The uppermost Pleistocene–Holocene mud drape across the Marmara Sea: quantification of detrital supply from southern Marmara rivers

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Abstract :

The Marmara Sea (area 11,350 km2; volume 3,378 km3; central basins >1100 m deep) straddles the North Anatolian Transform Fault separating the Eurasian and Aegean-Anatolian tectonic plates. Along with the shallow straits of Dardanelles and Bosphorus (depths ~63 m and ~40 m, respectively), the Marmara Sea forms the only marine connection between the Black Sea and the eastern Mediterranean. During Pleistocene glacial stages, the modern straits were subaerial valleys and the modern Marmara basin was occupied by the landlocked Propontis Lake. Previous researchers attributed major portions of a widely distributed uppermost Pleistocene–Holocene mud blanket (locally >10–25 m thick; volume 43–47 km3) to transport of suspended load through one or both of the straits, as either the Aegean Sea (at ~13.8 cal ka) or the Neoeuxine Lake (today's Black Sea, at ~11.1 cal ka) began to spill into the Marmara basin. To test these suggestions, the thicknesses and volume of the mud blanket were determined from >5000 line-km of airgun, sparker and boomer profiles and >100 cores, and compared with the contemporary supply from local rivers to decide, by difference, if the straits might have had a significant role. Volume calculations for the detrital supply from rivers rely on (1) decades of daily water- and sediment-discharge data from gauging stations, acquired before 20th century dam construction and, independently, (2) the BQART model which uses a variety of hydrological, geomorphic, geological and climate data. These calculations demonstrate that >85–90% of the detritus in the offshore mud blanket was supplied by steep rivers (Kocasu River and its tributaries) and mountainous streams draining the highlands of the southern Marmara region. Geochemistry of the <38 μm fraction supports this source. Any input through the Dardanelles has been sporadic and limited to perhaps ~5 Gt of suspended load (equivalent to ~5.2 km3 of porous mud when deposited) because of changing directions and rates of flow since the Last Glacial Maximum. Resedimentation through mass wasting and transgressive shoreface erosion appear to be minor compared with river supply. The isolated nature of the Marmara basin and its supply from mostly a single watershed afford an opportunity to verify the reliability of this type of hindcast analysis, based upon sediment-discharge data and catchment models – analysis which cannot be completed with a comparable level of certainty along open marine coastlines elsewhere.

Highlights

► Latest Pleistocene–Holocene Unit 1 forms a blanket of 43–47 km^3 in Marmara Sea ► Kocasu River and its tributaries can account for 85–90% of the volume of Unit 1 ► Sediment budget calculations rely on data from gauging stations and BQART model ► Isolated Marmara basins allow reliable quantification of the sedimentary budget

Keywords : marine sediment, mud, chemical analyses, sea-level changes, transgression, waterways

1. Introduction

Geomorphologists, sedimentologists and stratigraphers have long sought to understand the various factors which control the rate of delivery of sediment from river drainage basins to the sea (Syvitski and Milliman, 2007, and references therein). These factors include area and relief of the drainage basin; climate as it controls soil development, vegetation cover and water discharge (both its average and extremes); bedrock composition (which controls erosion rate), and human perturbations to any of the above (e.g., deforestation, agricultural practices, damming of rivers). Once the suspended load reaches the sea, it is generally redistributed over wide areas of the adjacent shelf and basin (e.g., Hill et al., 2007; Liu et al., 2009). Coarser sand fractions tend to be retained near the shore in littoral drift cells (Warrick, 2020). Off-shelf transport is greatest when sealevel is at lowstands, although some rivers with direct access to the nearshore heads of modern submarine canyons may continue to transfer sediment into deep water even at highstands (e.g., Sakarya River, Algan et al., 2002; Congo River; Azpiroz-Zabala et al., 2017).

The dispersal of suspended load generally makes it difficult to test and validate hindcast calculations and modelling results for the long-term sediment contributions of modern rivers. The small size and isolated nature of the Marmara Sea, western Turkey (Fig. 1) prevent escape of even the finest detritus. This allows quantification and tracking of the sediment supplied to its offshore since the rising Mediterranean Sea breached the ~ −75 m sill in the Strait of Dardanelles at ~13.8 cal ka (calibrated using Marine13 dataset). This sediment forms a mud blanket (or drape) locally >10–20 m thick, the facies and biota of which are well known from previous studies (Table 1). The Kocasu River draining the Susurluk Drainage Basin (Fig. 2) (Kazancı et al., 2019) accounts for ~80% of the freshwater and ~90% of the suspended-sediment discharge to the Marmara Sea (Aksu et al., 1999). The remainder is almost entirely accounted for by the Biga and Gönen streams which also drain the southern Marmara region (Fig. 2). Small quantities of

freshwater reach the sea via the Kemer creek near the Strait of Dardanelles, and via a number of mostly ephemeral streams which head north of the Marmara Sea, including the Ana, Kınıklı, Kula and Kurbağalıdere creeks (Fig. 1). None of these are significant sediment sources.

The tributaries of the Kocasu River (Fig. 2) have their own names (i.e., the Nilüfer, Orhaneli, Emet, Simav and Kocaçay branches). They coalesce to flow in a single channel through the Karacabey Gorge and into the Marmara Sea. Today, the Kocaçay and Emet branches have their seaward flow interrupted by the small Manyas and Uluabat Lakes, respectively, but this was not the case before ~4.1 cal ka (Kazancı et al., 2004). All tributaries have been dammed to varying degrees since ~1985 (with a few earlier exceptions).

There has never been a quantitative study to assess whether the Susurluk Drainage Basin (i.e., Kocasu River and its tributaries) could have provided the bulk of the mud in the uppermost Pleistocene to Holocene drape across the Marmara Sea. An alternative sediment supply through the straits of Dardanelles (Baştürk et al., 1986; Gökaşan et al., 2010) and Bosphorus (Aksu et al., 2016; Hiscott et al., 2017) has been advocated, but never compared with the local river supply. Past detrital supply from the Susurluk Drainage Basin can be reliably estimated for two main reasons: (1) there are numerous gauging stations along the various tributaries of the Kocasu River (as well as the Gönen and Biga streams), with records of daily water discharge over 30–60 years, as well as sediment-trap data to establish relationships between water discharge and suspended-sediment discharge; (2) although farming is practiced in the lowland area around the Manyas and Uluabat Lakes, the upland areas covering much of the inland portion of the drainage basin are forested with limited human disturbance (Google Earth™ satellite imagery acquired 30 December, 2016), so potentially experience erosion today as they did during the Holocene. With appropriate caveats, the enclosed nature of the Marmara basin might allow a quantitative

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comparison of the long-term sediment supply by rivers and streams of the Susurluk Drainage

Basin to the post-13.8 cal ka mud drape. There is no reason to suspect loss of sediment through

either the Strait of Dardanelles or Strait of Bosphorus (Fig. 1), which are narrow and presently

are controlled by sills only ~63 m and ~40 m deep, respectively, with saline Mediterranean water

moving northward along their floors since ~13.8 cal ka (Dardanelles; Aksu et al., 2016) and ~9.4

cal ka (Bosphorus; Yanchilina et al., 2017; Ankindinova et al., 2019).

In this study, we compare and contrast (1) the volume and fine-fraction (<38 μm)

geochemistry of uppermost Pleistocene and Holocene sediment deposited across the shelves and

basins of the Marmara Sea since its reconnection with the Aegean Sea at ~13.8 cal ka with (2)

the volume of sediment supplied over this same time interval by the Kocasu River and the Gönen

and Biga streams draining the southern Marmara region, as well as the fine-fraction

geochemistry of samples from the modern Kocasu Delta. Other possible sediment sources are

critiqued. The aim is to ascertain whether rivers of the Susurluk Drainage Basin can account for

the offshore mud drape.

2. Setting

The Marmara Sea is a ~220 km long and ~70 km wide east–west-trending inland

depression, developed over the western extension of the North Anatolian Fault zone (Fig. 1). It

has a surface area of ~11,350 km² and a water volume of 3,378 km^3. The Marmara Sea is

connected to the Black Sea and the Aegean Sea via the straits of Bosphorus and Dardanelles,

respectively (Fig. 1). The Black Sea receives continental runoff from European rivers and

exports ~300 km^3 yr^{-1} of brackish water to the Marmara Sea and onward into the eastern

Mediterranean Sea (Polat and Tuğrul, 1996; Özsoy et al., 2001). The present water exchange

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across both straits occurs as a two-layer flow. A cooler, lower salinity (S = 17–20 practical salinity units, hereafter unitless) surface layer 20–30 m thick exits the Black Sea and a warmer, higher salinity (S = 38–39) layer of Mediterranean water flows northward through the straits at greater depths (Polat and Tuğrul, 1996; Özsoy et al., 2001). Shear between the upper and lower layers provides a mechanism to enhance sediment suspension and advection in the Marmara Sea, as does agitation under storm waves. Today, the Marmara Sea is subjected to frequent gale-force winds (Deniz et al., 2013), with 100-year storms generating 23.9 m s^{-1} (or ~86 km hr^{-1}) winds (Öztunalı-Özbahçeci, 2020). Given the N–S fetch distance of ~60–70 km across the Marmara Sea, these 100-year storms can generate average wavelengths of 150–190 m (e.g., Thurman, 1988), with a storm wave base in the range 75–85 m (i.e., half the wavelength; Garrison, 2005).

The Marmara Sea straddles an elongated entire negative flower structure situated above a central master fault (e.g., Aksu et al., 2000). Between the northern and southern shelves and steep fringing slopes, there are four central basins separated by three intervening ridges. The western Tekirdağ and Central basins are elongate, southwest-trending rhombohedral depressions where water depths exceed 1100–1200 m (Fig. 1). The easternmost Çınarcık Basin is a west–northwest-trending elongate depression with water depth >1200 m. The smaller and significantly shallower Kumburgaz Basin (~800 m deep) is situated on the broad southwest-trending ridge separating the Central and Çınarcık basins (Fig. 1). A fifth basin outside the central zone is a shallow (~370 m deep), crescent-shaped depression perched high on the southern slope of the Çınarcık Basin, called the İmralı Basin (e.g., Sorlien et al., 2012). The three ridges which separate the Tekirdağ, Central and Çınarcık basins have water depths generally shallower than 600 m. The shelf edge occurs at water depths of 90–100 m around the Marmara Sea, and very steep slopes of 10º–30º lead to the abyssal depths of the central basins. The northern shelf is

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narrow (3–10 km wide), whereas an approximately 20–30 km-wide shelf occurs across the southern Marmara Sea (Fig. 1). The former is characterised by eroded Tertiary bedrock overlain by a very thin veneer of Holocene sediments, whereas the latter is characterised by a thick succession of Miocene–Recent stacked and north-prograded delta wedges (Aksu et al., 1999; Sorlien et al., 2012; Vardar et al., 2014).

The Marmara Sea experienced a lowstand from ~25–13.8 cal ka (e.g., Aksu et al., 1999, 2016; Yaltırak et al., 2002), when the global ocean was lower than the floor of the Strait of Dardanelles (Lambeck et al., 2007). Lowstand terraces at ~ –85 m elevation (Çağatay et al., 2009) and down-stepping of southern shelf deltas to an ~ –90 m topset–foreset transition point (Aksu et al., 1999) pin the lowest stand of the Marmara lake to an elevation of ~ –85 m to ~ –90 m, which is 10–15 m below the latest Pleistocene sill elevation of –75 m in the Strait of Dardanelles (Aksu et al., 2016). The drawdown of the lake level would only be possible if evaporation exceeded the input from local rivers and streams during the latest Pleistocene.

The Susurluk Drainage Basin (~22,000 km^2) consists of two distinct regions: (a) a rugged upland reaching elevations of >2500 m and drained by tributaries of the Kocasu River, and (b) a lowland depression (~1060 km^2) confined on its northern side by Karadağ Mountain and an adjacent prominent line of hills, with a single exit through the Karacabey Gorge (Kazancı et al., 2019) (Fig. 2). The depression has an elevation less than +15 m and, today, is a swampy inland extension of the Kocasu delta plain (together, they cover ~1350 km^2). Through the latest Pleistocene and early Holocene, the depression was an area of bypass until the level of the Marmara Sea approached its current highstand at ~7.0 cal ka (Lambeck et al., 2007). Subsequently, floodplain deposits started to infill the depression, and at ~4.1 cal ka (Kazancı et al., 2004) lacustine muds started to accumulate in what are now the Manyas and Uluabat Lakes.

Late Holocene mini-deltas have developed where branches of the Kocasu system enter the lakes. Hence, these lakes have been traps for both bedload and some suspended load since ~4.1 cal ka. Kazancı et al. (1999, 2004) documented upper Holocene sediments ~11 m thick and ~8 m thick in the Manyas and Uluabat Lakes, respectively, as well as 5–40 m thick fluviatile sediments across the surrounding lowland depression.

The bedrock in the Susurluk Drainage Basin is extremely varied (Fig. 3). Ehrmann et al. (2007) attributed high proportions of smectite in the western Marmara Sea to weathering and erosion of the widespread volcanic rocks in the southern Marmara region. Kazancı et al. (2014) linked high levels of boron in sediments of Lake Uluabat to valley incision into economic deposits crossed by the Emet branch of the Kocasu system. Although carbonates are not abundant in the Susurluk Drainage Basin, $CaCO_3$ accounts for 6–7% of the fine-grained infill of Lake Uluabat at core site LV 10 (Fig.2) (Kazancı et al., 2014).

3. Data and methods

3.1. Seismic reflection profiles and cores

This paper is based on (a) ~3300 line-km of single-channel airgun profiles with ~135 ms (~100 cm) vertical resolution and ~100–300 ms maximum penetration at shelf depths to ~500–900 ms penetration in deep basins, (b) 1940 line-km of Huntec deep-tow-system (DTS) boomer/sparker profiles with ~20–25 ms (~15–20 cm) vertical resolution and ~20–40 ms maximum penetration at shelf depths to ~50–90 ms penetration in deep basins, and (c) visual description of 74 gravity and 35 piston cores (Fig. 4). Piston cores were collected using a Mooring Systems Inc™ Kullenburg-type piston corer (original Benthos™ design) with split piston option, 1000 kg head weight, triggered ~4.5 m above the sediment-water interface using a

50 kg trigger weight. Core-table descriptions recorded texture, colour (Geological Society of America, 1984), approximate carbonate content from reaction to 1N HCl, physical sedimentary structures, bioturbation intensity, ichnotaxa wherever possible, mollusc and (rarely) echinoderm genus and species identity. The airgun profiles were acquired using a 40 cubic inch (655 cm^3) sleeve gun with the reflections received by a 21-element 9 m long streamer and filtered to 150–1500 Hz. The Huntec DTS profiles were collected using a 500 J boomer or periodically a 1000 J multi-tip sparker source, with echoes recorded using a single internal hydrophone as well as a 21-element 6 m long streamer and filtered to 600–6000 Hz. Navigational fixes from GPS (global positioning system) receivers were recorded every 10 minutes while surveying at a ship speed of ~10 km hr^{-1}. The two-way travel time was converted to sediment thickness using an acoustic velocity of 1500 m s^{-1} (consistent with Dal Forno and Gasperini, 2008).

The volume of the uppermost Pleistocene to Holocene Unit 1 was determined by tabulating and then contouring the thickness of the unit at each navigation fix (or fraction of a navigation fix where additional detail was needed). The contouring was done using two methods: (a) applying Global Mapper™ software to the xyz dataset (latitude, longitude, thickness) and (b) hand-contouring in AutoCad™ after the xyz data were imported into MapInfo™ and thicknesses placed at navigation points as tags. In both cases the volume of sediments in Unit 1 was determined as a last step by Global Mapper™. The two methods differed in their volume estimates by <5%.

3.2. Elemental composition of silt-to-clay fractions – laser ablation ICPMS

To compare the Kocasu source and both shelf and basinal examples of Unit 1 muds, 38 medium-grained silt to clay samples (<38 μm) were taken from cores acquired in 2002, and

seven samples were collected in 2017 on the subaerial top of the modern Kocasu Delta (samples M17-1 to M17-7; Fig. 2). The cores recovered typical Unit 1 facies (Table 1) so are not described here. Geographic coordinates, sample depths and ages of the deepest Unit 1 sample in each core are given in Supplementary material 1. Ages are based on calibrated radiocarbon dates (Supplementary material 2) and confirm that the samples come from Unit 1. At each sample depth, ~25 cm^3 of mud was extracted, treated with 10 mL 15% H_2O_2 and wet-sieved through a 38 μm sieve. The fine fractions were collected in 1 L beakers, and the suspensions were flocculated using 20 mL of saturated $CaCl_2$ solution. The excess water was decanted and the residues were transferred into 500 mL beakers and acidified using 25–50 mL 10% HCl to remove carbonates. The carbonate-free residues were rinsed several times with settling accelerated using a centrifuge. They were oven dried at 40ºC and powdered in the Memorial University of Newfoundland rock crushing facility using titanium rings. The powdered samples were sent to Actlabs^{TM} (Ancaster, Ontario, Canada) for the preparation of lithium borate glasses (or fused disks) for laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) analyses. Small chips from each fused disk were mounted on glass slides using double-sided sticky tape. For the LA-ICPMS analyses, a 193 nm ArF Excimer GeoLas™ laser system (Coherent, Germany) was coupled to an Element XR™ (Thermo Fisher, Germany). The laser ablation system at Memorial University of Newfoundland is equipped with a cylindrical ablation cell with a volume of ~10 cm^3 capable of holding one sample slide at a time. NIST SRM 610 was used as the primary standard and BCR-2G as the secondary standard.

The laser was operated with a repetition rate of 4 Hz, an energy density of 4 J cm^{-2} and five pre-ablation shots were carried out on each sample. Pre-ablation shots are used to remove the upper layer of each glass chip. The carrier gas flow was 1L/min He and the ICPMS was

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optimised for high sensitivity and a low oxide formation ratio (i.e., a ThO+/Th + ratio of <0.2 %). A crater size of 40 μm was used during the ablations. In each sample, three spots were ablated. The concentrations of 55 elements (Al, Ag, As, Ba, Ca, Cd, Ce, Co, Cr, Cs, Cu, Dy, Er, Eu, Fe, Ga, Gd, Ge, Hf, Ho, In, K, La, Lu, Mg, Mn, Mo, Na, Nb, Nd, Ni, P, Pb, Pr, Pt, Rb, Re, Sb, Sc, Si, Sm, Sn, Sr, Ta, Tb, Th, Ti, Tm, U, V, W, Y, Yb, Zn, Zr) were measured.

3.3. River and stream loads

The Turkish General Directorate of Waterworks (EIE, 1982–2011) has collected daily water discharge and sediment load data at strategic points along various tributaries since the 1950s. The EIE databases (EIE, 1982, 1987, 1993, 2000) relate water discharge (units $m^3 s^{-1}$) to suspended sediment load (units tonnes day^{-1}), but the only data which are useful for this study are from un-dammed water courses. Fortunately, the EIE measurements started before post-1985 dam construction. The 1972–1985 data from the Gönen stream (147 measurements), the 1971–1990 data from the Kocaçay branch of the Kocasu River (219 measurements) and the 1964–1979 data from the Emet branch (141 measurements) pre-date damming, and allow fitting of best-fit curves which can be used to estimate average loads of each watershed in the southern Marmara region (as tonnes yr^{-1}) and long-term contributions to the offshore area. On log-log plots of water discharge versus sediment load, data points define trends which can be summarised by second-order polynomial trends (Fig. 5), indicating that high-discharge events transport particularly high amounts of suspended sediment. Bedload is not accounted for in EIE datasets, so an increment is required to estimate the total sediment load. Data for similar watersheds elsewhere (e.g., Turowski et al., 2010) suggest an increment of ~20%, which has been applied in this study.

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Estimates of annual suspended-sediment contributions to Unit 1from the Kocasu, Gönen and Biga watersheds were calculated using long-term records of daily discharges. Day-by-day conversions to sediment discharges capture as much short-term variation as possible, for example the high sediment loads characteristic of flood events. Daily water discharges were reduced to 67% of modern measurements before 11.7 cal ka to acknowledge a drier climate during the Younger Dryas (Mudie et al., 2002a; Valsecchi et al., 2012). After accounting for bedload, the result (as solid mineral mass, in Gt) was reduced by 60% to recognise the loss of a significant amount of river detritus to floodplains and delta construction before the remainder escapes to offshore, marine areas. The figure of 60% is from Kukal (1971, p. 38), who argued that only ~40% of the river supply successfully bypasses floodplains and the delta proper to contribute to contemporary marine successions.

A second, independent method of estimating sediment contributions from rivers and streams is provided by a global database of river loads and yields. Equation 1, from Syvitski and Milliman (2007, their equation 7a) estimates the annual suspended-load discharge (Q_s, [Mt yr^{-1}]) of a river when a number of parameters are known or can be reliably estimated. In situations without anthropogenic effects, the difference between BQART estimates and directly measured values of annual load are ~5–7% of the estimate. The relationship is given by:

$$Q_s = 0.0006 \ B \ Q^{0.31} \ A^{0.5} \ R \ T \qquad\qquad (1)$$

where B is a substrate factor set to 1.0 or 1.2 in the Susurluk Drainage Basin based on the bedrock composition (Konak, 2002; Türkecan and Yurtsever, 2002) (Fig. 3), Q is the annual water discharge [km^3], A is the area of the drainage basin [km^2], R is the maximum drainage-basin relief in headwater areas [km], and T is the annual average temperature [°C]. Areas (A)

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drained by individual branches of the Kocasu system and the Gönen and Biga streams were

digitised from topographic maps; the maximum relief (R) was measured in cross sections from

upland areas acquired in Google Earth™ (December 30, 2016 imagery) using the pull-down

menu Ruler>Path>Profile, and values of average annual temperature were taken from Vidal et al.

(2010) and Valsecchi et al. (2012). As with calculations based on gauging-station data, daily

water discharges were reduced to 67% of modern measurements before 11.7 cal ka to

acknowledge a drier climate during the Younger Dryas (Mudie et al., 2002a; Valsecchi et al.,

2012), a bedload increment of 20% was applied to most calculations (see specifics in Results),

and only 40% of the resulting solid mass is estimated to have escaped the delta front to reach the

open sea (and Unit 1 depocentres).

Comparison of river contributions to Unit 1 sediment volumes requires conversion of

solid mineral masses (in Gt, for river loads) to equivalent volumes of porous mud (km^3,

consistent with seismic mapping of offshore successions). We first converted the sediment loads

[Gt] to solid volume [km^3] using an average assumed grain density of 2.65 t m^{-3} (silicate

bedrock), then expanded this to a notional equivalent volume of porous sediment having a

porosity of 60%, and finally incremented the result by an additional 10% to recognise the fact

that marine Unit 1 contains ~10% biogenic carbonate not supplied by rivers (e.g., nannofossils,

foraminifera; Çağatay et al., 2000; Ergin et al., 2007). This is in addition to the ~6–7% detrital

carbonate carried in the suspended load of the Kocasu River (Kazancı et al., 2014). The 60%

porosity is directly calculated from measurements of bulk density reported by Dal Forno and

Gasperini (2008) for core IM-05 acquired in 153 m of water ~8 km east of İmralı Island (Fig. 4).

There, Unit 1 is ~2.4 m thick and bulk density is 1.7 t m^{-3}. With a pore-water density of 1.028 t

m^{-3} and grain density of 2.67 t m^{-3} (silicates and minor carbonate), this value of bulk density translates to a porosity of 59%.

4. Results

4.1 Character and volume of uppermost Pleistocene to Holocene Unit 1

This paper deals entirely with the uppermost Pleistocene to Holocene sedimentary unit which was deposited after saline Mediterranean water entered the Marmara Sea at ~13.8 cal ka. Following Aksu et al. (1999), this is designated as Unit 1. The same unit has been recognised by Evans et al. (1989; their bioturbated muds), Wong et al. (1995; their unit 4), Smith et al. (1995; their unit A), Gökaşan et al. (2008, 2010, 2012; their unit 1), Çağatay et al. (2000; their unit 1), Çağatay et al. (2009; their seismostratigraphic unit C1, P1, lithologic unit L1), Eriş et al. (2007; their unit 1), Eriş et al. (2011; their seismostratigraphic unit S1, lithologic unit L1), Eriş et al. (2012; their unit 2), Vardar et al. (2014, their unit C3). Although authors have used different names in their publications (e.g., Çağatay et al., 2000, 2009; Eriş et al., 2007, 2011, 2012) the descriptions and the basal ages assigned to these units indicate that they are synonymous with the 'transparent' drape (i.e., Unit 1) of this paper (characteristics in Table 1). Unit 1 has also been recognised in Calypso giant piston cores acquired by the *RV Marion Dufresne* from the deep Marmara Sea basins as a marine hemipelagic unit overlying a pre-Holocene lacustrine succession (e.g., Beck et al., 2007, 2015; Campos et al., 2013). Above the ~ −85 m level of the last lowstand, Unit 1 sits on a transgressive surface (called α1) which returns a strong, high amplitude reflection because of subaerial exposure during the lowstand. This simplifies mapping of the drape unit. In basinal areas, cores confirm the base of the post-transgressive mud with the arrival of euryhaline to stenohaline organisms like foraminifera, or chemical and mineralogical

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changes (e.g., calcite spike of Reichel and Halbach, 2007; see also Beck et al., 2015). On steep basin-margin slopes, Holocene sediments are thin or absent because of unstable conditions which promote sliding and other mass flow processes (Beck et al., 2007; Zitter et al., 2012).

Away from the modern coastline, from active deltas of the southern basin margin (Aksu et al., 1999; Hiscott and Aksu, 2002), from the Holocene outflow delta immediately south of the Strait of Bosphorus (Hiscott et al., 2002, 2007a; Aksu et al., 2016), and from the saline inflow channel at the eastern exit of the Strait of Dardanelles (Aksu et al., 2018), Unit 1 is characterised by laterally continuous, high-frequency, acoustically weak reflections (leading to a 'transparent' appearance) which can be traced in Huntec DTS profiles across shelf portions of the entire study area (Fig. 6). This internal seismic architecture suggests that Unit 1 is composed predominantly of fine-grained sediments with thin beds of slightly coarser or denser material.

Across the southwestern Marmara Sea, Unit 1 onlaps the prominent erosional unconformity α1 (Fig. 6). The hiatus at α1 must be several millennia wherever it is mapped at elevations shallower than –75 m, because the underlying successions had to be subaerially exposed at those localities since marine isotopic stage (MIS) 3 (ending ~29 ka). This is the last time that the base level in the Marmara basin could have exceeded the sill depth in the Strait of Dardanelles because the base level in the Aegean Sea was in the vicinity of –30 m to –40 m (Skene et al., 1998; Svitoch et al., 2000).

An isopach map of Unit 1 (Fig. 7) allows the calculation of its volume (as sediment with water-filled pores). Graphic logs of Memorial University of Newfoundland cores (Fig. 8) and *Marion Dufresne* (Calypso) cores (Fig. 9) show the extent of thickness variations within Unit 1. Unit 1 is 15–25 m thick in the Tekirdağ, Central, Kumburgaz and Çınarcık basins along the central deep axis of the Marmara Sea (Figs. 7B, 9). It is equally thick within the İmralı Basin

perched on the southeastern slope. Across the southern shelf, Unit 1 is distributed as a broadly east–west-trending, >20 m-thick narrow belt which encircles the present-day mouth of the Kocasu River and which shows evidence of progradational clinoforms in water depths less than ~35 m (Vardar and Alpar, 2017, their fig. 5a). This lobe extends north and northwest for a considerable distance onto the southern Marmara shelf. Similar, but distinctly smaller and thinner prodelta lobes occur seaward of the mouths of the Gönen and Biga streams. Unit 1 is only ~5 m thick in Gemlik Bay.

The thickness of Unit 1 varies from 7–15 m along the northern and southern levées of the Dardanelles channel to <5 m across the northwestern portion of the Strait of Dardanelles (Fig. 7B) (Aksu et al., 2018). It is thickest along the northern levée where Unit 1 defines narrow, broadly east–west-trending, elongated lobes with central thicknesses locally exceeding 25 m. Similar narrow, but thinner lobes (15–20 m) also occur along the southern levée. Unlike broad areas of the middle and outer shelves, there is an internal unconformity (*d*-unconformity, Fig. 10) within Unit 1 in both levées of the Dardanelles channel, and the top of the unit is truncated along the northern side of the channel. These relationships suggest a more complex post-13.8 cal ka history of sedimentation in this area, with possible implications for significant sediment supply to Unit 1.

Atypical Unit 1 facies are also present at the southern exit of the Bosphorus Strait in the northeastern Marmara Sea. An older part of the unit (11.1–10.2 cal ka) constitutes an early Holocene mid-shelf delta characterised in Huntec DTS profiles by an elongate, lobate plan view shape and a climbing topset–foreset transition (Fig. 11). This deltaic wedge has a volume of ~0.53 km^3 and is sandwiched between the oldest muds of Unit 1 (between reflectors β3 and β2 of Aksu et al., 2016) and the post-delta Holocene drape (β1 to seabed, Fig. 11). This lobe

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developed as a result of vigorous Black Sea outflow through the strait (Aksu et al., 2016; Hiscott et al., 2017), and must have introduced prodelta mud at least to that local area and perhaps to the nearby Çınarcık Basin (Fig. 1). If 60% of the load of the Bosphorus outflow was retained in the delta itself (Kukal, 1971), the offshore contribution would have been rather small, at ~0.35 km^3.

For the purpose of a comparison between the volume of post-13.8 cal ka sediment across the Marmara Sea and the sediment contributions of the Kocasu River and the Gönen and Biga streams, the Marmara Sea and the Kocasu delta plain can be divided into five depositional domains: (1) the southern shelf, (2) the southwestern Marmara Sea at the entrance to the Strait of Dardanelles, (3) the northern shelf, including the early to middle Holocene mid-shelf delta at the Bosphorus exit, (4) the deep basins beyond the shelf edge and (5) the Manyas and Uluabat Lakes situated in the low-lying region south of Karadağ Mountain and adjacent coastal hills (Figs. 2, 7A). Except for the finest clay-sized particles in long-term suspension by waves, there is no sediment pathway between the southern Marmara water courses and the northern shelf, including the vicinity of the Bosphorus exit (e.g., Aksu et al., 2016).

The entire volume of post-13.8 cal ka water-saturated sediment in the Marmara basin exclusive of the northern shelf and Bosphorus exit (thus domains 1, 2 and 4) is estimated to be 45.0 km^3 (Table 2). Uncertainty is expected to be of the order of about ±5%, so a reasonable range for this volume is 43–47 km^3. The Manyas and Uluabat Lakes have accumulated an additional 1.1 km^3 (Table 2).

4.2. Quantity of detritus provided by Kocasu, Gönen and Biga watersheds

Two approaches to the quantification of sediment supply from rivers and streams along the southern margin of the Marmara Sea are considered: (1) utilisation of data from gauging

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stations on the Kocasu River and the Gönen and Biga streams to calculate long-term sediment supply, based on the assumption that latest Pleistocene to Holocene sediment loads of the various rivers and streams can be estimated from modern relationships (Fig. 5), and (2) application of the BQART model of Syvitski and Milliman (2007).

4.2.1. River supply using data from gauging stations

The Kocasu River and the Gönen and Biga streams are the only significant point sources in the Marmara Sea, with the Kocasu River dominating the sediment and water discharges. The equivalent volumes of porous sediment expected from these water courses, using pre-dam relationships (Fig. 5), are presented in Table 3 as sums over three time intervals: (a) 13.8–7.0 cal ka (transgression) when all load from the four Kocasu tributaries and from the Biga and Gönen streams escaped the river mouths, (b) 7.0–4.1 cal ka (early highstand) when only 40% of the load from all watersheds is hypothesised to have reached the open sea as the remainder constructed delta lobes and aggraded low-elevation floodplains, and (c) 4.1–0.0 cal ka (highstand) when no bedload from the Kocaçay, Emet and Orhaneli tributaries reached the sea, but other water courses continued to provide both 40% of their suspended load and 40% or their bedload. By this time, a portion of the bedload fraction would have likely remained in nearshore and shelf transport (Warrick, 2020) while the muddy suspended load would have more readily moved offshore to the outer shelf or deep basins.

Decades of data have been incorporated to improve the estimate of average annual suspended loads (e.g., 55 years of Kocaçay data = 20,075 conversions of water discharge to suspended-load discharge). Based on the scenarios explained above, the Kocasu, Gönen and Biga contributions of suspended load and bedload since 13.8 cal ka total ~44.3 km^3 (Table 3).

More than 95% of this is attributed to the Kocasu River and its tributaries (i.e., the Susurluk Drainage Basin). This is essentially identical to the 45.0 km^3 (with uncertainty, 43–47 km^3) required to account for Unit 1 in the Marmara Sea (Table 2).

Furthermore, calculations suggest fthat 14.0–16.8 km^3 of sediment were retained in floodplain and delta top settings, assuming in this case a porosity of 40–50% more typical of continental deposits (Manger, 1963). This can be tested against the measured volume around the Manyas and Uluabat Lakes and in the modern Kocasu Delta and its prodelta. If 14.0–16.8 km^3 of sediment were retained in this area since ~7.0 cal ka, when sea level reached a near-modern elevation, the average sediment thickness over an area of 1,350 km^2 would be ~10.4–12.5 m. This range is very reasonable based on what is known about sediment thicknesses in the southern Marmara onshore region (Kazancı et al., 2019).

4.2.2. River supply using BQART equation of Syvitski and Milliman (2007)

Using the same temporal constraints that were imposed on calculations based on gauging-station data, the BQART procedure gives a total sediment supply of 38.31 ± 2.68 km^3 over the last 13.8 cal ka (Table 4), with 12.2–14.7 km^3 retained on land and in the subaerial portion of the highstand Kocasu Delta. This is 86% of the amount calculated independently using data from modern gauging stations, and strongly supports those results, although in detail the BQART estimates are less than those using station data for most tributaries of the Kocasu River, and are larger for the Nilüfer, Biga, Gönen and Keçi streams (Tables 3, 4). In the remainder of the paper, the 13.8–0.0 cal ka sediment supply from rivers and streams of the southern Marmara region is set to 40 km^3 to simplify interpretations and discussion, and individual tributary and stream

values are taken to be averages of estimates in Tables 3 and 4. Uncertainty in the total supply is probably at least 5% of 40 km^3, so notionally ±2 km^3 of porous mud.

4.2.3. Fine-fraction geochemical linkages, Kocasu Delta to deep basins

Volumes aside, erosion products of the Susurluk Drainage Basin should have a composition consistent with Unit 1 muds to strengthen the case for a source-to-sink linkage. Given the dearth of sand in offshore portions of the mud drape, a geochemical approach has been selected. An exact match in the abundances and ratios of elements cannot be expected for several reasons: (1) although all samples consist of <38 μm silt and clay, the samples surely have different proportions of coarse, medium and fine silt and clay, so likely different adsorption characteristics for certain chemical elements; (2) the modern samples from the subaerial Kocasu Delta are probably contaminated due to agricultural fertilisers and industry in the watershed; (3) in some cases there is considerable scatter in elemental abundances at single core sites, so although the source and sink may have overlapping ranges for particular elements, the mean values may not correspond. For these reasons, comparison of the seven Kocasu Delta samples to the offshore shelf and deep basin samples will focus on general similarity of abundances, overlapping 95% confidence intervals for mean abundances, and in many cases simply an indication that abundances are in the same range.

LA-ICPMS data are available in Supplementary Material 3, except for platinum which showed a high degree of scatter attributed to the use of platinum crucibles for the preparation of the fused disks. Mean values were calculated for (a) the Kocasu delta top, (b) the western shelf drape (cores M02-103P and M02-106P) and (c) deep basins (cores M02-88P, M02-89P, M02-90P, M02-102P). The averages were then sorted from smallest to largest (for the Kocasu set;

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Table 5) and plotted as pairs of log_{10} values to assess how closely the pairs of delta and marine

values follow a straight line (Fig. 12). For both the western shelf and the deep basinal cores,

linear regression coefficients (R^2) exceed 0.97. This suggests an overall good similarity of mean

values across the spectrum of elements, although matches for a few elements are poor (e.g., As,

P, Sb, Mo in deep basin samples, and to some extent Cu and W) (Fig. 12). Arsenic is also high in

modern sediments of Uluabat Lake (Katip et al., 2012) and its elevated level in delta-top samples

M17-1 to M17-2 is similarily interpreted as an anthropogenic effect. Antimony (Sb) likewise is

probably high in modern samples because of use in industry and agriculture (Li et al., 2018).

As a second approach, 95% confidence intervals around means for the Kocasu Delta

samples versus (a) shelf samples and (b) basins samples were calculated with the objective of

looking for overlapping confidence intervals as an indicator of good correspondence between

means. This is similar to, but not as stringent as a two-tailed student t-test with the null

hypothesis of equality of mean values (e.g., Tan and Tan, 2010) but our goal is to demonstrate

general similarities between groups of samples, not that all samples are from the same

population. Overlap of 95% confidence intervals characterises 17 minor, trace and rare-earth

elements when the delta samples are compared with all marine samples (shelf and deep basin

samples together) and an additional nine elements if delta and shelf fine fractions are considered

alone (Table 5). For Kocasu Delta and shelf silt–clay samples alone (i.e., cores M02-103P and

M02-106P), 15 of the 47 minor, trace and rare-earth elements have statistically indistinguishable

means using a 2-tailed student t-test and the assumption of unequal variances.

Overall, there is nothing in the geochemical data to suggest that the Unit 1 silts and clays

could not have been derived entirely from the suspended load of the Kocasu River, but neither is

there a strong tracer element to prove this linkage. Kazancı et al. (2014) point to boron as a

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possible tracer because there are economic borate deposits currently mined in the Emet

watershed, and boron concentrations in lacustrine muds of Uluabat Lake sharply increase ~1000

years ago. However, lithium borate flux was used in the preparation of LA-ICPMS fused glasses,

so boron is not in our dataset. Also, the vast majority of the Unit 1 drape is older than ~1000 CE.

Molybdenum is 28 times more abundant in the average of basinal cores than in shelf

cores M02-103P and M02-106P (Fig. 12 outlier). Otherwise there is only one difference

marginally greater than a factor of two. Samples of Unit 1 from the early Holocene outflow delta

at the southern exit of the Strait of Bosphorus have not been included in this study because of

coarser mean grain size, some problems with authigenic gypsum raising Ca abundances in

several samples, and poor similarity to samples at our other sites, particularly for Re (57×

multiplier), Cd (15× multiplier), Sb (13× multiplier), and As (7× multiplier) (see data in

Supplementary Material 3). However, average Mo is 116× more abundant in the area of the

Bosphorus exit than elsewhere in Unit 1 (average of 38 shelf and basin samples), and 1670×

more abundant than in our 15 shelf samples alone. None of these samples are younger than ~2.5

cal ka, so apparently elevated Mo is a primary source characteristic of detritus from the

Bosphorus outflow.

4.2.4. Denudation rate

Table 4, column 5 provides areas for the watersheds contributing to each gauging station.

These areas pertain to the steeper upstream areas away from the lowland depression occupied by

the Manyas and Uluabat Lakes, and were obtained using Global Mapper™ software. The

denudation rate for the study area can be estimated from the summed solid volumes of suspended

and bedload (Table 4, with 20% increment for bedload and a grain density of 2.65 kg m^{-3})

divided by the total watershed area and duration of erosion (13,800 years). The results are slightly different for gauging-station volumes and BQART estimates, but fall in the range of 0.05–0.06 mm yr^{-1}. This is far less than rates for steep watersheds with high annual rainfall in Japan (Sueoka et al., 2016) where denudation rates can locally reach ~10 mm yr^{-1}. The lower rate in the Susurluk Drainage Basin results from a drier climate.

5. Interpretation

The volume of uppermost Pleistocene to Holocene sediments (13.8–0.0 cal ka) buried across the southern Marmara shelf (9.09 km^3; Table 2) can be accounted for by the Kocasu River (~38.0 km^3 as porous equivalent, averaging gauging-station and BQART results). However, the Gönen and Biga streams (contribution of ~3.3 km^3 equivalent since 13.8 cal ka) cannot account for the volume of contemporaneous sediments in the southwestern Marmara region near the entrance to the Strait of Dardanelles (4.5 km^3, Table 2). Some of the mud in that domain must have been imported from elsewhere – either from the Kocasu system which appears to have generated an excess supply or from the initial entry of Aegean water through the Strait of Dardanelles at the time of the first reconnection of the Marmara Sea to the global ocean (Section 6.2).

The deep Marmara basins contain ~31.4 km^3 of uppermost Pleistocene to Holocene sediments (Table 2). Based solely on mean values, if it is assumed that the Kocasu, Gönen and Biga sediment loads (averaged across Tables 3 and 4) first fully accounted for deposits on the southern and southwestern shelves, then there might be the possibility of a deficit in supply of ~3.5 km^3 in the deep basins since 13.8 cal ka.

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Uncertainties are likely in the calculations, stemming from the underlying assumptions, which might be sufficient to explain the apparent mismatch in riverine supply versus Unit 1 volumes. In this study, modern relationships between sediment loads of rivers and water discharges have been extrapolated back to 13.8 cal ka. However, there is good evidence from around Turkey that 20th century erosion rates are several times higher than previous rates because of intensive agriculture and urban development (Atalay, 1984; Özsoy et al., 2012). This would affect the equations used to calculate sediment loads if the water courses involved were in areas of intensive land use, but would increase rather than decrease the apparent mismatch in sediment volumes if pre-historical sediment yields were lower than today. Fortunately, the gauging stations for this study (Fig. 2) are close to the upstream, forested and little-developed parts of the various watersheds. It is implicit in the use of modern water and sediment discharges that the climate, in particular rainfall, has not varied significantly through the Holocene (whereas Younger Dryas dryness is incorporated into calculations; Tables 3, 4). Although some fluctuations are to be expected, pollen studies by Mudie et al. (2002a) and Roberts (2012) suggest persistent relatively warm and wet conditions in the region by ~9.7 cal ka and stability in these aspects of the climate by ~7.5 cal ka. If the period 13.8–9.7 cal ka was somewhat drier than assumed in the calculations of this study, then the possible deficit of supply to the central basins would be higher, not lower, than calculations suggest.

The fact that gauging station calculations (Table 3) seem to agree, within uncertainty, with the mapped volumes of Unit 1 (Table 2) should not be taken as proof that no other sources of silt and clay existed through the latest Pleistocene and Holocene. Certain critical assumptions on how to translate suspended-load detritus (in Gt) into equivalent volumes of porous marine sediment could be sufficiently incorrect to widen the disagreement between apparent supply and

the Unit 1 volume. The volume itself, calculated from seismic profiles, could be an underestimate. Strict adherence to a porosity of 60%, biogenic carbonate content of 10%, and retention of 60% of river load in floodplains and highstand deltas (Kukal, 1971) can all be questioned, certainly in detail. However, from what is known about marine muds (Kominz et al., 2011), the present average porosity is not likely to be lower than 60%, and anything higher than 60% would reduce (or even reverse) the already small difference in volume estimates between Kocasu, Gönen and Biga output and Unit 1. The same porosity, carbonate requirement and onland retention were applied to BQART estimates of sediment supply (Table 4), yet those are ~15% lower than the estimates based on gauging station relationships. Although no single parameter can explain the differences between Tables 3 and 4, it is our view that an estimate in the lower range of gauging station results makes most sense, and that the post-13.8 cal ka contribution from the Susurluk Drainage Basin and Gönen and Biga watersheds was likely in the range of 38–40 km^3 (at a notional porosity of 60% incremented by 10% biogenic carbonate introduced offshore).

6. Discussion

The primary reason for insisting on a deficiency in supply from rivers alone is that there are several other processes which must have contributed fine-grained sediment, even if the river supply was dominant. The deep central basins contain thick post-reconnection deposits (Figs. 7B, 9) with ubiquitous muddy turbidites and homogenites attesting to seismically-induced failures of portions of the very steep slopes which encircle these areas (e.g., Beck et al., 2007, 2015; Eriş et al., 2012; Zitter et al., 2012; Campos, 2014). Any material mobilised in this way from Pleistocene and older deposits would not be accounted for by the calculations targeting

contemporaneous river supply. Roberts (2012) documents common presence and several spikes

in reworked pre-Quaternary dinoflagellate cysts (600–1200 cysts g^{-1}) and pollen in core M02-

89P from the İmralı Basin, confirming some input from older seafloor erosion and failures.

Another potential source for sediment in the deep basins is lateral transfer of detritus

from the lowstand shelves during the 13.8–7.0 cal ka transgression. This process is called

transgressive shoreface erosion, or ravinement (Swift, 1968). Ravinement depends on energetic

waves to carve up to several metres of shoreface deposits away as the shoreline sweeps

landward. Modern storm wave base at ~ −75 to 85 m (Section 2) is compatible with shoreface

erosion, but the burial of pristine sand waves and delta lobes on parts of the southern shelf (Aksu

et al., 1999) suggests minor downcutting during the latest Pleistocene to early Holocene

transgression.

6.1. Possible Bosphorus contributions

A possible sediment source, at least from ~11.1–6.0 cal ka, is outflow from the Black Sea

through the Strait of Bosphorus. The strait carried bedload in a one-way flow until ~10.2 cal ka

(Aksu et al., 2016; Hiscott et al., 2017) and then suspended load until ~6.0 cal ka as a salt wedge

penetrated, at depth, into the strait, presaging the eventual full penetration of a bottom-hugging

saline density current into the Black Sea by ~9.4 cal ka (Flood et al., 2009; Ankindinova et al.,

2019).

The majority of the Unit 1 mud drape is younger than 10.2 cal ka when the Black Sea

outflow started to weaken as indicated by abandonment of the climbing delta at the southern exit

of the Strait of Bosphorus (Fig. 11). It is therefore likely that only small amounts of detritus

introduced by Holocene Bosphorus outflow reached central and western parts of the Marmara

Sea to contribute to the Unit 1 mud drape. Based on the volume of the delta lobe itself, the offshore contribution might have been ~0.35–0.50 km^3 (Section 4.1). Nevertheless, the Mo signature of deep basinal muds in this study (Fig. 12, average 20.8 μg g^{-1}) is consistent with the high Mo abundances in the silts and muds of the Holocene drape overlying the outflow delta itself (average 1243 μg g^{-1} for 10 samples, but as high as 3715 μg g^{-1} in M02-110P samples with an age of ~8.3 cal ka). A dedicated study based on deep basinal cores in the Çınarcık Basin would be required to investigate the extent of this source-to-sink relationship.

6.2. Possible Dardanelles contributions

Baştürk et al. (1986) calculated the suspended-sediment discharge into the Marmara Sea through the Strait of Dardanelles to be ~0.9 Mt yr^{-1}. This was based on discharge of the saline lower layer and the quantity of suspended sediment in water samples, captured on 1.2 μm filter papers during a limited number of casts in March, April and July of 1986. We do not find the sampling frequency nor density to be sufficient for strong conclusions, and are somewhat skeptical because no rivers or large streams enter the strait to provide suspended material to the Aegean inflow. Also, there are fields of bioherms across most of the floor and at the northern end of the strait, in water up to 60 m deep (Aksu et al., 2018), constructed in part by coralline red algae which require diffuse sunlight to survive (i.e., relatively clear water well down into the saline lower layer). Nevertheless, we assess the possible impact of a sediment contribution of ~0.9 Mt yr^{-1} below.

Melting of the Scandinavian ice sheet after the Last Glacial Maximum caused meltwater to spill from the Neoeuxine Lake (today's Black Sea), through the Propontis Lake (today's Marmara Sea) to the Aegean Sea (Grosswald, 1980; Tudryn et al., 2016; Yanchilina et al., 2019).

The overspill occurred ~17.2–15.7 cal ka (Tudryn et al., 2016). Then, the sill at the northern end of the Strait of Dardanelles had an elevation of ~ −76 to −77 m (Aksu et al., 2016) so the level of the Propontis Lake must have been somewhat higher, say ~−75 to −73 m. The Aegean Sea was significantly lower at ~ −120 m (Lambeck et al., 2014). The level of the Propontis Lake then fell to ~ −85 m when outflow from the Neoeuxine Lake ceased and net evaporation ensued (Aksu et al., 1999), resulting in isolation of the lake from the Aegean Sea.

Gökaşan et al. (2010) argued that the subsequent entry of Aegean water into the lowstand Marmara basin (starting 13.8 cal ka according to Aksu et al., 2016) carried large amounts of detritus eroded from the floor of the former Dardanelles valley. Because of the large difference in density between this Aegean inflow and the fresh to brackish waters of the pre-reconnection Propontis Lake, the initial inflow likely descended as an underflow into the westernmost Tekirdağ Basin, confining its sediment contribution to that basin. The spill depth between the Tekirdağ Basin and the Central Basin is at an elevation of −710 m, below which the Tekirdağ Basin has a volume of ~105 km^3. This volume is sufficient to have retained ~75% of the dense, inspilling Aegean seawater needed to equalise the water levels of the Aegean and Marmara seas.

The initial inflow of Aegean surface water into the Marmara basin through the Strait of Dardanelles likely introduced sediment (Gökaşan et al., 2010). However, that inflow would have ceased once the level of the Marmara Sea rose to coincide with the level in the Aegean Sea, because there was no connection to the Neoeuxine Lake at this time. The level of the Propontis Lake immediately before entry of Aegean water was about −85 m, and its level by 13.2 cal ka was no shallower than −60 m (Lambeck et al., 2007). The volume of Aegean Sea water needed to raise the base level from −85 m to −60 m is ~140 km^3, which is one-quarter to one-sixth of today's annual northward bottom flow (Polat and Tuğrul, 1996). At the time of first entry of

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saline water the cross-sectional area of the strait would have been much less than today, so the flow velocity and erosional potential might have been greater. The α1 unconformity marking the base of Unit 1 where the Dardanelles inflow channel enters the Marmara Sea (Fig. 10) might have been created by scour during this event. The lower portion of Unit 1 (i.e., the sediments between the α1 unconformity and the d reflector, Fig. 10) potentially accumulated atop this scoured surface as the initial inflow waned and as water level rose to coincide with the level in the Aegean Sea.

The incoming seawater had to overtop a sill located at the northern end of the strait (Aksu et al., 2016, their fig. 3a). Because the strait deepens by ~20 m on the Aegean side of this point, erosion would have been limited to the area of the sill itself, and the descent from the sill to the lowstand Marmara shoreline, then ~ 30 km away. Given the modest water volume (~140 km^3) and the short distance, the volume of sediment mobilised by the initial entry was likely considerably smaller than suggested by Gökaşan et al. (2010). If the suspended sediment concentration had been 1200 mg L^{-1} (0.0012 Gt km^{-3}), which is high for rivers in flood (e.g., Blanchard et al., 2011; Ellison et al., 2014) and double the typical wet-season concentration for mountainous tributaries of the Amazon River (Gibbs, 1967), then the 140 km^3 of Aegean inflow would have transported 0.17 Gt of detritus into the Tekirdağ Basin, equivalent at 60% porosity and 10% biogenic carbonate to only 0.18 km^3 of porous sediment.

After the equalisation of water levels, saline Aegean water probably continued to flow, albeit more slowly, into the Marmara basin (a) because of its density contrast with the pre-existing brackish water mass and (b) to replenish evaporative losses. This slow inflow expelled and replaced ~2740 km^3 of brackish water remaining from the lacustrine phase; at 10% of today's rate of inflow (so ~70 km^3 yr^{-1}) the replacement might have taken ~40 years to complete.

The Kocasu, Gönen and Biga watersheds only supply fresh water at ~5.2 km^3 yr^{-1}, so would not have significantly delayed the salination.

From ~13.2–11.1 cal ka, the Marmara Sea remained an embayment of the Mediterranean Sea, with full marine salinity (Table 6). The *d* reflector in the mud-prone levées at the eastern exit of the Strait of Dardanelles (Fig. 10) possibly records reduced sedimentation during stagnation of exchange at this time (i.e., little inflow or outflow). Subsequently, the rising Neoeuxine Lake began to export brackish water down the Bosphorus valley (Aksu et al., 2016), across the Marmara Sea and through the Dardanelles, but with no return flow because saline water had not yet gained access to the Neoeuxine Lake. The southward flow through both straits was one-way only, so opposite to the situation from ~13.2–13.2 cal ka (Table 6). The unidirectional outflow might have controlled deposition of the uppermost portion of Unit 1 above the *d* reflector (Fig. 10). This situation continued until ~9.5 cal ka when the first saline water began to enter the Neoeuxine Lake/Black Sea (Major et al., 2006; Yanchilina et al., 2017; Ankindinova et al., 2019). Sustained two-way flow along the Strait of Bosphorus began later at ~8.0 cal ka (Hiscott et al., 2007b) and it is only then that the northward flow of saline water along the floor of the Strait of Dardanelles might have become competent to transport ~0.9 Mt yr^{-1} of suspended sediment into the Marmara basin (Baştürk et al., 1986) and to cause the erosion seen in the modern walls of the Dardanelles channel (i.e., the α0 unconformity, Fig. 10). Integrated over 8,000 years, this flux would have been able to bring ~7.2 Gt of suspended sediment into the western Marmara Sea. As explained above, we believe this estimate is too large given other constraints, so 5 Gt might be a more realistic maximum, which is equivalent to ~5.2 km^3 of porous marine mud. A Dardanelles input of this amount, and only since ~8.0 cal ka, is ~12% of the 43–47 km^3 of siliciclastic detritus in Unit 1 (Table 2).

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7. Conclusions

The uppermost Quaternary, post-transgressive mud drape in the Marmara Sea provides a rare opportunity to quantitatively compare the supply of detritus from specific rivers and streams to the volume of material dispersed to offshore shelves and basins. More commonly, fine material of this calibre is advected far offshore and alongshore so that it cannot be confidently tracked backward to its source. In this study, sediment supply from rivers and streams along the southern margin of the Marmara Sea is estimated in two ways: (1) using hydrographic measurements of water and suspended-sediment discharge over several decades in these watersheds, and (2) using a multivariate model developed for a large number of rivers worldwide (BQART model), taking into consideration the area, relief and geology of the watersheds, climatic regime, and average water discharge. The predicted sediment supply since reconnection of the Marmara basin to the global ocean at ~13.8 cal ka approaches what is needed to account for all the coeval muddy Unit 1 across the modern marine shelves and basins, surpassing ~85–90% of the requirement. Although other sediment sources must have been active, this study strongly suggests their limited contribution. These include: (a) products of shoreface erosion during the –85 m to 0 m transgression; (b) episodic transport of detritus through the Strait of Dardanelles (13.8–13.2 cal ka; 8.0–0 cal ka); (c) early Holocene supply of detritus through the Strait of Bosphorus, and (d) failure of older slope material adding to deep basin successions.

The application of a cap of 40% on the amount of river-borne detritus available to marine areas fits nicely with reasonable estimates of the volume of latest Quaternary floodplain and lacustrine deposits in the lower part of the Susurluk Drainage Basin, around the Manyas and Uluabat Lakes. If the sequestering of 60% of river load in floodplains and delta lobes had not

been specified in calculations, the post-13.8 cal ka supply would have significantly exceeded the volume of contemporary marine mud.

For the first time, the detrital input through the Strait of Dardanelles has been critically evaluated based on the complex history of flow in that waterway (both changing magnitude and direction). Unconformity-bounded sedimentary wedges beneath the banks of the Dardanelles channel as it enters the Marmara Sea are tentatively assigned to five distinct phases of flow: (1) 13.8–13.2 cal ka erosion producing the late Pleistocene α1 unconformity; (2) accumulation of lenticular levée wedges under waning inflow from the Aegean Sea as the water level in the Marmara Sea approached the level of the global ocean; (3) development of an hiatus (~13.2–11.1 cal ka) and minor unconformity (*d* reflector) when the Marmara Sea was a marine embayment of the Aegean Sea with limited water exchange in either direction; (4) accumulation of the uppermost levée wedges under westward-directed outflow toward the Aegean Sea, 11.1–8.0 cal ka, as the Neoeuxine Lake spilled outward without significant return flow; (5) full establishment of two-way flow in both straits so that the bottom water in the Strait of Dardanelles now moves persistently northeastward, widening the saline inflow channel (Fig. 10) and locally truncating older strata along its banks.

The primary objective of explaining the origin of the thick mostly Holocene mud drape in the tectonically active Marmara Sea has been addressed successfully, with relatively steep rivers and streams of the southern margin contributing the majority of the detritus. Further coring and dating will be necessary to test hypotheses advanced here on the origin of wedges of sediment beneath the banks of the submarine channel at the northeastern exit of the Strait of Dardanelles.

Acknowledgments

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Figure Captions

Figure 1. (A) Location of the study area in western Turkey. (B) Map of the Marmara Sea as a

gateway linking the Black Sea to the Aegean Sea via the Straits of Bosphorus and

Dardanelles. Coastline and rivers are from NOAA National Geophysical Data Center,

extracted from http://www.ngdc.noaa.gov/mgg/shorelines/ shorelines.html. Topography

and bathymetry compiled using GeoMapApp (Ryan et al., 2009), bathymetry of the

Marmara Sea compiled from Rangin et al. (2001). White dashed line (long dashes)

represents the catchment area of water courses draining into the Marmara Sea, delineated

using the digital topography in Global Mapper. White lines with shorter dashes delimit the

portion of this catchment called the Susurluk Drainage Basin (Kazancı et al., 2019) which

includes tributaries of the Kocasu River. Note that the Kocasu River and the Gönen, Biga

and Kemer streams enter Marmara Sea from the south, and several much smaller streams

from the north, including A = Ana, Kı = Kınıklı, K = Kula, Ku = Kurbağalıdere streams.

Tekirdağ, Central, Kumburgaz, Çınarcık and İmralı are basins and Manyas and Uluabat are

Lakes discussed in text. 50 m and 100 m isobaths are from IOC (1981). Red box is

illustrated in Figure 4.

Figure 2. (A) Location of EIE stations (white filled circles with red numbers). Also shown are

the drainage areas of the Kocasu River and the Gönen and Biga streams (white dashed

lines). Emet, Kocaçay, Nilüfer, Orhaneli and Simav are branches of the Kocasu River. The

inset (B) shows an enlargement of the Karacabey gorge and sample locations in that area

(red filled circles). The gorge connects the low-lying alluvial plain (including the Manyas

and Uluabat Lakes) behind the shore parallel rugged hills, including Karadağ Mountain,

and the present-day coastal . Credits for the base maps are given in the Figure 1 caption.

Figure 3. Geological maps of the Marmara region, simplified and redrawn from the 1:500,000 İzmir and İstanbul Map Sheets of Konak (2002) and Türkecan and Yurtsever (2002), respectively. Because of the complicated nature of the map, the pertinent geology is presented in four panels: (A) ophiolitic mélange and plutonic rocks; (B) metamorphic rocks, (C) volcanic rocks and (D) sedimentary rocks. b = Biga Stream, g = Gönen Stream, k = Kocasu River, c = coastal erosion along the southeastern Gallipoli Peninsula and northern shoreline of the Marmara Sea. Credits for the base map are given in the Figure 1 caption.

Figure 4. Map of the study area showing the locations of the seismic reflection profiles (white lines) and piston and gravity cores used in this study. Red, aquamarine, white and yellow filled circles are gravity, piston, Calipso and Livingstone cores, respectively. Credits for the base map are given in the Figure 1 caption.

Figure 5. Graphs showing the pre-dam relationships between water discharges and suspended sediment loads in the Emet and Kocaçay branches (stations 302 and 314) of the Kocasu River and the Gönen stream (station 210). Data are compiled from EIE (1982, 1987, 1993, 2000) sources. The 2nd order quadratic equations are used for the calculation of post-13.8 cal ka sediment budgets.

Figure 6. High-resolution Huntec seismic reflection profile showing the internal architecture of the sedimentary successions across the southwestern Marmara Sea and the correlation of seismostratigraphic Unit 1 with core M02-106P. Vertical scale is calculated using 1500 m s^{-1} acoustic velocity. Red numbers = navigation fix positions. Location is shown in the inset and Figure 4.

Figure 7. (A) Depositional domains for the purpose of comparison between the volumes of Unit
1 sediments (latest Quaternary; post-13.8 cal ka) across the Marmara Sea and the sediment
contributions of the Kocasu River and the Gönen and Biga streams. **(B)** Thickness
distribution of Unit 1 (in metres, calculated using 1500 m s^{-1} acoustic velocity in seismic
reflection profiles) across the Marmara Sea. Thicknesses of Holocene sediments in the
Manyas and Uluabat Lakes are from Kazancı et al (2004). Also shown are the Kocasu
River and the Gönen, Biga, Ana (A), Kınıklı (Kı), Kula (K) and Kurbağalıdere (Ku)
streams. Note the thick accumulation of Unit 1 in the southwestern Marmara Sea at the
entrance to the Strait of Dardanelles, off the mouth of the Kocasu River, as well as in deep
water basins. Credits for the base map are given in the Figure 1 caption.

Figure 8. Lithostratigraphic logs of cores used in this study, showing radiocarbon dates (in cal yr
BP; see Supplementary Material 2 for raw ^{14}C ages). Gravity cores at the sites of M02-
103P, M02-104P and M02-106P demonstrate core-top losses of 74 cm, 100 cm and 36 cm
in those piston cores, respectively. The volcanic ash layer in cores M02-89P, M02-90P,
M02-102P and M14-16P is the regionally widespread Y2 ash dated at 21,945 cal yr BP.
Credits for the location map are given in the Figure 1 caption. On the map, a trailing P is
omitted from every piston core name to reduce clutter.

Figure 9. Summary core logs for Calypso cores raised from deep basins of the Marmara Sea,
extracted from Beck et al. (2007, 2015), Çağatay et al. (2009), Vidal et al. (2010), Eriş et
al. (2011, 2012), Campos (2014) and Aloisi et al. (2015). Unit 1 is the marine section,
shown either as a green fill or grey (sapropel M1). Credits for the location map are given in
the Figure 1 caption.

Figure 10. High-resolution Huntec seismic reflection showing the internal architecture of the sedimentary successions across the Dardanelles channel at the northeastern entrance to the Strait of Dardanelles. The *d* reflector (*d* for Dardanelles) is interpreted as an unconformity, explained in the text. Vertical scale in metres is calculated using 1500 m s^{-1} acoustic velocity; VE (vertical exaggeration) ~23×. Red numbers at base = navigation fix positions. Location is shown in the inset where red- and aqua-filled circles are gravity and piston cores, respectively.

Figure 11. Interpreted Huntec DTS profile across the 11.1–10.2 cal ka Holocene outflow delta at the southern exit of the Strait of Bosphorus, modified from Aksu et al. (2016). Subtle up-and-down wobbles along the seabed result from towfish flight variations created by strong currents in the area. Ages of reflectors are: β1, 10.2 cal ka; β2, 11.8 cal ka; β3, 12.0 cal ka; β4, 16.2 cal ka. Everything above β3 belongs to Unit 1 of this study.

Figure 12. Log-log plot of average elemental abundances in the modern versus abundances in shelf cores M02-103P and M02-100P (blue dots) and deep basinal cores (red dots). Blue dots are alway on top, so in places partially obscure other data points. The order of elements, left to right, is the same as the ranking in Table 5. Only the anomalies are labelled, but the identity of other pairs can be determined using the tabulated values. Blue and red dots for each element are aligned vertically, and their degree of separation is a qualitative indication of the difference in shelf and deep basin means.

Table Captions

Table 1. Characteristics of Unit 1 (uppermost Pleistocene to Holocene mud drape), from

literature sources.

Table 2. Volumes of porous, water-saturated Unit 1 mud in various parts of the Marmara Sea.

Thicknesses were determined from seismic profiles assuming an acoustic velocity of

1500 m s^{-1}. Values for domain 5 consider the lakes only and not adjacent portions of the

inland depression.

Table 3. Contributions of river and stream detritus to the Marmara basin as environmental

conditions changed since 13.8 cal ka. Column 4 gives annual suspended loads from

gauging station daily records and equations in Figure 5. The values in column 5 are the

result of reducing daily water discharges to 67% of modern measurements to

acknowledge a drier climate during the Younger Dryas. Columns 6–9 take account of

constraints noted in Section 4.2.1 (e.g., 20% increment at certain times for bedload).

Column 10 sums columns 6–9. Column 11 converts the values in column 10 to an

equivalent volume of sediment with 60% porosity, a grain density of 2.65 t m^{-3}, and a

10% increment for the biogenic carbonate added to river input in the marine setting. To

convert, divide by 2.65, then by 1.0–0.60 = 0.40 and finally multiply by 1.1.

Table 4. Input values (columns 3–7) for BQART determinations (columns 8–11) of river

suspended-load contributions. Column 4 presents averages over 30–60 years (gauging

stations), reduced in Younger Dryas to account for less rainfall. Column 6 is from Google

Earth™ (Ruler>Path>Profile). Column 7 is from Valsecchi et al. (2012) with an

interpolated value of temperature for 11.7–7.0 cal ka. Columns 8–10 are calculated using

equation 7a of Syvitski and Milliman (2007). The values in column 11 are based on totals

for each branch incremented by 20% for bedload (only before 4.1 cal ka for Emet and

Kocaçay branches due to Manyas and Uluabat trapping afterward), then reduced to 40%

of that amount (following the Kukal, 1971, suggestion that 60% remains on the

floodplain and in the delta during highstands), then inflated to an assumed porosity of

60% with an additional 10% increment for biogenic carbonate. These are estimates of the

volumes available to Unit 1 since 13.8 cal ka. The TOTAL includes ±7% expected bias

(Syvitski and Milliman, 2007).

Table 5. Mean abundances of elements from LA-ICPMS measurements, ranked in order of

decreasing abundance in the Kocasu Delta sample set. Column 1 averages samples M17-

1 to M17-7; column 3 averages samples from shelf cores M02-103P and M02-106P;

column 4 averages samples from basinal cores M02-88P, M02-89P, M02-90P, M02-

102P. Column 5 indicates if 95% confidence intervals (C.I.) overlap for Kocasu and all

marine data (shelves and basins); column 6 indicates if 95% confidence intervals overlap

for delta and shelf samples only; column 7 indicates if delta and shelf samples have

indistinguishable means based on a two-tailed student t-test.

Table 6. Inferred evolution of Dardanelles valley and channel after the Last Glacial Maximum.

SW flow is toward the Aegean Sea; NE flow is toward the Marmara Sea.

Table 1.

Feature or parameter	Value or description	Sources
Texture	Mixtures of generally <10–15% sand, clay >30% (rarely less near deltas and shorelines), silt >10% and <30% in deep basins. Silt–fine sandy laminae intercalated in deep basins.	Ergin and Bodur (1999), Beck et al. (2007), Hiscott et al. (2017)
Facies	On shelves, either thoroughly bioturbated mud with subtle colour mottling, or pinstripe laminated (at millimetre scale) sapropel with no burrows or rare *Chondrites* traces. Bioturbated facies are dominant. In deep basins, burrowing is less because of poor water-column oxygenation, finely laminated sapropel alternates with faintly banded mud punctuated by silt–sand laminae.	Çağatay et al. (2000), Aksu et al. (2002, 2016), Beck et al. (2007), Eriş et al. (2012)
Sand to coarse silt fraction	Predominantly biogenic carbonate grains (foraminifera, juvenile or fragmented molluscs) with increased siliciclastic detritus in very fine sand fraction, including many mafic minerals and metamorphic	Aksu et al. (2002), Beck et al. (2007), Hiscott et al. (2017)

	grains. In deep basins, silt is rich in fresh plagioclase, amphibole and pyroxene, brown mica, opaques minerals and detrital (rather than biogenic carbonate).	
Clay minerals	~50% smectite, ~25–30% illite, ~10–20% kaolinite, <10% chlorite	Bayhan et al. (2001), Kazancı et al. (2004), Ehrmann et al. (2007)

Table 1 continued.

Feature or parameter	Value or description	Sources
Organic carbon	TOC mostly 0.5–3.0%, highest in deep basins and laminated sapropels.	Çağatay et al. (2000), Abrajano et al. (2002), Ergin et al. (2007)
Carbonate content	7–20% but typically ~10–15% as $CaCO_3$	Çağatay et al. (2000), Ergin et al. (2007), Beck et al. (2007)
Macrofossils	Molluscs: *Corbula gibba*, *Mytilus galloprovincialis*, *Timoclia ovata*, *Paphia discrepans*, *Donax trunculus*, *Cardium (or Cerastoderma) edule*, *Parvicardium exiguum*, *Bittium reticureticulatum*, *Chlamys varia*, *Spisula subtruncata*, *Abra alba*, *Turritella communis*, *Myrtea spinifera*; species of *Vermetus*, *Patell*a, *Cyrcomphalus*, *Hydrobia*, *Dentalium*, *Nucula*.	Çağatay et al. (2000), Aksu et al. (2016), Büyükmeriç (2016)

	Rare serpulid tubes, echinoderm plates, crustacean claws.	
Microfossil groups	Benthic and planktonic foraminifera, nannofossils, dinoflagellate cysts, pollen and spores, diatoms	Çağatay et al. (2000), Aksu et al. (2002), Kaminski et al. (2002), Mudie et al. (2002a, 2002b), McHugh et al. (2008), Londeix et al. (2009), Valsecchi et al. (2012)
Colour	Dark yellowish brown (10YR4/2), olive gray (5Y4/1), brownish gray (5YR4/1), greenish gray (5GY4/1), to moderate olive gray (5Y4/2). Sapropel is colour banded 5Y4/1 and 5GY5/2.	Aksu et al. (2002)

Table 2.

Do m ai n	T h i c k n e s s r a n	V o l u m e w i t h p

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	ge (m)	ore water (km³)
(1) Southern shelf	0 – 25	9.09
(2) Southwestern shelf	0 – 25	4.55
(3) Northern	0 – 30	1.48

shelf and Bosphorus exit		
(4) Deep central basins	0 – 30	31.37
(5) Manyas and Uluabat lakes	0 – 15	1.10

onl y		
To tal (d om ain s 1, 2, 4)		4 5 . 0 1

Table 3.

1	2	3	4	5	6	7	8	9	10	11
System	Branch	yrs of daily data	t yr^{-1} Holocene	t yr^{-1} YD	13.8–11.7 Gt	11.7–7.0 Gt	7.0–4.1 Gt	4.1–0 Gt	Total Gt	Total at 60% porosity (km^3)
Kocasu	Kocaçay	55	487456	227314	0.573	2.749	0.6?9	0.799	4.800	4.98
	Emet + Orhaneli	61	1924683	886818	2.235	10.8??	?.679	3.156	18.926	19.64
	Simav	59	1597574	734200	1.?50	9.010	2.224	3.144	16.228	16.84
	Nilüfer	49	123311	6?5?8	0.158	0.695	0.172	0.243	1.267	1.32
Gönen	Gönen	37	73727	36215	0.091	0.416	0.103	0.145	0.755	0.78
	Keçi	29	2?46	1563	0.004	0.014	0.003	0.005	0.026	0.03
Biga	Biga	46	62426	30862	0.078	0.352	0.087	0.123	<u>0.640</u>	<u>0.66</u>
Grand total								42.642	44.25	

Table 4.

1	2	3	4	5	6	7	8	9	10	11
Branch	Range	B	Q	A	R	T	Qs	Duration	Total	Total at 60%
	cal ka		km^3 yr^{-1}	km^2	km	°C	Mt yr^{-1}	yr	Gt	porosity (km^3)
Emet & Orhaneli	13.8 – 11.7	¹.00	1.159	9629	1.125	10.0	0.693	2100	1.456	

(Stn 302)	11.7–7.0	1.00	1.755	9629	1.125	12.0	0.946	4700	4.448	
	7.0–4.1	1.00	1.755	9629	1.125	13.7	1.08	2900	3.133	
	4.1–0	1.00	1.755	9629	1.125	13.7	1.080	4100	4.430	10.75
Simav	13.8–11.7	1.25	0.893	6454	1.350	10.0	0.785	2100	1.649	
(Stn 316)	11.7–7.0	1.25	1.352	6454	1.350	12.0	1.072	4700	5.038	
	7.0–4.1	1.25	1.352	6454	1.350	13.7	1.224	2900	3.549	

	4.1−0	1.25	1.3552	6454	1.350	13.7	1.224	4100	5.017	12.59
Kocaçay	13.8−11.7	1.25	0.3777	2309	0.650	10.0	0.173	2100	0.3644	
(Stn 314)	11.7−7.0	1.25	0.5711	2309	0.650	12.0	0.2236	4700	1.1111	
	7.0−4.1	1.25	0.5711	2309	0.650	13.7	0.2270	2900	0.782	
	4.1−0	1.25	0.5711	2309	0.650	13.7	0.2270	4100	1.106	2.68
Nilüfer	13.8−11.7	1.50	0.298	1291	2.0000	10.0	0.445	2100	0.934	
(S	11	1.	0.	12	2.	12	0.	47	2.	

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tn 321)	.7 –7 .0	50	452	91	000	.0	607	00	852	
	7.0 –4.1	1.50	0.452	1291	2.0000	13.7	0.693	2900	2.009	
	4.1 –0	1.50	0.452	1291	2.0000	13.7	0.693	4100	2.840	7.13
Gönen	13.8 –11 1.7	1.25	0.416	167?	0.?00	10.0	0.187	2100	0.392	
(Stn 210 + 315)	11 1.7 –7.0	1.25	0.631	1670	0.800	12.0	0.255	4700	1.199	
	7.0	1.2	0.6	167	0.8	13.	0.2	290	0.8	

65

	−4.1	5	31	0	00	7	91	0	45	
	4.1−0	1.25	0.6311	1670	0.800	13.7	0.291	4100	1.194	3.00
Biga	13.8−11.7	1.25	0.3333	2096	0.550	10.0	0.13⊦	2100	0.282	
(Stn 212)	11.7−7.0	1.25	0.5055	2096	0.550	12.0	0.183	4700	0.862	
	7.0−4.1	1.25	0.505	2096	0.550	13.7	0.209	2900	0.607	
	4.1−0	1.25	0.505	2096	0.550	13.7	0.209	4100	0.8588	2.15
									TO0054A	38.31 ±

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									L	2 . 6 8

Table 5.

1	2	3	4	5	6	7
	Kocasu	103/106	basins	overlap	overlap	t-test
	$\mu g\ g^{-1}$	$\mu g\ g^{-1}$	$\mu g\ g^{-1}$	95%	95%	$\alpha = 0.05$
Elements	n = 7	n = 15	n = 23	C.I.	C.I.	
Re	0.055	0.075	0.078		X	
In	0.12	0.20	0.16			

	1	1	2			
A g	0 . 1 5 7	0 . 2 5 6	0 . 1 8 0			
T m	0 . 2 4 1	0 . 2 4 7	0 . 2 2 1	X	X	X
L u	0 . 2 6 1	0 . 3 3 8	0 . 2 3 3	X		
T b	0 . 4 6 5	0 . 4 4 4	0 . 3 3 9		X	X
H o	0 . 6 0 6	0 . 5 6 4	0 . 4 5 8		X	
M o	0 . 6 8 5	0 . 7 4 3	2 0 . 7 4 9		X	X
E u	0 . 7	0 . 8	0 . 5	X	X	X

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	84	38	63		
Cd	0.9322	1.5881	1.2443		
Sb	1.8566	0.4855	0.6888		
Yb	1.9100	2.0255	1.626	X	X	X
Er	1.9555	1.7211	1.596		
Ta	2.227	1.446	1.444		
U	2.442	2.998	2.558		
Ge	2.544	3.400	3.002		
G	3	4	1	X		

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d	.03	.04	.91			
Dy	3.18	2.82	2.48			
Sm	4.05	3.53	3.03			
Sn	4.65	5.40	2.60	X	X	
Hf	4.87	4.83	3.c8	X	X	X
Pr	5.07	5.61	4.63	X	X	
Th	12.4	1'.8	10.3		X	X
Pb	13.4	15.0	7.9		X	X
Ga	17.3	18.6	19.0		X	

1	2	3	4	5	6	7
Cs	17.5	12.0	12.8			
Y	18.2	16.9	14.6		X	
Nd	19.7	20.7	17.2	X	X	X
Nb	20.6	16.0	15.1			

Table 5. continued.

1	2	3	4	5	6	7
	Kocasu	103/106	basins	overlap	overlap	t-test
	$\mu g\ g^{-1}$	$\mu g\ g^{-1}$	$\mu g\ g^{-1}$	95%	95%	$\alpha = 0.05$

Elements	n = 7	n = 15	n = 23	C.I.	C.I.	
Sc	21.0	20.1	21.9	X	X	X
La	24.7	27.4	23.1	X	X	
Cu	28.1	38.1	45.9			
As	32.4	2.4	4.9			
Ce	45.9	49.0	41.5	X	X	
Co	50.	35.	42.	X	X	X

	9	9	8			
Zn	93.4	75.3	81.2			
Rb	121	135	143			
V	125	134	150	X	X	X
Sr	131	126	86		X	X
Ni	165	161	142	X	X	X
W	186	76	95	X		
Zr	189	190	157	X	X	X
Cr	285	171	186			
Ba	487	395	451			
Mn	49	29	32			

	3	4	7			
P	514	1776	784	X		
Ti	6345	5598	5225			
Na	11608	9671	6978			
Ca	20225	12188	14194			
K	21815	21310	22431	X	X	X
Mg	22136	14948	17280			
Fe	46	40	43	X	X	

	750	888	333			
Al	98762	111652	114967			
Si	293192	297301	292138	X	X	X

Table 6.

Age span (cal ka)	Flow direction	Underlying cause
17.2–15.7	100% SW	Melting of Scandinavian ice sheet swells Neoeuxine Lake which spills southward into the Marmara Sea and onward to the Aegean Sea. Levels were, respectively, ~−35m, −73 m and −120 m.
15.7–13.8	Subaerial	Evaporative drawdown of Marmara Sea to ~−84 m. Propontis valley Lake phase begins.
13.8–~13.2	100% NE	Rising global ocean overtops the Dardanelles sill, quickly raising the level of the Marmara Sea from ~−84 m to ~−60 m. Only ~140 km^3 would have been required, ~75% of this pooling in the

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		Tekirdağ Basin. The Marmara Sea became an embayment of the Aegean Sea.
~13.2–11.1	50:50	Slow and weak density-driven exchange of saline Aegean water but a trickle for brackish residual Propontis Lake water, in equal amounts, as base level in the Marmara Sea rises slowly from ~–60 m to ~–43 m.
11.1–10.2	100% SW	Vigorous outflow from the Neoeuxine Lake (at ~–35 m elevation) into the Marmara Sea (at –43 m rising to ~–35 m) caused by increased central Asian and European river discharges and possible glacial-lake outburst floods. Entry of Aegean water was blocked by this outflow as it continued southwestward through the Strait of Dardanelles.
10.2–9.4	100% SW	Weakening Neoeuxine Lake outflow is lifted off the floor of the Strait of Bosphorus as a 'salt wedge' penetrates up that strait against the surface brackish outflow. With no entry of saline water into the Neoeuxine Lake, flow through the Strait of Dardanelles continues to be entirely toward the Aegean Sea.
9.4–8.0	>80% SW	Significant but episodic intrusion of saline water into the Black Sea
	<20% NE	with a large initial pulse sufficient to sharply raise the $^{87}Sr/^{86}Sr$ ratio but not to elevate salinity enough to accommodate euryhaline

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		faunas. Aegean water started to move northeastward along the floor of the Dardanelles.
8.0–7.5	<80% SW	Establishment of two-way flow in both straits and eventual
	>20% NE	colonizing of the Black Sea by euryhaline molluscs and ostracods.
7.5–today	~67% SW	Modern two-way flow, so in the Strait of Dardanelles Aegean
	~33% NE	S=39 water runs northeastward along the bottom (5–25 cm s^{-1}) and predominantly Black Sea S=20 water exits to the Aegean Sea as an ~30 m-thick surface layer (10–30 cm s^{-1}).

Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

Highlights

- Latest Pleistocene–Holocene Unit 1 forms a blanket of 43–47 km^3 in Marmara Sea

- Kocasu River and its tributaries can account for 85–90% of the volume of Unit 1

- Sediment budget calculations rely on data from gauging stations and BQART model

- Isolated Marmara basins allow reliable quantification of the sedimentary budget
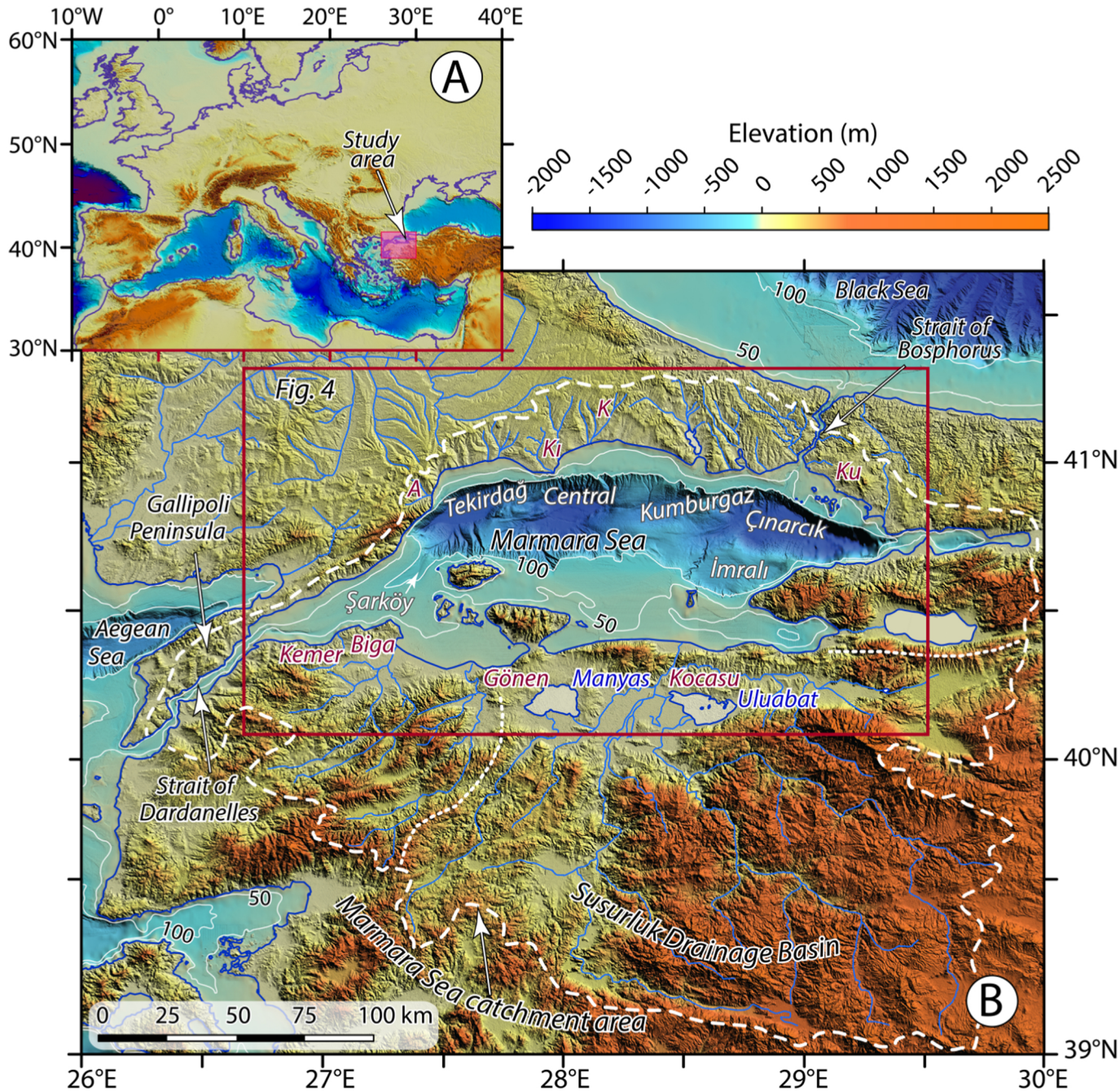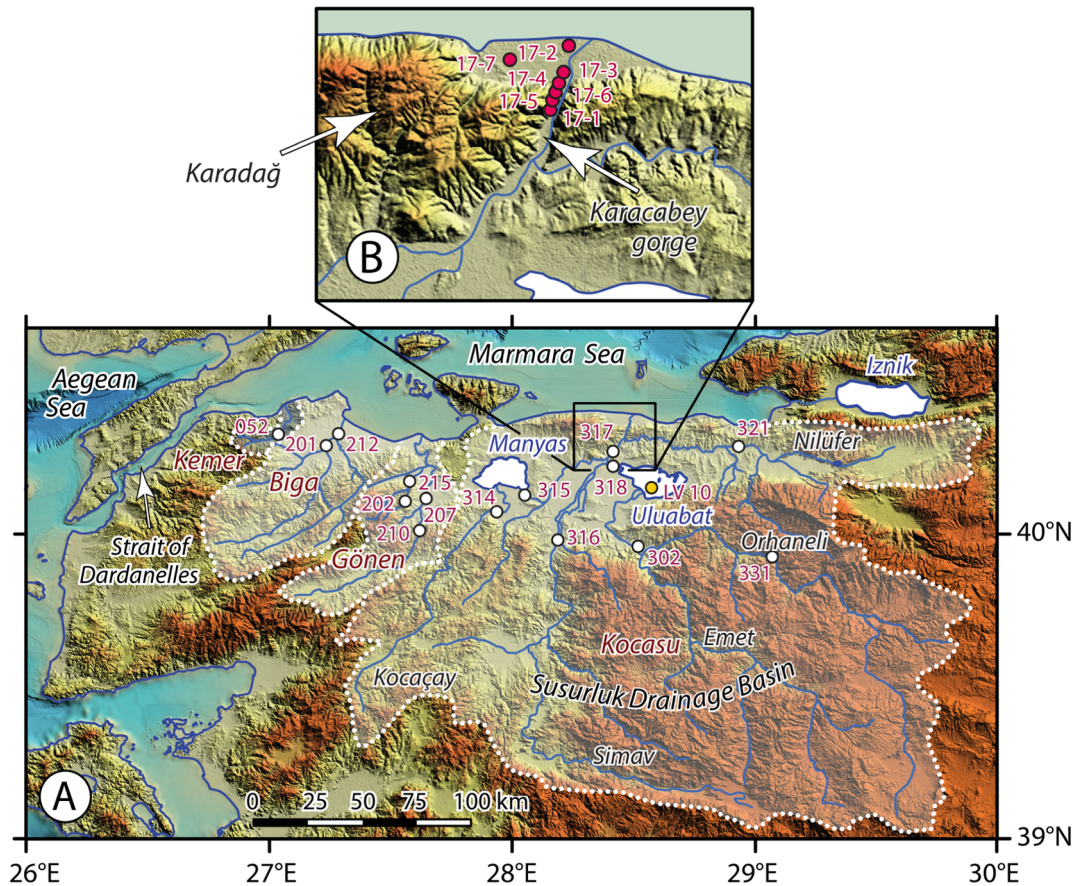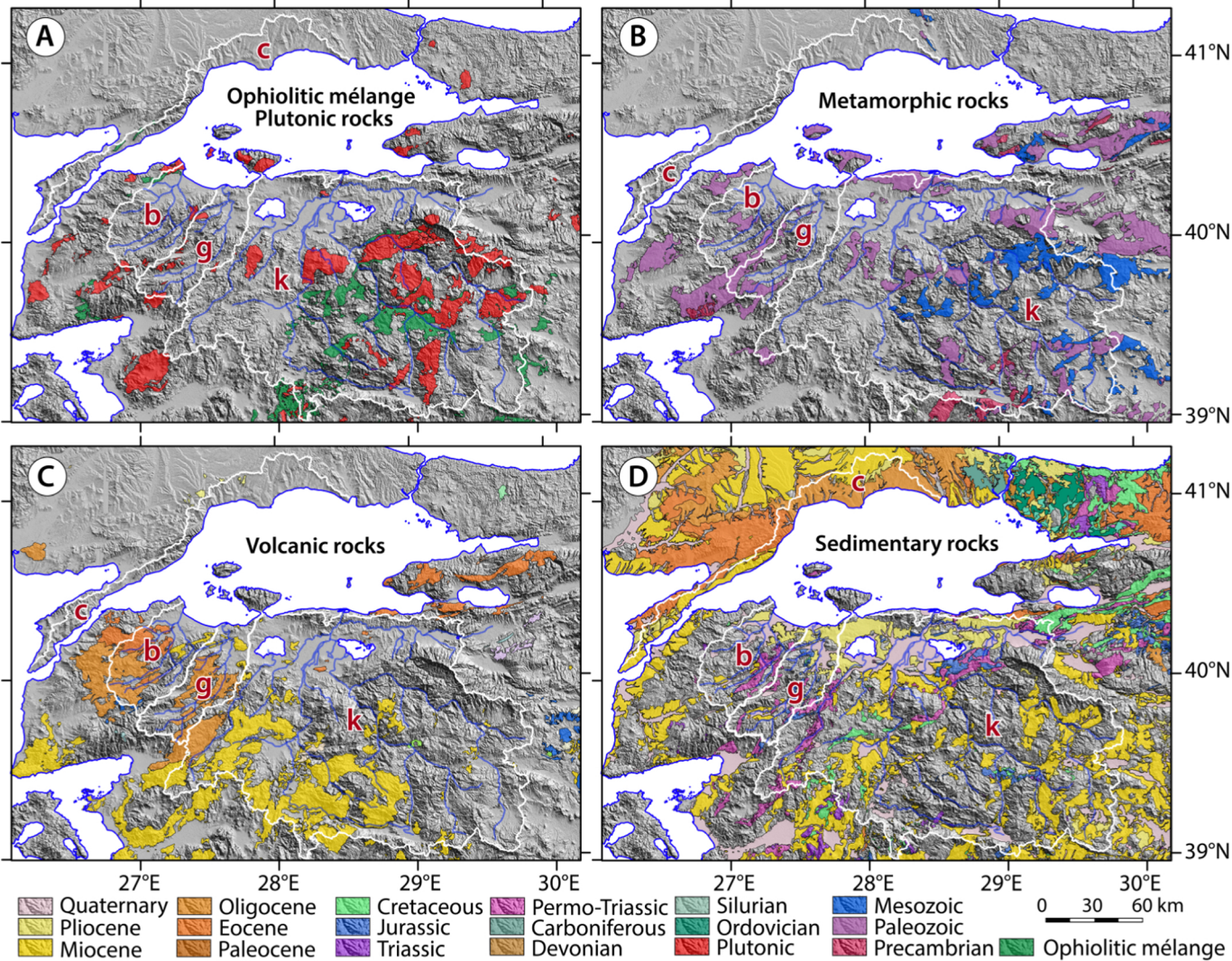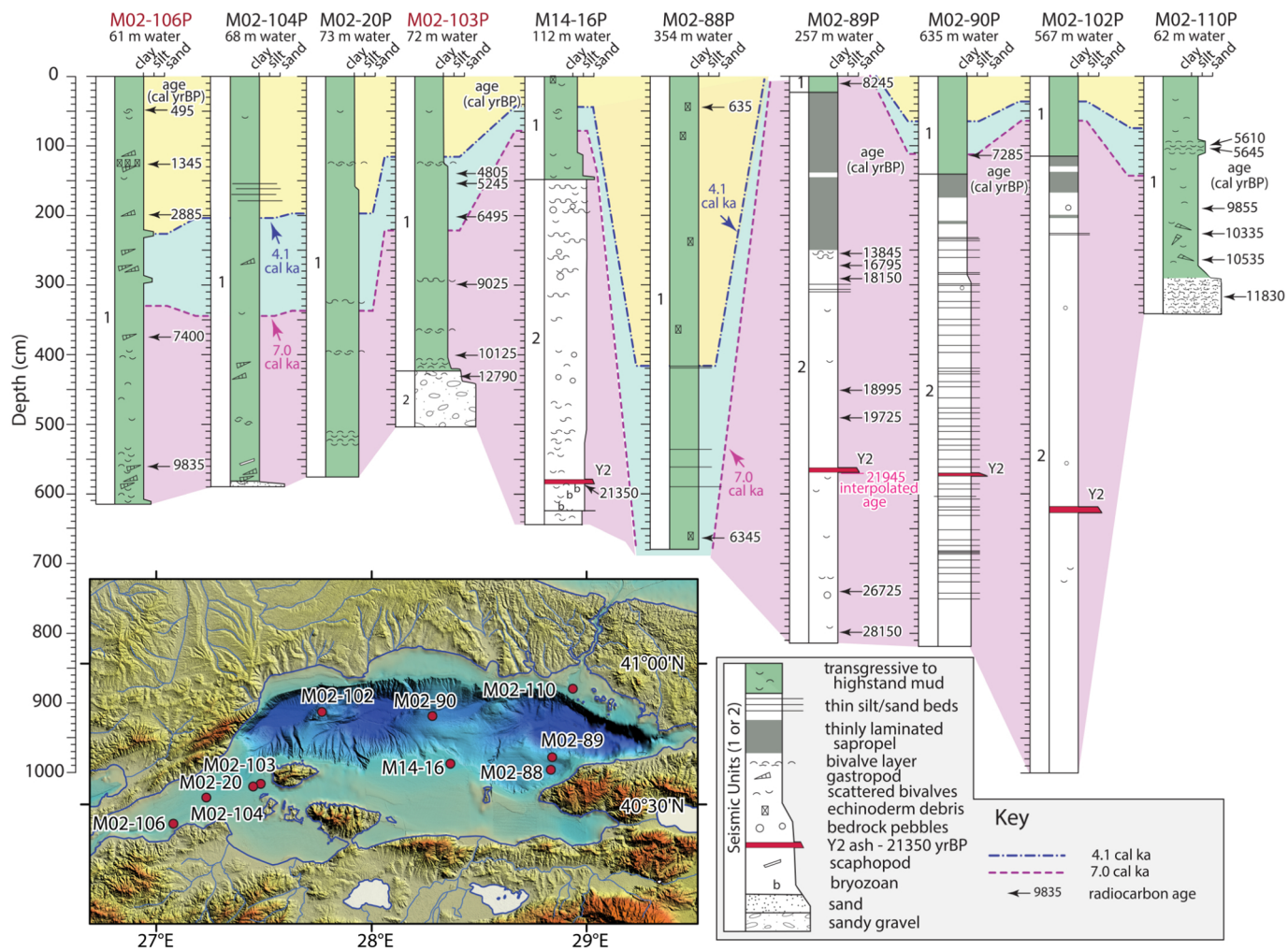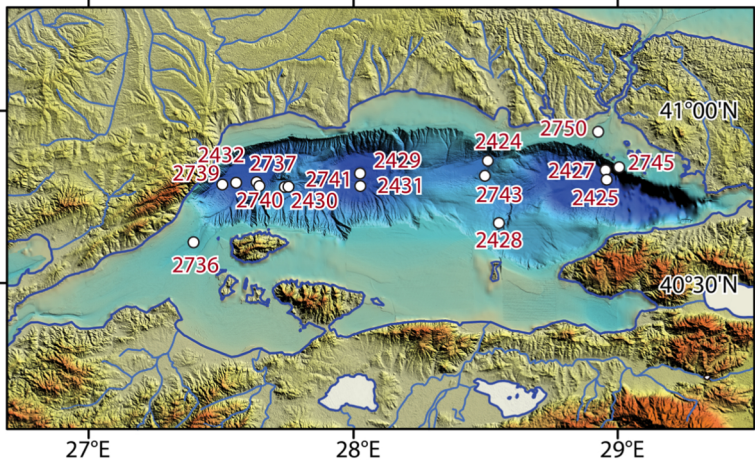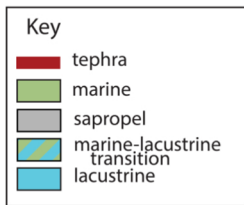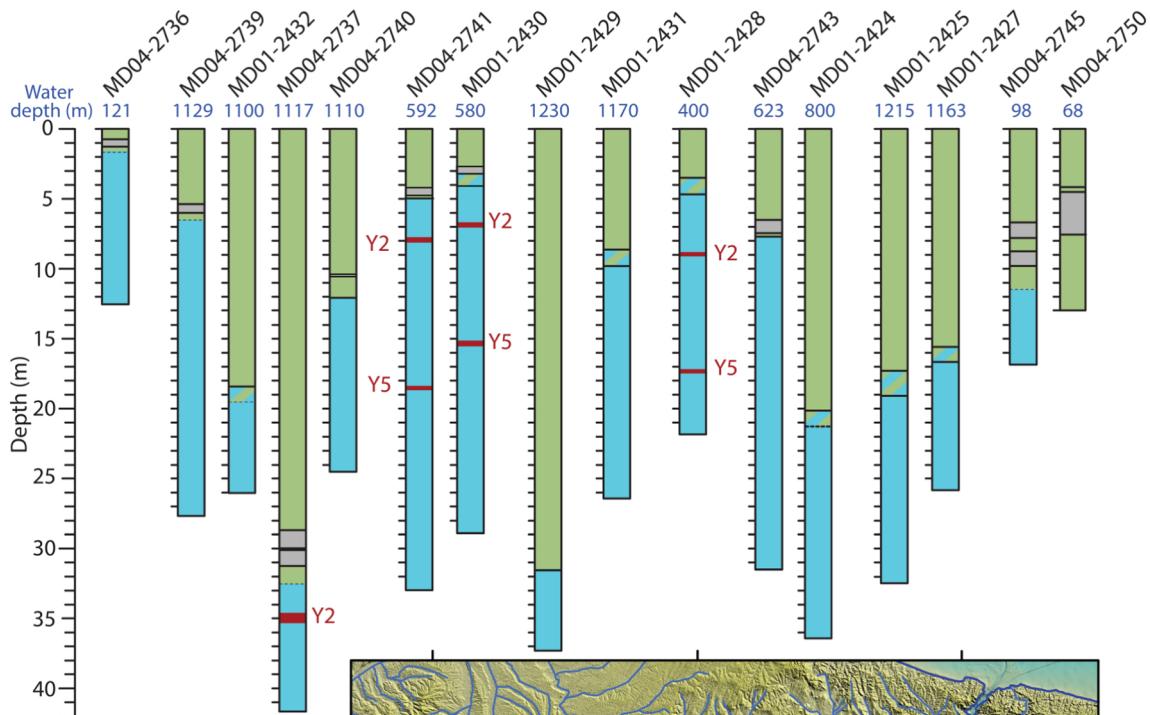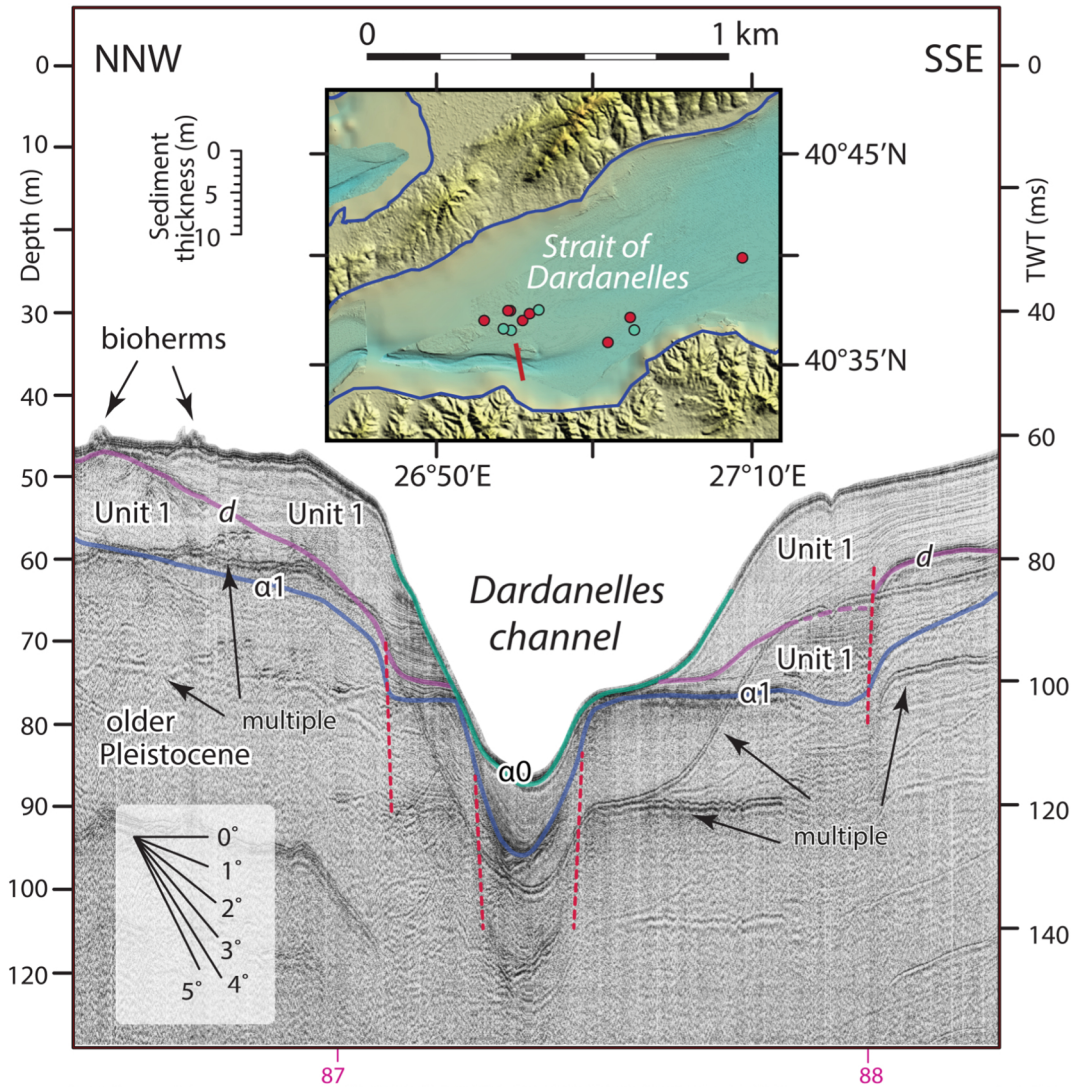
Figure 1

Figure 2

Figure 3
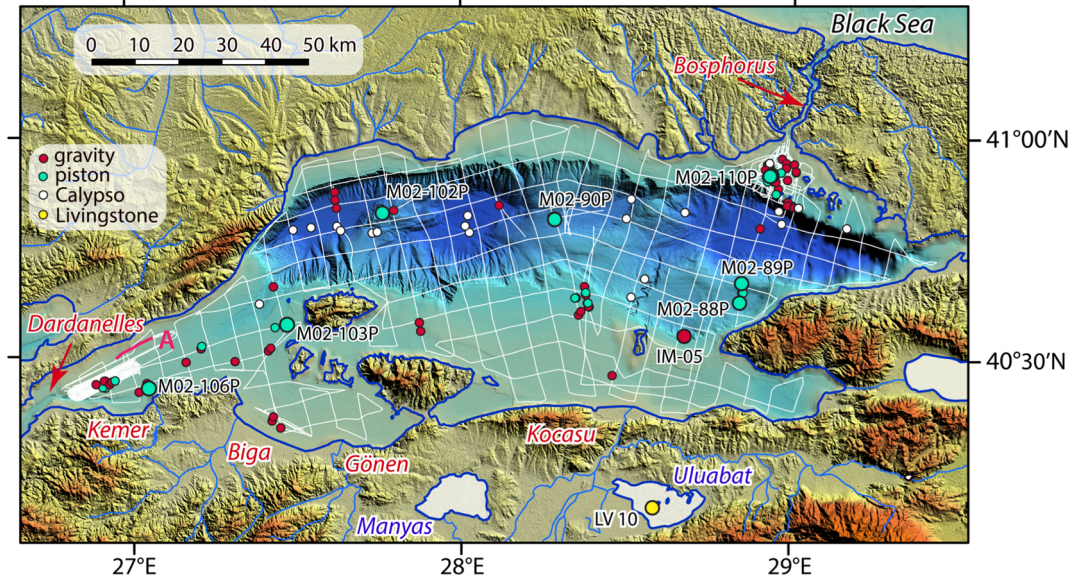
Figure 4
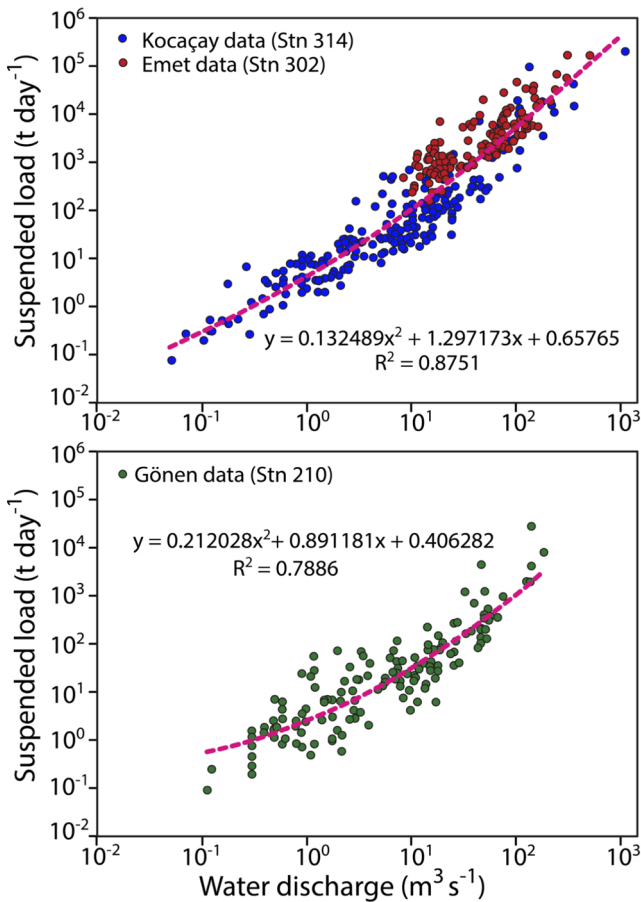
Figure 5
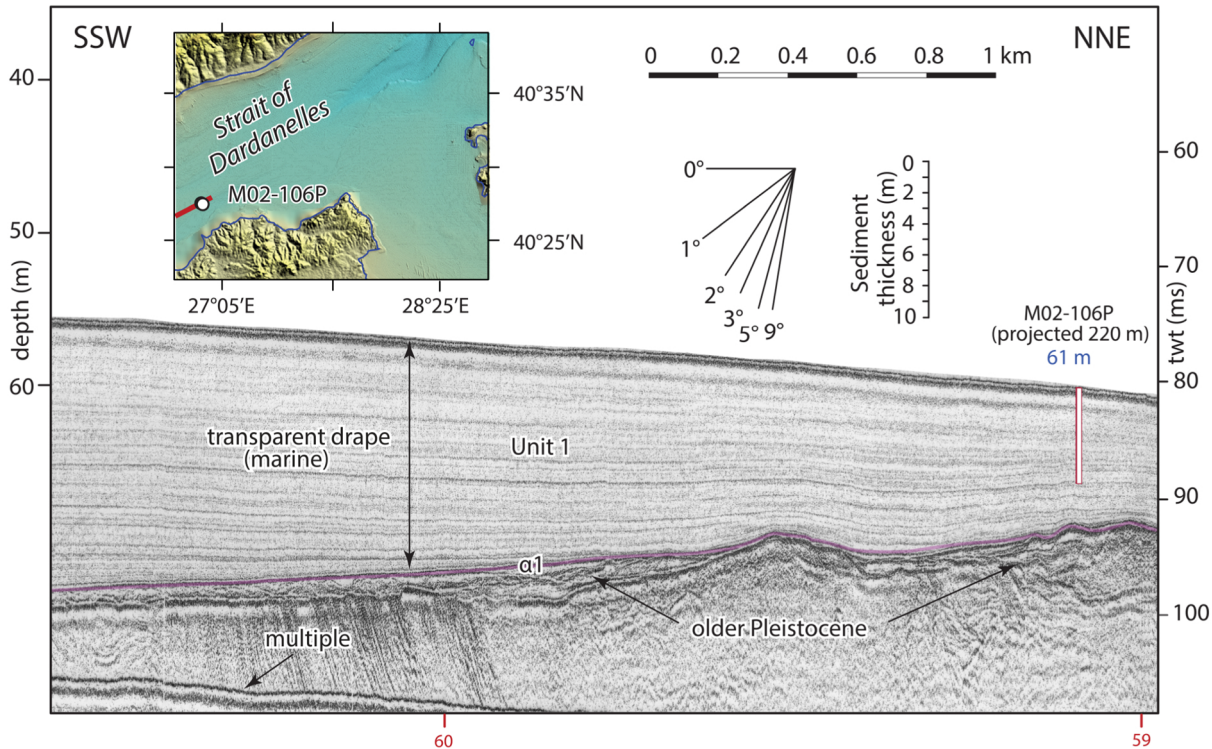
Figure 6
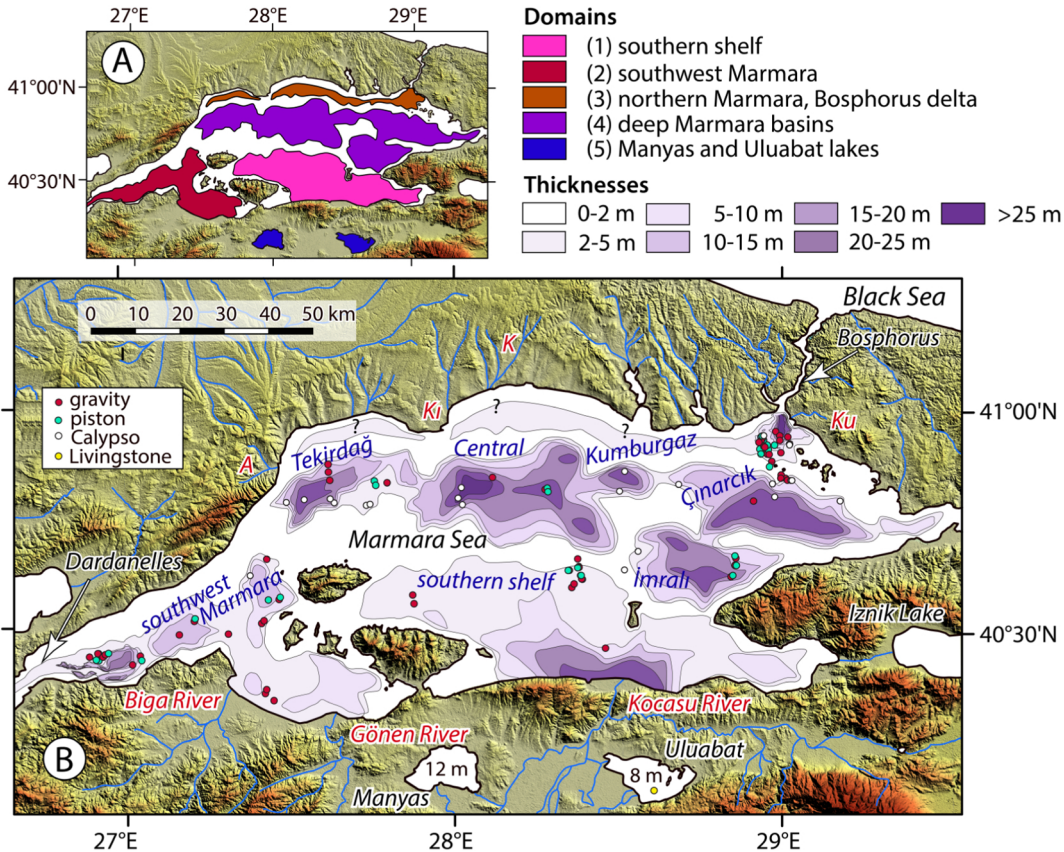
Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12