



Orbital- and millennial-scale environmental changes between 64 and 20 ka BP recorded in Black Sea sediments

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Abstract. High-resolution pollen and dinoflagellate cyst records from sediment core M72/5-25-GC1 were used to reconstruct vegetation dynamics in northern Anatolia and surface conditions of the Black Sea between 64 and 20 ka BP. During this period, the dominance of *Artemisia* in the pollen record indicates a steppe landscape and arid climate conditions. However, the concomitant presence of temperate arboreal pollen suggests the existence of glacial refugia in northern Anatolia. Long-term glacial vegetation dynamics reveal two major arid phases ~ 64–55 and 40–32 ka BP, and two major humid phases ~ 54–45 and 28–20 ka BP, correlating with higher and lower summer insolation, respectively. Dansgaard–Oeschger (D–O) cycles are clearly indicated by the 25-GC1 pollen record. Greenland interstadials are characterized by a marked increase in temperate tree pollen, indicating a spread of forests due to warm/wet conditions in northern Anatolia, whereas Greenland stadials reveal cold and arid conditions as indicated by spread of xerophytic biomes. There is evidence for a phase lag of ~ 500 to 1500 yr between initial warming and forest expansion, possibly due to successive changes in atmospheric circulation in the North Atlantic sector. The dominance of *Pyxidopsis psilata* and *Spiniferites cruciformis* in the dinocyst record indicates brackish Black Sea conditions during the entire glacial period. The decrease of marine indicators (marine dinocysts, acritarchs) at ~ 54 ka BP and increase of

freshwater algae (*Pediastrum*, *Botryococcus*) from 32 to 25 ka BP reveals freshening of the Black Sea surface water. This freshening is possibly related to humid phases in the region, to connection between Caspian Sea and Black Sea, to seasonal freshening by floating ice, and/or to closer position of river mouths due to low sea level. In the southern Black Sea, Greenland interstadials are clearly indicated by high dinocyst concentrations and calcium carbonate content, as a result of an increase in primary productivity. Heinrich events show a similar impact on the environment in the northern Anatolia/Black Sea region as Greenland stadials.

1 Introduction

The climatic conditions during the marine isotope stage 3 (MIS 3) are characterized by abrupt alternations of cold and warm intervals, known as Dansgaard–Oeschger (D–O) cycles or Greenland stadials (GS) and Greenland interstadials (GI), respectively (Dansgaard et al., 1993; Grootes et al., 1993; Svensson et al., 2006). During GS, ice sheets became unstable and armadas of ice drifted southwards in the North Atlantic, as indicated by increases in ice-rafted detritus (IRD) (Bond and Lotti, 1995). The most extreme phases of these cold intervals are the Heinrich events (HE) (Heinrich, 1988; Bond et al., 1993), which

are characterized by reduced sea-surface temperatures (SST) and sea-surface salinity (SSS) in the North Atlantic (Rahmstorf, 2002; Hemming, 2004). Although the ultimate physical mechanisms triggering D–O cycles are still under discussion (e.g. Kudrass et al., 2001; Rahmstorf, 2002; Wunsch, 2006), the climatic and environmental impacts of D–O cycles and Heinrich events are evident in numerous records across the Northern Hemisphere (Voelker et al., 2002), including pollen records from the Atlantic and western Mediterranean region (Allen et al., 1999, 2000; Sánchez Goñi et al., 2000, 2002; Combourieu-Nebout et al., 2002; Roucoux et al., 2005; see review by Fletcher et al., 2010). Pollen records from the western Mediterranean reveal a clear response of vegetation to D–O cycles with steppe landscapes prevailing during GS and temperate forests during GI (Voelker et al., 2002; Fletcher et al., 2010). In contrast, the effect of HE on vegetation are less clear, as several vegetation records from Mediterranean (ODP 976: Combourieu-Nebout et al., 2002, MD95-2043: Sánchez Goñi et al., 2002) and Atlantic (MD95-2039: Roucoux et al., 2005) provide evidence for a very strong impact, whereas the Atlantic record MD95-2042 (Sánchez Goñi et al., 2000, 2002) reveals a climatic impact of HE similar to that of GS. In contrast to the western Mediterranean, the environmental impacts of D–O cycles and HE in the eastern Mediterranean region are less well documented (Fletcher et al., 2010). High-resolution pollen records from Greece mainland (Tzedakis et al., 2004; Müller et al., 2011) and the Aegean Sea island Lesbos (Margari et al., 2009), reveal a high sensitivity and rapid responses of vegetation to D–O cycles, whereas records from the Caucasus (Arslanov et al., 2007) and the Cretan basins (Geraga et al., 2005) do not show a clear D–O-like pattern. These contradictory pollen data highlight the necessity for more high-resolution pollen studies in the eastern Mediterranean region, as pointed out by Fletcher et al. (2010).

Anatolia is a part of the Mediterranean region sensu lato (Lauer, 1993; Walter and Breckle, 1999; Weischet and Endlicher, 2000; Abrantes et al., 2012), and characterized by high plant diversity with ~8500 species (Kürschner et al., 1997), which results from the geographical complexity of this mountainous region and glacial–interglacial dynamics during the Pleistocene (Zohary, 1973). The southern and eastern Black Sea coastal mountainous regions, for instance, were most likely long-term glacial refugia for thermophilic trees (Leroy and Arpe, 2007). Occurrences of Paleogene/Neogene relicts such as e.g. *Pterocarya* and *Zelkova* in the Pontic Mountains and Caucasus suggest the presence of glacial refugia for thermophilic species in these regions (Zohary, 1973; Mayer and Aksoy, 1986). For the late glacial, the presence of temperate tree refugia in northern Anatolia was confirmed by pollen records (e.g. Bottema et al., 1995; Shumilovskikh et al., 2012). The Dziguta record in the Black Sea region shows also evidence for well-developed forests in the Caucasus during the last 50 ka BP (Arslanov et al., 2007). In contrast, the last glacial vegetation history of the Pontic

Mountains (northern Anatolia) remains largely unknown, as well-dated and highly resolved palaeobotanical records are missing. Sediment sequences from the Black Sea sediments are an excellent source of information on changes in vegetation as shown by palynological studies covering the late glacial period and the Holocene (e.g. Shumilovskikh et al., 2012; Filipova-Marinova et al., 2013) as well as the penultimate deglaciation and Eemian (Shumilovskikh et al., 2013a). To study the impact of D–O cycles and HE on the vegetation in northern Anatolia, we developed a pollen record for core 25-GC1 from the SE Black Sea, which covers MIS 3 continuously (Nowaczyk et al., 2012).

Furthermore, the last glacial history of the Black Sea basin is even less studied than the vegetation history of the surrounding areas, though it is well known that the Black Sea experienced marine and lacustrine phases during the Pleistocene due to marked global sea level changes (Degens and Ross, 1974; Ross et al., 1978; Neprochnov, 1980; Badertscher et al., 2011). While the last brackish–marine transition at ~8.3 ka BP is well studied in dinocyst and other records (Wall et al., 1973; Mudie et al., 2004; Atanassova, 2005; Bahr et al., 2006, 2008; Kwiecien et al., 2009; Marret et al., 2009; Buynevich et al., 2011; Bradley et al., 2012; Shumilovskikh et al., 2012, 2013b; Filipova-Marinova et al., 2013), details on the marine–brackish transition that occurred after the last interglacial, as well as of the sea-surface conditions of the Black Sea during the MIS 3, are largely unknown (Badertscher et al., 2011). Dinocyst studies for the Eemian suggest a sea-surface salinity in the Black Sea of ~30 (Shumilovskikh et al., 2013b), whereas studies for the late glacial reveal freshwater/brackish conditions (see references above). A reconstruction of the Black Sea surface temperatures (based on TEX₈₆) for the last 40 ka reveals clear imprints of HE, whereas clear GIs and GSs are not evident (Ménot and Bard, 2012). This is in contrast to well-expressed D–O cycles and HE in stalagmite oxygen and carbon isotopes records from Sofular Cave in northern Turkey (Fleitmann et al., 2009). Due to lack of Black Sea sea-surface reconstructions older than 40 ka BP, the timing and rate of hydrological changes of the basin during the last glacial period have remained poorly understood.

To fill this gap of knowledge, we present the first continuous high-resolution pollen and algal records from the SE Black Sea, covering the glacial interval 64–20 ka BP to reveal the long- and short-term changes in vegetation of northern Anatolia and sea-surface conditions of the Black Sea in great detail.

2 Environmental and climatic conditions of the Black Sea and northern Anatolia

Today, the Black Sea is the largest anoxic basin on Earth. The upper ~100–150 m water layer is oxygenated with a salinity of 17–18 PSU, whereas the whole water body below is

anoxic with a salinity of ~ 22 (Özsoy and Ünlüata, 1997). Such a characteristic vertical stratification of the Black Sea is caused by the interruption of vertical water circulation through different densities of the upper and lower water layers. The upper layer receives freshwater input from rivers and precipitation and is therefore less dense than the deeper layer, which is influenced by the inflow of saline (~ 37) Mediterranean water through the Bosphorus Strait (Kosarev and Kostianoy, 2008).

The present-day climate of northern Anatolia is controlled by seasonal changes in the position and strength of the Azores High and Icelandic Low, by the strength of the Siberian High in winter and associated changes in westerly storm tracks (Türkeş et al., 1996). In northern Anatolia, mean annual temperatures and precipitation range from ~ 10 to 14°C (average temperature in January 7°C and in July 23°C) and 600 to 2000 mm yr^{-1} , respectively (Kürschner et al., 1997). In contrast, a continental climate with extremely cold winters and long-lasting snow cover prevails in central Anatolia, where average temperatures in January range from around -3 to -8°C and the lowest temperatures can vary between -21 and -34°C . Summers are dry and hot with mean temperatures ranging between 20 and 23°C in July. Total rainfall varies from 280 and 420 mm yr^{-1} and peaks in the winter months (Kürschner et al., 1997; Kürschner and Parolly, 2012).

The modern distribution of the vegetation in the Pontic Mountains is closely related to orographic conditions (Atalay, 2002). The northwestern slopes of the Pontic Mountains receive precipitation from the Black Sea and are covered by temperate Euxinian forests. At altitudes between 0 and 800 m, *Fagus orientalis* associates with various deciduous trees (*Carpinus betulus*, *Castanea sativa*, *Tilia platyphyllos* (Rubra), *Corylus*, deciduous *Quercus* species etc.), evergreen shrubs (*Rhododendron ponticum*, *Ilex aquifolium*, *I. colchica*, *Daphne pontica*, *Buxus sempervirens*, *Prunus laurocerasus*) and lianas (*Smilax*, *Hedera*, *Clematis*, *Vitis*) (Zohary, 1973; Mayer and Aksoy, 1986). However, scattered Mediterranean trees or shrubs (*Phillyrea media*, *Laurus nobilis*, *Cistus creticus*, *Juniperus oxycedrus*, *Erica arborea*, *Myrtus communis*, *Arbutus andrachne*) can be found in warm and dry habitats along the Black Sea coast. At higher elevations, between ~ 800 and 1500 m, pure beech (*F. orientalis*) or hornbeam-beech (*Carpinus–Fagus*) forests predominate. Relatively low mountains (below 1500 m) explain the dominance of pure *Fagus* forests and the absence of *Abies* in the central part of northern Anatolia, whereas *Abies* dominates in the western and *Picea* in the eastern Pontic Mountains. Grasslands dominate in the subalpine and alpine mountain belt above ~ 2000 m. The south-eastern mountain slopes facing the Anatolian Plateau lie in the rain shadow and are therefore covered with drier vegetation, including open woodland dominated by eurythermic conifers (*Pinus nigra*, *Juniperus excelsa*, *J. oxycedrus*) and deciduous oaks (*Quercus pubescens*, *Q. trojana*). At the border between forest steppe

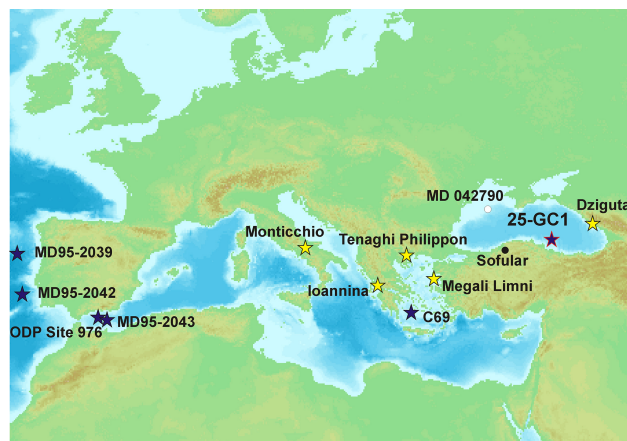


Fig. 1. Location of the Black Sea core 25-GC1 and records covering the last glacial as discussed in the text: stalagmites (black dot) – Sofular Cave (Fleitmann et al., 2009; Badertscher et al., 2011); TEX₈₆ temperature reconstructions (white dot) – MD042790 (Ménot and Bard, 2012); terrestrial pollen records (yellow stars) – Dziguta (Arslanov et al., 2007), Ioannina (Tzedakis et al., 2004), Megali Limni (Margari et al., 2009), Monticchio (Allen et al., 1999, 2000), and Tenaghi Philippon (Müller et al., 2011); marine pollen records (blue stars) – C69 (Geraga et al., 2005), MD95-2039 (Roucoux et al., 2005), MD95-2042 (Sánchez Goñi et al., 2000), MD95-2043 (Sánchez Goñi et al., 2002), ODP Site 976 (Combouret-Nebout et al., 2002).

and steppe, oaks and junipers grow. Above ~ 2000 m elevation subalpine thorny shrubs (*Astracantha* spec., *Sarcopoterium* spec.) are characteristic (Zohary, 1973; Mayer and Aksoy, 1986).

3 Material and methods

Sediment core 25-GC1 ($42^\circ 06.2' \text{ N}$, $36^\circ 37.4' \text{ E}$, 418 m water depth) was collected from the SE Black Sea (Fig. 1) during R/V *Meteor* cruise M 72/5 in 2007. The studied sediments consist of fine-grained siliciclastic material with calcium carbonate content varying between 15 and 40 % (Nowaczyk et al., 2012).

3.1 Chronological framework

The construction of the age–depth model for core 25-GC1 is described in detail by Nowaczyk et al. (2012) and presented here briefly. The basic chronology of core 25-GC1 is based on marker horizons such as the Laschamp geomagnetic excursion at 40.70 ± 0.95 ka BP (Bonhommet and Babkine, 1967; Gillot et al., 1979; Guillou et al., 2004; Plenier et al. 2007; Nowaczyk et al., 2012), the Campanian Ignimbrite tephra at 39.28 ± 0.11 ka (DeVito et al., 2001; Pyle et al., 2006), the Korudağ tephra at ~ 20 ka ± 6 ka (Kuzucuoglu et al., 1998), the Santorini tephra at ~ 1627 –1600 yr BC (Guichard et al., 1993; Friedrich et al., 2006),

and 10 AMS radiocarbon dates (Fig. 2). The final chronology as published in Nowaczyk et al. (2012) is based on tuning the 25-GC1 record to the GICC05 chronology of the North Greenland Ice core Project (NGRIP) (Svensson et al., 2008) and to the North Atlantic core MD95-2024 (Stoner et al., 2000). Tuning was carried out by the computer software “extended tool for correlation” (xtc), performing a simultaneous synchronization with interactive wiggle-matching routines of climate-related variations in the Black Sea records such as log-normalized IRD abundances, carbonate content, Ca XRF-counts, and high-resolution magnetic susceptibility.

3.2 Palynological investigations

In total, 163 subsamples of 1 cm³ of wet sediment were taken along the core and prepared using standard methods (Erdtman, 1960; Moore et al., 1999), including treatments with cold 10 % HCl, cold 71–74 % HF over two nights, acetolysis (three minutes) and final cleaning within an ultrasonic bath on nylon sieves of 1–2 µm mesh size. In order to calculate pollen concentrations (Stockmarr, 1971), one or two tablets of *Lycopodium* spores (Batch No. 177745, consisting of $18\,584 \pm 354$ spores) were added to each sample. We followed the pollen and spore nomenclature of Beug (2004) and Chester and Raine (2001). The pollen sum for calculation of pollen and spore percentages includes a minimum of 300–350 identified counted pollen grains per sample and consists of all identified taxa, excluding reworked pollen. In order to compare the pollen record from 25-GC1 with other European pollen records, we grouped pollen spectra to biomes, following the classification of Fletcher et al. (2010) and summarized typical Euxinian elements to a special group (Table 1). Complete pollen and dinocyst data are stored at PANGAEA (www.pangaea.de).

Based on preliminary results of a study of surface samples (Shumilovskikh, 2013), we suggest that pollen source areas for cores 25-GC1 and for 22-GC3/8 (Shumilovskikh et al., 2012; Shumilovskikh et al., 2013a) include the coastal regions close to the coring sites, namely the central part of northern Anatolia and especially the northern slopes of the Pontic Mountains.

In order to evaluate sea-surface conditions, dinoflagellate cysts (dinocysts), green algae (*Pediastrum*, *Botryococcus*, *Zygnemataceae*) and acritarchs (*Cymatiosphaera*, *Microhystridium*) were counted on the residues from preparation for palynological analyses. Due to the possible damage of cysts by acetolysis (Marret and Zonneveld, 2003) and therefore biased percentages of species in assemblages, we present records of dinocyst concentrations reflecting primary productivity (e.g. Shumilovskikh et al., 2012) or rather presence/absence of distinct species. Freshwater/brackish dinocysts include *Pyxidinospis psilata*, *Spiniferites cruciformis* and *Caspidinium rugosum*, and marine dinocysts include *Lingulodinium machaerophorum*, *Operculodinium centrocarpum*, *Tuberculodinium*

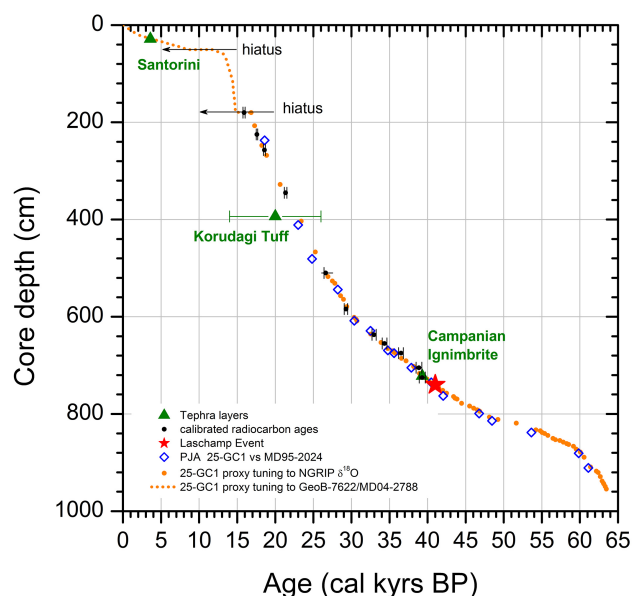


Fig. 2. Stratigraphical framework for Black Sea core 25-GC1: Santorini tephra (Guichard et al., 1993; Friedrich et al., 2006), Korudagi tephra (Kuzucuoglu et al., 1998), Campanian Ignimbrite tephra (De Vivo et al., 2001; Pyle et al., 2006), Laschamp Event (Bonhommet and Babkine, 1967; Gillot et al., 1979; Guillou et al., 2004; Plenier et al., 2007; Nowaczyk et al., 2012), PJA (paleointensity based on anhysteretic remanent magnetization) – relative paleointensity of 25-GC1 (Nowaczyk et al., 2012) tuned to MD95-2024 (Stoner et al., 2000), 25-GC1 proxy tuning to North Greenland Ice core Project (NGRIP) $\delta^{18}\text{O}$ record (Svensson et al., 2008) and to the Black Sea sediment records GeoB-7622 (Lamy et al., 2006) and MD04-2788 (Kwiecien et al., 2008).

vancampoe, *Tectatodinium pellitum*, *Bitectatodinium tepikiense*, *Pentapharsodinium dalei*, *Spiniferites pachydermus*, and *Spiniferites* spp. (Shumilovskikh et al., 2013b).

The pollen and algal diagrams shown in Figs. 3 and 4 were constructed using the software C2 (version 1.6.6) (Juggins, 2003). In order to highlight long-term trends, selected curves were smoothed using a simple moving average of 20 points.

4 Results

Pollen and algal records are presented in Figs. 3 and 4, respectively. Temporal resolution of the records varies between ~ 500 in the lower and ~ 100 yr in the upper part of the core section.

4.1 Pollen record

The pollen spectra 25-GC1 are dominated by herbs *Artemisia* (23–68 %), *Chenopodiaceae* (7–33 %) and *Poaceae* (3–16 %). Arboreal pollen percentages vary between ~ 6 and 47 %, in response to highly variable climatic conditions during the glacial period. The main tree pollen

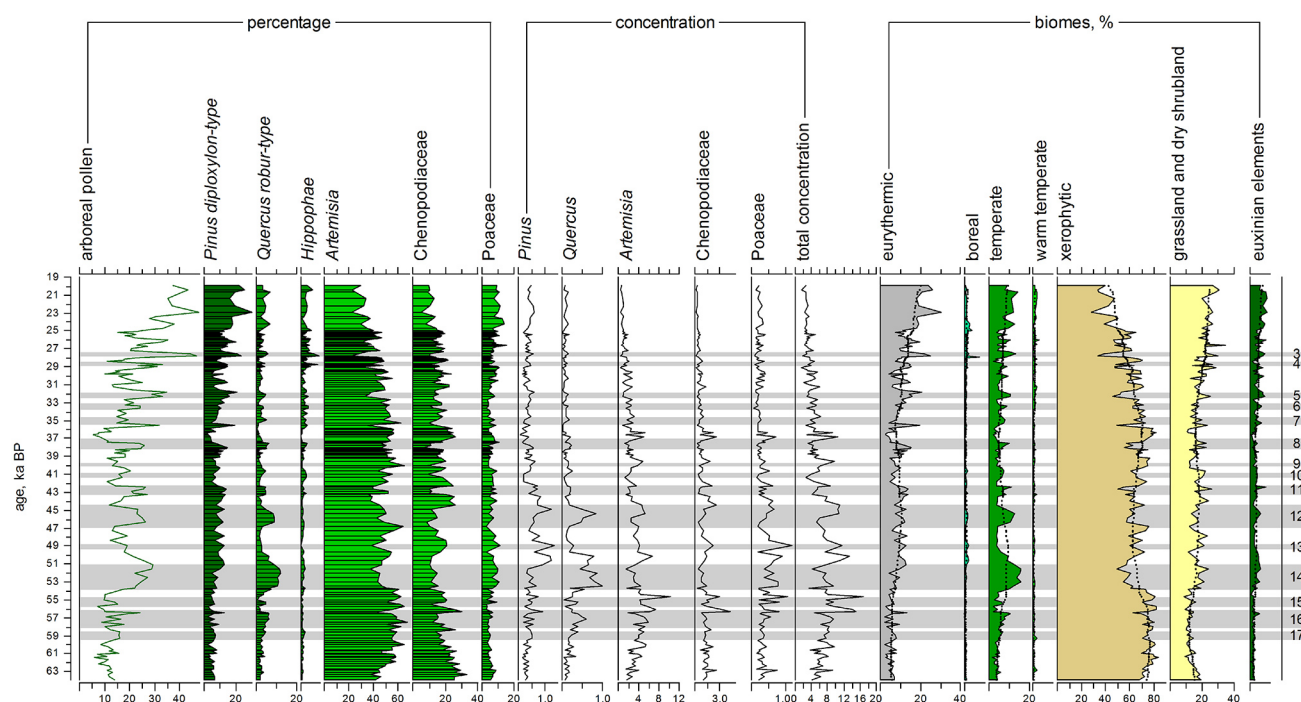


Fig. 3. Percentage (%) and concentration ($\text{grains} \times 10^3 \text{ cm}^{-3}$) diagram of main pollen taxa and percentages of biomes, grouped following Fletcher et al. (2010) (Table 1), versus age. Percentages were calculated on a total pollen sum. Eurythermic, temperate, xerophytic as well as grassland and dry shrubland biomes percentages are plotted with a smoothed curve (dotted curve, simple moving 20 points average). Horizontal grey bars with numbers indicate GI as identified by the 25-GC1 CaCO_3 (GI 17-5) and dinocyst (GI 4 and 3) records.

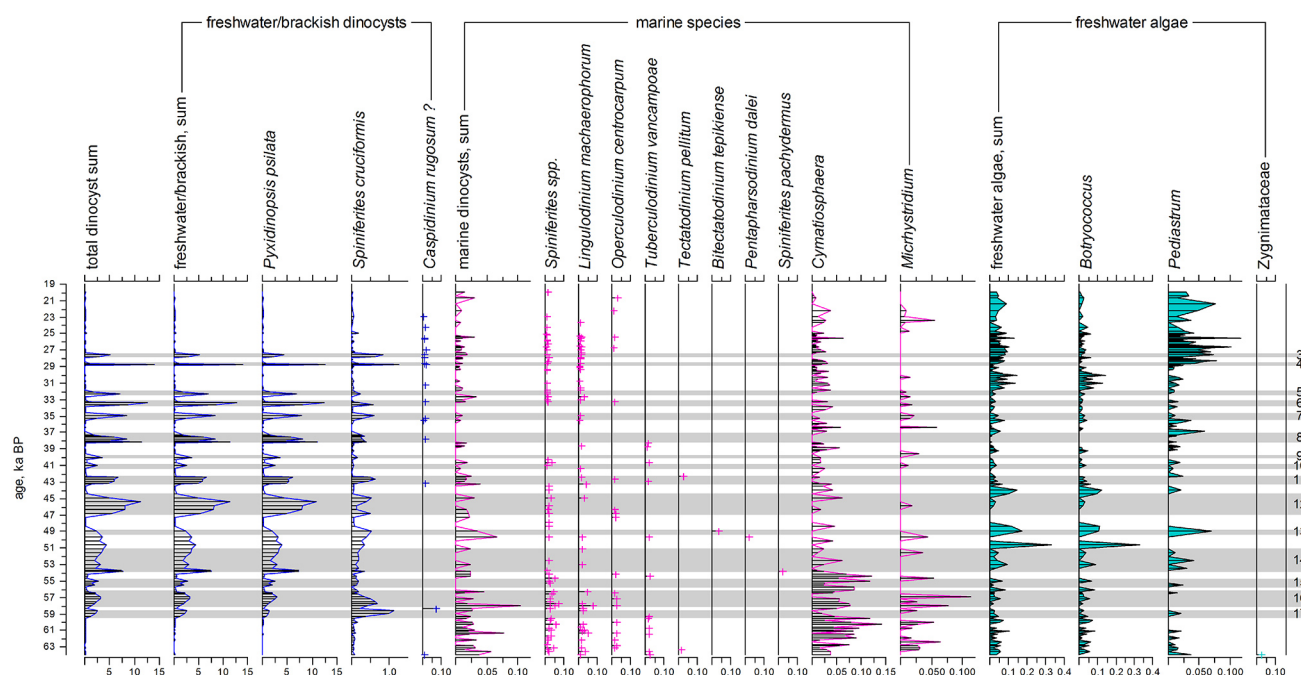


Fig. 4. Concentration diagram of dinoflagellate cysts, freshwater algae and acritarchs ($\text{specimens} \times 10^3 \text{ cm}^{-3}$) versus age. Horizontal grey bars with numbers indicate GI as identified by the 25-GC1 CaCO_3 (GI 17-5) and dinocyst (GI 4 and 3) records.

Table 1. Definition of mega-biomes for Europe based on Fletcher et al. (2010) and the northern Anatolian Euxinian forest.

Mega-biome	Characteristic pollen taxa
Warm temperate forest	<i>Buxus</i> , Cannabaceae, <i>Castanea</i> , <i>Celtis</i> , <i>Hedera helix</i> -type, <i>Ilex</i> , <i>Juglans</i> , <i>Olea</i> , <i>Ostrya</i> -type, <i>Phillyrea</i> , <i>Pistacia</i> , <i>Platanus</i> , <i>Pterocarya</i> , <i>Quercus</i> evergreen, <i>Quercus cerris</i> -type, <i>Rhamnus</i> -type, <i>Vitis</i> , <i>Tamarix</i> Additional pollen types: <i>Cotinus coggygria</i> , Cistaceae (<i>Helianthemum nummularium</i> group, <i>H. oelandicum</i> -type)
Temperate forest	<i>Acer</i> , <i>Alnus</i> , <i>Carpinus betulus</i> , <i>Cedrus</i> , <i>Cornus mas</i> -type, <i>Corylus</i> , <i>Fagus</i> , <i>Frangula alnus</i> , <i>Fraxinus</i> , <i>Myrica</i> , <i>Prunus</i> , <i>Quercus robur</i> -type, <i>Salix</i> , <i>Sorbus</i> , <i>Tilia</i> , <i>Ulmus</i> , <i>Viburnum</i> , <i>Sambucus</i> , <i>Taxus</i> Additional pollen types: <i>Helleborus foetidus</i>
Boreal	<i>Abies</i> , <i>Betula</i> , <i>Picea</i>
Eurythermic conifers	<i>Pinus</i> , <i>Juniperus</i>
Grassland and dry shrubland	Cyperaceae, Ericaceae, <i>Hippophae</i> , Poaceae, other NAP
Xerophytic steppe	<i>Artemisia</i> , Chenopodiaceae, <i>Ephedra</i> Additional pollen types: <i>Haloxylon</i>
Euxinian forest	<i>Abies</i> , <i>Picea</i> , <i>Fagus</i> , <i>Carpinus betulus</i> , <i>Corylus</i> , <i>Acer</i> , <i>Alnus</i> , <i>Ulmus/Zelkova</i> , <i>Tilia</i> , <i>Fraxinus excelsior</i> , <i>F. oxycarpa</i> , <i>Ilex</i> , <i>Cornus mas</i> -type, <i>Myrica</i> , <i>Frangula alnus</i> , <i>Viburnum lantana</i> , <i>Sambucus ebulus</i> , <i>Taxus</i> , <i>Castanea</i> , <i>Buxus</i> , <i>Hedera helix</i> -type, <i>Helleborus foetidus</i> , <i>Pterocarya</i>

taxa are dominated by *Pinus* (2–30 %) and *Quercus* (1–12 %). Pollen concentration vary between 1.6 and $13\text{--}17 \times 10^3$ grains cm^{-3} and peak between ~ 57 and 45 ka BP and then decline continuously towards 25 ka BP. Long-term (millennial) trends of biome changes are shown with smoothed curves (Fig. 3). They show an increase in eurythermic and grassland/dry shrubland biomes towards 25 ka BP, with a maximum in temperate biomes at $\sim 55\text{--}45$ ka BP and two maxima in xerophytic biomes at $\sim 60\text{--}55$ and 40–32 ka BP.

During GI, arboreal and temperate biome pollen increase by up to 15–47 and 10–16 %, respectively (Fig. 3). The most indicative taxon is the *Quercus robur*-type, which reaches 4–12 % and $100\text{--}1000$ grains cm^{-3} . The general proportion of the *Pinus Diploxylon*-type increases towards 20 ka BP, when they reach maxima of 14–24 %, whereas concentration maxima (up to 1300 grains cm^{-3}) occur during GI 14 to 12.

GS are characterized by a clear response of xerophytic biomes and an increase of non-arboreal pollen (NAP) percentages of up to 94 % (Fig. 3), mainly of *Artemisia* and Chenopodiaceae.

4.2 Algal record

Algal remains (Fig. 4) are dominated by dinocysts, reaching concentrations up to $11\,000\text{--}14\,000$ cysts cm^{-3} , with the main contribution from freshwater/brackish species *Pyxidinopsis psilata* and *Spiniferites cruciformis*. Concentrations of dinocysts are very low ($0\text{--}400$ cyst cm^{-3}) during GS and considerably higher ($1300\text{--}14\,000$ cyst cm^{-3}) during GI (Fig. 4).

Several marine dinocyst species occur occasionally in the record such as *Lingulodinium machaerophorum*, *Operculodinium centrocarpum*, *Spiniferites* ssp., *Tuberculodinium vancampoae*, *Tectatodinium pellitum*, *Bitectatodinium tepikiense*, and *Spiniferites pachydermus* (Fig. 4). The dinocyst record shows several maxima of marine species with a clear decreasing trend in concentrations since ~ 54 ka BP, similar to the marine acritarchs *Cymatiosphaera* and *Micrhystridium*. Freshwater algae *Pediastrum* and *Botryococcus* show several maxima with highest values prevailing between 51 and 47 ka BP (up to 70 and 330 specimens cm^{-3} , respectively; Fig. 4). A period of persistent occurrence of freshwater algae in the sediment begins after GI 5 at ~ 32 ka BP, whereas *Pediastrum* concentrations increase significantly from ~ 30 ka BP (up to $70\text{--}100$ specimens cm^{-3}).

5 Discussion

5.1 Vegetation

5.1.1 Long-term dynamics

The dominance of *Artemisia*, Chenopodiaceae, Poaceae, *Pinus* and *Quercus* in the glacial pollen spectra (Fig. 3) indicates a wide spread of steppe and forest-steppe in northern Anatolia between 64 and 20 ka BP, similar to modern ecosystems in central Anatolia. Since rainfall of less than 300 mm yr^{-1} favours treeless steppe vegetation in central Anatolia today (Kürschner and Parolly, 2012), the dominance of steppe is indicative of a considerable drop in precipitation from $600\text{--}1000\text{ mm yr}^{-1}$ today to possibly

$\sim 300\text{--}600\text{ mm yr}^{-1}$ between 64 and 20 ka BP. Furthermore, the glacial climate along the Turkish Black Sea coast was more continental and characterized by winter temperatures below 0°C . This assumption is also supported by the reduction of Euxinian vegetation (Fig. 3), which was wide spread during the last interglacial (Shumilovskikh et al., 2013a), due to generally cooler/drier conditions in the region and increased seasonality of precipitation. A reduction of summer temperatures during the glacial period is more than likely, but clear evidence for this is lacking in the pollen record. The presence of pollen of cool-temperate (*Fagus*, *Alnus*, *Corylus*, *Carpinus betulus*), warm-temperate (*Ostrya*-type, *Castanea*) and Mediterranean taxa (*Quercus ilex*-type, *Phillyrea*, *Pistacia*) (Fig. 3) suggests the existence of glacial refugia in regions where mean winter temperatures remained close to or above 0°C . In general, oscillating arboreal pollen values in the 25-GC1 record (Fig. 3) reveal fluctuating steppe and forest distribution in response to the generally unstable climatic conditions during the MIS4 to 2 in Eastern Mediterranean.

At around $\sim 64\text{--}60$ ka BP, low percentages of arboreal pollen (6–16 %), mainly *Pinus* and *Quercus*, and high percentages of *Artemisia* (Fig. 3) reveal rather open landscapes that are dominated by *Artemisia*-steppe with small populations of oaks and pines. The dominance of steppe taxa and subtle changes in the pollen spectra indicate rather stable cold/dry climate conditions.

During MIS 3, the glaciers retreated from their MIS 4 position and climate in Europe exhibited warmer conditions. MIS 3 GCM experiments show different patterns of sea level pressure and precipitation compared to today (Barron and Pollard, 2002). A substantial decrease in precipitation in southern Europe led to a significant expansion of temperate grasslands, as evidenced from pollen records (Fletcher et al., 2010). GCM simulations (Barron and Pollard, 2002) for northern Anatolia suggest decreases in summer temperature of 0 to 2°C at 40 ka BP and 2 to 4°C at 30 ka BP, and a drop in winter temperatures between 2 and 4°C for both periods. The dominance of open landscapes in northern Anatolia confirms lower precipitation between 64 and 25 ka BP.

Throughout MIS 3 ($\sim 57\text{--}29$ ka BP), the gradual decrease in the extension of temperate *Quercus* (Figs. 3 and 5) during GI and an increase in eurythermic *Pinus* in pollen assemblages of 25-GC1 (Figs. 3 and 5) indicates a general progressive cooling in northern Anatolia towards 28 ka BP. Modern pine forests in northern Turkey include *Pinus sylvestris* and *P. nigra*, both growing on the rain-shadow slopes of the Pontic Mountains at the upper forest limit of 1800 to 2000 m (Walter, 1956). The present-day oak woodlands and shrublands – represented in central Anatolia mainly by *Q. pubescens* and *Q. trojana* – together with *J. oxycedrus* and *J. excelsa* grow at lower forest limits, forming forest-steppe borders (Mayer and Aksoy, 1986; Uğurlu et al., 2012). The decreasing extent of oaks and expansion of pine in forests over the course of MIS 3 is thus possibly related to a downward migration of the forest line, leading to shrinking of oak

populations at northern slopes of the Pontic Mountains and the development of pine forests.

The smoothed biome curves reveal particular features in the records (Figs. 3 and 5). Xerophytic biome maxima (Fig. 5), characteristic of more arid conditions in northern Anatolia, occur at $\sim 64\text{--}55$ and $40\text{--}32$ ka BP, whereas the spread of temperate, eurythermic, boreal and grassland biomes (Figs. 3 and 5) occur at $\sim 54\text{--}45$ and at $28\text{--}20$ ka BP, suggesting more humid conditions during these periods. The high similarity of arid and humid periods in northern Anatolia with maxima and minima in summer insolation is remarkable. The spread of drought-adapted xerophytic vegetation occurs at times of higher summer insolation. Temperate biomes, however, seem to expand at times of lower summer insolation, suggesting a direct response of the environment in northern Anatolia to changes in summer insolation. We suggest that peaks in summer insolation resulted in higher temperatures and evaporation rates in summer, leading to higher water deficits in summer and the spread of steppe. Low summer insolation caused lower evaporation and higher moisture throughout the year, allowing forests to expand. Similar observations of enhanced seasonality at ~ 60 and ~ 33 ka BP in response to higher summer insolation were made by Fletcher and Sánchez Goñi (2008) for the western Mediterranean region.

The onset of MIS 2 is characterized by rather high AP percentages (15–48 %, Figs. 3 and 5) and indicates the expansion of woodlands related to an increase in moisture availability in northern Anatolia. The decrease in aridity at the onset of MIS 2 is also supported by the decreasing role of the xerophytic biome (Figs. 3 and 5), reflecting a contraction of steppe habitats. The increase in temperate biomes, grasslands and Euxinian elements (Fig. 3) suggests an increase in moisture availability throughout the year. However, the general climate in the region was considerably drier than today, but less arid compared to GS or MIS 4. In accordance, $\delta^{13}\text{C}$ isotopes from Sofular Cave vary between -7 and -8‰ (Fig. 5), which is more positive than $\delta^{13}\text{C}$ values during warm/wet GI (-8 to -10‰) and more negative than values of between -6 and -7‰ during cold/dry GS. This agrees with pollen records from the Sea of Marmara which also reveal rather high precipitation (600 mm yr^{-1}) during the LGM (Mudie et al., 2002).

Wet and cold conditions during MIS 2 are in good agreement with reconstructions of glacier dynamics in Anatolia during the last glacial maximum (LGM), which suggest that precipitation was fairly high (Akçar and Schlüchter, 2005; Sarıkaya et al., 2011). Using cosmogenic radioisotope dating, Akçar et al. (2007, 2008) found evidence that glacier expansion in the Kaçkar Mountains (NE Anatolia) started at $\sim 26 \pm 1.2$ ka BP and lasted until $\sim 18.3 \pm 0.9$ ka BP. Maximum glacier expansion occurred at $\sim 20.3 \pm 1.5$ ka BP in north-western Anatolia (Zahno et al., 2010), at $\sim 21.3 \pm 0.9$ ka BP in central Anatolia (Sarıkaya et al., 2009), and at $\sim 20.4 \pm 1.3$ ka BP in western Taurus

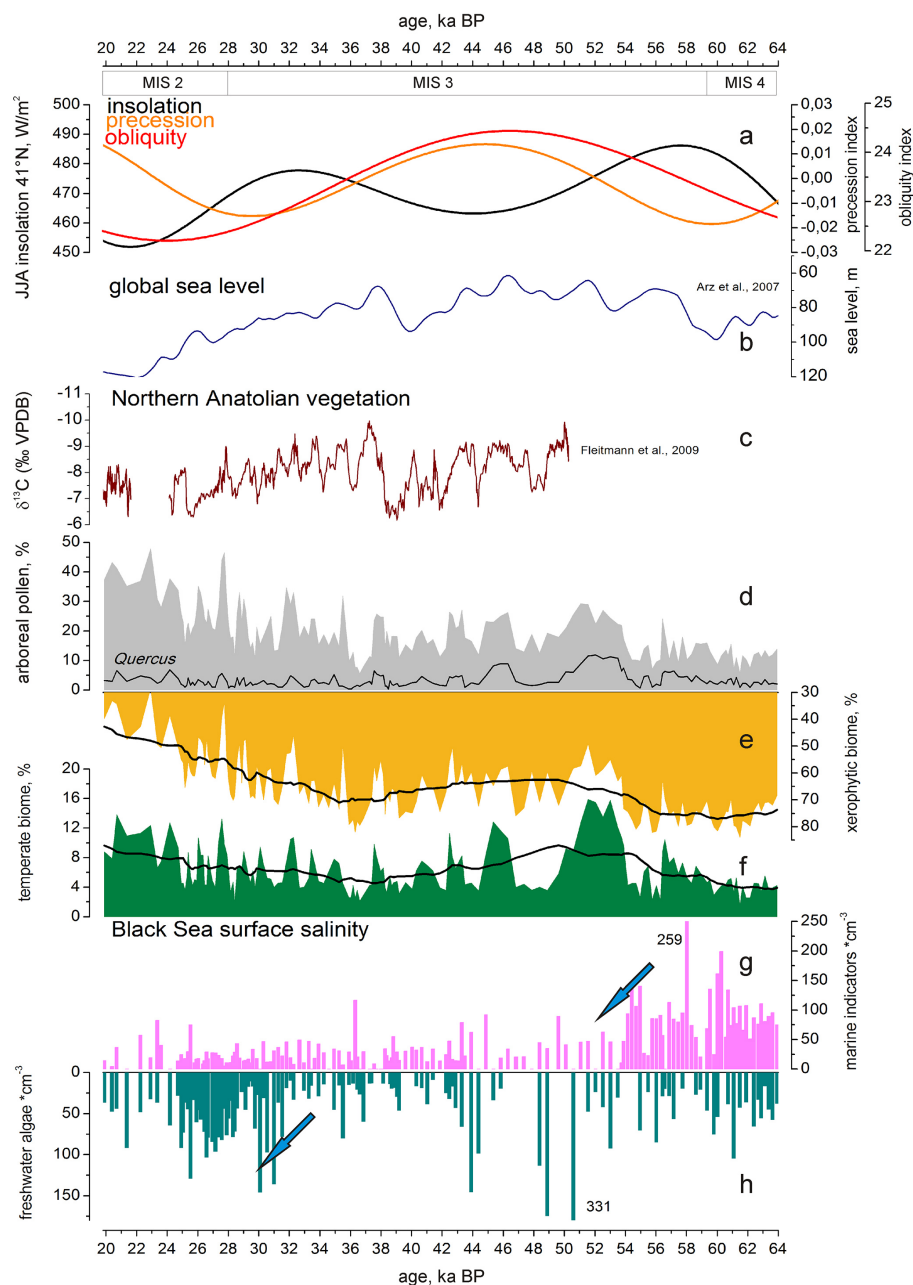


Fig. 5. Summary plot for northern Anatolian vegetation and Black Sea surface salinity in comparison to insolation and global sea level changes: **(a)** summer insolation at 41°N (W m^{-2}) (mean for June, July and August), precession and obliquity indices; **(b)** reconstruction of the global sea level (Arz et al., 2007); together with regional data from northern Anatolia: **(c)** $\delta^{13}\text{C}$ from Sofular Cave (Fleitmann et al., 2009) reflecting changes in C3 and C4 plants and from Black Sea core 25-GC1: **(d)** arboreal pollen record (%) and *Quercus* curve (thin line); **(e)** xerophytic biome (%) with smoothed curve (thick line; simple moving 20 points average) (please note the reverse scale); **(f)** temperate biome (%) with smoothed curve (thick line); **(g)** concentration of marine dinocysts and acritarchs (specimens cm^{-3}); **(h)** concentration of freshwater algae (specimens cm^{-3}). Blue arrows indicate two major freshening phases of the Black Sea surface. Marine isotope stages are indicated following Shackleton (1987).

(Sarikaya et al., 2008). Based on glacier expansions in western Taurus (Sarikaya et al., 2008), estimated LGM temperatures were between 8.5 and 11.5°C lower than modern and LGM precipitation up to 1.9 times higher than today. Since advances of glaciers during the MIS 2 required

snow accumulation, a higher amount of winter precipitation originating in the Black Sea in combination with shorter and cooler summers can be assumed for northern Anatolia. This scenario could potentially explain the reconstructed changes in vegetation towards cold/humid

conditions, occurring during a period when the Black Sea in general experienced a negative water balance that resulted in a lowered lake level (100 m b.s.l.; e.g. Popescu et al., 2004; Yanko-Hombach, 2007).

One striking features of the 25-GC1 pollen record is an increase of *Hippophae* since 36 ka BP (Fig. 3). *Hippophae*, a pioneer species that tolerates higher salt content in the air and soils, is capable of biological nitrogen fixation and typically grows in open, dry and sandy areas, such as coasts, river banks, subalpine elevation, and semi-desert sites. An increase of *Hippophae* indicates more open landscapes in the catchment area, which could be either the result of drier climatic conditions and/or increased delta dynamics and/or a possible decrease of the sea level of the Black Sea at the end of MIS 3. However, since the general trends in vegetation dynamics suggest development of wetter conditions in northern Anatolia, a combination of the two latter processes is very likely and are in agreement with reconstructed low levels of the Black Sea during the late Pleistocene (Popescu et al., 2004).

5.1.2 Abrupt climate changes

The high-frequency oscillations of AP and NAP in the 25-GC1 pollen record show a D–O-like pattern (Figs. 3 and 6).

GSs are characterized by an increase in NAP, mainly by Chenopodiaceae and partly by *Artemisia*, as well as by xerophytic biomes (Fig. 3), indicating arid conditions in northern Anatolia. In contrast, GIs show higher percentages and concentrations of *Quercus robur*-type as well as in Euxinian elements and temperate biomes (Figs. 3 and 6). This is probably related to a spread of temperate oak-dominated forests in northern Anatolia due to a warmer and more humid climate. Furthermore, a pronounced increase in *Quercus* percentages is characteristic for each GI, except for GI 13, which is only visible in the AP curve (Fig. 3). The important role of temperate *Quercus* during GI successions was documented by terrestrial (e.g. Müller et al., 2011) and marine (e.g. Combourieu-Nebout et al., 2002; Sánchez-Gómez et al., 2000, 2002; Roucoux et al., 2005) pollen records from the eastern and western Mediterranean regions and seems to be a typical feature for southern Europe (Fletcher et al., 2010).

The expansion of oaks (8–12 %; Figs. 3 and 6) and temperate biomes is most pronounced during GI 14 and 12, the longest and warmest intervals during MIS 4 to 2. In the nearby pollen records 22-GC3/8 from the SE Black Sea, such high percentages of *Quercus* were characteristic for initial warming intervals during termination 1 (Meiendorf/Bølling–Allerød) (Shumilovskikh et al., 2012) and termination 2 (Shumilovskikh et al., 2013a), suggesting very similar climatic conditions. Fletcher et al. (2010) demonstrate that the amplitude of vegetation responses to the longest GIs 17–16, 14, 12 and 8 depends on the geographical location of records, with pronounced vegetation responses below 40° N. In contrast, between 40 and 44° N only GI 14 and 12 had

a discernible impact of vegetation. In this aspect, the northern Anatolian vegetation changes derived from core 25-GC1 at 42° N seem to confirm this observation, although all GIs are clearly reflected. According to Sánchez Gómez et al. (2008) and Fletcher et al. (2010), extension of temperate forests during GIs was modulated by orbitally driven changes in atmospheric circulation. GI 14 and 12 occurred during obliquity maximum (preferential high-latitude warming) and precession maximum (reduced low-latitude warming) (Fig. 5), which might result in forest development above 40° N under warmer and wetter climatic conditions (Fletcher et al., 2010). Furthermore, such an orbital configuration favours a reduced seasonality and less strongly differentiated Mediterranean climate. Since the main pollen source region of 25-GC1 record is in northern Anatolia at ~41° N, both explanations should be considered.

The longest and the most pronounced GIs 17–16, 14, 12 and 8 in the 25-GC1 pollen record provide the unique possibility to compare potential phase relationships between changes in vegetation in northern Anatolia and sea-surface conditions of the Black Sea (Fig. 5). Our pollen record has a centennial resolution and shows a synchronous increase of arboreal pollen, temperate biome percentages and increase in productivity (increased calcium carbonate precipitation and dinocyst concentrations; Fig. 6) at the onset of GI. However, temperate biomes tend to reach their maxima ~500–1500 yr after the onset of warming. Such phase lags are also evident in pollen records near Iberia (Roucoux et al., 2005; Sánchez Gómez et al., 2009) as well as in the speleothem $\delta^{13}\text{C}$ record from Sofular Cave in northern Anatolia, where negative shifts of 1 to 3 ‰ (indicative of transition into interstadial conditions) occur within 252 ± 87 yr (Fleitmann et al., 2009). For GI 1 (14.5–12.9 ka BP), “delayed” response of temperate biomes to warming was also documented in northern Anatolia by pollen record 22-GC3 (Shumilovskikh et al., 2012). Studies on the evolution of Iberian vegetation within GI have shown that vegetation is responding to the successive atmospheric changes related with the position and strength of the westerlies within a GI (Sánchez Gómez et al., 2009). Such slow changes in atmospheric circulation could play an important role in the development of northern Anatolian vegetation. Alternatively, these phase lags between climate and vegetation development can be explained by the gradual soil development in response to increases in temperature/precipitation or simply by the slow ecosystem response of reaching of a new equilibrium, as suggested by Fleitmann et al. (2009).

5.2 Black Sea surface conditions

5.2.1 Long-term patterns

During the glacial period, the 25-GC1 dinocyst record reveals freshwater/brackish conditions with a salinity of ~7–12 as indicated by dominance of the *P. psilata* (with modern occurrence in the Baltic Sea; Matthiessen and Brenner,

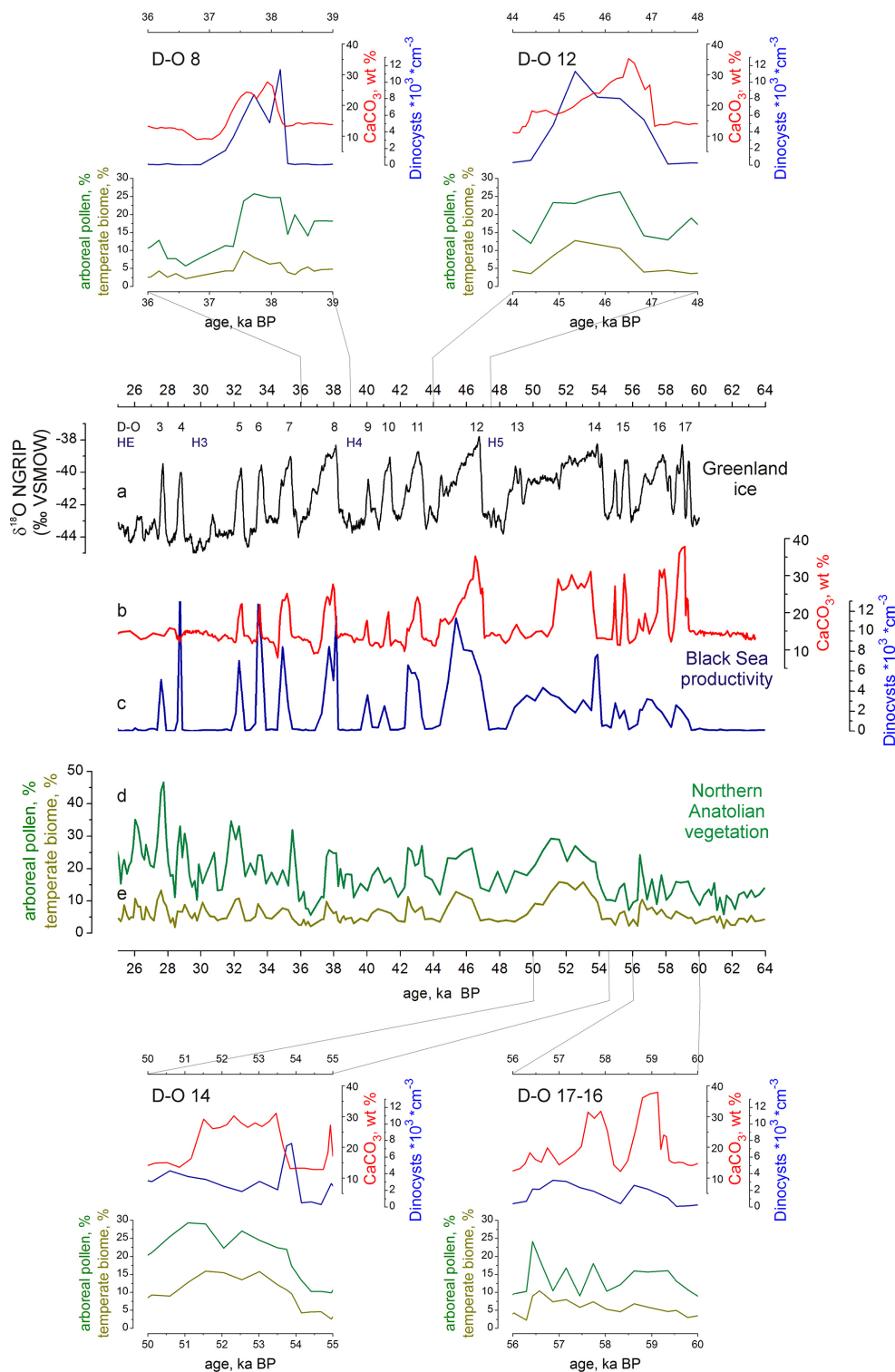


Fig. 6. Northern Anatolian vegetation and the Black Sea productivity in response to D–O and HE climatic variability: (a) $\delta^{18}\text{O}$ record (‰ VSMOW) from NGRIP (GICC05) ice core (Svensson et al., 2008) with numbers of GI and HE marked above the curve; (b) calcium carbonate record from core 25-GC1 (wt %); (c) total dinocyst concentration from core 25-GC1 (specimens $\times 10^3 \text{ cm}^{-3}$); (d) arboreal pollen (%) and (e) temperate biome (%) from core 25-GC1. The longest GI 17–16, 14, 12 and 8 are plotted in extra graphs, demonstrating a good correlation of northern Anatolian vegetation with the Black Sea productivity records.

1996; Yu and Berglund, 2007), and concurrent presence of *S. cruciformis*, known from the Caspian Sea (Marret et al., 2004) and modern freshwater lakes in Greece (Kouli et al., 2001) and Turkey (Leroy and Albay, 2010). The freshwater/brackish phase of the Black Sea indicates the absence of Mediterranean water inflow into the basin, because of the low global eustatic sea level during the last glacial (e.g. Siddall et al., 2003; Arz et al., 2007). This coincides with brackish to freshwater conditions reconstructed for the Sea of Marmara (Çağatay et al., 2009). A striking feature of the dinocyst record is the concurrent presence of the tropical/subtropical species *T. vancampoeae*, and fully marine indicators such as *S. pachydermus*, *B. tepikiense*, *T. pellitum* during the glacial brackish phase of the Black Sea (Fig. 4). Based on previous studies, these species entered the Black Sea basin during the Mediterranean transgression at the beginning of the Eemian and were abundant between 126.5–121 ka BP when subtropical temperatures and sea-surface salinity of ~ 28 –30 prevailed. Between 121 and 119 ka BP, their abundance decreased due to cooling and freshening of the sea-surface (Shumilovskikh et al., 2013b). The occurrence of subtropical and fully marine species between 64 and 25 ka BP prompts new questions, namely whether new dinoflagellate invasions occurred during intrusion of Mediterranean water into the Black Sea due to high global sea levels during MIS 5a and 5c (Dorale et al., 2010; Badertscher et al., 2011), or if these dinoflagellates had survived since the last interglacial. Further sediment archives from the Black Sea, covering the time window from 119 to 64 ka BP, are required to answer these questions. In any case, the presence of these tropical/subtropical species during the freshwater/brackish phase of the Black Sea is interesting in two respects. Firstly, their presence suggests SSS in the Black Sea at the maximum limit of *P. psilata*, possibly 12–15. Secondly, it could indicate a broader tolerance of these species than suggested from their modern distribution (Marret and Zonneveld, 2003). The occurrence of marine species in the brackish Black Sea environment during MIS 3 is also known from sediment cores from the Bulgarian shelf (Yanko-Hombach, 2007). While Mediterranean molluscs were absent in these cores, the occurrence of marine foraminifers such as *Ammonia tepida*, *Nonion matagordanus* and *Elphidium ponticum* was documented. In contrast to the author's explanation of inflow of Mediterranean water during the early MIS 3, we propose the absence of such a connection and the slow extinction of marine species through steady freshening of the Black Sea.

A general freshening of sea surface conditions towards 20 ka BP is indicated by a decreasing concentration of marine species and an increasing concentration of freshwater algae. However, this freshening was not gradual. While sediment records from the Sea of Marmara suggest an outflow of Black Sea water during the MIS 3 (Çağatay et al., 2009), the decrease in marine indicators at ~ 54 ka BP and increase in freshwater algae at ~ 32 –25 ka BP indicate two major freshening phases (Fig. 5). The decrease of marine species

concentrations at ~ 54 ka BP could indicate a rise of the Black Sea level at the beginning of MIS 3 through increased freshwater input by river runoff or precipitation, resulting in a positive hydrological balance of the Black Sea. This change in Black Sea hydrology correlates well with the spreading of temperate forests, suggesting more humid conditions on land (Fig. 5, Sect. 5.1.1). The second freshening phase indicated by the increase of freshwater algae concentrations at ~ 32 –25 ka also correlates with the establishment of more humid conditions on land, as suggested from pollen record 25-GC1 (Fig. 5, Sect. 5.1.1). This period occurs at the end of the high level of the Black Sea during ~ 40 –27 ka BP (Tarkhankutian beds), proposed from mollusc and foraminiferal reconstructions (Yanko-Hombach, 2007). Moreover, reconstructions of Volga discharge into the Caspian Sea suggest connection between the Caspian and Black Seas through the Manyč–Kerch strait at around 30 ka BP (Kroonenberg et al., 1997). The occurrence of the Caspian dinocyst *Caspidinium rugosum* between 29 and 23 ka BP (Fig. 4) is further evidence for the connection between the Caspian and Black Sea. Such a connection would result in an additional and considerable input of freshwater into the Black Sea basin and an outflow into the Sea of Marmara (Çağatay et al., 2009). An alternative explanation for freshening of the Black Sea surface at ~ 32 ka BP could be floating ice, indicated by high amounts of ice-rafted debris (Nowaczyk et al., 2012), which could have led to a slight seasonal freshening of the sea-surface in spring.

During MIS 2, the Black Sea level dropped to 100 m b.s.l., evidenced by formation of the Danube Canyon (Popescu et al., 2004). A lower Black Sea level is supported in the 25-GC1 record by an increase in *Hippophae* percentages (see Sect. 5.1.1). Such higher concentrations of the freshwater algae can be linked to the closer position of the Kızılırmak and Yeşilırmak river mouths to the core location as well as to a seasonal sea-surface freshening by floating ice.

5.2.2 Abrupt climate changes

High concentrations of dinocysts, mainly of the freshwater/brackish species *P. psilata* and *S. cruciformis* in core 25-GC1 (Fig. 4), indicate enhanced productivity of the Black Sea during GI. This assumption is supported by the high CaCO_3 concentrations (Fig. 6), which are indicative of authigenic calcium carbonate precipitation during phytoplankton blooms and resultant CO_2 depletion and pH increase (Bahr et al., 2005). Interestingly, both records show sometimes slightly different responses to GI. However, while dinocyst concentrations increase during each GI, GI 4 and 3 are not apparent in the CaCO_3 record. In contrast to CaCO_3 , the dinocyst records do not show interruptions between GI 14 and 13 (Fig. 5). Possibly both proxies reveal different thresholds to warming.

Periods of very low concentrations of dinocysts and calcium carbonate, indicative of cold temperatures, correlate

well with high amounts of ice-rafted debris in the sediment core 25-GC1, which are evidence for extreme cold winter temperatures (Nowaczyk et al., 2012).

Interestingly, dinocysts data from core 25-GC1 contradict TEX₈₆ reconstructions from the NW Black Sea (Ménot and Bard, 2012), which do not reveal the D–O-related dynamics. These differences could reveal geographical patterns in D–O dynamics or different sensitivity to warming. However, further studies on TEX₈₆ and other paleoenvironmental proxies in SE Black Sea as well as dinocysts studies in the NW Black Sea are needed to resolve this question.

6 Conclusions

The last glacial pollen and algal records from Black Sea core 25-GC1 clearly reveal that the climate in the Black Sea/northern Anatolian region is strongly influenced by orbital-, millennial- and multi-centennial-scale climate variability.

1. Between 64 and 20 ka BP, the dominance of *Artemisia* and *Chenopodiaceae* in the pollen record 25-GC1 indicates a steppe landscape and arid climate conditions. However, the concomitant presence of temperate and warm-temperate arboreal pollen suggests the existence of glacial refugia in northern Anatolia.
2. Long-term glacial vegetation dynamics in northern Anatolia reveal two major arid phases at ~64–55 and 40–32 ka and two major humid phases at ~54–45 and 28–20 ka BP. All of these phases correlate with maxima/minima of summer insolation, respectively. Based on this correlation we propose a direct influence of summer insolation on the evaporation rate and thus annual moisture distribution in northern Anatolia. Furthermore, a humid phase at ~54–45 ka BP occurred during the obliquity maximum, which favoured warmer/wetter conditions and forest spread in latitudes north of 40° N. Increase in AP and temperate biomes between 25 and 20 ka BP suggests that the LGM was apparently more humid than the end of MIS 4 or GS, even if it was colder/drier than modern conditions.
3. The general dominance of *Pyxidopsis psilata* and *Spiniferites cruciformis* in dinocyst assemblages reveal brackish surface conditions between 64 and 20 ka BP. Further freshening of the Black Sea surface occurred in two major phases at ~54 and ~32–25 ka BP, which might have been caused by humid phases in the Black Sea region and/or a possible Caspian Sea connection, influencing hydrology and sea-level changes the Black Sea. The increased input of freshwater by rivers through a closer position to the core location during low water levels of the Black Sea as well as seasonal surface freshening by floating ice could play additional role.
4. D–O cycles are clearly indicated by pollen and algal records in 25-GC1. GIs are characterized by the spreading of temperate forests in northern Anatolia, suggested from increased *Quercus* pollen and Euxinian elements in the pollen assemblages, and by enhanced primary productivity in the lake system of the Black Sea, indicated by increased concentrations of dinocysts and calcium carbonate. GSs are characterized by a low primary productivity in the Black Sea and spread of xerophytic vegetation in northern Anatolia, indicating cold climatic conditions.
5. Maximum forest development in response to GI warming is reached ~500–1500 yr after productivity changes in the Black Sea, possibly due to successive atmospheric changes during GIs and/or slow soil development.
6. Expansion of the forest in response to GI was likely modulated by orbital parameters, so that the major temperate forest development during the longest GI 14 and 12 occurred during obliquity maximum and precession maximum.
7. HEs show a similar impact on the northern Anatolia/southern Black Sea environment as the other GSs.

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