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Invited review

A 600,000 year long continental pollen record from Lake Van, eastern Anatolia (Turkey)

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ABSTRACT

Lake Van is the fourth largest terminal lake in the world (38.5°N, 43°E, volume 607 km^3, area 3570 km^2, maximum water depth 460 m), extending for 130 km WSW–ENE on the eastern Anatolian high plateau, Turkey. The sedimentary record of Lake Van, partly laminated, obtains a long and continuous continental sequence that covers multiple interglacial–glacial cycles. Promoted by the potential of the sedimentary sequence for reconstructing the paleoecological and paleoclimate development of the Near East, a deep drilling operation was carried out in 2010 supported by the International Continental Scientific Drilling Program (ICDP). The 219 m long continental pollen record presented here is based on a well-dated composite profile drilled on the so-called Ahlat Ridge in water depth of 360 m encompassing the last 600,000 years. It is the longest continuous continental pollen record of the Quaternary in the entire Near East and central Asia obtained to date. The glacial–interglacial cycles and pronounced interstadials are clearly reflected in the vegetation development based on millennial-scale time resolution. In general, the glacial/stadial vegetation is characterized by dwarf-shrub steppe and desert steppe, whereas the climax vegetation of past interglacials can be described as oak steppe-forest similar to the present interglacial in this sensitive semi-arid region between the Black, Caspian, and Mediterranean Seas. By comparing the Lake Van pollen record with other western Asian and southern European long continental pollen sequences as well as marine and ice-core records, the regional variability of the climate signals is also discussed.

1. Introduction

Long continental pollen records have fundamentally contributed to our understanding of millennial-scale paleoclimate variability on land in comparison to marine and ice-core records. In contrast to north-central Europe where glaciations have produced stratigraphic unconformities, complete records of terrestrial events over multiple glacial–interglacial cycles are documented in southern Europe in which a broad correspondence to the deep-sea oxygen isotope record has been noted (Tzedakis et al., 1997, 2001). Classical records are Bouchet/Praclaux in southern France encompassing the last four climatic cycles (Reille et al., 2000), Valle di Castiglione in Italy spanning the last 250 ka (Follieri et al., 1988), a 430 ka long record at Ioannina (Tzedakis, 1994a), and a 1.35 million year long record at Tenaghi Philippon, both in Greece (Wijmstra, 1969; Wijmstra and Smit, 1976; Tzedakis et al., 2006).

However, our knowledge about the vegetation and climate development during the past glacial–interglacial cycles in the Near East based on continental sedimentary sequences is rather poor. To date, the longest pollen record for the continental interior of the Near East has been described on two 100 m cores from Lake Urmia in north-western Iran spanning around 200 ka (Djamali et al., 2008). Lake Van, a terminal lake located on the high plateau of eastern Anatolia in Turkey, has therefore been chosen as an additional target to obtain a long continental record of multiple glacial–interglacial cycles.

First palynological studies from Lake Van were published by Van Zeist and Woldring (1978). However, uncertainties in the varve

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chronology established by Kempe and Degens (1978) did not allow a correlation with radiocarbon-dated Lateglacial/Postglacial pollen diagrams in the vicinity of Lake Van such as Lake Urmia and Lake Zeribar (Van Zeist and Bottema, 1977, 1991; Bottema, 1986). Further paleoecological investigations were made on annually laminated sediment cores drilled in 1990 (Wick et al., 2003). The pollen record, spanning about 13 ka documents evidence of Lateglacial and Holocene climatic changes as well as human impact in this semi-arid region of eastern Anatolia. As preparation for a deep drilling campaign under the umbrella of the International Continental Scientific Drilling Program (ICDP), a site survey was carried out in 2004 (Litt et al., 2009). Based on the seismic results, we cored different locations to water depths of up to 420 m. Multidisciplinary scientific work at position of a proposed ICDP drill site (Ahlat Ridge) included investigations on magnetic susceptibility, physical properties, stable isotopes, XRF scans, and pollen and spores (Litt et al., 2009). This core extends back to the Last Glacial Maximum (ca 20 ka). The results improved the potential of this locality for obtaining an almost continuous and undisturbed long continental paleoclimate record, which was finally drilled in 2010 (Litt et al., 2011, 2012). In this paper we present first palynological results of a 119 m long composite profile from Ahlat Ridge drilled in water depth of 375 m with millennial-scale time resolution.

2. Study area

Lake Van is situated on the eastern Anatolian high plateau (Turkey) close to the border to Iran (38.5°N, 43°E, Fig. 1). The 460 m deep lake lies within a tectonic depression with a maximum extension of 130 km ENE–WSW. The present lake level is at 1646 m above sea level (a.s.l.). With a surface area of about 3570 km^2 and a volume of ca 607 km^3, Lake Van is the fourth largest terminal lake and the largest soda lake on earth. High carbonate concentrations, active regional volcanism of Nemrut and Süphan (2948 m a.s.l. and 4058 m a.s.l. respectively) as well as subaqatic hydrothermal exhalations are responsible for the high alkalinity (pH 9.8, salinity 21.4‰) of the water (Kempe et al., 1991; Kaden et al., 2010). Therefore, Lake Van is located in a zone of complex tectonic movements, associated with the collision of Afro/Arabian plate from the south and the Eurasian plate from the north (Reilinger et al., 2006; Keskin, 2007).

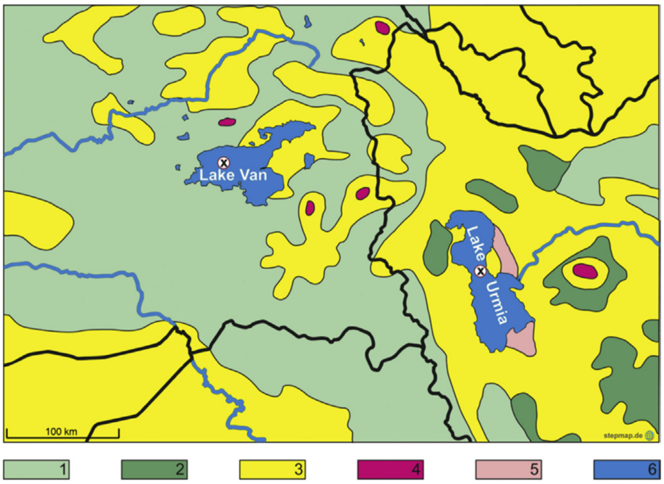

Fig. 1. Vegetation map of SE Anatolia and NW Iran at Lake Van and Lake Urmia (modified after Frey and Kürschner, 1989). 1 – Mixed formation of cold-deciduous broad-leaved montane woodland and xeromorphic dwarf-shrublands; 2 – Open tree and shrub vegetation; 3 – Dwarf-shrublands (steppe); 4 – Subalpine and alpine vegetation; 5 – Salt marshes; 6 – Lakes and rivers. Black lines mark state borders. The location of drill sites is marked in both lakes.

The regional environment at Lake Van is characterized by a continental climate. It is strongly influenced by the tracks of the westerly jet stream, the extension of the subtropical high pressure system, and the dry continental air masses of NE-Europe and Asia (Siberian high pressure system; Roberts and Wright, 1993; Litt et al., 2012). The local climate of Lake Van area shows a strong seasonality with cold winters (December to February) and warm summers (July to September). Average temperature ranges between 22 °C in July and below 0 °C in January (Table 1). During the summer period, sinking air masses of the subtropical high-pressure system control the climate of the region. The moisture from the Mediterranean Sea does not reach eastern Anatolian in summer, which is reflected in the lowest precipitation values during this season (Akçar and Schlüchter, 2005), whereas the highest precipitation falls during the winter and spring seasons (Sariş et al., 2010). This moisture originates from the westerly winds traveling east over the warm Mediterranean Sea (Roberts and Wright, 1993; Akçar and Schlüchter, 2005). A significant aspect in moisture transport is that the west–east oriented mountain ranges act as a precipitation barrier. Along the Bitlis Massif, the total amount of precipitation receives up to 1200 mm/year, whereas the precipitation in the northern region does not exceed 400 mm/year (Table 1). The diverse topography and the distribution of rainfall are further reflected in the vegetation.

According to the distribution of precipitation related to orography, the Lake Van region of south-eastern Anatolia is located in a transitional zone between two vegetation types (e.g. steppe forest and treeless steppe) of the Irano-Turanian plant territory after Zohary (1973). The so-called Kurdo-Zagrosian oak steppe forest belt (*Quercetea brantii*) extends from the Taurus mountains including the Bitlis complex, SW shore of Lake Van, to the southern part of the Zagros mountains (SW Iran). The forest-steppe consists of several deciduous oak species (*Quercus brantii*, *Q. infectoria*, *Q. ithaburensis*, *Q. libani*, *Q. robur*, *Q. petraea*, *Q. boissieri*, *Q. mannifera*) as well as *Pistacia atlantica*, *P. khinjuk*, *Acer monspessulanum*, *Juniperus oxycedrus*, *Pyrus syriaca*, *Crataegus* spp. and *Prunus*, *Amygdalus* spp. The upper tree line around 2500–2700 m is formed by *Betula verrucosa* (Zohary, 1973; Van Zeist and Bottema, 1991). This oak steppe forest has also been described as 'mixed formation of cold-deciduous broad-leaved montane woodland and xeromorphic dwarf-shrublands' by Frey and Kürschner (1989). Above the treeline whose limits vary depending on exposure, dwarf-shrub formations of the subalpine region occur. They are characterized by spiny, cushion-shaped *Astragalus*, *Gypsophila* species and other genera also appearing in a thorn-cushion form (Frey and Kürschner, 1989; Van Zeist and Bottema, 1991). The northern and northeastern parts of the lake are characterized by an almost treeless landscape, with semi-arid vegetation. The dwarf-shrub steppes (Frey and Kürschner, 1989) of the Irano-Turanian floral province in central and eastern Anatolia might be originally covered by herb-ridge *Stipa-Bromus* steppe, which has been replaced by *Artemisia* steppe ridge in thorn-cushions caused by intensive grazing (Van Zeist and Bottema, 1991). Today, the vegetation NE of Lake Van is dominated by an *Artemisia fragrans* steppe, different species of chenopods, and grasses and with some Sub-Euxinian oak forest elements (Frey and Kürschner, 1989; Van Zeist and Bottema, 1991). Remarkable is the high species number of different forb genera in the central and eastern Anatolian steppe vegetation such as *Achillea*, *Astracantha*, *Astragalus*, *Centaurea*, *Echinops*, *Thymus* and others (Kürschner et al., 1995).

3. Material and methods

3.1. Lithostratigraphy of the sediment cores

The Ahlat Ridge (AR) sediment cores of the 2010 drilling campaign were opened at the IODP core repository located at

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Table 1
Meteorological data at Lake Van (provided by the Turkish State Meteorological Service). Observation period: 1975–2008.

Station	Coordinates			Mean temperature [°C]			Mean precipitation [mm]		
	Latitude °N	Longitude °E	Altitude m a.s.l.	January	July	Year	January	July	Year
Erciş	39°20′	43°22′	1750	−6	21.8	7.7	31	7	421
Van	38°27′	43°19′	1661	−4	22.2	9	35	4	385
Bitlis	38°24′	42°06′	1551	−2	22	9.4	161	5	1232
Tatvan	38°30′	42°17′	1690	−3,2	21.9	8.7	95	7	816

MARUM, University of Bremen in spring 2011 by the Paleovan scientific team (Litt et al., 2012). After core description and correlation of the seven parallel cores (multiple coring) a composite profile recovering a sedimentary record of 219 mcblf (meter composite below sea floor; average discovery 91%) were constructed based on the sedimentary pattern of laminated intervals and volcaniclastic layers (Stockhecke et al., in press).

Present day sedimentation and the Holocene sequence is varved (Kempe and Degens, 1978; Landmann et al., 1996; Wick et al., 2003; Litt et al., 2009; Stockhecke et al., 2012). The AR composite profile consists of 76% lacustrine sediments (background sedimentation; Fig. 2), 2% fluvial deposits, 17% of volcaniclastic deposits (tephra) and 5% gaps (Stockhecke et al., in press). Within the background deposits eight major lacustrine sediment types were differentiated and separated from the event layers. Event deposits are graded beds (ca 300 layers) and volcaniclastic layers (about 300 layers; Stockhecke et al., in press). The volcaniclastic layers are reworked or fallout tephras from neighboring volcanoes at the northern lake shore (Sumita and Schmincke, 2013a,b,c). All event layers were subtracted from the 219-m-long composite AR record, which resulted in a 174-m-long event-corrected AR record ('metres composite below lake floor – no Events'; mcblf-nE, Stockhecke et al., in press). The lithological variability and sharp lithological boundaries documented by the lithostratigraphy of the AR record indicates that Lake Van recorded abrupt environmental and climate change in the past (Stockhecke et al., in press). Varved sediment layers reoccur at several sections throughout the AR record except the basal gravel units and diatomaceous muds (below ca 190 m). The diatomaceous muds were deposited in a freshwater environment, which lasted over about 150 ka after the initial lake flooding about 600 ka ago (Stockhecke et al., in press). The recurrence of laminated sediments, which were deposited during warm/wet climate periods and lake level rises in correspondence to interglacial/interstadial periods since Marine Isotope Stage (MIS) 11. Gray banded clayey silts oppose the finely laminated sections and were deposited during cold/dry climate conditions during lake level lowering in correspondence to glacial/stadial periods (Stockhecke et al., in press).

3.2. Chronology

The procedure how to compile a robust chronology for the 600,000 year old Lake Van record has been described and discussed in detail by Stockhecke et al. (in this volume). A brief summary is given in the following. The age model was constructed using climatostratigraphic alignment of proxy records, varve chronology, tephrostratigraphy, argon–argon single-crystal dating, radiocarbon dating, magnetostratigraphy, and cosmogenic nuclides. Reference curves for the climatostratigraphic alignment are the GICC05 (NGRIP members, 2004; Steffensen et al., 2008; Svensson et al., 2008; Wolff et al., 2010) from 0 to 116 ka BP, the speleothem-based synthetic Greenland record (GLT-syn; Barker et al., 2011) for the interval 116–400 ka BP and the same synthetic Greenland record but on the EDC timescale from Antarctica ice cores for 400 to 600 ka (GLT-syn; Barker et al., 2011; see also Fig. 3B). Eight geomagnetic tie points (from ~32 to ~250 ka), based on minima in the RPI record, and nine $^{40}Ar/^{39}Ar$ ages confirm the age model. However, uncertainties increase with depth as tectonic activity affected the drill site and the sedimentary regime was completely different during the early evolution of Lake Van (Stockhecke et al., in press). All control points to produce the age model for the Ahlat Ridge record we used to date the pollen spectra (Fig. 4) are listed in Stockhecke et al. (in this volume).

3.3. Pollen analysis

For the palynological analyses presented here, 188 samples (distance of 1.0 m each) have been selected for pollen analyses from the whole AR composite profile excluding coarse-grained tephra rich layers.

The preparation procedure of pollen samples of a specific sediment volume (4 cm^3) includes:

(1) treatment with 10% hot HCl to break down the sediment and to remove carbonates,
(2) treatment with 10% hot KOH to remove soluble humic acid,
(3) sieving to remove coarse detritus (mesh size: 200 μm),
(4) treatment with 39% cold HF to remove silicates (48 h),
(5) hot 10% HCl,
(6) glacial acetic acid,
(7) hydrolysis of cellulose with hot acetolysis mixture (9 parts acetic anhydrite and 1 part concentrated H_2SO_4),
(8) glacial acetic acid,
(9) ultrasonic sieving (mesh size: 10 μm) to concentrate the palynomorphs (Faegri and Iversen, 1989).

Lycopodium marker tablets were added to calculate the pollen concentrations per volume (cm^3) (Stockmarr, 1971). The residue was stained with safranin and mounted on slides in glycerol. For identification of pollen taxa, a pollen reference collection of Near Eastern plants (Steinmann Institute, Palaeontology, Bonn) as well as descriptions of the Mediterranean palynoflora were used (Reille, 1990, 1992, 1995, 1998; Van Zeist and Bottema, 1977; Chester and Raine, 2001). The number of pollen grains counted in each sample is at least 500 for the terrestrial pollen sum (joint analyses by N. Pickarski (0–60 m), T. Litt (61–155 m), G. Heumann (156–212 m)). Percent calculation, cluster analysis to define pollen zones, and printing of the diagrams (Figs. 2 and 4) was carried out using the TILIA computer programs including CONISS (Grimm, 1991-2011).

4. Results and discussion

4.1. Pollen zonation

The distribution of pollen in percent along the Ahlat Ridge record is illustrated in a simplified pollen diagram based on sediment depth of the composite profile including event layers (see Fig. 2). The pollen record can be subdivided into 11 biozones on the basis of changes in the trees and shrubs versus herbs ratio (arboreal pollen (AP) versus non-arboreal pollen (NAP)) and changes in the relative

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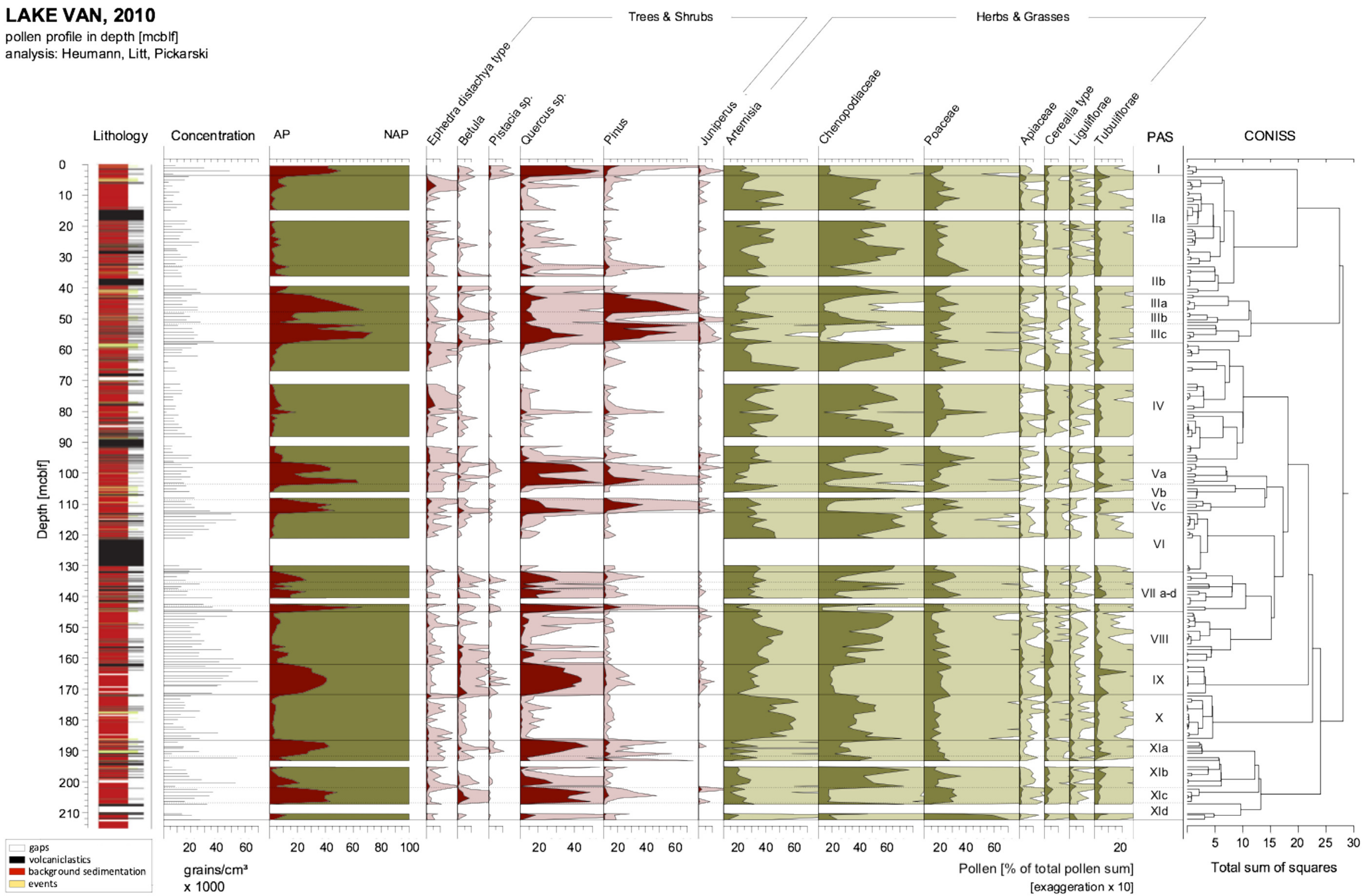

Fig. 2. Simplified percentage pollen diagram (right) related to the lithology (left) of the Lake Van sedimentary record at Ahlat Ridge presented on composite depth (meter composite depth below lake floor – mcblf, after Stockhecke et al., in press). AP – arboreal pollen; NAP – non-arboreal pollen; PAS – pollen assemblage superzones (based on cluster analysis), definitions see Subchapter 4.1.

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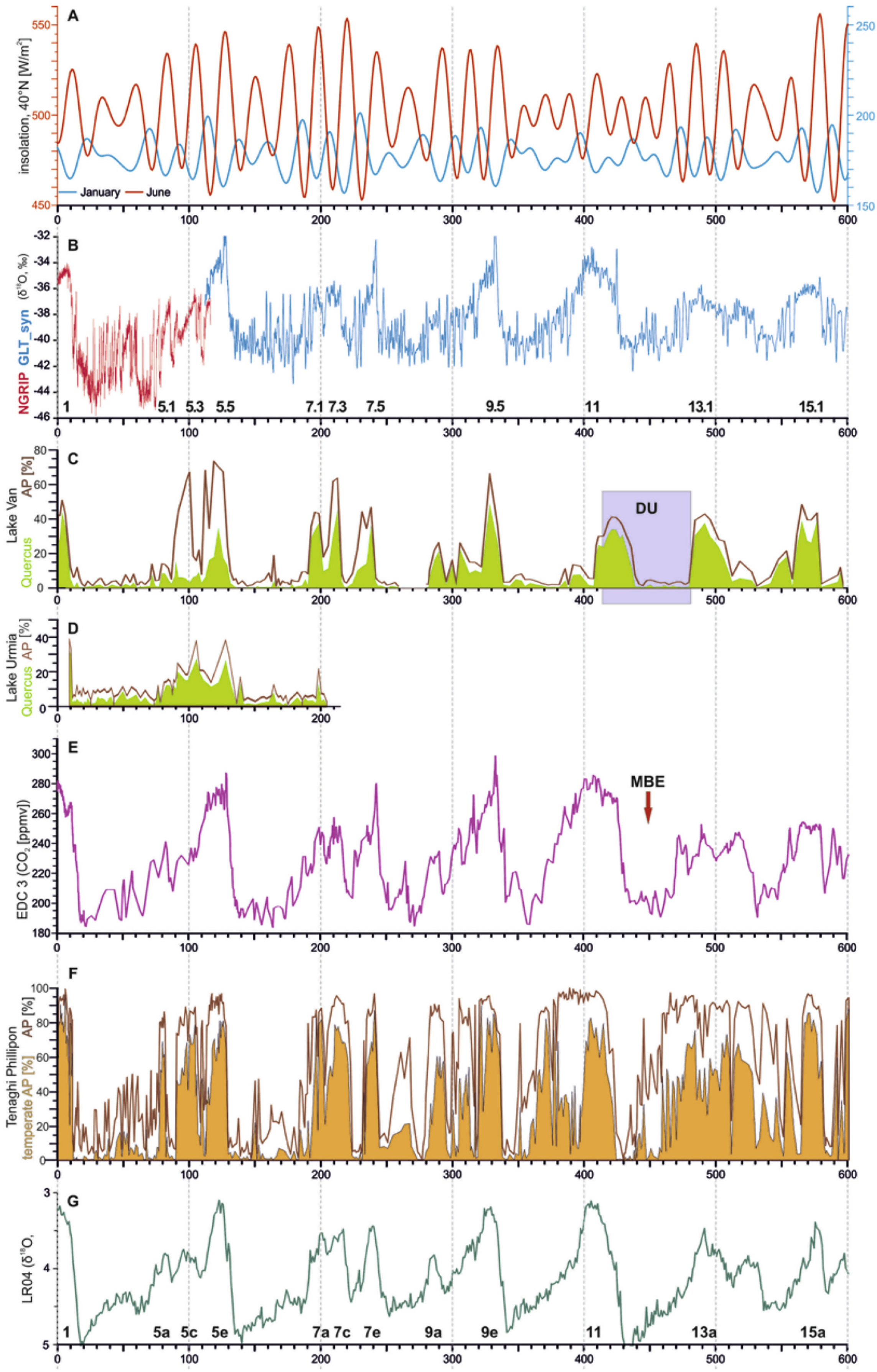

Fig. 3. A – mid-June and mid-January Insolation 40°N (Berger, 1978, data available as supporting material in: Berger et al., 2007); B – Greenland ice-core record NGRIP (Wolff et al., 2010) and synthetic ice-core record GLT_syn (Barker et al., 2011); C – Lake Van pollen record (total AP and *Quercus*), the Deformed Unit (DU) is indicated in gray (no time control); D – Lake Urmia pollen record (Total AP and *Quercus*, Djamali et al., 2008; synchronized with B and C); E – Atmospheric CO_2 concentration from Antarctic ice cores (EIPCA Dome Concordia – EDC, Jouzel et al., 2007), mid-Brunhes event MBE; F – Tenaghi Philippon pollen record (total AP and sum of thermophilous trees, Tzedakis et al., 2006); G – LR04 stack of benthic oxygen isotope records (Lisiecki and Raymo, 2005).

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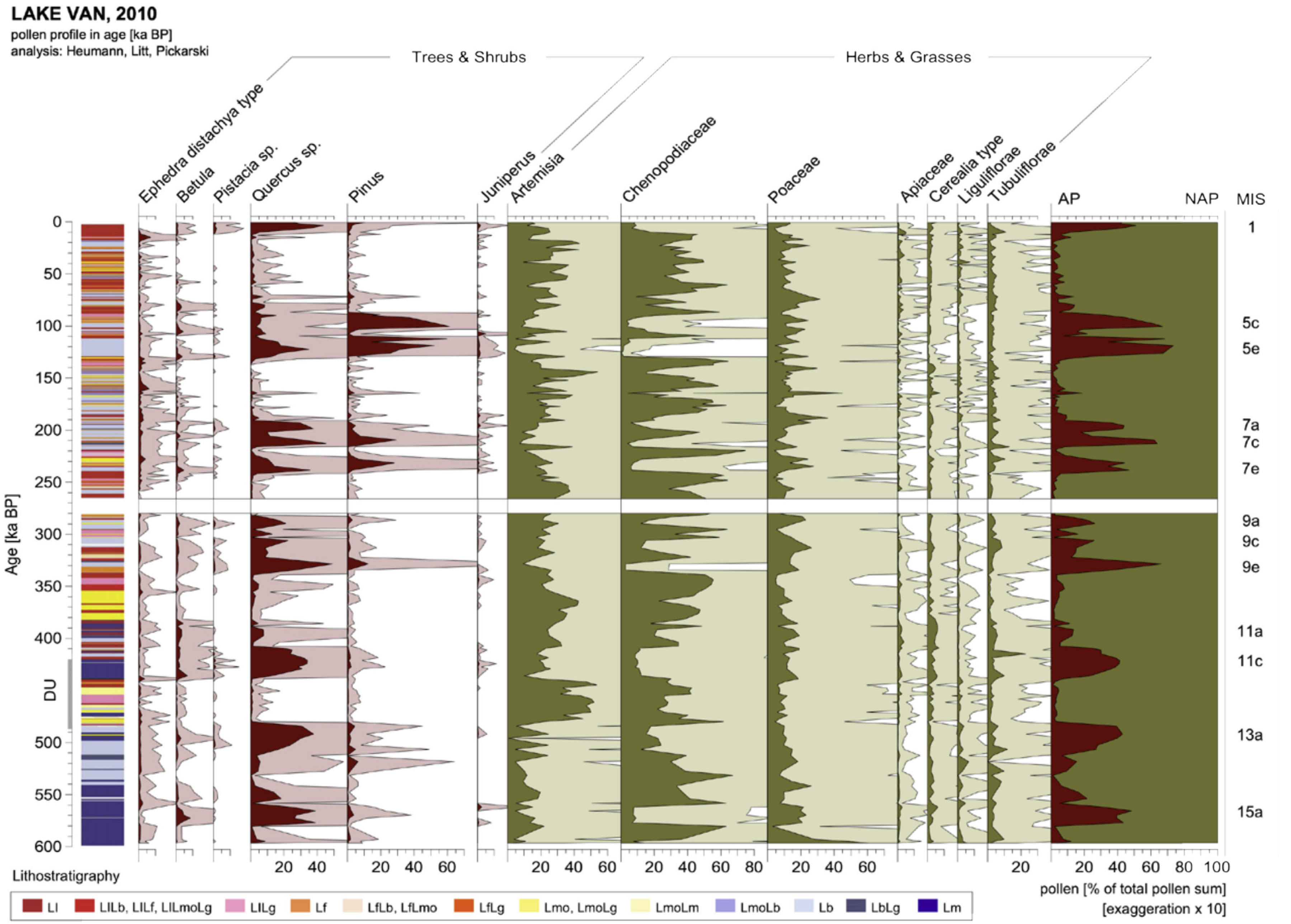

Fig. 4. Simplified pollen diagram of the Lake Van (Ahlat Ridge) record plotted against time scale following the age-depth model performed by Stockhecke et al. (in this volume) and related to detailed lithostratigraphy after Stockhecke et al. (in press). Ll – laminated clayey silt; Lf – faintly laminated clayey silt; Lmo – mottled clayey silt; Lb – banded clayey silt; Lm – massive clayey silt; Lg – graded bed. DU – interval of the deformed unit (no time control); AP – arboreal pollen; NAP – non-arboreal pollen; MIS – marine isotope stages.

Table 2
Pollen zonation of the Lake Van-Ahlat Ridge composite record.

Pollen Assemblage Superzones (PAS)	Sediment depth mcblf	Features
Quercus–*Pistacia* PAS (I)	3.49	AP: predominance of *Quercus*, *Pistacia* frequent NAP: remarkable amounts of *Artemisia* and Poaceae Lower boundary: increase of *Quercus* (>5%)
Artemisia–Poaceae–Chenopodiaceae PAS (II)	32.63	IIa AP: *Ephedra* frequent NAP: predominance of steppic elements
	41.85	IIb AP: remarkable amounts of *Quercus* (5–10%) NAP: predominance of steppic elements Lower boundary: rapid decrease of AP, mainly *Pinus*
Pinus–*Quercus* PAS (III)	47.71	IIIa AP: predominance of *Pinus*
	51.61	IIIb predominance of NAP such as *Artemisia* and chenopods
	57.79	IIIc AP: predominance of *Pinus* and *Quercus* Lower boundary: increase of *Quercus* (>5%)
Ephedra–*Artemisia* PAZ (IV)	96.64	AP: remarkable amounts of *Ephedra* NAP: predominance of steppic elements with changing maxima between *Artemisia*, chenopods and grasses Lower boundary: rapid decrease of AP, mainly *Quercus*
Quercus–*Pinus* PAS (V)	103.55	Va AP: double peak, lower part predominance of *Quercus* and *Pinus*, upper part predominance of *Quercus*
	108.60	Vb Predominance of NAP such as *Artemisia* and chenopods
	112.65	Vc AP: predominance of *Pinus* and *Quercus* Lower boundary: increase of *Quercus* (>5%)
Chenopodiaceae–*Artemisia* PAS (VI)	132	AP: some *Ephedra* NAP: predominance of steppic elements Lower boundary: decrease of AP, mainly *Quercus*
Quercus–Chenopodiaceae PAS (VII)	144.90	AP: predominance of *Quercus* with remarkable fluctuations (VIIa–d), maximum values in the lower part (VIId) Lower boundary: increase of *Quercus* (>5%)
Artemisia–Chenopodiaceae PAS (VIII)	161.90	AP: some *Ephedra* NAP: predominance of steppic elements with the general tendency of increasing Chenopodiaceae values Lower boundary: rapid decrease of *Quercus*
Quercus PAS (IX)	171.70	AP: predominance of *Quercus* Lower boundary: increase of *Quercus* (>5%)
Artemisia PAS (X)	186.56	NAP: predominance of *Artemisia* and further steppic elements such as chenopods Lower boundary: rapid decrease of *Quercus*
Quercus–Chenopodiaceae PAS (XI)	191.54	XIa AP: predominance of *Quercus*
	201.90	XIb NAP: predominance of Chenopodiaceae
	206.80	XIc AP: predominance of *Quercus*
	212.20	XId AP: some *Quercus* NAP: predominance of chenopods and grasses Lower boundary: not defined

frequency of individual taxa. The features of biozones as well as criteria for defining the lower boundaries are given in Table 2. Regarding the hierarchical biostratigraphical classification of long continental pollen records we follow the concept of pollen assemblage superzones (PAS) introduced by Tzedakis (1994b). Superzones can be defined such that they correspond to chronostratigraphic stages or substages, which would be in agreement with the zonation of the Lake Van pollen diagram presented here. The advantage of such procedure is that, based on a higher sample resolution in the future, there will be the potential to identify different successions of assemblages within a superzone each in theory constituting a pollen assemblage zone (PAZ). The hierarchical approach relying on pollen assemblage superzones (as defined in this paper), pollen assemblage zones and subzones (which will be defined in the future) is in agreement with the International Stratigraphic Guide (Salvador, 1994). The boundary definitions of the PAS are supported by cluster analysis that uses the main AP and NAP components (Fig. 2).

4.2. Biostratigraphy and boundary definition

It has long been appreciated that there is a broad correspondence between warm climatic intervals, respectively periods of low ice volume as defined by Marine Isotope Stages (MIS) and forest intervals on land as demonstrated in southern Europe based on long pollen records (compiled by Tzedakis et al., 1997, 2001). However, that boundaries defined by using different proxies do not necessarily occur at the same time has been demonstrated by investigation of marine sediment cores west of Portugal by Shackleton et al. (2002, 2003). Here the MIS 6/5e boundary has

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Table 3
Comparison between MIS terminations, forest expansions, insolation, and climate events in NGRIP/GLT_syn.

MIS	MIS lower boundaries (in ka BP)	Forest expansion (in ka BP)	Mid-June insolation max. 40°N in ka BP (Berger, 1978)	Events NGRIP/GLT_syn in ka BP
1	13 (TI)	11.6 (central Europe: Litt et al., 2001) 8 (Lake Van: Wick et al., 2003)	11	YD/Holocene: 11.7 ref. to 2000 (Walker et al., 2009)
5e	132 (TII)	126 (Iberian margin, Sanchez-Goñi et al., 1999) 127.2 (Monticchio, Brauer et al., 2007)	127	128.4 (peak 5.5) (Barker et al., 2011)
7a	200	198.5 (Iberian margin, Roucoux et al., 2007)	198	198.4 (peak 7.1) (Barker et al., 2011)
7c	216.8 (TIIIa)	215.3 (Iberian margin, Roucoux et al., 2007)	220	216 (max. peak 7.3) 220.1 (first peak 7.3) (Barker et al., 2011)
7e	246 (TIII)	243.2 (Iberian margin, Roucoux et al., 2007)	243	241.8 (peak 7.5) (Barker et al., 2011)
9e	338 TIV)	336.6 (Iberian margin, Roucoux et al., 2007)	334	334.2 (peak 9.5) (Barker et al., 2011)

been significantly earlier (ca 5000 years) than the glacial–interglacial stage boundary on land identified by pollen analysis of the same marine record. Comparable observations have been described by Tzedakis et al. (2004) and Roucoux et al. (2007) for MIS 8/7 and MIS 10/9 boundaries (Table 3). To avoid this problem of significant differences in the phasing of marine and terrestrial stage boundaries, Tzedakis (2005) tried to explore directly the relation between vegetation phases and orbital configurations. He hypothesized that the timing of expansion of Mediterranean sclerophylls (*Pistacia*, *Olea*, *Phillyrea*) during interglacials follows closely the peaks of the June insolation curve rather than having a fixed time lag relative to the mid-point of the deglaciation.

However, in the continental, semi-arid Lake Van region, it is difficult to use only the expansion of woodland as criterion for the lower boundary of a warm stage. As already shown in high-resolution pollen studies by Wick et al. (2003), and confirmed by Litt et al. (2009), a delay of ca 3000 years in the expansion of deciduous oak woodlands refer to the Pleistocene/Holocene boundary as defined in the Greenland ice core from NorthGRIP (NGRIP) stratotype (Walker et al., 2009) can be recognized. Therefore, a multi-proxy approach would be more suitable for boundary definitions as shown already in Litt et al. (2009) and Wick et al. (2003).

Ice cores from Greenland are unprecedented reference profiles for climate-stratigraphy in the Northern Hemisphere not only for the Pleistocene–Holocene transition. In addition, they document abrupt climate oscillations below orbital cycles known as Dansgaard–Oeschger oscillations (DO) or Greenland Interstadials (GI) over the last glacial stage (Wolff et al., 2010). They are also reflected in some high-resolution pollen sequences (Fletcher et al., 2010; Müller et al., 2011). Multi-proxy studies based on annually laminated sediments from Lake Van cores already obtained in 1990 and 2004 have demonstrated that climate events such as the Greenland Interstadial 1 (Bölling/Alleröd) or the Greenland Stadial 1 (Younger Dryas) can clearly be identified by abiotic and biotic proxies such as oxygen isotopes, Mg/Ca ratios, X-ray Fluorescence (XRF) measurements as well as pollen data (Wick et al., 2003; Litt et al., 2009). Even the relative duration of the cold interval interpreted as Younger Dryas in Lake Van is very similar to ice-laminae counting of the NGRIP core (Rasmussen et al., 2006) or varve-counted central European lacustrine deposits (Litt et al., 2001, 2003). Therefore it is legitimate to use the Greenland ice-core record as reference stratotype also for correlation and synchronization of the older climate events identified by multi-proxy analyses of the Lake Van record.

As far as the past interglacials are concerned, the peaks of the synthetic Greenland record just after each Termination have therefore been used as correlation points for forest expansion as documented in the Lake Van pollen sequence (Fig. 3). For the last interglacial, the peak of MIS 5.5 in GLT_syn is at 128.4 ka, which is close to the onset of oak expansion in the varve-counted record of Monticchio in Italy (127 ka, Brauer et al., 2007) as well as to the mid-June insolation maximum at 40°N (127 ka, Berger, 1978; see also Table 3).

4.3. Palynological results and discussion

4.3.1. Holocene

The *Quercus*–*Pistacia* PAS (I) marks the recent interglacial stage (Figs. 2 and 4), which is characterized by an oak steppe-forest (*Q. brantii* with *P. atlantica*, *J. oxycedrus*). Caused by the lower resolution of the pollen diagram in this study we refer to previous more detailed investigations of the Lateglacial and Holocene natural as well as anthropogenically influenced vegetation history (Wick et al., 2003; Litt et al., 2009).

4.3.2. Last interglacial–glacial cycle

The *Pinus*–*Quercus* PAS (IIIa–c) includes the last interglacial and the older early glacial interstadial interrupted by a stadial. The climate-stratigraphic terms 'interglacial' and 'interstadial' were first defined by Jessen and Milthers (1928) for periods with characteristic records of non-glacial climate, as indicated by paleobotanical evidence for major vegetation changes. Following these suggestions, interglacials are classified as periods with a climate optimum (climax vegetation) at least as strong as the present interglacial (Holocene) in the same region. Interstadials are assumed to have been either too short or too cold to reach the climate level of interglacial type in the same region. This definition is still valid in Quaternary climate stratigraphy (Gibbard and van Kolfschoten, 2004) and can also be applied for the Lake Van region in the following interpretation.

PAS IIIc marks clearly the last interglacial stage, the Eemian, correlative mainly to MIS 5e, which is characterized by the highest AP values of the whole cycle. The succession in the transition zone between penultimate glacial and last interglacial (e.g., Termination II) is similar to that of Termination I: higher *Ephedra* values are followed by *Artemisia* peak and increase of *Betula* as pioneer tree. The expansion of an open mixed-oak forest around or just after 128 ka (peak MIS 5.5; Stockhecke et al., in this volume) in the mountain range of the Bitlis Massif seems to be comparable to the development during the Postglacial (Kurdo-Zagrosian oak steppe-forest after Zohary, 1973). However, there is no independent proof of a possible millennial-scale delay between forest expansion and interglacial onset, as documented for the early Holocene.

In contrast to the Holocene is the relatively high amount of pine of the woodland component during the last interglacial, which does not grow naturally in the Lake Van region today. Pine (e.g. *Pinus nigra*) is a main arboreal component of the so-called Xero-Euxinian steppe-forest after Zohary (1973), which recently occurs in more continental western and central Anatolia, in the rain shadow of the coastal mountain range including south of the Pontic mountain range (Van Zeist and Bottema, 1991). Two other pine species occur

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in Turkey today: *P. brutia* and *P. pinea* (Kürschner et al., 1995) and their distribution is mainly related to Mediterranean climate. Compared to the present spatial distribution of *Pinus nigra*, it seems that the boundary in Anatolia during the Eemian reached much further to the East. It is interesting to note that pine did not occur at Lake Urmia, SE of Lake Van (Fig. 1), during the last interglacial where an *Artemisia* steppe-forest dominated by *Quercus* and *Juniperus* marks this thermophilous interval (so-called Sahand interglacial after Djamali et al., 2008).

The expansion and predominance of pine might be a hint at an even higher continentality during the last interglacial in the Lake Van region compared to the Holocene. It might be a result of insolation during MIS 5e as discussed in Berger et al. (2007). The pattern of latitudinal distribution over time shows positive deviation of the mid-month insolation at June from 135 ka to 120 ka for 40°N with a maximum around 127 ka and a negative deviation at January from the present-day values (see also Fig. 3). By using a coupled ocean–atmosphere model, Kaspar and Cubasch (2007) performed climate simulations for the Eemian as equilibrium experiments with orbital parameters and greenhouse gas concentrations set to 125 ka. The mean simulated difference (Eemian – present day) in northern summer (June, July, August) shows positive near-surface temperature anomalies, however in northern winter (December, January, February) negative temperature anomalies for the Anatolian region (40°N). The latter contrasts to the north-central European region at 60°N, where the simulated difference Eemian – present day in northern winter shows positive surface-temperature anomalies which can be confirmed by an independent model-data comparison with climate reconstructions based on paleobotanical data (Kaspar et al., 2005). It is interesting to note that neither the simulations for precipitation in summer nor for precipitation in winter at 125 ka performed by Kaspar and Cubasch (2007) show deviations compared to present-day in the Anatolian region. Therefore a higher continentality index during the last interglacial compared to today might be an explanation for the role of pine (probably *P. nigra*) as strong competitor to oak in the woodland components. The latter one may occur in meso-climatically more favorable higher altitudes caused by increasing, orography related precipitation values.

PAS IIIb can be interpreted as an almost treeless stadial (MIS 5d) with high values of steppe elements in the pollen flora followed by a pronounced warm phase (PAS IIIa), which is characterized by similar high pine percentages, however, much lower oak values as in the previous interglacial. Based on the definition of interglacials and interstadials described above, PAS IIIa is clearly an interstadial stage caused by the reduced amplitude of thermophilous tree components (e.g. deciduous oak) and can be correlated with the Brörup in north-central Europe, more or less equivalent to MIS 5c (onset GI24 at 108 ka, Stockhecke et al., in this volume). The mid-month insolation maximum of June near the onset of the interstadial is similar high as that of the previous interglacial, which could, again, explain the predominant role of pine.

The *Artemisia*–Poaceae–Chenopodiaceae PAS (IIa, b) documents the continuation of the last glacial vegetation and climate development until Termination I. IIb can be correlated with the second Early Weichselian stadial (MIS 5b) and interstadial (MIS 5a or Odderade Interstadial in NE Europe, onset GI21 at 85 ka, Stockhecke et al., in this volume). The second Early Weichselian interstadial is not as pronounced as the first one in PAS IIIa caused by the absence of pine. It should be mentioned that an intercalated ca 2-m-thick tephra layer (V-60, Stockhecke et al., in press; Incekaya-Dibeli Tephra, Sumita and Schmincke, 2013c) in this interstadial might have lead to an incomplete documentation of the vegetation signal. In southern European pollen records (e.g. Tenaghi Philippon) the interstadial 5a is usually not subdued in amplitude. However, reduced AP values in the Lake Van record in comparison with the older interstadial in MIS 5c correspond to the pattern described in the Lake Urmia region (Kaboudan Interstadials I and II, see Djamali et al., 2008; see Fig. 3).

PAS IIa is characterized by steppe and desert steppe vegetation with *Artemisia*, chenopods, grasses and forbs. In general, dryness in the high plateau seems to be the limiting factor for tree growth during the Pleniglacial. For the mountain ranges of the Bitlis Massif, it would have been too cold for trees, and it might be covered by plant communities dominated by thorn-cushions as reconstructed also for the north-western Zagros region (Van Zeist and Bottema, 1991). For higher elevations there is evidence that Bitlis Massif had been covered by glaciers during the last ice age (Akçar and Schlüchter, 2005).

The sample resolution at this stage of investigation does not allow identifying rapid vegetation dynamics correlative to DO events, which are already documented in the higher resolved total organic carbon TOC record reflecting aquatic productivity and preservation controlled by lake-level variations (Stockhecke et al., in press). Ongoing pollen analyses based on a higher time resolution for the last interglacial–glacial cycle will contribute to this aspect in the near future (Pickarski, Doctoral Thesis in prep.).

4.3.3. *Penultimate interglacial–glacial cycle*

The *Quercus*–*Pinus* PAS (Va–c) can be described as an interglacial complex with three remarkable AP peaks. This general pattern of triplicate warm phases interrupted by two stadials is characteristic both in marine and ice-core records (MIS 7a, 7c, 7e, see Fig. 3 and Table 3), as well as for continental pollen sequences in southern Europe correlated and synchronized by Tzedakis et al. (2001). The pollen record of Lake Urmia does not encompass the complete penultimate interglacial stage (Djamali et al., 2008). Only the younger part of MIS 7 with a modest expansion of steppe-forest is documented (Fig. 3).

Recent high-resolution pollen analyses at Ioannina in Greece by Roucoux et al. (2008) suggest that during all of these three warm intervals of MIS 7 winter temperatures seem to be lower than during the Holocene and the last interglacial, as indicated by smaller populations of sclerophyllous taxa. A similar tendency of reduced thermophilous components in pollen spectra has already been discussed for the Velay region (Massif Central, France) by Reille et al. (2000), where the warm phases Bouchet 1, 2, and 3 as equivalents to MIS 7e, c, and a (see Tzedakis et al., 1997) are described as interstadials rather than interglacials.

The observation of cooler MIS 7 interglacials made in southern Europe is at variance with the Lake Van pollen record where the vegetation development during all three warm intervals reach the level of the last interglacial and the Holocene. The oldest and middle warm phases are characterized by an oak–pine forest-steppe, whereas in the youngest one only *Quercus* predominates among AP percentages but also reaches maximum Holocene values. However, regarding the relative amplitude of the warm intervals within MIS 7, the general pattern shows similarities to the southern European records as discussed in Roucoux et al. (2008) and Tzedakis (2005). In the ice-core records of Antarctica, MIS 7e (MIS 7.5) is characterized as warm interglacial and MIS 7c, 7a (MIS 7.3, 7.1) are rather cooler interstadials (Jouzel et al., 2007, see also Fig. 3). The largest forest expansion in the Lake Van region during the penultimate interglacial complex, however, occurred during the terrestrial equivalent of MIS 7c, which corresponds to the occurrence of the highest forest tree diversity in southern European pollen diagrams at the same time (Follieri et al., 1988; Tzedakis et al., 2003; Roucoux et al., 2008). Tzedakis (2005, p. 1590) suggested that "during temperate intervals the extent of forest development in this geographical region may be more closely related to the amplitude of insolation and associated climate regimes (e.g. high temperatures and winter precipitation), but may diverge from the extent of residual ice volume". At Lake Van, the vegetation seems to respond even stronger to changes in insolation. The vegetation development especially of oak mimics not only the high-amplitude changes in June insolation during MIS 7d and 7c, but also that during MIS 7b and 7a, which differs slightly from southern European records where no real stadial phase between MIS 7c and 7a can be observed (Tzedakis, 2005; see Fig. 3; Roucoux et al., 2008).

In the following penultimate glacial (*Ephedra–Artemisia* PAS (IV)), the interglacial oak forest-steppe was replaced by an *Artemisia* dwarf-shrub steppe and desert steppe with grasses, chenopods and *Ephedra*. The vegetation type is very similar to that of the last glacial. Based on a pollen record from the Portuguese margin (Margari et al., 2010) and confirmed by a high-resolution terrestrial pollen record at Ioannina, NW Greece (Roucoux et al., 2011) the vegetation history of MIS 6 in both regions can be divided in an early period (MIS 6e) with pronounced AP oscillations and a later period with subdued oscillations. Even if the Lake Van pollen record has not the same time resolution, a similar pattern can be observed based on fluctuations of birch, oak and pine curves.

4.3.4. Interglacial–glacial cycle-3

The *Quercus–Poaceae* PAS (VIIa–d) marks the next older interglacial–interstadial complex correlative to MIS 9. VIId is characterized by a remarkable oak peak (near or just after 9.3 peak at 334 ka, Stockhecke et al., in this volume). The AP values are comparable to those of the Holocene. Pine does not play any role in the woodland components. The percentages around 10% are rather low, similar to the Postglacial, and can be interpreted as long-distance transport. The climate optimum of the interglacial with open mixed-oak forest (as equivalent of MIS 9e or Litochoris interglacial in Tenaghi Philippon after Wijmstra, 1969) is followed by two interstadials with less pronounced *Quercus* peaks up to 20% in PAS VIIa and VIIc which is similar to the general pattern documented in the marine as well as southern European pollen records (MIS 9a and 9c; Krimenes and Kavalla interstadials in Tenaghi Philippon after Wijmstra, 1969).

The interstadial oak-steppe vegetation was replaced by a steppe or desert steppe in the following Chenopodiaceae–*Artemisia* PAS (VI), which marks the cold stage (MIS 8) prior to the penultimate interglacial complex. Within that glacial, a 10-m-thick volcanoclastic layer in the AR record caused a ca 15 ka long gap due to poor recovery of the adjacent lacustrine sediments (V-206, Stockhecke et al., in press; in this volume).

4.3.5. Interglacial–glacial cycle-4

The *Quercus* PAS (IX) is clearly related to interglacial deposits correlative to MIS 11c. Following sedimentological studies by Stockhecke et al. (in press), the lower part of the interglacial belongs to a 21-m-thick deformed unit of overturned and overthrusted sediment packages overlain by a 6-m-thick megaturbidite of reworked lacustrine sediment due to a seismically triggered megaevent, which occurred ~413 ka (Stockhecke et al., in press). To minimize the problems related to the disturbed unit, we sampled only parts of this critical interval, which show a clear lamination or stratification. Slumps or homogenites have been excluded from the analysis. Even if the internal structure of the unit is disturbed and can thus not be interpreted in detail, the general picture of the pollen data (predominance of *Quercus* as thermophilous woodland component) is comparable to previous interglacial stages. Generally, the Termination V shows a characteristic succession documented also in Terminations I and II: an *Ephedra* peak is followed by *Betula* before the remarkable *Quercus* increase can be observed. However, the *Quercus* increase corresponds to the reworked megaturbidite that the succession cannot be dated and limits further interpretation. The final phase of the *Quercus* PAS (IX) belongs to an undisturbed sedimentary unit after Stockhecke et al. (in press), which is characterized by a last peak of AP values (more or less correlative to 11.3 peak at 407 ka, Stockhecke et al., in this volume) followed by decreased AP values.

In general, the MIS 11, which is considered as a potential analogue for the Holocene and its future based on its insolation (Loutre and Berger, 2003; Tzedakis, 2010), is an exceptional interglacial complex because in the deep-sea isotope record of North Atlantic the duration with ca 30,000 years is much longer than the succeeding interglacials (e.g. ODP 980, McManus et al., 1999; see also Tzedakis et al., 2012). In addition, weak changes in insolation are associated with substantial climate changes and greenhouse gas variations showing a prolonged duration of high CO_2 concentrations during MIS 11 in comparison to the younger interglacial stages (Jouzel et al., 2007; Tzedakis et al., 2009; see Fig. 3). Direct correlation between marine (benthic/planktonic foraminifera, isotopes) and terrestrial signals (pollen) in NW Iberian margin deep-sea cores also show that this interglacial was more or less twice as long as the Last Interglacial in that region (Vigo interglacial after Desprat et al., 2007; see also Tzedakis, 2010).

Due to the sediment deformation that only allows to date the lower and upper boundary (Stockhecke et al., in this volume), we can only hypothesize that the *Quercus* PAZ was also prolonged in the Lake Van region.

In the following *Artemisia*–Chenopodiaceae PAS (VIII), the forest-steppe is replaced by steppe and desert-steppe as indicated by the pollen spectra. The lowermost *Quercus* peak (10%) might be related to an interstadial phase correlative to MIS 11a, whereas the following part belongs mainly to MIS 10.

4.3.6. Interglacial–glacial cycles-5 and -6

Artemisia PAS (X) characterizes a cold stage with typical assemblages of steppe and desert-steppe elements. This interval belongs also to the deformed unit and the original deposition before the secondary re-deposition took place during MIS 12 (Stockhecke et al., in press). Thus, the pollen data confirms the results based on the lithostratigraphic framework of the Lake Van sediments.

The lowermost *Quercus*–Chenopodiaceae (PAS XIa–d) encompasses probably MIS 13–15. XIc is a full interglacial stage as indicated by the predominance of oak mainly correlative to MIS 15a (peak MIS 15.1 at 579 ka, Stockhecke et al., in this volume), whereas XIa supposes to be a younger interglacial correlative to MIS 13a (peak MIS 13.1 at 488 ka, Stockhecke et al., in this volume). Both warm intervals are interrupted by a glacial stage as shown by the steppe and desert steppe pollen assemblages in XIb. The formation of the lacustrine deposits at Ahlat Ridge started around 600 ka during the middle part of the interglacial complex of MIS 15 (Stockhecke et al., in press).

Overall, the Lake Van pollen record is in line with results from Tenaghi Philippon (Tzedakis et al., 2006) that both the AP maxima and the predominance of thermophilous trees are similar among the interglacial stages before and after the mid-Brunhes-event (MBE; ca 430 ka; Fig. 3). As such the AP values of the AR record of thermophilous elements are comparable for the warm periods before 430 ka with those of the younger interglacial stages including the Holocene. This is in contrast to the differences in interglacial intensities before and after the MBE documented in marine and ice-core records (Jouzel et al., 2007; Tzedakis et al., 2009). Therefore, it seems that the high mid-June insolation maxima and the continental setting of Lake Van outweighed the effects of reduced CO_2 concentrations and/or residual ice volume in

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determining the strength of vegetational development, as already discussed for MIS 7.

5. Conclusions

Lacustrine sediments of Lake Van drilled in 2010 yield the longest continental pollen record in the entire Near East and central Asia obtained to date encompassing the last 600 ka. The multi-millennial-scale resolved (average ca 3 ka, see Fig. 4) Lake Van pollen record documents the glacial and interglacial stages as well as the most pronounced interstadials clearly as increase of thermophilous oak and/or additional arboreal pollen types.

In general, the glacial/stadial vegetation in the Lake Van region during the last 600 ka can be described as dwarf-shrub steppe and desert steppe with *Ephedra*, *Artemisia*, chenopods, grasses and forbs very similar to each other.

The climax vegetation of each interglacial stage (apart from the Holocene, MIS 5e, 7a, 7c, 7e, 9e, 11c, 13a, and 15a) in the Lake Van region is characterized by an oak steppe-forest with pistachio and juniper. The diversity of tree genera among the main woodland components in the pollen diagram seems to be rather low compared to southern Europe interglacials and their forest successions. However, it must be stressed that a validation of this pattern is only possible based on high-resolution pollen analyses in the future.

Pinus as additional arboreal component resistant to cold and continental conditions played a major role only during the last (MIS 5e) and the penultimate interglacials (MIS 7c and MIS 7e). Even if pine values are also high during the MIS 5c, this warm interval has to be classified as interstadial caused by much lower values of thermophilous, deciduous oaks compared to full interglacials.

In contrast to the atmospheric CO_2 concentration from Antarctic ice cores or marine isotope values based on benthic foraminifera, there is no clear subdivision in the Lake Van pollen record between low-amplitude interglacials (cooler cycles) prior the mid-Brunhes event (MBE) at 430 ka and high-amplitude, post MBE interglacials (oak values as reference). Lower CO_2 concentrations in the atmosphere might be compensated by stronger insolation forcing during MIS 13a and 15a (Fig. 3). A similar pattern can be observed during the triplicate interglacial complex MIS 7 where AP and oak values reach maximum values during 7c instead of 7e and match the results of pollen records from Greece. This underlines the different environmental correspondence to global climate change in the continental interior of the Near East compared to the global ice volume and/or greenhouse gas.

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