020 \text{ kg/m}^3; g = 9.81 \text{ m/s}^2; \mu_s = 0.7$$

The boulders with different characteristics of the study area in different localities have been examined using those formulations based on their physical properties and initial

positions before the initial motion. Then, we compared our results with the boulder transport histogram of Nandasena et al. (2011). The results were discussed in Ch. 3.6.

3.5.1 Identification and measurements of out-of-place boulders

To identify the out-of-place boulders, we took the criteria of Nott (1997) as a reference that are explained above. We have measured several boulders in different regions along the coast of southern Anatolian Plateau which are Yeşilovacık, Narlıkuyu and Aydıncık localities in Silifke district (Mersin, Turkey). In Narlıkuyu and Yeşilovacık localities, the coastal area is characterized by a rocky coast with frequent embayments in a narrow and relatively steeply sloping shelf. This rugged morphology is mainly due to Holocene fault scarps related to the active Ecemiş Fault Zone in Mersin-Silifke area (Okyar et al., 2005; Ozel et al., 2007; Öğretmen et al., 2015).

Along Yeşilovacık rocky coast, which is characterized by remnants of marine abrasion platforms at different elevations (from 0.7 up to 7.2 m), we found several remains from marine environments, such as ellipsoidal boulders with marine biomarkers, well-rounded pebbles from shore deposits, and remains of marine mollusks from an infralittoral marine environment (0–40 m depth). From the frequent cm-sized remains of present marine mollusk, we collected balanids, *Ostrea edulis*, *Patella caerulea*, *Conus mediterraneus*, *Fusinus pulchellus*, *Monodonta turbinata*, *Palliolium cf. incomparabile*, and *Cypraeidae* indet. from abrasion platforms at 3.0 and 7.2 m a.s.l. In the study area, in contrast to the frequent occurrence of onshore transported marine mollusk remains, only two places show m-scale boulders with marine biomarkers (e.g., boulders with vermetids and/or pierced by marine perforating bivalves), which required higher-energy waves to be transported onshore.

One place where we found out-of-place boulders is from the coast south of Yeşilovacık, close to the new harbor under construction (36°11'1.6"N, 33°39'36.32"E).

There, the rocky coast is mainly characterized by outcrops of highly deformed shallow-water Triassic limestones and Upper Pleistocene breccias from a base slope detrital fan, which unconformably overlies the Mesozoic carbonates. The Upper Pleistocene breccias exhibit tidal notches and remnants of abrasion platforms at different elevations above the present sea level (Fig 3.6). Ellipsoidal shaped out-of-place boulders with marine biomarkers lie above remnants of abrasion platforms, pointing to onshore transport by storm waves and/or tsunami events. The boulders are characterized by Triassic limestones and Upper Pleistocene breccias with biomarkers such as lithophaga borings (*Petricola lithophaga*) and vermetids (*B. triquetra*) (Fig 3.6), indicating a provenance from an intertidal marine environment.

A second place where we found an onshore transported boulder is in the vicinity of Narlıkuyu (36°26'53.77"N, 34°7'25.43"E) (Figs 3.4 and 3.5). In this area, the step-like shape rocky coast is mainly characterized by outcrops of deeper-water marls and marly limestones, possibly Pliocene in age. These deeper-water deposits are rich in small echinoids and planktonic and benthic foraminifera, and they unconformably lie above middle Miocene shallow-water limestones. Also in this area, some of the ellipsoidal out-of-place boulders show lithophaga borings with *P. lithophaga* as a marker of an intertidal marine environment.

On the other hand, another region with out-of-place boulders is the Aydıncık coast that has a beachrock morphology with gently seaward-dipping conglomeratic slabs which has a different coastal morphology than the other locations (Narlıkuyu and Yeşilovacık) and different boulder characteristics. The modern coast of the location consists of rounded pebbles different from the beachrock lithology. In the region, the out-place-boulders that have been studied are positioned on top of the beachrock slabs which as well have the same lithology of beachrocks (Figs. 3.8 and 3.9). By shape, they are platy, rectangular or triangular and their corners are preserved which is also a clue to examine the transportation mode. The boulders are scattered starting from around 10 to 13 m inland up to 25 m (Fig 3.9). In total, 35 out-of-

place boulders have been measured in this locality and volumes have been calculated (Fig 3.9, Table 3.6). The altitude of the boulders is always less than a meter (avg. 0.33 m) and in general upper part of the boulders contain marine mollusks of an infralittoral marine environment (0-40 m) such as balanids, *Mytilaster minimus*, *Chama gryphoides*, *Bivonia triquetra*, *Ostrea* sp. and also serpulids, which is significant to interpret that the boulders have been transported onshore from offshore domains by high-energy waves (Dawson, 1999; Goff et al., 2012). In the locality, there is only one boulder (AYD-A) that has a different lithology and form which is found at the SW-part of the bay. It is a compact calcareous tufa boulder, consistent with the geology of the locality, with infralittoral marine environment biomarkers such as *Mytilaster minimus*, *Chama gryphoides*, *Bivonia triquetra* and has an ellipsoidal-shape.

Table 3.5: Physical properties of the out-place-boulders in Narlıkuyu (NAR) and Yeşilovacık (YEŞ) localities.

Boulder ID	Density (t/m ³)	Volume (m ³)	Weight (t)	Lithology	Age (cal yrs)
YEŞ 0	2.7	0.816	2.20	Triassic limestone	-
YEŞ 00	2.4	1.289	3.09	Compacted calcareous breccia	-
YEŞ 3	2.7	0.113	0.30	Triassic limestone	13±35 BP
YEŞ 4	2.7	1.004	2.71	Triassic limestone	130±40 BP
YEŞ 5	2.4	3.014	7.23	Compacted calcareous breccia	>1950 AD
NAR 0	2.6	1.143	2.97	Miocene limestone	-

Boulder ID	a (m)	b (m)	c (m)	Volume (m ³)	Density (t/m ³)	Weight (tons)	Lithology	Samples	Age (cal yrs)
A	1,6	1	0,8	1,28	2	2,56	Calcareous tufa	AYD-A	younger than AD 1950
B	2,4	1,85	0,8	3,552	2,3	8,1696	Beachrock	AYD-B	younger than AD 1950
C	1,2	1,1	0,55	0,726	2,3	1,6698	Beachrock		
D	0,65	0,45	0,3	0,08775	2,3	0,201825	Beachrock		
E1	1,8	1,5	0,35	0,945	2,3	2,1735	Beachrock		
E2	2,5	1,35	0,5	0,84375	2,3	1,940625	Beachrock		
E3	2,3	1,35	0,45	1,39725	2,3	3,213675	Beachrock		
E4	2,1	1,1	0,5	1,155	2,3	2,6565	Beachrock	AYD-E4	younger than AD 1950
E5	1,7	1,15	0,4	0,782	2,3	1,7986	Beachrock		
E5a	0,7	0,7	0,2	0,098	2,3	0,2254	Beachrock		
E5b	1	0,95	0,13	0,1235	2,3	0,28405	Beachrock		
E5c	0,75	0,4	0,13	0,039	2,3	0,0897	Beachrock		
E5d	0,9	0,6	0,13	0,0702	2,3	0,16146	Beachrock		
E6	2,15	1,8	0,45	1,7415	2,3	4,00545	Beachrock		
E7	1,3	1,05	0,23	0,31395	2,3	0,722085	Beachrock		
E8	1,6	0,7	1,18	0,2016	2,3	0,46368	Beachrock		
E9	1,4	0,9	0,3	0,378	2,3	0,8694	Beachrock		
E10	1,4	0,9	0,3	0,378	2,3	0,8694	Beachrock	AYD-E10	
E11	1,6	1,4	0,25	0,56	2,3	1,288	Beachrock		
E11a	0,8	0,6	0,25	0,12	2,3	0,276	Beachrock		
E11b	1,2	0,8	0,25	0,24	2,3	0,552	Beachrock		
E12	0,9	0,6	0,4	0,216	2,3	0,4968	Beachrock		
E13	1,6	0,9	0,13	0,1664	2,3	0,38272	Beachrock		
E13a	0,5	0,6	0,1	0,03	2,3	0,069	Beachrock		
E14	0,9	0,7	0,2	0,063	2,3	0,1449	Beachrock		
E15	0,6	0,35	0,13	0,0273	2,3	0,06279	Beachrock		

E16	1,1	0,8	0,25	0,22	2,3	0,506	Beachrock		
E17	0,65	0,45	0,2	0,0585	2,3	0,13455	Beachrock		
E17a	1,2	0,7	0,15	0,126	2,3	0,2898	Beachrock		
E18	1,3	1	0,22	0,286	2,3	0,6578	Beachrock		
E19	1	0,7	0,22	0,154	2,3	0,3542	Beachrock		
E20	1,45	1,15	0,15	0,1250625	2,3	0,28764375	Beachrock	AYD-E20	
E21	1,6	1,6	0,35	0,896	2,3	2,0608	Beachrock		
E22	2,1	1,4	0,3	0,441	2,3	1,0143	Beachrock		
E23	1,5	1,3	0,25	0,4875	2,3	1,12125	Beachrock		

Table 3.6: Physical properties of out-of-place boulders that were measured in Aydıncık district and AMS ^{14}C results.

3.5.2 Classification of the out-of-place boulders (tsunami vs storm)

Eventhough it is still on debate the classificaion of tsunami and storm boulders, recent observations and simulation are strong tools to make this discrimination. Following the boulder transport histogram of Nandasena et al. (2011), based on our measurements of the out-of-place boulders, it is possible to calculate the required wave velocity and height to transport certain masses from offshore to onshore. Although storms originating from the west significantly influence the sea level of the Mediterranean, especially along the coastal area of southern Turkey (Krestenitis et al., 2011), according to the Wind and Wave Atlas of the Mediterranean Sea (WEAORC, 2004), the probability that a storm surge with waves higher than 4 m could affect southern Turkey is negligible (0.1 event per year), whereas 1.5 events/year are expected for storms with waves higher than 2.5 m. Using the formulations that are given in methodology chapter, we were able to classify the origin event of the boulder transportaion of studied localities.

3.6 Results

To evaluate the boulders as tsunamiites that have been found in the study area, we consider the recent studies regarding the storm events in the Eastern Mediterranean and combine with the position of boulders on the coastline using Nott (1997)'s criteria. Along the coastal area of southern Turkey, according to the Wind and Wave Atlas of the Mediterranean Sea, the statistical results show that the occurrence probability of storm surge waves higher than 4 m is 0.1 event per year which is negligible, whereas the probability for storms with waves higher than 2.5 m is 1.5 events per year (WEAORC, 2004).

3.6.1 Narlıkuyu

In Narlıkuyu region, a 1-m-long (ca. 1.3-t) boulder (NAR 0) of Miocene limestone shows lithophaga borings containing *P. lithophaga*. The boulder lies 6 m inland from the

shoreline near the cliff slope, above an abrasion platform cut at 0.8 m a.s.l. within the Pliocene marls and marly limestones (Figs 3.4 and 3.5). The physical properties of NAR 0 (Table 3.5), together with the step-like morphology of the Narlıkuyu coastal area, point to onshore transport by saltation in high-energy waves after the applying of all the three formulations that are given in Ch.3.2. Our calculation result for NAR0 boulder for sliding mode is 2,8 m/s, for overturning/rolling mode is 4,14 m/s and for saltation mode is 8,76 m/s flow velocity is required to initiate the motion. Consequently, by using the boulder transport histogram of Nandasena et al. (2011), based on the velocities that we have calculated, sliding or overturning/rolling initial motions are not possible; the boulders remain stable. On the other hand, the velocity that is calculated to initiate the motion with saltation is possible. Since in the Eastern Mediterranean, no storm event has been observed with that wave speed, the transportation of the boulder should be a result of tsunami waves.



Figure 3.4: Out-of-place boulder in Narlıkuyu locality. The boulder in the picture is from Miocene limestone and is characterized by intense boring by *P.lithophaga*, showing an onshore transport from the intertidal zone.

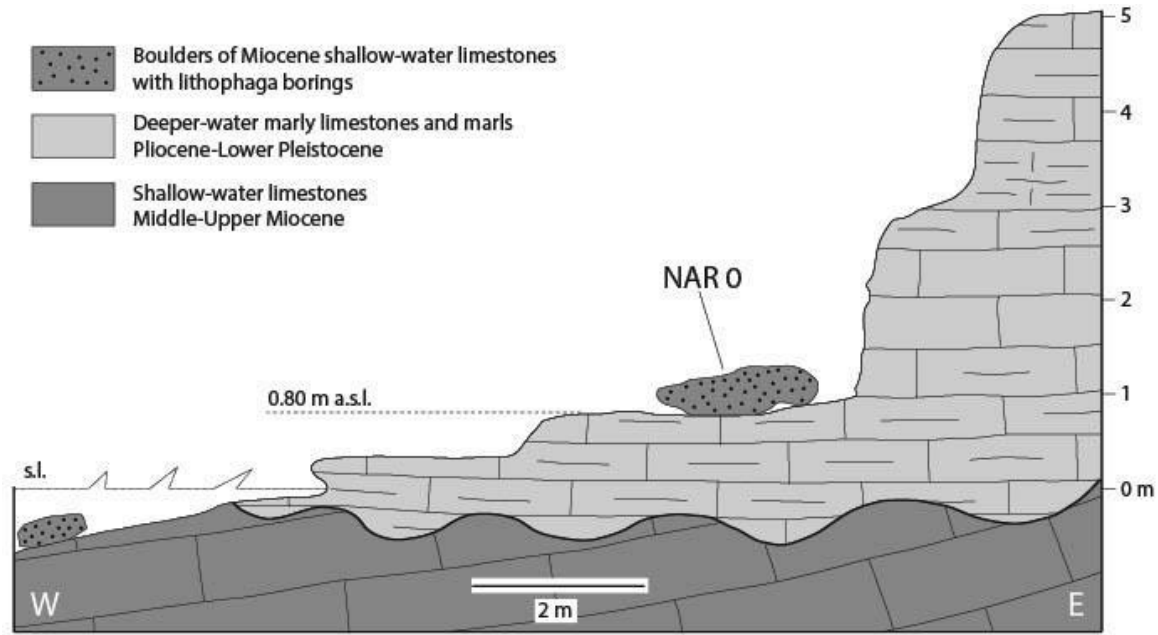


Figure 3.5: Geological cross section of the Narlıkuyu outcrop of NAR 0 boulder.

3.6.2 Yeşilovacık

The physical properties, ages, and locations of at least some of the out-of-place boulders from Yeşilovacık (Table 3.5) are consistent with onshore transport due to high-energy waves. In particular, given the step-like shape of the coast, we consider that the location of the YEŞ 5 boulder, a more than 3-t block of Upper Pleistocene breccia encrusted by vermetids (*B. triquetra*), which lies above an abrasion platform at 2.6 m a.s.l. (Figs 3.6 and 3.7), is consistent with a transport by saltation, instead of sliding or rolling. We applied all the formulas also for YEŞ5 boulder for an evident result. The numbers that we obtained are 4.5 m/s for rolling/overturning mode; 3.2 m/s for sliding mode whereas 10.9 m/s for saltation mode of initial motion. Accordingly, following the boulder transport histogram of Nandasena et al. (2011), in the first 2 options the boulder must remain stable. To start moving the YEŞ 5 boulder (1.4 x 1.2 x 0.8 m) from a shallow-water marine environment onto the shore requires a minimum flow velocity of about 11 m/s which is greater than the observed wave speed of the area as well. We found two additional m-scale boulders (YEŞ 0 and YEŞ 00) on the same 2.6

m abrasion platform where we found the YEŞ 5 boulder. Although one of the Yeşilovacık boulders found at 2.6 m a.s.l. could have been derived from cliff collapse (YEŞ 00, Table 3.5), the rest of the m-scale boulders are indisputably from the sea since they are either from Triassic limestones, which underlie the Pleistocene calcareous breccias of the local cliff and define the seafloor in most of the area (Fig 3.6 and 3.7), or characterized by infralittoral marine biomarkers such as lithophaga borings (YEŞ 4) and/or encrusting marine organisms like vermetids (YEŞ 3, 4, 5). AMS ^{14}C dating of *B. triquetra* from YEŞ 5 yielded an age younger than 1950 AD, whereas *B. triquetra* from boulders YEŞ 3 and YEŞ 4 gave radiocarbon ages of 13 ± 35 years BP and 130 ± 40 years BP, respectively (Table 3.1).



Figure 3.6: Out-of-place boulders with serpulids and marine notches in Yeşilovacık locality.

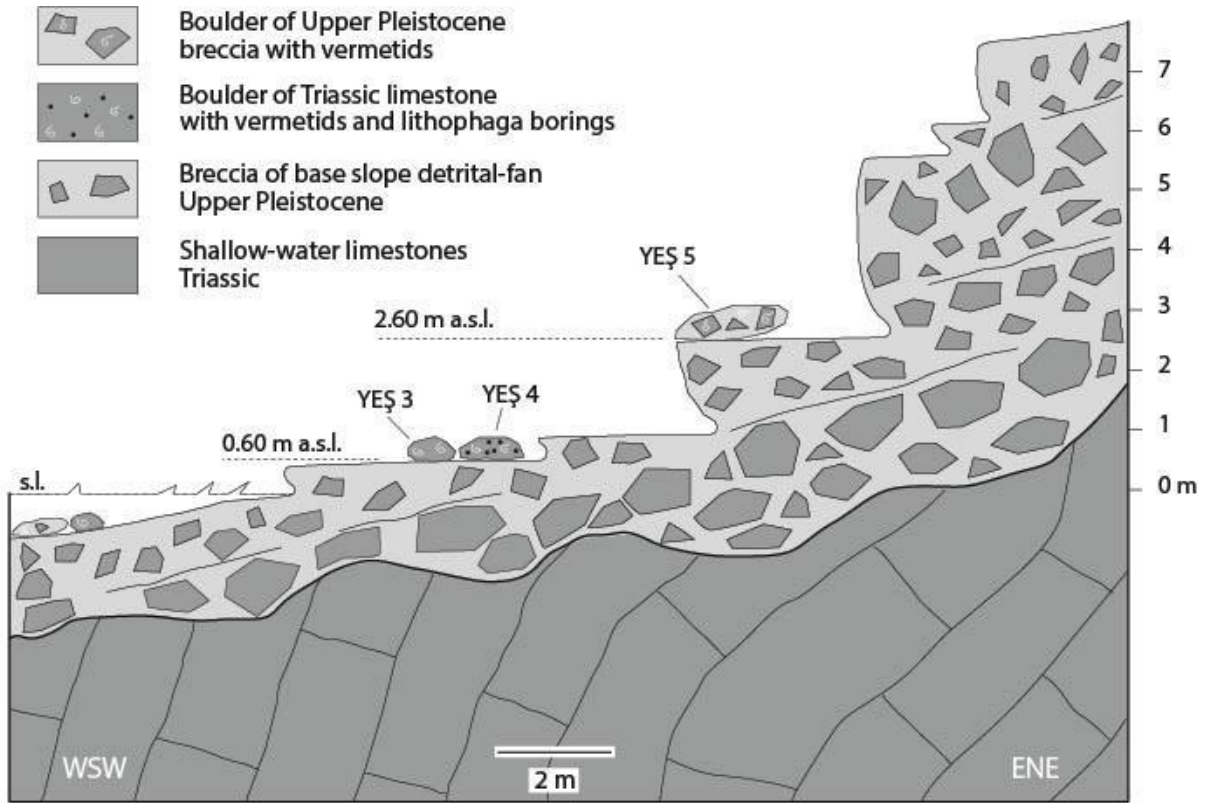


Figure 3.7: Geological cross section of the Yeşilovacık locality with out-of-place boulders.

3.6.3 Aydıncık

The absence of a local cliff in Aydıncık district allows us to eliminate the possibility of offshore-sourced transportation such as a cliff collapse. In that case, we focus on high energy-wave transportation. The other evidence for tsunamiite definition of out-of-place boulders in Aydıncık district is that the boulders are seaward-dipping imbricated which is considered as an evidence of a tsunami event (Young et al., 1996) (Fig 3.8).

Among the transport mechanisms of coastal boulders that have been given above (see Ch. 3.4), in Aydıncık district, considering the coastline geomorphology is beachrock and the boulders lying on the beachrock are slabs of the same lithology, among the possible initial position models given, the best fit model is that they were joint-bounded in their initial pre-transport setting (Etienne et al., 2011; Lau et al., 2015; Nandasena et al., 2011). On the basis

of the physical properties of Aydıncık boulders, we used the formulation (3) (see Ch. 3.5) of Nandasena (2011) that has been established for joint-bounded-boulders.

The variable values regarding the study area that the equation is applied for the largest boulder AYD-B are:

$$\rho_s = 2030 \text{ kg/m}^3; c = 0.8 \text{ m}; \Theta = 3^\circ$$

As a result of our test on AYD-B boulder, the minimum flow velocity that is required to initiate the boulder transportation should be at least 10.7 m/s, that is much greater than the storm flow velocity records of the region similar to our observation in Narlıkuyu locality. When our field observations and the test result combined together, we come up with the interpretation that the sustained transportation mode after the initial motion should be sliding.



Fig 3.8: A) Aydıncık beachrock boulders with imbrication. White arrow shows the direction of imbrication. B) Photo showing the lithology of the beachrocks-slab boulders.

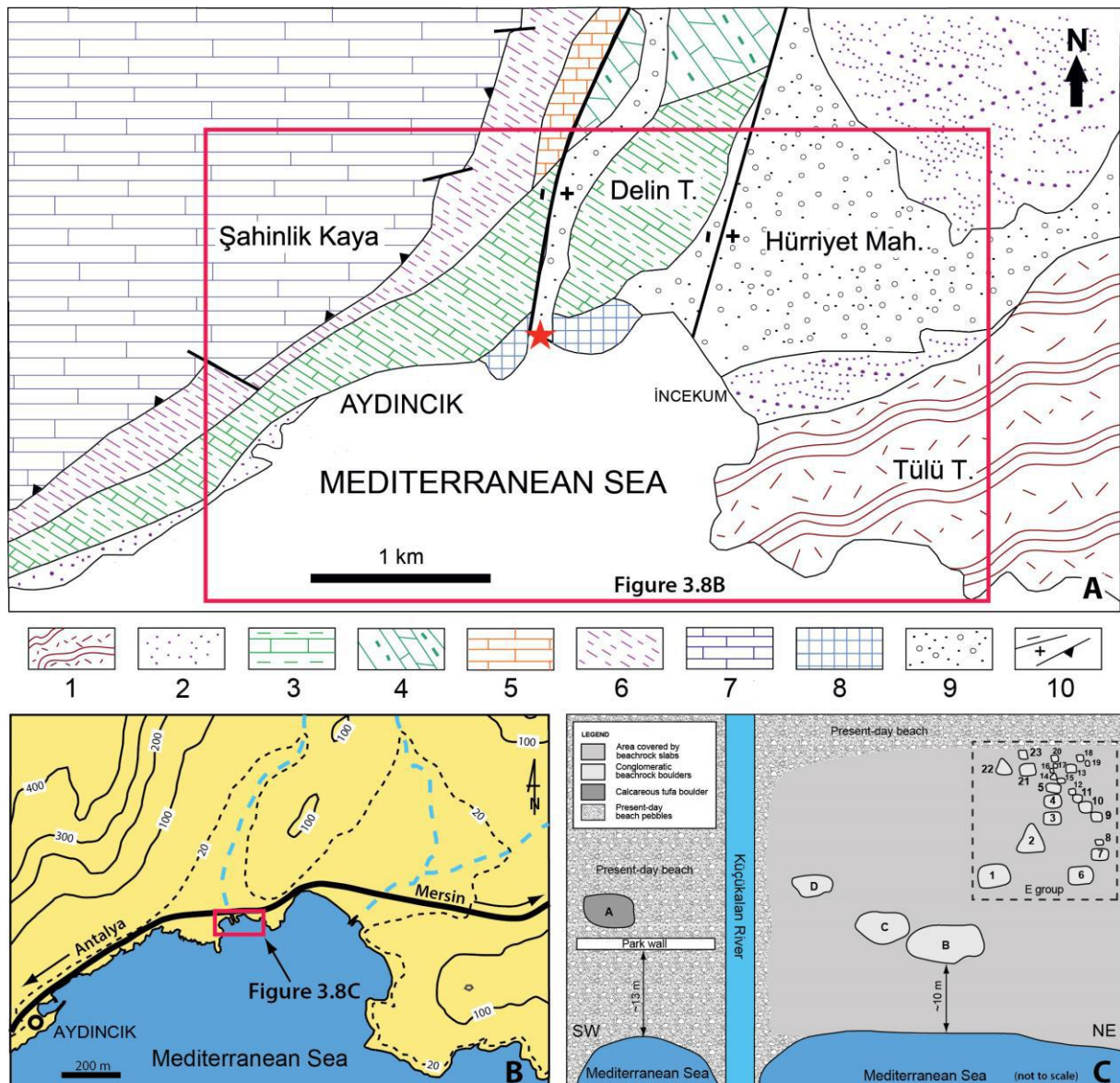


Fig 3.9: Geological map and sketch showing the position of the out-of-place boulders in Aydıncık locality (different sizes of the boulders were considered).

3.7 Discussion

To definitively link geological evidence for tsunamis to a specific tsunamigenic event, it is necessary to know the age of the tsunamites. For the out-of-place boulders along the coast of the Silifke area, the AMS ^{14}C dating of *B. triquetra* (Table 3.1) point to a tsunamigenic event younger than 1950 AD. According to the updated Eastern Mediterranean tsunami catalog compiled by Altınok et al. (2011) and taking into account the simulations by Yolsal et al.

(2007), only two tsunamigenic events after 1950 AD [see Table 1 in Altinok et al (2011)] could have potentially struck the southern Anatolian coast: 10 September 1953 (SW Cyprus) and 9 July 1956 (Southern Aegean Sea). It has been reported that the latter earthquake was a tsunamigenic earthquake with $M > 7.5$ that occurred between Amorgos and Astypalea islands (Southern Aegean Sea). Accordingly, the only remaining tsunamigenic event within the tsunami catalog that could have impacted southern Anatolia after 1950 AD is the 10 September 1953 earthquake, with an epicenter offshore of SW Cyprus. For this event, Altinok et al. (2011) provided detailed information on the tsunamigenic source, noting that the earthquake ($M = 6.2$) was triggered by 0.35 cm of displacement along a right-lateral transpressional fault, which is consistent with the kinematics of the Paphos transform fault.

Combining our data with previous studies, we deduce that in the Eastern Mediterranean region in the past and in the recent past there have been tsunami events that triggered by earthquakes, volcanic eruptions and/or submarine landslides (Altinok et al., 2011; Öğretmen et al., 2015; Papadopoulos, 2016a). Given that the approximate offshore morphology of southern Anatolia is characterized by narrow and relatively steeply sloping shelf where some offshore slumps and landslides affecting the Holocene sequence were identified on high-resolution seismic reflection profiles (Fig. 3.3), to explain the occurrence of the out-of-place boulders at 2.6 m a.s.l., we cannot exclude a local landslide-generated tsunami in the last 65 years along the southern Anatolia coast. Also by previous researchers, in Mediterranean region landslide tsunamis that were caused by earthquakes or rockfalls were reported (Papadopoulos, 2016b). Taking into account that former reports, slope failures affecting this coastal area could be triggered by the seismic and tectonic activity recently recognized along the southwestern prolongation of the Ecemiş Fault Zone (Okyar et al. 2005; Ozel et al. 2007; Öğretmen et al. 2015).

Our coastal geomorphologic evidence shows that in the region, to have out-of-place tsunami boulders, not necessarily big magnitude earthquakes are required but also small magnitude earthquakes could generate tsunami events to transport tsunamiites from offshore domains to onland with different transportation modes.

3.8 Conclusion

Based on the occurrence of onshore, m-scale out-of-place boulders derived from shallow marine environments in the Silifke district, we show for the first time geological evidence of tsunamis affecting the coast of southern Anatolia (Öğretmen et al, 2015). AMS ^{14}C dating on vermetids (*B. triquetra*) from those boulders point to a tsunami event younger than 1950 AD. From the updated Eastern Mediterranean tsunami catalog (Altinok et al. 2011), the only tsunamigenic event that could have impacted southern Anatolia after 1950 AD is the 10 September 1953 $M = 6.2$ earthquake from the tsunamigenic zone of the southwestern Cyprus arc and the Paphos transform fault. The wave impacting the southern Anatolian coast at Yeşilovacık had a minimum run-up of 3.0 m, as indicated by out-of-place boulders found at 2.6 m a.s.l., which yield AMS ^{14}C dates younger than 1950 AD. Given the low magnitude and the small co-seismic displacement in a predominantly strike-slip rupture associated with the 10 September 1953 earthquake, unless strong near-shore amplification occurred, this event is very likely incapable of producing a tsunami with at least 3.0 m of run-up at the southern Anatolian coast. Accordingly, a local landslide-generated tsunami in the last 65 years cannot be ruled out to explain the observed tsunamigenic deposits. The acquisition of high-resolution bathymetry maps for the Anatolian margin of the Cilicia trough to accurately model the near-shore amplification of tsunami waves could help to resolve this question of whether an earthquake or a local landslide generated the tsunami responsible for the m-scale out-of-place boulders found at Yeşilovacık and Narlıkuyu.

In Aydıncık district, the importance of coastal morphology shows itself as not only big magnitude earthquakes and submarine landslides, but also small magnitude earthquakes could be tsunamigenic regarding the fact that convenient morphology (e.g. beachrock morphology) would assist in transporting slab-boulders from offshore domains to onshore.

Eventually, reasoning the strategic importance of the Narlıkuyu, Yeşilovacık and Aydıncık district with their proximity to the planned infrastructure (Nuclear Power Plant), it should not be ruled out the likelihood of tsunamigenic events to prevent the possible damages in the future.

Acknowledgements

This study is part of the Anatolian Plateau Climate and Tectonic Hazards (ALeT) project, funded by the European Union Marie Curie Actions “Initial Training Networks” program (FP7-PEOPLE-2013-ITN)” (grant agreement number 607996). I would like to thank to Prof. Dr. Doğan Kalafat for his scientific contributions.

GENERAL CONCLUSIONS

This study is part of the Anatolian Plateau Climate and Tectonic Hazards (ALeT) project, funded by the European Union Marie Curie Actions “Initial Training Networks” program (FP7-PEOPLE-2013-ITN)”, and represents a comprehensive work on the latest Cenozoic evolution of the southern margin of the Central Anatolian Plateau (CAP) and the main natural hazards that can affect this unstable area.

The disaster connected with the single 1999 Mw7.4 Izmir earthquake in Turkey reminds us of the greater vulnerability of populations in steep, seismically active regions, such as the margins of the CAP. Natural hazards become disasters only when vulnerable populations and infrastructure are involved. To this end, it is worth to note that at the CAP southern margin strategic infrastructures (Akkuyu Nuclear Power Plant) and major expansions of residence complexes have been planned to be built, increasing its vulnerability to natural disasters.

The main task of this study is to better understand the geological processes affecting the CAP southern margin, their triggering mechanisms, and possible feedbacks with surface processes, for contributing to ensure sustainable development and public safety of the region, reducing the risk of natural hazards.

Although the thesis has been organized in three independent different chapters, which present the long-term uplift rate, the short-term uplift rate, and the natural hazards of the area, the issues discussed in each of them are deeply interconnected. The partial conclusions of each chapter can be combined together to give a complete knowledge on the dynamic processes that have been affecting the CAP southern margin since the early Middle Pleistocene and to better understand the triggering mechanisms of the natural hazards in the region.

The biostratigraphical analyses carried out on the youngest marine deposits of the CAP southern margin, taking into account planktonic and benthic foraminifers, ostracods, and calcareous nannofossils, gave us astonished results useful to constrain the timing of the long-term uplift of the CAP southern margin and to answer the crucial question on when the major elevation of the CAP southern margin has been acquired. This work provides unprecedented results on the stratigraphy of the youngest marine deposits capping the Central Taurides units compared to the stratigraphy reported in the official geological map of Turkey (e.g., Şenel, 2002). At the CAP southern margin, in the area surrounding both Gülnar and Tol villages (1000 to 1200 m a.s.l.) the geological map of Turkey indicates the presence of Middle Miocene (ca. 12 Ma) deep-water marine marls (e.g., Şenel, 2002). In contrast, combining the results of this study from both Gülnar East and Tol sections, we can conclude that the youngest marine deposits at the CAP southern margin can be referable to the Mediterranean Plio-Pleistocene marine transgression, which reached the Gülnar district in the Early Pleistocene (ca. 1.7 Ma), with the deposition of Early Pleistocene epibathyal marine deposits onto Middle Miocene shallow-water limestones of the Mut Formation. In addition, given that in the Tol section neogloboquadrinids are not present, the youngest bioevent recognizable in the Pleistocene deposits capping the CAP southern margin is the LCO of *Neogloboquadrina* spp. (sin), which is dated 0.61 Ma. This last biostratigraphical result lead to consider the youngest marine deposits in the Gülnar district as at least early Middle Pleistocene. This means that 0.61 Myr ago the majority of the CAP southern margin was at an elevation close to the sea-level. Qualitative and quantitative analyses on the benthic fauna from both the Gülnar East and Tol sections point to a marine environment from the epibathyal zone (400-500 m of paleodepth), with a palaeo-coastline attested at ca. 1500-1600 m a.s.l.

Considering the elevation of the early Middle Pleistocene palaeo-coastline (1500-1600 m a.s.l.) and the maximum age (0.61 Ma) for the youngest marine deposits at the CAP

southern margin, fast uplift rates reaching values up to 2.45 - 2.62 mm/yr can be evaluated. These uplift rates are higher than those suggested by different authors for the same area based on both different deposits and even shorter time span. Given that the maximum uplift rate predicted by numerical model for shallow slab break-off is only 0.8 mm/yr (Duretz et al., 2011), this result on the uplift rate call into question the effectiveness of a lithospheric slab break-off as the principal mechanisms for triggering the uplift of the CAP southern margin (Cosentino et al., 2012; Schildgen et al., 2012; 2014). In the framework of a dynamic uplift of the Central Anatolian Plateau, a more complex mantle flow system could be responsible for the post 0.61 Ma uplift rates of the CAP southern margin. In the framework of a dynamic uplift of the Central Anatolian Plateau, a more complex mantle flow system could be responsible for the post 0.61 Ma uplift rates of the CAP southern margin, possibly involving the arrival of hot material through the mantle flow sourced by the Afar deep mantle upwelling (Faccenna et al., 2013).

To evaluate the short-term uplift rate of the CAP southern margin and its present day uplift state, we analyzed the coastal area between Mersin and Anamur, looking at the geological features correlatable with the development of a coastline (e.g., marine notches, wave-cut platforms, rock-pools; algae trottoirs, beachrocks, archaeological coastal infrastructures, etc.) and collecting samples suitable for AMS ^{14}C datings. In the study area, all those sea-level markers shows indisputable late Holocene uplift. Using the late Holocene sea-level predicted by Lambeck & Purcell (2005) for the coastal area of the CAP southern margin, in the last 4 kyr the uplift rate of the study area has been ca. 1 mm/yr. This result from our AMS ^{14}C datings is in agreement with the GPS vertical velocity field measured from the GPS station in Silifke, which shows uplift between 0.8 to 1.4 mm/yr (Serpelloni et al., 2013).

In the study area, the late Holocene sea-level markers are affected by extensional tectonics, with offsets of a few centimeters up to 0.70 m. The late Holocene uplift of the CAP

southern margin is coupled with a daily intense seismic activity along a NE-SW tectonic lineament. Although the present seismic activity is mainly due to earthquakes with $M < 3.0$, in the historical archives there are evidence for destructive earthquakes along the coastal area of southern Anatolia (e.g., Corycus, 4th century; Elaioussa-Sebaste, 7th century). The occurrence of extensional faults that cut through late Holocene sea-level markers is an evidence of active tectonics linked to the local seismicity of the study area, even capable to give destructive earthquakes. For this reason, we propose to add in the Active Fault Map of Turkey a seismogenic active fault zone, as a NE-SW fault strand of the Ecemiş Fault in its southwestern prolongation.

Based on the occurrence of onshore, m-scale out-of-place boulders derived from shallow marine environments in the Silifke district (Aydıncık, Yeşilovacık, and Narlıkuyu), we show for the first time geological evidence of tsunamis affecting the coast of southern Anatolia (Öğretmen et al, 2015). AMS ^{14}C dating on vermetids (*B. triquetra*) from those boulders point to a tsunami event younger than 1950 AD. The only tsunamigenic event that could have impacted southern Anatolia after 1950 AD is the 10 September 1953 $M = 6.2$ earthquake from the tsunamigenic zone of the southwestern Cyprus arc and the Paphos transform fault. Given the low magnitude and the small co-seismic displacement in a predominantly strike-slip rupture associated with the 10 September 1953 earthquake, this event is very likely incapable of producing a tsunami at the southern Anatolian coast. Accordingly, a local landslide-generated tsunami in the last 65 years cannot be ruled out to explain the observed tsunamigenic deposits. The acquisition of high-resolution bathymetry maps for the Anatolian margin of the Cilicia trough to accurately model the near-shore amplification of tsunami waves could help to resolve this question of whether an earthquake or a local landslide generated the tsunami responsible for the m-scale out-of-place boulders found at Aydıncık, Yeşilovacık, and Narlıkuyu.

Summing up, the CAP southern margin, as the margins of other orogenic plateaus, is a really unstable area. In the late Messinian (5.5 Ma), it started active growth, experiencing a first phase of uplift (Radeff et al., 2016; 2017). The geological instability of the study area is possibly supported through time by the dynamic uplift due to a deep rooted triggering mechanism connected with a complex mantle flow system.

In terms of natural hazards, the CAP southern margin is a seismically active region that in the past was struck by at least two historical destructive earthquakes (Corycus, 4th century; Elaioussa-Sebaste 7th century). This study also provide evidence of active tectonics affecting late Holocene sea-level markers along the coastal area of the CAP southern margin, suggesting to add in the Active Fault Map of Turkey a new seismogenic active fault zone, in the offshore of the study area, as a NE-SW fault strand of the Ecemiş Fault. In addition to the seismic hazard, the study area is also affected by tsunami hazard, given that this study shows for the first time geological evidence of tsunamis that struck the coast of southern Anatolia in the last 65 years.

In conclusion, this study lead to consider the CAP southern margin as a highly unstable area susceptible to different natural hazards. Unfortunately, the previously scarce geological knowledge on the study area brought the decision makers to plan strategic infrastructures such as Nuclear Power Plant and major expansion of residence complexes, increasing the vulnerability of the area to the natural hazards that already affected the area in the past.



REFERENCES

- Abbott, M.B., Binford, M.W., Brenner, M. & Kelts, K.R., 1997. A 3500 C14 yr High-Resolution Record of Water-Level Changes in Lake Titicaca, Bolivia/Peru. *Quat. Res.* 47, 169–180. doi:10.1006/qres.1997.1881
- AGIP, 1982. *Foraminiferi Padani (Terziario e Quaternario)*, 2nd ed. Milano.
- Åhlfeldt, J., 2015. Digital Atlas of the Roman Empire [WWW Document]. Digit. Atlas Rom. Emp. URL <http://dare.ht.lu.se/> (accessed 10.20.16).
- Aiello, G. & Barra, D., 2001. Pliocene ostracod assemblages at the MPI3- MPI4 boundary in the Capo Rossello borehole (Agrigento, Sicily). *Boll. della Soc. Paleontol. Ital.* 40, 93–103.
- Aiello, G., Barra, D. & Bonaduce, G., 2000. Systematics and biostratigraphy of the ostracoda of the Plio-Pleistocene Monte S. Nicola section (Gela, Sicily).pdf. *Boll. della Soc. Paleontol. Ital.* 39, 83–112.
- Akhun, S.D., Çifçi, G., Dondurur, D., Piercey, T., Kurtboğan, B., Aksu, A.E. & Hall, J., 2008. Kilikya Baseni' nin Plio-Kuvaterner Tektonik Yapıları (Kozan Fayı).
- Aksu, A.E., Hall, J. & Yaltırak, C., 2009. Miocene--Recent evolution of Anaximander Mountains and Finike Basin at the junction of Hellenic and Cyprus Arcs, eastern Mediterranean. *Mar. Geol.* 258, 24–47. doi:10.1016/j.margeo.2008.04.008
- Aksu, A.E., Walsh-Kennedy, S., Hall, J., Hiscott, R.N., Yalt, C., Akhun, S.D. & Çifçi, G., 2014. The Pliocene – Quaternary tectonic evolution of the Cilicia and Adana basins , eastern Mediterranean : Special reference to the development of the Kozan Fault zone. *Tectonophysics* 622, 22–43.
- Aksu, a. E., Calon, T., Hall, J., Kurtboğan, B., Gürçay, S. & Çifçi, G., 2014. Complex interactions fault fans developed in a strike-slip system: Kozan Fault Zone, Eastern Mediterranean Sea. *Mar. Geol.* 351, 91–107. doi:10.1016/j.margeo.2014.03.009
- Aksu, a. E., Calon, T.J., Hall, J., Mansfield, S. & Yaşar, D., 2005. The Cilicia–Adana basin complex, Eastern Mediterranean: Neogene evolution of an active fore-arc basin in an obliquely convergent margin. *Mar. Geol.* 221, 121–159. doi:10.1016/j.margeo.2005.03.011
- Akyuz, H.S., Altunel, E., Karabacak, V. & Yalciner, C.C., 2006. Historical earthquake activity of the northern part of the Dead Sea Fault Zone, southern Turkey. *Tectonophysics* 426, 281–293. doi:10.1016/j.tecto.2006.08.005
- Allen, M.B., Saville, C., Blanc, E.J.P., Talebian, M. & Nissen, E., 2015. Orogenic plateau growth: Expansion of the Turkish-Iranian Plateau across the Zagros fold-and-thrust belt. *Tectonics* 32, 171–190. doi:10.1002/tect.20025
- Allmendinger, R.W., Jordan, T.E., Kay, S.M. & Isacks, B.L., 1997. the Evolution of the Altiplano-Puna Plateau of the Central Andes. *Annu. Rev. Earth Planet. Sci.* 25, 139–174. doi:DOI: 10.1146/annurev.earth.25.1.139
- Altenbach, A.V., Lutze, G.F., Schiebel, R. & Schönfeld, J., 2003. Impact of interrelated and interdependent ecological controls on benthic foraminifera: An example from the Gulf of Guinea. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 197, 213–238. doi:10.1016/S0031-0182(03)00463-2
- Altınok, Y., Alpar, B., Özer, N. & Aykurt, H., 2011. Revision of the tsunami catalogue affecting Turkish coasts and surrounding regions. *Nat. Hazards Earth Syst. Sci.* 11, 273–291. doi:10.5194/nhess-11-273-2011
- Altınok, Y. & Ersoy, Ş., 2000. Tsunamis observed on and near the Turkish coast. *Nat. Hazards* 21, 185–205.
- Amore, O.F., Caffau, M., Massa, B. & Morabito, S., 2004. Late Pleistocene – Holocene paleoclimate and related paleoenvironmental changes as recorded by calcareous nannofossils and planktonic foraminifera assemblages in the southern Tyrrhenian Sea (Cape Palinuro, Italy). *Mar. Micropaleontol.* 52, 255–276. doi:10.1016/j.marmicro.2004.05.004
- Anthonissen, D.E. & Ogg, J.G., 2012. Appendix 3. Cenozoic and Cretaceous Biochronology of Planktonic Foraminifera and Calcareous Nannofossils, in: Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M. (Eds.), *The Geologic Time Scale 2012*. Elsevier, Amsterdam, pp. 1083–1127.
- Antonioli, F., Ferranti, L., Lambeck, K., Kershaw, S., Verrubbi, V. & Dai, G., 2006. Late Pleistocene to Holocene record of changing uplift rates in southern Calabria and northeastern Sicily (southern Italy),

- Central Mediterranean Sea). *Tectonophysics* 422, 23–40. doi:10.1016/j.tecto.2006.05.003
- Antonioli, F., Lo Presti, V., Rovere, A., Ferranti, L., Anzidei, M., Furlani, S., Mastronuzzi, G., Orru, P.E., Scicchitano, G., Sannino, G., Spampinato, C.R., Pagliarulo, R., Deiana, G., de Sabata, E., Sansò, P., Vacchi, M. & Vecchio, A., 2015. Tidal notches in Mediterranean Sea: A comprehensive analysis. *Quat. Sci. Rev.* 119, 66–84. doi:10.1016/j.quascirev.2015.03.016
- Anzidei, M., Antonioli, F., Benini, A., Lambeck, K., Sivan, D., Serpelloni, E. & Stocchi, P., 2011. Sea level change and vertical land movements since the last two millennia along the coasts of southwestern Turkey and Israel. *Quat. Int.* 232, 13–20. doi:10.1016/j.quaint.2010.05.005
- Auriemma, R. & Solinas, E., 2009. Archaeological remains as sea level change markers: A review. *Quat. Int.* 206, 134–146. doi:10.1016/j.quaint.2008.11.012
- Babinot, J., 2002. Ostracodes miocènes de séries annexes aux Bassin de Koprucay et de Manavgat, région d'Antalya (Sud-Turquie): Systématique-Implication stratigraphiques et paléoenvironnementales. *Rev. Paléobiologie* 21, 735–757.
- Baldrighi, E., Lavaleye, M., Aliani, S., Conversi, A. & Manin, E., 2014. Large Spatial Scale Variability in Bathyal Macrobenthos Abundance, Biomass, a- and b-Diversity along the Mediterranean Continental Margin. *PLoS One* 9, 1–16.
- Barbano, M.S., Gerardi, F. & Pirrotta, C., 2011. Differentiation between boulders deposited by tsunamis and storm waves along the South-Eastern ionian coast of sicily (Italy). *Boll. di Geofis. Teor. ed Appl.* 52, 707–728. doi:10.4430/bgta0022
- Barka, A. & Reilinger, R., 1997. Active tectonics of the Eastern Mediterranean Region: deduced from GPS, neotectonic and seismic data. *Ann. di Geofis.* XL, 587–610. doi:10.1111/j.1365-2486.2011.02460.x
- Barra, D., Aiello, G. & Bonaduce, G., 1998. The genus *Argilloecia* Sars, 1866 (Crustacea: Ostracoda) in the Pliocene-Early Pleistocene of the M. San Nicola Section (Gela, Sicily), in: *Proceedings of 2nd European Ostracodologists Meeting*. British Micropaleontological Society, Glasgow, pp. 129–134.
- Bartol, J. & Govers, R., 2014. A single cause for uplift of the Central and Eastern Anatolian plateau? *Tectonophysics* 637, 116–136. doi:10.1016/j.tecto.2014.10.002
- Bartol, J., Matenco, L., Garcia-Castellanos, D. & Leever, K., 2012. Modelling depositional shifts between sedimentary basins: Sediment pathways in Paratethys basins during the Messinian Salinity Crisis. *Tectonophysics* 536–537, 110–121. doi:10.1016/j.tecto.2012.03.007
- Bassant, P., 1999. The high-resolution stratigraphic architecture and evolution of the Burdigalian carbonate-siliciclastic sedimentary systems of the Mut Basin, Turkey. *l'Université de Fribourg*.
- Bassant, P., Van Buchem, F.S.P., Strasser, A. & Görür, N., 2005. The stratigraphic architecture and evolution of the Burdigalian carbonate - Siliciclastic sedimentary systems of the Mut Basin, Turkey. *Sediment. Geol.* 173, 187–232. doi:10.1016/j.sedgeo.2004.01.017
- Ben-Avraham, Z., Woodside, J., Lodolo, E., Gardosh, M., Grasso, M., Camerlenghi, A. & Vai, G.B., 2006. Eastern Mediterranean basin systems. *Geol. Soc. London, Mem.* 32, 263–276. doi:10.1144/GSL.MEM.2006.032.01.15
- Benson, R.H., 1977. Evolution of Oblitacythereis from Paleocosta (Ostracoda: Trachyleberididae) during the Cenozoic in the Mediterranean and Atlantic. *Smithson. Contrib. Paleobiol.* 33, 1–47.
- Blanc-Vernet, L., 1969. Contribution a l'étude des foraminifères du Méditerranée. *Recl. Trav. la Stn. Mar. Stat. Mar.d'Endoume* 64, 1–315.
- Blow, W.H., 1969. Late middle Eocene to Recent planktonic foraminiferal biostratigraphy, in: *Bronnimann, P., Renz, H.H. (Eds.), Proceedings of the First International Conference on Planktonic Microfossils, Geneva 1967*. Geneva, pp. 199–422.
- Bolli, H.M. & Premoli Silva, I., 1973. Oligocene To Recent Planktonic Foraminifera and Stratigraphy of the Leg 15 Sites in the Caribbean Sea. *DSDP Initial Reports* 15, 578–584.
- Bolli, H.M. & Saunders, J.B., 1985. Oligocene to Holocene low latitude planktic foraminifera, in: *Bolli, H.M., Saunders, J.B., Perch-Nielsen, K. (Eds.), Plankton Stratigraphy*. Cambridge University Press, pp. 155–262.

- Bonaduce, G., 1983. The deep-water benthic ostracodes of the Mediterranean, in: Maddocks, R.F. (Ed.), *Application of Ostracoda*. University Houston Geosciences, pp. 459–471.
- Bonaduce, G. & Barra, D., 2002. The ostracods in the palaeoenvironmental interpretation of the Late Langhian–Early Serravallian section of Ras il-Pellegrin (Malta). *Riv. Ital. di Paleontol. e Stratigr.* doi:10.13130/2039-4942/5471
- Bonaduce, G. & Sprovieri, R., 1985. The appearance of *Cytheropteron testudo* Sars (Crustacea: Ostracoda) is a Pliocene event. Evidences from a sicilian sequence (Italy). *Boll. della Soc. Paleontol. Ital.* 23, 131–136.
- Boomer, I., Horne, D.J. & Slipper, I.J., 2003. The use of ostracods in palaeoenvironmental studies, or, What can we do with an ostracod shell?, in: *Bridging the Gap. Trends in the Ostracode Biological and Geological Sciences*. Paleontological Society Papers, pp. 153–180.
- Boulton, S.J. & Stewart, I.S., 2013. Holocene coastal notches in the Mediterranean region: Indicators of palaeoseismic clustering? *Geomorphology*. doi:10.1016/j.geomorph.2013.11.012
- Bourgeois, J., 2009. Geologic Effects and Records of Tsunamis, in: Robinson, A.R., Bernard, E.N. (Eds.), *The Sea*. Harvard University Press, Cambridge, pp. 53–91.
- Bown, P.R. & Young, J.R., 1998. Chapter 2: Techniques, in: Bown, P.R. (Ed.), *Calcareous Nannofossil Biostratigraphy*. Kluwer Academic Publisher, pp. 16–28.
- Bozkurt, E. & Mittweide, S.K., 2001. Introduction to the Geology of Turkey—A Synthesis. *Int. Geol. Rev.* 43, 578–594. doi:10.1080/00206810109465034
- Bull, W.B., 1972. Recognition of alluvial-fan deposits in the stratigraphic record, in: Rigby, J.K. (Ed.), *Recognition of Ancient Sedimentary Environment*. SEPM Special Publications, pp. 63–83.
- Burke, K. & Şengör, A.M.C., 1986. Tectonic escape in the evolution of the continental crust, in: Barazangi, M., Brown, L.D. (Eds.), *Reflection Seismology: The Continental Crust*. American Geophysical Union, pp. 41–53.
- Burton-Ferguson, R., Aksu, A. E., Calon, T.J. & Hall, J., 2005. Seismic stratigraphy and structural evolution of the Adana Basin, eastern Mediterranean. *Mar. Geol.* 221, 189–222. doi:10.1016/j.margeo.2005.03.009
- Burton, P.W., McGonigle, R., Makropoulos, K.C. & Üçer, S.B., 1984. Seismic risk in Turkey, the Aegean and the eastern Mediterranean. *Geophys. J. R. Astr. Soc.* 78, 475–506.
- Carobene, L., 2015. Marine Notches and Sea-Cave Bioerosional Grooves in Microtidal Areas : Examples from the Tyrrhenian and Ligurian Coasts — Italy. *J. Coast. Res.* 31, 536 – 556. doi:10.2112/JCOASTRES-D-14-00068.1
- Caruso, A., 2004. Climatic changes during Late Pliocene and Early Pleistocene at Capo Rossello (Sicily, Italy): response from planktonic foraminifera, in: Coccioni, R., Galeotti, S., Lirer, F. (Eds.), *Proceedings of the First Italian Meeting on Environmental Micropaleontology*. Grzybowski Foundation Special Publication, pp. 17–35.
- Casale, V., 1982. I foraminiferi bentonici dell'Abbeveratoio Conzo (Grammichele, Catania). *Geol. Rom.* 21, 545–557.
- Chappell, J., Omura, A., Esat, T., McCulloch, M., Pandolfi, J., Ota, Y. & Pillans, B., 1996. Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth Planet. Sci. Lett.* 141, 227–236. doi:10.1016/0012-821X(96)00062-3
- Ciampo, G., 1986. Ostracodi del limite Tortonian/Messiniano in alcune sezioni italiane. *Boll. della Soc. Paleontol. Ital.* 24, 29–110.
- Cipollari, P., Cosentino, D., Radeff, G., Taylor, F., Faranda, C., Grossi, F., Gliozzi, E., Smedile, A., Gennari, R., Darbas, G., Dudas, Ö., Gürbüz, K., Nazik, A., Ehtler, H., Cipollari, P., Cosentino, D., Radeff, G., Schildgen, T.F., Faranda, C., Grossi, F. & Darbas, G., 2012. Easternmost Mediterranean evidence of the Zanclean flooding event and subsequent surface uplift : Adana Basin, southern Turkey, in: Robertson, A.H.F., Parlak, O. & Ünlügenç, U.C. (Eds.), *Geological Development of Anatolia and the Easternmost Mediterranean Region*. Geological Society, London, Special Publications, 372, pp. 473–494. doi:10.1144/SP372.5

- Cipollari, P., Halászová, E., Gürbüz, K. & Cosentino, D., 2013. Middle-Upper Miocene paleogeography of southern Turkey: insights from stratigraphy and calcareous nannofossil biochronology of the Olukpinar and Başıyayla sections (Mut-Ermenek Basin). *Turkish J. Earth Sci.* 22, 820–838. doi:10.3906/yer-1208-2
- Cita, B., Capraro, L., Ciaran, N., Stefano, E. D., Lirer, F., Maiorano, P., Marino, M., Raffi, I., Rio, D., Sprovieri, R., 2008. The Calabrian Stage redefined. *Episodes* 31, 408–419.
- Colalongo, M.L. & Pasini, G., 1980. La Ostracod fauna plio-pleistocenica della sezione Vrica in Calabria (con considerazioni sul limite Neogene/Quaternario). *Boll. della Soc. Paleontol. Ital.* 27, 277–289.
- Colalongo, M.L., Pasini, G., Poluzzi, A. & Sprovieri, R., 1990. Relationship between the benthic foraminifers and the ostracodes in the Pliocene-Pleistocene Tyrrhenian deep-sea record (ODP Leg 107, Site 654), in: Kastens, K.A., Mascle, J., Al., E. (Eds.), *Proc. ODP, Sci. Results.* pp. 479–493.
- Colalongo, M.L., Pasini, G. & Sartoni, S., 1981. Remarks on the Neogene Quaternary boundary and the Vrica Section (Calabria, Italy). *Boll. della Soc. Paleontol. Ital.* 20, 99–120.
- Cooper, F.J., Roberts, G.P. & Underwood, C.J., 2007. A comparison of 103–105 year uplift rates on the South Alkyonides Fault, Central Greece: Holocene climate stability and the formation of coastal notches. *Geophys. Res. Lett.* 34, 1–6. doi:10.1029/2007GL030673
- Cooper, J.A.G., 1991. Beachrock formation in low latitudes: implications for coastal evolutionary models. *Mar. Geol.* 98, 145–154. doi:10.1016/0025-3227(91)90042-3
- Coppa, M.G., Russo, B. & Siani, G., 1994. The Holocene foraminiferal assemblages of the continental margin between Agropoli and Capo Palinuro (Tyrrhenian Sea), in: Matteucci, R. (Ed.), *Studies on Ecology and Paleocology of Benthic Communities.* Boll. Soc. Paleont. Ital., pp. 67–91.
- Cosentino, D., Cipollari, P., Di Bella, L., Esposito, A., Faranda, C., Giordano, G., Gliozzi, E., Mattei, M., Mazzini, I., Porreca, M. & Funiciello, R., 2009. Tectonics, sea-level changes and palaeoenvironments in the early Pleistocene of Rome (Italy). *Quat. Res.* 72, 143–155. doi:10.1016/j.yqres.2009.03.003
- Cosentino, D. & Gliozzi, E., 1988. Considerazioni sulle velocità di sollevamento di depositi eutirreniani dell'Italia meridionale e della Sicilia. *Mem. Soc. Geol. It.* 41, 653–665.
- Cosentino, D., Radeff, G., Darbaş, G., Dudas, F., Gürbüz, K. & Schildgen, T., 2010. Late Miocene geohistory of the Mut and Adana Basins (southern Turkey): Insight for the uplift of the southern margin of the Central Anatolia Plateau, in: *Tectonic Crossroads: Evolving Orogens of Eurasia-Africa-Arabia.* Geological Society of America, pp. 34–35.
- Cosentino, D., Schildgen, T.F., Cipollari, P., Faranda, C., Gliozzi, E., Hudackova, N., Lucifora, S. & Strecker, M.R., 2012. Late Miocene surface uplift of the southern margin of the Central Anatolian Plateau, Central Taurides, Turkey. *Geol. Soc. Am. Bull.* 124, 133–145. doi:10.1130/B30466.1
- Costa, P.J.M., Andrade, C., Freitas, M.C., Oliveira, M. a., da Silva, C.M., Omira, R., Taborda, R., Baptista, M. A. & Dawson, A.G., 2011. Boulder deposition during major tsunami events. *Earth Surf. Process. Landforms* 36, 2054–2068. doi:10.1002/esp.2228
- Dawson, A.G., 1994. Geomorphological effects of tsunami run-up and backwash. *Geomorphology* 10, 83–94. doi:10.1016/0169-555X(94)90009-4
- Dawson, A.G., 1999. Linking tsunami deposits, submarine slides and offshore earthquakes. *Quat. Int.* 60, 119–126. doi:10.1016/S1040-6182(99)00011-7
- Dawson, A.G., Stewart, I., 2007. Tsunami deposits in the geological record. *Sediment. Geol.* 200, 166–183. doi:10.1016/j.sedgeo.2007.01.002
- de Kaenel, E., Siesser, W.G. & Murat, A., 1999. Pleistocene calcareous nannofossil biostratigraphy and the western Mediterranean sapropels, Sites 974 to 977 and 979. *Proc. Ocean Drill. Program, 161 Sci. Results* 161, 159–183. doi:10.2973/odp.proc.sr.161.250.1999
- De Rijk, S., Jorissen, F.J., Rohling, E.J. & Troelstra, S.R., 2000. Organic flux control on bathymetric zonation of Mediterranean benthic foraminifera. *Mar. Micropaleontol.* 40, 151–166. doi:10.1016/S0377-8398(00)00037-2
- De Rijk, S., Troelstra, S.R. & Rohling, E.J., 1999. Benthic foraminiferal distribution in the Mediterranean Sea. *J.*

- Foraminifer. Res. 29, 93–103.
- de Stigter, H.C., van der Zwaan, G.J. & Langone, L., 1999. Differential rates of benthic foraminiferal test production in surface and subsurface sediment habitats in the southern Adriatic Sea. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 149, 67–88.
- del Carmen Sánchez Ariza, M., 1988. Study of agglutinated foraminifera from the Motril-Nerja littoral region, Spain. *Abh.Geol.B.-A* 41, 301–305.
- Dercourt, J., Zonenshain, L.P., Ricou, L.-E., Kazmin, V.G., Le Pichon, X., Knipper, A.L., Grandjacquet, C., Sbertshikov, I.M., Geyssant, J., Lepvrier, C., Pechersky, D.H., Boulin, J., Sibuet, J.-C., Savostin, L.A., Sorokhtin, O., Westphal, M., Bazhenov, M.L., Lauer, J.P. & Biju-Duval, B., 1986. Geological evolution of the tethys belt from the atlantic to the pamirs since the LIAS. *Tectonophysics* 123, 241–315. doi:10.1016/0040-1951(86)90199-X
- Desruelles, S., Fouache, É., Ciner, A., Dalongeville, R., Pavlopoulos, K., Kosun, E., Coquinot, Y. & Potdevin, J.L., 2009. Beachrocks and sea level changes since Middle Holocene: Comparison between the insular group of Mykonos-Delos-Rhenia (Cyclades, Greece) and the southern coast of Turkey. *Glob. Planet. Change* 66, 19–33. doi:10.1016/j.gloplacha.2008.07.009
- Dewey, J.F. & Şengör, A.M.C., 1979. Aegean and surrounding regions: complex multiplate and continuum tectonics in a convergent zone. *Geol. Soc. Amer. Bull.* 90, 84–92.
- Dhont, D., Chorowicz, J. & Yürür, T., 1999. The Bolkar Mountains (Central Taurides, Turkey): a Neogene extensional thermal uplift? *Geol. Bull. Turkey* 42, 69–87.
- Di Bella, L., 2010. Plio-Pleistocene foraminiferal assemblages of the Monte Mario site (Rome, Italy). *Boll. Della Soc. Paleontol. Ital.* 49, 145–161.
- Di Bella, L., Carboni, M.G. & Bergamin, L., 2000. Pliocene-Pleistocene foraminiferal assemblages of the middle and lower Tiber Valley: stratigraphy and paleoecology. *Geol. Rom.* 36, 129–145.
- Duman, T.Y., Çan, T., Emre, Ö., Keçer, M., Doğan, A., Ateş, Ş. & Durmaz, S., 2005. Landslide inventory of northwestern Anatolia, Turkey. *Eng. Geol.* 77, 99–114. doi:10.1016/j.enggeo.2004.08.005
- Duretz, T., Gerya, T. V. & May, D.A., 2011. Numerical modelling of spontaneous slab breakoff and subsequent topographic response. *Tectonophysics* 502, 244–256. doi:10.1016/j.tecto.2010.05.024
- Ehlers, T.A., 2013. The influence of orogenic plateau formation on climate and erosion.
- Ehlers, T.A. & Poulsen, C.J., 2009. Influence of Andean uplift on climate and paleoaltimetry estimates. *Earth Planet. Sci. Lett.* 281, 238–248. doi:10.1016/j.epsl.2009.02.026
- Emeis, K. & Sakamoto, T., 1998. The sapropel theme of leg 160 1, in: Robertson, A.H.F., Emeis, K.-C., Richter, C., Camerlenghi, A. (Eds.), *Proceedings of the Ocean Drilling Program. ODP Texas A&M University, Texas*, pp. 29–36. doi:10.2973/odp.proc.sr.160.058.1998
- Emre, Ö., Duman, T.Y., Duman, Ş. & Özalp, S., 2012. Türkiye Diri Fay Haritası (Yenilenmiş). Maden Tetkik ve Arama Yayınları, Ankara.
- Equini Schneider E. (2008). Elaiussa-Sebaste, a port city between east and west. *Archaeological guide*, printed by Altan Basım Ltd. Şti.
- Erener, A. & Düzgün, H.B.S., 2013. A regional scale quantitative risk assessment for landslides: Case of Kumluca watershed in Bartın, Turkey. *Landslides* 10, 55–73. doi:10.1007/s10346-012-0317-9
- Eriş, K.K., Bassant, P. & Ülgen, U.B., 2005. Tectono-stratigraphic evolution of an Early Miocene incised valley-fill (Derinçay Formation) in the Mut Basin, Southern Turkey. *Sediment. Geol.* 173, 151–185. doi:10.1016/j.sedgeo.2003.12.011
- Etienne, S., Buckley, M., Paris, R., Nandasena, A.K., Clark, K., Strotz, L., Chagué-Goff, C., Goff, J. & Richmond, B., 2011. The use of boulders for characterising past tsunamis: Lessons from the 2004 Indian Ocean and 2009 South Pacific tsunamis. *Earth-Science Rev.* 107, 76–90. doi:10.1016/j.earscirev.2010.12.006
- Etienne, S. & Terry, J.P., 2012. Coral boulders, gravel tongues and sand sheets: Features of coastal accretion and

- sediment nourishment by Cyclone Tomas (March 2010) on Taveuni Island, Fiji. *Geomorphology* 175–176, 54–65. doi:10.1016/j.geomorph.2012.06.018
- Evelpidou, N., Kampolis, I., Pirazzoli, P.A. & Vassilopoulos, A., 2012. Global sea-level rise and the disappearance of tidal notches. *Glob. Planet. Change* 92–93, 248–256. doi:10.1016/j.gloplacha.2012.05.013
- Faccenna, C., Becker, T.W., Conrad, C.P. & Husson, L., 2013. Mountain building and mantle dynamics. *Tectonics* 32, 80–93. doi:10.1029/2012TC003176
- Faccenna, C., Bellier, O. & Martinod, J., 2006. Slab detachment beneath eastern Anatolia: a possible cause for the formation of the North Anatolian Fault Earth and Planetary. *Sci. Lett.* 242, 1–2. doi:10.1029/2002JB001757.
- Faranda, C., Gliozzi, E., Cipollari, P., Grossi, F. & Darbaş, G., 2013. Messinian paleoenvironmental changes in the easternmost Mediterranean Basin: Adana Basin, southern Turkey. *Turkish J. Earth Sci.* 22, 839–863.
- Faranda, C., Gliozzi, E. & Mazzini, I., 2007. Palaeoenvironmental evolution of the Plio-Pleistocene Monte Mario succession (Rome, Italy) inferred from ostracod assemblages. *Riv. Ital. di Paleontol. e Stratigr.* 113, 473–485.
- Fiorini, F. & Vaiani, S.C., 2001. Benthic foraminifers and transgressive-regressive cycles in the Late Quaternary subsurface sediments of the Po Plain near Ravenna (Northern Italy). *Boll. della Soc. Paleontol. Ital.* 40, 357–403.
- Flores, J.-A., Gersonde, R. & Sierro, F.J., 1999. Pleistocene fluctuations in the Agulhas Current Retroflexion based on the calcareous plankton record. *Mar. Geol.* 37, 1–22.
- Fontanier, C., Jorissen, F.J., Licari, L., Alexandre, A., Anschutz, P. & Carbonel, P., 2002. Live benthic foraminiferal faunas from the Bay of Biscay: Faunal density, composition, and microhabitats. *Deep. Res. Part I Oceanogr. Res. Pap.* 49, 751–785. doi:10.1016/S0967-0637(01)00078-4
- Frankel, E., 1968. Rate of formation of beach rock. *Earth Planet. Sci. Lett.* 4, 439–440. doi:10.1017/CBO9781107415324.004
- Frezza, V. & Carboni, M.G., 2009. Distribution of recent foraminiferal assemblages near the Ombrone River mouth (Northern Tyrrhenian Sea, Italy). *Rev. micropaléontologie* 52, 43–66.
- Frezza, V., Carboni, M.G. & Matteucci, R., 2005. Recent foraminiferal assemblages near Ponza Island (Central Tyrrhenian Sea, Italy). *Boll. della Soc. Paleontol. Ital.* 44, 155–173.
- Frezza, V., Pignatti, J.S. & Matteucci, R., 2010. Benthic foraminiferal biofacies in temperate carbonate sediment in the Western Pontine Archipelago (Tyrrhenian Sea, Italy). *J. Foraminifer. Res.* 40, 313–326.
- Galili, E., Zviely, D. & Weinstein-Evron, M., 2005. Holocene sea-level changes and landscape evolution on the northern Carmel coast (Israel). *Méditerranée* 104, 79–86.
- Gardner, T.W., 1992. Quaternary uplift astride the aseismic Cocos Ridge, Pacific coast, Costa Rica. *Geol. Soc. Am. Bull.* 104, 219–232. doi:10.1130/0016-7606(1992)104<0219:QUATAC>2.3.CO;2
- Gebhardt, H., 1999. Middle to Upper Miocene benthonic foraminiferal palaeoecology of the Tap Marls (Alicante Province, SE Spain) and its palaeoceanographic implications. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 145, 141–156. doi:10.1016/S0031-0182(98)00110-2
- Gelati, R., 1975. Miocene marine sequence from Lake Van: Eastern Turkey. *Riv. Ital. Paleontol. Strat.* 81, 477–490.
- Geological Survey Department of Cyprus, 2014. Historic Earthquakes [WWW Document]. *Hist. Earthquakes*. URL <http://www.moa.gov.cy/moa/> (accessed 10.9.14).
- Goff, J., Chagué-Goff, C., Nichol, S., Jaffe, B. & Dominey-Howes, D., 2012. Progress in palaeotsunami research. *Sediment. Geol.* 243–244, 70–88. doi:10.1016/j.sedgeo.2011.11.002
- Goff, J., Dudley, W.C., DeMaintenon, M.J., Cain, G. & Coney, J.P., 2006. The largest local tsunami in 20th century Hawaii. *Mar. Geol.* 226, 65–79. doi:10.1016/j.margeo.2005.09.017

- Goff, J., McFadgen, B.G. & Chagué-Goff, C., 2004. Sedimentary differences between the 2002 Easter storm and the 15th-century Okoropunga tsunami, southeastern North Island, New Zealand. *Mar. Geol.* 204, 235–250. doi:[http://dx.doi.org/10.1016/S0025-3227\(03\)00352-9](http://dx.doi.org/10.1016/S0025-3227(03)00352-9)
- Goodwin, I.D., 2009. Beachrock. *Encycl. Paleoclimatology Anc. Environ.* doi:10.1007/978-1-4020-4411-3_21
- Goto, K., Chavanich, S.A., Imamura, F., Kunthasap, P., Matsui, T., Minoura, K., Sugawara, D. & Yanagisawa, H., 2007. Distribution, origin and transport process of boulders deposited by the 2004 Indian Ocean tsunami at Pakarang Cape, Thailand. *Sediment. Geol.* 202, 821–837. doi:10.1016/j.sedgeo.2007.09.004
- Göğüş, O.H. & Pysklywec, R.N., 2008. Mantle lithosphere delamination driving plateau uplift and synconvergent extension in eastern Anatolia. *Geology* 36, 723. doi:10.1130/G24982A.1
- Göncüoğlu, M.C., 2010. Introduction to the geology of Turkey: Geodynamic evolution of the Pre-Alpine and Alpine terranes. *MTA Monogr. Ser.* 5, 1–69.
- Gradstein, F.M., Ogg, J.G. & Smith, A.G., 2004. A geologic time scale 2004. Cambridge University Press.
- Gregory-Wodzicki, K.M., 2000. Uplift history of the Central and Northern Andes: A review. *GSA Bull.* 112, 1091–1105. doi:10.1130/0016-7606(2000)112<1091:UHOTCA>2.0.CO;2
- Griffin, D.L., 2002. Aridity and humidity : two aspects of the late Miocene climate of North Africa and the Mediterranean. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 182, 65–91.
- Guidoboni, E., 1989. I terremoti prima del Mille in Italia e nell'area mediterranea. *Storia archeologia sismologia*. Istituto Nazionale di Geofisica, Bologna.
- Guidoboni, E. & Comastri, A., 2005. Catalogue of earthquakes and tsunamis in the Mediterranean area from the 11th to the 15th century. INGV-SGA, Bologna.
- Guidoboni, E., Comastri, A. & Traina, G., 1994. Catalogue of ancient earthquakes in the Mediterranean area up to the 10th century. Roma.
- Guilcher, A. & Pont, P., 1957. Etude expérimentale de la corrosion littorale des calcaires. *Bull. Assoc. Geogr. Fr.* 48–62.
- Hammer, Ø., Harper, D.A.T. & Ryan, P.D., 2001. PAST: Paleontological Statistics Software Package for Education and Data Analysis. *Palaeontol. Electron.* 4(1), 1–9. doi:10.1016/j.bcp.2008.05.025
- Haq, B.U., Hardenbol, J. & Vail, P.R., 1988. Mesozoic and Cenozoic chronostratigraphy and eustatic cycles, in: *Sea-Level Changes-An Integrated Approach*. SEPM Special Publications, pp. 71–108. doi:10.2110/pec.88.01.0071
- Harris, N., 2006. The elevation history of the Tibetan Plateau and its implications for the Asian monsoon. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 241, 4–15. doi:10.1016/j.palaeo.2006.07.009
- Hartley, A., 2003. Andean uplift and climate change. *J. Geol. Soc. London.* 160, 7–10. doi:10.1144/0016-764902-083
- Hatzfeld, D. & Molnar, P., 2010. Comparisons of the kinematics and deep structures of the Zagros and Himalaya and of the Iranian and Tibetan plateaus and geodynamic implications. *Rev. Geophys.* 1–48. doi:10.1029/2009RG000304.1.INTRODUCTION
- Hemleben, C., Spindler, M. & Anderson, O.R., 1989. *Modern Planktonic Foraminifera*. Springer-Verlag. doi:10.1016/S0016-6995(89)80108-1
- Hohenegger, J., 2005. Estimation of environmental paleogradient values based on presence / absence data : a case study using benthic foraminifera for paleodepth estimation 217, 115–130. doi:10.1016/j.palaeo.2004.11.020
- Hopley, D., 1986. Beachrock as a sea-level indicator, in: van de Plassche, O. (Ed.), *Sea-Level Research: A Manual for the Collection and Evaluation of Data*. Free University, Amsterdam, pp. 157–174.
- Hopley, D., 2005. Trottoirs. *Encycl. Coast. Sci.*
- Hsü, K.J., Montadert, L., Bernoulli, D., Cita, M.B., Erickson, A., Garrison, R.E., Kidd, R.B., Mèlierés, F., Müller, C. & Wright, R., 1977. History of the Messinian salinity crisis. *Nature* 267, 399–403.

- Hsü, K.J., Montadert, L., Bernoulli, D., Cita, M.B., Erickson, A., Robert, E., Kidd, R.B., Mélières, F., Müller, C. & Wright, R., 1978. History of the Mediterranean salinity crisis. *Initial Reports Deep Sea Drill. Progr.* 42, 1053–1078.
- Hsü, K.J., Ryan, W.B.F. & Cita, M.B., 1973. Late Miocene Desiccation of the Mediterranean. *Nature* 242, 240–244. doi:10.1038/242240a0
- Hüsing, S.K., Dekkers, M.J., Franke, C. & Krijgsman, W., 2009. The tortonian reference section at monte dei corvi (Italy): Evidence for early remanence acquisition in greigite-bearing sediments. *Geophys. J. Int.* 179, 125–143. doi:10.1111/j.1365-246X.2009.04301.x
- Iaccarino, S.M. & Premoli Silva, I., 2007. *Practical Manual of Neogene Planktonic Foraminifera*.
- Imamura, F., Goto, K. & Ohkubo, S., 2008. A numerical model for the transport of a boulder by tsunami. *J. Geophys. Res.* 113, C01008. doi:10.1029/2007JC004170
- Jaffey, N. & Robertson, A.H.F., 2004. New sedimentological and structural data from the Ecemis Fault Zone, southern Turkey: implications for its timing and offset and the Cenozoic tectonic escape of Anatolia. *J. Geol. Soc. London* 158, 367–378. doi:10.1144/jgs.158.2.367
- Jenkins, J., Turner, B., Turner, R., Hayes, G.V. et al, 2013. Seismicity of the Earth 1900–2010 Middle East and Vicinity (ver. 1.1, Jan. 28, 2014). U.S. Geological Survey Open File Report 2010-1083-K, scale 1:7,000,000.
- Jolivet, L. & Faccenna, C., 2000. Mediterranean extension and the Africa-Eurasia collision. *Tectonics* 19, 1095–1106. doi:10.1029/2000TC900018
- Jordan, T.E., Nester, P.L., Blanco, N., Hoke, G.D., Dávila, F. & Tomlinson, A.J., 2010. Uplift of the Altiplano-Puna plateau: A view from the west. *Tectonics* 29. doi:10.1029/2010TC002661
- Jorissen, F.J., 1988. Benthic foraminifera from the Adriatic Sea: principles of phenotypic variation. *Utr. Micropaleontological Bull.* 37, 1–174.
- Kaiho, K., 1994. Benthic foraminiferal dissolved-oxygen index and dissolved-oxygen levels in the modern ocean. *Geology* 22, 719–722. doi:10.1130/0091-7613(1994)022<0719:BFDIOIA>2.3.CO
- Kalafat, D., Kekovalı, K., Güneş, Y.M., Yılmaz, M., Kara, P.D. & Berberoğlu, M., 2009. *A Catalogue of Source Parameters of Moderate and Strong Earthquakes for Turkey and its Surrounding Area (1938-2008)*, Boğaziçi University Publication. İstanbul.
- Kay, R.W. & Kay, S.M., 1993. Delamination and magmatism. *Tectonophysics* 219, 177–189. doi:10.1016/0040-1951(93)90295-U
- Kelletat, D., 2006. Beachrock as Sea-Level Indicator? Remarks from a Geomorphological Point of View. *J. Coast. Res.* 226, 1558–1564. doi:10.2112/04-0328.1
- Kelling, G., Robertson, A. & Van Buchem, F., 2005. Cenozoic sedimentary basins of southern Turkey: an introduction. *Sediment. Geol.* 173, 1–13. doi:10.1016/j.sedgeo.2004.03.013
- Kennett, J.P. & Srinivasan, S.M., 1983. *Neogene Planktonic Foraminifera: A Phylogenetic Atlas*. Hutchinson Ross Publishing Company, Pennsylvania.
- Keskin, M., 2003. Magma generation by slab steepening and breakoff beneath a subduction-accretion complex: An alternative model for collision-related volcanism in Eastern Anatolia, Turkey. *Geophys. Res. Lett.* 30, 7–10. doi:10.1029/2003GL018019
- Keskin, M., 2007. Eastern Anatolia: a hot spot in a collision zone without a mantle plume. *Geol. Soc. Am. Spec. Publ.* 1–40. doi:10.1029/2003GL018019
- Kortekaas, S. & Dawson, A.G., 2007. Distinguishing tsunami and storm deposits: An example from Martinhal, SW Portugal. *Sediment. Geol.* 200, 208–221. doi:10.1016/j.sedgeo.2007.01.004
- Kowalik, Z., 2012. *Introduction to Numerical Modeling of Tsunami Waves*. University of Alaska, Fairbanks.
- Krestenitis, Y.N., Androulidakis, Y.S., Kontos, Y.N. & Georgakopoulos, G., 2011. Coastal inundation in the north-eastern mediterranean coastal zone due to storm surge events. *J. Coast. Conserv.* 15, 353–368.

- Krijgsman, W., Hilgen, F.J., Raf, I., Sierro, F.J. & Annunzio, Á.G.D., 1999. Chronology , causes and progression of the Messinian salinity crisis. *Nature* 400, 652–655.
- Laborel, J., 2010. Algal Rims. *Encycl. Mod. Coral Reefs*.
- Laborel, J. & Laborel-Deguen, F., 2016. Biological Indicators of Relative Sea-Level Variations and of Co-Seismic Displacements in the Mediterranean Region. *J. Coast. Res.* 10, 395–415.
- Lambeck, K., Antonioli, F., Purcell, A. & Silenzi, S., 2004a. Sea-level change along the Italian coast for the past 10,000 yr. *Quat. Sci. Rev.* 23, 1567–1598. doi:10.1016/j.quascirev.2004.02.009
- Lambeck, K., Anzidei, M., Antonioli, F., Benini, A. & Esposito, A., 2004b. Sea level in Roman time in the Central Mediterranean and implications for recent change. *Earth Planet. Sci. Lett.* 224, 563–575. doi:10.1016/j.epsl.2004.05.031
- Lambeck, K. & Chappell, J., 2001. Sea level change through the last glacial cycle. *Science* (80-.). 292, 679–686. doi:10.1126/science.1059549
- Lambeck, K. & Purcell, A., 2005. Sea-level change in the Mediterranean Sea since the LGM: Model predictions for tectonically stable areas. *Quat. Sci. Rev.* 24, 1969–1988. doi:10.1016/j.quascirev.2004.06.025
- Lau, A.Y.A., Terry, J.P., Switzer, A.D. & Pile, J., 2015. Advantages of beachrock slabs for interpreting high-energy wave transport: Evidence from Luda Island in south-eastern Taiwan. *Geomorphology* 228, 263–274. doi:10.1016/j.geomorph.2014.09.010
- Liu, H., Sakashita, T. & Sato, S., 2014. An experimental study on the tsunami boulder movement, in: *Proceedings of the 34th International Conferene of the Coastal Engineering*, ASCE, 2014.
- Lourens, L., Hilgen, F., Shackleton, N.J., Laskar, J. & Wilson, J., 2004. Appendix 2: Orbital tuning calibrations and conversions for the Neogene Period, in: Gradstein, F., Ogg, J.G., Smith, A. (Eds.), *A Geologic Time Scale 2004*. Cambridge University Press, pp. 469–484.
- Lourens, L.J., Antonarakou, A., Hilgen, F.J., Van Hoof, A.A.M., Vergnaud-Grazzini, C. & Zachariasse, W.J., 1996a. Evaluation of the Plio-Pleistocene astronomical timescale. *Paleoceanography* 11, 391–413. doi:10.1029/96PA01125
- Lourens, L.J., Hilgen, F.J. & Raffi, I., 1998. Base of large Gephyrocapsa and astronomical calibration of early Pleistocene sapropels in Site 967 and Hole 969D: solving the chronology of the Vrica Section (Calabria, Italy), in: Robertson, A.H., Emeis, K.-C., Richter, C., Camerlenghi, A. (Eds.), *Proc. ODP, Sci. Results*. pp. 191–197.
- Lourens, L.J., Hilgen, F.J., Raffi, I. & Vergnaud-Grazzini, C., 1996b. Early Pleistocene chronology of the Vrica Section (Calabria, Italy). *Paleoceanography* 11, 797. doi:10.1029/96PA02691
- Malz, H. & Jellinek, T., 1994. Podocopide Tiefsee-Ostracoden aus Kastengreifer-Proben im ostlichen Mittelmeer ('Meteor'-Fahrt 25/leg 1: Ionisches bis Levantinisches Becken. *Senckenbergiana Lethaea* 74, 9–32.
- Maouche, S., Morhange, C. & Meghraoui, M., 2009. Large boulder accumulation on the Algerian coast evidence tsunami events in the western Mediterranean. *Mar. Geol.* 262, 96–104. doi:10.1016/j.margeo.2009.03.013
- Marriner, N. & Morhange, C., 2007. Geoscience of ancient Mediterranean harbours. *Earth Sci. Rev.* 80, 137–194. doi:10.1016/j.earscirev.2006.10.003
- Mastronuzzi, G., 2010. Tsunami in Mediterranean Sea. *Egypt J Env. Chang.* 2, 1–12.
- McQuarrie, N. & Van Hinsbergen, D.J.J., 2013. Retrodeforming the Arabia-Eurasia collision zone: Age of collision versus magnitude of continental subduction. *Geology* 41, 315–318. doi:10.1130/G33591.1
- Meier, T., Becker, D., Endrun, B., Rische, M., Bohnhoff, M., Stockhert, B. & Harjes, H.-P., 2007. A model for the Hellenic subduction zone in the area of Crete based on seismological investigations. *Geol. Soc. London, Spec. Publ.* 291, 183–199. doi:10.1144/SP291.9
- Mendes, I., Dias, J.A., Schönfeld, J. & Ferreira, O., 2012. Distribution of living benthic foraminifera on the

- northern Gulf of Cadiz continental shelf. *J. Foraminifer. Res.* 42, 18–38.
- Meriç, E., Avşar, N., Yokeş, M.B. & Dinçer, F., 2014. Atlas of recent benthic foraminifera from Turkey. *Micropaleontology* 60, 211–398.
- Merritts, D. & Bull, W.W., 1989. Interpreting Quaternary uplift rates at the Mendocino triple junction, northern California, from uplifted marine terraces. *Geology* 17, 1020–1024.
- Milker, Y. & Schmiedl, G., 2012. A taxonomic guide to modern benthic shelf foraminifera of the western Mediterranean Sea. *Palaeontol. Electron.* 15, 134.
- Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N. & Pekar, S.F., 2005. The Phanerozoic Record of Global Sea-Level Change. *Science* (80-.). 310, 1293–1298. doi:10.1126/science.1116412
- Milliman, J.D. & Emery, K.O., 1968. Sea Levels during the Past 35,000 Years. *Science* (80-.). 162, 1121–1123. doi:doi: 10.1126/science.162.3858.1121
- Molnar, P., England, P. & Martinod, J., 1993. Mantle dynamics, uplift of the Tibetan Plateau, and the Indian Monsoon. *Rev. Geophys.* 31, 357. doi:10.1029/93RG02030
- Montgomery, D.R., Balco, G. & Willett, S.D., 2001. Climate tectonic and the morphology of the Andes. *Geology* 29, 579–582. doi:10.1130/0091-7613(2001)029<0579:CTATMO>2.0.CO;2
- Morhange, C., Pirazzoli, P.A., Marriner, N., Montaggioni, L.F. & Nammour, T., 2006. Late Holocene relative sea-level changes in Lebanon, Eastern Mediterranean. *Mar. Geol.* 230, 99–114. doi:10.1016/j.margeo.2006.04.003
- Morton, R.A., Gelfenbaum, G. & Jaffe, B.E., 2007. Physical criteria for distinguishing sandy tsunami and storm deposits using modern examples. *Sediment. Geol.* 200, 184–207. doi:10.1016/j.sedgeo.2007.01.003
- Murray, J., 2006. *Ecology and Applications of Benthic Foraminifera*. Cambridge University Press, Cambridge.
- Nanayama, F., Shigeno, K., Satake, K., Shimokawa, K., Koitabashi, S., Miyasaka, S. & Ishii, M., 2000. Sedimentary differences between the 1993 Hokkaido-nansei-oki tsunami and the 1959 Miyakojima typhoon at Taisei, southwestern Hokkaido, northern Japan. *Sediment. Geol.* 135, 255–264. doi:http://dx.doi.org/10.1016/S0037-0738(00)00076-2
- Nandasena, N.A.K., Paris, R. & Tanaka, N., 2011. Reassessment of hydrodynamic equations: Minimum flow velocity to initiate boulder transport by high energy events (storms, tsunamis). *Mar. Geol.* 281, 70–84. doi:10.1016/j.margeo.2011.02.005
- Nazik, A., 2004. Planktonic foraminiferal biostratigraphy of the Neogene sequence in the Adana Basin , Turkey , and its correlation with standard biozones. *Geol. Mag.* 141, 379–387. doi:10.1017/S0016756804009148
- Noller, S.J., Zomeni, Z. & Panayides, I., 2011. Preliminary field and GIS based assessment of tsunami hazard on Cyprus. *CIESM Work. Monogr* 42, 49–58.
- Noormets, R., Crook, K.A.W. & Felton, E.A., 2004. Sedimentology of rocky shorelines: 3. Hydrodynamics of megaclast emplacement and transport on a shore platform, Oahu, Hawaii. *Sediment. Geol.* 172, 41–65. doi:10.1016/j.sedgeo.2004.07.006
- Nott, J., 2003a. Tsunami or Storm Waves ? -Determining the Origin of a Spectacular Field of Wave Emplaced Boulders Using Numerical Storm Surge and Wave Models and Hydrodynamic Transport Equations. *J. Coast. Res.* 19, 348–356.
- Nott, J., 2003b. Waves, coastal boulder deposits and the importance of the pre-transport setting. *Earth Planet. Sci. Lett.* 210, 269–276. doi:10.1016/S0012-821X(03)00104-3
- Okay, A.I., 2008. Geology of Turkey : A synopsis. *Anschnitt* 21, 19–42.
- Okay, A.I., Tansel, I. & Tüysüz, O., 2001. Obduction, subduction and collision as reflected in the Upper Cretaceous–Lower Eocene sedimentary record of western Turkey. *Geol. Mag.* 138, 117–142. doi:10.1017/S0016756801005088
- Okay, A.I., Zattin, M. & Cavazza, W., 2010. Apatite fission-track data for the Miocene Arabia-Eurasia collision.

- Okyar, M., Ergin, M. & Evans, G., 2005. Seismic stratigraphy of Late Quaternary sediments of western Mersin Bay shelf, (NE Mediterranean Sea). *Mar. Geol.* 220, 113–130. doi:10.1016/j.margeo.2005.06.024
- Ola, P. & Adewale, B., 2014. Palynostratigraphy and Paleoclimate of the Sequences Penetrated by Meren 31 Side Tract-2 Well, Offshore Niger Delta. *Int. J. Geosci.* 5, 1206–1218.
- Ozel, E., Ulug, A. & Pekcetinöz, B., 2007. Neotectonic aspects of the northern margin of the Adana-Cilicia submarine basin, NE Mediterranean. *J. Earth Syst. Sci.* 116, 113–124. doi:10.1007/s12040-007-0011-9
- Öğretmen, N., Cosentino, D., Gliozzi, E., Cipollari, P., Iadanza, A. & Yıldırım, C., 2015. Tsunami hazard in the Eastern Mediterranean: geological evidence from the Anatolian coastal area (Silifke, southern Turkey). *Nat. Hazards* 79, 1569–1589. doi:10.1007/s11069-015-1916-2
- Özgül, N., 1976. Toroslar'ın bazı temel jeolojik özellikleri (Some geological aspects of the Taurus orogenic belt (Turkey)). *Bull. Geol. Soc. Turkey* 19, 65–78.
- Özgül, N., 1984. Stratigraphy and tectonic evolution of the Central Taurides. *MTA Derg.* 77–90.
- Papadopoulos, G., 2016a. Historical and Geological Evidence of Tsunamis in Europe and the Mediterranean, in: *Tsunamis in the European-Mediterranean Region*. Elsevier Inc., pp. 39–76. doi:10.1016/B978-0-12-420224-5/00002-8
- Papadopoulos, G., 2016b. Tsunamigenic Sources and Generation Mechanisms, in: *Tsunamis in the European-Mediterranean Region*. Elsevier Inc., pp. 101–119. doi:10.1016/B978-0-12-420224-5/00004-1
- Papadopoulos, G.A. & Kortekaas, S., 2003. Characteristics of landslide generated tsunamis from observational data, in: L, L., J, M. (Eds.), *Submarine Mass Movements and Their Consequences*. Kluwer Academic Publisher, Dordrecht, pp. 367–374.
- Papadopoulos, G. A., Gràcia, E., Urgeles, R., Sallares, V., De Martini, P.M., Pantosti, D., González, M., Yalciner, A.C., Mascle, J., Sakellariou, D., Salamon, A., Tinti, S., Karastathis, V., Fokaefs, A., Camerlenghi, A., Novikova, T. & Papageorgiou, A., 2014. Historical and pre-historical tsunamis in the Mediterranean and its connected seas: Geological signatures, generation mechanisms and coastal impacts. *Mar. Geol.* 354, 81–109. doi:10.1016/j.margeo.2014.04.014
- Paris, R., Naylor, L.A. & Stephenson, W.J., 2011. Boulders as a signature of storms on rock coasts. *Mar. Geol.* 283, 1–11. doi:10.1016/j.margeo.2011.03.016
- Paris, R., Wassmer, P., Sartohadi, J., Lavigne, F., Barthomeuf, B., Desgages, E., Grancher, D., Baumert, P., Vautier, F., Brunstein, D. & Gomez, C., 2009. Geomorphology Tsunamis as geomorphic crises : Lessons from the December 26 , 2004 tsunami in Lhok Nga , West Banda Aceh (Sumatra , Indonesia). *Geomorphology* 104, 59–72. doi:10.1016/j.geomorph.2008.05.040
- Parker, F.L., 1958. Eastern Mediterranean Foraminifera. *Rep. Swed. Deep. Exped.* 1947–48, 219–283.
- Parker, F.L., 1962. Planktonic foraminiferal species in Pacific sediments. *Micropalontology* 8, 219–254.
- Parlak, O., Höck, V. & Delaloye, M., 2002. The supra-subduction zone Pozanti - Karsanti ophiolite, southern Turkey: Evidence for high-pressure crystal fractionation of ultramafic cumulates (PDF Download Available).pdf. *Lithos* 1–2, 205–224.
- Parlak, O. & Robertson, A., 2004. The ophiolite-related Mersin Melange, southern Turkey: its role in the tectonic–sedimentary setting of Tethys in the Eastern Mediterranean region. *Geol. Mag.* 141, 257–286. doi:10.1017/S0016756804009094
- Perg, L.A., Anderson, R.S. & Finkel, R.C., 2001. Use of a new ¹⁰Be and ²⁶Al inventory method to date marine terraces, Santa Cruz, California, USA. *Geology* 29, 879–882. doi:10.1130/0091-7613(2001)029<0879:UOANBA>2.0.CO;2
- Pignatelli, C., Sansò, P. & Mastronuzzi, G., 2009. Evaluation of tsunami flooding using geomorphologic evidence. *Mar. Geol.* 260, 6–18. doi:10.1016/j.margeo.2009.01.002
- Pirazzoli, P.A., 1986a. Marine notches, in: Van de Plassche, O. (Ed.), *Sea-Level Research: A Manual for the Collection and Evaluation of Data*. Free University, Amsterdam, pp. 361–400.

- Pirazzoli, P.A., 1986b. The Early Byzantine tectonic paroxysm. *Zeitschrift fur Geomorphol.* 62, 31–49.
- Pirazzoli, P.A. & Evelpidou, N., 2013. Tidal notches : A sea-level indicator of uncertain archival trustworthiness. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 369, 377–384. doi:10.1016/j.palaeo.2012.11.004
- Pirazzoli, P.A., Laborel, J., Saliège, J.F., Erol, O., Kayan, İ. & Person, A., 1991. Holocene raised shorelines on the Hatay coasts (Turkey): Palaeoecological and tectonic implications. *Mar. Geol.* 96, 295–311.
- Poirier, J.P. & Taher, M.A., 1980. Historical seismicity in the Near and Middle East, North Africa, and Spain from Arabic documents (VIIth-XVIIIth century). *Bull. Seismol. Soc. Am.* 70, 2185–2201.
- Poole, A.J. & Robertson, A.H.F., 1991. Quaternary uplift and sea-level change at an active plate boundary, Cyprus. *J. Geol. Soc. London* 148, 909–921. doi:10.1144/gsjgs.148.5.0909
- Pujol, C. & Vergnaud Grazzini, C., 1995. Distribution patterns of live planktic foraminifers as related to regional hydrography and productive systems of the Mediterranean Sea. *Mar. Micropaleontol.* 25, 187–217. doi:10.1016/0377-8398(95)00002-I
- Radeff, G., Cosentino, D., Cipollari, P., Schildgen, T.F., Iadanza, A., Streker, M.R., Darbaş, G. & Gürbüz, K., 2016. Stratigraphic architecture of the upper Messinian deposits of the Adana Basin (southern Turkey): implications for the Messinian Salinity Crisis and the Taurus petroleum system. *Ital. J. Geosci.* 135, 408–424. doi:http://doi.org/10.3301/IJG.2015.18
- Radeff, G., Schildgen, T.F., Cosentino, D., Strecker, M.R., Cipollari, P., Darbaş, G. & Gürbüz, K., 2017. Sedimentary evidence for late Messinian uplift of the SE margin of the Central Anatolian Plateau: Adana Basin, southern Turkey. *Basin Res.* 1–27. doi:10.1111/bre.12159
- Raffi, I., 2002. Revision of the early-middle pleistocene calcareous nannofossil biochronology (1.75-0.85 Ma). *Mar. Micropaleontol.* 45, 25–55.
- Raffi, I., Backman, J., Fornaciari, E., Pa, H. & Rio, D., 2006. A review of calcareous nannofossil astrobiochronology encompassing the past 25 million years. *Quat. Sci. Rev.* 25, 3113–3137. doi:10.1016/j.quascirev.2006.07.007
- Reilinger, R., McClusky, S., Vernant, P., Lawrence, S., Ergintav, S., Cakmak, R., Ozener, H., Kadirov, F., Guliev, I., Stepanyan, R., Nadariya, M., Hahubia, G., Mahmoud, S., Sakr, K., ArRajehi, A., Paradissis, D., Al-Aydrus, A., Prilepin, M., Guseva, T., Evren, E., Dmitrova, A., Filikov, S. V., Gomez, F., Al-Ghazzi, R. & Karam, G., 2006. GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions. *J. Geophys. Res. Solid Earth* 111, 1–26. doi:10.1029/2005JB004051
- Rio, D., Raffi, I. & Villa, G., 1990. Pliocene-Pleistocene calcareous nannofossil distribution patterns in the Western Mediterranean, in: Kastens, K.A., Mascle, J. (Eds.), *Proc. ODP, Sci. Results*. pp. 513–533.
- Robertson, A.H.F., Unlüğenç, Ü.C., İnan, N. & Taşlı, K., 2004. The Misis–Andırın Complex: a Mid-Tertiary melange related to late-stage subduction of the Southern Neotethys in S Turkey. *J. Asian Earth Sci.* 22, 413–453. doi:10.1016/S1367-9120(03)00062-2
- Robertson, A.H.F. & Dixon, J.E., 1984. Introduction: aspects of the geological evolution of the Eastern Mediterranean. *Geol. Soc. London, Spec. Publ.* 17, 1–74. doi:10.1144/GSL.SP.1984.017.01.02
- Robertson, A.H.F. & Shipboard Scientific Party, 1996. Role of Eratosthenes Seamount in collisional processes in the Eastern Mediterranean, in: Emeis, K.-C., Robertson, A.H.F., Richter, C. (Eds.), *Proc. ODP, Initial Reports*. pp. 513–520.
- Robertson, A.H.F., Ustaomer, T., Pickett, E.A., Collins, A.S., Andrew, T. & Dixon, J.E., 2004. Testing models of Late Palaeozoic-Early Mesozoic orogeny in Western Turkey: support for an evolving open-Tethys model. *J. Geol. Soc. London*. 161, 501–511. doi:10.1144/0016-764903-080
- Rouchy, J.M., Orszag-Sperber, F., Blanc-Valleron, M.M., Pierre, C., Rivière, M., Combourieu-Nebout, N. & Panayides, I., 2001. Palaeoenvironmental changes at the Messinian-Pliocene boundary in the Eastern Mediterranean (southern Cyprus basins): Significance of the Messinian Lago-Mare. *Sediment. Geol.* 145, 93–117. doi:10.1016/S0037-0738(01)00126-9
- Royden, L.H., Burchfiel, B.C., King, R.W., Wang, E., Chen, Z., Shen, F. & Liu, Y., 1994. Surface deformation and lower crustal flow in Eastern Tibet. *Science* (80-.). 276, 788–790.

- Ruddiman, W.F. & Kutzbach, J.E., 1989. Forcing of late Cenozoic northern hemisphere climate by plateau uplift in southern Asia and the American west. *J. Geophys. Res.* 94, 18409. doi:10.1029/JD094iD15p18409
- Ruggieri, G., 1967. Due ostracofaune del Miocene alloctono della Val Marecchia (Appennino settentrionale). *Riv. Ital. di Paleontol. e Stratigr.* 73, 352–384.
- Russo, B., Curcio, E. & Iaccarino, S., 2007. Paleoeecology and paleoceanography of a Langhian succession (Tremi Islands, southern Adriatic Sea, Italy) based on benthic foraminifera. *Boll. della Soc. Paleontol. Ital.* 46, 107–124.
- Ryan, W.B.F., 2009. Decoding the Mediterranean salinity crisis. *Sedimentology* 56, 95–136. doi:10.1111/j.1365-3091.2008.01031.x
- Salamon, A., Rockwell, T., Guidoboni, E. & Comastri, A., 2009. A critical evaluation of tsunami records reported for the Levant Coast from the second millennium BCE to the present. *Isr. J. Earth Sci.* 58, 327–354. doi:10.1560/IJES.58.2
- Salamon, A., Rockwell, T., Ward, S.N., Guidoboni, E. & Comastri, A., 2007. Tsunami hazard evaluation of the Eastern Mediterranean : historical analysis and selected modeling. *Bull. Seismol. Soc. Am.* 97, 5–9. doi:10.1785/0120060147
- Salamon, A., Hofstetter, A., Garfunkel, Z. & Ron, H., 1996. Seismicity of the eastern Mediterranean region: Perspective from the Sinai subplate. *Tectonophysics* 263, 293–305. doi:10.1016/S0040-1951(96)00030-3
- Schattner, U., 2010. What triggered the early-to-mid Pleistocene tectonic transition across the entire eastern Mediterranean? *Earth Planet. Sci. Lett.* 289, 539–548. doi:10.1016/j.epsl.2009.11.048
- Scheffers, A., 2004. Tsunami imprints on the Leeward Netherlands Antilles (Aruba, Curaçao, Bonaire) and their relation to other coastal problems. *Quat. Int.* 120, 163–172. doi:http://dx.doi.org/10.1016/j.quaint.2004.01.015
- Scheffers, A., 2008. Tsunami Boulder Deposits, in: Shiki, T., Tsuji, Y., Yamazaki, T., Minoura, K. (Eds.), *Tsunamiites - Features and Implications*. Elsevier B.V., pp. 299–317. doi:10.1016/B978-0-444-51552-0.00017-5
- Scheffers, A. & Kelletat, D., 2003. Sedimentologic and geomorphologic tsunami imprints worldwide - A review. *Earth-Science Rev.* 63, 83–92. doi:10.1016/S0012-8252(03)00018-7
- Schildgen, T.F., Cosentino, D., Bookhagen, B., Niedermann, S., Yıldırım, C., Echtler, H., Wittmann, H. & Strecker, M.R., 2012a. Multi-phased uplift of the southern margin of the Central Anatolian plateau, Turkey: A record of tectonic and upper mantle processes. *Earth Planet. Sci. Lett.* 317–318, 85–95. doi:10.1016/j.epsl.2011.12.003
- Schildgen, T.F., Cosentino, D., Caruso, A., Buchwaldt, R., Yıldırım, C., Bowring, S. A., Rojay, B., Echtler, H. & Strecker, M.R., 2012b. Surface expression of eastern Mediterranean slab dynamics: Neogene topographic and structural evolution of the southwest margin of the Central Anatolian Plateau, Turkey. *Tectonics* 31, n/a-n/a. doi:10.1029/2011TC003021
- Schildgen, T.F., Yıldırım, C., Cosentino, D. & Strecker, M.R., 2014. Linking slab break-off, Hellenic trench retreat, and uplift of the Central and Eastern Anatolian plateaus. *Earth-Science Rev.* 128, 147–168. doi:10.1016/j.earscirev.2013.11.006
- Sciuto, F., 2003. Dati preliminari sulla ostracofauna pliocenica di Capo Milazzo (Sicilia NE). *Boll. della Soc. Paleontol. Ital.* 42, 179–184.
- Sciuto, F. & Rosso, A., 2008. Distribution pattern of deep-water ostracod assemblages from Lower Pleistocene sediments from Furnari, Sicily. *Boll. della Soc. Paleontol. Ital.* 47, 33–43.
- Sciuto, F., Rosso, A. & Torrisi, S., 2003. Associazioni ad ostracodi del Pliocene di Centuri interpretazione paleoambientale. *Boll. della Soc. Paleontol. Ital.* 42, 267–280.
- Sen Gupta, B.K. & Machain-Castillo, M.L., 1993. Benthic foraminifera in oxygen-poor habitats. *Mar. Micropaleontol.* 20, 183–201.
- Serpelloni, E., Faccenna, C., Spada, G., Dong, D. & Williams, S.D.P., 2013. Vertical GPS ground motion rates in the Euro-Mediterranean region: New evidence of velocity gradients at different spatial scales along the

- Nubia-Eurasia plate boundary. *J. Geophys. Res. Solid Earth* 118, 6003–6024. doi:10.1002/2013JB010102
- Sgarrella, F. & Moncharmont Zei, M., 1993. Benthic Foraminifera of the Gulf of Naples (Italy): systematics and autoecology. *Boll. della Soc. Paleontol. Ital.* 32, 145–264.
- Shackleton, N.J. & Crowhurst, S., 1997. Sediment fluxes based on an orbitally tuned time scale 5 Ma to 14 Ma, Site 926, in: Shackleton, N.J., Curry, W.B., Richter, C., Bralower, T.J. (Eds.), *Proc. ODP, Sci. Results.* pp. 69–82.
- Shiki, T. & Yamazaki, T., 2008. The Term ‘Tsunamiite’, in: Shiki, T., Tsuji, Y., Yamazaki, T. & Minoura, K. (Eds.), *Tsunamiites - Features and Implications.* Elsevier B.V., pp. 5–7. doi:10.1016/B978-0-444-51552-0.00002-3
- Sissingh, W., 1972. Late Cenozoic Ostracoda of the South Aegean Island Arc. *Utr. Micropaleontological Bull.* 6, 1–187.
- Spezzaferri, S. & Tamburini, F., 2007. Paleodepth variations on the Eratosthenes Seamount (Eastern Mediterranean): sea-level changes or subsidence? *eEarth Discuss.* 2, 115–132. doi:10.5194/eed-2-115-2007
- Stampfli, G.M., 2000. Tethyan oceans. *Geol. Soc. London, Spec. Publ.* 173, 1–23. doi:10.1144/GSL.SP.2000.173.01.01
- Stefanelli, S. & Capotondi, L., 2008. Foraminiferal response to the deposition of insolation cycle 90 sapropel in different Mediterranean areas. *J. Micropaleontology* 27, 45–61. doi:10.1144/jm.27.1.45
- Stewart, I. & Morhange, C., 2009. Coastal geomorphology and sea-level change, in: Woodward, J.C. (Ed.), *The Physical Geography of the Mediterranean.* Oxford University Press, Oxford, pp. 385–414.
- Strecker, M.R., Alonso, R.N., Bookhagen, B., Carrapa, B., Hilley, G.E., Sobel, E.R. & Trauth, M.H., 2007. Tectonics and Climate of the Southern Central Andes. doi:10.1146/annurev.earth.35.031306.140158
- Şafak, Ü., 1997. Karaman yöresi üst Miyosen-Pliyosen istifinin ostrakod faunası ve ortamsal yorumu. *MTA Derg.* 119, 89–102.
- Şafak, Ü., Kelling, G., Gökçen, N.S. & Gürbüz, K., 2005. The mid-Cenozoic succession and evolution of the Mut basin, southern Turkey, and its regional significance. *Sediment. Geol.* 173, 121–150. doi:10.1016/j.sedgeo.2004.03.012
- Şenel, M., 2002. Geological Map of Turkey, Konya (No. 14) scale 1:500,000, 1 sheet. *MTA Derg.*
- Şengör, A.M. & Kidd, W.S., 1979. Post-collisional tectonics of the Turkish-Iranian plateau and a comparison with Tibet. *Tectonophysics* 55, 361–376. doi:10.1016/0040-1951(79)90184-7
- Şengör, A.M.C., 2003. East Anatolian high plateau as a mantle-supported, north-south shortened domal structure. *Geophys. Res. Lett.* 30, 2–5. doi:10.1029/2003GL017858
- Şengör, A.M.C., Görür, N. & Şaroğlu, F., 1985. Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as an example, in: Biddle, K.T., Christie-Blick, N. (Eds.), *Strike-Slip Deformation, Basin Formation, and Sedimentation.* The SEPM, pp. 227–264. doi:10.2110/pec.85.37.0227
- Şengör, A.M.C., Görür, N. & Şaroğlu, F., 1985. Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as a case study. *SEPM* 37, 227–264.
- Şengör, A.M.C., Özeren, M.S., Keskin, M., Sakıncı, M., Özbakır, A.D. & Kayan, I., 2008. Eastern Turkish high plateau as a small Turkic-type orogen: Implications for post-collisional crust-forming processes in Turkic-type orogens. *Earth-Science Rev.* 90, 1–48. doi:10.1016/j.earscirev.2008.05.002
- Şengör, A.M.C. & Yılmaz, Y., 1981. Tethyan evolution of Turkey: A plate tectonic approach. *Tectonophysics* 75, 181–241. doi:10.1016/0040-1951(81)90275-4
- Tanar, Ü. & Gökçen, N., 1990. Mut-Ermenek Tersiyer istifinin stratigrafisi ve mikropaleontolojisi. *MTA Derg.* 175–180.
- Thomas, M.L.H., 1996. Origin and community structure of the harrington sound notch, Bermuda. *Bull. Mar. Sci.* 58, 753–763.

- Triantaphyllou, M.V., Antonarakou, A., Drinia, H., Dimiza, M.D., Kontakiotis, G., Tsolakis, E., & Theodorou, G., 2010. High resolution biostratigraphy and paleoecology of the early Pliocene successions of Pissouri Basin (Cyprus Island), in: *Proceedings of the 12th International Congress. Bulletin of the Geological Society of Greece, Patras*, pp. 763–772.
- Trudgill, S.T., 1987. Bioerosion of intertidal limestone, Co. Clare, Eire - 3: zonation, process and form. *Mar. Geol.* 74, 111–121. doi:10.1016/0025-3227(87)90009-9
- Tsourou, T., Drinia, H., Antonarakou, A. & Lianou, V., 2012. A rich ostracod fauna from Cape Kalogeras (Zakynthos Island, Greece), in: *9th Pan-Hellenic Symposium of Oceanography and Fisheries. Athens*.
- Turner, R. j., 2005. Beachrock. *Encycl. Coast. Sci.*
- Tuttle, M.P., Ruffman, A., Anderson, T. & Jeter, H., 2004. Distinguishing tsunami from storm deposits in eastern America: The 1929 Grand Banks tsunami versus the 1991 Halloween storm. *Seismol. Res. Lett.* 75, 117–131.
- van Couvering, J.A. & Miller, J.A., 1971. Late Miocene Marine and Non-marine Time Scale in Europe. *Nature* 230, 559–563.
- Van Harten, D. & Droste, H.J., 1988. Mediterranean Deep-sea Ostracods, the Species Poorness of the Eastern Basin as a Legacy of an Early Holocene Anoxic Event, in: Hanai, T., Ikeya, N., Ishizaki, K. (Eds.), *Evolutionary Biology of Ostracoda: Its Fundamentals and Applications*. Elsevier, Amsterdam, pp. 721–737.
- van Hunen, J & Allen, M.B., 2011. Continental collision and slab break-off: A comparison of 3-D numerical models with observations. *Earth Planet. Sci. Lett.* 302, 27–37. doi:10.1016/j.epsl.2010.11.035
- Villanueva-Guimerans; Canudo, I., 2008. Assemblages of recent benthic foraminifera from the northeastern Gulf of Cadiz. *Geogaceta* 44, 263–278.
- Violanti, D., Pierre, F. Dela, Trenkwalder, S., Lozar, F., Clari, P., Irace, A. & D’Atri, A., 2011. Biostratigraphic and palaeoenvironmental analyses of the Messinian/Zanclean boundary and Zanclean succession in the Moncucco quarry (Piedmont, northwestern Italy). *Bull. la Soc. Geol. Fr.* 182, 149–162. doi:10.2113/gssgfbull.182.2.149
- Violanti, D., Trenkwalder, S., Lozar, F. & Gallo, L.M., 2009. Micropalaeontological analyses of the Narzole core: Biostratigraphy and palaeoenvironment of the late messinian and early zanclean of piedmont (Northwestern Italy). *Boll. della Soc. Paleontol. Ital.* 48, 167–181.
- Wade, B.S. & Bown, P.R., 2006. Calcareous nannofossils in extreme environments : The Messinian Salinity Crisis, Polemi Basin, Cyprus. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 233, 271–286. doi:10.1016/j.palaeo.2005.10.007
- WEAORC, 2004. Western European Armaments Organization Research Cell-Western European Union: Wind and Wave Atlas of the Mediterranean Sea, April 2004 [WWW Document]. URL http://users.ntua.gr/mathan/pdf/Pages_
- Weiss, R., 2012. The mystery of boulders moved by tsunamis and storms. *Mar. Geol.* 295–298, 28–33. doi:10.1016/j.margeo.2011.12.001
- Westaway, R., 1993. Quaternary uplift of southern Italy. *J. Geophys. Res. Solid Earth* 98, 21741–21772. doi:10.1029/93JB01566
- Yıldız, A., Toker, V., Demircan, H. & Sevim, S., 2003. Mut Havzası Pliyosen-Pleyistosen nannoplankton, planktik foraminifera, iz fosil bulguları ve paleoortam yorumu. *Yerbilimleri* 28, 123–144.
- Yolsal, S., Taymaz, T. & Yalciner, A.C., 2007. Understanding tsunamis, potential source regions and tsunami-prone mechanisms in the Eastern Mediterranean. *Geol. Soc. London, Spec. Publ.* 291, 201–230. doi:10.1144/SP291.10
- Young, R.W., Bryant, E.A. & Price, D.M., 1996. Catastrophic wave (tsunami?) transport of boulders in southern New South Wales, Australia. *Zeitschrift fur Geomorphol.* 40, 191–207.
- Zazo, C., Goy, J.L., Dabrio, C.J., Bardají, T., Hillaire-Marcel, C., Ghaleb, B., González-Delgado, J.Á. & Soler, V., 2003. Pleistocene raised marine terraces of the Spanish Mediterranean and Atlantic Coasts: Records of

coastal uplift, sea-level highstands and climate changes. *Mar. Geol.* 194, 103–133. doi:10.1016/S0025-3227(02)00701-6

Zhisheng, A., Kutzbach, J.E., Prell, W.L. & Porter, S.C., 2001. Evolution of Asian monsoons and phased uplift of the Himalaya-Tibetan plateau since Late Miocene times. *Nature* 411, 62–66.



SUPPLEMENTARY MATERIAL

Table 1: Stratigraphic markers recognized in Gülnar East and Tol sections used to constrain the age model.

Stratigraphic markers	Occurrence in the Gülnar Est section (m)	Occurrence in the Tol section (m)	Correlated events	Age (Ma)	Literature
<i>Neogloboquadrina</i> spp. (sin)		base	dissapperance	0.61	Gradstein et al., 2004
<i>G. calida</i>	15.40	4.10	bottom	0.78	Iaccarino and Premoli Silva, 2007
Paracme <i>Neogloboquadrina</i> spp.(sin)	14.50		top paracme	1.209	Lourens et al., 1998
Paracme <i>Neogloboquadrina</i> spp.(sin)	7.80		bottom paracme	1.37	Lourens et al., 1998
<i>N. pachyderma</i> sx	6.45		bottom paracme	1.367-1.370	Iaccarino and Premoli Silva, 2007
<i>B. marginata</i>	4.40		bottom	1.543	Cita et al, 2008
large <i>Gephyrocapsa</i>	3.95		first occurrence in the section	< 1.617	Raffi, 2002
5th sapropel	3.95		ODP 967 Sapropel 38	1.564	Lourens et al., 1998
4th sapropel	3.55		ODP 967 Sapropel 39	1.584	Lourens et al., 1998
3rd sapropel	2.90		ODP 967 Sapropel 40	1.603	Lourens et al., 1998
2nd sapropel	2.00		ODP 967 Sapropel 41	1.522	Lourens et al., 1998
1st sapropel	1.50		ODP 967 Sapropel 42	1.542	Lourens et al., 1998
<i>G. crassaformis</i>	0.30		first influx	1.722-1.723	Lourens et al., 1998
medium <i>Gephyrocapsa</i>	0.30		first occurrence in the section	< 1.73	Raffi, 2002

Table 2: Autochthonous benthic foraminifers from Gülnar East Section

Benthic foraminifer species	Bathymetric range	Optimum bathymetric zones	Literature
<i>Amphicoryna scalaris</i>	infralittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Amphicoryna sublineata</i>		circalittoral	Di Bella et al., 2000
<i>Angulogerina angulosa</i>	infralittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Astacolus crepidulus</i>	infralittoral-epibathyal	epibathyal	Blanc-Vernet, 1969; Sgarrella & Moncharmont Zei, 1993
<i>Astacolus</i> sp.			
<i>Astrononion stelligerum</i>	infralittoral-upper epibathyal		Spezzaferri & Tamburini, 2007
<i>Bolivina alata</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bolivina albatrossi</i>	circalittoral-epibathyal	lower epibathyal	Sgarrella & Moncharmont Zei, 1994
<i>Bolivina catanensis</i>	infralittoral-epibathyal	circalittoral-epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Di Bella 2010
<i>Bolivina dilatata</i>	infralittoral-epibathyal		Spezzaferri & Tamburini, 2007; Goineau et al., 2012; Milker & Schmiedl, 2012
<i>Bolivina</i> aff. <i>dilatata</i>	infralittoral-epibathyal		Spezzaferri & Tamburini, 2007; Goineau et al., 2012; Milker & Schmiedl, 2012
<i>Bolivina italica</i>	infralittoral-circalittoral		Mendes et al., 2012
<i>Bolivina plicatella</i>	circalittoral-epibathyal	circalittoral-upper epibathyal	Milker & Schmiedl, 2012
<i>Bolivina pseudoplicata</i>	infralittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bolivina</i> cf. <i>pseudoplicata</i>	infralittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bolivina punctata</i>	infralittoral-epibathyal	infralittoral-circalittoral	Blanc-Vernet, 1969; Coppa et al., 1994; Fiorini and Vaiani, 2001
<i>Bolivina spathulata</i>	infralittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bolivina variabilis</i>	infralittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993 ; Milker & Schmiedl, 2012
<i>Bolivina</i> sp.	infralittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993 ; Milker & Schmiedl, 2012
<i>Bulimina costata</i>			
<i>Bulimina inflata</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bulimina marginata</i>	infralittoral-epibathyal	lower circalittoral-lower epibathyal	Jorissen, 1988; Sgarrella & Moncharmont Zei, 1993; De Rijk et al., 1999; Frezza et al., 2005, 2010; Frezza & Carboni, 2009
<i>Bulimina</i> cf. <i>marginata</i>	infralittoral-epibathyal	lower circalittoral-lower epibathyal	Jorissen, 1988; Sgarrella & Moncharmont Zei, 1993; De Rijk et al., 1999; Frezza et al., 2005, 2010; Frezza & Carboni, 2009
<i>Bulimina minima</i>		epibathyal	Violanti et al., 2009
<i>Cassidulina carinata</i>	circalittoral-epibathyal	lower circalittoral-lower epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Frezza et al., 2005, 2010; Frezza & Carboni, 2009

<i>Cassidulina crassa</i>	circalittoral-epibathyal	lower circalittoral-lower epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Frezza et al., 2005
<i>Cassidulina laevigata</i>	circalittoral-epibathyal		Milker & Schmiedl, 2012
<i>Cassidulinoides bradyi</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Cibicidoides pachyderma</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Cibicidoides pseudoungerianus</i>	infralittoral-epibathyal	infralittoral-lower circalittoral	Milker & Schmiedl, 2012
<i>Cibicidoides robertsonianus</i>		epibathyal	Violanti et al., 2009
<i>Cibicidoides wüllerstorfi</i>	circalittoral-epibathyal	epibathyal	van Hinsbergen et al., 2005; Russo et al., 2007
<i>Cibicidoides</i> sp.			
<i>Dentalina inflexa</i>	circalittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Dentalina leguminiformis</i>	circalittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Dentalina mucronata</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Dentalina</i> sp.			
<i>Discorbinella bertheloti</i>	infralittoral-epibathyal	circalittoral-epibathyal	Jorissen, 1988; Sgarrella & Moncharmont Zei, 1993
<i>Ehrenbergina bradyi</i>	epibathyal	epibathyal	Di Bella et al., 2015
<i>Epistominella lecalvezi</i>	epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006
<i>Eponides concameratus</i>	infralittoral-epibathyal	circalittoral	Murray, 2006; Sgarrella & Moncharmont Zei, 1993
<i>Eponides repandus</i>	infralittoral-epibathyal	circalittoral	Murray, 2006
<i>Fissurina bradyana</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina marginata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina orbignyana</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina quadricostulata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina sequenziana</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina</i> sp.			
<i>Gavelinopsis lobatulus</i>	infralittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006
<i>Gavelinopsis praegeri</i>	infralittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993; Milker and Schmiedl, 2012; Di Bella, 2010; Murray 2006
<i>Globocassidulina oblonga</i>	circalittoral-epibathyal	lower circalittoral-lower epibathyal	Sgarrella & Moncharmont Zei, 1993; Frezza et al., 2005
<i>Globocassidulina subglobosa</i>	circalittoral-epibathyal	lower epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007; Milker & Schmiedl, 2012
<i>Gyroidina altiformis</i>	circalittoral-epibathyal	lower epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Gyroidina neosoldanii</i>	circalittoral-epibathyal	lower epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Gyroidina umbonata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993

<i>Hanzawaia boueana</i>	infralittoral-epibathyal		Russo et al., 2007
<i>Heterolepa floridana</i>	infralittoral-epibathyal		Ola and Adewale, 2014
<i>Hoeglundina elegans</i>	circalittoral-epibathyal	lower epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Karreriella bradyi</i>	circalittoral-epibathyal	epibathyal	van Hinsbergen et al., 2005; Spezzaferri & Tamburini, 2007
<i>Karreriella chilostoma</i>	epibathyal	epibathyal	van Hinsbergen et al., 2005
<i>Lagena</i> cf. <i>apiopleura</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Lagena gibbera</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Lagena nebulosa</i>	circalittoral-epibathyal	lower epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Lagena striata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Ola and Adewale, 2014
<i>Lagena strumosa</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Lagena</i> sp.			
<i>Lenticulina cultrata</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Lenticulina curvisepta</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006
<i>Lenticulina gibba</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Lenticulina inornata</i>	circalittoral-epibathyal	epibathyal	Spezzaferri & Tamburini, 2007; Ola and Adewale, 2014
<i>Lenticulina orbicularis</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Lenticulina rotulata</i>	circalittoral-epibathyal	epibathyal	Spezzaferri & Tamburini, 2007
<i>Lenticulina stellata</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006; Gebhardt, 1999 (with references)
<i>Lenticulina</i> sp.			
<i>Marginulina glabra</i>	circalittoral-epibathyal	epibathyal	Blanc-Vernet, 1969; Sgarrella & Moncharmont Zei, 1993
<i>Marginulina</i> cf. <i>glabra</i>	circalittoral-epibathyal	epibathyal	Blanc-Vernet, 1969; Sgarrella & Moncharmont Zei, 1993
<i>Marginulina</i> sp.			
<i>Martinottiella communis</i>	circalittoral-epibathyal		Gebhardt, 1999 (with references); Murray, 2006; Spezzaferri & Tamburini, 2007
<i>Melonis barleeanus</i>	circalittoral-epibathyal	circalittoral-epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Melonis pompilioides</i>	circalittoral-epibathyal		Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Neolenticulina peregrina</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Nodosaria</i> sp.			
<i>Oridorsalis umbonatus</i>	epibathyal		Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Pandaglandulina dinapolii</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Parafissurina lateralis</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Planulina ariminensis</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999

			(with references); Spezzaferri & Tamburini, 2007
<i>Planulina cf. ariminensis</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Plectrofrondicularia raricosta</i>	circalittoral-bathyal		Hohenegger, 2005.
<i>Pleurostomella alternans</i>	infralittoral-epibathyal		Spezzaferri & Tamburini, 2007
<i>Polymorphina</i> sp.			
<i>Praeglobobulimina pupoides</i>	epibathyal		Gebhardt, 1999 (with references)
<i>Pullenia bulloides</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Pullenia quinqueloba</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Pullenia salisburyi</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Murray, 2006
<i>Saracenaria italica</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Sigmoilinita tenuis</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri and Tamburini, 2007
<i>Siphonina reticulata</i>	infralittoral-epibathyal		Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Sphaeroidina bulloides</i>	infralittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Stilostomella consobrina</i>	epibathyal		Gebhardt, 1999 (with references)
<i>Stilostomella consobrina</i> var. <i>emaciata</i>	epibathyal		Gebhardt, 1999 (with references)
<i>Stilostomella monilis</i>	circalittoral-epibathyal		Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Textularia</i> sp.			
<i>Uvigerina auberiana</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Uvigerina bradyana</i>	circalittoral-epibathyal		Murray, 2006
<i>Uvigerina canariensis</i>	circalittoral-epibathyal		Murray, 2006
<i>Uvigerina cf. canariensis</i>	circalittoral-epibathyal		Murray, 2006
<i>Uvigerina mediterranea</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Jorissen, 1988; Sgarrella & Moncharmont Zei, 1993; de Stigter et al., 1999; De Rijk et al., 1999, 2000; Fontanier et al., 2000; Frezza et al., 2005, 2010; Frezza and Carboni, 2009
<i>Uvigerina peregrina</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; De Rijk et al., 1999; Altenbach et al., 2003; Fontanier et al., 2002; Spezzaferri & Tamburini, 2007; Frezza & Carboni, 2009
<i>Uvigerina proboscidea</i>	circalittoral-epibathyal		Fontanier et al., 2002
<i>Uvigerina pygmaea</i>	circalittoral-epibathyal		Spezzaferri & Tamburini, 2007
<i>Uvigerina</i> sp.			
<i>Valvulineria bradyana</i>	infralittoral-epibathyal	opportunistic species	Jorissen, 1988; Sgarrella & Moncharmont Zei, 1993; Coppa et al., 1994; Fontanier et al., 2002; Frezza & Carboni, 2009

Table 3: Autochthonous ostracods from Gülnar East section

Ostracod species	Stratigraphic range	Bathymetric range	Literature
<i>Argilloecia acuminata</i> Müller, 1894	Serr.-Rec.	50-2500 m	Colalongo et al., 1981; Malz and Jellinek, 1994; Barra et al., 1998; Aiello et al., 2000; Aiello & Barra, 2001; Sciuto & Rosso, 2008
<i>Argilloecia fatua</i> Barra, Aiello & Bonaduce, 1996	Serr.-Cal.	50-1000 m	Ciampo, 1986; Aiello et al., 2000; Aiello & Barra, 2001
<i>Bairdoppilata profunda</i> Aiello, Barra & Bonaduce, 2000	Langh.-Gel. (MP15)	300-1000 m	Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce and Barra, 2002; Sciuto, 2003; Sciuto et al., 2003; Faranda et al., 2007
<i>Bythocypris producta</i> (Seguenza, 1880)	Langh.-Cal.	500-1000 m	Colalongo et al., 1981; Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce & Barra 2002;
<i>Cytherella robusta</i> Colalongo & Pasini, 1980	Langh.-Cal.	300-1000 m	Colalongo et al., 1990, 1981; Bonaduce & Sprovieri, 1985; Aiello and Barra, 2001; Bonaduce & Barra 2002; Faranda et al., 2007
<i>Cytherella russoi</i> Sissingh, 1972	Langh.-Gel. (MP15)	50-1000 m	Sissingh, 1972; Aiello et al., 2000;
<i>Eucytherura</i> sp.		s	
<i>Henryhowella sarsii profunda</i> Bonaduce, Barra & Aiello, 1999	Aqui.-Rec.	200-1000 m	Bonaduce and Sprovieri, 1985; Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce & Barra, 2002; Faranda et al., 2007; Sciuto, 2003; Sciuto et al., 2003; Sciuto & Rosso, 2008
<i>Krithe iniqua</i> Abbate, Barra, Aiello & Bonaduce, 1993	Langh.-Cal.	50-1000 m	Ciampo, 1986; Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce & Barra, 2002; Sciuto and Rosso, 2008
<i>Oblitacythereis mediterranea</i> Benson, 1977	Zanc. (MP12)-Rec.	250-1000 m	Bonaduce & Sprovieri, 1985; Barra et al., 1998; Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce & Barra, 2002
<i>Paijenborchella iocosa</i> Kingma, 1948	Langh.-Gel. (MP15)	500-1000 m	Bonaduce & Sprovieri, 1985; Aiello & Barra, 2001; Bonaduce & Barra, 2002
<i>Parakrithe ariminensis</i> (Ruggieri, 1967)	Langh.-Piac. (MP15)	100-1000 m	Ruggieri, 1967; Aiello et al., 2000; Aiello & Barra, 2001; Bonaduce & Barra, 2002; Faranda et al., 2007
<i>Parakrithe declivis</i> Ciampo, 1980	Tort.-Zancl. (MP14a)	300-350 m	Faranda et al., 2007
<i>Parakrithe rotundata</i> Aiello, Barra, Abate & Bonaduce, 1993	Langh.-Cal.	500-1000 m	Aiello et al., 2000; Bonaduce and Barra, 2002
<i>Paracytherois</i> sp.			

Table 4: Autochthonous planktonic foraminifers from Gülnar East and Tol sections

Planktonic foraminifer species	Stratigraphic range	Literature
<i>Globigerina bulloides</i>	MMi7-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerina falconensis</i>	Burdigalian-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinella calida</i>	Middle Pleistocene-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinella obesa</i>	Oligocene-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinella pseudobesa</i>	MMi8-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinella siphonifera</i>	MMi12b-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinita glutinata</i>	Oligocene-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinita parkerae</i>	N13-Recent	Blow, 1969
<i>Globigerinita</i> sp.		
<i>Globigerinoides immaturus</i>	Oligocene-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinoides quadrilobatus</i>	MMi2-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinoides ruber</i>	MP14a-Recent	Iaccarino & Premoli Silva, 2007
<i>Globigerinoides trilobus</i>	Burdigalian-Recent	Iaccarino & Premoli Silva, 2007
<i>Globorotalia crassaformis crassaformis</i>	MP14b-MP1e1	Iaccarino & Premoli Silva, 2007
<i>Globorotalia scitula</i>	MMi5b-Recent	Iaccarino & Premoli Silva, 2007
<i>Globoturbototalita rubescens</i>	MP1e1-Recent	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina acostaensis</i>	MMi7c-MP1e1	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina dutertrei</i>	MP15b-Recent	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina humerosa</i>	MMi12b-MP16	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina pachyderma</i> dx	MP16a-Recent	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina pachyderma</i> sx	MP1e1-0.61 Ma	Iaccarino & Premoli Silva, 2007
<i>Neogloboquadrina</i> sp.		
<i>Orbulina bilobata</i>	MMi5a-Recent	Iaccarino & Premoli Silva, 2007
<i>Orbulina suturalis</i>	MMi5a-Recent	Iaccarino & Premoli Silva, 2007
<i>Orbulina universa</i>	MMi5b-Recent	Iaccarino & Premoli Silva, 2007
<i>Tenuitellinata iota</i>	MP16a-Recent	Iaccarino & Premoli Silva, 2007

Table 5: Autochthonous benthic foraminifers from Tol sections

Benthic foraminifer species	Bathymetric range	Optimum bathymetric range	References
<i>Amphicoryna scalaris</i>	infralittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Amphicoryna sublineata</i>		circalittoral	Di Bella et al., 2000-2002
<i>Astacolus crepidulus</i>	infralittoral-epibathyal	epibathyal	Blanc-Vernet, 1969; Sgarrella & Moncharmont Zei, 1993
<i>Bolivina alata</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bolivina punctata</i>	infralittoral-epibathyal	infralittoral-circalittoral	Blanc-Vernet, 1969; Coppa et al., 1994; Fiorini & Vaiani, 2001; Spezzaferri and Tamburini, 2007
<i>Bulimina elongata</i>	circalittoral-bathyal	epibathyal	Hohenegger, 2005
<i>Bulimina inflata</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bulimina sp.</i>	circalittoral-epibathyal	upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Bulimina stirata</i>	circalittoral-bathyal	lower epibathyal	Sgarrella and Moncharmont Zei, 1993; Jorissen 1988
<i>Cancris auriculus</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Casale, 1982
<i>Cassidulina crassa</i>	circalittoral-epibathyal	lower circalittoral-lower epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Frezza et al., 2005; Milker & Schmiedl, 2012
<i>Cassidulinoides bradyi</i>	circalittoral-epibathyal	lower circalittoral-upper epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Cibicides wüllerstorfi</i>	circalittoral-epibathyal	epibathyal	van Hinsbergen, 2005; Russo et al., 2007
<i>Cibicidoides pseudoungerianus</i>	infralittoral-epibathyal	infralittoral-lower circalittoral	Milker & Schmiedl 2012; Murray 2006
<i>Cribrorobulina serpens</i>	circalittoral-epibathyal	lower circalittoral-lower epibathyal	Spezzaferri & Tamburini, 2007
<i>Dentalina aciculata</i>			
<i>Dentalina leguminiformis</i>	circalittoral-epibathyal	circalittoral-epibathyal	Sgarrella & Moncharmont Zei, 1993, Spezzaferri and Tamburini 2007
<i>Dentalina mucronata</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Dentalina sp.</i>			
<i>Fissurina quadricostulata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Fissurina sequenziana</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Gyroidina longispira</i>	upper circalittoral-bathyal		Violanti et al., 2011
<i>Gyroidina neosoldanii</i>	infralittoral-epibathyal	lower epibathyal	Sgarrella & Moncharmont Zei, 1993, Spezzaferri and Tamburini 2007
<i>Hansenica soldanii</i>	upper circalittoral-upper epibathyal	upper epibathyal	Wright, 1978

<i>Hanzawaia boueana</i>	infralittoral-epibathyal		Russo et al., 2000, Villanueva-Guimerans; Canudo, 2008
<i>Lagena elongata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993
<i>Lagena striata</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Ola & Adewale, 2014
<i>Lenticulina calcar</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Lenticulina cultrata</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Lenticulina curvisepta</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006
<i>Lenticulina gibba</i>	infralittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Lenticulina inornata</i>	circalittoral-epibathyal	epibathyal	Spezzaferri & Tamburini, 2007; Ola & Adewale, 2014
<i>Lenticulina orbicularis</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Lenticulina rotulata</i>	circalittoral-bathyal	epibathyal	Spezzaferri & Tamburini, 2007
<i>Lenticulina sp.</i>			
<i>Lenticulina stellata</i>	circalittoral-bathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Murray, 2006; Gebhardt, 1999 (with references)
<i>Marginulina glabra</i>	circalittoral-epibathyal	epibathyal	Blanc-Vernet, 1969; Sgarrella & Moncharmont Zei, 1993
<i>Martinottiella communis</i>	circalittoral-epibathyal		Gebhardt, 1999 (with references); Murray, 2006; Spezzaferri & Tamburini, 2007
<i>Melonis barleanus</i>	circalittoral-epibathyal	circalittoral-epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Melonis pompilioides</i>	circalittoral-epibathyal		Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Oridorsalis umbonatus</i>	epibathyal		Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Planulina ariminensis</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Plectofrondicularia raricosta</i>	circalittoral-epibathyal	epibathyal	Hohenegger, 2005
<i>Plectofrondicularia semicostata</i>			
<i>Pleurostomella alternans</i>	infralittoral-epibathyal		Spezzaferri & Tamburini, 2007
<i>Polymorphina sp.</i>			
<i>Praeglobobulimina pupoides</i>	epibathyal		Gebhardt, 1999 (with references)
<i>Pullenia bulloides</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Pullenia quinqueloba</i>	circalittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Pullenia salisburyi</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993, Murray 2006
<i>Pyrgo oblonga</i>	infralittoral-bathyal		Casale, 1982

<i>Saracenaria italica</i>	infralittoral-epibathyal	epibathyal	Sgarrella & Moncharmont Zei, 1993
<i>Sigmoilinita tenius</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007
<i>Siphonina reticulata</i>	infralittoral-epibathyal		Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007, Violanti et al., 2011
<i>Sphaeroidina bulloides</i>	infralittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993; Spezzaferri & Tamburini, 2007,
<i>Spiroloculina depressa</i>	infralittoral-upper epibathyal		Casale, 1982; Murray, 2006
<i>Spiroplectammia wrighti</i>	upper epibathyal	upper epibathyal	del Carmen Sánchez Ariza, 1988
<i>Stilostomella monilis</i>	circalittoral-epibathyal		Gebhardt, 1999 (with references); Spezzaferri & Tamburini, 2007
<i>Textularia agglutinans</i>	circalittoral	upper circalittoral	Sgarrella & Moncharmont Zei, 1993
<i>Uvigerina auberiana</i>	circalittoral-epibathyal	epibathyal	Parker, 1958; Sgarrella & Moncharmont Zei, 1993
<i>Uvigerina bradyana</i>	circalittoral-epibathyal		Murray, 2006
<i>Uvigerina canariensis</i>	circalittoral-epibathyal		Murray, 2006
<i>Uvigerina longistriata</i>	circalittoral-epibathyal		Murray 2006
<i>Uvigerina peregrina</i>	circalittoral-epibathyal		Sgarrella & Moncharmont Zei, 1993; De Rijk et al., 1999; Altenbach et al., 2003; Fontanier et al., 2002; Spezzaferri & Tamburini, 2007; Frezza & Carboni, 2009
<i>Uvigerina proboscidea</i>	circalittoral-epibathyal		Fontanier et al., 2002
<i>Uvigerina pygmaea</i>	circalittoral-epibathyal		Spezzaferri & Tamburini, 2007