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Discussions

Orbital and millennial-scale environmental changes between 64 and 25 ka BP recorded in Black Sea sediments

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Abstract

High-resolution pollen and dinoflagellate cyst records from marine sediment core 25-GC1 were used to reconstruct vegetation dynamics in Northern Anatolia and surface conditions of the Black Sea between 64 and 25 ka BP. During this period, the domi-
5 nance of *Artemisia* in the pollen record indicates a steppe landscape and arid climate conditions. However, the presence of temperate and warm-temperate arboreal pollen suggests the existence of glacial refugia in Northern Anatolia. A general cooling trend towards 25 ka BP is evidenced by the decrease of *Quercus* and increase of *Pinus*. There is evidence of orbital-driven vegetation dynamics in Northern Anatolia during
10 64–25 ka BP with spread of steppe during precession minima (insolation maxima) and development of forests during precession maxima (insolation minima). Dansgaard–Oeschger (D–O) events are characterized by a marked increase in temperate tree pollen, indicating a spread of forests due to warm and wet conditions in Northern Anatolia. The dominance of *Pyxidinopsis psilata* and *Spiniferites cruciformis* in the dinocyst
15 record indicates a rather brackish Black Sea during the last glacial period. The decrease of marine indicators (marine dinocysts, acritachs) 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, related to orbital-driven arid/humid phases in the region, influencing hydrology and level changes of the Black Sea. D–O interstadials are char-
20 acterized by high dinocyst concentrations and calcium carbonate content, as a result of an increase in primary productivity in the Black Sea. Heinrich events show a similar impact on the environment in Northern Anatolia/Black Sea region as D–O stadials.

1 Introduction

The climatic conditions during the marine isotope stage 3 (MIS 3) are characterized
25 by dramatic and rapid alternations of cold and warm intervals, known as Dansgaard-Oeschger (D–O) cycles (Dansgaard et al., 1993; Grootes et al., 1993). During cold

phases of D–O cycles, the Greenland ice sheet became instable and armadas of ice drifted southwards in Northern Atlantic, as indicated by marked 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 character-
5 ized by low sea-surface temperatures (SST) and reduced 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 records across the northern hemisphere
10 (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 cold and arid stadials and temperate
15 forests during more warm and humid interstadials (Voelker et al., 2002; Fletcher et al., 2010). In contrast to well-expressed D–O cycles in the pollen records, the effect of HE on Mediterranean vegetation are unclear, as several vegetation records show evidence for a very strong impact (e.g. Combourieu Nebout et al., 2002; Roucoux et al., 2005; Margari et al., 2009), whereas others do not (e.g. Sánchez Goñi et al., 2000, 2002).
20 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). The high-resolution terrestrial pollen records from the Balkans (Tzedakis et al., 2004; Müller et al., 2011) and the Aegean Sea island Lesvos (Margari et al., 2009), for instance, reveal a high sensitivity and rapid responses of vegetation to D–O cycles,
25 whereas terrestrial pollen record from the Caucasus (Arslanov et al., 2007) and the Cretan Basin (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 also pointed out by Fletcher et al. (2010).

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Anatolia is a part of the Mediterranean region (Abrantes et al., 2012), and characterized by high plant diversity with ∼ 8500 species (Kürschner et al., 1997). Such a high diversity results from the geographical complexity of this mountainous region and glacial-interglacial dynamics during the Pleistocene (Zohary, 1973). Indeed, the southern and eastern Black Sea coastal mountainous regions were most likely long-term glacial refugia for thermophilic trees (Leroy and Arpe, 2007), and occurrences of tertiary 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, presence of temperate trees refugia in Northern Anatolia was confirmed by pollen records (e.g. Bottema et al., 1995; Shumilovskikh et al., 2012). Longer glacial vegetation records in the Black Sea region are represented by the Dziguta record, which evidences 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 high resolution palaeobotanical records are missing. Sediment sequences from the Black Sea sediments are an excellent source of information on changes in vegetation as revealed 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 vegetation in Northern Anatolia, we established 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 during the Pleistocene, the Black Sea experienced marine and lacustrine phases 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 by 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.,

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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 still unknown. Indeed, 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). The only one TEX$_{86}$ reconstruction of the Black Sea surface temperatures for the last 40 ka reveals clear imprints of HE, whereas D–O events 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 the 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 largely unknown.

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 64–25 ka BP to reveal the long- and short-term changes in vegetation of Northern Anatolia and sea-surface conditions of the Black Sea in unprecedented 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 ~ 200 m water layer is oxygenated with a salinity of 17–18, 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 therefore is less dense than the deeper layer, which is influenced by the inflow of saline (~ 37) Mediterranean water through the Bosporus 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 temperature 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, long-lasting snows and regular frosts prevails in central Anatolia, where average temperatures in January range from –2.7 to –7.8 °C and minimum temperatures can vary between –21.0 and –34.4 °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 forests. At altitudes between 0 and 800 m, *Fagus orientalis* associates with various deciduous trees (*Carpinus betulus, Castanea sativa, Tilia 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, Myrthus 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 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

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

3 Material and methods

The sediment core 25-GC1 (42°06.2′ N, 36°37.4′ E, 418 m water depth) was collected from the SE Black Sea (Fig. 1) during RV *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 Korudağıtephra 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 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-normalised IRD abundances, carbonate content, Ca XRF-counts, and high-resolution magnetic susceptibility.

3.2 Palynological investigations

In total, 152 subsamples of $1\,cm^3$ 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 ± 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 include a minimum of 300–350 identified counted pollen grains per sample. 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).

In order to evaluate sea-surface conditions, dinoflagellate cysts (dinocysts), green algae (*Pediastrum, Botryococcus*, Zygnemataceae) and acritarchs (*Cymatiosphaera, Micrhystridium*) were counted on the residues from preparation for palynological analyses. Due to the possible damage of cysts by acetolysis and therefore biased percentages of species in assemblages, we present records of dinocyst concentrations, which reflect primary productivity (e.g. Shumilovskikh et al., 2012) or rather presence/absence of distinct species. Freshwater/brackish dinocysts include *Pyxidinopsis psilata, Spiniferites cruciformis* and *Caspidinium rugosum*, and marine dinocysts include *Lingulodinium machaerophorum, Operculodinium centrocarpum, Tuberculodinium vancampoae, Tectatodinium pellitum, Bitectatodinium tepikiense, Penthapharsodinium dalei, Spiniferites pachydermus*, and *Spiniferites* spp. (Shumilovskikh et al., 2013b).

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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 smooth out short-term fluctuations and highlight long-term trends, we smoothed selected curves by using a simple moving average of 20 points.

4   Results

Pollen and algal records are presented in Figs. 3 and 4, respectively. Grey shaded bars denote D–O events as identified by the 25-GC1 $CaCO_3$ record (Fig. 6). Because D–O events 3 and 4 are absent in $CaCO_3$ record, they are indicated based on 25-GC1 dinocyst record (Figs. 4 and 6).

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 taxa are dominated by *Pinus* (2–14 %) and *Quercus* (1–12 %). The pollen concentration vary between 1.6 and 13–17 $\times 10^3$ 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 ~ 55–45 ka BP and two maxima in xerophytic biomes at ~ 60–55 and 40–32 ka BP.

During the D–O events, 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 100–1000 grains cm^{-3}. The general proportion of the *Pinus diploxylon*-type increases towards 25 ka BP, when they reach maxima of 14–

24 %, whereas concentration maxima (up to 1.300 grains cm^{-3}) occur during D–O 14 to 12.

Cold phases of D–O cycles are characterized by a clear response of xerophytic biomes and an increase in non-arboreal pollen (NAP) percentages of up to 94 %
5 (Fig. 3), mainly by *Artemisia* and Chenopodiaceae.

4.2 Algal record

Algal remains (Fig. 4) are dominated by dinocysts, reaching 11 000–14 000 cysts cm^{-3}, with the main contribution of freshwater/brackish species *Pyxidinopsis psilata* and *Spiniferites cruciformis*. Concentrations of dinocysts are very low (0–400 cyst cm^{-3})
10 during cold phases of D–O cycles and relatively high (1300–14 000 cyst cm^{-3}) during their warm phases (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
15 *Spiniferites pachydermus* (Fig. 4). The dinocyst record shows several maxima of marine species with a clear decreasing trend in concentrations towards 25 ka BP, similar to the marine acritarchs *Cymatiosphaera* and *Micrhystridium*. The smoothed curve for marine indicators exhibits the strongest decrease in concentrations at ∼ 54 ka BP (Fig. 5).
20 Freshwater algae *Pediastrum* and *Botryococcus* show several maxima with highest values prevailing between 53 and 44 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 D–O 5 at ∼ 32 ka BP, whereas *Pediastrum* concentrations increase from ∼ 30 ka BP significantly (up to 70–100 specimens cm^{-3}). The smoothed curve for
25 freshwater algae reveals two intervals of elevated concentrations: from ∼ 54 to 45 ka BP and 34 to 25 ka BP (Fig. 5).

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) indicate a wide spread of steppe and forest-steppe in Northern Anatolia between 64 and 25 ka BP. This is very similar to modern ecosystems in central Anatolia, where generally drier and more continental climatic conditions prevail than in the Pontic Mountains (Atalay, 2002). Since rainfall of less than 300 mm yr^{-1} today favours treeless steppe vegetation in central Anatolia (Kürschner and Parolly, 2012), the dominance of steppe between 64 and 25 ka BP at the coast indicates a considerable drop in precipitation from 600–1000 mm yr^{-1} today to possibly ∼ 300–600 mm yr^{-1} during the glacial period. Furthermore, the climate along the Turkish Black Sea coast was more continental and characterized by considerably lower winter temperatures of below 0 °C. Moreover, the reduction of euxinian vegetation (Fig. 3), which was wide spread during the last interglacial (Shumilovskikh et al., 2013a), indicates a seasonally heterogenous distribution of precipitation throughout the year with possible summer droughts. A reduction in summer temperatures during the glacial period is very likely, but clear evidence for it 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) suggest the existence of glacial refugia in the Pontic Mountains where mean winter temperatures stayed above 0 °C. In general, oscillating arboreal pollen values in the 25-GC1 record (Fig. 3) reveal fluctuating steppe and forest distribution, confirming the generally assumed unstable climatic conditions during the last glacial.

At the end of MIS 4 (∼ 64–60 ka BP), low percentages of arboreal pollen, mainly *Pinus* and *Quercus*, and high percentages of *Artemisia* suggest rather open landscapes (Fig. 3), dominated by *Artemisia*-steppe with small populations of oaks and pines. The

dominance of steppe taxa and rather subtle changes in the pollen spectra indicate cold/dry and stable climate conditions.

Throughout MIS 3 (∼ 60–28 ka BP), a gradual decrease in the extension of temperate *Quercus* (Figs. 3 and 5) during warm phases of the D–O cycles and an increase in eurythermic *Pinus* (Figs. 3 and 5) indicates a general cooling in Northern Anatolia towards 28 ka BP, suggested from comparison with modern distributions of pine and oak woodlands. 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 (1800–2000 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 the lower forest limits, forming forest-steppe borders (Mayer and Aksoy, 1986; Uğurlu et al., 2012). The overall trend of decreasing extension of oaks during D–O events and increase of pine in forests during MIS 3 is possibly related to a downward migration of the forest line due to decreasing temperatures, thus leading to shrinking of oak populations at northern slopes of Pontic Mountains and the development of pine forests.

The visual similarity of the smoothed curves of xerophytic and temperate biomes with summer insolation at 30° N (Fig. 5) is remarkably high. Xerophytic biome maxima, characteristic for more arid conditions in Northern Anatolia, between ∼ 60 and 55 and 40 and 32 ka BP coincide with summer insolation maxima. The maxima of temperate biomes between ∼ 55 and 45 ka BP reveals more humid conditions at a time of minimum summer insolation.

The general increase in arboreal pollen values from 6–16 % at 64–60 ka BP to 15–45 % at 28–25 ka BP (Figs. 3 and 5) indicates the expansion of woodlands and is most likely related to an increase in moisture availability for tree growth in Northern Anatolia. The general decrease in aridity at the end of MIS 3 is also supported by the decreasing role of the xerophytic biome since ∼ 34 ka BP (Figs. 3 and 5), reflecting shrinking of steppe habitats. The increase in euxinian elements (Fig. 3) suggests an increase in moisture availability throughout the year.

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One of the striking features of the 25-GC1 pollen record is an increase of *Hippophae* since 36.5 ka BP (Fig. 3). *Hippophae*, a pioneer species that tolerates higher salt content in the air and soil, is capable of biological nitrogen fixation and typically grows in open, dry and sandy areas, such as coasts, river banks, subalpine elevation, and dry semi-desert sites. An increase of *Hippophae* indicates more open landscapes in the catchment area, as a 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. Since the general trends in vegetation dynamics suggest development of wetter conditions in Northern Anatolia, a combination of the two latter processes are 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 show a distinct D–O like pattern in the 25-GC1 pollen record (Figs. 3 and 6).

Cold phases during D–O cycles are characterized by an increase in NAP, mainly by Chenopodiaceae and partly by *Artemisia*, and by the xerophytic biomes (Fig. 3), indicating arid conditions in Northern Anatolia. In contrast, warm phases of D–O cycles 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 an onset of warmer and more humid conditions. Furthermore, a pronounced increase in *Quercus* percentages is characteristic for each particular D–O event, except for events 17-16 and 10-9 and D–O 13, which is only visible in the AP curve (Fig. 3). The important role of temperate *Quercus* during D–O interstadial successions was documented by terrestrial (e.g. Müller et al., 2011) and marine (e.g. Combourieu Nebout et al., 2002; Sánchez-Goñi et al., 2000, 2002; Roucoux et al., 2005) pollen records from eastern and western Mediterranean regions and seems to be a typical feature for southern Europe (Fletcher et al., 2010).

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The expansion of oaks and temperate biomes is the most pronounced for D–O events 14 and 12, and both events are the longest warm intervals during the last glacial, when *Quercus* reaches 8–12 % (Figs. 3 and 6). In the nearby pollen records 22-GC3/8 from the SE Black Sea, such high percentages of *Quercus* were characteristic for ini-

5   tial warming intervals during termination 1 (Meiendorf/Bølling-Allerød) (Shumilovskikh et al., 2012) and 2 (Shumilovskikh et al., 2013a) suggesting similar climatic conditions. Fletcher et al. (2010) demonstrate that the amplitude of vegetation responses to the longest D–O events 17-16, 14, 12 and 8 depends on the geographical location of records. At latitudes below 40°N there is a pronounced vegetation response to all

10   considered events, while this is only the case for D–O events 14 and 12 in records between 40° and 44° N. In this aspect, the Northern Anatolian vegetation changes derived from core 25-GC1 seem to confirm this observation, although all D–O events are clearly reflected.

In contrast to terrestrial pollen records from eastern Mediterranean (e.g. Margari et

15   al., 2009; Müller et al., 2011) and TEX$_{86}$ measurements from the Black Sea (Ménot and Bard, 2012), the Black Sea record 25-GC1 does not show pronounced responses of vegetation to HE, which seems to have a similar impact on the Northern Anatolian vegetation to the cold phases of D–O cycles. Furthermore, the most severe climatic conditions in Northern Anatolia occurred after D–O 8 between ∼ 37 and 36.2 ka BP,

20   when AP reaches the lowest values of ∼ 6–8 % and when the dominance of xerophytic *Artemisia* and Chenopodiaceae (Fig. 3) suggests the spread of steppe under arid conditions. HE are associated with a southward extension of polar waters and ice drifting in the North Atlantic (Bond et al., 1993; Bond and Lotti, 1995), initiating the shutdown of North Atlantic Deep Water (NADW) formation and consequently of the thermoha-

25   line circulation (Stocker, 1999; Elliot et al., 2002). This led to an additional cooling of the North Atlantic (Clark et al., 2002) and a southward migration of the polar front (Margari et al., 2009). During HE, polar water reached more southern positions in the North Atlantic, causing more arid conditions on the adjacent land areas. This had a visible impact on the northern Iberian vegetation (Roucoux et al., 2005), but was less

pronounced in southern Iberia (Sánchez-Goñi et al., 2000). Ingression of the polar wa-
ter into Mediterranean Sea through the Strait of Gibraltar led to more severe climatic
conditions in western Mediterranean region during HE rather than during D–O stadials
(Combourieu Nebout et al., 2002; Sánchez-Goñi et al., 2002). However, the absence
of a clear vegetation response in Northern Anatolia to HE suggests a weaker influence
of North Atlantic cooling in the Black Sea region, hence a lack of significant change in
25-GC1 pollen assemblages during HE. In this case, local terrestrial records from the
Eastern Mediterranean (e.g. Margari et al., 2009; Müller et al., 2011) are more sensitive
to detect a cooling associated with HE.

The longest and the most pronounced D–O events 17-16, 14, 12 and 8 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, indicated
by calcium carbonate precipitation and dinocyst concentrations (Fig. 6), at the onset
of a D–O event. However, temperate biomes reach their maxima ~ 500–1500 yr after
the onset of warming. Similar trends are documented by pollen records from western
Mediterranean (Roucoux et al., 2005) as well as by the δ^{13}C record in the Sofular
speleothem (So-1) from Northern Anatolia, where negative shifts of 1 to 3 ‰ (indica-
tive of transition into interstadial conditions) occur within 252 ± 87 yr (Fleitmann et al.,
2009). Possibly, these phase lags between climate and vegetation development can
be explained by the gradual soil development in response to increases in tempera-
ture/precipitation, leading a to rather slow spread of temperate forests.

5.2 Black Sea surface conditions

5.2.1 Long-term patterns

During the glacial period, the dinocyst record reveals freshwater/brackish conditions
with a salinity of ~ 7–12 ‰ as indicated by dominance of *P. psilata* with modern oc-

currence in the Baltic Sea (Matthiessen and Brenner, 1996; Yu and Berglund, 2007), and *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 Mediter-

5 ranean water inflow into the basin, because the global sea level was below the modern Bosporus depth of −37 m during the last glacial (e.g. Siddall et al., 2003; Arz et al., 2007).

   A striking feature of the dinocyst record is the concurrent presence of the tropi-cal/subtropical species *T. vancampoae,* and fully marine indicators such as *S. pachy-*

10 *dermus, 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, the abundance of these species de-

15 creased 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 a new question, namely whether a new dinoflagellate invasions occurred dur-ing a Mediterranean water intrusion into the Black Sea due to high global sea levels during MIS 5a and 5c (Badertscher et al., 2011), or if these dinoflagellates had survived

20 since the last interglacial. Further Black Sea archives, in particular covering the time window from 119 to 64 ka BP, would be needed to answer this question. In any case, presence of these tropical/subtropical species during the freshwater/brackish phase of the Black Sea is interesting in two respects. Firstly, they suggest SSS in the Black Sea at the maximum limit of *P. psilata*, possibly 12–15. Secondly, it could indicate a broader

25 tolerance of these species than suggested from their modern distribution (Marret and Zonneveld, 2003).

   A further striking feature of the record is a decreasing concentration of marine species and increasing concentration of freshwater algae, suggesting a general fresh-ening of sea surface conditions towards 25 ka BP, which, however, was not gradual.

The decrease in marine indicators at ~54 and increase in freshwater algae at ~34–25 ka BP reveals freshening phases (Fig. 5), which could be explained by different processes. On one hand, a decrease of the marine indicators at ~54 correlate with spreading of temperate forests between 55 and 45 ka BP, suggesting more humid summers during the summer insolation minima. Furthermore, the sedimentological records from the Marmara Sea indicate a rather high sea level until ~35 ka BP (Çağatay et al., 2009), suggesting possible Black Sea discharge through the MIS 3 as a result of a positive hydrological balance of the Black Sea. On the other hand, high freshwater algae concentrations at ~34–25 ka could be linked to the closer position of the Kızılırmak and Yeşilırmak river mouths to the core location as a result of a decrease of the Black Sea level. This assumption is supported by an increase in *Hippophae* percentages (see Sect. 5.1.1) and by the formation of the Danube Canyon during the late glacial period (Popescu et al., 2004). Moreover, increasing arboreal pollen values towards 25 ka BP (Figs. 3 and 5) suggest rather wet conditions at the end of the MIS 3 (see Sect. 5.1.1). Floating ice, indicated by high amounts of ice rafted debris, could have been an additional source of the freshwater (Nowaczyk et al., 2012), leading to freshwater inputs during springs and therefore to the seasonal freshening of the sea-surface. It is likely that the decrease of the sea-surface salinity during the MIS 3 may have been caused by a combination of different factors and further records such as geochemical, sedimentological etc. are required to verify our assumptions.

5.2.2 Abrupt climate changes

High concentrations of dinocysts, mainly of freshwater/brackish species *P. psilata* and *S. cruciformis* in core 25-GC1 (Fig. 4), reveal an enhanced productivity of the Black Sea during D–O events. 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 D–O events. However, while dinocyst concentrations increase during each D–O event, D–O events

4 and 3 are not apparent in the $CaCO_3$ record. In contrast to $CaCO_3$, the dinocyst records do not show interruptions between D–O 14 and 13 (Fig. 5). Possibly both proxies reveal different thresholds to warming (Shumilovskikh et al., 2012). Furthermore, 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 TEX_{86} reconstructions from 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 TEX_{86} in SE Black Sea as well as dinocysts studies in NW Black Sea are needed to resolve this question.

5.3 Land-sea correlation and regional comparison

In general, the last glacial environmental reconstructions of the Black Sea region from core 25-GC1 reveal orbitally-driven dynamics as a response to summer insolation variations and ice volume changes as well as millennial-scale variability triggered by abrupt changes in the North Atlantic. In this section we will discuss possible causes and mechanisms of these terrestrial and aquatic reactions in comparison to other Mediterranean records.

MIS 4, 3, and 2 are based on benthic oxygen isotopes from the North Atlantic and were suggested to reflect the global ice volume and relative sea level (Shackleton, 1987). During MIS 4 and 2, characterized by the highest positive isotope values throughout the past 100 ka (Fig. 5c), ice sheets were the most advanced and sea level was the lowest. During MIS 4, pollen-based climatic reconstructions suggest cold and arid conditions in the western (Roucoux et al., 2005) and eastern Mediterranean (Margari et al., 2009). In accordance with this, the Black Sea pollen record from core 25-GC1 reveals a dominance of arid steppe conditions in Northern Anatolia and low concentrations of dinocysts reveal low sea-surface temperatures in the Black Sea dur-

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ing 64–60 ka, which are also indicated by high amounts of ice-rafted debris in the sediment (Nowaczyk et al., 2012).

In contrast to MIS 4, pollen and algal records reveal rather wet conditions towards MIS 2. Spread of forests and especially of euxinian formations (Fig. 5) reveal a reduced seasonality of precipitation, whereas freshening of the Black Sea surface suggests a change in the Black Sea water balance related to higher precipitation, higher river input or lower evaporation. In general, reconstructions of wet and cold conditions at the beginning of MIS 2 from the Black Sea pollen record 25-GC1 are in good agreement with reconstructions of glacier dynamics in Anatolia during the last glacial maximum (LGM), when precipitation was fairly high (Akçar and Schlüchter, 2005; Sarikaya et al., 2011). Using cosmogenic radioisotope dating, Akçar et al. (2007, 2008) found that glacier advances at Kaçkar Mountains (NE Anatolia) started at least $\sim 26 \pm 1.2$ ka BP and lasted up to $\sim 18.3 \pm 0.9$ 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 (Sarıkaya et al., 2008). Based on glacier expansions in western Taurus (Sarıkaya 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 in combination with shorter and cooler summers can be assumed for Northern Anatolia. Rather wet conditions during the MIS 2 were also documented for the western Mediterranean region by pollen records from north-west Iberia (Roucoux et al., 2005) and high lake levels from Lake Lisan (Bartov et al., 2003), correlating with an increasing global ice volume, leading to Northern Hemisphere cooling (Shackleton, 1987), and decreasing summer insolation (Berger, 1978).

During MIS 3, the glaciers retreated and climate in Europe experienced interstadial conditions. The 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

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grasslands, as evidenced from pollen records (Fletcher et al., 2010). GCM simulations (Barron and Pollard, 2002) for Northern Anatolia suggest a summer temperature decrease of between 0° and 2 °C at 40 ka BP and 2° and 4 °C at 30 ka BP, and a drop in winter temperatures between 2° and 4 °C for both time periods. The dominance of open

5 landscapes in Northern Anatolia confirms lower precipitation between 64 and 25 ka BP.

The pollen and algal records from core 25-GC1 reveal close correlations of vegetation and sea-surface salinity changes with summer insolation and precession (Fig. 5). Maxima in summer insolation coincide with spread of xerophytic vegetation under arid conditions and a reduced freshening of the Black Sea, whereas distributions of the tem-

10 perate biomes correspond to summer insolation minima. A direct response of the environment to the summer insolation changes is therefore likely. For Northern Anatolian vegetation, maxima in summer insolation might suggest higher temperatures and consequently higher evaporation rates during the warm period of the year, thus leading to minima of available moisture in summer, enabling forest spreading, and to an increase

15 of the Black Sea salinity. And vice versa, summer insolation minima should cause a decrease in evaporation and more equal moisture conditions around the year, thus allowing forest spreading and providing more freshwater input. However, an alternative explanation proposed by Fletcher and Sánchez Goñi (2008) suggests that higher summer insolation at low latitudes, driven by precession minima at ∼ 60 and ∼ 33 ka BP,

20 enhanced seasonality in the Mediterranean region. This seems to be also true for the Black Sea region as is suggested from Northern Anatolian vegetation changes (Fig. 5e, f) and from surface salinities in the Black Sea (Fig. 5h). However, the western Mediterranean and the Black Sea regions reveal sensitivity of different vegetation types to precession (insolation) changes. Precession minima (insolation maxima) led to

25 the development of Mediterranean vegetation in the western Mediterranean (Fletcher and Sánchez Goñi, 2008) and steppe in Northern Anatolia (Fig. 5f), whereas precession maxima (insolation minima) induced spread of Atlantic vegetation in the western Mediterranean (Fletcher and Sánchez Goñi, 2008) and temperate forests in Northern Anatolia (Fig. 5e). These precessional-driven changes were probably linked to atmo-

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spheric teleconnections between the Asian monsoon and the eastern Mediterranean region (Fletcher and Sánchez Goñi, 2008).

Orbitally-driven changes in atmospheric circulation had an obvious impact on the extension of temperate forests in Northern Anatolia associated with D–O interstadials. D–O events 14 and 12 occurred during precession-driven summer insolation minima and coincide with the strongest development of temperate forests in Northern Anatolia (Figs. 5e and 6e), whereas D–O 17-16 and 8 occurred near summer insolation maxima and are characterized by lower extent of temperate forest in Northern Anatolia. This observation is in accordance with the hypothesis of D–O forest modulations suggested by Sánchez Goñi et al. (2008) and Fletcher et al. (2010). An enhanced Mediterranean climate with warmer and drier summers and wetter winters was suggested for D–O 16-17 and 8, which are near precession minima and preferential low latitude warming. D–O 14 and 12 occurred during obliquity maximum and precession maximum, which suggest less strongly differentiated Mediterranean climate with summer cooler/wetter and reduced seasonality in precipitation (Fletcher et al., 2010).

6   Conclusions

The last glacial pollen and algal records from Black Sea core 25-GC1 clearly indicate that the Black Sea/Northern Anatolian region is a highly sensitive area for orbital, millennial and multi-centennial scale climate variability.

1. Long-term vegetation changes between 64 and 25 ka BP reveal a decrease in temperatures towards MIS 2, indicated by the decreasing role of *Quercus* and increasing role of *Pinus*. Moreover, development of euxinian vegetation towards 25 ka BP suggests that the beginning of MIS 2 was apparently more humid than the end of MIS 4.

2. The increase in concentrations of freshwater algae and decrease in concentrations of marine dinocysts and acritarchs suggest a lowering of the Black Sea sur-

face salinity towards MIS 2. Freshening of the Black Sea surface was not gradual and might have been caused by orbital-driven arid/humid phases in the Black Sea region, influencing hydrology and sea-level changes the Black Sea.

3. Long-term glacial vegetation dynamics in Northern Anatolia are strongly suggested to be driven by the precession cycle, indicated by spreading of temperate forests and more humid conditions during summer insolation minima as well as xerophytic biomes and more arid situation during summer insolation maxima. These changes could have been caused by a direct influence of summer insolation and/or through atmospheric teleconnections.

4. D–O interstadials are characterized by spreading of temperate forests in Northern Anatolia, suggested from increased *Quercus* pollen and euxinian elements, and by enhanced primary productivity in the lake system of the Black Sea, indicated by increased concentrations of dinocysts and calcium carbonate.

5. D–O stadials are characterized by a low primary productivity in the Black Sea and spread of xerophytic vegetation in Northern Anatolia, indicating cold climatic conditions. The impact of cold HEs on the Northern Anatolian vegetation is not substantially different than the impact of other D–O stadials.

6. Extension of the forest in response to D–O events was likely modulated by precession: the major forest development during the longest D–O 14 and 12 occurred during summer insolation minimum.

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Table 1. Definition of megabiomes for Europe based on Fletcher et al. (2010) and the Northern Anatolian euxinian forest.

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

Fig. 1. Map of the Mediterranean region with the location of the Black Sea core 25-GC1 as well as records covering the last glacial as discussed in the text: stalagmites (black dot) – Sofular Cave (Fleitmann et al., 2009; Badertscher et al., 2011); TEX$_{86}$ 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), OPD Site 976 (Combourieu Nebout et al., 2002).

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Fig. 2. Stratigraphical framework for Black Sea core 25-GC1: Santorini tephra (Guichard et al., 1993; Friedrich et al., 2006), Korudağıtephra (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 – 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) δ^{18}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).

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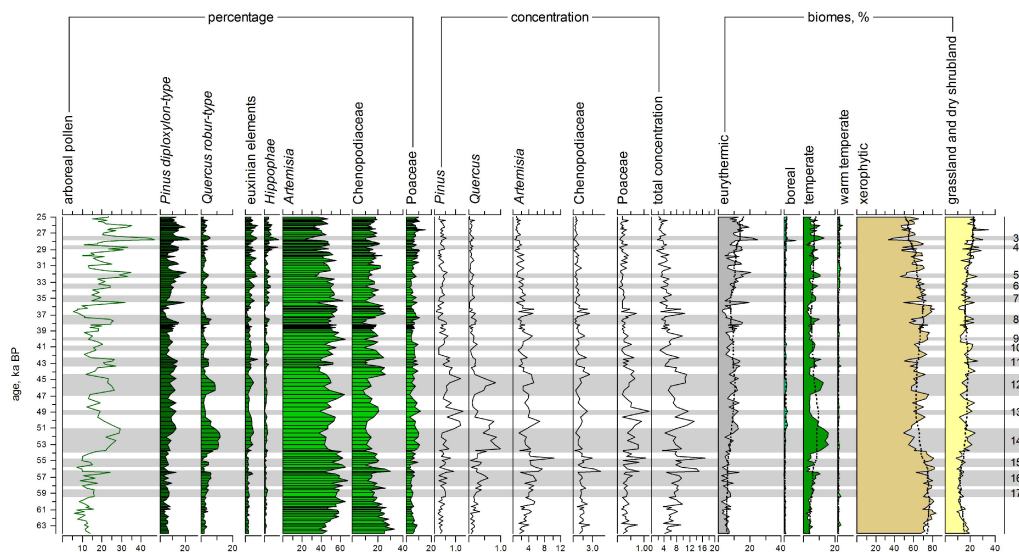
Fig. 3. Percentage (%) and concentration (grains $*cm^{-3}\, 10^{-3}$) diagram of main pollen taxa and percentages of biomes, grouped following Fletcher et al. (2010) (Table 1), versus age. Percentages were calculated on a total pollen sum. Eurythermic, temperate, xerophytic as well as grassland and dry shrubland biomes percentages are plotted with smoothed curve (dotted curve, simple moving 20 points average). Horizontal grey bars with numbers indicate D–O events as identified by the 25-GC1 CaCO$_3$ (D–O 17-5) and dinocyst (D–O 4 and 3) records.
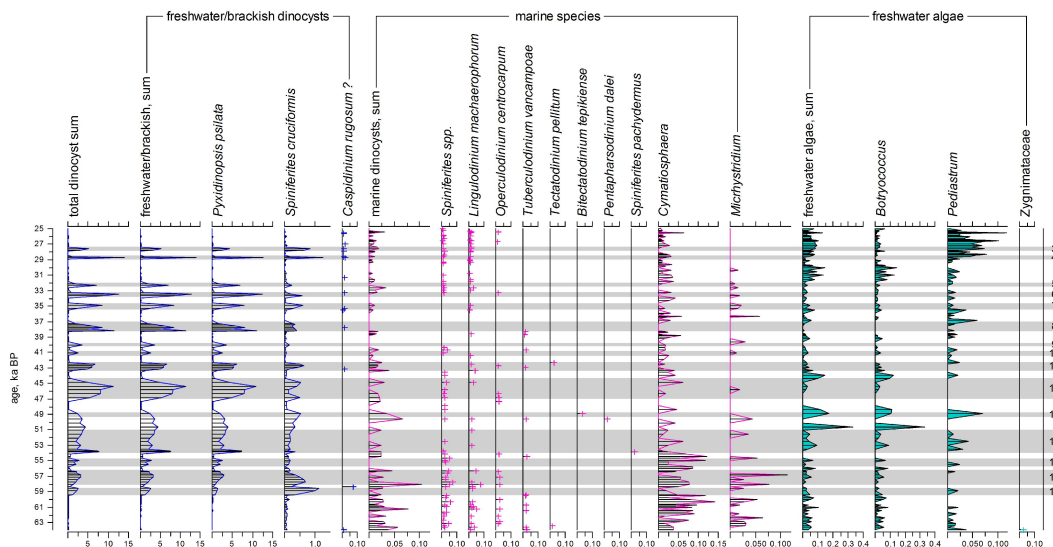
Fig. 4. Concentration diagram of dinoflagellate cysts, freshwater algae and acritarchs (specimens *cm^{-3} 10^{-3}) versus age. Horizontal grey bars with numbers indicate D–O events as identified by the 25-GC1 CaCO$_3$ (D–O 17-5) and dinocyst (D–O 4 and 3) records.

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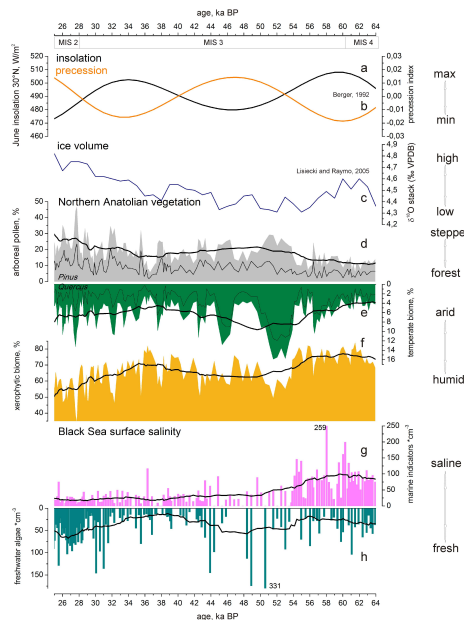
Fig. 5. Summary plot for Northern Anatolian vegetation and the Black Sea surface salinity in comparison to insolation and global ice volume changes: **(a)** June insolation at 30° N (W m^{-2}) (Berger, 1992); **(b)** precession index (Berger, 1992); **(c)** δ^{18}O stack (‰VPDB) (Lisiecki and Raymo, 2005) representing global ice volume; together with data from Black Sea core 25-GC1 with: **(d)** arboreal pollen record (%) with smoothed curve (thick line; simple moving 20 points average) and *Pinus* curve (thin line); **(e)** temperate biomes (%) with smoothed curve (thick line; simple moving 20 points average) and *Quercus* curve (thin line) (please note the reverse scale); **(f)** xerophytic biome (%) with smoothed curve (thick line); **(g)** concentration of marine dinocysts and acritarchs (speciments *cm^{-3}) with smoothed curve (thick line; simple moving 20 points average); **(h)** concentration of freshwater algae (speciments *cm^{-3}) with smoothed curve (thick line; simple moving 20 points average). Marine isotope stadia are indicated following Shackleton (1987).

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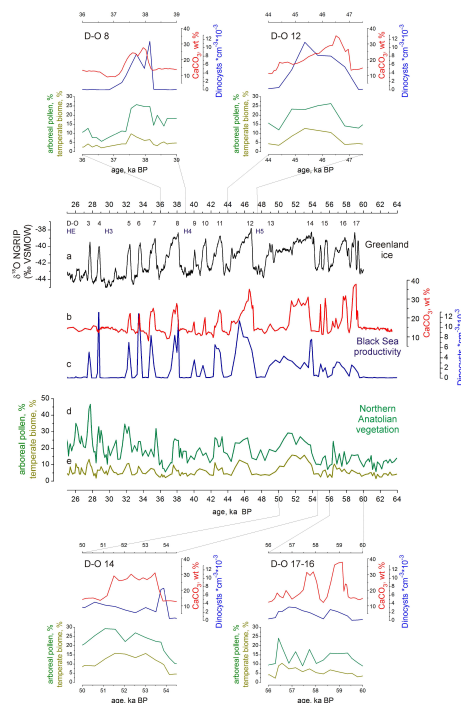
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Fig. 6. Northern Anatolian vegetation and the Black Sea productivity in response to D–O and HE climatic variability: **(a)** δ^{18}O record (‰ VSMOW) from NGRIP (GICC05) ice core (Svensson et al., 2008) with numbers of D–O events and HE marked above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