

ROMA TRE UNIVERSITY

**THE MIDDLE PLEISTOCENE UPLIFT OF THE CENTRAL ANATOLIAN
PLATEAU (CAP) SOUTHERN MARGIN: NEW EVIDENCE FROM MUT AND
ANTALYA BASINS (SOUTH TURKEY)**

Ph.D. Thesis

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**IL SOLLEVAMENTO NEL PLEISTOCENE MEDIO DEL MARGINE
MERIDIONALE DELL'ALTOPIANO ANATOLICO CENTRALE (CAP):
NUOVE EVIDENZE DAI BACINI DI MUT E ANTALYA (TURCHIA
MERIDIONALE)**

Tesi di Dottorato di Ricerca

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As the first doctoral holder in my family, I dedicate this success to my mother and sisters.

"Woman, Life, Freedom"

TABLE OF CONTENTS

	<u>Page</u>
PREFACE.....	1
SUMMARY.....	3
INTRODUCTION	5
1. THE WORLD’S HIGHEST LATE MIDDLE PLEISTOCENE MARINE DEPOSITS: INSIGHTS FROM THE SOUTHERN MARGIN OF THE CENTRAL ANATOLIAN PLATEAU (SOUTHERN TURKEY).....	9
ABSTRACT.....	9
1.1 INTRODUCTION.....	9
1.2 METHODS.....	10
1.3 RESULTS.....	12
1.4 DISCUSSION.....	14
1.4.1 Biostratigraphic and paleoenvironmental signals	14
1.4.2 Implications for the uplift history of the CAP southern margin	15
1.4.3 Elevation of MIS 7 marine deposits in the world.....	15
1.5 CONCLUSIONS	18
1.6 SUPPLEMENTAL MATERIAL	18
1.6.1 Refined Geological Mapping	18
1.6.2 Calcareous nannofossils	18
1.6.3 Ostracods	24
2. TECTONO-SEDIMENTARY EVOLUTION AND UPLIFT OF THE ANTALYA BASIN DURING THE LATE MIDDLE PLEISTOCENE: INSIGHTS FROM THE MANAVGAT FAN-DELTA COMPLEX.....	38
ABSTRACT.....	38
2.1 INTRODUCTION.....	38
2.1.1 Geological setting and stratigraphy	41
2.2 MATERIAL AND METHODS	45
2.3 RESULTS.....	47
2.3.1 Lithostratigraphy of the MNV composite section	47
2.3.1.1. MNV4 section	47
2.3.1.2. MNV1 and MNV2 sections.....	47
2.3.2 Micropaleontological results	52
2.3.2.1 Calcareous nannofossils.....	52
2.3.2.2 Foraminifera	54
2.4 DISCUSSION.....	57
2.4.1 Depositional environments and facies evolution	57
2.4.1.1. Lower Interval (0–24 m): proximal upper delta-front.....	57
2.4.1.2. Middle Interval (24–111 m): distal lower delta-front	59
2.4.1.3. Upper Interval (111–125 m): upper delta-front.....	60
2.4.1.4 Depositional Context.....	60
2.4.2 Micropalaeontological constraints of the Manavgat Sub-basin	61
2.4.3 Impact on the chronostratigraphy of the Antalya Basin.....	63
2.4.4 Tectonic controls on the basin evolution.....	55
2.5 CONCLUSIONS	68

3. QUATERNARY DIFFERENTIAL UPLIFT IN MUT AND ANTALYA BASINS (SOUTHERN TURKEY): A RESPONSE TO DEEP-SEATED DYNAMIC PROCESSES.....	70
ABSTRACT	70
3.1 INTRODUCTION	70
3.2 METHODS.....	73
3.2.1. Calcareous nannofossil analyses	73
3.2.2. Foraminifera and ostracods	73
3.2.3. Geomorphological modelling	74
3.3 RESULTS.....	75
3.3.1 Late Middle Pleistocene marine deposits from Mut and Antalya basins	75
3.3.2 Integration with geomorphological analyses	78
3.3.3 Compiled uplift history along the southern CAP margin.....	79
3.4 DISCUSSION.....	80
3.4.1 Differential uplift along the CAP southern margin and the Isparta Angle....	80
3.4.2 Integration with geomorphological analyses.....	81
3.4.3 Geodynamic Implications.....	82
4. GENERAL CONCLUSIONS.....	86
5. REFERENCES.....	88

PREFACE

This thesis is the outcome of a three-year doctoral project titled “The Middle Pleistocene uplift of the Central Anatolian Plateau (CAP) southern margin: New evidence from Mut and Antalya basins (South Turkey)”, carried out under the supervision of Prof. Paola Cipollari and Prof. Domenico Cosentino. I am deeply grateful to my supervisors for their support and trust, which greatly improved the quality of my work and manuscripts.

The project greatly benefited from collaborations with Prof. Elsa Gliozzi, Dr. Simone Racano, Prof. Claudio Faccenna, Prof. Fabrizio Lirer, Prof. Süheyla Kanbur, Prof. Massimo Mattei, Prof. Francesca Cifelli, Dr. Nazik Öğretmen, and Costanza Faranda. I would also like to express my sincere thanks to Prof. Paolo Ballato for his constant encouragement throughout my PhD.

This research investigates the Mut and Antalya Basins of southwestern Turkey to reconstruct the tectonic evolution based on Middle Pleistocene marine deposits. The Tauride orogenic belt, a major segment of the Alpine–Himalayan system, preserves a long-lived record of Paleozoic–Cenozoic subduction, collision, and post-orogenic deformation. In the Central Taurides, post-orogenic extension and subsequent marine deposition within the intramontane Mut–Ermenek region document multiphase uplift that continued into the Pleistocene. Similarly, in the Isparta Angle (IA), where the western and central Tauride domains converge, ongoing Neogene–Quaternary deformation allowed marine sedimentation in the Antalya Basin to persist into the Pleistocene.

In this thesis, I present new biostratigraphic, paleomagnetic, and paleobathymetric data from the TOL-1 section (1177 m a.s.l.) in the Mut Basin and the Manavgat (MNV) fan-delta succession (143 m a.s.l.) in the Antalya Basin. These data refine the timing and magnitude of Middle Pleistocene uplift along the CAP southern margin and the IA, quantify the differential uplift between these regions, and evaluate the geodynamic processes responsible for this rapid surface response.

This thesis comprises three manuscripts, each representing a draft intended for peer-reviewed publication. Manuscripts 1 and 2 are planned for submission within the current month, while Manuscript 3 remains in preparation. It follows a short summary of each manuscript.

Manuscript 1 – The world’s highest late Middle Pleistocene marine deposits: insights from the southern margin of the Central Anatolian Plateau (southern Turkey)

The first manuscript investigates a well-constrained marine sedimentary succession documenting a shallowing-upward sequence of marls, clayey marls, and calcarenite intercalations deposited during the late Middle Pleistocene (ca. 0.53–0.23 Ma). We applied qualitative calcareous nannofossil and planktonic foraminiferal biostratigraphy, while paleobathymetry was reconstructed using ostracod taphonomy. By integrating biostratigraphy, paleobathymetry (~50 m b.s.l.) and sea-level estimates at 0.23 Ma (~–18 m), we calculate an uplift rate of ~5.2 m/kyr for the southern margin of the CAP. Our results document the highest known late Middle Pleistocene marine deposits globally and provide new insights into the geodynamic evolution of the region.

Manuscript 2 – Tectono-sedimentary evolution and uplift of the Antalya Basin (southern Turkey) during the late Middle Pleistocene: insights from the Manavgat fan-delta complex

The second manuscript provides a revised stratigraphic and chronostratigraphic framework for the Manavgat Sub-basin. We investigate a well-constrained fan-delta succession records deposition within a deltaic system comprising upper delta-front, lower delta-front, and renewed upper delta-front facies, reflecting alternating high- and low-energy shallow-marine conditions deposited during the late Middle Pleistocene (<0.26–0.23 Ma). The stratigraphy of the Antalya Basin has long been debated, as several formations have been assigned conflicting ages and depositional interpretations. Calcareous nannofossil and planktonic foraminiferal assemblages indicate that marine deposition persisted until ~0.23 Ma, with uplift rates of ~0.8 m/kyr. This record offers new constraints on the timing and magnitude of Middle Pleistocene uplift and refines our understanding of the tectono-sedimentary evolution of both the Antalya Basin and the broader Isparta Angle.

Manuscript 3 – Quaternary differential uplift in Mut and Antalya basins (southern Turkey): a response to deep-seated dynamic processes

The third manuscript synthesizes the differential uplift patterns recorded in the Mut Basin and the Manavgat Sub-basin. Although these sections were deposited synchronously, they record markedly different uplift rates, 5.2 m/kyr and ~0.8 m/kyr, respectively, indicating strongly divergent mechanisms driving vertical deformation. By integrating these findings with marine-terrace modelling and river-profile inversions, we refine the late Quaternary uplift evolution of the Central Anatolian Plateau (CAP) southern margin. The synthesis identifies two discrete slab break-off events beneath Cyprus, one in the Late Miocene and another in the Middle Pleistocene, and demonstrates how these events generated contrasting uplift patterns across southern Anatolia.

SUMMARY

This thesis investigates the Mut and Antalya basins in southern Turkey to understand how deep Earth processes, particularly slab segmentation, slab break-off, and associated mantle upwelling, shape surface uplift across Anatolia. These basins preserve the highest late Middle Pleistocene marine deposits known worldwide, offering a rare natural laboratory for linking mantle-scale dynamics to rapid surface deformation.

The Mut Basin, located along the Central Anatolian Plateau (CAP) southern margin, contains a marine succession now uplifted to 1177 m a.s.l. Foraminiferal biostratigraphy, marked by the consistent presence of *Globigerinella calida* from the base upward, indicates that deposition began after ~0.53 Ma. The calcareous nannofossil biostratigraphic markers, specifically the last occurrence of *Pseudoemiliana lacunosa* (0.46 Ma) and the first occurrence of *Emiliana huxleyi* (0.26 Ma), combined with, paleomagnetic data confirm a late Middle Pleistocene age. Ostracod taphonomy indicates a shallowing marine environment from upper epibathyal to littoral conditions. Integrating paleobathymetry (~50 m b.s.l.) with global sea-level estimates at 0.23 Ma (~-18 m) yields an exceptional uplift rate of ~5.2 m/kyr, among the highest ever documented for this time period. This rapid uplift likely reflects asthenospheric upwelling and lithospheric rebound following slab break-off beneath Cyprus.

In contrast, the Antalya Basin, situated within the tectonically complex Isparta Angle (IA) where the Aegean and Cyprus subduction systems converge, records a more moderate uplift history. New lithostratigraphic and micropalaeontological analyses from the Manavgat Sub-basin (MNV) composite succession show that deposition continued until ~0.23 Ma, significantly revising previously assigned Miocene–Early Pleistocene ages. The deltaic succession records shifting upper and lower delta-front environments controlled by tectonics, sediment supply, and sea-level fluctuations. The calcareous nannofossil biostratigraphy constrains the onset of deposition to after ~0.26 Ma, based on the presence of *E. huxleyi*, while the termination coincides with late Middle Pleistocene marine strata (TOL-1) in the Mut Basin. These constraints imply an average uplift rate of ~0.8 m/kyr, reflecting the combined influence of slab retreat and thermal–viscous relaxation of the upper plate.

This study presents new integrated stratigraphic and geomorphological constraints from the Mut Basin to reconstruct the timing, magnitude, and geodynamic drivers of Middle Pleistocene uplift in southern Anatolia. Geomorphological analyses further clarify these patterns. Marine-

terrace modelling, incorporating the new TOL-1 age and elevation, links seven terraces (30–1400 m a.s.l.) to major sea-level highstands and identifies a sharp uplift pulse peaking during MIS 5 (~120 ka) with rates of ~6.2 m/kyr, followed by slower Holocene uplift. River-profile inversions from seventeen catchments, independently confirm the timing and spatial focus of this uplift, showing coherent kilometre-scale growth of the southern Central Anatolian Plateau margin. When placed within the long-term vertical-motion history, two major subsidence phases (respectively at ~17.0 and 6.6 Ma) and two uplift pulses (~5.3 and 0.4 Ma) emerge, with the new Middle Pleistocene record documenting the fastest uplift rates yet identified. These contrasting uplift magnitudes between Mut and Antalya reflect fundamentally different deep-mantle forcing. The Mut Basin sits above an inferred Middle Pleistocene shallow slab break-off, which triggered asthenospheric upwelling and rapid lithospheric rebound, producing >1.2 km of uplift since ~0.40 Ma. Conversely, the Antalya Basin, positioned above a still-attached and retreating Antalya slab, experienced only modest uplift, driven by upper-plate processes such as influence of slab retreat, and crustal relaxation. Together, these results demonstrate that repeated slab segmentation beneath the Cyprus subduction system, first during the late Messinian (~5.4 Ma) and again in the Middle Pleistocene (~0.40 Ma), governed the contrasting uplift histories of the Mut and Antalya basins. The sharp differential uplift of coeval deposits, highlights the strong influence of deep mantle dynamics on rapid surface deformation in southern Anatolia.

INTRODUCTION

Orogenic plateaus such as Tibet, the Altiplano–Puna, and Anatolia offer exceptional natural laboratories for investigating how lithosphere–mantle interactions shape Earth’s surface (Garzione et al., 2017; Göğüş et al., 2017; Zhang et al., 2021). The Central Anatolian Plateau (CAP) forms part of a complex neotectonic system comprising the East Anatolian Contractual Province, the West Anatolian Extensional Province, and the CAP, which together accommodate the westward extrusion of Anatolia driven by Arabia–Eurasia collision and African slab rollback since the Miocene (McKenzie, 1970, 1978; Şengör et al., 1985; Reilinger et al., 2006). Plateau evolution has been strongly influenced by both crustal-scale faulting (Şengör and Yılmaz, 1981; Yıldırım et al., 2011, 2013) and deep lithospheric-scale processes including slab tearing, delamination, break-off, and mantle flow reorganization (Faccenna et al., 2006; Biryol et al., 2011; Cosentino et al., 2012; Schildgen et al., 2012; Göğüş et al., 2017; Fernández-Blanco et al., 2020; Andersen et al., 2024; Aykut et al., 2023, 2025). Understanding the timing and mechanisms of CAP uplift is essential for reconstructing interactions between deep geodynamic drivers, crustal deformation, sediment supply, and climate, yet remains challenging due to the sparse chronostratigraphic control and the partial erosion of syn-tectonic sedimentary records.

Sedimentary basins along the southern margin of the CAP and the adjacent Isparta Angle (IA), including the Adana, Antalya, and Mut–Ermenek basins, preserve critical markers of surface uplift (Cosentino et al., 2012; Schildgen et al., 2012; Öğretmen et al., 2018; Racano et al., 2020; Aykut et al., 2023). Uplift rates across Anatolia vary substantially: the Central Pontides record 0.2–0.3 m/kyr over the last ~400 ka (Yıldırım et al., 2011), the plateau interior shows incision rates of 0.05 m/kyr since 1.9 Ma (Çiner et al., 2015) and volcanic sequences indicate 0.12 m/kyr from 5 to 2.5 Ma decreasing to 0.04 m/kyr thereafter (Aydar et al., 2013). By contrast, the Central Taurides (CT) reach 2–3.5 km elevation and are unconformably overlain by Neogene marine deposits now found above ~1.5 km, reflecting multiphase uplift (Cosentino et al., 2012; Cipollari et al., 2013; Racano et al., 2020). In the southeastern CT, uplift in the Adana Basin may have begun in the Late Messinian (5.45 Ma) based on the correlation of 34 seismic reflection profiles of the Handere Formation which consists mainly of fluvial conglomerates and marls (Radeff et al., 2016, 2017).

A major revision to the regional uplift chronology came from [Öğretmen et al. \(2018\)](#), who demonstrated that the youngest marine deposits in the Mut–Ermenek Basin are not Miocene, in age, as commonly assumed, but Early to Middle Pleistocene. The Güle section (983 m a.s.l.) was assigned an Early Pleistocene age, while the Tol section (1087 m a.s.l.) a Middle Pleistocene one (~460 ka). Together with their present elevations (~1500–1600 m a.s.l.), these markers imply rapid uplift rates of ~3.2–3.4 m/kyr, with geomorphological evidence for even faster pulses of 5.2–6.2 m/kyr over the last 0.4–0.6 Ma ([Racano et al., 2020, 2021](#)), and ongoing uplift of 0.8–1.5 m/kyr during the late Holocene ([Liberatore et al., 2022](#)). Building on this framework, the present study incorporates a nearby reference section, the Tol-1 succession (1177 m a.s.l.), to refine the Middle Pleistocene uplift history of the CAP southern margin.

To the southwest, the Antalya Basin occupies the structurally complex Isparta Angle at the transition between the Central and Western Taurides. Its Neogene–Quaternary evolution reflects alternating extension and contraction linked to the Hellenic–Cyprus subduction systems ([Glover and Robertson, 1998](#); [Hall et al., 2014](#); [Aykut et al., 2023](#)). Traditionally, marine successions across the Aksu, Köprüçay, and Manavgat sub-basins were assigned Miocene ages ([Akay et al., 1985](#); [Flecker et al., 1998](#)). Recent biostratigraphic work has significantly revised this view. [Kanbur and Öğretmen \(2022\)](#) used planktonic foraminiferal assemblages to show that the Eskiköy Formation in the Aksu Sub-basin is Early Pleistocene (1.54–1.36 Ma) rather than Pliocene, demonstrating that marine conditions in the IA persisted far longer than previously recognized. This revision highlights major gaps in the regional chronostratigraphic framework and underscores the need for high-resolution, multi-proxy dating across neighboring basins.

Within this broader context, the Manavgat Sub-basin occupies the eastern limb of the IA and unconformably overlies the Tauride basement. Its >1-km-thick Neogene–Quaternary sedimentary infill consists of marine clastics and carbonates, beginning with the Burdigalian Tepekli Formation ([Karabıykoğlu et al., 2000](#)) and continuing through the Oymapınar Limestone, Çakallar Breccia, Geceleme Formation, and the thick Serravalian–Tortonian Karpuzçay Formation. While traditionally considered Miocene, parts of the Karpuzçay and overlying units have now been reinterpreted as Plio–Pleistocene based on nannofossil biostratigraphy ([Sagular and Özer, 2024](#)). Nevertheless, the chronostratigraphy of the Manavgat Sub-basin remains poorly constrained due to reworking, localized hiatuses, and limited high-resolution micropalaeontological control. This uncertainty persists despite recent efforts to refine the regional stratigraphy, including the reassignment of the Manavgat Formation by [Sagular and Özer \(2024\)](#), who demonstrated that its upper units comprise Plio–Pleistocene shallow-marine to deltaic deposits rather than exclusively Miocene successions. To

address these gaps, the present study focuses on the MNV fan-delta succession (143 m a.s.l.) within the Manavgat Formation, with the aim of establishing a robust biostratigraphic age framework and demonstrating that shallow-marine conditions persisted in the eastern IA into the Middle Pleistocene.

Together, the Mut (CAP southern margin) and Manavgat (IA) basins preserve coeval Middle Pleistocene marine deposits that now differ in elevation by more than 1 km—>1177 m a.s.l. at Mut versus ~143 m a.s.l. at Manavgat. This striking contrast provides a unique opportunity to examine how upper-crustal deformation and deep-mantle processes interact across two tectonically distinct sectors of southern Anatolia. By integrating new chronostratigraphic, sedimentological, and geomorphological datasets, this study reconstructs the timing and magnitude of Middle Pleistocene uplift and evaluates whether the region's rapid vertical motions reflect processes typical of orogenic plateaus or instead result from unique slab-fragmentation dynamics beneath the Cyprus–Antalya subduction system.

Despite extensive regional research, several key questions remain unresolved:

- Chronostratigraphic uncertainty persists for the youngest marine deposits in both the Mut and Antalya basins due to inconsistent micropaleontological and lithostratigraphic interpretations.
- The timing and magnitude of Middle Pleistocene uplift along the CAP southern margin have not been constrained using integrated, high-resolution, multi-proxy datasets.
- The relationship between CAP southern margin uplift and IA deformation remains unclear, particularly why coeval Middle Pleistocene marine deposits now differ by >1 km in elevation.
- It is unresolved whether uplift resulted solely from Miocene slab break-off or whether a younger, shallow Middle Pleistocene slab-detachment event also contributed.
- A coherent tectono-sedimentary correlation across the Mut and Antalya basins is lacking due to conflicting stratigraphic frameworks.

The overarching aim of this thesis is to reconstruct the Middle Pleistocene uplift history of southern Anatolia by integrating records from the Mut (CAP southern margin) and Antalya (IA) basins. This aim is addressed through the following objectives:

- Establishing high-resolution chronologies for the TOL-1 (Mut Basin) and MNV (Manavgat Sub-basin) successions using calcareous nannofossil and foraminiferal biostratigraphy, supported by paleomagnetic data.

- Reconstructing depositional environments through detailed facies analysis and ostracod-based paleobathymetry.
- Quantifying uplift magnitudes and rates using deposit elevations, paleobathymetric reconstructions, sea-level estimates, and geomorphological modelling of terraces and river profiles.
- Refining the Middle Pleistocene tectono-sedimentary evolution of the Mut and Antalya basins.
- Developing a geodynamic model that links uplift pulses to slab segmentation, break-off beneath the Cyprus–Antalya subduction system.

1. THE WORLD'S HIGHEST LATE MIDDLE PLEISTOCENE MARINE DEPOSITS: INSIGHTS FROM THE SOUTHERN MARGIN OF THE CENTRAL ANATOLIAN PLATEAU (SOUTHERN TURKEY)

ABSTRACT

The Mut Basin in southern Turkey hosts the world's highest late Middle Pleistocene marine deposits, offering new constraints on the geodynamic evolution of the southern margin of the Central Anatolian Plateau (CAP). At an elevation of 1177 m a.s.l., the TOL-1 section preserves two key calcareous nannofossil bioevents: the last occurrence of *Pseudoemiliana lacunosa* (0.46 Ma) and the first occurrence of *Emiliana huxleyi* (0.26 Ma). Foraminiferal biostratigraphy, marked by the consistent presence of *Globigerinella calida* from the base upward, indicates that deposition began after ~0.53 Ma. Ostracod taphonomy reveals a transition from upper epibathyal to littoral environments, with a shallowing-upward trend beginning around 0.40 Ma. Paleomagnetic data indicates normal polarity, consistent with deposition during the Brunhes Chron. By integrating biostratigraphy, paleobathymetric reconstructions (~50 m b.s.l.), and sea-level estimates at 0.23 Ma (~ -18 m), a mean uplift rate of ~5.2 m/kyr is calculated. This exceptionally high rate is attributed to mantle upwelling following slab break-off of the still subducting Cyprus slab. These findings provide a critical temporal benchmark for modeling the interplay between deep Earth processes and surface uplift at the western margin of the Africa-Eurasia collision zone.

1.1. INTRODUCTION

Orogenic plateaus such as Tibet, the Altiplano-Puna, and Anatolia provide natural laboratories for investigating how deep Earth processes shape surface evolution (e.g., [Faccenna et al., 2006](#); [Garzione et al., 2017](#); [Göğüş et al., 2017](#); [Zhang et al., 2021](#)). Situated within the Africa–Eurasia convergence zone, the CAP exemplifies a tectonic setting where uplift reflects complex mantle-driven processes including slab segmentation, break-off, lithospheric dripping, and delamination ([Cosentino et al., 2012](#); [Schildgen et al., 2012, 2014](#); [Andersen et al., 2024](#)).

The CAP interior is a low-relief surface surrounded by high-relief margins, most prominently the Central Taurides (CT) to the south. In this region, post-orogenic extension followed by marine sedimentation (preserved in the Antalya, Mut, and Adana basins) records multiple phases of uplift ([Cosentino et al., 2012](#); [Schildgen et al., 2012, 2014](#); [Cipollari et al., 2013](#);

Öğretmen et al., 2018). Notably, marine sedimentation persisted in several sub-basins, and recent micropaleontological studies demonstrate that the youngest marine units in the Aksu (Kanbur and Öğretmen, 2022), Manavgat (Sagular and Özer, 2024), and Mut (Yıldız, 2003) basins extend into the Quaternary. These young marine deposits are crucial for quantifying surface uplift and for constraining the timing and mechanisms of mountain range evolution.

In the Mut Basin, high-elevation marine deposits have been recently reinterpreted as Lower to Middle Pleistocene (Öğretmen et al., 2018). They established the Gülnar East section (983 m a.s.l.) as Lower Pleistocene (from 1.73 Ma to 1.10 Ma) and the Tol section (1087 m a.s.l.) as Middle Pleistocene (from ~0.61 Ma to ~0.46 Ma), implying mean uplift rates of 3.2–3.4 m/kyr. These findings highlight rapid Quaternary surface uplift of the CAP southern margin, which continues today at rates of 0.9–1.5 m/kyr as constrained by Late Holocene sea-level markers (Liberatore et al., 2022).

In this study, we present new biostratigraphic, paleomagnetic, and paleobathymetric data from the TOL-1 section (1177 m a.s.l., top of the section) in the Mut Basin, which points to a rejuvenation of the youngest marine deposits capping the CAP southern margin to the late Middle Pleistocene [Marine Isotopic Stage (MIS) 7; ~0.24–0.19 Ma; Lisiecki and Raymo, 2005]. MIS 7 deposits are preserved globally across diverse tectonic settings, providing benchmarks to evaluate regional uplift rates.

Our objectives are to (i) refine the timing and magnitude of the Middle Pleistocene uplift along the CAP southern margin, (ii) place the CAP's MIS 7 marine record into a global framework by comparing it with coeval stratigraphic sections from other tectonic settings, and (iii) hypothesize possible geodynamic processes responsible for this rapid surface response.

1.2. METHODS

The TOL-1 section (36.385225°N, 33.424114°E; 1177 m a.s.l.) is in the Mut Basin of southwestern Turkey (Fig. 1). This approximately 71-m-thick succession comprises a shallowing-upward marine sequence of marls, clayey marls, and interbedded calcarenite (Fig. 2). We collected 68 bulk samples for micropaleontological, geochemical, and paleomagnetic analyses. The results on ostracods, stable isotopes, and paleomagnetic analyses are shown in the Supplemental Material. Calcareous nannofossils were analyzed quantitatively using smear slides on a per-unit-area basis (Backman and Shackleton, 1983). To improve the detection of *E. huxleyi*, the uppermost 23 samples (labelled TOL-1A 1-23) were processed using concentration techniques (De Capoa et al., 2004). The results obtained were classified according to the biostratigraphic scheme of Di Stefano et al. (2023). Planktonic foraminifera were studied qualitatively for biostratigraphic support, following the biostratigraphic zonation in Lirer et al.

(2019). Ostracods were analyzed quantitatively to provide paleobathymetric constraints. Stable isotope analyses ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) were performed on planktonic foraminifera *Trilobatus trilobus*. Between 10 and 15 specimens of *T. trilobus* were picked from 30 samples, and the analyses were performed using a Thermo Delta V Advantage mass spectrometer at the Geochemical Laboratory of Earth Science Department of Sapienza University of Rome. Calibration to Vienna Pee Dee Belemnite (VPDB) was carried out following NBS-19 standards (Coplen, 1996). The results were compared with reference *Globigerinoides ruber* stable isotope data from ODP Site 967 (Grant et al., 2022). A total of 67 oriented samples were collected for magnetostratigraphy. Alternating field (AF) demagnetization (up to 100mT) was performed with a 2G SQUID cryogenic magnetometer at the National Institute of Geophysics and Volcanology (INGV, Rome) to isolate the characteristic remanent magnetization. Chronological and stratigraphic data were integrated to refine the geological map of the study area. Sentinel-2 satellite imagery was processed using principal component analysis in ENVI software to discriminate the Quaternary lithologies and to update the boundaries between Lower and Middle Pleistocene deposits (Fig. S1).

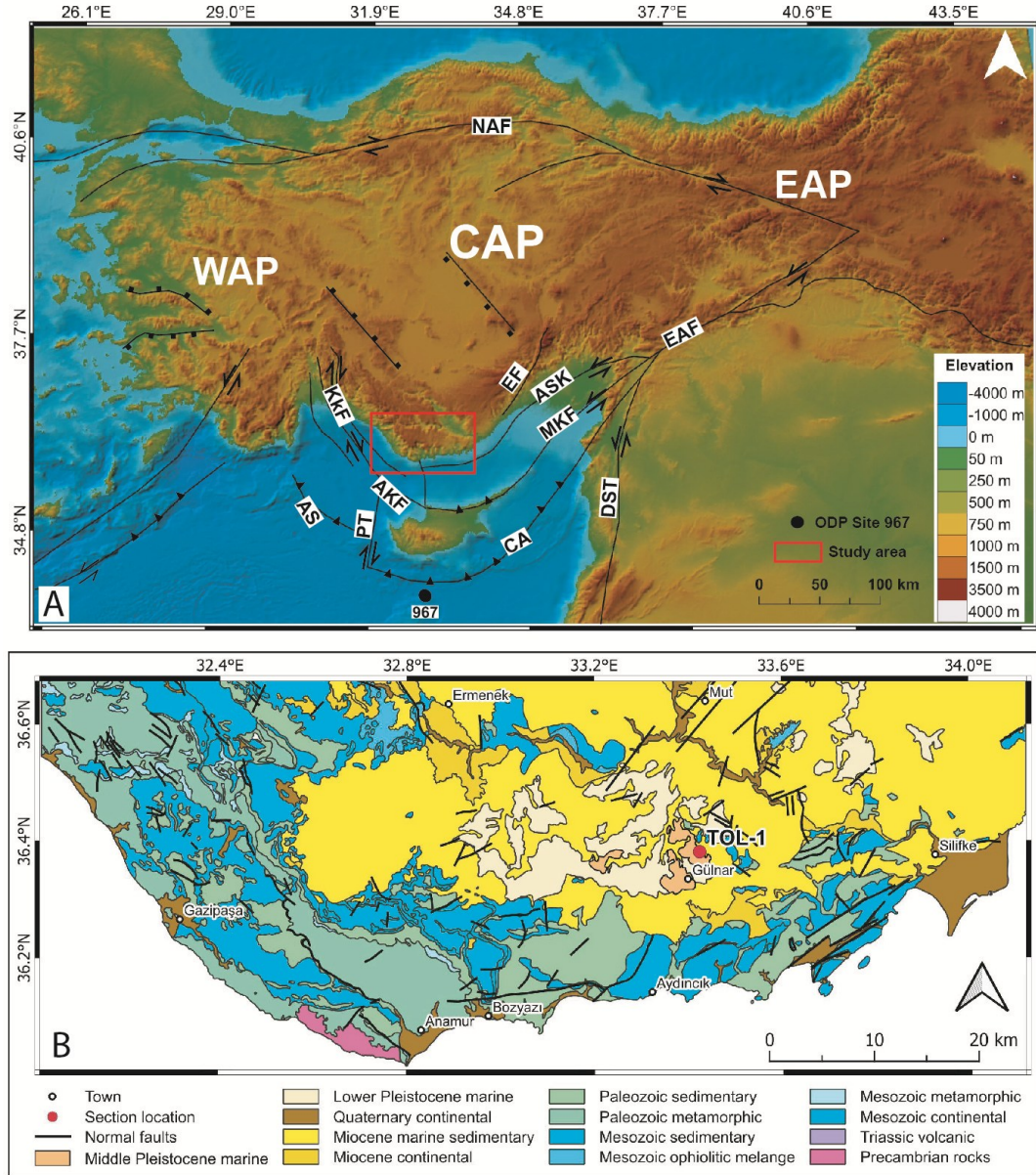


Figure 1. (A) Simplified regional tectonic map of Anatolia showing the location of ODP Site 967. Eastern Anatolian Plateau (EAP); Western Anatolian Province (WAP); North Anatolian Fault (NAF); East Anatolian Fault (EAF); Death Sea Transform Fault (DST); Cyprus Arc (CA); Paphos Transform (PT); Antalya Slab (AS); Aksu-Kyrenia Fault (AKF); Kırkkavak Fault (KkF); Ecemiş Fault (EF); Anamur-Silifke-Kozan Fault (ASK); Misis-Kyrenia Fault (MKF); (B) Geological map of the southern margin of the CAP showing the location of the TOL-1 section (modified after [Racano et al., 2021](#)).

1.3. RESULTS

Calcareous nannofossil assemblages are dominated by long-range and reworked taxa (see Table S1), whereas marker species (*Emiliania huxleyi*, *Gephyrocapsa* spp., and *Pseudoemiliania lacunosa*) are scarce (Figs. 2A and S2). Long-range forms occur consistently (Fig. S3), while reworked taxa comprise ~30% on average, locally reaching ~60% of the total assemblage. *Gephyrocapsa* spp., and *E. huxleyi* rarely exceed 1–2% (Fig. 2B). Two bioevents were recognized: the last occurrence (LO) of *P. lacunosa* (dated at 0.46 Ma) and the first occurrence

(FO) of *E. huxleyi* (dated at 0.26 Ma). Isolated *P. lacunosa* specimens above its LO are interpreted as reworked.

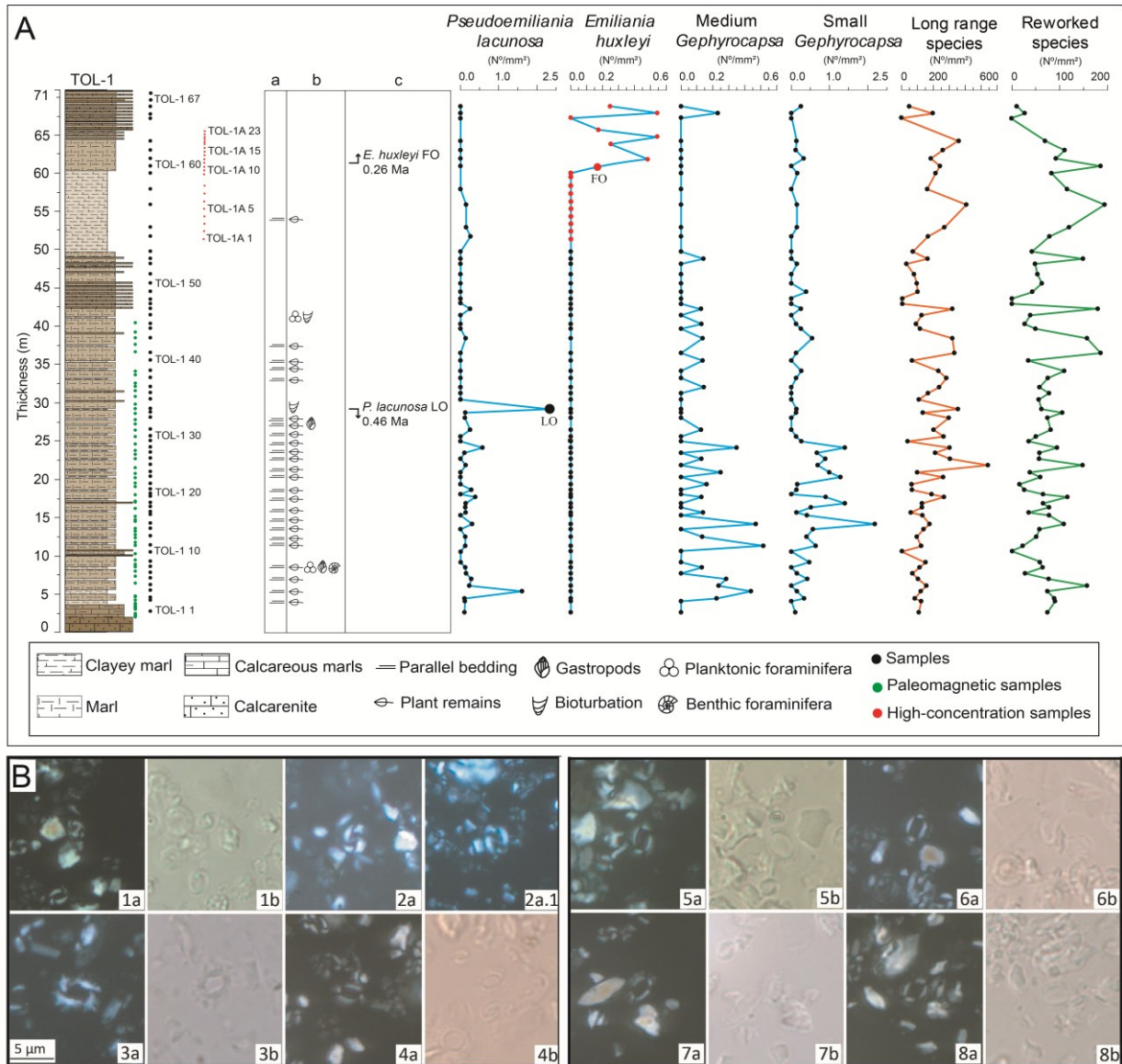


Figure 2. (A) Stratigraphy of the TOL-1 section showing the abundance of calcareous nannofossil marker species. Blue curves represent index species; orange curves indicate long-range taxa; green curves show reworked species. (a) Sedimentary structures; (b) Fossil components; (c) Biostratigraphic events. (B) Microphotographs of calcareous nannofossils from the TOL-1 section. (1) small *Gephyrocapsa* (TOL-1A 12); (2a, 2a.1) medium *Gephyrocapsa* (TOL-1 21, TOL-1 28); (3) *P. lacunosa* (TOL-1 6); (4–8) *E. huxleyi* (TOL-1A 21, TOL-1A 14, TOL-1 65). (a) Crossed nicols; (b) Parallel nicols. Scale bar is uniform across all images.

The planktonic foraminifera in the TOL-1 section are generally poorly to moderately preserved and moderately diversified, resulting in several indeterminate specimens. The assemblage includes *Globigerinodes quadrilobatus*, *G. obliquus obliquus*, *T. trilobus*, *Globigerinella calida*, *Globoturborotalita rubescens*, *Globigerinita glutinata*, *Neogloboquadrina pachyderma*, *N. incompta*, *N. dutertrei*, *Globigerinoides ruber* (white), *G. tenellus*, *Orbulina universa* and *Globigerina bulloides* (Table S2; Fig. S4). The consistent occurrence of *G. calida* from the

base of the section upward constrains deposition to younger than ~0.53 Ma (FO of *G. calida*). Although this age would typically coincide with the common presence of *Globorotalia truncatulinoides* and frequent *N. pachyderma*, *G. truncatulinoides* is absent and *N. pachyderma* occurs only sporadically within the studied interval.

Forty-nine autochthonous ostracod species have been detected. They are included in three different groups: littoral, epibathyal/bathyal, and eurybathic species (Tabs. S3 and S4; Figs. S5 and S6). The paleoenvironmental analysis of ostracods (Tab. S5, Figs. S7 and S8) indicates that the deposition of the TOL-1 section started in a littoral, vegetated environment that deepened along the section down to the upper epibathyal environment and became shallow littoral again (around 50 m of depth) at the top of the section.

Concerning the planktonic foraminiferal stable oxygen isotope data of the TOL-1, $\delta^{18}\text{O}$ values range from -2.23‰ to -0.96‰ , consistent with MIS 13 (Fig. S9; Table S6).

According to the provided biostratigraphic age constrains, different sedimentation rates have been evaluated: >0.41 m/kyr at the base, 0.16 m/kyr between the LO of *P. lacunosa* and FO of *E. huxleyi*, and ~ 0.45 m/kyr in the uppermost ~ 5 m (Fig. S10). Consequently, the top of the TOL-1 section is dated at ~ 0.23 Ma (MIS 7). All paleomagnetic samples show normal magnetic polarity, consistent with deposition during the Brunhes Chron (Fig. S11). By considering that the top of the TOL-1 section (1177 m a.s.l.) was deposited at ~ 50 m b.s.l., when the sea level was ~ 18 m lower than the present (Antonioli et al., 2021), we calculate a mean uplift rate of ~ 5.2 m/kyr.

1.4. DISCUSSION

1.4.1. Biostratigraphic and paleoenvironmental signals

Despite the evident reworking, calcareous nannofossil analysis identified two robust bioevents that fix the age of the TOL-1 marine succession. Importantly, the FO of *E. huxleyi* (Fig. 2) constrains the age of the top of TOL-1 marine succession extend marine conditions in the Mut Basin up to ~ 0.23 Ma.

Planktonic foraminifera provide independent support for the Middle Pleistocene age. The presence of *G. calida* from the base of the section aligns with its known range beginning at an age younger than 0.53 Ma (Fig. S4). Notably, *G. ruber* (pink) was absent from the section, although it is commonly reported during interglacial periods elsewhere in the Mediterranean (e.g., MIS 11 in the east, MIS 9 in the west, and Holocene/MIS 5e in the west; Pérez-Folgado et al., 2004; Numberger et al., 2009). A proposed FO at ~ 0.33 Ma in the eastern Mediterranean (Lirer et al., 2019) remains uncertain due to the lack of published supporting evidence. Its sporadic distribution suggests ecological factors may influence its occurrence.

Ostracod assemblages provide complementary paleoenvironmental insights, documenting an oscillating shallowing-up trend from ~400 m to ~50 m water depth. This pattern reflects a gradual uplift beginning around 0.40 Ma, culminating in final emergence after ~0.23 Ma, when the youngest marine sediments correlate with MIS 7. At TOL-1, multiproxy signals converge on the onset of uplift around 0.40 Ma, when ostracod paleobathymetry records the start of shallowing, and the increase of reworked nannofossils and ostracods suggests enhanced erosion of older strata. According to the calcareous plankton bioevents, we propose a possible correlation between $\delta^{18}\text{O}$ *T. trilobus* from TOL-1 section with $\delta^{18}\text{O}$ *G. ruber* of ODP Site 967 (eastern Mediterranean, [Grant et al., 2022](#)). This correlation enables constraining the study section between MIS 14 and MIS 7 (Fig. S9).

1.4.2. Implications for the uplift history of the CAP southern margin

Although both sea-level fluctuations and tectonic activity can influence bathymetric records, the substantial elevation gain at TOL-1, from ~50 m b.s.l. at ~0.23 Ma to 1177 m a.s.l. today, indicates a rapid and sustained deep-seated geodynamic forcing, as already suggested for this region ([Faccenna et al., 2006](#); [Cosentino et al., 2012](#); [Schildgen et al., 2012; 2014](#); [Duretz and Gerya, 2013](#); [Göğüş et al., 2017](#); [Andersen et al., 2024](#); [Aykut et al., 2025](#)). Subduction dynamics in the eastern Mediterranean, including rollback, segmentation, and slab break-off, consistent with seismic tomography evidence along the Cyprus Arc, are well known to generate cycles of uplift, subsidence, and volcanism (e.g., [Biryol et al., 2011](#); [Portner et al., 2018](#); [Aydin et al., 2022](#)). The surface uplift documented at TOL-1 coincides with evidence for accelerated incision and surface uplift elsewhere in southern Anatolia between ~0.45 Ma and 0.25 Ma ([Racano et al., 2020, 2021](#); [Aykut et al., 2025](#)), supporting a link to slab break-off and subsequent mantle rebound. This underscores the critical role of deep-seated geodynamics in shaping the southern CAP, with TOL-1 serving as a key section that records the rapid Quaternary uplift of southern Anatolia.

1.4.3. Elevation of MIS 7 marine deposits in the world

MIS 7 marine deposits are preserved across a wide range of tectonic settings globally (Fig. 3; Table S7) and serve as key benchmarks for reconstructing Quaternary sea-level and uplift histories. On passive margins, MIS 7 deposits generally occur at low elevations commonly less than 60 m. Examples include the northern margin of the CAP (58 m), the Hampshire coast of the UK (~10 m), the eastern seaboard of the USA (~10 m), and multiple localities in South America (e.g., Patagonia, 20–35 m). Even in regions with repeated uplift signals, such as South Korea, MIS 7 elevations typically range between 30 m and 65 m.

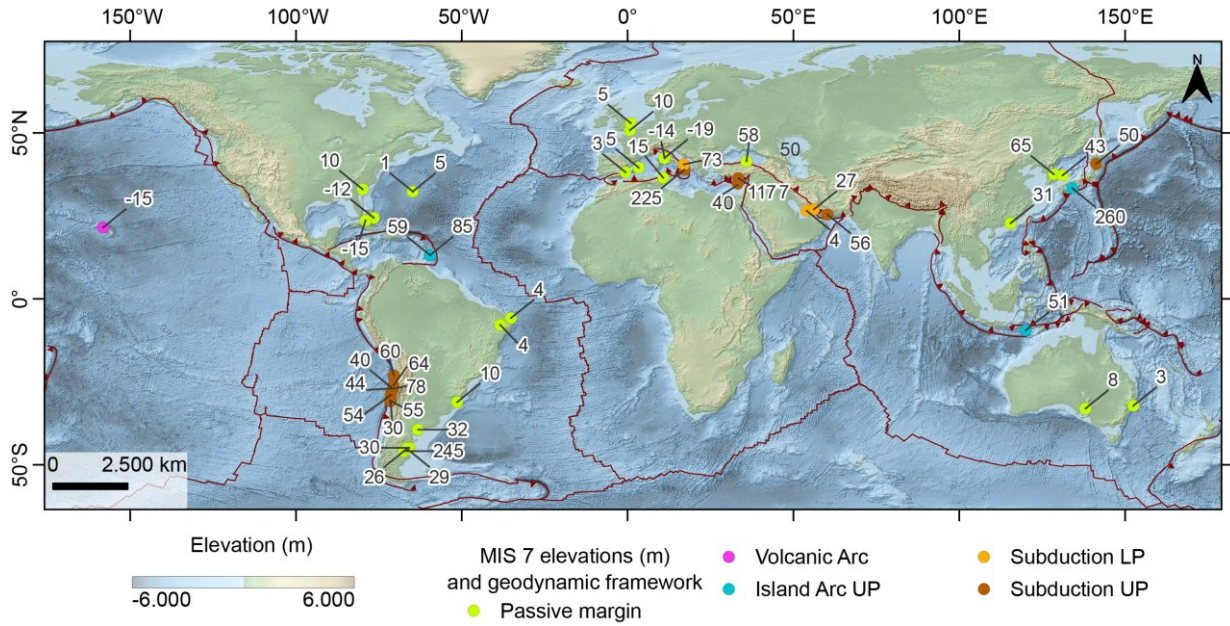


Figure 3. Global distribution of MIS 7 marine record, highlighting their varying elevations across different tectonic contexts (data set in Table S7).

On subduction-related margins, MIS 7 deposits are more variable and can reach significantly higher elevations than on passive margins (Fig. 3). In Calabria, marine terraces attributed to MIS 7 stand at ~225 m (Nalin et al., 2020). While along the Chilean coast, MIS 7 deposits occur between ~40 m and 80 m (Martinod et al., 2016). Island arc settings yield some of the highest values, such as the MIS 7 terraces in the Muroto Peninsula (Japan), which reach ~260 m (Matsu'ura, 2015). In contrast, volcanic arc settings such as Oahu preserve MIS 7 deposits below present sea level (–10 to –20 m), reflecting subsidence rather than uplift (Sherman et al., 2014).

Against this global distribution, the elevation of the TOL-1 section is extraordinary: MIS 7 marine deposits are preserved at 1177 m, an order of magnitude higher than any other known MIS 7 shoreline worldwide. This extreme elevation highlights a pronounced asymmetry between the southern (1177 m; this study) and northern (58 m; Berndt et al., 2018) margins of the CAP. While the southern margin experienced rapid Middle Pleistocene uplift driven by mantle upwelling following slab detachment beneath the Cyprus subduction zone, the northern margin evolved under a contrasting tectonic regime influenced primarily by transpressional deformation associated with the North Anatolian Fault and the lateral growth of the Pontide orogenic wedge. Studies of Quaternary fluvial and coastal terraces along the northern CAP indicate long-term, linked to activity of the North Anatolian Fault Zone and associated crustal thickening and wedge tectonics in the Pontides rather than to deep mantle upwelling processes (Yildirim et al., 2011). Consequently, uplift along the northern CAP margin has been more distributed and moderate in magnitude, whereas the southern margin records short-lived, high-

magnitude uplift pulses tied to deep mantle dynamics (Cosentino et al., 2012; Aykut et al., 2025). This north–south contrast is unusual among orogenic plateaus and reflects the unique tectonic setting of the CAP, bounded by an actively retreating/tearing slab system to the south and a transpressional fault system to the north.

Comparison with Cyprus further emphasizes this differential behavior. MIS 7 shorelines in Cyprus occur at relatively low elevations, reaching ~8–11 m a.s.l. along the western and southeastern coasts (Poole et al., 1990) and up to ~40 m a.s.l. in the Kyrenia Range (Palamakumbura et al., 2016). These elevations are one to two orders of magnitude lower than MIS 7-equivalent deposits along the southern CAP margin. Despite both regions occupying the upper plate of the Cyprus subduction system, vertical motions are therefore strongly partitioned across the plate boundary zone. This partitioning is likely controlled by the Anamur–Silifke–Kozan fault system, a major Quaternary normal-fault zone (Cosentino et al., 2012; Hall et al., 2014) that mechanically decouples the rapidly uplifting southern CAP from the more moderately uplifted Cyprus forearc. South of this structure, Cyprus records limited uplift related to slab rollback and flexural responses, whereas north of it, slab break-off and focused asthenospheric upwelling beneath the southern CAP produced extreme and spatially localized surface uplift.

Within the Mediterranean subduction zones, the stark contrast between MIS 7 elevation in the Calabrian Arc (225 m) and southern Anatolia (1177 m) can be explained by differences in slab geometry. P-wave model across the Cyprus subduction zone and the CAP southern margin reveals slab break-off (Portner et al., 2018), whereas tomography of the Calabrian Arc (Argnani et al., 2016) shows a continuous slab of the subducting Ionian plate (Fig. 4). Then, the extreme uplift at the CAP southern margin is best explained by mantle upwelling, which drove exceptionally rapid surface uplift during the late Middle Pleistocene.

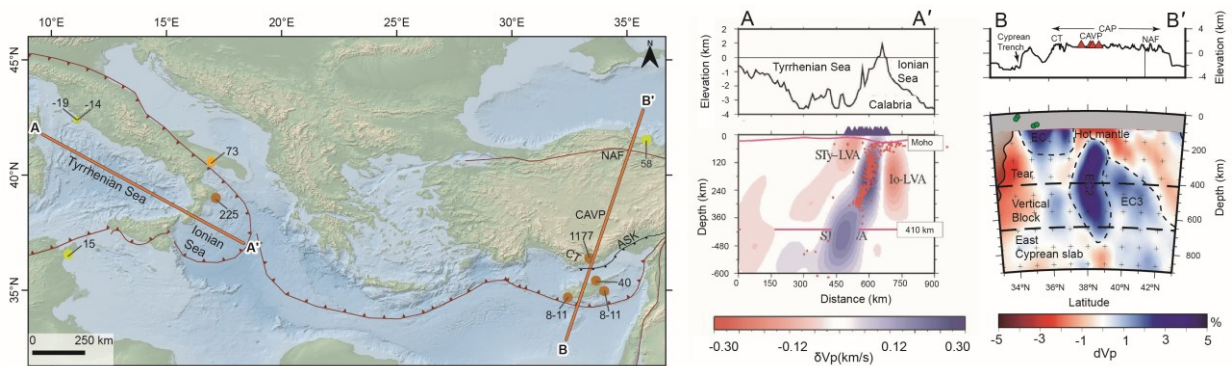


Figure 4. Map showing elevation and tectonic context of MIS 7 deposits (see Fig. 3 for legend), and locations of tomographic cross sections (A–A') across the Calabrian slab (Argnani et al., 2013) and (B–B') across the eastern Cyprian slab (Portner et al., 2018). P-wave velocity perturbation sections are shown, with a colour scale indicating velocity anomalies. CAVP - Central Anatolian Volcanic Province; NAF - North Anatolian Fault; EC - East Cyprian slab; STy-HVA - Southern Tyrrhenian high-velocity anomaly; STy-LVA - Southern Tyrrhenian low-velocity anomaly.

1.5. CONCLUSIONS

Along the southern margin of the CAP, in the western part of the Africa–Eurasia collision zone, late–middle Pleistocene (post-0.53 Ma) marine sediments of TOL-1 section occur at the highest elevation on Earth (~1177 m a.s.l.), indicating an average uplift rate of ~5.2 m/kyr. This exceptional rate of uplift is best explained by mantle upwelling triggered by slab break-off beneath the Cyprus Arc. These findings contribute significantly to our understanding of the geodynamic evolution at the western margin of the Africa–Eurasia collision zone. Precision estimates of rates of surface uplift of mountain ranges by dating uplifted marine deposits can provide useful constraints for defining and modeling the driving forces for uplift.

SUPPLEMENTAL MATERIAL

Sarı et al.

Refined Geological Mapping

To improve the delineation of Quaternary lithostratigraphic units, a Sentinel-2 satellite image (Copernicus dataset) was processed using ENVI software. Principal Component Analysis (PCA) was applied to enhance spectral variations within Quaternary deposits. The PCA results reveal distinct spectral domains corresponding to different lithological units. The spectral signature of the Middle Pleistocene deposits identified at the TOL-1 section was used as a reference to trace correlative units across the study area. Based on these results, unit boundaries were manually refined to produce the updated geological map (Fig. S1).

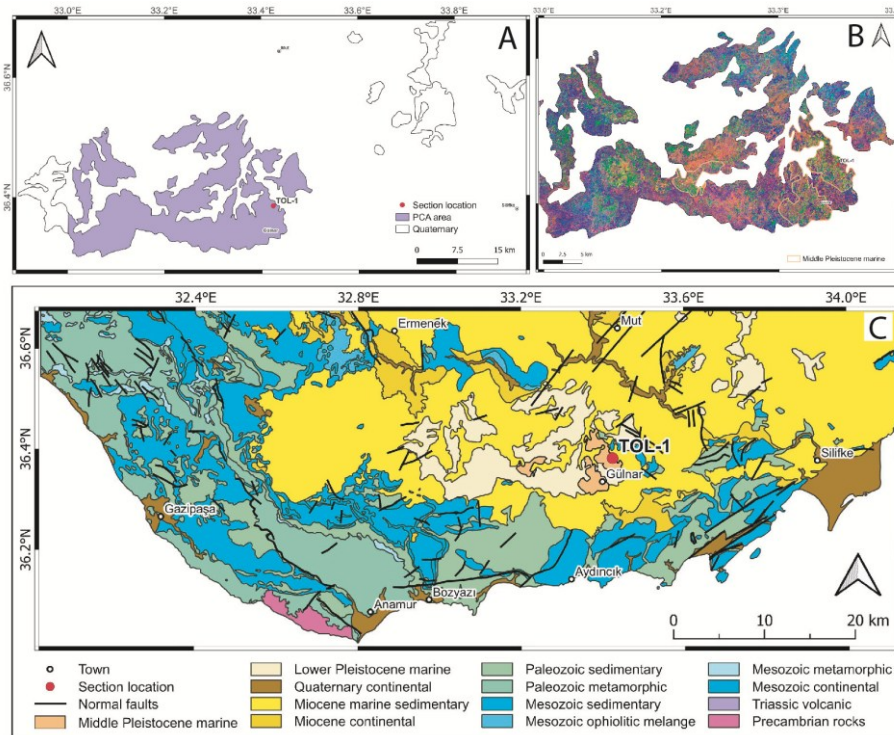


Figure S1. (A) Area selected for the PCA analysis, restricted to Quaternary deposits. (B) PCA result highlighting spectral contrasts, with manually redrawn boundaries distinguishing Lower and Middle Pleistocene lithologies. (C) Revised geological map of the southern margin of the Central Anatolian Plateau (CAP).

Table S1. Quantitative calcareous nannofossils biostratigraphy.

Sample ID	Depth (cm)	<i>Pseudoemiliania lacunosa</i>	Medium <i>Gephyrocapsa</i>	Small <i>Gephyrocapsa</i>	<i>Emiliania huxleyi</i>	<i>Rhabdosphaera clavigera</i>	<i>Syracosphaera</i> sp.	<i>Pontosphaera</i> spp.	<i>Umbilicosphaera jafari</i>	<i>Calcidiscus leptoporus</i>	<i>Coccolithus pelagicus</i>	<i>Umbilicosphaera sibogae</i>	<i>Helicosphaera</i> spp.	Reworked
TOL-1 66	6875	0	0	2	0	4	13	6	138	14	109	18	82	80
TOL-1 65	6785	0	2	0	3	23	66	40	978	37	259	90	200	233
TOL-1 64	6715	0	0	0	1	0	0	0	0	0	0	0	0	0
TOL-1 63	6420	0	0	1	4	23	82	73	1206	54	552	124	448	496
TOL-1 62	6300	0	0	1	2	50	130	32	460	224	925	106	180	901
TOL-1 61	6190	0	0	2	3	18	65	29	474	75	283	30	133	556
TOL-1 60	6095	0	0	0	1	16	126	56	314	204	636	26	110	1146
TOL-1 59	6000	0	0	1	0	25	41	64	346	28	420	31	335	502
TOL-1 58	5790	0	0	0	0	8	74	31	324	59	390	10	83	694
TOL-1 57	5590	1	0	1	0	48	89	30	1098	323	978	52	143	1316
TOL-1 56	5290	1	0	1	0	28	70	53	480	114	780	53	176	786
TOL-1 55	5170	2	0	0	0	3	40	38	492	35	372	35	235	596
TOL-1 54	4975	0	0	0	0	3	12	6	378	3	17	17	82	301
TOL-1 53	4880	0	1	0	0	9	32	39	144	28	660	37	211	1056
TOL-1 52	4810	0	0	1	0	4	6	17	35	34	72	8	41	328
TOL-1 51	4675	0	0	0	0	9	45	13	242	10	65	17	76	322
TOL-1 50	4555	0	0	0	0	13	47	21	270	28	232	58	79	490
TOL-1 49	4450	0	0	3	0	13	21	9	408	32	126	113	52	329
TOL-1 48	4325	0	0	0	0	0	0	0	0	0	0	0	0	0
TOL-1 47	4265	0	0	0	0	0	0	0	0	0	0	0	0	0
TOL-1 46	4225	2	1	2	0	34	87	93	875	64	852	119	393	1420
TOL-1 45	4140	0	0	0	0	30	34	32	542	34	174	108	85	321
TOL-1 44	4025	0	1	1	0	22	30	19	131	41	384	21	76	213
TOL-1 43	3965	0	0	2	0	1	17	39	168	16	522	9	129	385
TOL-1 42	3840	1	1	4	0	25	52	75	732	78	1146	92	145	1158
TOL-1 41	3645	0	0	1	0	59	69	53	928	13	1164	78	129	1399
TOL-1 40	3550	0	1	0	0	2	10	21	114	7	274	13	34	252
TOL-1 39	3410	0	0	2	0	12	35	43	116	57	1304	61	109	828
TOL-1 38	3320	0	0	1	0	20	128	70	388	46	1126	80	243	563
TOL-1 37	3200	0	1	0	0	16	56	43	298	77	892	48	215	404
TOL-1 36	3120	0	0	0	0	10	24	21	276	85	664	34	88	561
TOL-1 35	3035	0	0	0	0	3	28	14	268	67	252	41	79	385
TOL-1 34	2910	16	0	1	0	54	175	90	1272	45	350	61	405	432

TOL-1 33	2865	1	0	1	0	23	41	39	270	30	279	48	275	795
TOL-1 32	2800	1	0	0	0	166	62	67	310	25	1302	84	442	620
TOL-1 31	2640	2	1	0	0	26	52	24	222	52	1031	38	146	640
TOL-1 30	2555	0	0	1	0	26	43	34	816	25	787	55	262	390
TOL-1 29	2490	0	0	2	0	4	5	31	37	7	120	9	75	269
TOL-1 28	2410	5	3	12	0	88	132	96	672	72	1110	191	217	812
TOL-1 27	2340	1	0	6	0	15	51	72	74	49	1329	91	187	514
TOL-1 26	2260	0	1	7	0	90	102	45	1482	84	117	280	185	447
TOL-1 25	2180	1	0	5	0	45	119	121	387	300	2124	576	236	1070
TOL-1 24	2085	0	2	8	0	7	37	62	45	59	383	110	86	303
TOL-1 23	2020	0	0	9	0	9	67	90	576	106	221	307	458	418
TOL-1 22	1930	0	1	1	0	2	18	18	42	19	239	22	36	102
TOL-1 21	1855	2	0	1	0	2	10	34	24	3	276	36	75	185
TOL-1 20	1800	0	0	0	0	5	16	68	29	5	1068	43	193	493
TOL-1 19	1760	3	1	7	0	36	93	146	167	48	888	187	510	904
TOL-1 18	1680	1	0	10	0	19	71	37	251	10	241	141	160	465
TOL-1 17	1625	1	0	4	0	19	37	87	91	24	478	79	171	596
TOL-1 16	1560	1	1	1	0	4	7	28	17	2	254	17	95	252
TOL-1 15	1520	0	0	3	0	11	24	49	46	77	570	62	102	558
TOL-1 14	1410	2	3	14	0	7	41	47	168	198	388	132	153	692
TOL-1 13	1340	0	0	4	0	31	26	22	342	133	158	113	153	405
TOL-1 12	1240	1	1	3	0	12	42	59	179	71	182	54	132	387
TOL-1 11	1120	1	4	5	0	15	26	24	586	59	54	104	96	179
TOL-1 10	1020	0	0	0	0	0	0	0	0	0	0	0	0	0
TOL-1 9	910	0	0	4	0	26	149	105	205	69	379	192	142	498
TOL-1 8	840	1	1	0	0	24	34	58	210	17	204	109	216	495
TOL-1 7	760	1	0	1	0	3	13	25	37	25	234	55	77	192
TOL-1 6	690	2	2	3	0	18	36	55	67	9	163	258	134	545
TOL-1 5	600	2	2	0	0	36	21	61	59	25	632	244	230	1324
TOL-1 4	530	11	3	1	0	12	67	20	116	17	207	247	141	510
TOL-1 3	435	1	2	3	0	25	157	38	88	36	234	108	66	790
TOL-1 2	400	1	0	0	0	17	72	136	86	14	183	240	283	746
TOL-1 1	250	1	0	1	0	4	62	65	229	62	202	198	143	658

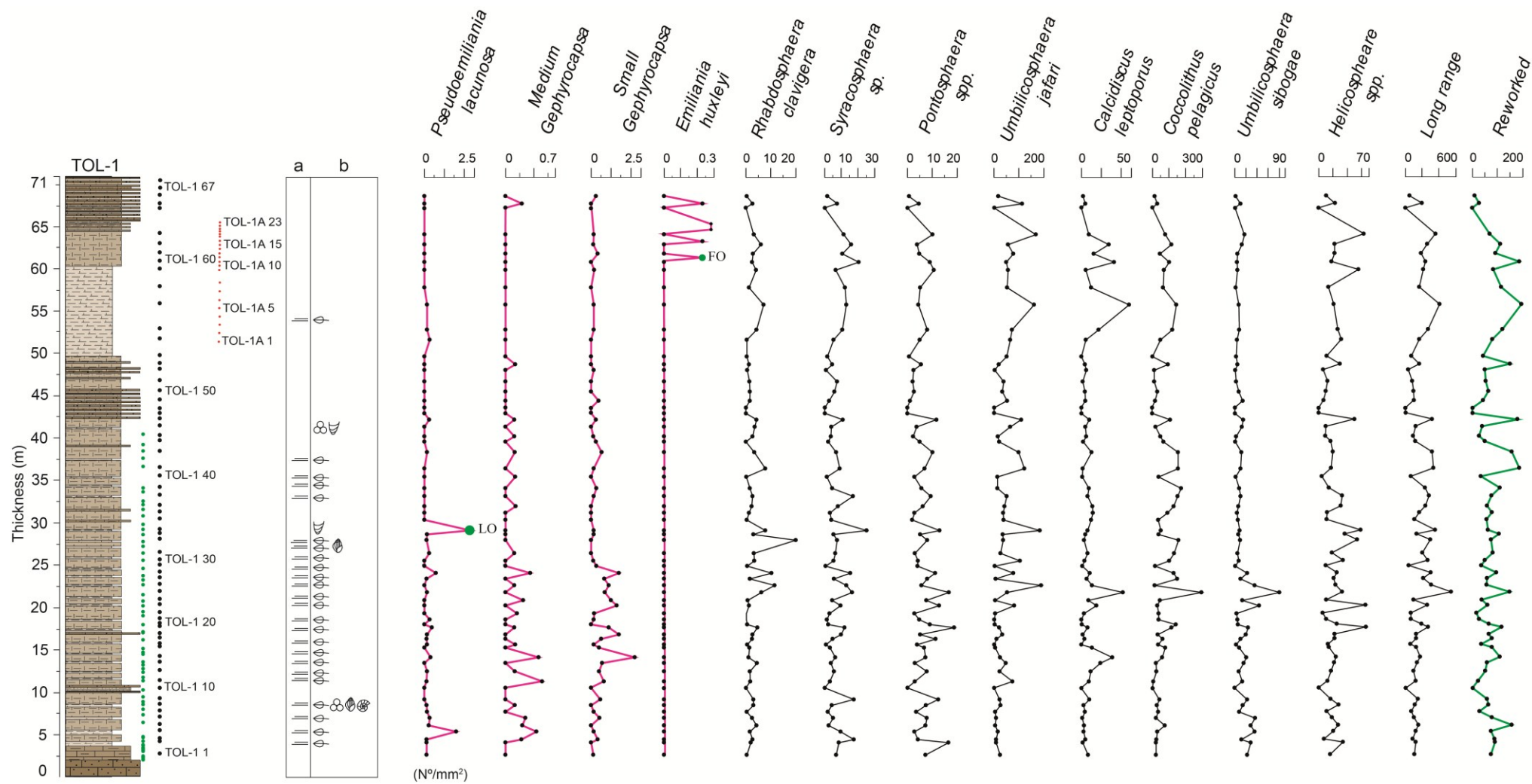


Figure S2. Calcareous nannofossil biostratigraphy of the TOL-1 section. Pink lines show index species, red circles mark concentrated samples, black lines indicate long-range species, and green lines show reworked species. Panels: (a) sedimentary structures; (b) fossil components.

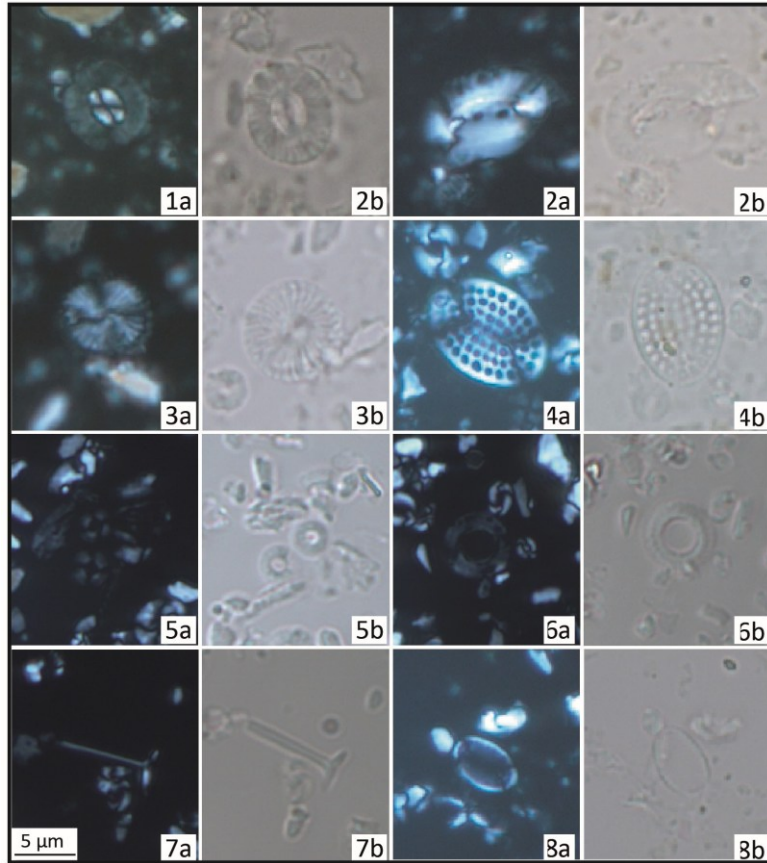


Figure S3. Microphotographs of calcareous nannofossils long-range taxa from the TOL-1 section.—(1) *C. pelagicus* (TOL-1 5); (2) *H. carteri* (TOL-1 2); (3) *C. leptoporus* (TOL-1 1); (4) *Pontosphaera* spp. (TOL-1 2); (5) *U. jafari* (TOL-1 1); (6) *U. sibogae* (TOL-1 1); (7) *R. clavigera* (TOL-1 1); (8) *Syracosphaera* spp.- (TOL1 1). (a) Crossed nicols; (b) Parallel nicols. Scale bar uniform for all images.

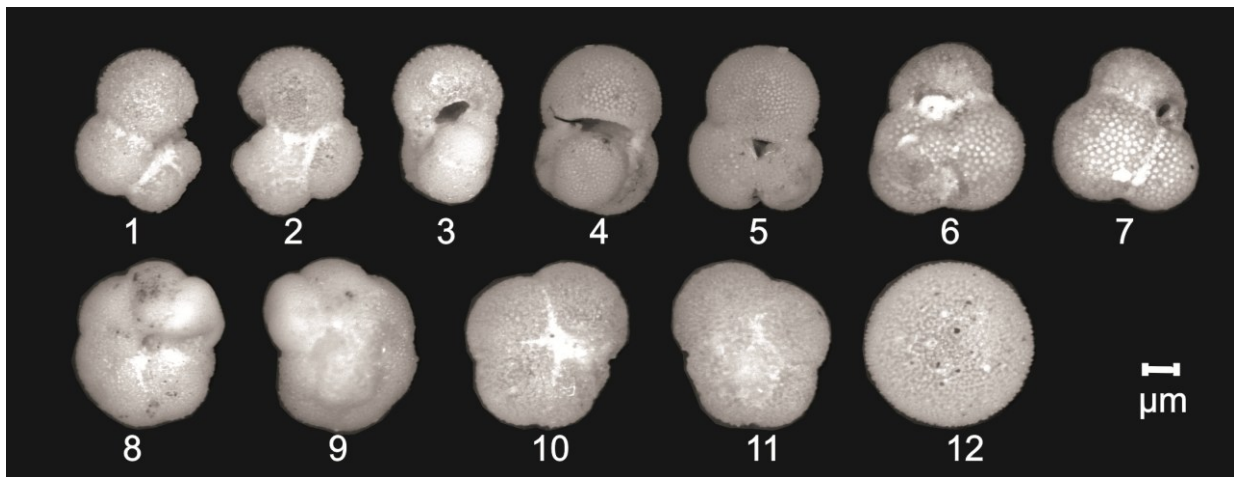


Figure S4. Scanning electron microscope (SEM) images of selected planktonic foraminiferal taxa from the TOL-1 section. (1) *Globigerinella calida* ventral view; (2) *Globigerinella calida* dorsal view; (3) *Globigerinella calida* axial view; (4) *Globigerinoides quadrilobatus* ventral view; (5) *Globigerinoides quadrilobatus* dorsal view; (6) *Globigerinoides sacculifer* ventral view; (7) *Globigerinoides sacculifer* dorsal view; (8) *Neogloboquadrina dutertrei* ventral view; (9) *Neogloboquadrina dutertrei* dorsal view; (10) *Neogloboquadrina pachyderma* ventral view; (11) *Neogloboquadrina pachyderma* dorsal view; (12) *Orbulina universa*.

Table S2. The semi-quantitative analysis on planktonic foraminifera was based on surveying a standard number of fields in the washing residue >125micron (according to Hilgen et al., 2003), distinguishing the following abundance categories: Present (P)<3 specimens, Rare (R) <3-10 specimens, Few (F) 10-30 specimens, Common (C) 30-50 specimens and Abundant (A) > 50 specimens. Preservation: low preservation (-); moderate preservation (+); good preservation (++). Indeterminate species are referred to incrustated and recrystallized species.

Sample ID	Thickness (cm)	Remarks on abundance	Preservation	Indeterminate species	<i>G. quadrilobatus</i>	<i>T. trilobus</i>	<i>G. glutinata</i>	<i>Orbulina universa</i>	<i>G. calida</i>	<i>N. incompta</i>	<i>N. pachyderma</i>	<i>N. dutertrei</i>	<i>G. ruber (white)</i>	<i>G. bulloides</i>	<i>G. obliquus obliquus</i>	<i>G. rubescens</i>	<i>G. tenellus</i>
TOL1-67	6995	P	-	P	R	R	P									P	
TOL1-66	6875	P	-	P	P	P		R								P	
TOL1-65	6785	P	-	P	R	R		P						P		P	
TOL1-64	6715	F	+	C				F								P	
TOL1-63	6420	F	+	C	R	R	F	C	F	R			P			P	
TOL1-62	6300	C	+	C	C	C		C	P				P	P	R	R	
TOL1-61	6190	A	++	C	C	C		F						R	C	F	
TOL1-60	6095	A	++	F	D	D			F	R	R			P	C	F	F
TOL1-59	6000	A	++	F	D	D	F	C	F	F	F			R	C	F	
TOL1-58	5790	A	+	C	D	D		C					R		C	F	
TOL1-57	5590	A	+	C	C	C	F	C							C	F	
TOL1-56	5290	A	+	C	D	D		C		P				P	F	R	
TOL1-55	5170	A	+	C	C	C		D						C	C	F	
TOL1-54	4975	A	+	C	C	C		C							C	F	
TOL1-50	4555	A	+	C	C	C		C							C	F	
TOL1-49	4450	A	+	C	C	C		C							F	R	
TOL1-46	4225	A	++	F	C	C		D						F	C	F	
TOL1-41	3645	A	++	F	D	D	F	C						F	C	F	
TOL1-38	3320	A	++	F	D	D		C	C					F			
TOL1-35	3035	A	++	F	D	D		F	C	R							
TOL1-32	2800	A	++	F	D	D		C	C						F	F	
TOL1-27	2340	A	++	F	D	D		C					C	F	F	F	
TOL1-25	2180	A	++	F	C	C		D						F	C	F	
TOL1-23	2020	A	++	F	C	C		D	F						C	F	
TOL1-21	1855	A	++	F	C	C		C				C			F	R	
TOL1-19	1760	A	++	F	C	C		D	R	R					C	F	
TOL1-16	1560	A	++	F	C	C		D	R	R					C	F	
TOL1-13	1340	C	++	F	C	C	R	D	R	R	R	F		R	F		
TOL1-6	690	C	+	A	C	C	R	D		P	P	F		R	F		
TOL1-4	530	C	+	C	C	C	R	D		P	P	F		R	F		
TOL1-2	400	A	++	C	C	C		F	R		R	F		F	F		

Ostracods

Overall, the TOL-1 section yielded 107 ostracod taxa. Thirty-six species were considered reworked since their known stratigraphic distribution ended before Chibanian (Tab. S3). They are mainly shallow water species reworked from the Messinian marly limestones of the Mut Fm (i.e., Sarıalan and Başyayla sections – Babinot, 2002; Cosentino et al. 2012), from the Zanclean-Gelasian section of Kargıcak (Yıldız et al., 2003), and from the Calabrian section of Gülnar East (Öğretmen et al., 2018).

Twenty-two species were considered displaced, possibly from the nearby Tol section (Öğretmen et al., 2018) based on the population structure of each taxon, which, being unbalanced between adult and juvenile valves, are referable to transported death assemblages in high or low energy environments (Boomer et al., 2003). Among them, *Bythocypris bosquetiana*, *Neglecandona* sp., *Cistacythereis emaciata*, *Cytherelloidea* sp. *Cyprideis* sp. *Kejiella* sp. *Macrocyprissa arcuata*, *Palmoconcha turbida* and *Sagmatocythere* sp. are surely displaced being represented only by few juveniles. Moreover, in some samples also some species represented by fragmentary or abraded valves were considered as displaced, i.e. among others, in samples TOL-1 2, TOL-1 13, TOL-1 28, TOL-1 29, TOL-1 35, TOL-1 39, and TOL-1 44 *Grinioneis haidingeri*, in samples TOL-1 6 and TOL-1 41 *Xestoleberis communis*, in samples TOL-1 7, TOL-1 9, TOL-1 11, TOL-1 12, TOL-1 13, TOL-1 14, TOL-1 24, TOL-1 25, TOL-1 26 *Aurila convexa*, in sample TOL-1 17 *Bythocypris producta*.

Forty-nine taxa (45 of which were identified at the species level) were considered autochthonous for their rather good preservation, the presence of carapaces and the balanced population age structures. Their percentage frequencies in each sample are reported in Tab. S4. Frequencies of autochthonous vs. reworked taxa and the percentages of the most representative autochthonous species are illustrated in Fig. S5. Selected specimens are reported in Fig. S6.

On the autochthonous taxa, the unconstrained cluster analysis (Bray-Curtis similarity and the unweighted pair group method using arithmetic average-UPGMA) in Q-mode was performed using the PAST software, ver. 4.06b (Hammer et al., 2001). Cluster analysis dendrogram, whose suitability was checked through the cophenetic correlation coefficient (c) (Sokal and Rohlf, 1962; Mouchet et al., 2008), has been used to determine sample groups based on their ostracod community composition. At a similarity level of 0.23, the obtained dendrogram reported in Fig. S7 (c=0.75) discriminates six clusters.

Cluster 1 includes only sample TOL-1 63 characterised by the presence of the eurybathic species *Argilloecia acuminata*, accompanied by the infralittoral species *Paracytheridea triquetra*.

Cluster 2 groups samples TOL-1 2, TOL-1 6, TOL-1 10, TOL-1 14, TOL-1 35, TOL-1 41, TOL-1 42, TOL-1 45, TOL-1 46, TOL-1 51, TOL-1 53, TOL-1 55, TOL-1 56, TOL-1 57, and TOL-1 62, dominated by the eurybathic species *Acanthocythereis hystrix* and the littoral species *Grinioneis haidingeri*.

Cluster 3 gathers samples TOL-1 3, TOL-1 5, TOL-1 7, TOL-1 8, TOL-1 9, TOL-1 11, TOL-1 12, TOL-1 13, TOL-1 15, TOL-1 16, TOL-1 17, TOL-1 18, TOL-1 19, TOL-1 20, TOL-1 21, TOL-1 22, TOL-1 23, TOL-1 24, TOL-1 25, TOL-1 26, TOL-1 27, TOL-1 28, TOL-1 29, TOL-1 30, TOL-1 31, TOL-1 32, TOL-1 33, TOL-1 36, TOL-1 37, TOL-1 38, TOL-1 39, TOL-1 40, TOL-1 43, TOL-1 44, TOL-1 49, TOL-1-50, TOL-1-52, TOL-1-54, TOL-1 58, TOL-1 59, TOL-1 60, and TOL-1 61, characterised by the co-dominance of the epibathyal species *Bythocypris producta*, *Cytherella robusta* and *Henryhowella sarsi profunda*, occasionally accompanied by the eurybathic species *Krithe iniqua*.

Cluster 4 includes sample TOL-1 1, characterised by the dominant littoral species *Aurila convexa* accompanied by the eurybathic species *Krithe iniqua*.

Cluster 5 is made only by sample TOL-1 34 characterized by the only presence of the eurybathic species *Acanthocythereis hystrix*.

Finally, Cluster 6 includes only the topmost sample TOL-1 65 characterized by the presence of only the littoral species *Hemicytherura defioei*.

Using the distribution of the samples in the dendrograms and considering the dominant species characteristic of each cluster and their autoecology, the paleobathymetric reconstruction of the TOL-1 section, reported in Fig. S8 has been proposed.

Table S3. List of the reworked ostracod species collected in the TOL-1 section.

Species	Stratigraphic range	Bathymetry	Literature
<i>Aurila albicans</i> (Ruggieri, 1958)	Middle Serravallian-Early Messinian	0-25 m	Bonaduce et al., 1992; Conesa and Babinot, 1999; Faranda et al., 2008; Ruggieri, 1958, 1962, 1967, 1976; Sissingh, 1972.
<i>Aurila calciplena</i> (Uliczny, 1969)	Pliocene	littoral	Uliczny, 1969; Mostafawi and Matzke-Karasz, 2006.
<i>Aurila cicatricosa</i> (Reuss, 1850)	Serravallian- Piacenzian	0-25 m	Faranda et al., 2008.
<i>Aurila freudenthali</i> (Sissingh, 1972)	Middle Serravallian-Early Messinian	0-25 m	Aruta, 1983; Babinot and Boukli-Hacene, 1998; Bonaduce et al., 1987, 1992; Doruk, 1973; Faranda et al. 2008; Miculan, 1992; Sissingh, 1972.
<i>Aurila longa</i> (Ruggieri, 1962)	Tortonian- Early Messinian	0-50 m (<i>Porites</i>)	Faranda et al., 2008.
<i>Aurila punctata</i> (Munster, 1830)	Langhian-Calabrian	0-50 m	Ducasse and Cahuzac, 1996; Gross, 2006; Ruggieri, 1962, 1975;
<i>Bairdoppilata profunda</i> (Aiello, Barra and Bonaduce, 2000)	Langhian- Gelasian	300-1000 m	Aiello and Barra, 2001; Aiello et al., 2000; Bonaduce and Barra 2002; Faranda et al., 2007; Sciuto, 2003; Sciuto et al., 2003.
<i>Buntonia dertonensis</i> (Ruggieri, 1954)	Zanclean-Middle Gelasian	50-1000 m	Aiello et al., 2000; Ciampo 1986.
<i>Callistocythere montana</i> (Doruk, 1980)	Serravallian-Early Messinian	0-50 m	Doruk, 1980; Faranda et al., 2008
<i>Callistocythere perfossa</i> (Ciampo, 1984)	Tortonian -Early Messinian	0-50 m	Ciampo, 1984.
<i>Callistocythere</i> sp. 1 (Babinot, 2002)	Messinian Inf	0-50 m	Babinot, 2002.
<i>Cardobairdia</i> sp. 1 (Dall'Antonia, 2003)	Early Langhian-Early Tortonian	epibathial	Dall'Antonia, 2002, 2003.
<i>Cimbaurila diecii</i> (Sissingh, 1972)	Late Serravallian-Early Messinian	0-25 m	Faranda et al., 2008
<i>Cnestocythere lamellicosta</i> (Triebl, 1950)	Langhian.-Tortonian	0-50 m	Aiello and Szczechura, 2004; Brestenska and Jiricek, 1978; Gross, 2006; Zorn, 2004
<i>Costa (Cuneocosta) tricolostata pliocenica</i> (Ruggieri 1992)	Zanclean	500 m	Ruggieri 1992
<i>Cytherella cercinata</i> (Aiello, Barra, Bonaduce and Russo, 1996)	Tortonian - Pliocene	300-1000 m	Aiello et al., 1996; Doruk, 1974; Sissingh, 1972.
<i>Cytherella rusoi</i> (Sissingh, 1972)	Langhian- Gelasian	50-1000 m	Sissingh, 1972; Aiello et al., 2000; Bonaduce and Russo, 1985.
<i>Eucytherura lamina</i> (Ciampo, 1986)	Tortonian -Early Messinian	50-500	Ciampo, 1986.
<i>Flexus triebeli</i> (Ruggieri, 1962)	Zanclean - Calabrian	5-40 m	Faranda et al., 2007
<i>Graptocythere intricata</i> (Terquem, 1878)	Tortonian - Calabrian	0-50 m	Faranda et al., 2007
<i>Helioocythere vejhonensis</i> (Prochazka, 1893)	Langhian.-Early Messinian	0-25 m	Aiello and Szczechura, 2004; Bonaduce et al., 1998a; Conesa and Babinot, 1999; Faranda et al., 2008; Gross, 2006.
<i>Loxoconcha punctatella</i> ; (Reuss, 1850)	Aquitania–Early Messinian	0-50 m (<i>Porites</i>)	Aiello and Szczechura, 2004; Conesa and Babinot, 1999; Faranda et al., 2008;
<i>Loxocorniculum quadricornis</i> (Ruggieri, 1962)	Serravallian - Tortonian	0-50 m	Babinot 2002; Conesa and Babinot, 1999; Dall'Antonia, 2002. Ruggieri, 1962.
<i>Oblitacythereis mediterranea</i> (Benson, 1977)	Zanclean	250-1000 m	Aiello and Barra, 2001; Aiello et al., 2000; Barra et al., 1998; Bonaduce and Sprovieri, 1985; Violanti et al., 2009.
<i>Occultocythereis bituberculata</i> (Reuss, 1850)	Langhian –Early Messinian	0-50 m	Aiello and Szczechura, 2004; Babinot 2002; Dall'Antonia and Bossio, 2001; Gross, 2006;
<i>Orionina tegminata</i> (Doruk, 1974)	Tortonian	0-50 m	Babinot, 2002
<i>Pachicaudites attenuatus</i> (Ruggieri, 1980)	Tortonian–Early Messinian	0-50 m	Ciampo, 1986.
<i>Paracytheridea inscisa</i> (Doruk, 1980)	Tortonian - Messinian inf.	0-50 m	Doruk, 1980; El-Waer, 1988
<i>Parakriithe ariminensis</i> (Ruggieri, 1967)	Langhian. - Piacenzian	100-1000 m	Aiello et al., 2000; Aiello and Barra, 2001; Bonaduce and Barra, 2002; Faranda et al., 2007; Ruggieri, 1967;
<i>Pokorniella devians</i> (Bonaduce, Ruggieri and Russo, 1985)	Tortonian–Early Messinian	0-50 m	Bonaduce and Russo, 1985; Faranda et al., 2008; Sissingh, 1972;
<i>Pokorniella italica</i> (Ruggieri, Russo and Bossio, 1977)	Tortonian–Early Messinian	0-25 m	Aruta, 1983; Bonaduce et al., 1992; Conesa and Babinot, 1999; Faranda et al., 2008; Miculan, 1992; Ruggieri et al., 1977; Sissingh, 1972;
<i>Rethybythere (Bathybytere) vandenboldi</i> , (Ruggieri, 1960)	Gelasian-Calabrian	500-1000 m	Aiello et al., 2000; Bonaduce et al., 2000; Colalongo, 1966; Colalongo and Pasini, 1980.
<i>Ruggieria tetraptera</i> (Seguenza, 1880)	Aquitania - Pliocene	0-1000 m	Aruta, 1983; Bonaduce and Barra, 2002; Bossio et al., 2006; Ciampo, 1980; Conesa and Babinot, 1999; Miculan, 1992; Ruggieri, 1957.
<i>Sagmatocythere cristatissima</i> , (Ruggieri, 1967)	Tortonian–Early Messinian	0-50 m (<i>Porites</i>)	Faranda et al., 2008
<i>Sagmatocythere variesculpta</i> (Ruggieri, 1962)	Tortonian - Messinian	littoral	Ruggieri, 1962; Miculan, 1992;
<i>Sylvestra</i> sp. (Mostafawi, 1990)	Tortonian -Early Messinian	0-50 m	Mostafawi, 1990
<i>Tenedocythere cruciata</i> (Bonaduce, Ruggieri and Russo, 1986)	Serravallian-Early Messinian	0-25 m	Bonaduce et al., 1984; Faranda et al., 2008; Mostafawi, 2002.

Table S4. Percentage frequencies of the autochthonous ostracod species collected in the TOL-1 section.

Samples	Thickness (m)	<i>A. hystrix</i>	<i>A. acuminata</i>	<i>A. fatua</i>	<i>A. micra</i>	<i>A. convexa</i>	<i>A. nevanii</i>	<i>B. conformis</i>	<i>B. berchoni</i>	<i>B. producta</i>	<i>C. cf. rastrifera</i>	<i>C. carinata</i>	<i>C. calceolatus</i>	<i>C. biflexa</i>	<i>C. quadridentata</i>	<i>C. rubra</i>	<i>Cistacythereis</i> sp.	<i>C. edwardsii</i>	<i>C. robusta</i>	<i>E. pustulata</i>	<i>E. angulata</i>	<i>G. haidingeri</i>	<i>H. deflorei</i>	<i>H. sarsi profunda</i>
65	67.8																						100	
63	64.1		67																					
62	63										22					11						50		
61	61.8	1	2			5			2	37		1	2	5					22					
60	60.9		3	6						48				5					6		2			
59	59.9			3						95														
58	57.7		9							54									4		2			
57	55.8	29												32								29		
56	52.6	34	1			4	1	5		14		1		7								23		
55	51.5	18	2	4				4		7				5					15			25		
54	49.5	20	4			10													22		4			24
53	48.6	7				26	4		1			2		14		3	2			1	1	19		19
52	48.2									38				50										
51	46.7	9				9	5	1		2		4		11					3			23		16
50	45.5	7					2		1	10		3		7					1			10		54
49	44.3													21					50					29
46	42.2	8				5	2	2		3				16					4			15		27
45	41.3	13	2			13				10				5								21		16
44	40.2							19		38									10			10		
43	39.5	19										4				7			11			7		37
42	38.2	12	1				4												3		2	26		32
41	36.5	8		2		4		3		13		1		5					9			20		20
40	35.5			38															38					25
39	34	9						6		34									9			3		24
38	33	3						6		55		3							23					6
37	32									24									13					41
36	31.1									50									20					13
35	30.3																		33			33		
34	29.2	100																						
33	28.6	3	9							25									6					
32	28	0	4							23									30					13
31	26.5	1								15									15	1		1		33

30	25.6									21						1			24			1		
29	24.8									15									3			3		43
28	24									15									34			7		22
27	23.4	3	10							7									10		1			1
26	22.6			4			4			37									26					26
25	21.7						3			13									46					7
24	20.8						3			15									21					46
24	20.8									14		5							19					57
23	20.1																		3		2			48
22	19.4																		13					80
21	18.6									10									17					
20	18						1			16									30		1			1
19	17.5	2								10				2					14		1			14
18	16.9		5							3									12					38
17	16.2																	13	13					75
16	15.6		3		3					7						2			9				2	40
15	15.2					8	5	3		6		3	3	13					5			28		15
14	14.1		4			11	5	5		22		2	1		6				11			5		15
13	13.4					9		4		7									18					39
12	12.4					9		6		14									11		11	11		
11	11.2						13												7			67		13
10	10.2		2			5				8									20			7		23
9	9					2				26									20					33
8	8.4					4				28									11					19
7	7.5	15														8				15				31
6	6.8									20									28			2		13
z5	5.7																							
4	5.1									14				1					22			1		14
3	4.3																				40			20
2	3.7					71																		
1	2.3																						100	

Table S4 (continued). Percentage frequencies of the autochthonous ostracod species collected in the TOL-1 section.

Continued: Percentage frequencies of the autochthonous ostracod species collected in the FOL 1 section.																								
Samples	Thickness (m)	<i>H. albomaculata</i>	<i>K. compressa</i>	<i>K. iniqua</i>	<i>K. pernoides</i>	<i>K. perpulcra</i>	<i>L. ovulata</i>	<i>M. xenomis</i>	<i>M. mediterranea</i>	<i>P. depressa</i>	<i>P. triquetra</i>	<i>P. tetrapteron</i>	<i>P. ambigua</i>	<i>P. dimorpha</i>	<i>P. rotundata</i>	<i>P. semilunaris</i>	<i>P. turbida</i>	<i>P. reniformis</i>	<i>P. reniformis</i>	<i>Sagmatocythere sp. 1</i>	<i>S. alifera</i>	<i>Semicytherura sp. 1.</i>	<i>X. communis</i>	<i>X. ispar</i>
65	67.8																							
63	64.1										33													
62	63																				5	12		
61	61.8			11			7													4	1			
60	60.9						2								19									
59	59.9																							
58	57.7														26	6								
57	55.8														6									
56	52.6			5																1				
55	51.5		13						2		2													
54	49.5						12																	
53	48.6																							1
52	48.2																							
51	46.7		1				4													1		1		
50	45.5		1								1													
49	44.3																							
46	42.2		1				3													1				
45	41.3	3	5						3														5	
44	40.2																							
43	39.5											11												
42	38.2			3				1												1				
41	36.5			2					1					1	6					1	1		4	
40	35.5																							
39	34																							
38	33		3																					
37	32			1			1								1									
36	31.1																							
35	30.3																							
34	29.2																							

33	28.6		13										6	22										
32	28			6										13										
31	26.5		3	6									1											
30	25.6		53																					
29	24.8			28										10										
28	24			20																				
27	23.4		18	6										30			12							
26	22.6						2			2														
25	21.7			23																				
24	20.8			5			2																	
24	20.8																							
23	20.1		40				2																	
22	19.4			7																				
21	18.6			73																				
20	18		19	30																				
19	17.5		32	7			2							8				2						
18	16.9		5	30			1																	
17	16.2																							
16	15.6		2	19			2																	
15	15.2						1										3						5	
14	14.1		2	1			1																5	2
13	13.4			21																				
12	12.4			26																				
11	11.2																							
10	10.2		21	8	2																		1	
9	9		7																					
8	8.4		5	31								1												
7	7.5			8																			15	
6	6.8				38																			
5	5.7																							
4	5.1		17	16		15																		
3	4.3								20															
2	3.7			29																				
1	2.3																							

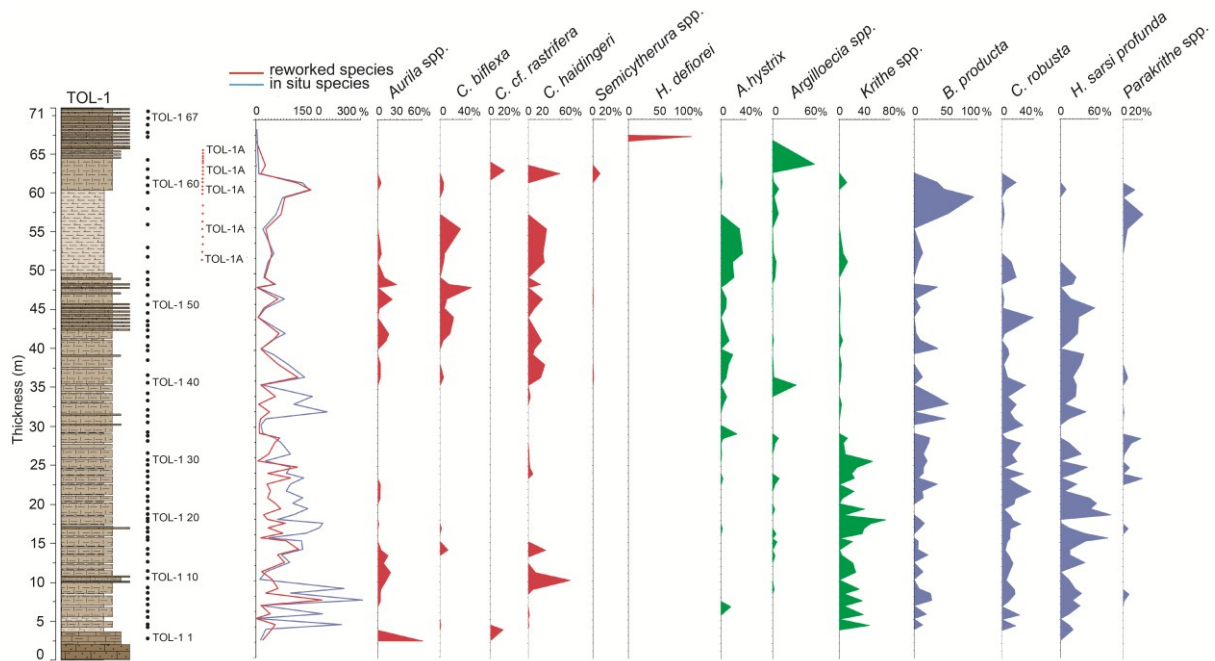


Figure S5. Autochthonous vs. reworked ostracod curve (on the left) and percentage frequencies of the main ostracod species collected in TOL-1 section. In red, littoral species, in green, eurybathic species, in blue, epibathyal species. Autoecological characteristics of the species derived from specialized literature (Tab. S5).

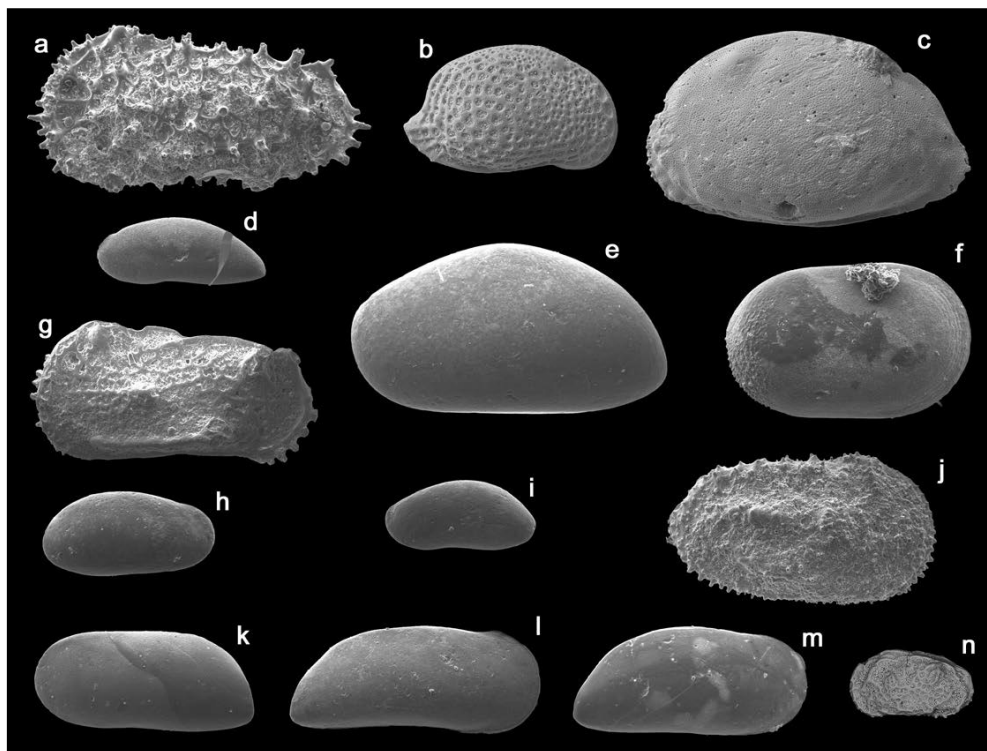


Figure S6. SEM images of selected ostracod species—(a) *Acanthocythereis hystrix*, VS, TOL-1 6. (b) *Aurila convexa*, VD, TOL-1 13. (c) *Aurila neviani*, VS, TOL-1 51. (d) *Argilloecia acuminata*, VS, TOL-113. (e) *Bythocypris producta*, VS, TOL-1 15. (f) *Cytherella robusta*, VD, TOL-1 12. (g) *Grinioneis haidingeri*, VS, TOL-1 3. (h) *Parakrithe rotundata*, VD, TOL-1 33. (i) *Argilloecia fatua*, VS, TOL-1 40. (j) *Henryhowella sarsi profunda*, VD, TOL-1 15. (k) *Krithe iniqua*, VS, TOL-1 15. (l) *Krithe perpulchra*, VD, TOL-1 3. (m) *Krithe compressa*, VD, TOL-1 15. (n) *Callistocythere cf. rastrifera*, VS, TOL-1 64. VD, valve in dorsal view; VS, valve in side (lateral) view. Scale bar: 0.1 mm.

Table S5. Autecological characteristics of the autochthonous ostracod species.

Species	Stratigraphic range	Bathymetry	Literature
<i>Acanthocythereis hystrix</i> (Reuss, 1850)	Langhian - Recent	50-1000 m	Aruta, 1983; Babinot and Boukli-Hacene, 1998; Bonaduce and Barra, 2002; Bonaduce et al., 1976; Dall'Antonia and Bossio, 2001; Faranda and Gliozzi, 2008; Miculan, 1992.
<i>Argilloecia acuminata</i> (Müller, 1894)	Serravallian - Recent	50-2500 m	Aiello et al., 2000; Aiello and Barra, 2001; Barra et al., 1996; Colalongo et al., 1981; Malz and Jellinek, 1984; Sciuto and Rosso, 2008
<i>Argilloecia fatua</i> (Barra, Aiello and Bonaduce, 1996)	Serravallian - Cal. (Sic.)	50-1000 m	Aiello et al., 2000; Aiello and Barra, 2001; Aiello et al., 2015; Ciampo, 1986.
<i>Argilloecia micra</i> (Bonaduce, Ciampo and Masoli, 1976)	Pliocene - Rec	170-1000 m	Barra et al., 1996; Bonaduce et al., 1976;
<i>Aurila convexa</i> (Baird, 1850)	Tortonian - Recent	0-122 m	Aiello et al., 2015; Barra, 1997; Breman, 1976; Bonaduce et al., 1976; Conesa and Babinot, 1999; Lachenal, 1989; Mostafawi, 1990; Nachite et al., 1993; Ruggieri, 1975, 1983; Zaggar and Malz, 1989;
<i>Aurila nebianii</i> (Ruggieri, 1953)	Gelasian - Recent	0-50 m	Aranki et al., 1992; Ruggeri, 1953, 1959, 1962, 1980.
<i>Bairdopplata conformis</i> (Terquem, 1878)	Gelasian (MP15b) - Recent	70-200(1000) m	Aiello et al., 2000; Aruta, 1983; Bonaduce and Barra, 2002; Conesa and Babinot, 1999; Colalongo 1966; Colalongo e Pasini 1980; Ruggieri, 1962; Terquem, 1878.
<i>Basslerites berchoni</i> (Brady, 1870)	Zanclean - Recent	0-50 m	Lachenal, 1989; Faranda and Gliozzi, 2008
<i>Bythocypris bosquetiana</i> (Brady, 1866)	Zanclean - Recent	500-1000 m	Aiello et al., 2000; Breman, 1975; Colalongo and Pasini, 1980.
<i>Bythocypris producta</i> (Seguenza, 1880)	Langhian- Chiban	500-1000 m	Aiello and Barra, 2001; Aiello et al., 2000; Bonaduce and Barra 2002; Colalongo et al., 1981; this paper
<i>Callistocythere rastrifera</i> (Ruggieri, 1953)	Piacenz. - Recent	inner infralittoral	Mostafawi, 1981; Ruiz et al., 1997.
<i>Carinocythereis carinata</i> (Roemer, 1838)	Langhian- Recent	0-104 m	Faranda and Gliozzi, 2008; Lachenal, 1989
<i>Caudites calceolatus</i> (Costa, 1853)	Tortonian - Recent	0-1 m	Arbulla et al., 2004; Lachenal, 1989
<i>Celtia biflexia</i> (Terquem, 1878)	Pliocene - Recent	0-50 m	Terquem, 1878; Wood, 2005
<i>Celtia quadridentata</i> (Baird, 1850)	Zanclean (Mpl1) – Calabrian (Sant.)	0-50 m	Faranda and Gliozzi, 2008; Wood, 2005
<i>Cistacythereis emaciata</i> (Brady, 1867)	Pliocene - Recent	50-100 m	Yassini, 1979; Wood, 2005.
<i>Cistacythereis rubra</i> (Müller, 1894)	Tortonian - Recent	0-18 m	Arbulla et al., 2004; Lachenal, 1989
<i>Costa edwardsii</i> (Roemer, 1838)	Langhian- Recent	13-120 m (40-60 FF)	Arbulla et al., 2004; Lachenal, 1989
<i>Cytherella robusta</i> (Colalongo and Pasini, 1980)	Langhian- Late Pleist.	300-1000 m	Aiello and Barra, 2001; Bonaduce and Barra 2002; Bonaduce and Sprovieri, 1985; Colalongo et al., 1981, 1990; Faranda et al., 2007. Pirkenseer et al., 2018.
<i>Cytheretta subradiosa</i> (Roemer, 1838)	Zanclean (MPI2) - Recent	0-60 m	Faranda and Gliozzi, 2008; Malz and Jellinek, 1984; Montenegro et al, 1998; Lachenal, 1989; Yassini, 1979;
<i>Echinocythereis (Rhodicythereis) pustulata</i> (Namias, 1900)	Zanclean (MPI1)- Calabrian (Sic)	30-40 m	Faranda and Gliozzi, 2008
<i>Eucytherura angulata</i> (Müller, 1894)	Tortonian - Recent	0-50 m	Aruta, 1983; Bonaduce et al., 1976; Ciampo, 1986; Miculan, 1992; Müller, 1894; Ruggieri, 1976.

<i>Grinioneis haidingeri</i> (Reuss, 1850)	Burdigalain - Chibanian	0-100 m	Aiello and Szczechura, 2004; Aranki et al., 1992; Brestenska and Jiricek, 1978; Conesa and Babinot, 1999; Faranda and Gliozzi, 2008; Faranda et al., 2007; Ruggieri, 1953, 1962, 1976a; Sissingh, 1972, 1973; this paper.
<i>Hemicytherura defiorei</i> (Ruggieri 1953)	Serravallian- Recent	0-100 m	Aiello and Szczechura, 2004; Brestenska and Jiricek, 1978; Aiello et al., 2015; Faranda et al., 2008; Lachenal, 1989
<i>Henryhowella sarsii profunda</i> (Bonaduce, Barra and Aiello, 1999)	Aquitanian - Recent	110-1000 m	Aiello and Barra, 2001; Aiello et al., 2000; Aiello et al., 2015; Bonaduce and Barra, 2002; Bonaduce and Sprovieri, 1985; Faranda et al., 2007; Sciuto, 2003; Sciuto and Rosso, 2008; Sciuto et al., 2003;
<i>Heterocythereis albomaculata</i> (Baird, 1850)	Piacenzian - Recent	10-70 m	Athersuch et al., 1989; Bassiouni, 1979; Faranda and Gliozzi 2008; Guillaume et al., 1985; Sciuto, 2003; Yassini, 1979;
<i>Krithe compressa</i> (Seguenza, 1880)	Zanclean-Late Pleistocene	160-1000 m	Aiello et al., 2000; Aiello et al., 2015; Bassetti et al., 2010; Barra et al., 1998; Ciampo, 1986; Ruggieri, 1991; Sciuto and Rosso, 2008;
<i>Krithe iniqua</i> (Abate, Barra, Aiello and Bonaduce, 1993)	Langhian- Chibanian	50-1000 m	Aiello et al., 2000; Aiello and Barra, 2001; Bonaduce and Barra, 2002; Ciampo, 1986; Sciuto and Rosso, 2008; this paper
<i>Krithe pernoides</i> (Bornemann, 1855)	Tortonian - Late Pleistocene	50-1000 m	Aiello et al., 1993; Aiello et al., 2000; Bassetti et al., 2010; Ciampo 1986;
<i>Krithe perpulchra</i> (Abate, Barra, Aiello and Bonaduce, 1993)	Piacenzian (MPI5a)- Chibanian	1000 m	Abate, et al., 1993; Aiello et al., 2000; this paper
<i>Loxoconcha ovulata</i> (Costa, 1853)	Late Tortonian- Recent	2-55 m	Arbulla et al., 2001, 2004; Babinot and Boukli-Hacene, 1998; Faranda and Gliozzi, 2008; Lachenal, 1989
<i>Macrocyprissa arcuata</i> (Colalongo e Pasini, 1980)	Calabarian- Recent	500- 1500 m	Colalongo and Pasini, 1980; Coles et al., 1996; Maddocks 1979, 1990;
<i>Microxestoleberis xenomis</i> (Barbeito-Gonzales, 1971)	Recent	50-110 m	Bonaduce et al., 1988
<i>Monoceratina mediterranea</i> (Sissingh, 1972)	Tortonian - Recent	circalitorale (>160 m)	Bonaduce et al., 1976; Ertekin and Tonoglu, 2008
<i>Paracytheridea depressa depressa</i> (Müller, 1894)	Pliocene- Recent	10-120 m	Bonaduce et al., 1976; Ciampo, 1976; Danatsas, 1994; Müller, 1894; Yassini, 1979; Zangger and Malz, 1989.
<i>Paracytheridea triquetra</i> (Reuss, 1850)	Aquitanian- Recent	8-26 m	Aiello and Szczechura, 2004; Babinot, 2002; Babinot and Boukli-Hacene, 1998; Faranda and Gliozzi, 2008; Mostafawi, 1989; Ruggieri, 1962;
<i>Palmoconcha turbida</i> (Müller, 1894)	Zanclean - Recent	20-210 m	Bonaduce et al., 1976; Lachenal, 1989; Yassini, 1980.
<i>Parahemingwayella tetrapterion</i> (Bonaduce, Ciampo and Masoli, 1976)	Late Tortonian- Recent	120-1000 m	Aiello et al., 2000; Bonaduce et al., 1976.
<i>Parakrithe ambigua</i> (Ciampo, 1980)	Piacenzian(MPI4)- Recent	1000 m	Aiello et al., 1993; Ciampo, 1980.
<i>Parakrithe rotundata</i> (Aiello, Barra, Abate and Bonaduce, 1993)	Langhian- Chibanian	500-1000 m	Aiello et al., 1993; Aiello et al., 2000; Bonaduce and Barra, 2002; this paper.
<i>Parakrithe dimorpha</i> (Bonaduce, Ciampo and Masoli, 1976)	Late Langhian - Recent	70-1000 m	Aiello et al., 1993; Bonaduce et al., 1976; Faranda and Gliozzi, 2008;
<i>Pseudopsammocythere reniformis</i> (Brady, 1868)	Late Pleistocene- Recent	0-25 m	Bonaduce et al., 1988; Zagger and Malz, 1989.
<i>Xestoleberis communis</i> (Müller, 1894)	Early Messinian-Recent	0-100 m	Babinot and Boukli-Hacene, 1998; Faranda and Gliozzi, 2008; Lachenal, 1989.
<i>Xestoleberis dispar</i> (Müller, 1894)	Zanclean - Recent	0-40 m	Aiello et al., 2020, 2021; Arbulla et al., 2004; Faranda and Gliozzi, 2008.

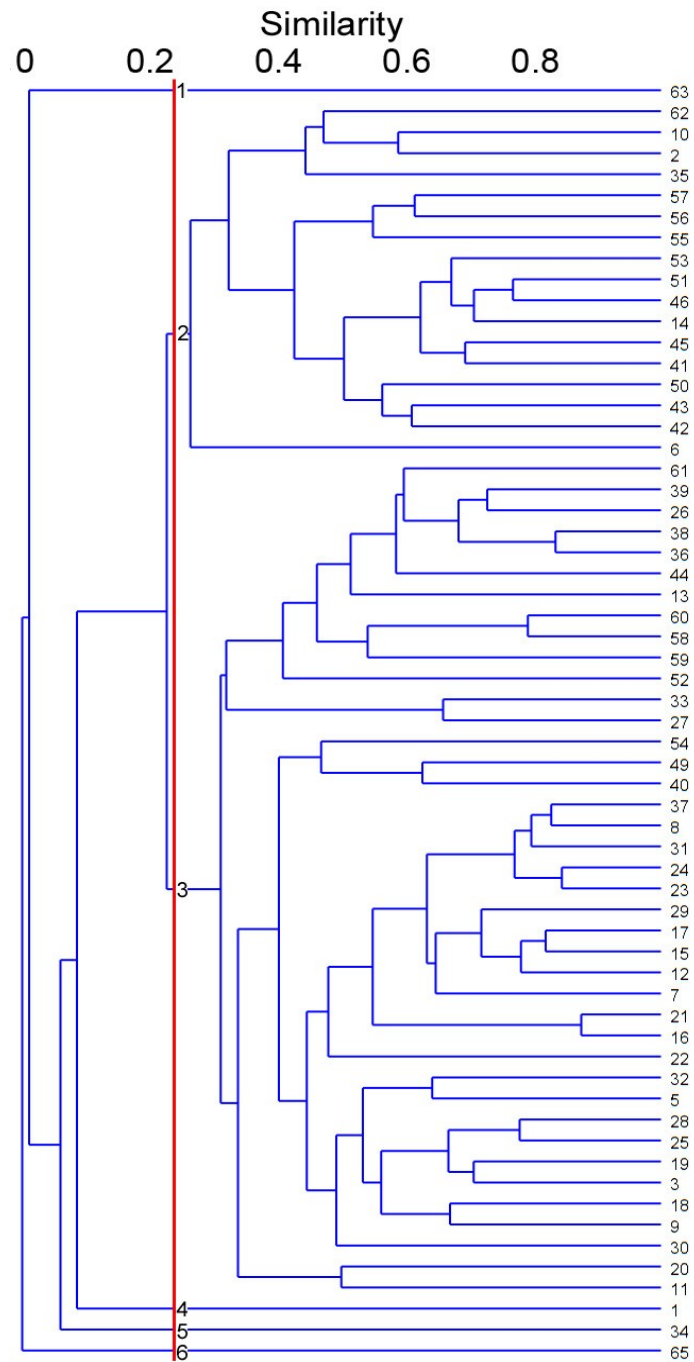


Figure S7. Dendrogram of the Cluster Analysis of ostracod assemblages in Q-mode applied on the samples of TOL-1 section.

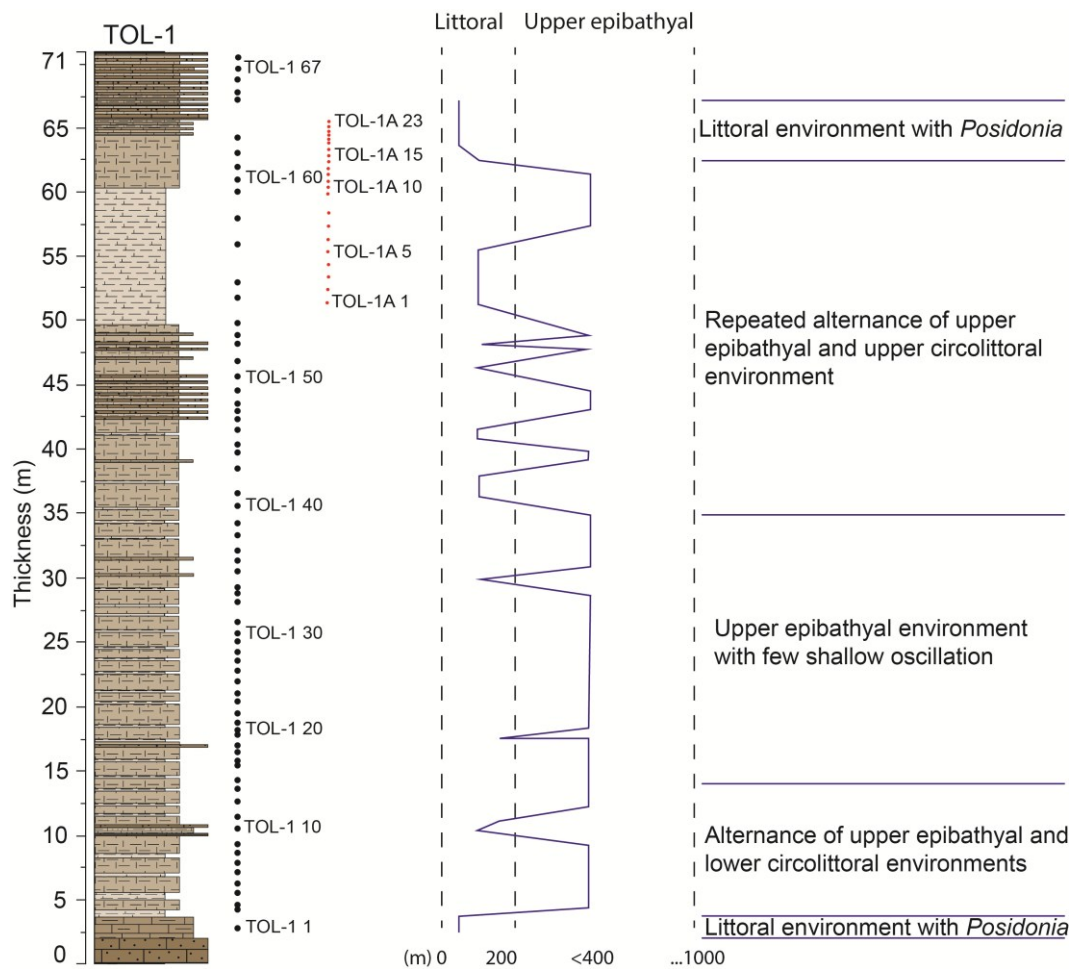


Figure S8. Palaeobathymetric reconstruction of the TOL-1 section based on ostracods.

Table S6. Oxygen-isotope ($\delta^{18}\text{O}$) stratigraphy of the TOL-1 section.

Thickness (cm)	$\delta^{18}\text{O}$
5170	-0.60
4140	-0.52
3035	-1.25
2865	-0.62
2800	-1.10
2640	-0.69
2490	-1.19
2260	-0.90
1800	-0.25
1760	-0.53
1625	-1.27
1520	-0.87
1240	-0.61
760	-0.26
690	-0.43
600	-0.59
400	-2.20

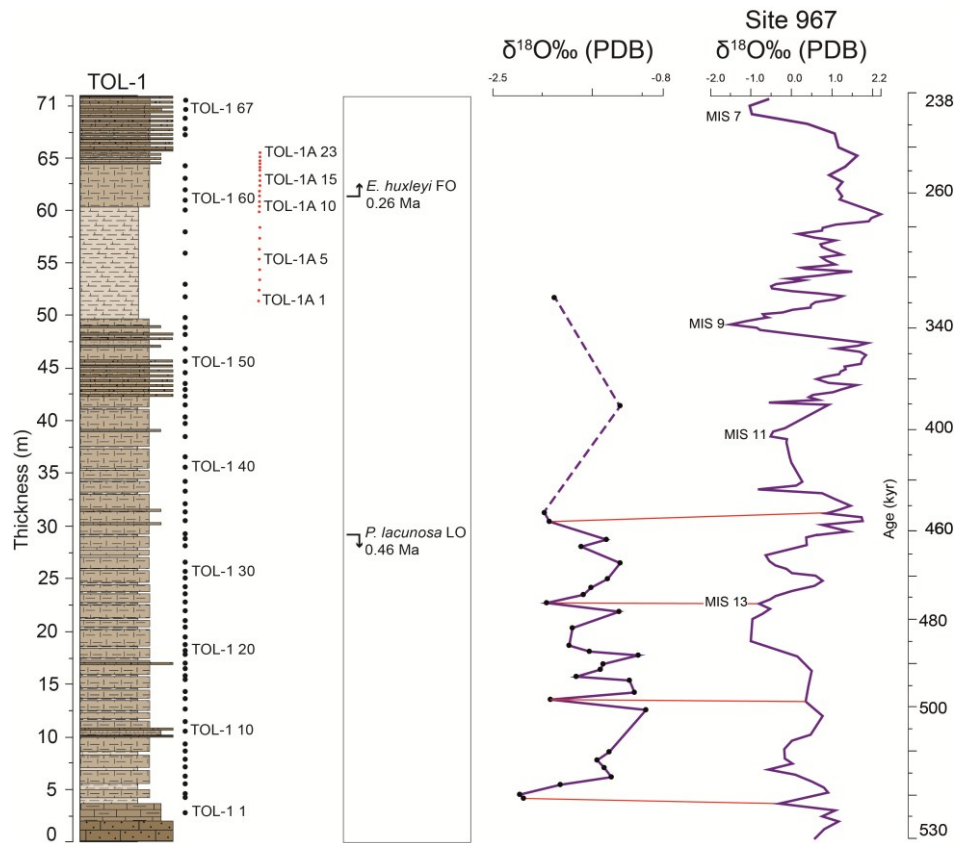


Figure S9. Stable oxygen isotope ($\delta^{18}\text{O}$) data from *Trilobatus trilobus* tests in the TOL-1 section compared with ODP Site 967 (Grant et al., 2022). Red lines indicate correlations between interglacial peaks, including MIS 13. Age-correlation tie points based on $\delta^{18}\text{O}$ are shown on the right-hand side of the curves.

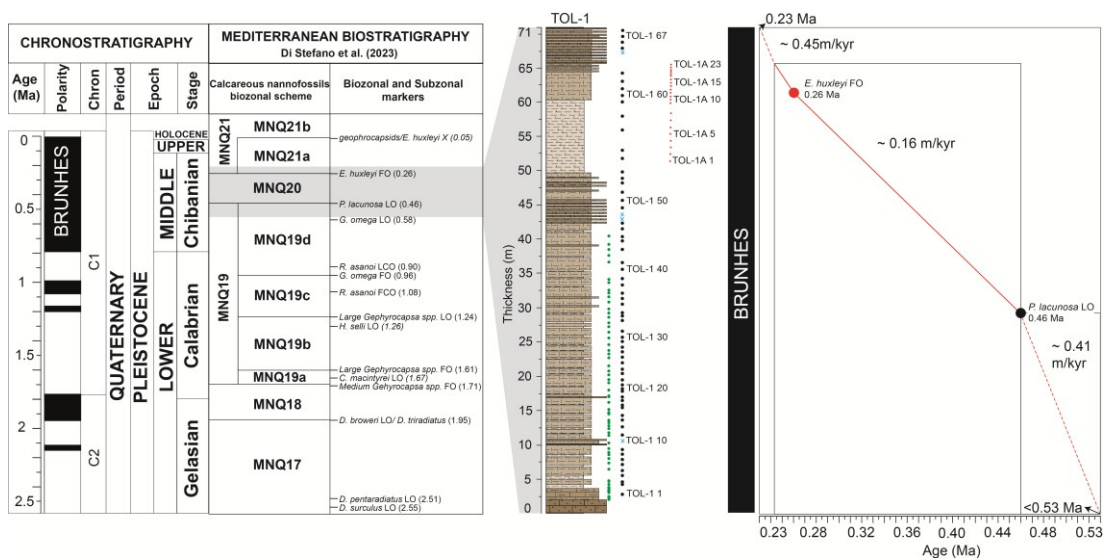


Figure S10. Integrated Mediterranean biostratigraphic scheme for the Pleistocene (Di Stefano et al., 2023), which includes astronomically calibrated magnetic reversal records. Key biostratigraphic markers include first common occurrences (FCO) and last common occurrences (LCO). The grey strip indicates the proposed age for the TOL-1 section. To the right, age constraints and sedimentation rates of the TOL-1 section as evidenced in this work.

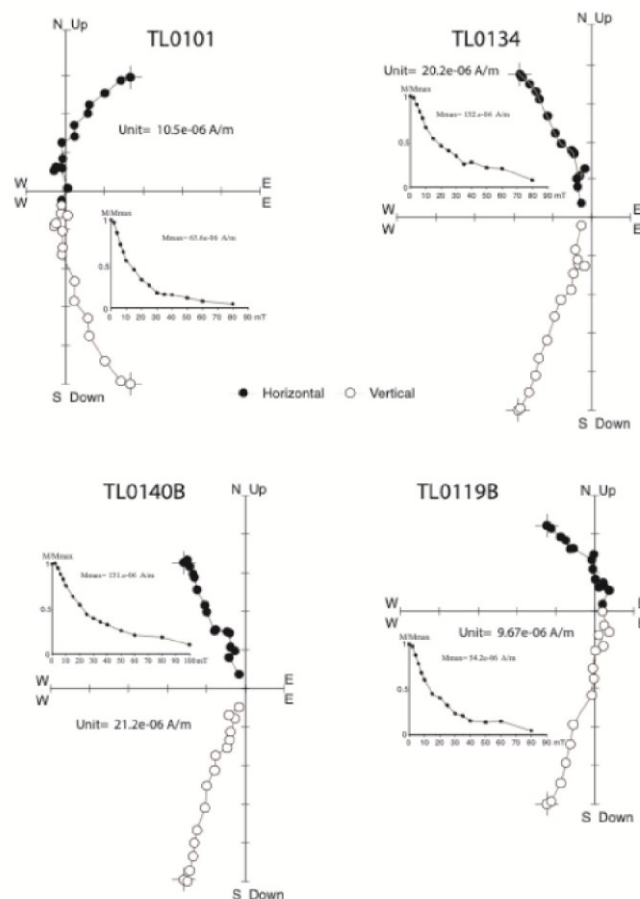


Figure S11. Orthogonal vector diagrams for the progressive alternating field (AF) demagnetization of representative samples. Open and solid symbols represent projections on the vertical and horizontal planes, respectively. N/UP, W, E, and S/DOWN indicate horizontal (north, south, west, and east) and vertical (up and down) projections combined into a single vector component diagram. For each sample, progressive AF demagnetization data in mT are reported.

Table S7. Global compilation of Marine Isotope Stage (MIS) 7 records from around the world, showing deposits at different elevations, dating methods, and tectonic contexts. Abbreviations: MIS = Marine Isotope Stage; OSL = Optically Stimulated Luminescence; U-Th = Uranium-Thorium dating; AAR = Amino Acid Racemization; ESR = Electron Spin Resonance; TL = Thermoluminescence; IRSL = Infrared Stimulated Luminescence; bsf = below sea floor; TR = Turkey; TN = Tunisia; IT = Italy; GB = Great Britain; ROK = Republic of Korea; BRA = Brazil; USA, HI = United States of America, Hawaii; UP = Upper plate; LP = Lower plate

Locality	MIS	Method for dating	Elevation (m)	Tectonic context	Reference
CAP N-margin (TR)	7	inferred	58	Passive margin	Berndt et al. (2018)
Hampshire (UK)	7b-7c	OSL	10	Passive margin	Bates et al. (2010)
Cap Bon (TN)	7	inferred	15	Passive margin	Elmejdoub and Jedoui (2009)
South Carolina (USA)	7	U-Th, OSL	10.7	Passive margin	Sanders et al. (2009); Doar III and Kendall (2017)
Argentarola Cave (IT)	7	U-Th	-9/-18	Passive margin	Bard et al. (2002)
Morston (GB)	7/6	OSL	5	Passive margin	Hoare et al. (2009)
Central East Coast of South Korea (ROK)	7	Cosmogenic ^{10}Be	60–80	Passive margin	Lee et al. (2015); Hong et al 2024
Central East Coast of South Korea (ROK)	7	OSL	40	Passive margin	Ko et al. (2009)
Central East Coast of South Korea	7	OSL	32-55	Passive margin	Kim et al. (2007)

(ROK)					
Rio Grande do Sul (BRA)	7e	TL + ESR	10	Passive margin	Lopes et al. (2014)
Bahamas	7c, 7e	U-Th	-10 to -15	Passive margin	Li et al. (1989)
Bahamas, Bermuda	7a, 7e	U-Th, AAR	2.5 to 0	Passive margin	Hearty and Kindler (1995)
Australia	7a-c	TL	4-2	Passive margin	Roy et al. (1997)
Spain	7e	U-Th	5	Passive margin	Vesica et al. (2000)
NE Brazil	7c	TL, OSL	5-3	Passive margin	Barreto et al. (2002)
Bahamas	7a-e	U-Th	12-20 m bsf	Passive margin	Robinson et al. (2002)
Australia	7a	TL, AAR	8	Passive margin	Murray-Wallace (2002)
Alicante, Spain	7e	U-Th	4.5-1.5	Passive margin	Goy et al. (2006)
Argentatola Cave, Italy	7a, 7c, 7e	U-Th	-18/-21	Passive margin	Dutton et al. (2009)
NE Brazil	7c	OSL, TL	>4 to >1.7	Passive margin	Suguio et al. (2011)
Patagonia	7	inferred	35-26	Passive margin	Pedoja et al. (2011)
Patagonia Pto Piramide	7	inferred	29	Passive margin	Schellmann and Radtke (2000)
Patagonia Caleta Valdez	7	U-Th	32	Passive margin	Codignotto et al. (1992)
Patagonia Bahia Camarones	7	U-Th, Paleontology	24.5	Passive margin	Rostami et al. (2000)
Patagonia Bahia Bustamente	7	ESR, U-Th	26	Passive margin	Rostami et al. (2000)
Patagonia Caleta Olivia	7	ESR, U-Th	32	Passive margin	Codignotto et al. (1992)
South China Sea	7	inferred	40-22	Passive margin	Pedoja et al. (2008)
Bermuda	7e	U-Th	6-4.5	Passive margin	Rowe et al. (2014)
Oahu (USA, HI)	7	U-Th	-10/-20	Volcanic Arc	Sherman et al. (1999); Mush and Szabo (1994)
Muroto Peninsula Japan	7	Tephrochronology	260	Island Arc UP	Matsu'ura (2015)
Sumba Island, Indonesia	7e	ESR, U-Th	51	Island Arc Subduction UP	Pirazzoli et al. (1991)
Barbados	7a, 7e	U-Th	78-92	Island Arc Subduction UP	Gallup et al. (1994)
Kendal Hill, Aberdare, Barbados	7a, 7c	ESR, U-Th	52-47 70-65	Island Arc Subduction UP	Schellmann et al. (2004)
Kish Island Persian Gulf	7	U-Th, ESR	1.5-6	Subduction LP	Pirazzoli et al. (2004)
Kish Island Persian Gulf	7	ESR	25.5-28	Subduction LP	Pirazzoli et al. (2004)
Taranto (IT)	7e	AAR	73	Subduction LP	De Santis et al. (2018)
Pan de Azucar, Chile	7	Cosmogenic ¹⁰ Be	64	Subduction UP	Martinod et al. (2016)
Calabrian Arc (IT)	7-9?	AAR + IRSL	225	Subduction UP	Belluomini et al. (1988); Nalin et al. (2020)
Bahia Iglesia, Chile	7	ESR	78	Subduction UP	Radtke (2020?)
Caleta Chanaral, Chile	7	Cosmogenic ¹⁰ Be	54	Subduction UP	Martinod et al. (2016)
Altos de Talinay, Chile	7	Cosmogenic ¹⁰ Be	55	Subduction UP	Saillard et al. (2009)
Caldera, Chile	7	inferred	44	Subduction UP	Marquardt et al. (2004)
Bahia Iglesia, Chile	7	inferred	40	Subduction UP	Marquardt et al. (2004)
Hornitos, Chile	7	AAR	60	Subduction UP	Ortlieb et al. (1996)
Kamikita Plain, NE Honshu (Japan)	7c-7e	Tephra and phytolith stratigraphy	50	Subduction UP	Matsu'ura et al. (2019)
Makran Coast (SE Iran and SW Pakistan)	7	Stratigraphy, terrace morphology	41-72	Subduction UP	Normand et al., 2018
Altos de Talinay, Chile	7	Cosmogenic ¹⁰ Be	30	Subduction UP	Martinod et al. (2016)
W and SE Cyprus	7	U-Th	8-11	Subduction UP	Poole et al., 1990
NE Cyprus	7	U-Th	40	Subduction UP	Palamakumbura et al., 2016

2. TECTONO-SEDIMENTARY EVOLUTION AND UPLIFT OF THE ANTALYA BASIN (SOUTHERN TURKEY) DURING THE LATE MIDDLE PLEISTOCENE: INSIGHTS FROM THE MANAVGAT FAN-DELTA COMPLEX

ABSTRACT

The Antalya Basin occupies within the Isparta Angle (IA), where the junction of the Aegean and Cyprus subduction systems produces a complex interaction of extensional and compressional regimes that strongly influences sedimentation patterns in southwestern Turkey. The Manavgat Sub-basin, at the eastern margin of the Antalya Basin, preserves Quaternary deltaic successions that record the interplay of tectonics, sediment supply, and sea-level fluctuations. To refine the stratigraphy of the Manavgat Sub-basin and improve our understanding on the tectono-sedimentary evolution of the Antalya Basin, we present new lithostratigraphic and micropalaeontological data from a composite section (MNV) of the Manavgat Formation. The MNV composite section records deposition within a deltaic system comprising upper delta-front, lower delta-front, and renewed upper delta-front facies, reflecting alternating high- and low-energy shallow-marine conditions. Calcareous nannofossil and planktonic foraminiferal assemblages indicate that the base of the MNV section is upper Middle Pleistocene, younger than ~0.26 Ma based on the presence of *Emiliania huxleyi*. These biostratigraphic constraints extend the age of marine deposition in the Manavgat Sub-basin into the late Middle Pleistocene, which is significantly younger than previously reported Miocene or Lower Pleistocene marine deposits. This revised regional chronology for the Antalya Basin allow us to define a mean uplift rate of ~0.8 m/kyr since the late Middle Pleistocene. This record offers new insights into the tectono-sedimentary evolution of the Antalya Basin (Isparta Angle), at the junction between the southwestern margin of the Central Anatolian Plateau and the Western Anatolian Extensional Province.

2.1. INTRODUCTION

The Tauride orogenic belt in southern Turkey forms part of the Alpine–Himalayan system (Fig. 1) and preserves a long-lived record of Paleozoic–Cenozoic subduction,

collision, and post-orogenic deformation (Okay et al., 2010; Parlak et al., 2016; Schmid et al., 2020). The western and central sectors of this belt converge within the Isparta Angle (IA), which is a transitional zone that links the Hellenic Arc, to the west, with the Cyprus Arc, to the east. The Isparta Angle represents one of the most structurally complex region in the Eastern Mediterranean (Blumenthal 1951; Özgül, 1976; Glover and Robertson, 1998; Deynoux et al., 2005; Kelling et al., 2005; Hall et al., 2014; Över et al., 2016).

The Isparta Angle is bounded by the Fethiye-Burdur Fault Zone to the west and the Akşehir Fault Zone to the east (Fig. 1). The Fethiye-Burdur Fault Zone is a major strike-slip fault that extends from the Fethiye Bay to Eğirdir Lake and features transtensional movement (e.g., Kocyiğit, 2000; Bozkurt, 2001; Tiryakioğlu et al., 2013; Över et al., 2016). Today, the IA is affected by an extensional tectonic regime, with widespread seismicity along its major active faults (Fig. 2).

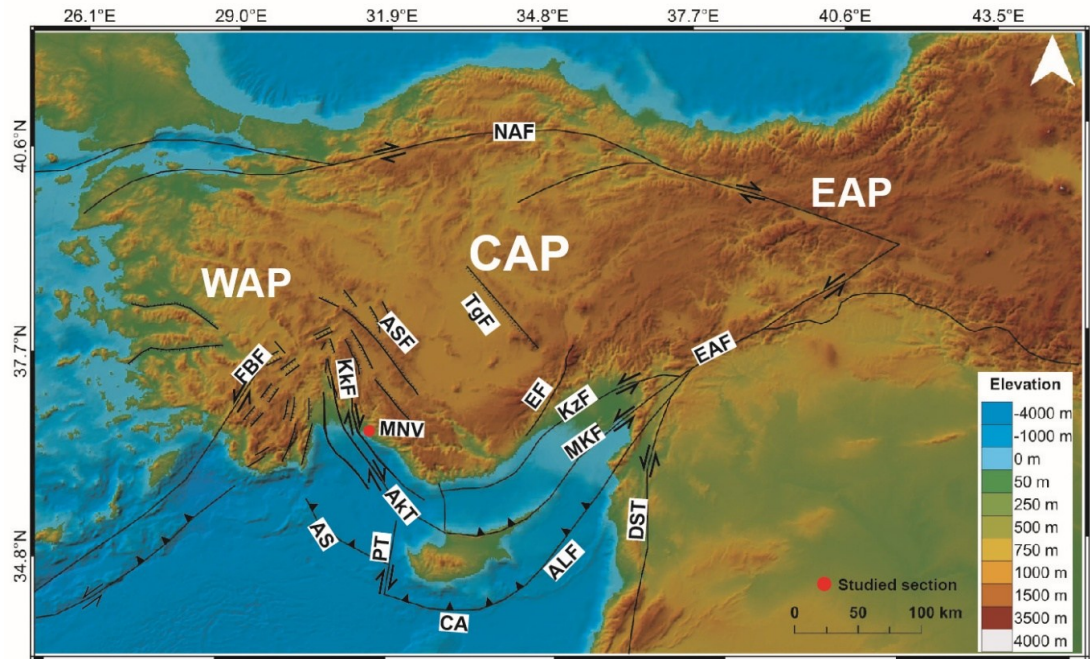


Figure 1. Simplified regional topography and tectonic map of Anatolia showing the location of the studied section (MNV). CAP, Central Anatolian Plateau; EAP, Eastern Anatolian Plateau; WAP, Western Anatolian Province; NAF, North Anatolian Fault; EAF, East Anatolian Fault; DST, Death Sea Transform Fault; CA, Cyprus Arc; PT, Paphos Transform; AS, Antalya Slab; ALF, Amos-Larnaka Fault; AkT, Aksu Thrust; KkF, Kırkkavak Fault; EF, Ecemiş Fault; KzF, Kozan Fault; MKF, Misis-Kyrenia Fault; TgF, Tuz Gölü Fault, FBF Fethiye-Burdur Fault; ASF, Akşehir Fault.

The IA recorded alternating extensional and compressional tectonic regimes during the Neogene–Quaternary. Stratigraphic, structural, and paleostress data define three major tectonic phases: 1) pre-Tortonian extensional phase; 2) late Tortonian–Early

Pliocene compressional phase; and 3) Late Pliocene-Quaternary extensional phase (e.g., [Poisson et al., 2003a, 2003b, 2011](#); [Hall et al., 2014](#); [Koç et al., 2016a, 2016b, 2017](#); [McPhee et al., 2018, 2022](#); [Wasoo et al., 2020](#)).

The first tectonic phase, prior to the Tortonian, corresponds to E–W-directed extension that initiated the early Antalya Basin and coeval rifting in the Yalvaç, Altınapa, and Ilgın continental basins (Fig. 2). In the late Tortonian-early Messinian, a transition to compressional tectonics produced the Lycian and Aksu deformation phases along the Aksu Thrust. The Lycian phase (~11–7 Ma) generated E–W contraction within the IA core, whereas the Aksu phase (Messinian–Early Pliocene) involved NE–SW to NW–SE shortening and open folding, interpreted to result from gravitational sliding of the Lycian Nappes and/or differential uplift and subsidence across the IA ([van Hinsbergen et al., 2010](#)). A third phase began in the Late Pliocene and persists to the present day, characterized by renewed N–S and E–W-directed extension ([Glover and Robertson 1998](#); [Aykut et al., 2023, 2025, 2026](#)), coeval with Middle Pleistocene uplift along the southeastern IA.

The pre-Tortonian extensional phase that affected the IA was responsible for the development of the Antalya Basin, which contain thick Neogene marine clastic and carbonate successions ([Flecker et al., 1998, 2005](#); [Karabıyıkoglu et al., 2000, 2005](#); [Poisson et al., 2003](#); [Deynoux et al., 2005](#); [Çiner et al., 2008](#)). The Antalya Basin has been extensively investigated for its Late Cenozoic evolution, deformation history, and deltaic basin-fill architecture ([Akay et al., 1985](#); [Flecker et al., 1998, 2005](#); [Glover, 1995](#); [Glover and Robertson, 1998](#); [Karabıyıkoglu et al., 2000, 2005](#); [Poisson et al., 2003](#); [Deynoux et al., 2005](#)).

Recent studies have further refined its stratigraphy, emphasizing the interplay between sediment supply, relative sea-level change, and tectonic activity in controlling deltaic progradation and basin evolution ([Sagular and Özer, 2024](#); [İlgar et al., 2024](#); [Larsen et al., 2024a, 2024b](#); [Wathne et al., 2024](#); [Yılmaz et al., 2024](#)).

Many earlier studies of the Antalya Basin assigned the majority of the marine deposits to the Miocene (e.g., [Flecker, 1995](#); [Glover and Robertson, 1998](#); [Karabıyıkoglu et al., 2000](#)). Only a few recent studies recognized younger marine deposits within the Antalya Basin. For instance, [Sagular and Özer \(2024\)](#) and [Kanbur and Öğretmen \(2022\)](#) identified Early Pleistocene marine deposits within Manavgat and Aksu sub-basins, respectively, suggesting a more prolonged marine influence in the Antalya Basin than previously thought.

As a consequence, the stratigraphy of the Antalya Basin is marked by conflicting age assignments and inconsistent formation boundaries. Particularly, one of its key successions, the Karpuzçay Fm within the Manavgat Sub-basin, has been variously interpreted as Langhian-Messinian ([Çiner et al., 2008](#)), Serravallian–Tortonian ([Yılmaz et al., 2024](#)), and Pliocene-Pleistocene ([Sagular and Özer, 2024](#)), reflecting major uncertainties in formation boundaries, depositional environments, and the timing of sedimentation. These inconsistencies have hindered a coherent understanding of the basin’s tectono-sedimentary evolution.

Our multi-proxy dataset tries to resolve these discrepancies by demonstrating that the Antalya Basin hosted marine depositional environments at least until the late Middle Pleistocene (late Chibanian), considerably younger than previously inferred. The exposed fan-delta succession of the Quaternary Manavgat Fm, in the Manavgat Sub-basin, provides an exceptional record for reconstructing its tectono-sedimentary evolution, close to the end of the existence of a marine environment in the Antalya Basin, before its forced regression as a consequence of the uplift of the region.

Through integrated facies and micropaleontological analyses, we intend to (i) refine the Quaternary stratigraphy, facies architecture, and depositional history of the sub-basin; (ii) establish a revised chronostratigraphic framework based on calcareous nannofossil and foraminiferal assemblages; (iii) quantify the timing and rate of post-depositional uplift; (iv) propose geodynamic mechanisms that may explain the tectono-sedimentary evolution of the sub-basin.

Collectively, our results improve our understanding of the late Quaternary tectono-sedimentary evolution of the Antalya Basin and the broader IA and provide new constraints on the geodynamics of the Eastern Mediterranean.

2. 1.1. Geological setting and stratigraphy

The southwestern margin of the CAP is underlain by Mesozoic autochthonous carbonates of the Geyikdağı Unit ([Özgül, 1976](#); [Akay and Uysal, 1988](#); [Schmid et al., 2020](#)). These are tectonically overlain by the allochthonous Beyşehir-Hoyran, Antalya, and Alanya nappes, which consist of ophiolitic mélanges, metamorphic rocks, platform carbonates, and Eocene carbonate–clastic sequences ([Özgül, 1976](#); [Şengör and Yılmaz, 1981](#)). The Antalya Basin developed unconformably atop these nappes, within the IA, during the Late Cenozoic.

The Antalya Basin is bounded by two major fault zones: the Fethiye-Burdur Fault Zone, to the west, and the Sultandağı-Akşehir Fault Zone, to the east (Fig. 2). Moreover, within the Antalya Basin, the activity of other tectonic elements, such as the N–S-trending Kırkkavak Fault (KkF) and the W- to SW-verging Aksu Thrust (AkT), have led to its division into three sub-basins (Fig. 2): Aksu, Köprüçay, and Manavgat (Dumont and Kerey, 1975; Akay et al., 1985; Monod et al., 2006; Çiner et al., 2008; Poisson et al., 2011; Hall et al., 2014). The Kırkkavak Fault, originally a paleotectonic structure, was reactivated as a normal fault during the Quaternary, reflecting a regional transition to extensional tectonics linked to southward retreat of the subducting Neo-Tethyan slab (Doğan et al., 2017; Koçyiğit et al., 2024; Aykut et al., 2023, 2025, 2026).

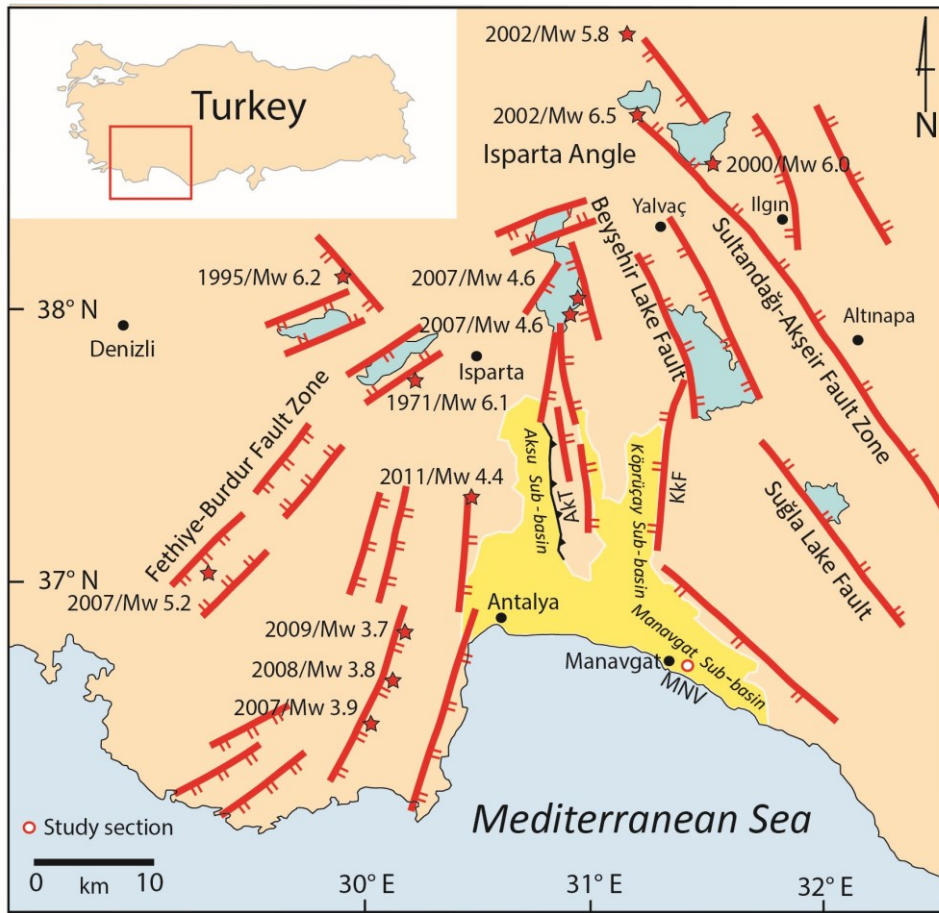


Figure 2. Seismotectonic setting of the Isparta Angle, Eastern Mediterranean region (modified from Över et al., 2016), showing the extension of the Neogene Antalya Basin (in yellow) and relative sub-basins (Aksu, Köprüçay, and Manavgat). Red stars indicate location and year of earthquakes with $M_w > 3.5$. KkF – Kırkkavak Fault; AkT – Aksu Thrust.

The stratigraphy of the Manavgat Sub-basin has been documented in detail by numerous studies, and a total of eight main formations have been recognized, spanning from the Lower Miocene to the Quaternary (Fig. 3).

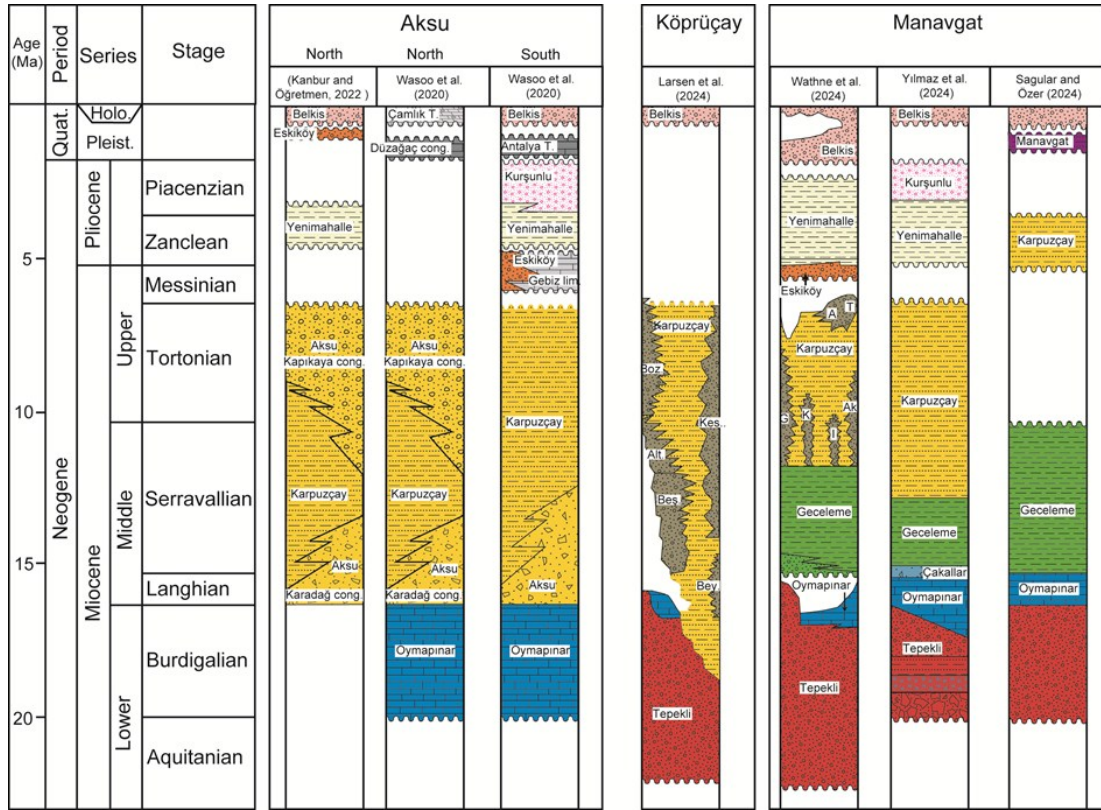


Figure 3. Lithostratigraphic framework of three major sub-basins of the Antalya Basin (Aksu, Köprüçay, and Manavgat) during the Neogene–Quaternary. The figure summarizes the documented lithostratigraphic units, their age assignments, and the major depositional systems recorded across the basin. Delta complexes are labelled using the following abbreviations: Beydiğın (Bey), Altınkaya (Al), Beşkonak (Bes), Bozburun (Boz), Kesme (Kes), Göktaş (G), Katırınemeği (K), İhtiyarlı (I), Akseki (Ak), and Taşlık (T), Asar (A).

The lowermost unit, the Tepekli Fm (Burdigalian), consists predominantly of alluvial-fan and fan-delta conglomerates (Karabıykoğlu et al., 2000; Çiner et al., 2008). It is overlain by the Oymapınar Fm (Upper Burdigalian–Langhian), which is composed of reefal limestones (Flecker et al., 1998; Karabıykoğlu et al., 2000; İslamoğlu, 2002; Çiner et al., 2008). Above this, the Çakallar Formation (Langhian) comprises fore-reef talus deposits (Akay et al., 1985). The overlying Geceleme Fm (Upper Langhian–Serravallian) consists of gray claystone and marl rich in planktonic foraminifers, interbedded with thin sandstone layers (Akay et al., 1985; Karabıykoğlu et al., 2000; İslamoğlu, 2002; Çiner et al., 2008). The Karpuzçay Fm (Serravallian–Tortonian) conformably overlies the Geceleme Fm and is characterized by alternating mudstone, siltstone, sandstone, and conglomerate (Akay et al., 1985; Çiner et al., 2008; Karabıykoğlu et al., 2005; Yılmaz et al., 2024). The Yenimahalle and Kurşunlu Fms (Pliocene) consist of alternating limestone–claystone–sandstone

and conglomerate–sandstone–mudstone successions, respectively (Yılmaz et al., 2024).

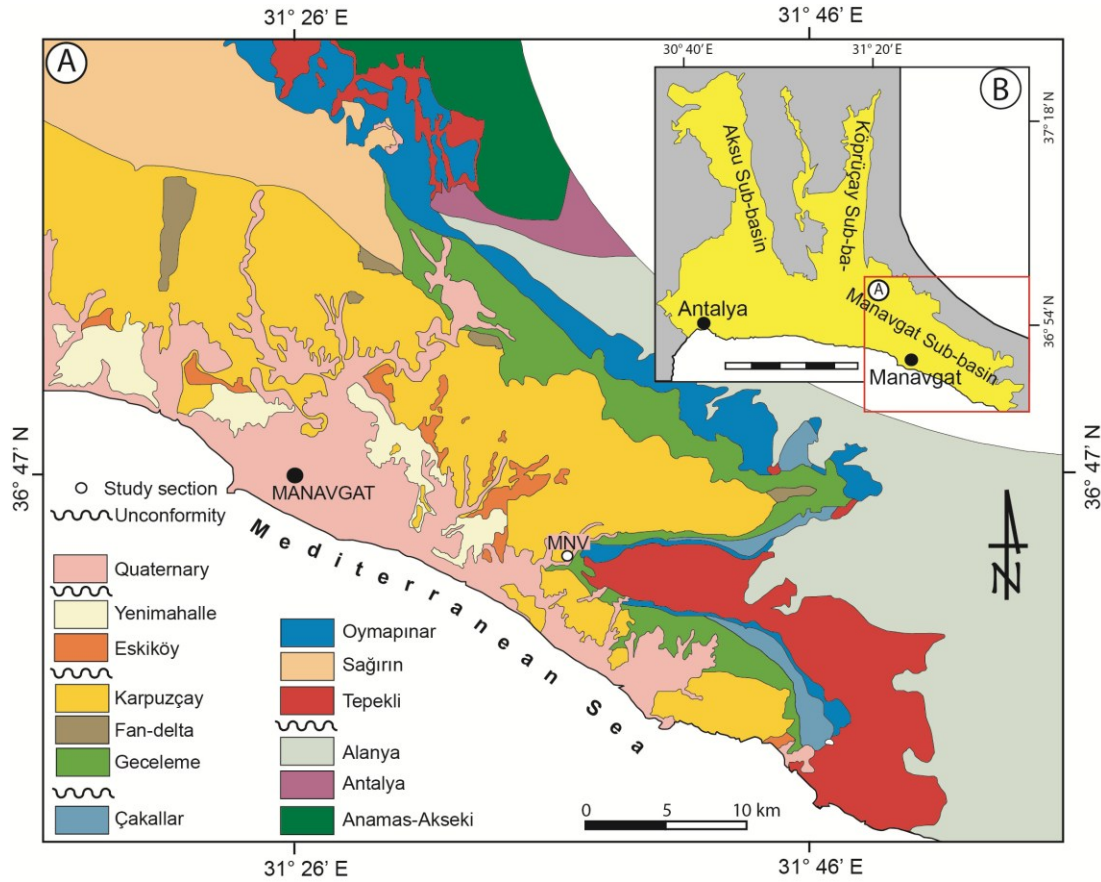


Figure 4. (A) Geological map of the Manavgat Sub-basin (modified after Wathne et al., 2024) showing Neogene and Quaternary sedimentary units, with locations of the studied composite sections (MNV). (B) Simplified geological map of the Antalya Basin, showing its three sub-basins, in yellow, and the bedrock successions, in grey. The red rectangle shows the position of Fig. 4A within the Antalya Basin.

In the Manavgat Sub-basin (Fig. 4), in contrast to these previous attributions, Sagular and Özer (2024) consider the lower part of the Karpuzçay Fm of Zanclean age, due to the occurrence of calcareous nannofossils attributable to the *Reticulofenestra pseudumbilica* zone (NN15, late Early Pliocene), while defining a new formation (Manavgat Fm) to indicate the youngest portion of the Karpuzçay Fm in which they recognized calcareous nannofossils attributable to the *Gephyrocapsa caribbeanica* subzone (NN19a, Early Pleistocene). The Manavgat Fm is composed of shallow marine and coastal clastic deposits (Sagular and Özer, 2024). The Belkis Fm (Quaternary) consist of fluvial conglomerates exposed mainly along the southern parts of the basin, near the present-day coastline (Yılmaz et al., 2024)

From a comparison among the different stratigraphies suggested by various authors for the Antalya Basin (Fig. 3), the main difference that emerges is the temporal

attribution of formations such as the Karpuzçay Fm (Sagular and Özer, 2024) and the Eskiköy Fm (Kanbur and Öğretmen, 2022). These new chronostratigraphic attributions proposed by Kanbur and Öğretmen (2022) for the Eskiköy Fm and by Sagular and Özer (2024) for the Karpuzçay Fm derive from the discovery of planktonic foraminifera and calcareous nannofossils referable to biozones that are much younger than previously thought for the same formations (e.g., Akay et al., 1985; Çiner et al., 2008; Karabıyıkoglu et al., 2005; Yılmaz et al., 2024).

These new findings call into question the chronology of the tectono-sedimentary evolution proposed by various authors who have studied the Antalya Basin so far, significantly rejuvenating marine deposits previously considered to be from Middle to Upper Miocene.

The Karpuzçay Fm and the Manavgat Fm (as defined by Sagular and Özer, 2024) contain the deltaic successions targeted in the present study.

2.2. MATERIAL AND METHODS

The present study focuses on the analysis of three stratigraphic sections, which we here combined into one composite, located in the Manavgat Sub-basin near the Uzunlar neighbourhood of Antalya, southwestern Turkey. According to previous studies on the Manavgat Sub-basin, our composite section should fall within the Karpuzçay Fm (Serravallian-Tortonian; Akay et al., 1985; Çiner et al., 2008; Karabıyıkoglu et al., 2005; Yılmaz et al., 2024) and should contain the boundary between the Karpuzçay Fm (Lower Pliocene) and the Manavgat Fm (Lower Pleistocene according to the stratigraphic arrangement proposed by Sagular and Özer (2024) for the Manavgat Sub-basin.

The composite section integrates two continuous successions: MNV4 and MNV1–2 sections (Fig. 5). The MNV4 section (36°44'28.74"N, 31°37'21.07"E, base of section; 36°44'18.84"N; 31°36'52.87"E, top of section) is approximately 65 m thick and exposed at ~50 m above sea level (a.s.l.). It is stratigraphically overlain by the MNV1–2 succession (36°44'30.76"N, 31°36'50.77"E, base of section; 36°44'38.20"N; 31°37'6.03"E, top of section), which comprises an additional 35 m of strata. The top of the MNV2 interval reaches ~143 m a.s.l., and field work indicates an estimated covered gap of ~25 m between the MNV4 and MNV1-2 measured sections.

Based on lithological correlation, stratigraphic position, and lateral continuity, the MNV4 and MNV1–2 successions are interpreted as a single depositional sequence forming the MNV composite section. Marl layers were systematically sampled throughout the MNV composite section for the micropaleontological analyses. A total of 46 samples were collected, 25 from MNV4, 9 from MNV1, and 12 from MNV2.

To establish the chronological framework of the composite succession, quantitative analyses of calcareous nannofossils were performed using a concentrated preparation method ([De Capoa et al. 2004](#)). As the concentration procedure markedly increases the number of specimens per field of view, typically ~ ten times higher than in standard smear-slide preparations, a reduced number of fields is sufficient to capture the full assemblage.

Accordingly, two counting strategies were applied. For the general assemblage, ten representative fields of view were examined per slide, which provided a stable estimate of species abundances due to the high specimen density of the concentrated preparations. In contrast, to ensure that even the rarest forms of the assemblages are sampled (like *Emiliana huxleyi*) approximately 350 fields of view were scanned. Specimen frequencies were recorded as the number of individuals per mm². Biostratigraphic interpretations and age determinations were based on the standard Mediterranean calcareous nannofossil zonation of [Di Stefano et al. \(2023\)](#).

Qualitative foraminiferal biostratigraphic analyses were conducted to complement the nannofossil data and possibly refine the age model. All samples were wet-sieved through a 63-µm mesh using tap water, without any prior acid treatment, and residues were dried at room temperature. Sample preparation, identification of the species, and photographing some species were performed at the Süleyman Demirel University (SDU), Department of Geological Engineering and at SDU, Innovative Technologies Application and Research Centre facilities. The biostratigraphic interpretation and age determinations were based on comparison with the standard Mediterranean foraminiferal biostratigraphic zonation proposed by [Lirer et al. \(2019\)](#).

Although the same preparation procedure was applied to recover ostracods for paleobathymetric analysis, no ostracod specimens were recovered from any of the samples.

2.3. RESULTS

2.3.1. Lithostratigraphy of the MNV composite section

The MNV composite section is approximately 125 m thick, encompassing the integrated MNV4, MNV1 and MNV2 sections (Figs. 5 and 6). It exhibits a rhythmic alternation of marl, silty marl, fine- to coarse-grained sandstone, and conglomerate, interbedded with slumped horizons and soft-sediment deformation structures.

2.3.1.1. MNV4 section

The basal part (0–2.4 m; base of MNV4) consists predominantly of silty marls and includes a thin, distinct matrix-supported conglomerate bed with decimetric thickness and clast sizes range from millimetres to centimetres (Figs. 5A and 6F). The 2.4–13.4 m interval is dominated by clast-supported conglomerates with mm- to cm-sized clasts, a poorly sorted medium-grained sandstone matrix (~30% of the unit), and interbedded coarse-grained sandstones with crude lamination, fine-grained sandstones with parallel and cross-lamination (Fig. 6D), medium-grained sandstones with clay chips and sole marks, silty marls. A 2.5-m-thick slump horizon (Sl-1) occurs within this interval (Fig. 6E). It shows an erosional basal surface and is composed of chaotically deformed marls and carbonate pebbles.

Between 13.4 and 20.9 m, the succession consists mainly of finely laminated sandstones and clays, interbedded with thin marly layers. This interval is overlain by coarse-grained sandstones and clast-supported conglomerates (20.9–24.7 m), which is poorly to moderately sorted and contains ~ 25% of sandy matrix. From 24.7 to 44.7 m, the section alternates between fine-grained sandstones exhibiting parallel and cross-lamination, coarse-grained sandstones crudely laminated, and marls (Fig. 6C). A second major slump (Sl-2) occurs in the upper part of this interval (~7 m thick), consisting of chaotically deformed marls and pebbles, with shell fragments, sparse *bryozoan* colonies, and fragments of coral colonies (Fig. 6B). Between 44.7 and 47.5 m, deposits are dominated by clast-supported conglomerates interbedded with fine- to coarse-grained sandstones; the conglomerates display channel-like basal erosional surfaces. The uppermost part of the MNV4 section (47.5–65 m) consists of alternating fine- to coarse-grained sandstones, minor matrix-supported conglomerates (poorly sorted, mm–cm clasts), and marls (Fig. 6A).

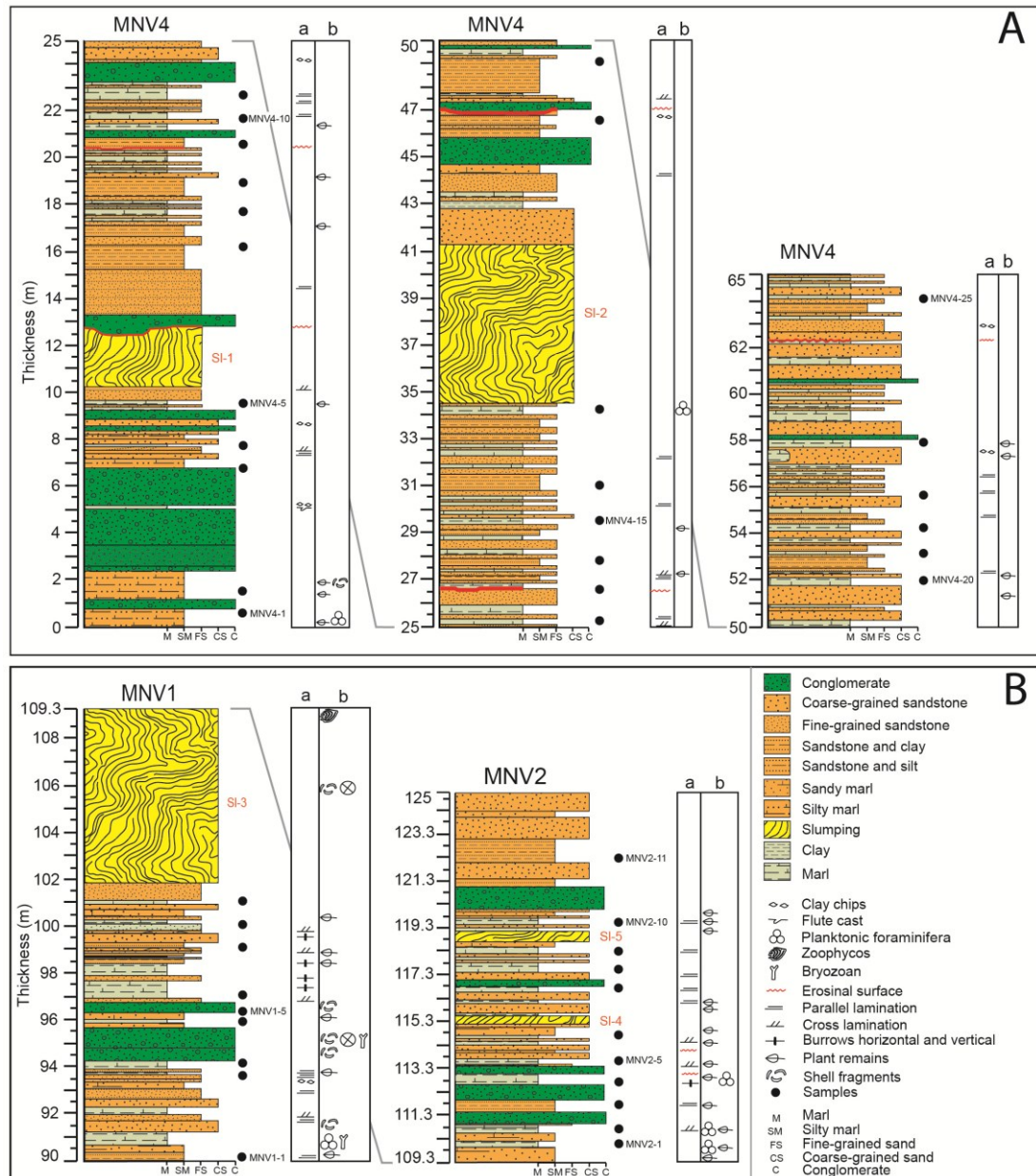
2.3.1.2. MNV1 and MNV2 sections

The MNV1–2 succession (Fig. 5B), which overlies the MNV4 section after an estimated ~25 m of covered and unmeasured section, spans from 90 to 125 m in the

composite log. The lower interval (90–94.25 m; MNV1) is composed mainly of marls and silty sandstones, and fine- to medium-grained sandstones with parallel and cross-lamination (Fig. 7H). This interval fines upward and is overlain by a clast-supported conglomerate (94.25–96.8 m) with a channel-like erosional base. Between 96.8 and 108.1 m, the succession is dominated by marls and silty marls, interbedded with fine- to medium-grained sandstones showing parallel and cross-lamination; brown to black organic-rich laminae are common, particularly between 96.8 and 101 m (Figs. 7D-F-G). This interval contains the third major slump horizon, Sl-3 (~7.5 m thick; Figs. 7A-B-C), composed of chaotic deposits formed of marls with oxidised horizons, sand, and conglomerate, with abundant shell fragments, bryozoans, and fragments of coral colonies. (Fig. 7C). In the MNV1 section, medium- to fine-grained sandstones with cross and parallel laminations are present.

The MNV2 section (109.3–125 m) continues above the Sl-3 slumped horizon of MNV1. The 110.9–117 m interval is dominated by clast-supported conglomerates with sandy matrix, interbedded with marls and silty sandstones, and fine- to medium-grained sandstones with crude parallel- and cross-lamination (Fig. 7A-B). The uppermost unit (117–125 m) consists of thick, coarse-grained sandstones with crude laminations, marking the top of the MNV composite section. Within MNV2, two thin slump layers (~0.5 m each) occur, composed of chaotic marls, sand, and pebble mixtures (Sl-4 and Sl-5; Fig. 5B).

Throughout the MNV composite section, marls are rich in planktonic foraminifera and, secondarily, benthonic foraminifera. Bioturbation structures are widespread, including *Zoophycos* sp., *Thalassinoides* sp., and *Planolites* sp. (Fig. 7E).



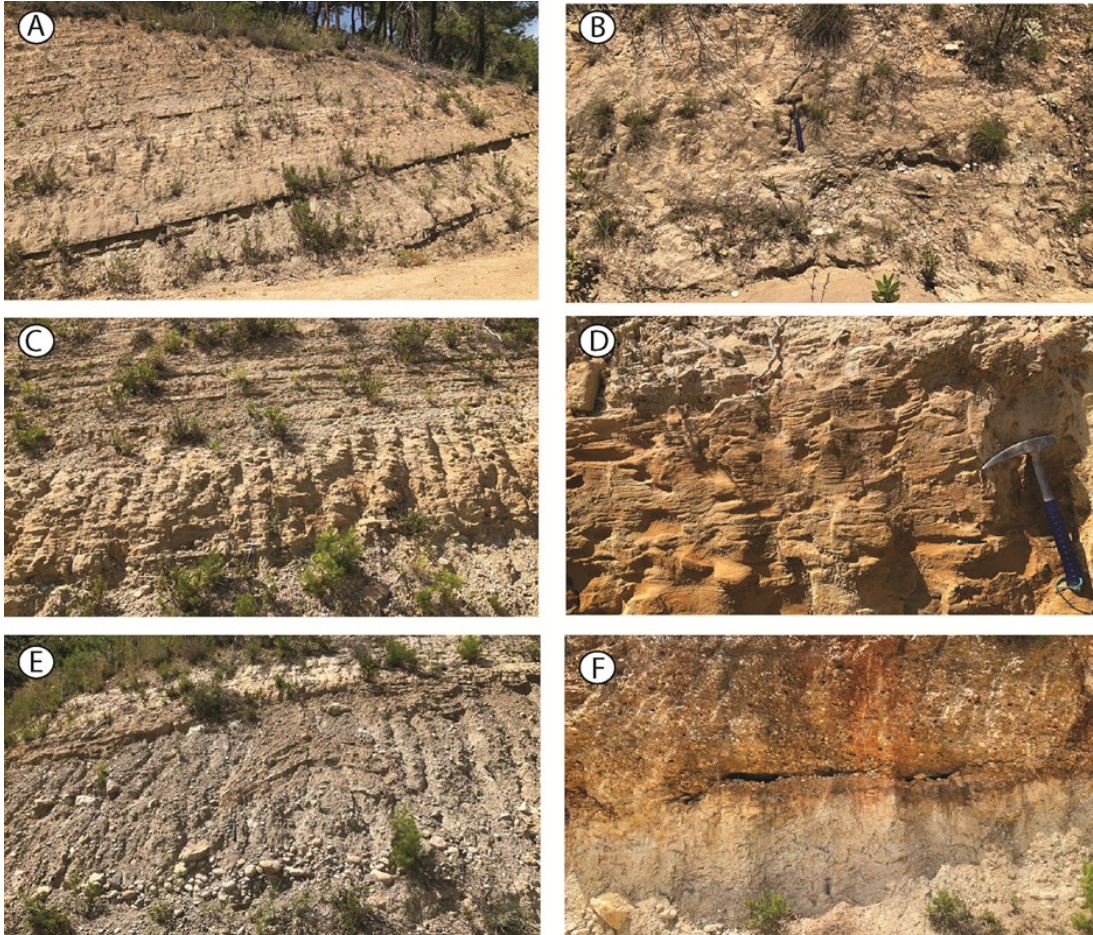


Figure 6. Field photographs showing representative physical sedimentary structures from top to bottom of the MNV4 section. (A) well-bedded alternations of marls and fine- to coarse-grained sandstones from the uppermost part of the section; (B) 7-m-thick slumped horizon (SL-2) from the middle portion of the section; (C) well-bedded alternation of conglomerates and coarse- to fine-grained sandstones from the lower part of the measured section; (D) cross-laminations (ripples) in fine-grained sandstones from the lower part of the measured section; (E) 2.5-m-thick slumped horizon (SL-1) from the lower part of the section; (F) lowermost portion of the section showing a disrupted marl horizon overlain by a thick conglomerate bed.

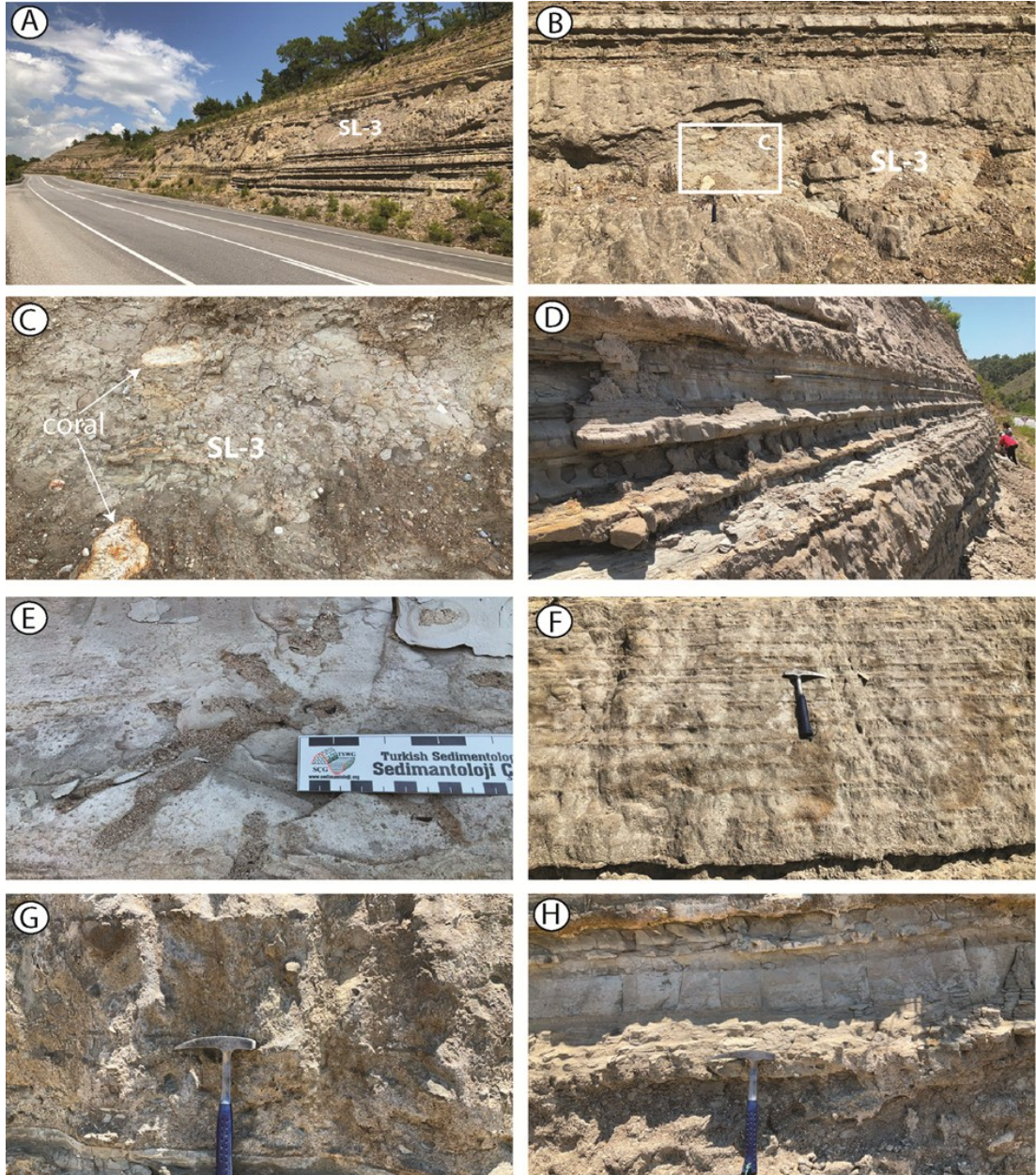


Figure 7. Field photographs showing representative physical sedimentary structures (from top to bottom) of the MNV1-2 composite section. (A) Panoramic view of the MNV1-2 composite section, along the D695 road (Manavgat-Konya road), showing the SL-3 slumped horizon; (B) topmost part of the SL-3 slumped horizon; (C) close-up of the SL-3 slumped horizon (see box in B) showing a muddy-supported conglomerate with fragments of coral colonies; (D) middle part of the MNV1 section showing well-bedded succession of marls, sandy marls, fine- and coarse-grained sandstones; (E) bioturbation (*Thalassinoides* sp.) in a marly horizon from the middle part of the MNV1 section; (F) coarse- to fine-grained sandstone bed from the lower part of the MNV1 section showing well-developed parallel-lamination; (G) sandy marl bed rich in clay chips from the lowermost part of the MNV1 section; (H) well-bedded lowermost portion of the MNV1 section showing alternations of marls, sandy marls, fine- and coarse-grained sandstones, and conglomerates.

2.3.2. Micropaleontological results

2.3.2.1. Calcareous nannofossils

Calcareous nannofossil assemblages from the MNV composite section are dominated by long-range and reworked taxa, whereas biostratigraphically significant species are rare (Figs. 8 and 9). The most common taxa include *Coccolithus pelagicus*, *Umbilicosphaera jafari*, *U. sibogae*, *Helicosphaera* spp., *Calcidiscus leptoporus*, *Pontosphaera* spp., *Syracosphaera* spp., *Calciosolenia* sp., and *Rhabdosphaera clavigera*. Small *Gephyrocapsa* specimens occur sporadically, while medium-sized *Gephyrocapsa* were not observed. Importantly, *Emiliana huxleyi* was identified throughout both MNV4 and MNV1-2 sections, although in low abundance. The first occurrence (FO) of *E. huxleyi* is dated to approximately 0.26 Ma in the Mediterranean region (Di Stefano et al., 2023), marking a key biostratigraphic boundary within the Middle Pleistocene. Therefore, its consistent presence within the studied sections indicates that deposition of the MNV composite section took place after ~0.26 Ma, placing the sedimentation of both sections in the late Middle Pleistocene.

Long-range and reworked species display similar vertical distribution patterns throughout the section suggesting they are largely reworked. In the MNV1 and MNV2 sections the relative abundance of reworked taxa decreases, contrasting with an increase in sedimentation rate. This pattern can be explained by considering that also many long-range species are reworked, since they increase during the same stratigraphic interval (Fig. 8).

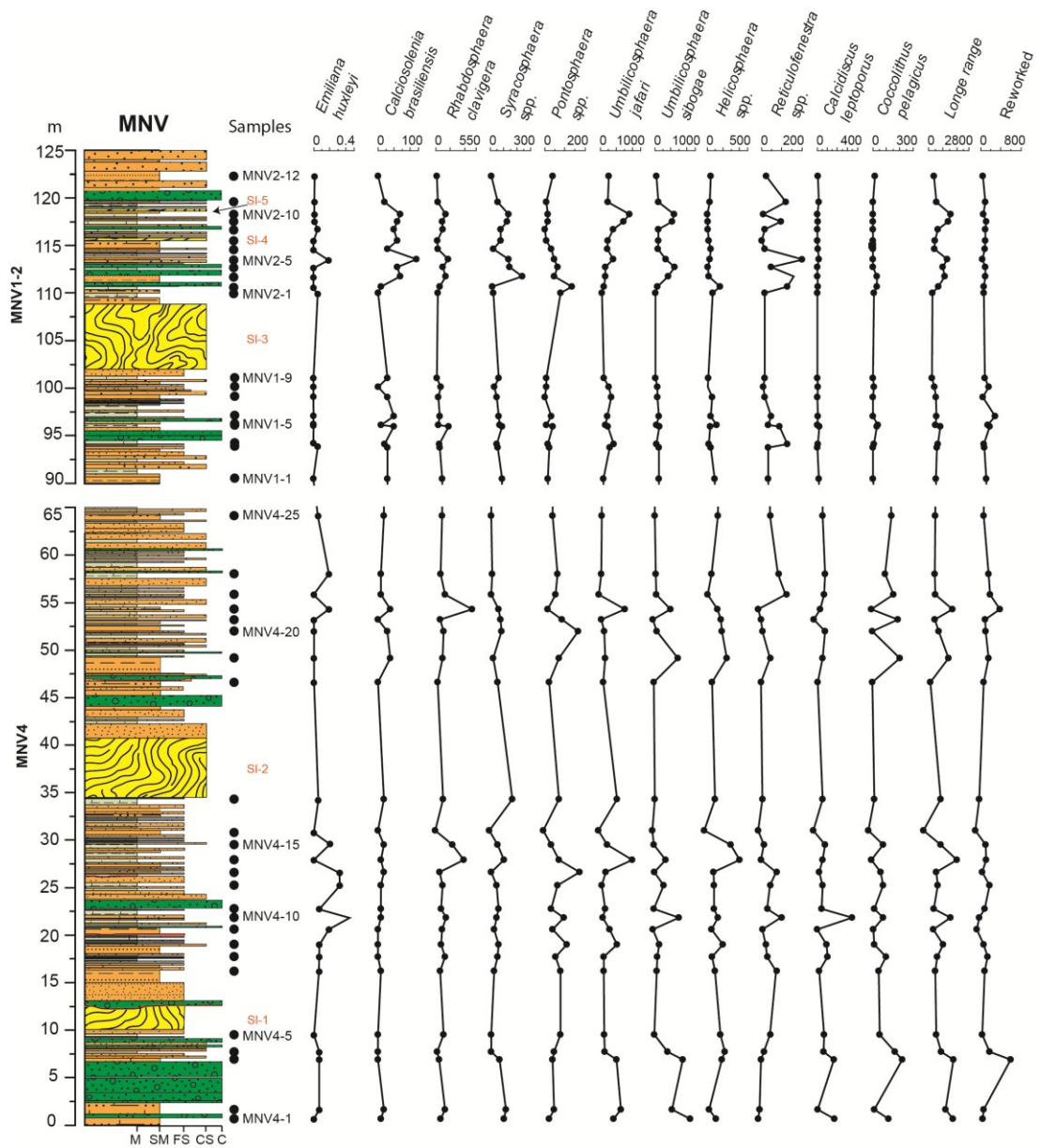


Figure 8. Quantitative calcareous nannofossil biostratigraphy at MNV composite section. Abundances are expressed as number of specimens per mm².

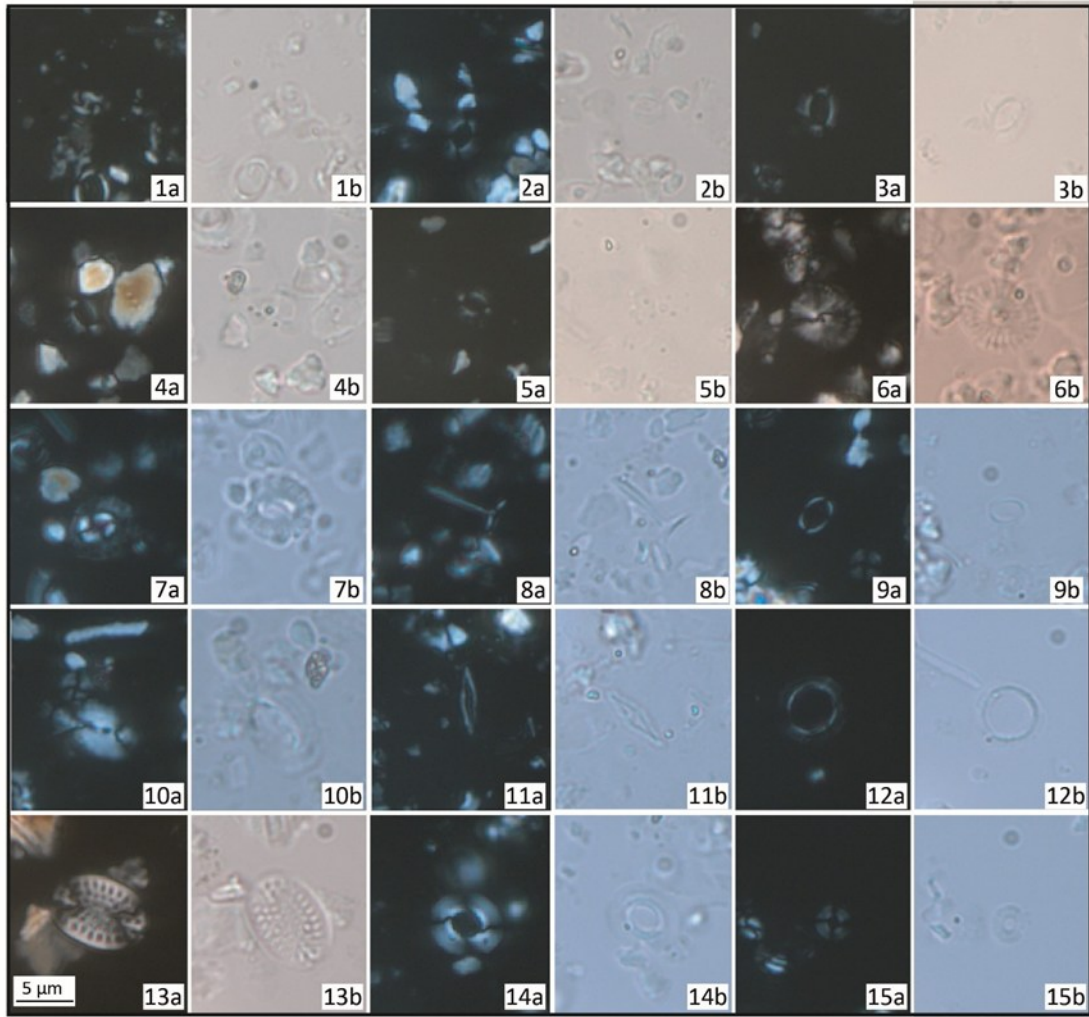


Figure 9. Microphotographs of calcareous nannofossils from studied sections; (1) small *Gephyrocapsa* (MNV2-2); (2-5) *Emiliana huxleyi* (MNV4-15, MNV4-7, MNV4-11, MNV2-8); (6) *Calcidiscus leptoporus* (MNV4-10); (7) *Coccolithus pelagicus* (MNV4-1); (8) *Rhabdosphaera clavigera* (MNV1-8); (9) *Syracosphaera* spp. (MNV4-12); (10) *Helicosphaera carteri* (MNV1-18); (11) *Calciosolenia* sp. (MNV1-1); (12) *Umbilicosphaera sibogae* (MNV2-5); (13) *Pontosphaera carteri*, (MNV4-11); (14) *Reticulofenestra* sp. (MNV1-4); (15) *Umbilicosphaera jafari* (MNV4-10). (a) Crossed nicols; (b) Parallel nicols. Scale bar uniform for all images.

2.3.2.2. Foraminifera

The planktonic foraminiferal assemblage in the MNV composite section includes several stratigraphically significant species occurring during the Pleistocene (Fig. 10). *Globoturbotalita rubescens* and *Neogloboquadrina* spp. are found frequently throughout the entire section, and to a lesser extent, *Globigerinella calida* is also found frequently. *Globigerinoides ruber* (pink) is frequently present throughout the MNV4 section and sporadically appears in MNV 1 and MNV2 sections. *Globoconella inflata* is encountered only at the bottom of the MNV1 section, whereas *Globorotalia crassaformis*, *G. truncatulinoides*, and *G. tenellus* are

found sporadically in a limited number of samples. An important Pleistocene marker species is the benthic foraminifera *Hyalinea balthica*, which is also found rarely in the MNV4 and MNV1 sections (Figs. 10 and 11).

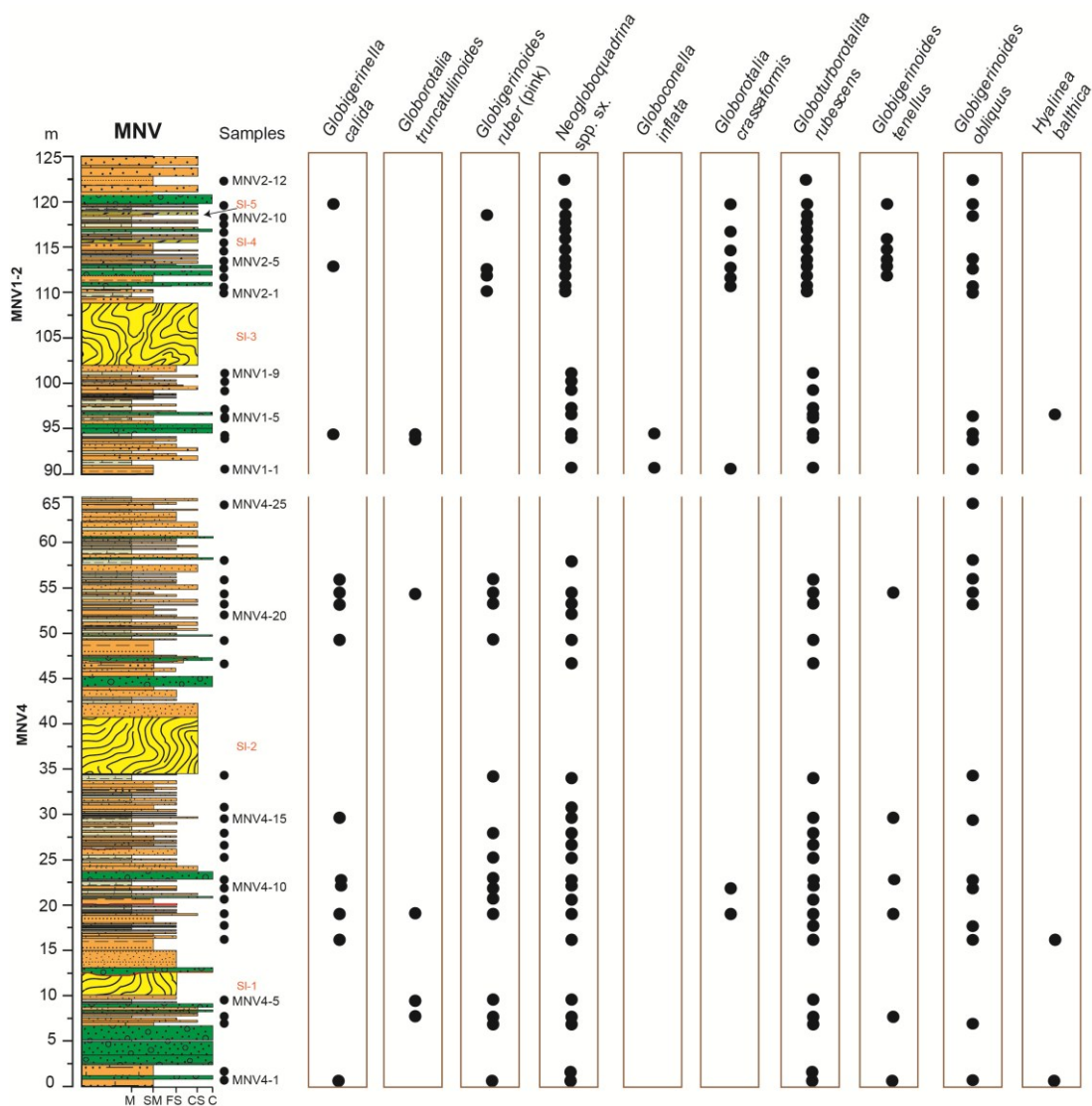


Figure 10. Benthic and planktonic foraminifera of the MNV composite section. Dots indicate presence of each species in the corresponding sample, illustrating their vertical distribution patterns.

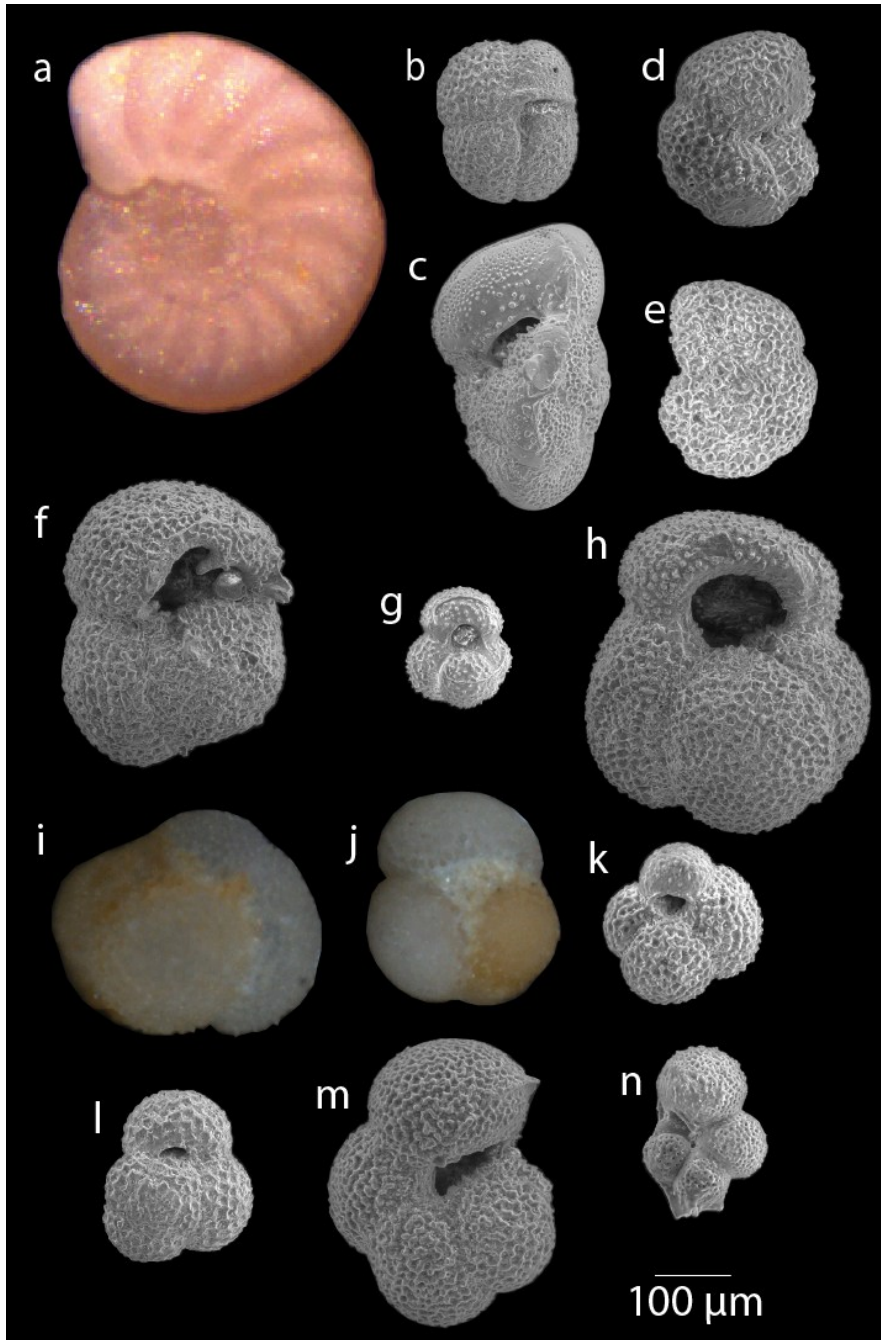


Figure 11. SEM and optical microscope photos of foraminifer marker species. (a) *Hyalinea balthica* (umbilical view) MNV4-1; (b) *Neogloboquadrina pachyderma* (umbilical view) MNV4-9; (c) *Globorotalia crassaformis* (lateral view) MNV2-11; (d) *Globorotalia truncatulinoides* (umbilical view) MNV3-6; (e) *G. truncatulinoides* (spiral view) MNV3-6; (f) *Globoconella inflata* (umbilical view) MNV1-1; (g) *Globoturborotalita rubescens* (umbilical view) MNV4-3; (h) *Globigerinoides obliquus* (umbilical view) MNV4-1; (i) *Globigerinoides ruber* (pink) (umbilical view) MNV4-1, j. *G. ruber* (pink) (umbilical view) MNV2-10, k. *Globigerinoides tenellus* (umbilical view) MNV4-22; (l) *G. tenellus* (spiral view) MNV4-15; (m) *Globigerinella calida* (umbilical view) MNV4-15; (n). *Globigerinella calida* (umbilical view) MNV4-19. Scale bar corresponds to 100-μm for all specimens.

2.4. DISCUSSION

2.4.1. Depositional Environments and Facies Evolution

Based on lithological and sedimentological characteristics, three distinct facies associations are recognized along the MNV composite section (Fig. 12): 1) proximal upper delta-front; 2) distal lower delta-front; and 3) upper delta-front.

2.4.1.1. Lower Interval (0–24 m): proximal upper delta-front

This interval is dominated by channelized, clast-supported conglomerates, with subordinate matrix-supported conglomerates interpreted as marine debris-flow deposits. Medium- to coarse-grained sandstones are also present, along with scarce cross- and parallel-laminated sandstones indicative of low-density turbidity flows. These sandstones are interbedded with marls rich in planktonic foraminifera, reflecting vigorous sediment supply to a proximal upper delta-front setting. This lower interval is also characterized by the presence of a slumped horizon (2.5 m-thick) mainly consisting of matrix-supported carbonate pebbles, cobbles and boulders (rounded, sub-rounded, and angular), which show a chaotic setting with disrupted layers and minor folds. The upper part of this slumped horizon contains also angular and sub-angular grey marly boulders from the Geceleme Fm. The sedimentary structure of this slumped horizon indicates deposition by a powerful but short-lived event like debris-flow occurred in the proximity of the basin margin. This slumping is affected by a syn-sedimentary normal fault (N150°; 60°) that presents a well-developed hangingwall clastic wedge, showing clear growth strata geometries (Fig. 13).

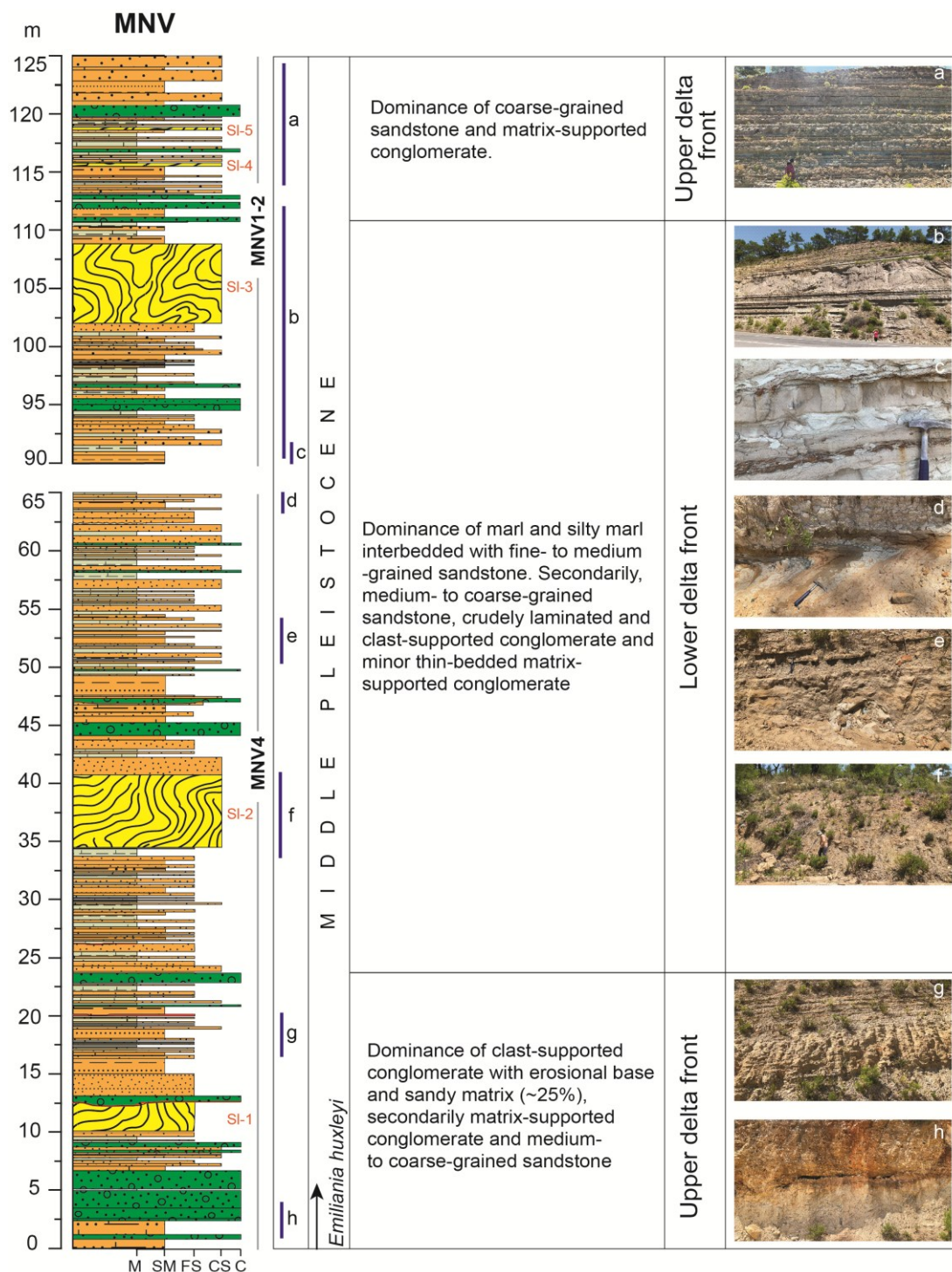


Figure 12: Depositional environment interpretation along the Middle Pleistocene MNV composite section. To the right, some field photographs of the main lithofacies representative of the different depositional environments are shown.

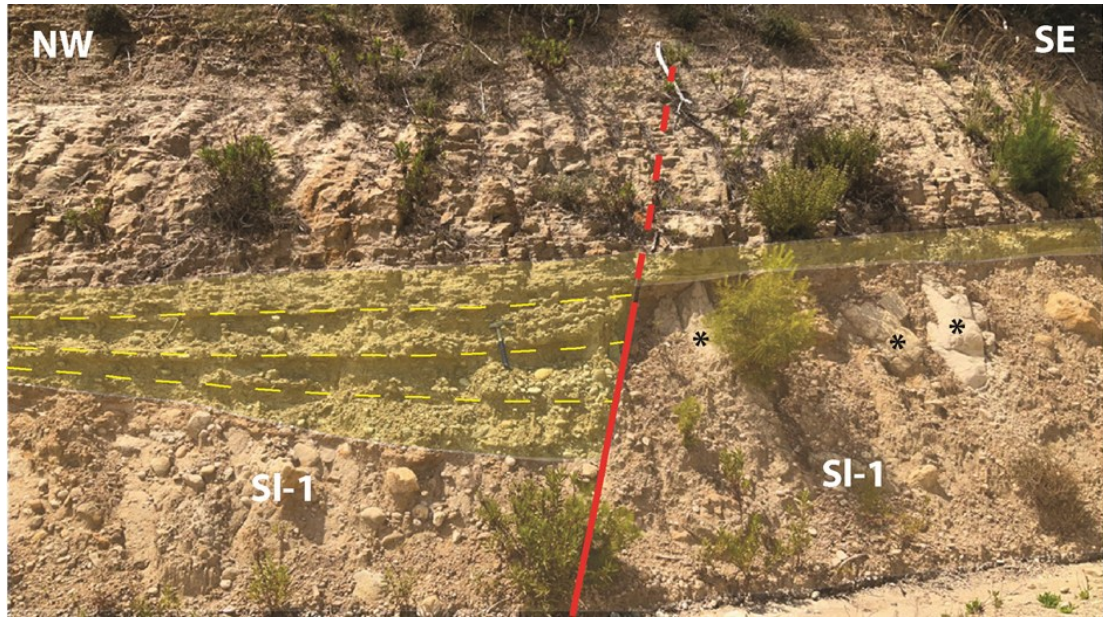


Figure 13. Meso-scale syn-sedimentary normal fault showing a well-developed clastic-wedge, with growth-strata (yellow overlap), at the hangingwall of a fault plane (N150°, 60°) that affects the SI-1 slumping horizon, in the basal portion of MNV4 section. Asterisks indicate some sub-angular marly boulders from the Geceleme Fm.

2.4.1.2. Middle Interval (24–111 m): distal lower delta-front

This interval shows a marked increase in marl with abundant calcareous plankton and well-laminated fine sandstones, indicating a deeper depositional environment. Conglomerate beds, both clast- and matrix-supported, are less frequent and thinner. We interpret this facies association as distal lower delta-front sedimentation, influenced by episodic turbidity flows (e.g., [Wright, 1977](#); [Coleman et al., 1988](#)) and occasionally recording short-lived proximal marine debris flows. Two major slump complexes occur here: SI-2, composed of thick, contorted marl–pebble masses, and SI-3, corresponding to the paleotsunami deposit of [Sagular and Özer \(2024\)](#), which is exceptionally thick and extends for hundreds of meters along the outcrop, serving as a key marker for long-distance correlation.

Limestone pebbles within the slump, together with shell fragments and large clasts of bryozoan and coral colonies, suggest an extra-basinal source, likely due to the erosion of the reefal limestones of the Middle Miocene Oymapınar Fm, which forms the substratum of the MNV section and characterizes the Manavgat Sub-basin margin ([Flecker et al., 1998](#); [Karabıyıkoglu et al., 2000](#); [İslamoğlu, 2002](#); [Çiner et al., 2008](#)).

2.4.1.3. Upper Interval (111–125 m): upper delta-front

This interval is dominated by clast-supported conglomerates, with subordinate matrix-supported conglomerates and medium- to coarse-grained sandstones, showing coarsening- and thickening-upward trends. The interval contains also two thin slumped horizons (SI-4 and SI-5). The return to coarser sedimentation, with a coarsening-upward trend that includes debris flow deposits, indicates renewed upper delta-front progradation.

2.4.1.4 Depositional Context

In the study section (MNV composite section), the deltaic sedimentary complex rests in angular unconformity on the Middle Miocene shallow-water limestones of the Oymapınar Fm, as shown in the scheme of Fig. 14, revealing a significant stratigraphic gap. This gap is undoubtedly due to the complex tectonics that affected the Antalya Basin from the Tortonian to the Lower Pleistocene, with phases of compressional (Lycian phase, late Tortonian-early Messinian; and Aksu phase, Messinian-Zanclean) and extensional tectonics (Late Pliocene-Present), which induced uplift and subsidence at the basin margin.

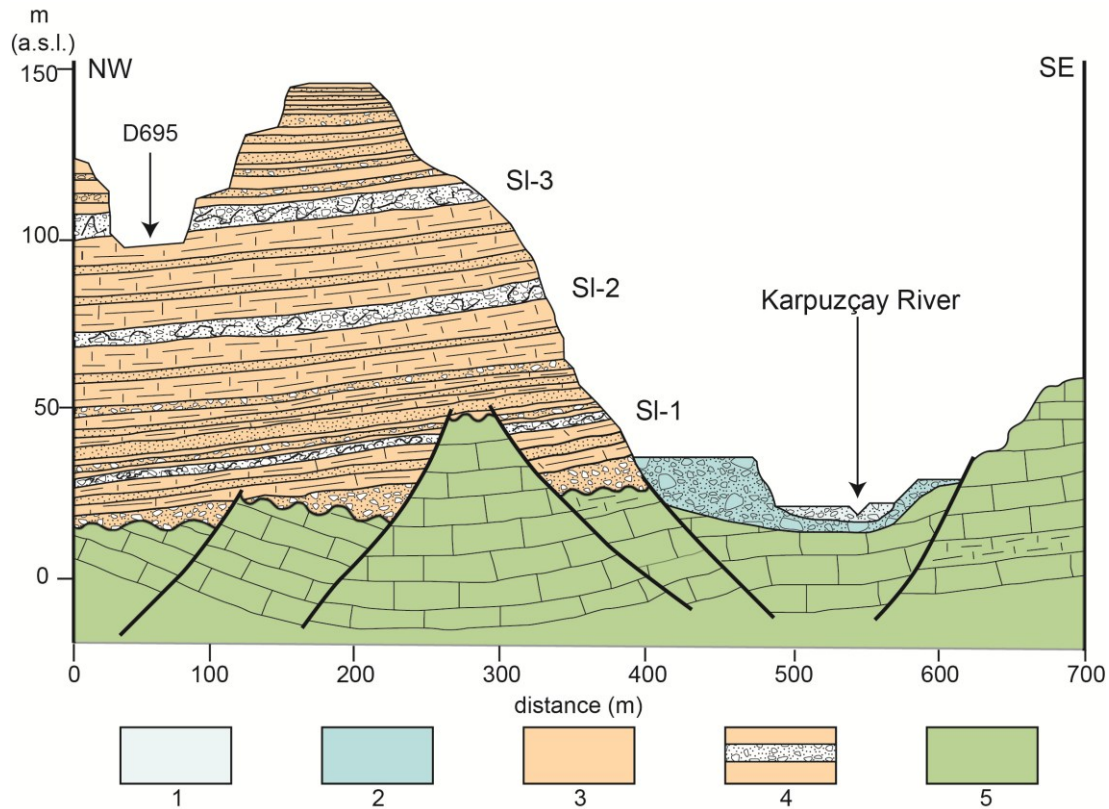


Figure 14. Field sketch showing the relationship between the sedimentary succession of the MNV composite section (Middle Pleistocene p.p., Chibanian p.p.) and the carbonate bedrock consisting of the Oymapınar Fm, (Middle Miocene p.p., Langhian). (1) Holocene fluvial

deposits; (2) Upper Pleistocene p.p. fluvial deposits; (3) marine deposits of the Manavgat Fm (Middle Pleistocene p.p.); (4) slumped horizon within the Manavgat Fm (Middle Pleistocene p.p., Chibanian); (5) shallow-water limestones of the Oymapınar Fm, (Middle Miocene p.p., Langhian).

In reconstructing the depositional context of the Manavgat sub-basin, it is interesting to note that the cobbles/boulders present in the different intervals of the delta complex show an inverse stratigraphy, with younger rocks in the basal interval of the delta complex (e.g., marls from the Geceleme Fm, Serravallian) and older rocks in the upper intervals (corals and bryozoan colonies from the Oymapınar Fm, Langhian). A stratigraphy with inverted clast ages often is due to tectonic uplift, basin inversion, erosion of uplifted blocks, or recycling. In the context of the Antalya Basin, the tectonic inversion that occurred between the last compressive phase (Aksu phase, Messinian-Zanclean) and the currently ongoing extension phase (Upper Pliocene-Present) certainly induced the uplift of the basin margin, which reversed the sediment supply.

Overall, the MNV succession records a regressive–transgressive–regressive pattern, typical of fan deltas responding to variations in basin subsidence and relative sea level. The frequent and thick slump deposits suggest rapid sediment delivery and significant basin instability.

The occurrence of syn-sedimentary normal faults affecting the Manavgat sedimentary succession suggests that the instability of the sedimentary basin was certainly triggered by shaking induced by seismicity associated with the movement along those normal faults, which controlled the evolution of the delta complex within the Manavgat Sub-basin.

On this regard, [Aykut et al., \(2025, 2026\)](#) show U-Th and U-Pb ages of fault-related calcites indicating continuous faulting from the Middle/Late Pleistocene to the Holocene, at the transition between the subsiding Köprüçay and Manavgat sub-basins and the uplifting area of the CAP western margin.

2.4.2. Micropalaeontological constraints of the Manavgat Sub-basin

The analysis of the micropalaeontological assemblages record at the MNV succession reveals the presence of pervasive reworking, a feature typical of tectonically active fan-delta systems, where sediment is repeatedly remobilized. The succession exhibits a parallel the co-occurrence of both reworked and long-range species, suggesting that

also many of these last are allochthonous rather than in situ, thereby hampering their precision as age indicators. Such taxonomic mixing likely contributed to the previously inconsistent chronological interpretations of the Manavgat Sub-basin (e.g., Akay et al. 1985; Akay and Uysal 1988; Flecker et al. 1995, 1998; Karabıyıkoglu et al. 2000; Deynoux et al. 2005; Çiner et al. 2008; Yılmaz et al., 2024). Despite the wide reworking, careful micropalaeontological analysis enabled the identification of reliable biostratigraphic markers. The consistent presence of *Emiliania huxleyi*, whose FO in the Mediterranean is dated to approximately 0.26 Ma (Di Stefano et al., 2023), indicates that the MNV succession was deposited after ~0.26 Ma. This demonstrates that the Manavgat Sub-basin remained an active marine depocentre until the late Middle Pleistocene.

Planktonic foraminiferal assemblages provide additional support for this chronostratigraphic framework. The identified species include *Globigerinoides ruber* (pink), *Neogloboquadrina* spp., *Globigerinella calida*, and *Globorotalia truncatulinoides*, whose stratigraphic ranges extend from the Middle Pleistocene to the present. The occurrence of *G. ruber* (pink) offers valuable paleoenvironmental information, although its biostratigraphic utility remains debated due to regional variation in its temporal distribution. Although an age of ~0.33 Ma has been suggested by Lourens et al. (2004) for the FO of the taxon, this result is not directly supported by published data. In the Western Mediterranean, it has been observed occurring during warm intervals such as MIS 5e and the Holocene (Pérez-Folgado et al., 2004), while Linares et al. (1999) reported its earliest occurrence during MIS 9 (ODP Sites 975 and 976). In the Eastern Mediterranean, *G. ruber* (pink) appears as early as MIS 11 (~0.47 Ma; Numberger et al., 2009). These regional discrepancies likely reflect ecological and taphonomic influences rather than diachronous evolutionary events. Importantly, these reinforce a chronologic framework centred on a post-0.26 Ma depositional history.

No ostracod specimens were recovered from the analyzed samples. This absence is consistent with deposition under high-energy, nearshore conditions, where rapid sedimentation, variable salinity, and strong hydrodynamic activity create environments unsuitable for ostracod colonization (Curtis, 1960). Such conditions reflect low environmental stability typical of shallow-marine fan-delta settings.

2.4.3. Impact on the chronostratigraphy of the Antalya Basin

Marine sedimentation within the central Isparta Angle is recorded in the Aksu, Köprüçay, and Manavgat sub-basins starting from the Early Miocene (Burdigalian), where Neogene–Quaternary sequences lie unconformably atop the Tauride carbonates (Glover and Robertson, 1998; Poisson et al., 2003, 2011; van Hinsbergen et al., 2010; van Hinsbergen and Schmid, 2012; Robertson et al., 2013).

The Aksu Basin preserves one of the most complete Late Cenozoic successions, comprising Miocene marine clastics with patch reefs, Pliocene shallow-marine deposits, and Lower Pleistocene–Recent marine to terrestrial clastics and travertines (Akay et al., 1985; Şenel et al., 1992, 1996; Monod, 2001; Glover and Robertson, 2003; Poisson et al., 2003, 2011; Çiner et al., 2008; Wasoo et al., 2020). Updated foraminiferal biostratigraphy from the topmost Eskiköy Fm (Kanbur and Öğretmen, 2022) extended the age of the succession from the previously assumed Pliocene to the Early Pleistocene (1.54–1.36 Ma), based on diagnostic planktonic taxa including *Neoglobobulimina pachyderma*, *Globigerinoides tenellus*, and *Globoturborotalita rubescens*.

The Köprüçay Sub-basin similarly contains a mixed siliciclastic–carbonate Miocene infill, including reefal shelf carbonates, mudstones, turbidites, and thick marginal conglomerates, documented in earlier regional mapping and facies studies (Blumenthal, 1951; Flecker, 1995; Poisson et al., 2003a, 2003b; Deynoux et al., 2005; Çiner et al., 2008; Koç et al., 2016).

The Manavgat Sub-basin has a long history of stratigraphic investigation (Blumenthal, 1951; Bizon et al., 1974; Akay et al., 1985; Flecker et al., 1998, 2005; Karabıyıkoglu et al., 2000, 2005; Deynoux et al., 2005; Çiner et al., 2008), with recent studies refining its Miocene carbonate and clastic delta facies architecture (İlgar et al., 2024; Larsen et al., 2024a, 2024b; Wathne et al., 2024; Yılmaz et al., 2024).

A major re-evaluation of the stratigraphy of the Manavgat Sub-basin was presented by Sagular and Özer (2024), who proposed a significant revision of the lithostratigraphy based on calcareous nannofossil assemblages. As other previous authors, they placed in the Middle Miocene the Oymapınar Fm (Langhian) and the overlying Geceleme Fm (Serravallian), but crucially argued that deposits of the Karpuzçay Fm, previously assigned to the Middle-Upper Miocene (Serravallian-

Tortonian), in fact represent two distinct successions: (i) a Lower Pliocene clastic shelf interval, which retains the name of the Karpuzçay Fm, and (ii) a Lower Pleistocene shallow-marine unit, newly defined as the Manavgat Fm. These Lower Pleistocene strata (Manavgat Fm) are interpreted to rest unconformably on the Lower Pliocene deposits (Karpuzçay Fm) and begin with a distinctive chaotic mass-transport unit, which they attribute to a paleotsunami event. This reinterpretation extends marine sedimentation to ~1.6 Ma and reframes the stratigraphic architecture of the Manavgat Sub-basin, providing an important but debated foundation for subsequent chronostratigraphic analyses.

Our new sedimentological and micropalaeontological observations provide a different perspective on both age and process. Deformed and chaotic units in the MNV composite section lack the diagnostic features of regionally sourced tsunami deposits and instead resemble internally generated slump packages typical of unstable delta settings consistent with the architecture of the MNV succession.

Furthermore, the MNV composite section includes both units reported in the stratigraphic model by [Sagular and Özer \(2024\)](#) for the Manavgat Sub-basin (Karpuzçay and Manavgat Fms). In particular, the mega slumping Sl-3 at the top of MNV1 corresponds to the paleotsunami event that, according to [Sagular and Özer \(2024\)](#), separates the Lower Pliocene deposits (Karpuzçay Fm) from those of the Lower Pleistocene (Manavgat Fm). As reported before, the MNV composite section represents a late Middle Pleistocene succession without hiatuses. These observations, together with their distribution on the geological map of the Manavgat Sub-basin provided by [Sagular and Özer \(2024\)](#), allow us to extend the Pleistocene age to the entire area previously indicated as affected by the Karpuzçay Fm, reaching its unconformable boundary with the Geceleme Fm (Middle Miocene, Serravallian). Therefore, our results lead us to amend the definition of Manavgat Fm proposed by [Sagular and Özer \(2024\)](#), including in the Manavgat Fm also the portion of succession that these authors attributed to the Karpuzçay Fm.

This conclusion is also supported by the observation that the top of the MNV composite section (~143 m a.s.l.) is associated with a gentle seaward dip (0.5–3°) palaeosurface, defining a late Middle Pleistocene delta front, which extends from ~170 m (inner margin) down to ~20 m (external edge) above present sea level, elongating perpendicular to the coast over a distance from 9 to 11 km (Fig. 15).

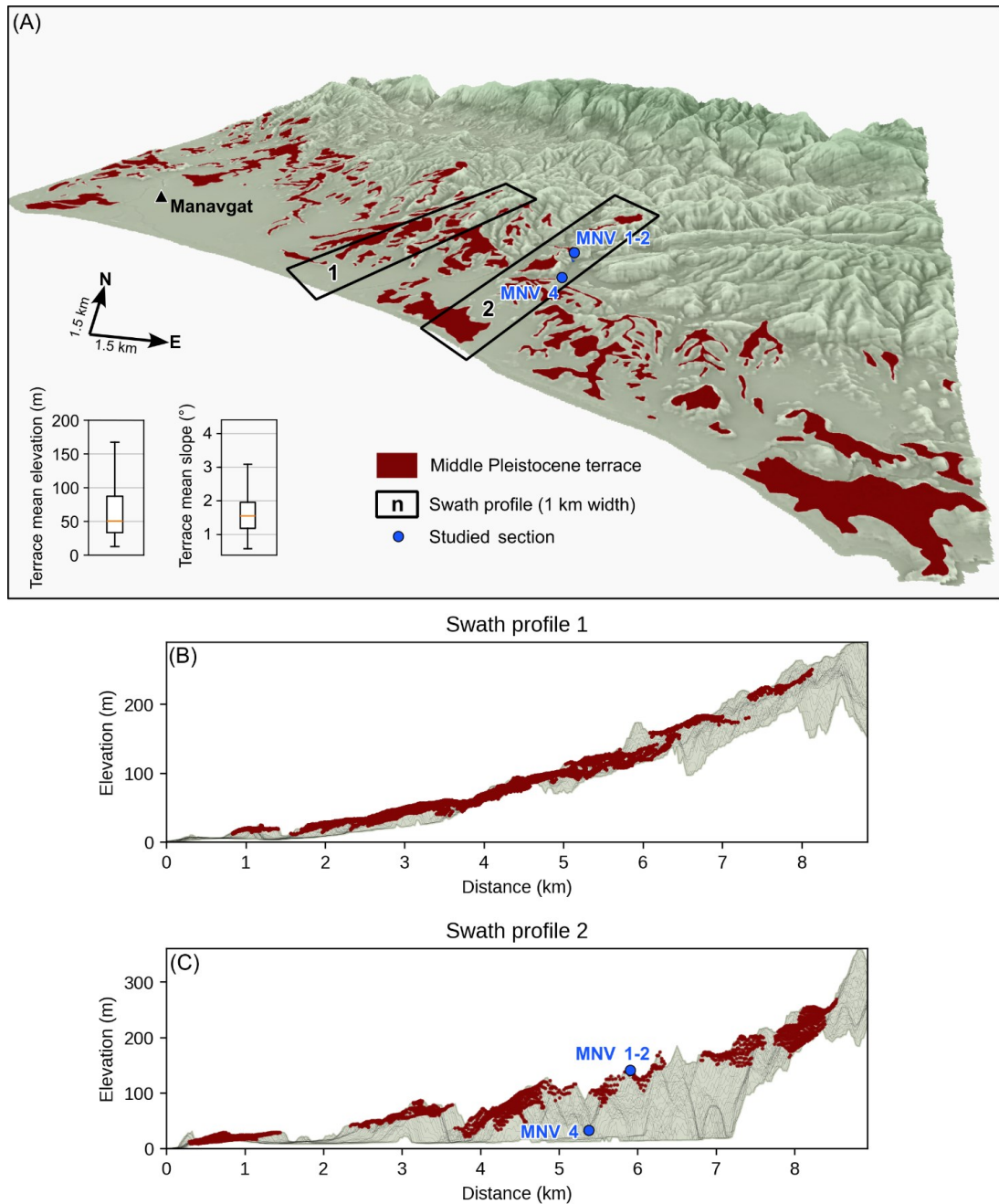


Figure 15. (A) Three-dimensional view of the late Middle Pleistocene marine terrace identified in the Manavgat area, boxplots of the mean elevation and slope, and boxes used to trace the swath profiles. The terrace is mapped and visualized in QGIS through direct digital mapping, using the FABDEM v1.2 digital terrain model (<https://data.bris.ac.uk/data/dataset/>) as base topography and red relief image map (RRM) analysis following Chiba et al. (2008); (B) Swath profile along the box 1 highlighting the position of the Middle Pleistocene terrace; (C) Swath profile along the box 2, highlighting the position of the Middle Pleistocene terrace and the location of the studied sections.

Calcareous nannofossil evidence, together with planktonic foraminiferal assemblages, places the base of the MNV composite section younger than ~0.26 Ma (FO of *Emiliania huxleyi*), correlating the uppermost marine deposits in the Manavgat Sub-basin with the late Middle Pleistocene (~0.23 Ma, MIS 7) strata of the

TOL-1 section in the Mut Basin (Sarı et al., 2025, under review). The presence of *E. huxleyi* at the base of the MNV section and near the top of TOL-1 demonstrates that both successions overlap within the ~0.26–0.23 Ma interval. Although the MNV record represents a higher-energy fan-delta system characterized by variable sediment accumulation, this correlation provides a strong regional chronostratigraphic anchor. Accordingly, the age of the MNV composite section is constrained between ~0.26 Ma and ~0.23 Ma, making it nearly 1 Myr younger than the age proposed by Sagular and Özer (2024) and much younger than the previously reported marine deposits in the Antalya Basin (e.g. Karabıyıklıoğlu et al., 2000, 2005; Deynoux et al., 2005; Çiner et al., 2008).

Despite extensive research across all the three sub-basins, the chronostratigraphy of the Antalya Basin remains complicated by reworking, localized hiatuses, variable sedimentation rates, and sparse high-resolution biostratigraphic control. These results refine the chronostratigraphic framework of the Manavgat Sub-basin and demonstrate that shallow-marine conditions, at least at the eastern limb of the Isparta Angle, persisted well into the late Middle Pleistocene.

2.4.4. Tectonic controls on the basin evolution

The Antalya Basin preserves a multi-stage tectonic history characterized by alternating phases of extension and compression throughout the Neogene–Quaternary. Regional structural, geomorphic, and paleostress datasets indicate four principal phases: (1) pre-Tortonian E–W extension initiated the early Antalya Basin and established the Manavgat, Köprüçay, and Aksu sub-basins (Hall et al., 2014; Koç et al., 2017; McPhee et al., 2018); (2) Tortonian–Pliocene compression, expressed by the Lycian and Aksu deformation phases along the Aksu Thrust, uplifted and folded basin strata (Glover and Robertson, 1998a; Poisson et al., 2003a, 2011; Deynoux et al., 2005); (3) from the Late Pliocene onward, the region entered a renewed phase of N–S-directed extension during which inherited thrusts were reactivated as normal faults, and NW–SE-oriented fault systems, including the Kırkkavak Fault, began accommodating NE–SW extension (Glover and Robertson, 1998).

Geomorphic indicators such as deeply incised valleys, active tectono-karst depressions, and fault-controlled gorges confirm that this extensional regime persisted throughout the Quaternary (Üner et al., 2013; Doğan et al., 2017; Aykut et

al., 2023, 2025, 2026). This prolonged extensional setting also allowed shallow-marine sedimentation to continue into the late Middle Pleistocene, as demonstrated by our new chronostratigraphy.

Syn depositional deformation within the MNV composite section is recorded by meso-scale syn-sedimentary extensional faults (Fig. 13), abundant turbidity flows, and five major slump complexes (from SI-1 to SI-5). The disrupted internal architecture of the slump deposits and the presence of extra-bacinal carbonate blocks, derived from erosion of uplifted Miocene units, indicates abrupt fault-driven slope failure, most plausibly triggered by renewed displacement along the Manavgat Basin boundary faults.

The presence of slumpings in the Middle Pleistocene succession of the Manavgat Sub-basin further links its tectono-sedimentary evolution with that of the Mut Basin, where Öğretmen et al. (2018) report a mega-slump in the Middle Pleistocene deposits at the end of the marine succession capping the CAP southern margin. These observations demonstrate that the fan-delta system of the Manavgat Sub-basin evolved in a tectonically active setting where extension, subsidence, and sediment supply interacted dynamically.

Yılmaz et al. (2024) identified two fault systems in the Manavgat Basin. Small normal faults affecting the Tepekli Fm and the larger Çardakköy Thrust, which they interpreted to place the Karpuzçay Fm onto the Tortonian Köprüçay Conglomerate. They also proposed that early basin deepening in the east was controlled by a buried NW-trending normal fault responsible for differential subsidence across the basin (Gömülü Normal Fay in Yılmaz et al., 2024).

However, their tectono-stratigraphic basin evolution depends on assigning the Karpuzçay Fm a Serravallian–Tortonian age. Our revised chronostratigraphy shows that these units actually contain mainly late Middle Pleistocene shallow-marine deposits, not Upper and Middle Miocene strata, which requires a fundamental re-evaluation of both slip rates and timing of the tectono-sedimentary basin evolution.

After ~0.23 Ma, the basin experienced rapid uplift under continued extension. The correlation of the uppermost marine deposits in the MNV succession with the late Middle Pleistocene strata in the Mut Basin (~0.23 Ma, MIS 7; Sarı et al., 2025), combined with paleodepth constraints and MIS 7 sea-level estimates, yields a mean uplift rate of ~0.8 m/kyr. This places the Manavgat Sub-basin among the most rapidly uplifting sectors of the eastern Isparta Angle.

The timing of the Manavgat Sub-basin uplift aligns with the deformation episodes at ~450 ka and 100–250 ka documented by U–Th–dated calcite veins in the western Central Anatolian Plateau (Aykut et al., 2025), indicating that uplift and extension were synchronous during the Middle Pleistocene, deeply influencing the stratigraphy of the Antalya Basin with debris flows and slumped thick horizons.

Sea level markers along the Eastern Mediterranean coast show that differential vertical motion continues into the late Holocene: the crustal block west of Antalya exhibits subsidence of –0.8 to –2.3 m/kyr (Çiner et al., 2009; Desruelles et al., 2009; Anzidei et al., 2011), whereas the Antalya–Alanya corridor shows moderate uplift of 0.17–0.35 m/kyr (Liberatore et al., 2022). This pattern is consistent with continued crustal adjustment under an extensional regime.

The underlying drivers of this Quaternary extension/uplift have been interpreted by Schildgen et al. (2012) and Howell et al. (2017), who proposed that extension reflects lithospheric warming and relaxation linked to slab tearing and associated asthenospheric upwelling. In contrast, other studies attribute the extensional tectonic deformation to rollback of a remnant segment of the southern Neo-Tethyan slab (Flecker et al., 2005; Glover and Robertson, 1998b; Kelling et al., 2005; Koçyiğit and Özacar, 2003; Üner et al., 2015). The most plausible interpretation is that both processes contributed: slab fragmentation and asthenospheric upwelling promoted regional uplift and crustal weakening, while continued slab rollback drove NE–SW extension and the activation of NW–SE-striking normal faults. Together, these mantle and crustal processes shaped the late Middle Pleistocene–Holocene evolution of the Manavgat Sub-basin and the whole Antalya Basin.

2.5. CONCLUSIONS

Integrated lithostratigraphic and micropalaeontological analyses of the MNV composite section reveal a deltaic succession comprising upper delta-front, lower delta-front, and renewed upper delta-front facies, reflecting alternating high- and low-energy shallow-marine conditions. Calcareous nannofossils, together with planktic foraminiferal assemblages, constrain the base of the succession to be younger than ~0.26 Ma (FO of *Emiliania huxleyi*), while the uppermost marine deposits possibly correlate the late Middle Pleistocene strata (~0.23 Ma, MIS 7) in the Mut Basin (Sarı et al., 2025). These findings extend the age of marine deposition in the Manavgat Sub-basin well into the late Middle Pleistocene, significantly

younger than previously reported (Middle Miocene or Early Pleistocene) assignments. The present elevation of the section top (~143 m a.s.l.), combined with paleodepth and MIS 7 sea-level reconstructions, indicates an average uplift rate of ~0.8 m/kyr since ~0.23 Ma, providing new quantitative constraints on the basin's late Quaternary tectono-sedimentary evolution. This rapid surface rise occurred under continued NE–SW-directed extension, facilitated by reactivation of the tectonic activity along NW–SE-oriented fault systems. Based on previous studies, the observed uplift/subsidence history suggests that both crustal deformation and slab-related geodynamic processes might be key drivers of the Antalya Basin's late Quaternary tectono-sedimentary evolution.

3. QUATERNARY DIFFERENTIAL UPLIFT IN MUT AND ANTALYA BASINS (SOUTHERN TURKEY): A RESPONSE TO DEEP-SEATED DYNAMIC PROCESSES

ABSTRACT

Along the southern margin of the Central Anatolian Plateau (CAP), we document the world's highest late Middle Pleistocene marine deposits, providing an exceptional record of rapid vertical uplift. New biostratigraphic constraints show that the Tol-1 section in the Mut Basin was deposited between <0.53 Ma and 0.23 Ma and now stands at 1177 m a.s.l., whereas the coeval MNV fan-delta succession in the Antalya Basin, situated within the Isparta Angle, was deposited between <0.26 Ma and 0.23 Ma remains at 143 m a.s.l. Despite being deposited synchronously, these successions record strongly divergent uplift rates of 5.2 and 0.8 m/kyr, respectively, revealing fundamental differences in the processes governing vertical movements.

The deep processes beneath the southern margin of the CAP are dominated by asthenospheric upwelling and crustal rebound following slab break-off, whereas uplift in the IA reflects the combined influence of slab retreat and thermal–viscous relaxation of the upper plate. This different vertical movement of the upper plate is needed to be accommodated by the action of a slab tear fault separating the Cyprus and the Antalya slabs.

By integrating these results with marine-terrace modelling and river-profile inversions, we refine the evolution of the late Quaternary uplift of the CAP southern margin. When combined with seismic tomography, the compiled constraints reveal two discrete slab break-off events beneath the Cyprus subduction system: an initial Late Miocene slab break-off and a renewed Middle Pleistocene shallow slab break-off. We propose that repeated slab fragmentation along the continuous subduction governed the surface evolution of Anatolia and the eastern Mediterranean.

3.1. INTRODUCTION

Differential vertical motions of Earth's crust are governed by tectonic deformation, surface loading and unloading, and thermal re-equilibration within the lithosphere as well as by deep-seated processes (Watts, 2001; Faccenna et al., 2006; Zhang et al., 2021). Although the fastest present-day uplift occurs in regions dominated by

glacial-isostatic rebound (Richter et al., 2017), tectonic and deep-seated processes can also generate rapid uplift, such as the 6 ± 3 m/kyr along Chile's Liquiñe–Ofqui Fault Zone, driven by oblique ridge subduction (Perroud and De Pascale, 2025), and ~ 8 m/kyr in eastern Tibet associated with intracontinental subduction (Pan and Shen, 2017). Along the southern margin of the Central Anatolian Plateau (CAP), where the African Plate subducts beneath the Anatolian Plate, similarly high rates were documented (from 4.1 m/kyr to 3.5 m/kyr; Öğretmen et al., 2018; Racano et al., 2020). Thus, the CAP southern margin is a key natural laboratory for evaluating how mantle-driven dynamics shape Earth's surface (Reilinger et al. 2006; Faccenna et al., 2006; Schildgen et al., 2012 and 2014; Öğretmen et al., 2018; Sarı et al., 2025a).

The Neogene–Quaternary evolution of the Mut and Antalya basins preserves a complementary record of how deep-seated mantle processes and upper-plate deformation co-evolved along the CAP southern margin and the Isparta Angle (IA) (Cosentino et al., 2012; Schildgen et al., 2012a, 2012b; Delph et al., 2017; Meijers et al., 2018; Fernández-Blanco et al., 2019). Both basins contain thick Miocene–Pleistocene marine successions resting unconformably on Tauride basement, yet they formed within markedly different tectonic domains (Fig.1b). The Mut Basin, situated on the upper plate of the eastern segment of the Cyprus subduction zone, preserves marine deposits now uplifted to ~ 2 km. They show kilometre-scale vertical movements (both uplift and subsidence) beginning with the deposition of upper Burdigalian deep marine marls (see in Fig.1b Başyayla and Olukpinar sections; Cosentino et al., 2012; Cipollari et al., 2013). Marine sedimentation on the CAP southern margin continued until the latest Messinian (<6.7 Ma; see in Fig.1b Sarıalan section; Schildgen et al., 2012b).

Its largely undeformed Miocene–Pleistocene marine succession supports interpretations that uplift was driven primarily by deep-seated processes beneath the Cyprus Arc (Cosentino et al., 2012; Schildgen et al., 2012a; 2012b; 2014), in contrast with the interpretation of the CAP southern margin as the forearc high of the Cyprus subduction system (Fernandez-Blanco et al., 2019). Conversely, the Antalya Basin developed within the structurally complex Isparta Angle, a transfer zone linking the Hellenic and Cyprus arcs (Glover and Robertson, 1998; Hall et al. 2014). Neogene–Quaternary deformation in the IA was punctuated by alternating extension and contraction: (i) pre-Tortonian E–W extension that generated the Antalya Basin, including Aksu, Köprüçay, and Manavgat sub-basins (Flecker, 1995; Çiner et al.,

2008); (ii) Messinian–early Pliocene contraction during the Lycian and Aksu deformation phases driven by gravitational sliding of the Lycian Nappes and/or differential uplift and subsidence across the IA (van Hinsbergen et al., 2010); and (iii) renewed N–S extension from the Late Pliocene to the present, coeval with Middle Pleistocene uplift (Glover and Robertson, 1998; Aykut et al., 2023, 2025; Sari et al., 2025b). This tectono-sedimentary history produced a mosaic of evolving depocentres hosting repeated marine incursions, uplift, and widespread Quaternary faulting (Poisson et al., 2003; Wasoo et al., 2020).

The contrast between these two southern Turkey regions becomes most pronounced in the late Middle Pleistocene. The TOL-1 section in the Mut Basin records exceptional uplift (~5.2 m/kyr) and hosts the world’s highest late Middle Pleistocene marine deposits (>1177 m a.s.l.; Sari et al., 2025a), whereas the coeval Manavgat succession in Antalya experienced subdued uplift (~0.8 m/kyr, 143 m a.s.l.; Sari et al., 2025b). Although deposited synchronously, these successions differ in present elevation by more than 1 km, reflecting fundamentally different deep geodynamic settings.

After the late Messinian uplift phase at the CAP southern margin (ca. 4 m/kyr, Öğretmen et al., 2018), which following the Duretz and Gerya (2013) model could be referred to crustal delamination processes that started after 7 Myr of continental crust subduction (Öğretmen et al., 2018), the Mut province recorded a second uplift phase focused in the late Middle Pleistocene (ca. 3 m/kyr, Öğretmen et al., 2018; ca 5 m/kyr, Sari et al., 2025a). As already suggested by Öğretmen et al. (2018), the late Middle Pleistocene uplift phase could be related to the slab break-off of the subducting plate, which induced rapid asthenospheric upwelling and crustal rebound beneath the CAP southern margin. In contrast, the lower uplift of the Antalya Basin was controlled by the subduction of a not detached and actively retreating Antalya slab, as evidenced by the tomography model of Portner et al. (2018) for the West Cyprian slab.

Together, the subsidence–uplift histories of both basins demonstrate that vertical movements along southern Anatolia can be explained by a multiphased uplift history (e.g., Schildgen et al., 2012; Öğretmen et al., 2018).

As stated before, the multiphased uplift history that affected the CAP southern margin has been explained by Öğretmen et al. (2018) with two uplift phases triggered by different geodynamic processes: 1) a late Messinian (ca. 5.45 Ma)

crustal delamination process, and 2) a late Middle Pleistocene (ca. 0.4 Ma) slab break-off.

Here, we use well-dated marine successions to quantify the magnitude, timing, and spatial differential uplift along the southern margin of the CAP and the IA (Fig. 1) to better reconstruct the Neogene uplift history of southern Anatolia and possibly refine the geodynamic model responsible for the multiphased vertical movements recorded in their marine deposits. By integrating new biostratigraphic constraints with geomorphological analyses, including marine-terrace modelling and river-profile inversion, we propose a geodynamic framework able to link both uplift phases and intervening subsidence with mantle-driven processes related to the Africa-Arabia-Eurasia subduction zone.

3.2. METHODS

Two stratigraphic sections were investigated: the TOL-1 section in the Mut–Ermenek Basin (36.385225°N, 33.424114°E; 1177 m a.s.l.), and the MNV composite section in the Manavgat Sub-basin (MNV4: 36°44′28.74″N, 31°37′21.07″E; MNV1–2: 36°44′30.76″N, 31°36′50.77″E).

At TOL-1, 68 bulk sediment samples were collected through a ~71-m-thick deeper-to-shallow-marine succession. At MNV, 46 marl samples were collected from the 125-m-thick composite section, including MNV4 and MNV1-2 sub-sections.

3.2.1. Calcareous nannofossil analyses

Calcareous nannofossils were analysed using standard smear-slide methods (Backman and Shackleton, 1983) and concentrated preparations following De Capoa et al. (2004) to improve recovery of *Emiliana huxleyi*. Taxonomic identification and biostratigraphic assignments follow the Mediterranean zonation of Di Stefano et al. (2023). For the analysis of TOL-1 section, quantitative counts were performed to define the calcareous nannofossil assemblages along the section, whereas for the MNV section, semi-quantitative counts were used, with ~350 fields of view scanned for rare *E. huxleyi*.

3.2.2. Foraminifera and ostracods

Samples were washed over a 63-μm sieve, dried, and examined using a Leica M205C stereomicroscope. Selected specimens were imaged using SEM. Planktonic

foraminifera were analysed semi-quantitatively and qualitatively for biostratigraphic support after [Lirer et al. \(2019\)](#).

3.2.3. Geomorphological modelling

The uplift history of the CAP southern margin was reconstructed by integrating marine-terrace analysis and river-profile inversion, following the methodologies of [Racano et al. \(2020, 2021\)](#). In this study, these models are updated using new age–elevation constraints derived from the TOL-1 marine section.

Marine terraces were mapped using 12 m resolution TanDEM-X digital elevation models. Terrace elevations were measured at shoreline angles, which represent former sea-level positions, using the TerraceM software ([Jara-Muñoz et al., 2016, 2019](#)). Initial uplift estimates were obtained by comparing terrace elevations with coeval sea-level positions. To account for time-variable uplift, a landscape evolution model (LEM) was applied to three well-preserved terrace sequences along the eastern sector of the southern CAP margin. The LEM was constrained by fixing the age of the uppermost terrace using the new biostratigraphic ages from the Mut Formation and by testing published sea-level curves ([Bintanja et al., 2005; Rohling et al., 2014](#)). Best-fitting uplift histories were identified using a Random Optimization approach (after [Racano et al., 2020](#)).

Independent constraints on rock-uplift history were obtained from inversion of river long profiles draining the southern CAP margin, following [Racano et al. \(2021\)](#). Drainage networks were extracted from the 30 m resolution FABDEM using LSDTopoTools2 ([Mudd et al., 2023](#)). Channel steepness indices were calculated using regionally representative concavity values, and erodibility parameters were calibrated using uplift rates derived from marine terraces. A total of 17 catchments were inverted using a bootstrap approach, and individual results were combined to produce a spatio-temporal model of rock uplift along the southern CAP margin.

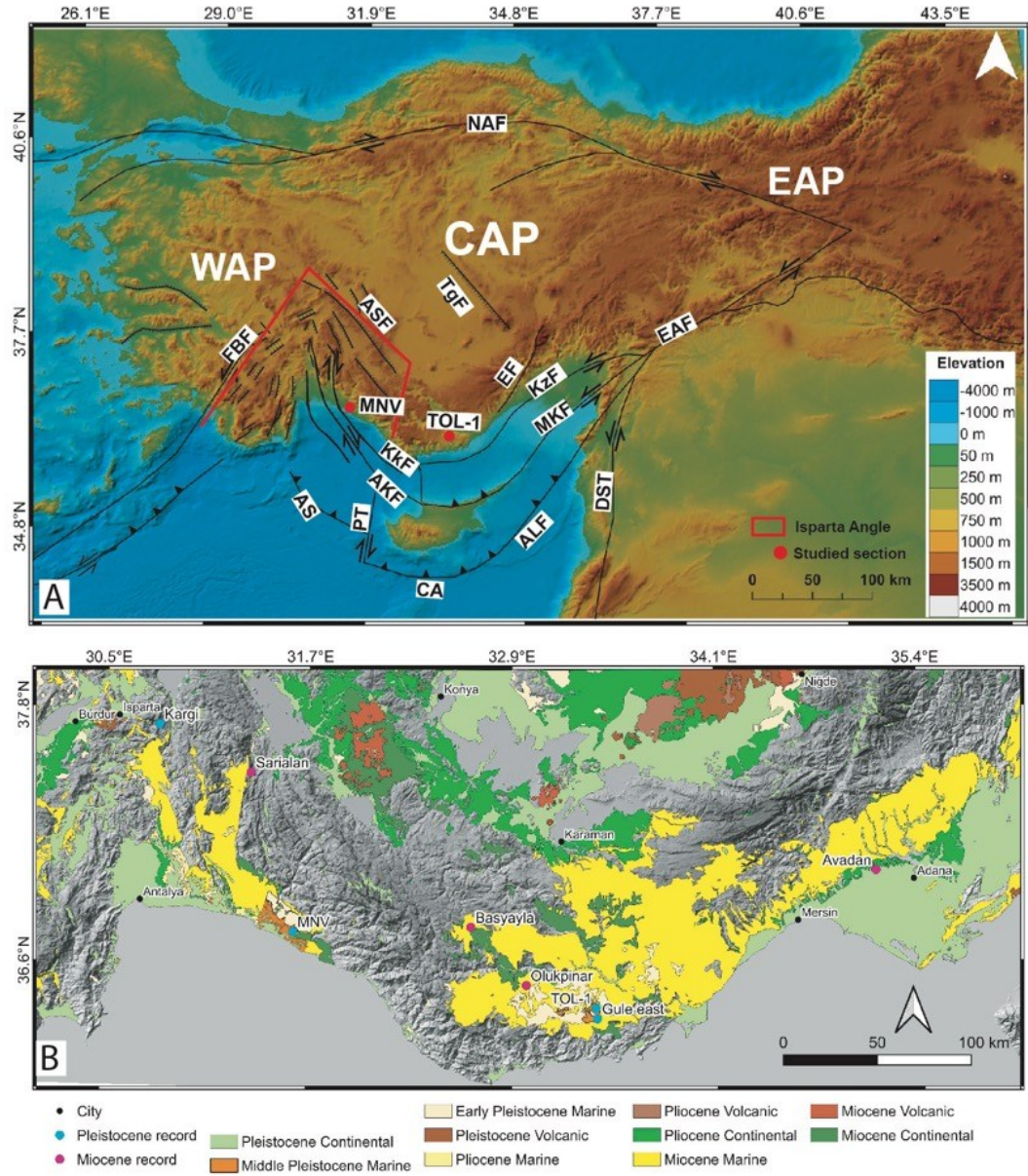


Figure 1. (A) Simplified tectonic map and topography of Anatolia showing major tectonic domains and fault systems. EAP—Eastern Anatolian Plateau; WAP—Western Anatolian Province; NAF—North Anatolian Fault; EAF—East Anatolian Fault; DST—Dead Sea Transform Fault; CA—Cyprus Arc; PT—Paphos Transform; AS—Antalya Slab; KA—Kyrenia Range; AKF—Aksu–Kyrenia Fault; KkF—Kırkkavak Fault; SAF—Silifke–Anamur Fault; EF—Ecemiş Fault; TgF—Tuz Gölü Fault; KzF—Kozan Fault; MKF—Misis–Kyrenia Fault; ACB—Adana–Cilicia Basin. (B) Geological map of the southern margin of the CAP (modified the geological map of Turkey (1:500.000) and after [Sarı et al., 2025a, 2025b](#)) showing the distribution of Miocene to Pleistocene deposits and the locations of documented marine sequences from the literature.

3.3. RESULTS

3.3.1. Late Middle Pleistocene marine deposits from Mut and Antalya basins

At the CAP southern margin, the youngest unconformable marine deposits capping the tectonic units of the Central Taurides, believed to be Middle Miocene in age

(Özgül, 1976; Dhont et al., 1999; Şenel, 2002), have been used to constrain the uplift of the CAP southern margin. Recently, they were significantly rejuvenated by the recognition in the Gülnar East and Tol sections (Mut Basin) of some Calabrian and Chibanian calcareous nannofossil and planktonic foraminiferal bioevents. Based on these findings, it was possible to reconstruct a deep marine environment that persisted at least until part of the Middle Pleistocene (ca. 0.45 Ma, Öğretmen et al., 2018).

Based on new results from analyses of calcareous nannofossil and planktonic foraminiferal associations conducted on new stratigraphic sections of the Mut Basin (TOL-1 section) and Antalya Basin (MNV1-2 and MNV4 sections), it has been possible to further rejuvenate the age of the marine deposits present in these two basins and, consequently, the uplift event that caused the entire southern Turkey to emerge (Sarı et al., 2025a, 2025b). In the Mut Basin, the TOL-1 section preserves the world's highest late Middle Pleistocene marine deposits at 1177 m a.s.l., with the last occurrence of *Pseudoemiliana lacunosa* (~0.46 Ma) recorded mid-section and the first occurrence of *Emiliana huxleyi* (~0.26 Ma) near the top. The inferred sedimentation rate for the upper portion of the TOL-1 section points to extend the age of the marine deposition up to 0.23 Ma. Foraminiferal biostratigraphy independently constrains the base of the section for the occurrence of *Globigerinella calida* since the basal samples. This finding indicates that the TOL-1 deposition post-dates the first occurrence of the *Globigerinella calida* (~0.53 Ma, Lirer et al., 2019). Ostracod assemblages documenting an oscillating shallowing-upward trend, from ~400 m to ~50 m water depth, was hypothesized to reflect gradual uplift beginning around 0.40 Ma. Integration of micropalaeontological, palaeomagnetic, and palaeobathymetric data indicates ~1245 m uplift, at rate of ~5.2 m/kyr (Fig. 2; Sarı et al., 2025a). In the Manavgat Basin, the MNV composite fan-delta succession, with a top at 143 m a.s.l., contains *E. huxleyi* starting from the base, which indicates ages younger than the FO of *E. huxleyi* (0.26 Ma). Considering that a regional uplift event was responsible for the surfacing of both the Mut Basin and the Antalya Basin, we speculate that the top of the Manavgat fan-delta succession (MNV composite section) may also have an age of 0.23 Ma (top of the TOL-1 section). The result indicates ~181 m uplift with a rate of ~0.8 m/kyr (Fig. 2; Sarı et al., 2025b).

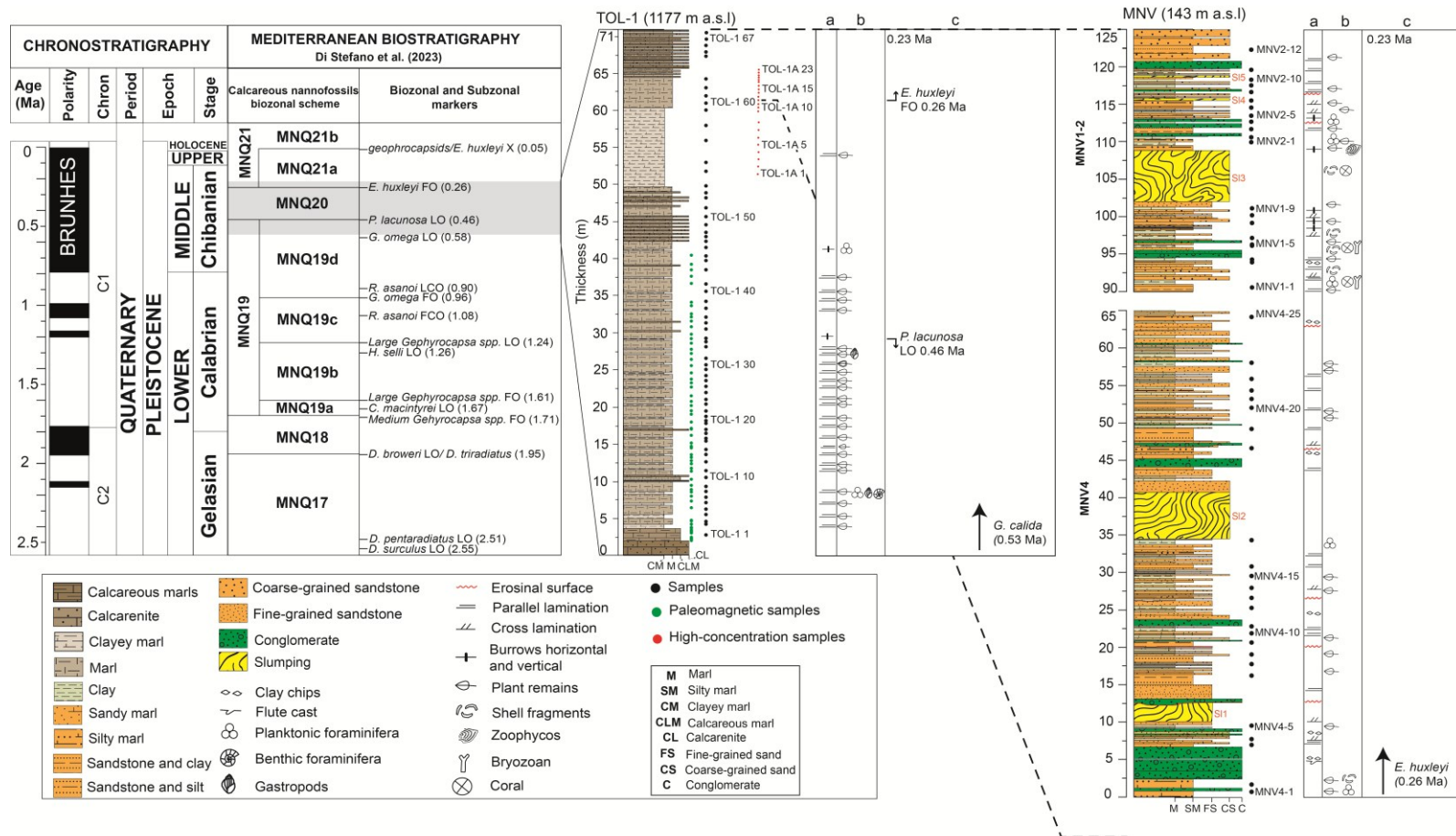


Figure 2. Integrated Mediterranean Pleistocene biostratigraphic framework (modified from Di Stefano et al., 2023) showing astronomically calibrated magnetic reversal chronology. Key calcareous nannofossil bioevents include first and last common occurrences (FCO, LCO) of diagnostic taxa. The grey band marks the inferred age interval of the Tol-1 section; the dashed line indicates the correlative level for the MNV section. (a) sedimentary structures, (b) fossil components, and (c) stratigraphic positions of key calcareous nannofossil and foraminiferal biostratigraphic events. Lithological and structural symbols for both sections are shown in the legend.

3.3.2. Integration with geomorphological analysis

Incorporating the new TOL-1 constraints (0.23 Ma / 1177 m a.s.l.) into the LEM, we correlate seven marine terraces (T1–T7; 30–1400 m a.s.l.) with major sea-level highstands (modified after [Racano et al., 2020](#)). The model identifies a pronounced Middle Pleistocene uplift pulse peaking during MIS 7 (~180 ka), with rates of ~6.2 m/kyr, followed by deceleration to 1.5–1.1 m/kyr at present (Fig. 3).

To refine the spatial and temporal patterns of uplift, we couple this terrace-based model with linear inversion of river profiles from seventeen catchments draining the southern CAP margin (modified after [Racano et al., 2021](#)). Both methods independently show that the kilometre-scale growth of the plateau margin was temporally coherent. Maximum uplift is centred between the Gazipaşa and Anamur districts (Fig.3).

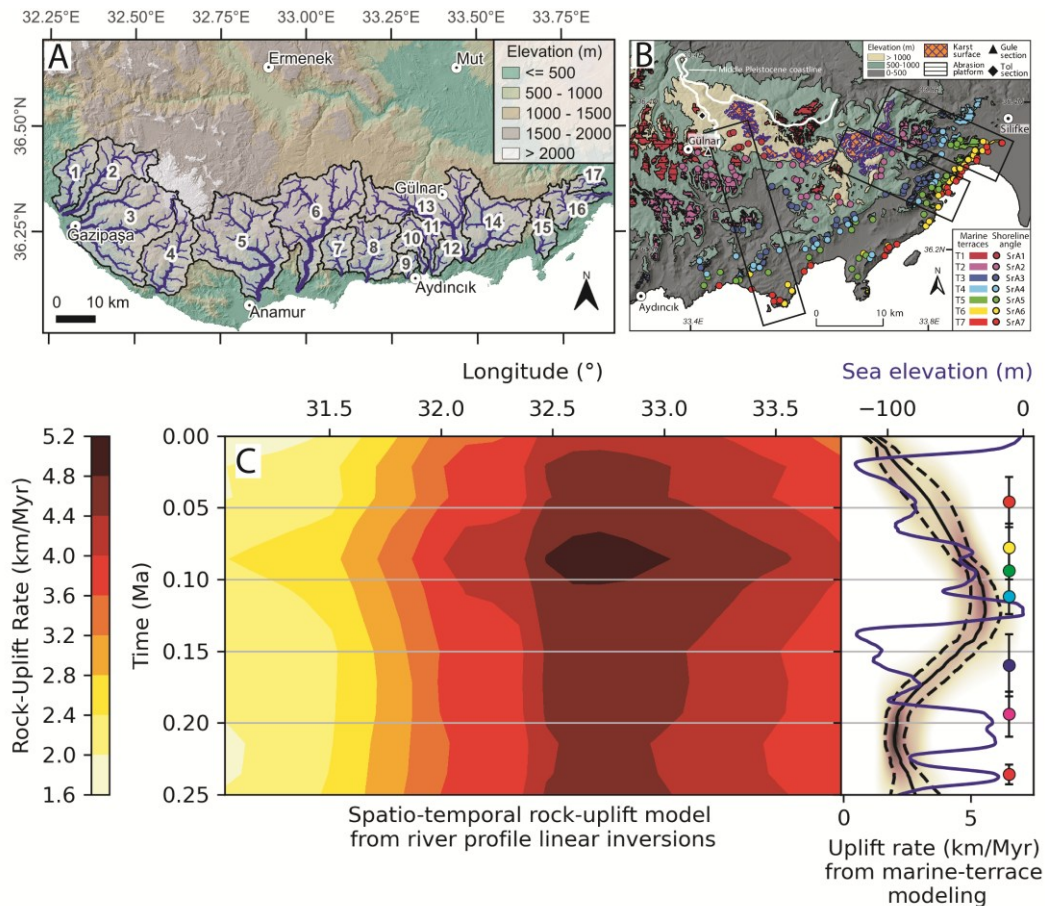


Figure 3. A) Locations of seventeen catchments used for river-profile analysis. (B) Classified Quaternary marine terraces, providing reference surfaces for uplift and sea-level correlation. (C) Spatio-temporal rock-uplift model integrating river-profile linear inversions (filled-contour plot) with marine-terrace modelling (right vertical plot), illustrating uplift rates and their relationship to sea-level fluctuations along the CAP southern margin (modified after [Racano et al., 2020, 2021](#)) Circles with error-bars represent marine terrace ages inferred with LEM simulations, coloured according with (B).

3.3.3. Compiled uplift history along the CAP southern margin

To place our new Middle Pleistocene uplift constraints in the long-term geodynamic context, we synthesized all available vertical-motion datasets from the CAP southern margin. The combined record shows two distinct phases of subsidence and two major uplift pulses since the late Early Miocene. Early subsidence occurred from the late Burdigalian to the early Messinian (ca. 17.0–6.6 Ma) with the deposition of the Köseleli and Mut formations, up to the 6.7–6.8 Ma abrupt shift in lithofacies patterns from carbonate (Mut Fm) to sapropelitic/diatomitic deposition of the Tripoli-like Fm of the Sarıalan section (Schildgen et al., 2012b). This first subsidence phase developed with average rates of ~ 0.03 m/kyr (Cosentino et al., 2012; Schildgen et al., 2012b).

A second subsidence phase affected the CAP southern margin from the early Zanclean to the Middle Pleistocene (5.3–0.4 Ma), when about 1 km of marine deposits (Sarıkavak Fm), with deepening- and shallowing-upward trends, sedimented on top of the Messinian erosional surface that affected both Mut and Köseleli formations (Öğretmen et al., 2018). This second subsidence phase developed with average rates of ~ 0.22 m/kyr.

These two phases of long-term downward motions were interrupted by two major uplift pulses. The first occurred, with a possible mean uplift rate of ~ 4 m/kyr, at the end of the Messinian salinity crisis (latest Messinian, ~ 5.45 Ma), when the Neogene stratigraphy of the Adana-Cilicia Basin recorded a sedimentary event with ~ 1.7 km thick clastic deposits coming from the erosion of the Central Taurides (Radeff et al., 2017; Öğretmen et al., 2018). A second, more pronounced uplift pulse culminated in the late Middle Pleistocene, when our new constraints on the TOL-1 section indicates rapid uplift at a rate of 5.9 m/kyr, which began ~ 400 kyr ago. Uplift then decelerated to 0.8–1.5 m/kyr in the late Holocene, as evidenced by uplifted sea level markers along the Mediterranean coastline of the CAP between Antalya and Adana (Liberatore et al., 2022). Together, these datasets document a complex vertical-motion history characterized by: 1) two long-term subsidence phases during the Early-Late Miocene (20–8 Ma) and the Early Pliocene-Middle Pleistocene (5.3–0.4 Ma), with the deposition of thick marine successions at the CAP southern margin; 2) two major uplift phases occurred during the late Messinian (~ 5.4 Ma) and the late Chibanian (<0.4 Ma), characterized by severe erosion and river incision, culminating in the fastest rates observed along the CAP margin southern (Fig. 4).

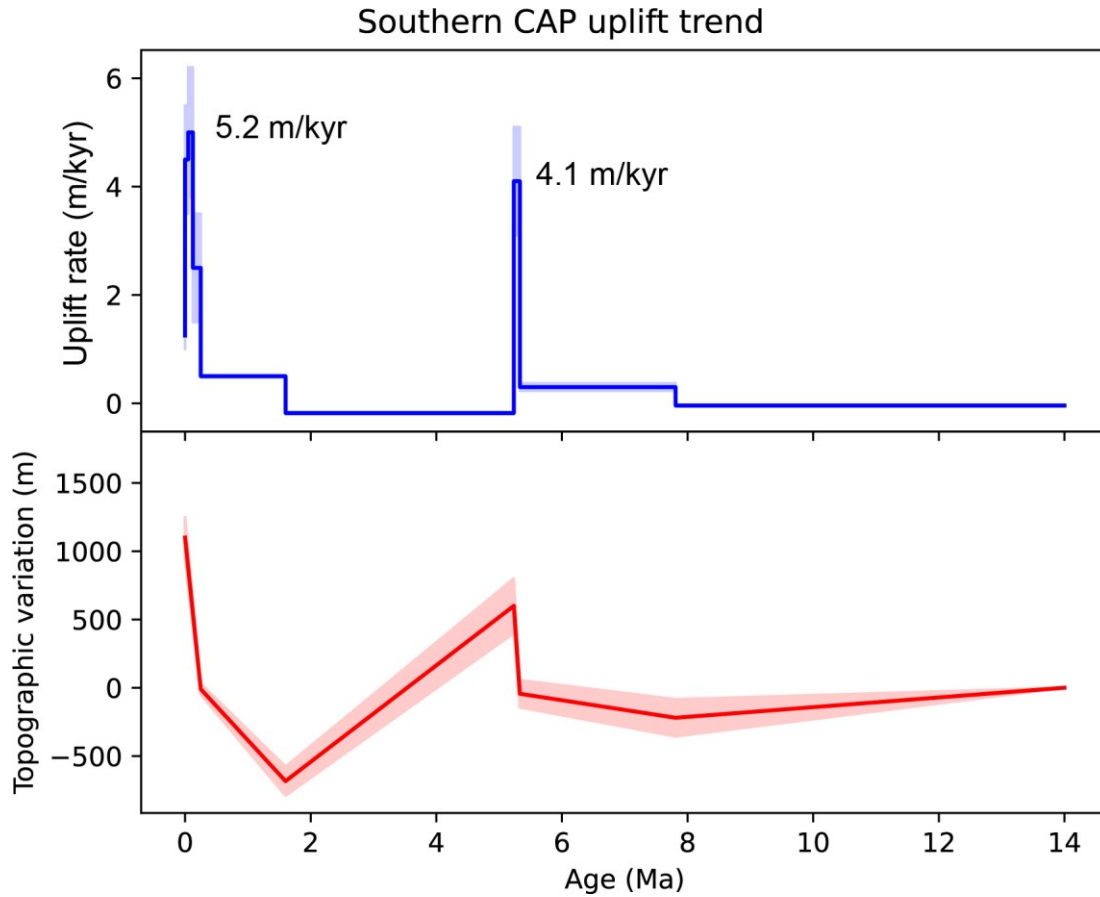


Figure 4. Compiled uplift history along the southern margin of the Central Anatolian Plateau. Uplift evolution from the Early Miocene to the present synthesizing stratigraphic, geomorphological, and geochronological constraints from the Mut Basin, illustrating the timing and magnitude of major subsidence–uplift phases and topographic variation through time.

3.4. DISCUSSION

3.4.1. Differential uplift along the CAP southern margin and the Isparta Angle

New biostratigraphic constraints from the Mut and Manavgat basins substantially refine the timing of the latest marine deposition along the CAP southern margin and the IA. At Mut, the TOL-1 section preserves a marine succession sedimented between ~ 0.53 Ma and ~ 0.23 Ma, currently at 1177 m a.s.l, which requires ~ 1.2 km of post-depositional uplift. This younger age for the top of the marine deposits capping the CAP southern margin, in comparison with previous data (e.g., [Öğretmen et al., 2018](#)), raises the inferred uplift rate to ~ 5.2 m/kyr, highlighting the timing and magnitude of the Middle Pleistocene uplift pulse.

In the Antalya Basin, calcareous nannofossils biostratigraphy demonstrate that the base of the MNV succession is younger than ~ 0.26 Ma, for the presence, since the basal samples of the MNV-4 sub-section, of *E.huxleyi*, whose first occurrence is 0.26

Ma (Di Stefano et al., 2023). As stated before, we consider that a regional uplift event was responsible for the contemporary uplift of both the Mut and the Antalya basins, then we speculate that the top of the Manavgat MNV composite section may also have an age of 0.23 Ma as the TOL-1 section. This revised chronology allows us to better constrain the age of the youngest marine deposits of the Antalya Basin almost 1 Ma younger than previously suggested by Sagular and Özer (2024), extending it to the late Middle Pleistocene. The present elevation of the top of the MNV composite section (143 m a.s.l.) implies a mean uplift rate of ~ 0.8 m/kyr, demonstrating markedly differential uplift between the Antalya and the Mut basins. Geomorphological analysis also reveal this difference (Aykut et al., 2023). Magnitude and timing of uplift of the Mut Basin require a short-lived shallow-depth slab break-off, as already suggested by Ögretmen et al. (2018), whereas the subdued uplift in the Antalya Basin reflects the SW-retreating Antalya slab with dominance of upper-crustal deformation.

3.4.2. Integration with geomorphological analyses

The revised 0.23 Ma age for the youngest marine deposits of the TOL-1 section fundamentally allow us to refine the previous geomorphological interpretations of late Quaternary uplift along the CAP southern margin. Earlier terrace-evolution models (Racano et al., 2020) relied on the TOL deposits (assigned to ~ 0.45 Ma; Ögretmen et al., 2018) to anchor the age of the highest terrace surface (~ 1400 – 1500 m a.s.l.), yielding peak uplift rates of 3.4 – 3.8 m/kyr between MIS 11 and MIS 7. The younger age of TOL-1 shifts the reference horizon forward by >200 kyr, compressing the time available for uplift and requiring substantially faster vertical motions to generate the same terrace staircase. When integrated into the LEM framework, the new chronology produces a sharper, higher-magnitude uplift pulse during MIS 7, with peak rates of ~ 6.2 m/kyr, nearly double previous estimates, and a more abrupt deceleration to Holocene values of 1.1 – 1.5 m/kyr.

River-profile inversion results, when recalibrated to the 0.23 Ma datum, no longer support the previously inferred west–east diachroneity of uplift with peaks at 400 ka (6.1 m/kyr) and 250 ka (4.3 m/kyr); instead, both fluvial and marine terrace datasets indicate a synchronous late Middle Pleistocene uplift pulse across the southern margin, with the spatial maximum centred between Gazipaşa and Anamur. This convergence of geomorphological evidence reveals a temporally narrower, spatially

more focused uplift event than previously recognised, refining both the magnitude and distribution of uplift and strengthening interpretations that the CAP southern margin responded rapidly and coherently to late Middle Pleistocene deep-seated forcing.

New fault-kinematic and U–Th geochronological data from [Aykut et al. \(2025\)](#) independently corroborate this pattern. Their analyses reveal that coeval upper crustal extension and surface uplift in the CAP southern margin has been accommodated by NW–SE-striking normal faults since at least the Middle Pleistocene. Calcite-vein deformation phases at ~0.45 Ma and 0.25 Ma (Aykut et al., 2025) partly overlap with the uplift pulse we reconstruct for the CAP southern margin. Importantly, their structural reconstructions identify the zone of strongest mantle-driven crustal deformation precisely between Gazipaşa and Anamur ([Portner et al., 2018](#)), matching the locus of maximum uplift indicated by our geomorphological datasets.

3.4.3. Geodynamic Implications

The geodynamic evolution of southern Anatolia reflects the interaction between Arabia–Eurasia collision, westward extrusion of Anatolia, and long-lived African subduction beneath the Cyprus Arc ([Dewey et al., 1986](#); [Faccenna et al., 2006](#)). Seismic tomography confirms that the Cyprus slab is segmented into eastern and western domains, separated by the Paphos Transform Fault ([Portner et al., 2018](#); [Güvercin et al., 2021](#)). According to [Portner et al. \(2018\)](#), the eastern Cyprean slab supports a buoyancy-driven slab deformation mechanism and vertically dipping block of the Cyprean slab (anomaly EC2; Fig.5). This segmented geometry governs the strongly contrasting uplift histories of the Mut and Antalya basins ([van Hinsbergen et al., 2010](#); [Göğüş and Pysklywec, 2008](#); [Bartol and Govers, 2014](#); [Meijers et al., 2018](#)). The Mut Basin lies above the eastern Cyprean slab fragment, where our new biostratigraphic and geomorphological constraints indicate two major uplift episodes (Fig. 5): (1) a late Messinian, long-wavelength phase generating ~0.8 km of plateau uplift at rates of ~4 m/kyr, recently interpreted as due to delamination processes ([Schildgen et al., 2012a](#); [Öğretmen et al., 2018](#)), and (2) a late Middle Pleistocene uplift pulse exceeding 1.2 km since ~0.40 Ma, with peak rates of ~6.2 m/kyr. Numerical models predict rapid, short-lived vertical motion (7–9 m kyr⁻¹ for <100 kyr) following slab detachment ([Duretz and Gerya, 2013](#); [Magni et al., 2017](#)),

matching the timing and magnitude of uplift recovered from TOL-1 terrace correlation and river-profile inversion. These results require a younger and shallower break-off in the eastern Cyprus slab that opened a transient slab window and facilitated focused asthenospheric upwelling beneath the Mut Basin by ~ 0.18 Ma.

Magmatic evidence supports the idea of mantle influx: at Karadağ (~ 90 km north of TOL-1), adakitic andesites at 825 ± 17 ka, absarokite–shoshonites at 393 ± 153 ka, and adakitic dacites at 45 ± 6 ka reflect variable-depth melting of a metasomatized mantle–lower crustal source during progressive lithospheric disruption and enhanced mantle influx (Çoban et al., 2022). Together, uplift and volcanism define a coherent signature of east–west–propagating slab tear within the African slab the initial Bitlis detachment fronts advanced toward the Cyprus segment (e.g., Biryol et al., 2011; Cosentino et al., 2012; Schildgen et al., 2014; Rabayrol et al., 2019).

In contrast, the upper plate of the western Cyprus (Antalya) slab, which hosted the development of the Neogene Antalya Basin, shows lower vertical movements due to a different geodynamic response to the subduction processes. In the western Cyprus subduction system, the subducting slab is still continuous, as shown by P-wave tomography down to 600 km (Portner et al., 2018).

In the Antalya Basin, persistent NE–SW extension, dense brittle faulting (e.g., Kırkkavak Fault Zone), and topographic dissection reflect a neotectonic stress field shaped by slab rollback, inherited crustal weaknesses, and differential motion between the fast-retreating Hellenic Arc and slow-retreating Cyprus Arc (Glover and Robertson, 1998; Salaün et al., 2012; Howell et al., 2017; McPhee et al., 2018). This kinematic mismatch has long been interpreted as evidence for slab tearing beneath the IA (Dilek and Sandvol, 2009; Güvercin et al., 2021). The evolution of deformation, from Late Miocene contraction to Pliocene–Quaternary strike-slip and normal faulting, records the shift from convergence-dominated to extension-dominated tectonic regimes (Glover and Robertson, 1998). Present-day subsidence may further reflect vertical motions induced by time-dependent mantle convection and plate motion across steady convective patterns (Houseman and McKenzie 1982). The westward retreat of the Antalya slab by ~ 150 km since the Eocene suggests that most of the offset was accommodated by small-scale peeling delamination (McPhee et al., 2019), combined with only ~ 25 km of extensional strain since the Miocene (Koç et al., 2018).

Numerical models of this process predict a wave of vertical motions: subsidence at the point of peeling where the mantle lithosphere is still connected to the upper crust, followed by uplift where the lithospheric mantle has peeled-off (e.g., Göğüş, and Ueda, 2018). Quaternary volcanism within the IA is consistent with this more moderate mantle reorganization. Gölcük volcanism defines an entirely renewed Pleistocene cycle decoupled from older Pliocene activity, initiated at ~ 200 ka and continuing through explosive eruptions between 115 ± 3 ka and 62 ± 2 ka, followed by tuff-ring formation from 72.7 ± 4.7 ka to 24 ± 2 ka (Platevoet et al., 2008; Elitok et al., 2010). Unlike the deeply sourced magmas at Karadağ, these eruptions reflect shallow decompression melting during regional NE–SW extension, consistent with a mechanically weakened crust rather than focused asthenospheric upwelling.

Taken together, geomorphological, stratigraphic, magmatic, and tomographic datasets reveal a fundamental geodynamic bifurcation along the southern Anatolian margin. The Mut Basin records uplift and volcanism driven by two-stage slab break-off, with a deep late Miocene detachment followed by a late Middle Pleistocene propagating tear, which was responsible for a km-scale surface uplift that produced the highest late Middle Pleistocene marine terrace known globally. In contrast, the Antalya Basin and IA represent the upper plate of a subduction system characterized by a still attached subsiding slab, where rollback, peeling delamination, and inherited weakness zones-controlled deformation, subsidence, and extension. The segmentation of the Cyprus slab thus provides a unified framework that reconciles the starkly divergent uplift patterns, structural styles, and magmatic histories of the CAP southern margin and the IA (Antalya Basin), while explaining the asymmetric evolution of topography across southern Anatolia.

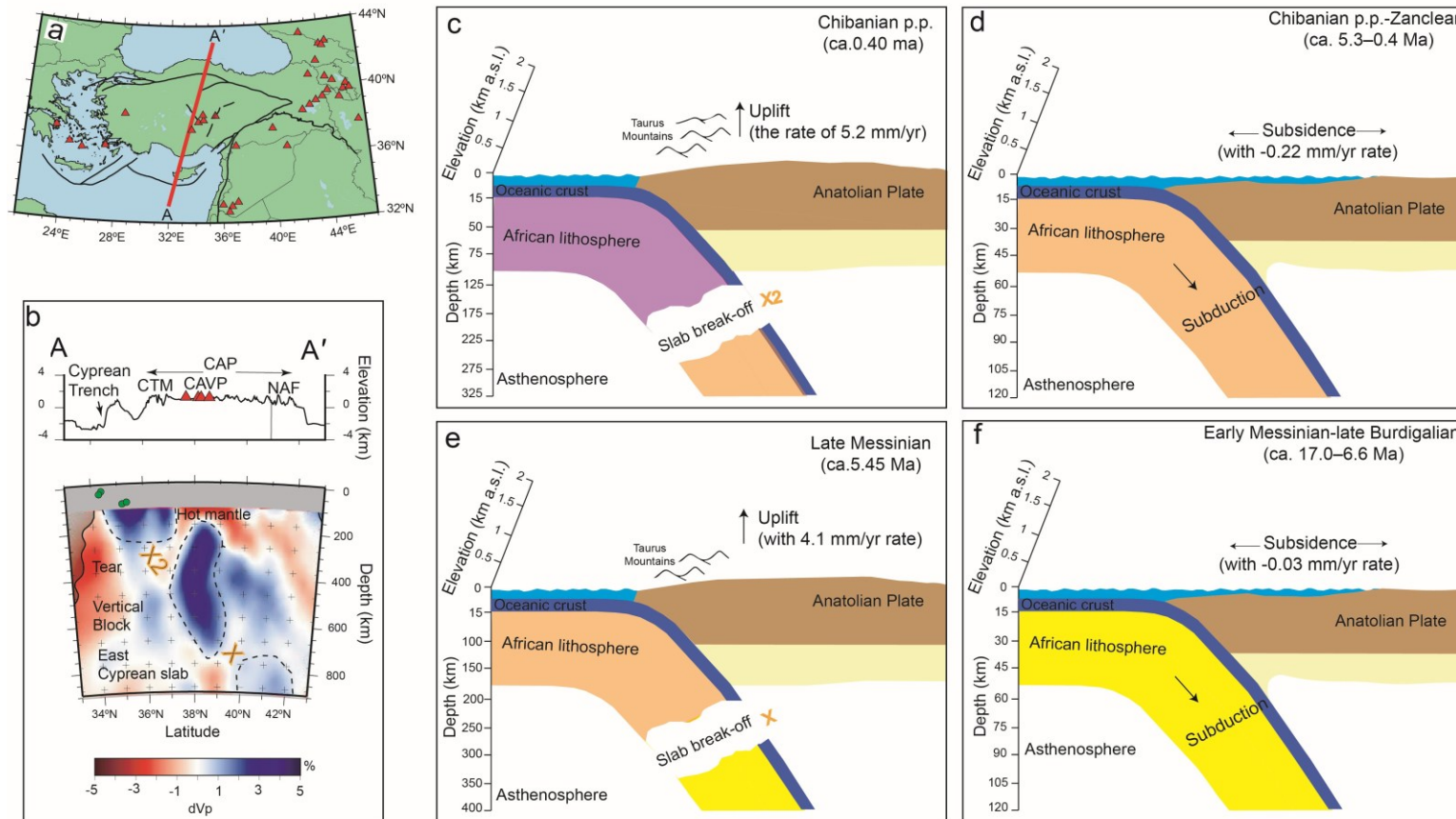


Figure 5. Conceptual model of deep-seated geodynamic processes beneath the southern margin of the Central Anatolian Plateau (CAP) within the Cyprus subduction system. (A) Regional map showing the location of the tomographic cross-section. (B) Tomographic section oriented down-dip of the eastern Cyprus slab, highlighting high-velocity anomalies interpreted as slab segments (Portner et al., 2018). Dashed lines indicate inferred slab geometry; X and X2 mark the inferred locations of slab break-off events. (C–F) Schematic panels illustrating the temporal evolution of subduction, slab segmentation, and associated surface deformation from the Early Miocene to the present. Abbreviations: Bitlis–Zagros Suture Zone (BZSZ), Central Taurus Mountains (CTM), Central Anatolian Volcanic Province (CAVP), East Anatolian Plateau (EAP).

4. GENERAL CONCLUSIONS

This thesis reconstructs the Middle Pleistocene uplift history of southern Anatolia through an integrated stratigraphic and geomorphological analysis of the Mut and Antalya basins. By directly comparing coeval marine successions along the southern margin of the CAP and within the IA, it provides new quantitative constraints on the timing, magnitude, and geodynamic drivers of surface uplift.

New calcareous nannofossil, foraminiferal, and paleomagnetic data significantly refine the chronostratigraphy of the youngest marine deposits in both basins. In the Mut Basin, the TOL-1 section records late Middle Pleistocene marine deposition between <0.53 Ma and 0.23 Ma and is now preserved at 1177 m a.s.l., representing the highest late Middle Pleistocene marine deposits known worldwide. In the Antalya Basin, the Manavgat (MNV) composite section demonstrates that marine deposition persisted until ~ 0.23 Ma, revising previous interpretations that assigned these deposits to the Miocene or Early Pleistocene.

Elevation, paleobathymetric reconstructions, and sea-level estimates indicate markedly different uplift rates for the two basins. Since ~ 0.23 Ma, the Mut Basin experienced exceptionally rapid uplift at an average rate of ~ 5.2 m/kyr, resulting in over 1 km of surface uplift. In contrast, the Antalya Basin records a much lower mean uplift rate of ~ 0.8 m/kyr. The >1 km elevation difference between synchronously deposited marine successions highlights pronounced differential uplift across southern Anatolia during the Middle Pleistocene.

Geomorphological analyses independently confirm these results. Marine-terrace modelling constrained by the new TOL-1 age–elevation data identifies a major uplift pulse along the CAP southern margin during the late Middle Pleistocene, with peak uplift rates during MIS 5.

These contrasting uplift histories reflect fundamentally different geodynamic processes beneath the Cyprus subduction system. The extreme uplift of the Mut Basin is best explained by asthenospheric upwelling and lithospheric rebound following a shallow Middle Pleistocene slab break-off beneath the southern CAP margin. In contrast, uplift in the Antalya Basin reflects continued slab retreat and thermal–viscous relaxation of the upper plate. The sharp lateral gradient in uplift requires decoupling between these domains, likely accommodated by a slab tear separating the Cyprus and Antalya slabs.

Overall, this thesis supports the interpretation that repeated slab segmentation and break-off beneath southern Anatolia, first during the late Miocene and again during the Middle Pleistocene, played a key role in controlling rapid surface uplift. The new stratigraphic and geomorphological constraints are consistent with strong coupling between deep mantle processes and Quaternary surface deformation, while not excluding alternative scenarios involving slab delamination and progressive detachment.

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