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The ELSA-Vegetation-Stack: Reconstruction of Landscape Evolution Zones (LEZ) from laminated Eifel maar sediments of the last 60,000 years

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A B S T R A C T

Laminated sediment records from several maar lakes and dry maar lakes of the Eifel (Germany) reveal the history of climate, weather, environment, vegetation, and land use in central Europe during the last 60,000 years. The time series of the last 30,000 years is based on a continuous varve counted chronology, the MIS3 section is tuned to the Greenland ice — both with independent age control from ^{14}C dates. Total carbon, pollen and plant macrofossils are used to synthesize a vegetation-stack, which is used together with the stacks from seasonal varve formation, flood layers, eolian dust content and volcanic tephra layers to define Landscape Evolution Zones (LEZ). LEZ 1 encompasses the landscape dynamics of the last 6000 years with widespread human influence. The natural oak and hazel forests of the early Holocene back to 10,500 b2k define LEZ 2. LEZ 3, the late glacial between 10,500 and 14,700 b2k, shows the development of a boreal forest with abundant grass and shallow water biomass in the lakes. The maximum of the last glaciation (LEZ 4: 14,700–23,000 b2k) was characterized by sparse vegetation of moss and characeae. These sediments are generally devoid of clay and sand and reveal no indication of snow-meltwater events. Accordingly, the Last Glacial Maximum (LGM) must have been extremely arid in central Europe. The sediments of the subsequent LEZ 5 from 23,000–28,500 b2k preserve distinct layers of clay and coarse sand, which indicates running water with clay in suspension and ephemeral coarse-grained fluvial sediment discharge. Abundant Ranunculaceae macroremains (used for ^{14}C dating), insects, moss and fungi sclerotia reflect a tundra environment during a time of frequent strong snowmelt events. Total carbon content, *Betula–Pinus* pollen and diatoms reach increased concentrations during Marine Isotope Stage (MIS) 3 interstadials that occurred between 28,500 and 36,500 b2k (LEZ 6). The entire MIS3 interstadials are well documented in the organic carbon record from the Auel dry maar. The main paleobotanical indicators of MIS3 are, however, grass pollen and heliophytes, which indicate a steppe environment with scattered/patchy trees during the interstadials. The stadial phases inferred during LEZ 6 reveal initiation of eolian dust deflation. The change of the early MIS 3 forested landscape to a steppe occurred with the LEZ 7–LEZ 6 transition. This is when modern man spread in central Europe. The principle vegetation change to a steppe at 36,500 b2k must have favoured the spread of horses, an important hunting prey of modern humans. We propose accordingly that the migration of the modern humans into central Europe might have been at least partly driven by climate and associated vegetation change. The LEZ 7 encompassed the time interval 36,500 to 49,000 b2k and was characterized by a boreal forest with high abundance of pine, birch, as well as spruce during the interstadial events. Abundant charcoal fragments indicate that this taiga was under frequent drought stress with regular burning. The most unexpected finding, but corroborated by all our maar records is the dominance of thermophilous tree taxa from 49,000 to 55,000 b2k (LEZ 8). Greenland interstadials 13 and 14 were apparently the warmest of MIS 3 according to the Eifel pollen records. The preceeding LEZ 9 from 55,000 to 60,000 b2k is also dominated by spruce, but thermophilous trees were sparse. A warm early MIS3 appears plausible, because summer insolation (at 60° N) was higher in the early MIS 3 than today, ice cover was low in Scandinavia and sea-surface temperatures of the North Atlantic were almost comparable to modern values during GI-14.

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1. Introduction

The global climate of the past 60,000 years is well known from polar ice cores (e.g. Grootes et al., 1993; Johnsen et al., 2001; Petit et al., 2004; NGRIP Community Members, 2004; EPICA, 2004) and ocean sediments (e.g. CLIMAP, 1981; Imbrie et al., 1984; SPECMAP, 1994; Lisiecki and Raymo, 2005). The stadial/interstadial succession of the last glacial cycle was first documented in the Greenland ice cores, but became soon visible also in marine records (e.g. McManus et al., 1994; Behl and Kennett, 1996; Schulz et al., 1999; van Kreveld et al., 2004), reflecting that these sharp climate anomalies are a common feature of the entire Northern Hemisphere ice–ocean–land climate system.

The best chronologies for the last 60,000 years of climate change comes from the layer counted Greenland ice cores (e.g. Parrenin et al., 2001; Svensson et al., 2008) and from Eifel maar sediments for the last 15,000 years (e.g. Zolitschka et al., 2015). Chronologies from the marine, land and glacier records have been recently synthesized into a common chronological framework by the INTIMATE project (e.g. Rasmussen et al., 2014) and summaries on the vegetation and environment of all of Europe have been presented by Moreno et al., 2014; Feurdean et al., 2014; Heiri et al., 2014.

The extent of the European MIS6 and MIS2 maximum moraines is well mapped (CLIMAP, 1981; Ehlers and Gibbard, 2003, 2004) but the ice extent during MIS4 is still poorly understood, even in central Europe. Fluvial deposits (e.g. Kasse et al., 1995, 2003), paleosol sequences (e.g. Haesaerts et al., 2010; Schirmer, 2012) and loess deposits (e.g. Antoine et al., 2001; Frechen et al., 2001; Marković et al., 2008, 2015) have revealed the main stadial and interstadial characteristics in Europe, dated by luminescence methods (Kadereit et al., 2013; Lang et al., 2003; Timar-Gabor et al., 2011; Zöller et al., 2014).

Classical pollen analysis defines the late glacial and Holocene into distinct pollen zones, which have an almost common and identifiable succession all over central Europe (e.g. Allen and Huntley, 2000; Litt et al., 2001; Guiter et al., 2003, 2005; Peyron et al., 2005; Tantau et al., 2005; Magny et al., 2006; Fletcher et al., 2010; Helmens, 2014). Pollen however, cannot be used to describe the climate of the Last Glacial Maximum (LGM) because most glacial lake sediments do not contain pollen or do not preserve them. Accordingly, we will later in this paper introduce "Landscape Evolution Zones" (LEZs). The LEZ are a new approach to combine the paleobotanic evidence with indicators of eolian dust activity, flood events, seasonal varve composition and volcanic activity to arrive at a synthesis of all factors determining a landscape. In addition to these natural environmental processes, we include the archeological evidence for past land use. Accordingly, LEZs are an approach to integrate the various sources of information and reach a synthesis on the nature and evolution of past environments in central Europe.

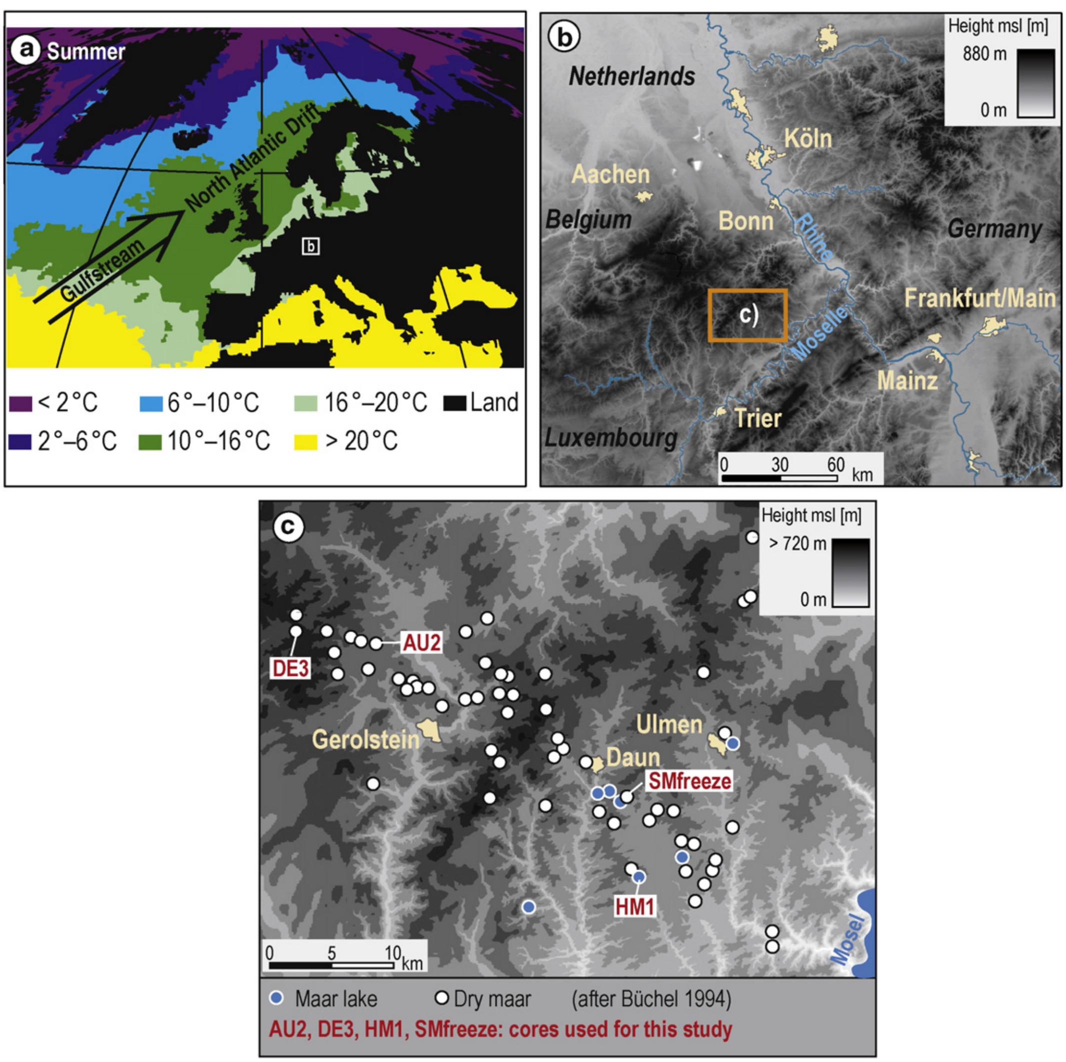

Fig. 1. a) Map of the modern North Atlantic sea surface temperatures and continental boundaries of Europe. b) Digital elevation map of the "Rheinisches Schiefergebirge". c) Digital elevation model of the northwestern volcanic field with drainage system, maar positions (after Büchel, 1994) and coring sites.

1.1. Introduction to the ELSA project

The west Eifel volcanic field (Germany) has six extant maar lakes and about 60 dry maar lakes (former maar lakes filled up with sediments) (Fig. 1). Most of these lakes and dry lakes have been systematically cored since 1998 by the ELSA Project (Eifel Laminated Sediment Archive) of the Institute for Geoscience, Johannes Gutenberg University Mainz, Germany. Today the core repository at Mainz hosts a total of 2700 m of laminated lake sediments. Twenty-one scientific publications, one scientific book and one book for the general public have already been published from this core material. A full list of all peer-reviewed ELSA papers is given by Sirocko (this volume) in the Introduction Paper to this Special Section.

All ELSA cores are laminated, but only the last 30,000 years, the early part of MIS3 and the last interglacial sediments show continuous seasonal layering that can be varve counted (Rein et al., 2007; Sirocko et al., 2005, 2013, this paper). The absolute age control of all ELSA cores is based on ^{137}Cs, ^{210}Pb, ^{14}C, flood event tuning, magnetostratigraphy, ice core tuning, luminescence techniques and argon/argon dating (Sirocko et al., 2013).

An important further step for the ELSA chronology is accomplished with this work, because the new total carbon record of core AU2 from the dry maar of Auel is now the best (highest sediment rate) record from the Eifel to replicate the complete Greenland interstadial succession of the last 60,000 years. The other five Greenland ice tuned ELSA cores had used only the greyscale record, which is not always unequivocal (Sirocko et al., 2013).

The tuning of the AU2 time series to the Greenland ice core timescale in this paper is now the latest update of the age/depth relations for all ELSA cores, because we transfer the ages of the distinct marker tephra in Auel to all other cores. Accordingly, this paper discusses the already published ELSA stacks, but on the new AU2 stratigraphy. All ELSA cores are now correlated to each other by means of tephrostratigraphy (see also Förster and Sirocko, 2016–in this volume) to arrive at a consistent matrix of sediment records aligned within a common multi-method temporal framework.

The basins of the modern maar lakes are between 20 and 70 m deep. The average annual sedimentation rate is ~1 mm/year, which allows a typical maar sediment record of about 70 m length to cover about 70,000 years. Accordingly, it is necessary to splice overlapping cores into a composite stack in order to cover longer time series. At present, the stacked ELSA cores reach over the last 60,000 years (this study), but with potential to extend back to the middle Pleistocene (Förster and Sirocko, 2016–in this volume). The oldest ELSA sediment records date to the time interval 400,000–500,000 years before present (Sirocko et al., 2013).

A table with all published and on-going research is accessible on the web page of the ELSA-project http://www.klimaundsedimente.geowissenschaften.uni-mainz.de.

2. Coring sites and coring methods

Sediment cores have been recovered from open maar lakes with both Niederreiter piston core and freeze core technology (www.uwitec.at), while infilled dry maar basins have been drilled by the company "Stölben-Bohr" using ICDP drilling technology, however, not from a barge (www.stoelben-GmbH.de). Digital elevation models for all ELSA coring sites were presented in Sirocko et al. (2013); the drillings sites for AU2 and DE3 cores are shown with the larger catchment by Brunck et al. (2016–in this volume).

2.1. Schalkenmehrener maar

Lake Schalkenmehrener Maar is part of the Dauner Maar group. It has a diameter of 528 m, an average depth of 14.5 m, and a maximum depth of 21 m. The ELSA Project has drilled a total of 9 cores from this maar lake, including 3 freeze cores. Freeze core SMF1 is dated by ^{137}Cs and ^{210}Pb and shows clearly the 1986 and 1963 nuclear fallout maxima (Sirocko et al., 2013) (Fig. 2). Both cores reveal a unique thick layer with abundant botanical macroremains in the lower section of the cores, ^{14}C dated to the first half of the 14th century (Sirocko et al., 2013);

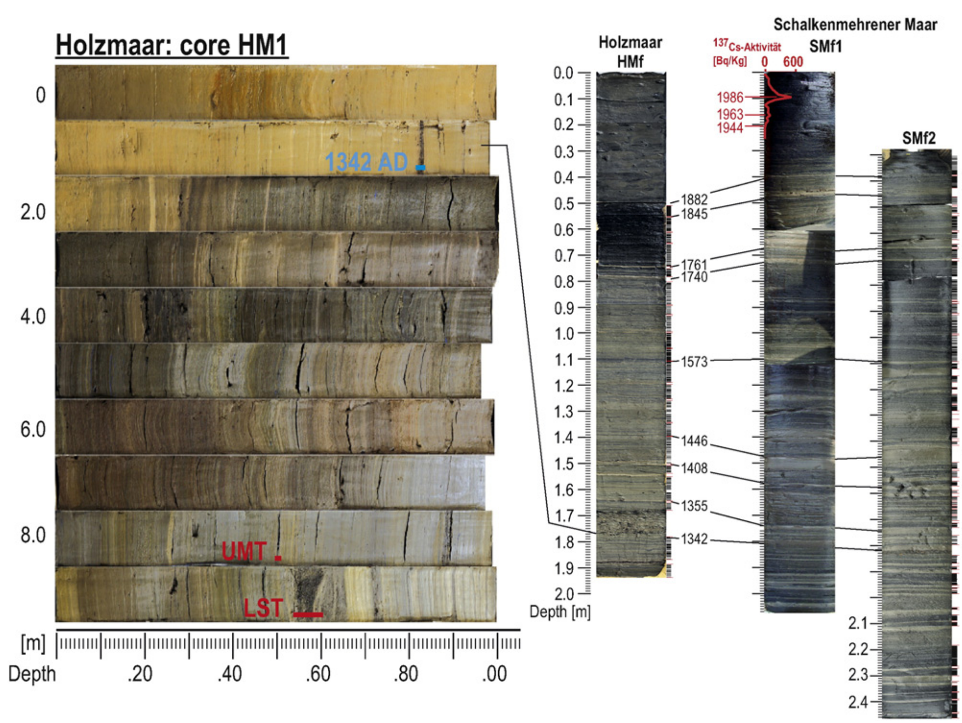

Fig. 2. Photos of varve counted cores HM1 (last 13,000 years) and HMfreeze from Lake Holzmaar (last 1000 years) together with varve counted freeze cores from Schalkenmehrener Maar. Historically documented flood layers (years after Glaser, 2001, and Federal Waterways and Shipping Administration, 2015) are indicated when visible in both maar lakes.

accordingly, this thick layer must represent the millennium flood of summer 1342 CE (Fig. 2). Freeze core SMF2 reveals a superb preservation of laminations, which have been shown by Fritz (2011) to be varves and reach back to the 11th century, countable with a precision of $\pm 5\%$. Some of the visible flood layers in the SMfreeze cores have been correlated with this error margin to known historical flood events (Brunck et al., 2016–in this volume) and are highligthed in Fig. 2. Accordingly, the freezes cores from Schalkenmehren are best suited to monitor the modern and medieval landscape history and are used for the last 1000 years of the ELSA chronology.

2.2. Holzmaar

Holzmaar is well known since long for chronologies from annual varve counting (Negendank et al., 1990; Zolitschka, 1998; Brauer et al., 1999, 2001), however, constrained by ^{14}C-dating (Hajdas et al., 1995). The ELSA project drilled a 2 m long freeze core, which was varve counted by Fritz (2011). This core together with core SMf2 from Schalkenmehrener Maar is presented in Fig. 2, however with a slightly updated stratigraphy to match the observation of locust macroremains, which are known to have invaded central Europe in 1338 and 1408. The other core from Hozmaar is a 10 m long piston core from the Holzmaar, which is used in this study for the Holocene varve chronology, pollen and plant macrofossil analyses (Fig. 2). The piston core was retrieved in two-meter long sections, which were cut into one-meter lengths, so that minor sediment loss may have occurred between meters 2/3, 4/5, 6/7 and 8/9 in the core. The sediment record of HM1 reaches the Laacher See tephra (LST) at 9.57 m depth (Fig. 2).

The varve time series for ELSA core HM1 is presented in this study. However, problems of missing varves and floating sections were encountered as in the earlier Holzmaar studies and solved by anchoring the floating varve chronologies to the well-dated flood layers at 1342 CE and 800 BCE and the Laacher See tephra at 12,880 b2k (Fig. 2). The ELSA-pollen record from core HM1 is of century scale resolution, but resembles all important stratigraphical features of the decadal scale resolution pollen results from Kubitz (2000); Stebich (1999); Litt and Stebich (1999) and Litt et al. (2001, 2003) for Meerfelder Maar (Brauer et al., 1999, 2001).

2.3. Dehner maar

The Dehner Maar (Figs. 1 and 3) is located in the northwestern part of the Western Eifel volcanic zone, north of the town of Reuth. The Dehner maar is still very well recognizable in the landscape, because it was infilled only during the early Holocene. It has a diameter of 950 m and lies at an elevation of 565 m. The maar basin is near circular and does not display any recognizable traces of past inflow, however there is an outflow to the south. Accordingly, the Dehner Maar is perfectly located to study the eolian input over the Eifelregion between the end of MIS 4 and the MIS 1/2 transition.

The core DE3 from the Dehner Maar is 87 m long and preserves the LST at 3.47 m depth (Fig. 3) and is paralleled by core DE2 drilled nearby. Both cores have a total of 18 radiocarbon ages, which are in agreement with chronological estimates derived from paleomagnetic data (Sirocko et al., 2013).

The DE3 sediments display annual laminations from 12,900 b2k back to 32,000 b2k. Abundant thermophilous tree pollen occur between 60 and 74 m depth, and pollen continue during all of MIS3 (Supplement 1). The stratigraphy of the *Picea*-dominated DE3 pollen profile was already persented by Sirocko et al. (2013), but is slightly modified in

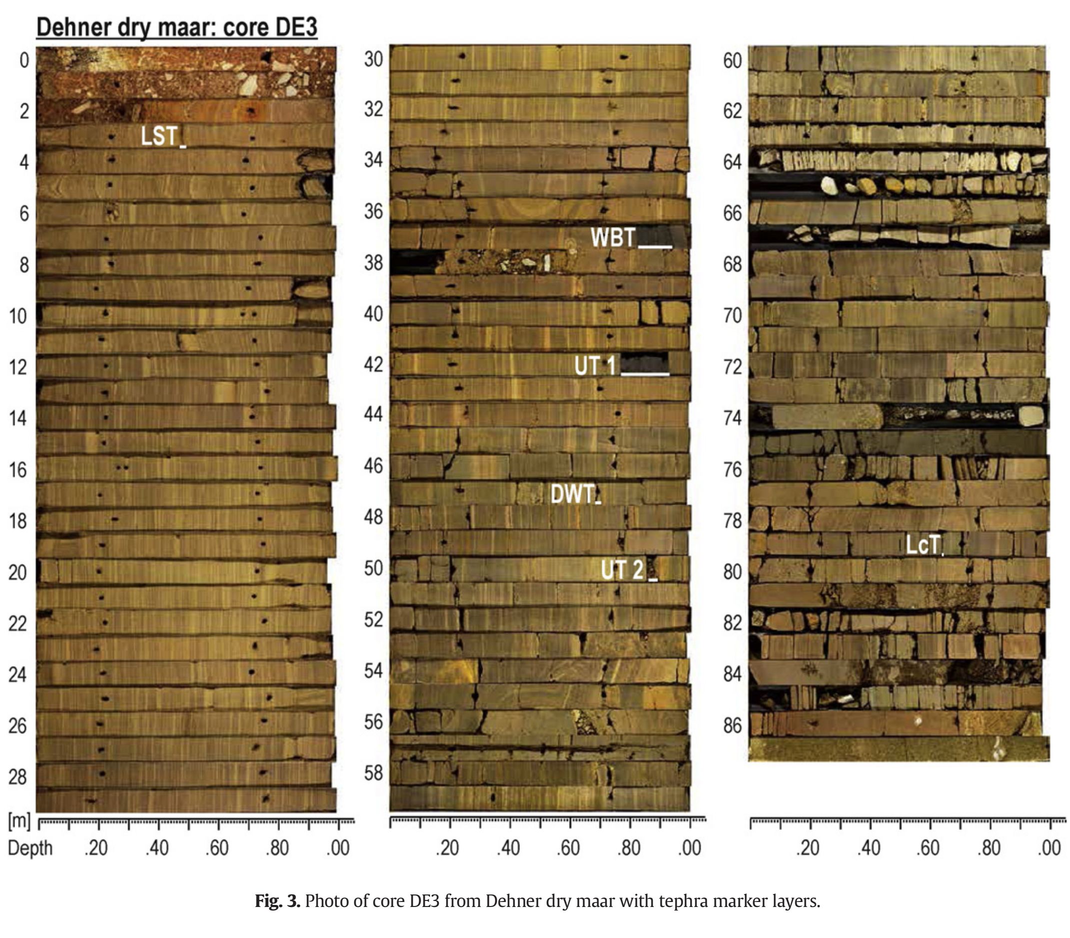

Fig. 3. Photo of core DE3 from Dehner dry maar with tephra marker layers.

this paper, because the upper section back to 30,000 years is now on the varve counted time scale and the lower part on the new Greenland ice core-tuned Auel time scale.

550 samples from the Dehner Maar were investigated for plant macroremains from 15 to 70 m depth (Supplement 2). Samples cover 10 c*m*-long core sections and each represent 200 g of dry sediment. The DE3 samples contained only 61 non-carbonized and carbonized plant macrofossils, of which only a fraction could be identified to the species level. The macroremains in the Dehner Maar sediment are recorded as present or absent, because the amount of 200 g of sediment was not large enough to find enough remains for calculation of concentration values.

The core DE3 is also used for the diatom analysis presented in this paper, because the diatoms can be well seen in the petrographic thin sections used for varve counting, which allows for annual resolution diatom analysis (Supplement 3). The data presented in this paper are, however, done with discrete samples in 10 cm intervals.

2.4. Auel dry maar

The Auel dry maar is one of the largest infilled maar lakes in the Eifel with a diameter of 1325 m. The Tieferbach brook flows today through the centre of the maar. This river has a large catchment area and a length of 9.4 km so that the sediment record contained in Auel maar is well-suited for reconstructing the extent of past riverine input into the lake (see Brunck et al., 2016–in this volume). Core AU2 is 123 m long (Figs. 1 and 4a, b) and has the highest sedimentation rate (2 mm/year on average) of all ELSA cores, due to the fluvial sediment influx to the basin. Half of each meter of core from AU2 (about 4 kg for each sample) was sieved and 3804 plant macrofossils were identified in total, including 2951 Characeae oogonia, 820 seed and fruits, of which 89 were identified to species level (Supplement 4). Pollen in core AU2 were determined only in low resolution in 1 m-increments (mix over entire 1 m core section) (Supplement 5).

The succession of pollen and tephras is identical between the Auel and the Dehner Maar, but AU2 has much higher number of plant macrofossilss due to the sample mass of 4 kg; accordingly macroremains are presented in counts/kg. Frequently encountered Ranunculaceae/*Ranunculus* seeds were used to obtain four ^{14}C dates from the glacial section, all of which reveal ages around 21,000 BP, possibly the GI2 interstadial.

2.5. The ELSA core stack

It was the intention of the ELSA Project to drill systematically all maar lakes and dry maar structures of the Eifel to obtain an understanding, which sites are best suited to form a matrix of correlated cores reaching from the modern back to the beginning of the Eifel volcanism in the middle Pleistocene.

The Core-Stack$_{2016}$ is presented in this paper. It starts with the ^{137}Cs dated core SMF1 for the last 50 years (Fig. 2). It continues with the varve counted cores SMF2 and HMf (Fig. 2) down to the 1342 CE millennium flood layer, which is used to anchor the varve counted Holocene time series of HM1. The Holocene varve counts in Holzmaar reach to the Laacher See Tephra at 12,900 b2k. This well visible layer with a distinct geochemical and mineral composition is found in the upper sections of core DE3 and AU2 from the dry maar sites of Dehner Maar (Fig. 3) and Auel maar (Fig 4). The record from DE3 is varve counted from 12,900 back to 32,000 b2k. Accordingly, the ELSA varve chronology is continuous from today back to 32,000 b2k. The record of AU2 from Auel cannot be varve counted, despite varves are visibly occasionally. AU2, however, shows a succession of total carbon maxima, which is used to tune AU2 to the Svensson et al. (2008) time scale for the time interval 25,000 to 60,000 b2k, thus overlapping with the varve counted DE3 varve chronology.

The eruptions of the Auel and Dehner maars occurred during the MIS4–3 transition at 60,000 b2k, which is the lowest date included in the ELSA-Stacks$_{2016}$. A continuation of the stack approach is shown for the Tephra-Stack$_{2016}$ already in the paper by Förster and Sirocko (2016–in this volume) and will be finally presented for all other stacks back to 220,000 b2k.

3. Proxies and methods

3.1. Total carbon (C_{total})

C_{total} was determined in core AU2 using elemental analyses. Dry combustion of subsamples (up to 20 mg) was carried out with a Hekatech elemental analyzer coupled via a Conflo III Interface to a Delta V advantage IRMS. For 105 samples taken at lower-resolution, an additional pre-treatment using 24 h HCl fumigation in silver capsules at 60 °C was followed by 24 h KOH treatment at 60 °C under vacuum in order to remove carbonates and determine total organic carbon (TOC). C_{total} and TOC are highly correlated ($R = 0.98$) for the low-resolution samples, and the carbonate fraction does not exceed 2%. Thus, only C_{total} is determined for the high-resolution time series and is interpreted as primarily presenting organic carbon.

3.2. Petrography of tephra

The West Eifel volcanic field is part of the "Schiefergebirge" in central Germany and shows a total of about 280 volcanic structures, 68 of which are Pleistocene Maar lakes (Fig. 1, after Büchel and Lorenz, 1982). The East Eifel volcanic field is geochemically and geodynamically characterized by much more differentiated magmas compared to the primitive West Eifel volcanic field magmas. Van den Boogard and Schmincke (1989) developed the general chronological framework of past volcanic activity in the East Eifel by argon/argon dating of the most differentiated tephras from the East Eifel volcanic field. These tephras can be clearly identified petrographically in the lake sediment records of the West Eifel volcanic field (Förster and Sirocko, 2016–in this volume). Accordingly the ash plumes from these explosive East Eifel eruptions reached the maar lakes and provide robust marker tephras. The main tephra were distinguished geochemically by Sirocko et al. (2013), but can now also be quantified by fast and efficient analysis of the sand fraction mineralogy only (Förster and Sirocko, 2016–in this volume).

Förster and Sirocko (2016–in this volume) examined the visible tephra layers >1 cm-thickness in all of the ELSA sediment core records. Bulk tephra samples of 2 g were sieved at 250–100 μm for petrographic characterization under a binocular microscope. A total of 100 identifiable crystals and grains were counted into ten groups including reddish sandstone, grayish sandstone, quartz, amphibole, pyroxene, scoria and pumice, sanidine, leucite and mica. The selection of these minerals and rock fragments was chosen based on their weathering resistance and characteristic appearance. Histograms of the count-results represent the volume percent (%) abundances for each grain type in the total from the 10 defined classes (Förster and Sirocko, 2016–in this volume).

3.3. Varve counting

Varve counting was carried out in three different ways. Visible mm-thick laminae were counted directly from photographs of freeze core SMf from Schalkenmehrener Maar (Fritz, 2011). The photos were taken immediately after core recovery on the drilling platform, when the colour of the dark and light seasonal laminae was clearly visible. Direct counting from photographs is most efficient when sedimentation rate is >1 mm/yr and sediment laminae are preserved by the freezing process up to the sediment/water interface. The main advantage of this approach is the 20 cm core width of completely undisturbed sediment, which often allows the identification of problematic layers much better than on a 2 cm-wide petrographic thin section.

10 cm-long petrographic thin sections were prepared for cores HM1 and DE3. Both cores have an excellent preservation of laminae and

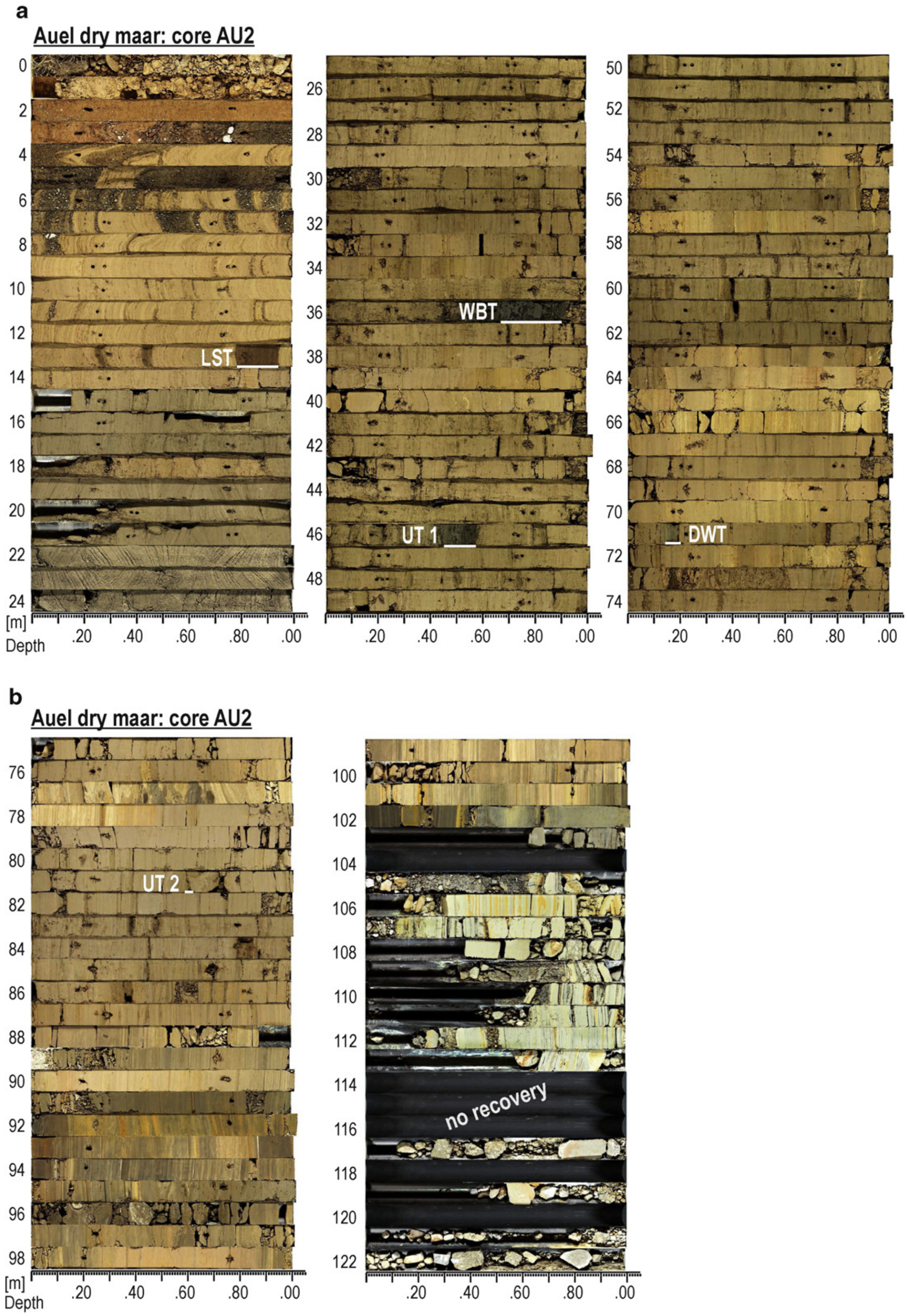

Fig. 4. a Photo of core AU2 from Auel (0-75 m) dry maar with tephra marker layers. b Photo of core AU2 from Auel (75-123 m) dry maar with tephra marker layers.

allow counting of annual layers (varves) already at 20 times magnification under the petrographic microscope. The average thickness of well identifiable varves was determined and then applied to the entire length of the 10 cm thin section to calculate the number of years below the LST tephra; an approach which resulted a continuous record from 1342 CE back to 32,000 b2k.

The varve chronology between today and 1342 CE is based on the freeze cores from Schalkenmehren and Holzmaar, which were continuously counted by Fritz (2011). Sirocko et al. (2013) anchored these varve time series to the several cm-thick 1342 CE flood layer (658 b2k). Brunck et al. (2016–in this volume) have now included historical flood layers at 1882, 1845, 1761, 1740, 1573, 1446, 15,408 and 1355 into an updated varve chronology of the last 1000 years (Fig. 2).

The next deeper anchor point is the 800 BCE flood (2800 b2k) (see also Martin-Puertas al., 2012). The subsequent tie points were defined by pollen analysis. In particular, the spread of beech trees in the Eifel occurred at 3800 b2k (1800 BCE) (Kubitz, 2000; Litt et al., 2001) whereas the elm decline is centred between 6300 and 6000 b2k. The 10 cm-thick sand layer with abundant grey pumice of the LST (12,880 b2k) is the tie point for the older cores from the dry maars covering MIS 2 and 3 (Figs. 2, 3, 4).

The only varve-countable sediment record below the LST is DE3. The DE3 varve chronology show 19,000 layers below the LST (12,900 b2k) and place the UT1 at 28,900 b2k. The Greenland ice tuned age of the UT1 in AU2 is, however, 30,200 b2k. Accordingly, the missing varves in the glacial section are 4% in comparison to the ice core tuned age model.

3.4. Pollen

Paleobotanical investigations of the volcanic lake sediments in the Eifel region have a long scientific tradition (Straka, 1975; Usinger, 1982) and were concentrated on the extant maar lakes (Litt et al., 2001, 2003; Kubitz, 2000; Litt and Stebich, 1999; Stebich, 1999). Palynological data from older maar structures (infilled maar lakes) have been published for MIS 5e (Sirocko et al., 2005) and MIS 11 (Diehl and Sirocko, 2005).

Pollen preparation followed the techniques of Berglund and Ralska-Jasiewiczowa (1986) and Faegri and Iversen (1989). Each pollen sample spans a depth range of 1 cm (except for AU2, which was sampled as a mix for each complete core meter representing 500 years) and was of about 1 cm^3 volume. The sediment was treated with potassium hydroxide solution (KOH), hydrochloric acid (HCl) and hydrofluoric acid (HF). For acetolysis, acetic acid ($C_2H_4O_2$) and a mixture (9:1) of acetic anhydride ($C_4H_6O_3$) and sulphuric acid (H_2SO_4) were used. Centrifugation was done at 3000–3500 rpm for 5 min. The samples were sieved at 200 μm and later filtered at 10 μm. *Lycopodium*-spore tablets were added for calibration of absolute pollen concentration per cm^3. The samples were mounted with liquid, anhydrous glycerol ($C_3H_8O_3$). Pollen counting was done under a maximum of six hundred-fold magnification. 300 pollen grains are counted for each sample. Percentages (% of all pollen) are presented only for those samples were the sum of all taxa exceeds 50 pollen grains. In addition, total pollen content (#/cm^3) have been calculated using the known numbers of lycopodium spores. The count data are presented, because the percentage data alone of *Pinus* and *Betula* are almost constant during MIS 3, but the pollen counts per ccm show clear maxima in the MIS 3 interstadials. All pollen data are documented in Supplements 1 and 5 and plotted versus depth in Figs. 5–8.

3.5. Plant macroremains

Pollen grains are dispersed by the wind over short or long distances and it can never be unambiguously inferred that the plants which produced a pollen grain grew near to the site of pollen deposition. This is different with plant macrofossils which always represent plants from within the catchment area of the lake. Thus, the combined evidence of a certain plant from both pollen and macroremains provides reliable evidence that this plant grew near the lake. However, plant macroremains and pollen can be reworked by slumping and turbidites, and thus could represent old components incorporated into younger sediment. Such events are episodic and have been identified and taken into account for the interpretation of isolated, rather unexpected spikes in all of the datasets presented.

Macroremains of aquatic plants or submerged macrophytes, like Characeae, which can drift, are dispersed within the lake naturally by surface currents. The other plant macroremains are transported by the fluvial inflow and represent the plants in the entire catchment. Thus, plant remains found in the sediments represent a mix of habitats, transported by different processes, but originate from within the catchment area of the maar lake.

Our study also includes macroscopic charcoal fragments resulting from lake proximal fires linked to human activity in the Eifel during

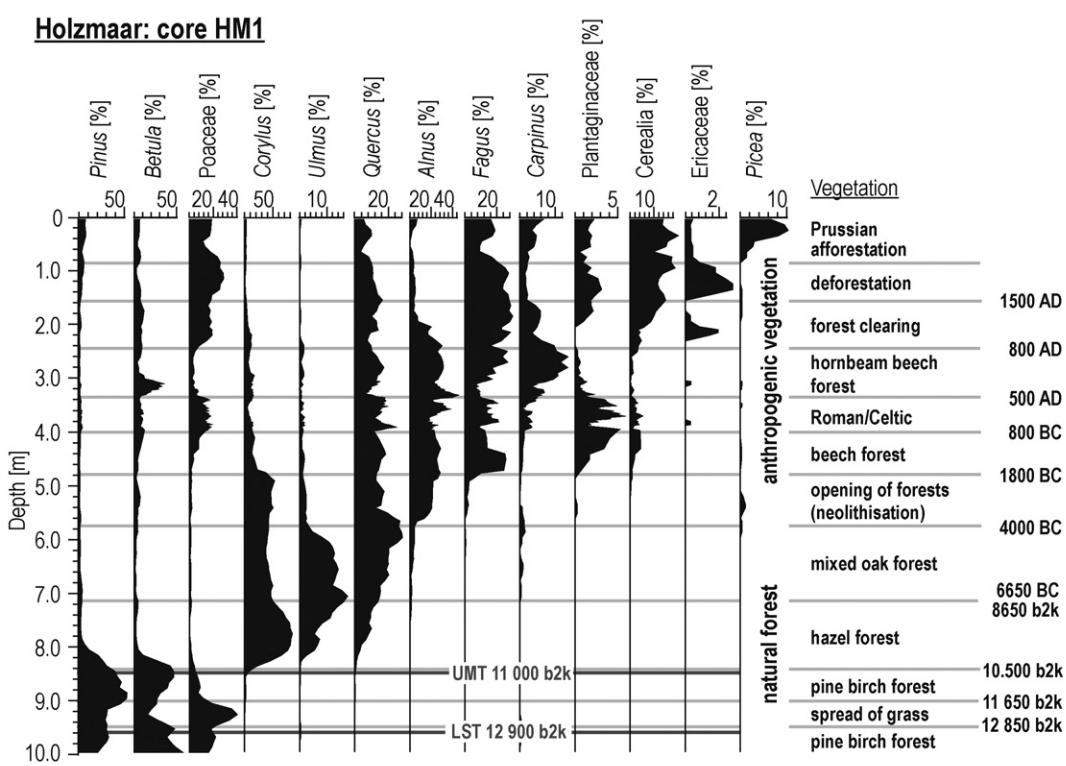

Fig. 5. Pollen concentrations, macroremains and marker tephras of core HM1 from Holzmaar versus depth.

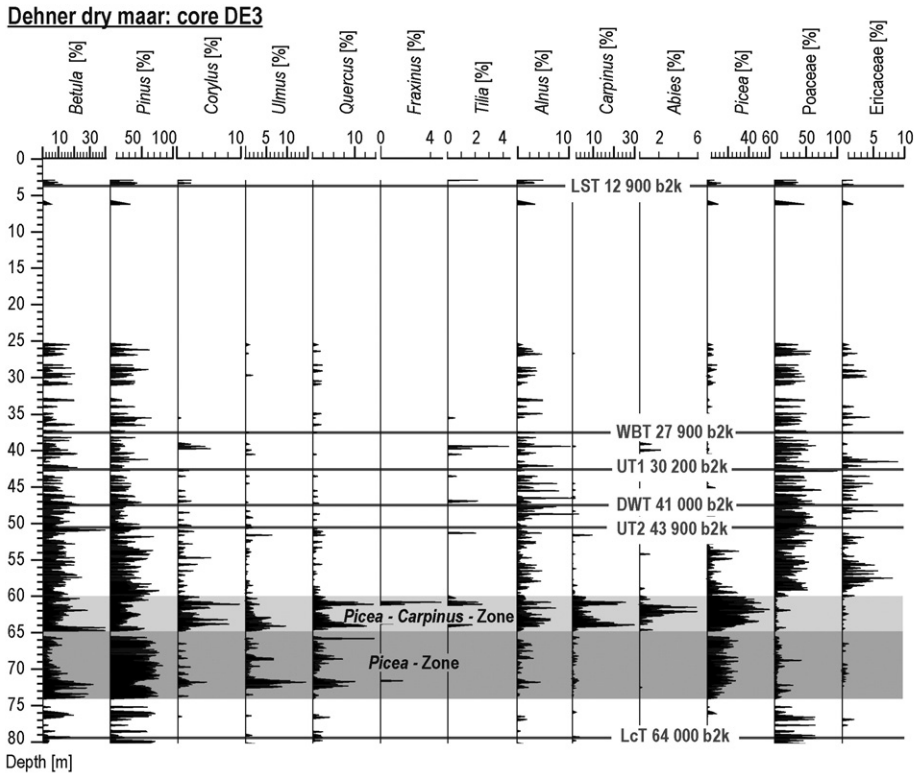

Fig. 6. Pollen percent concentrations and marker tephras of core DE3 from Dehner dry maar versus depth.

late Holocene (Herbig and Sirocko, 2012; Bandowe et al., 2014). The first MIS3 charcoal maxima should represent natural forest fires, because the strongest maxima occurred before anatomically modern humans moved into central Europe.

Sample preparation for plant macrofossil analysis followed Jacomet and Kreuz (1999). The sediment was soaked in water for several hours and then wet-sieved with at 200 μm. Plant macrofossils were identified and counted at 12 to 40 times magnification under a stereo-

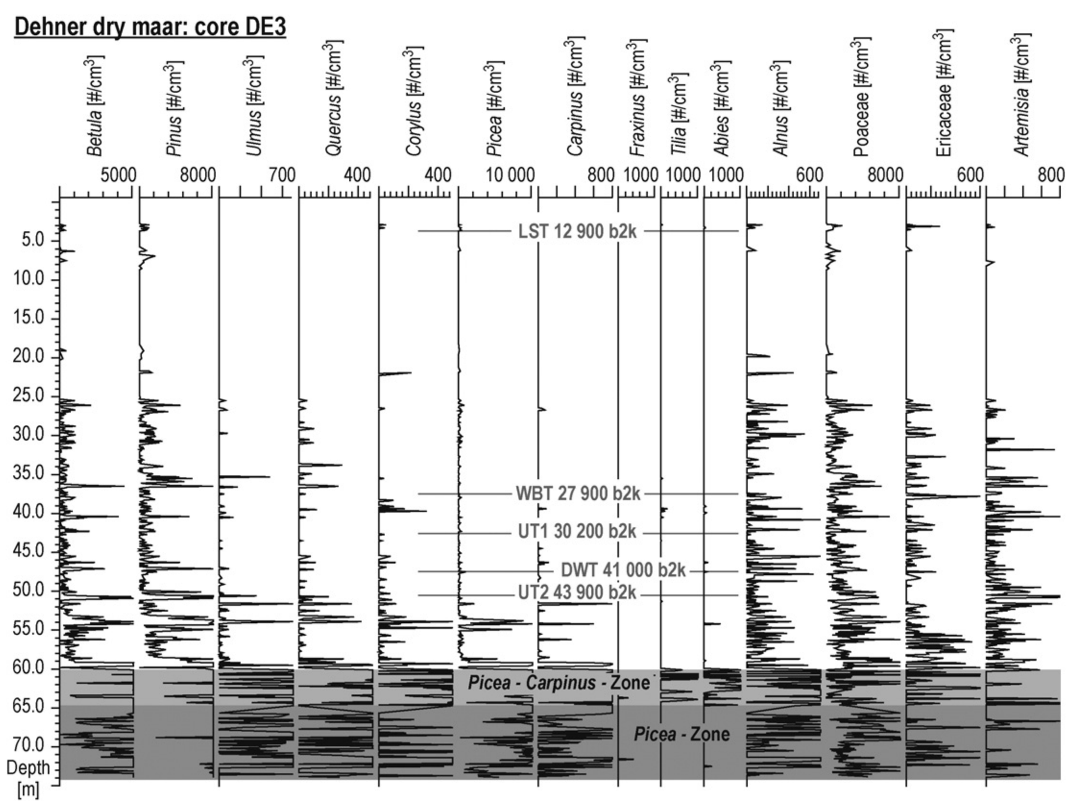

Fig. 7. Pollen counts [#/cm^3] of sediment and marker tephras of core DE3 from Dehner dry maar versus depth.

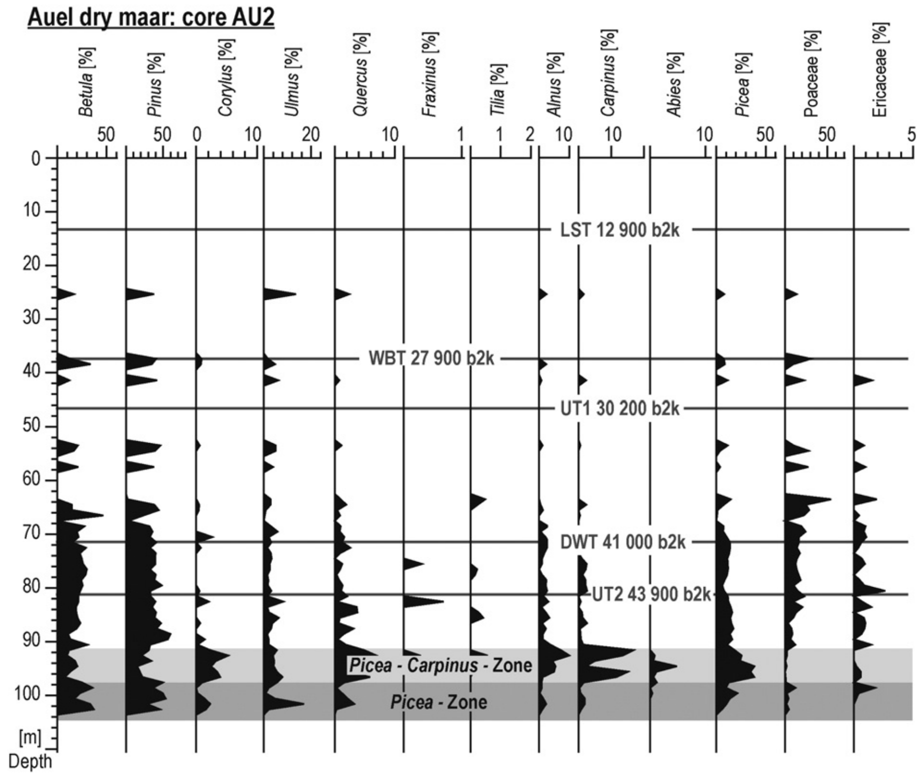

Fig. 8. Pollen concentrations of core AU2 from Auel dry maar versus depth.

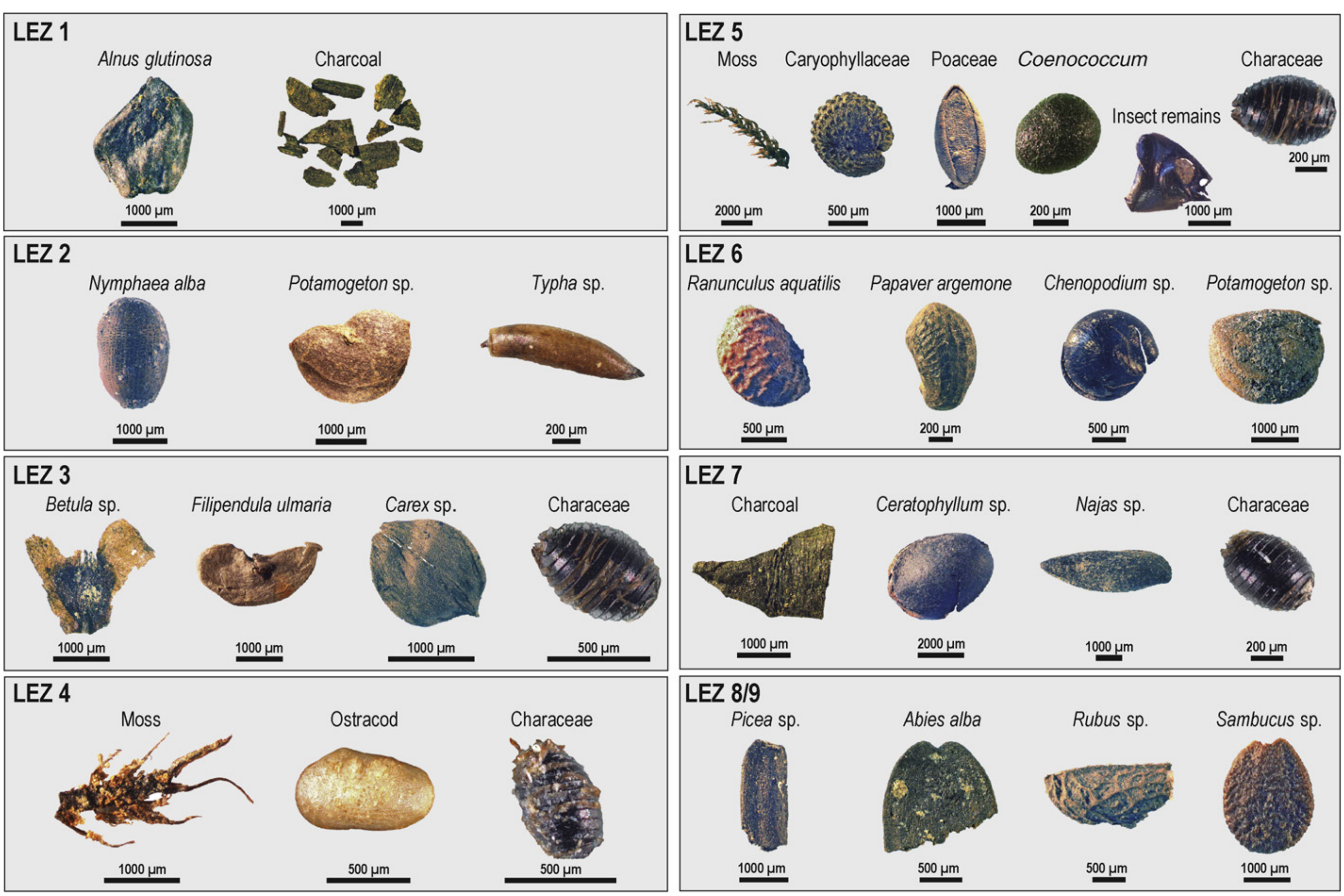

Fig. 9. Photos of selected macroremains typical for the LEZ.

microscope. For identification, guides (Jacomet et al., 1989; Mauquoy and van Geel, 2007; Cappers et al., 2006) as well as different reference collections in Kiel and Frankfurt were used.

The amount of plant macroremains was insufficient to calculate percentage values, so that they are presented as count numbers per kilogram, or presence/absence of a respective species, or as the sum of species belonging to a certain ecological group (e.g. crop weeds, littoral/reed, submerge plants).

Other identifiable organic components frequently found in the sediments are mosses, oospores and ostracods. The dataset listing all plant

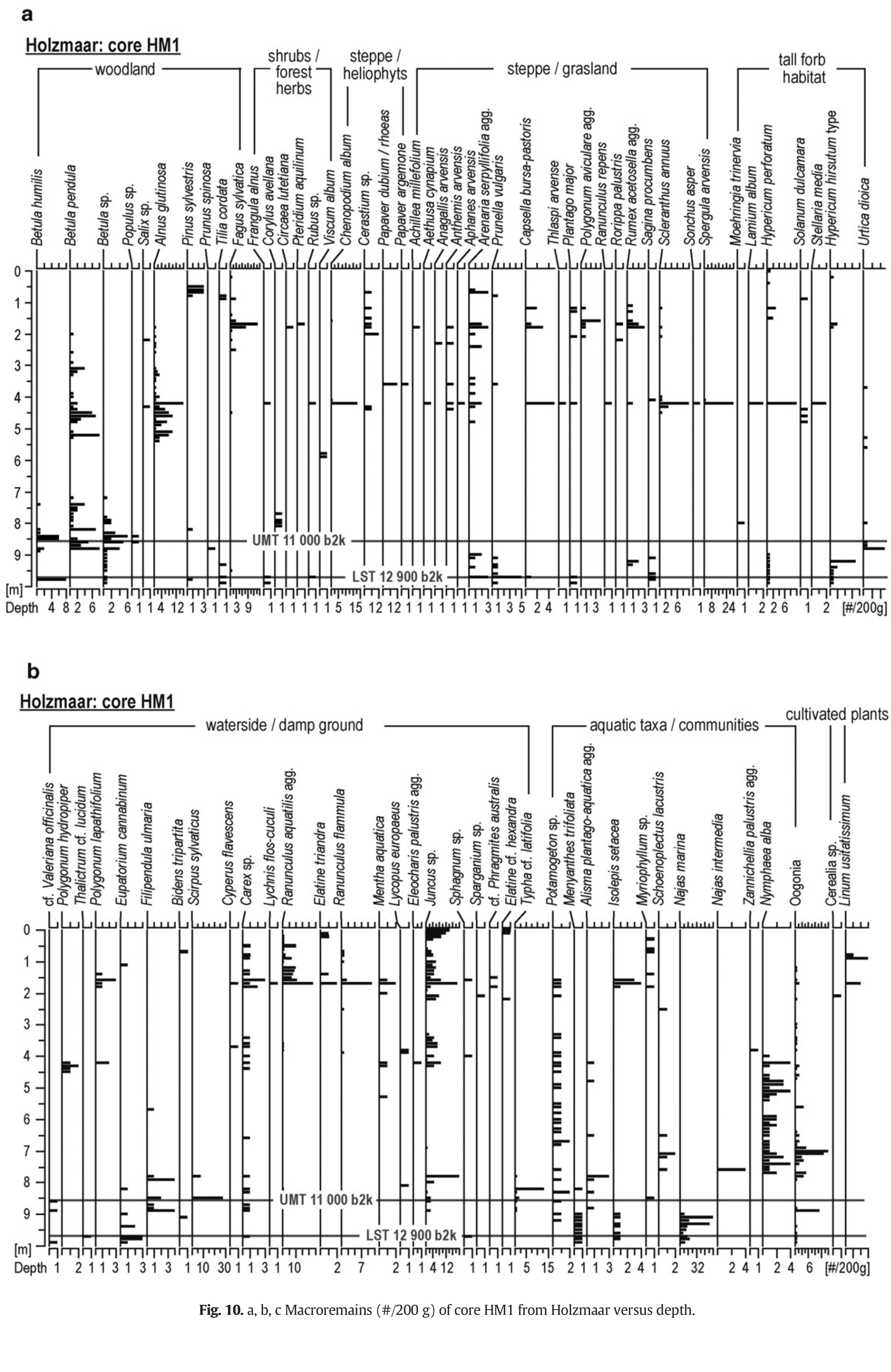

Fig. 10. a, b, c Macroremains (#/200 g) of core HM1 from Holzmaar versus depth.

macroremains is provided in Supplementes 2 and 4; whilst photographs of selected macroremains are presented in Fig. 9; respective downcore plots for HM1, DE3 and AU2 are given in Figs. 10-12.

Paleotemperature estimates (mean July temperature for LEZ 8) were carried out based on the climate indicator plant species method (Iversen, 1944). The indicator species method (Iversen, 1944) links the distribution of specific plant taxa to climate parameters (i.e. mean January and mean July temperature). Only the presence of selected indicator species with defined climate requirements were recorded and interpreted (e.g. *Typha* sp. *Ceratophyllum demersum*, *Schoenoplectus lacustris* and *Najas marina*).

3.6. Diatom analysis

250 samples of about 50 mg collected from core DE3 were sieved and treated with concentrated HCl and H_20_2 to remove the carbonates and the organic matter. The minerogenic fraction was then removed by decantation, and further sieving in order to concentrate the diatom tests. 235 different diatom species were identified and documented (Fig. 13 and Supplement 3). The taxa were grouped into planktonic and benthic species and presented as total sums. The benthic diatoms are further divided into species that live on the sediment and rocks (epilithic) and those that live on underwater vegetation (epiphytic). Maxima in the ratio of epiphytic to epilithic are thus an indicator of the abundance of underwater vegetation and of relatively high water temperature; minima in contrast indicate cold lakes with clastic bottom sediments — as today typical of alpine glacier lakes.

3.7. Eolian dust

Eolian dust can be best reconstructed from the Dehner Maar records, because this maar has no fluvial inlet, but an outlet, which stabilizes the water level to some extent. The records from Dehner Maar were used to construct the ELSA dust stack (Seelos et al., 2009; Dietrich and Seelos, 2010; Dietrich and Sirocko, 2011). The method of dust analysis is based on the sorting of the quartz silt fraction, detected from petrographic thin sections under a microscope with crossed nicols. The sorting is then used to approximate the percentage of the eolian fraction in the bulk quartz fraction between 20 and 63 μm size.

The resolution of quartz grain size detection is in 0.5 mm steps, which allowed a seasonal resolution of the primary ELSA-Dust-Stack (Seelos et al., 2009). We include the DE3 part of the dust stack in this paper, however using the new stratigraphy. The offset of the new DE3 record from the MIS2 stratigraphy of Seelos et al. (2009) is a maximum of 1000 years, because the record is now presented on the new varve time scale, not on the tuned time scale.

3.8. Flood events

Flood events can be best reconstructed from Auel maar because a large creek flows into the maar before leaving it on the opposite side.

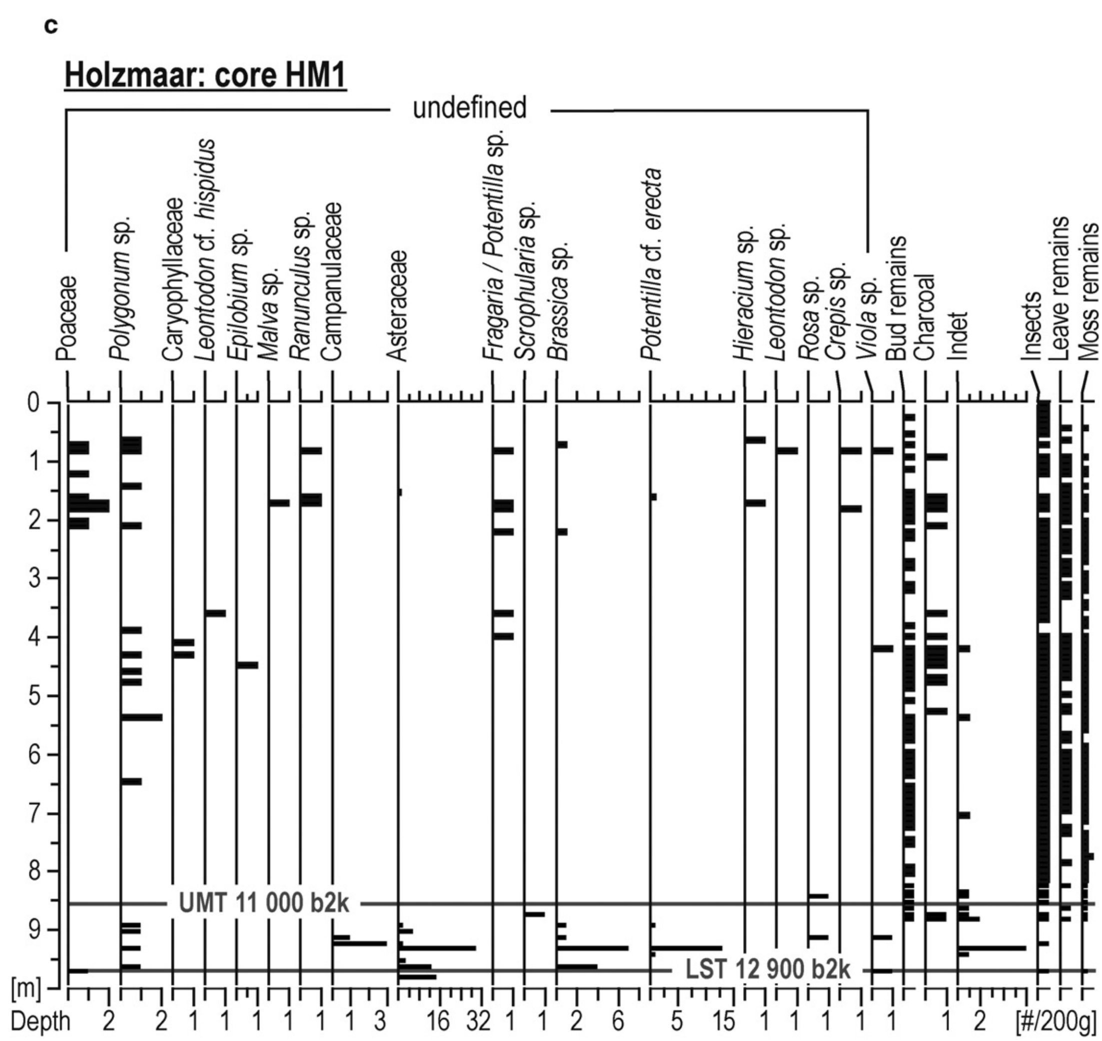

Fig. 10 (*continued*).

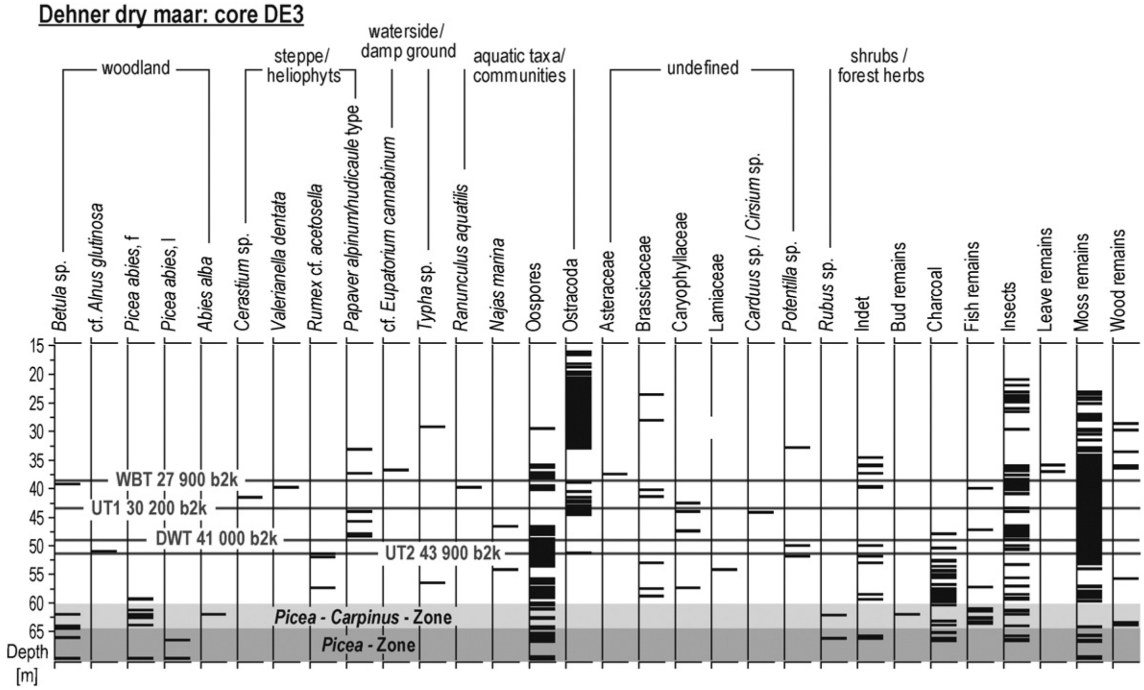

Fig. 11. Macroremains (presence/absence) of core DE3 from Dehner dry maar versus depth.

During the lake phase of this maar, all sediment influx must have settled in the deep basin. The fluvial inflow led to the generally very high sedimentation rate, as well as discrete flood layers emplaced following rainfall events.

10 cm-long thin sections were analysed by µXRF for elemental geochemistry at 100 µm resolution and for grain size on a petrographic thin section in order to distinguish flood layers from distal turbidites. Turbidites have a continuous grain size gradation whereas the grain-size profile of flood events is in contrast characterized by several grain size maxima over the entire layer thickness, because a flood event over several days shows numerous plumes of intense discharge leading to a discontinuous grain size gradient (see Brunck et al., 2016–in this volume). In addition, the thickness of each flood layer was measured for the classification of the event intensity. Accordingly, 88 flood layers over 7.5 mm thick were detected over the last 60,000 b2k (Brunck et al., 2016–in this volume). The normalised flood index was calculated by the number of flood layers per millennium divided by the maximal number of flood layers per millennium in each core.

4. Results

4.1. Chronology

4.1.1. Tephrachronology

Ar/Ar ages for the highly differentiated volcanic eruptions of the Laacher See Tephra, Dümpelmaar Tephra, Glees Tephra and Hüttenberg Tephra were established by van den Bogaard et al. (1989). Sirocko et al. (2013) presented geochemical evidence that the ash from all of these classical Eifel tephra were indeed deposited in the Eifel Maar lakes. Förster and Sirocko (2016–in this volume) characterized these tephra by petrographic analysis of the coarse-grained mineral fraction and transferred the tephra Ar/Ar ages of van den Bogaard et al. (1989) to the ELSA sediment cores. Accordingly, the tephra from Laacher See (LST), Dümpelmaar (DMT), Glees (GT) and Hüttenberg (HT) have a characteristic mineral composition, which can be used to correlate cores on an absolutely dated age scale (Fig. 14).

The ELSA cores also revealed numerous other tephra, all of which were analysed with the coarse-grained mineral method of Förster and Sirocko (2016–in this volume). This increased the number of correlation tie points significantly (Fig. 14). The ages for these non-Ar/Ar dated tephra are taken from core AU2, and are thus on the Greenland ice core stratigraphy.

The same petrographic method was also applied to samples from the tuff walls of 30 maars in the West Eifel volcanic field (see also Förster and Sirocko, 2016–in this volume). Apparently, the petrographic composition of the tephra from Wartgesberg (WBT) and Dreiser Weiher (DWT) were repeatedly found in numerous sediment cores (Figs. 3, 4, 14). The Dreiser Weiher ash is most clearly distinguished, because this is the only very primitive tephra, which contains abundant pyroxene together with 10% sanidine (a composition only explained if the sanidines are xenoliths, Förster and Sirocko, 2016–in this volume). Accordingly, the Dreiser Weiher Tephra (DWT) at 41,000 b2k stands out as the best petrographically distinguishable stratigraphic markers of MIS 3 lake sediments in the Eifel.

All cores with WBT and DWT also contained two ash layers with a very distinct primitive composition for which the site of eruption location is unknown (UT1, UT2), but could still be used as marker tephra if WBT and DWT were present as well. The last tephra of importance for the time period studied in this paper is a unique leucite bearing tephra (Lc1), which is found below the early MIS 3 *Picea*-Zone and serves as the oldest marker tephra of the last 60,000 years.

4.1.2. Varve counting

Annually laminated lacustrine sediments are excellent archives for reconstruction of the climate and environmental changes of the past since the pioneering work of de Geer (1912) up to the most recent review of Zolitschka et al. (2015). The annually laminated sediments of the Eifel maar lakes in central Germany were one of the first to allow a reconstruction of a varve time series for the Holocene. The first ^{14}C adjusted varve counts were completed for Holzmaar (Negendank et al., 1990; Zolitschka, 1991, 1998; Hajdas et al., 1995) followed by continuous varve counting of Meerfelder Maar (Brauer and Negendank, 1993; Brauer et al., 1999, 2001). Holzmaar covers the last 23,220 years, Meerfelder Maar the last 14,200 years. Varve counts in Eifel maar sediments below 23,000 b2k has not been reported yet, but varves have

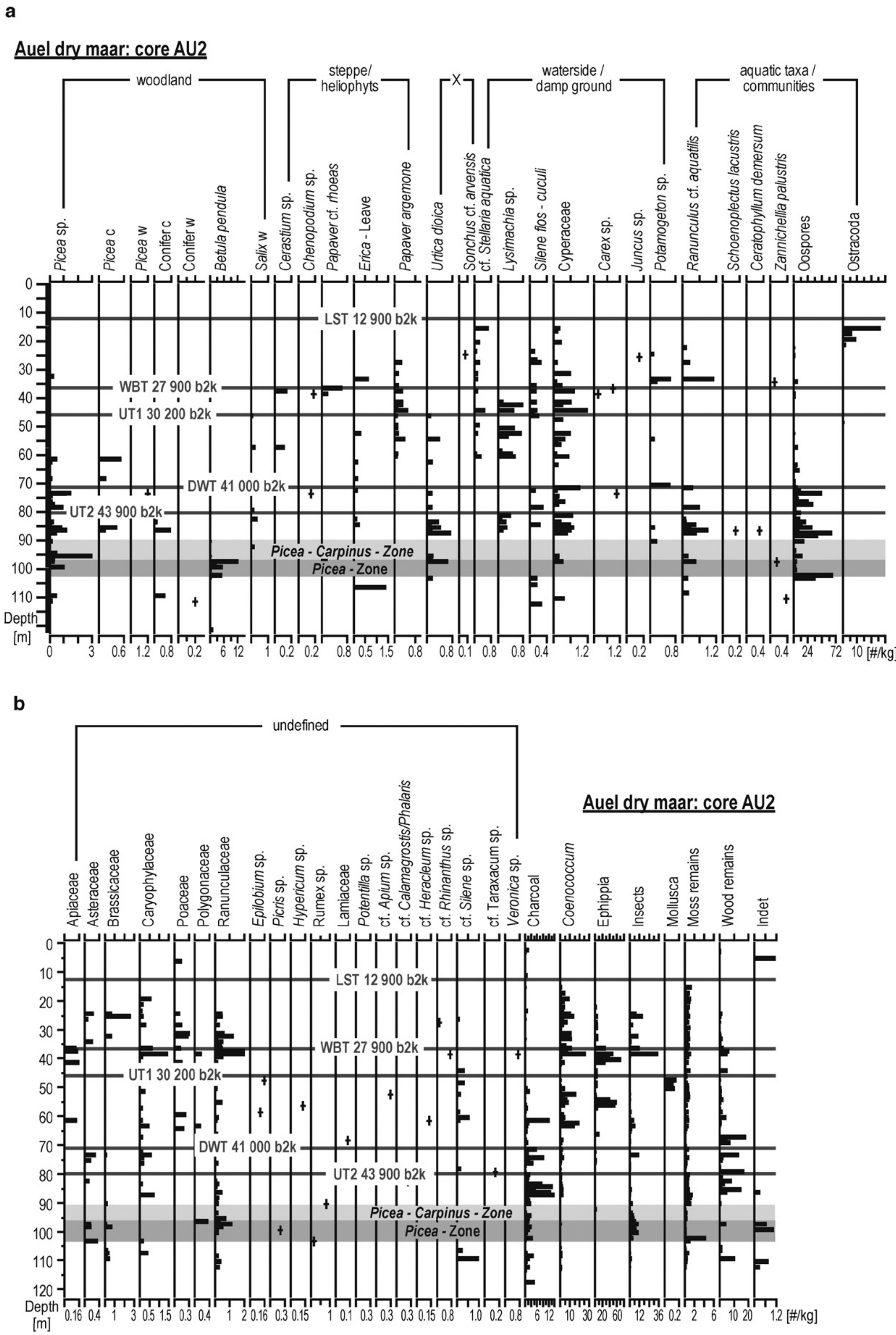

Fig. 12. a, b Macroremains (#/kg) of core AU2 from Auel dry maar versus depth.

been counted for the last interglacial in the dry Maar westlich Hoher List (Sirocko et al., 2005; Rein et al., 2007).

All of the cores presented in this paper are laminated, but only few, or selected time intervals, allow the production of a varve chronology (Figs. 15, 16). The freeze cores from Schalkenmehren (Fritz, 2011; Sirocko et al., 2013) and Holzmaar (HM1) together provide a continuous varve counting during the entire Holocene from the present down to the Laacher See Tephra (LST: 12,880 b2k) (Fig. 2). Varve thickness during the Holocene mainly documents the erosion efficiency in the catchment and show a first spike up core at 6300 b2k (4300 BCE) at Holzmaar when first Neolithic farmers began to clear the natural forest around the maar (Fig. 16). The second spike is centred after the introduction of the iron plough and the flood at 2800 b2k (800 BCE). The third Holocene varve maximum was during the medieval period when mining and charcoal production increased strongly, leading again to extensive forest clearing (Fig. 16).

The varve chronology below the LST was extended with core DE3 from Dehner Maar, which is annually-laminated from the LST down to the Wartgesberg Tephra (WBT) at 27,900 b2k and further down to 30,200 b2k. The respective age/depth relation is given in Figs. 17, 18. Examples of the thin section itself are shown on the ELSA web page. The ELSA varve stack chronology is accordingly continuous from modern times back to 30,200 b2k. The major characteristic of the glacial section is a spike during the times of permafrost melt near 14,000 b2k, which must have led to strong erosion on the maar flanks. The second glacial feature is a pronounced increase in varve thickness at 23,000 b2k, marking the onset of intense LGM dust activity. Finally, the varve counting corroborates independently the age of the ice core -tuned age of WBT at 27,900.

4.1.3. Radiocarbon dating

Schaber and Sirocko (2005); Sirocko et al. (2005, 2009); Dietrich (2011) and Dietrich and Sirocko (2011) developed a ^{14}C-based age model for cores DE2/DE3 from the Dehner Maar, which provided the backbone of the stratigraphy of all ELSA cores presented by Sirocko et al. (2013). This work included two ^{137}Cs and ^{210}Pb profiles, 270 ^{14}C-dates, two sediment records constrained with by a paleomagnetism dataset, four luminescence dates and two Ar/Ar dates and shows that the sediment records used in this study (HM1, DE3, AU2) indeed document MIS 1–4.

Unfortunately, radiocarbon ages can usually not be used to date the Eifel maar core sections precisely. Sirocko et al. (2013) showed in a comparison of varve counting with the numerous ^{14}C ages obtained from the Ulmener Maar that the plant macrofossils in the Holocene sediments consists mainly of reworked pieces of wood, seeds and even small twigs. Twigs (one year old branches) proved to be the optimal material for dating the sediments because they decay fast if not rapidly buried in the sediment. The same observation was made for the Holzmaar (Fig. 17), where all Holocene ^{14}C ages were much too old for the varve counted age model, which is, however, consistent with the palynostratigraphy.

Bulk sediment ^{14}C ages are even less reliable in the Eifel maar lakes because the anoxic bottom waters contain abundant old CO_2 from remineralisation of old organic matter in the sediments. In addition, outgassing of mantle CO_2 occurs even today at many locations in the Eifel (May, 2002) and questions the applicability of ^{14}C dating for Eifel maar sediments in general.

The Holocene ^{14}C ages are thus problematic and the same is expected for the MIS2/3 radiocarbon ages. Indeed, all ^{14}C ages presented in Fig. 17 are always older than the tephra-based age model. These ^{14}C ages thus only show that wood, seeds and bulk sediment must be younger than 55,000 b2k. Accordingly, we document the ^{14}C ages only to support other evidence for MIS3 affinity for these sediments, and rely on ice core tuning and tephra correlation to date the entire ELSA sediments precisely.

The main sediment record discussed here (AU2) was not included in Sirocko et al. (2013), because it was drilled in 2013. Core AU2 clearly archives the marker tephra horizons labelled as Laacher See Tephra (LST), unknown tephra 1 (UT1), Wartgesberg Tephra (WBT), Dreiser Weiher Tephra (DWT) and unknown tephra 2 (UT2), as well as the dominance of *Picea* during early MIS 3 (Figs. 16–19). Several kilograms of sediment from core AU2 were sieved from the depths corresponding to the MIS 2 to obtain several mg of pure Ranunculaceae seeds for dating; this resulted in four new ^{14}C dates constraining a short phase of tundra environment during the GI-2, but seeds were apparently eroded and redeposited also in the LGM and late glacial sediments (Fig. 19).

4.1.4. Greenland ice core tuning

The most important record for the entire ELSA stratigraphy is AU2 with a length of 123 m (Fig. 4). The average sedimentation rate is 2 mm/year and, accordingly each core meter represents about 500 years in the pollen and plant macrofossil records. AU2 has the highest abundance of plant macroremains, because the dry maar at Auel is/was fed by a stream. These specific conditions could explain why only AU2 closely resemble all Greenland interstadials as marked increases in the C_{total} record (Fig. 19).

The total carbon content of AU2 was measured at 20 cm-intervals to obtain a record of approximately 100-year resolution, which was then compared to the NGRIP stadial-interstadial succession (Svensson et al., 2008). The similarity of the two records is so consistent that the late and middle MIS 3 record from AU2 allowed continuous direct tuning of AU2 to the Greenland ice core chronology (Fig. 19). The ages for all tephra in AU2 where then applied to all other ELSA cores. Accordingly, the results from all cores presented in this study directly can be compared to the Greenland climate history on the b2k time scale (Svensson et al., 2008).

Maxima in AU2 C_{total} curve matche to some extent the diatom spikes in core DE3 (Figs. 13, 19). The diatoms reflect apparently in-lake productivity (e.g. Veres et al., 2008, 2009) and must influence the C_{total} content during MIS3 strongly. The MIS 3 diatom inferred interstadial succession is also visible in the DE3 tree pollen, but only in the absolute numbers of pollen counts (Fig. 7), not in the relative percentage values (Fig. 6).

The early MIS 3 linking is, however, still somewhat problematic because we observe a pronounced increase in total carbon of AU2 at 49,000 b2k (Fig. 19). Most likely, a dramatic change had occurred in the Eifel landscape at that time, but a respective signal in the Greenland ice is not apparent.

4.1.5. Anchoring the tuned Auel time series to speleothem dates

High-resolution U/Th dating of stalagmites in Europe (Moseley et al., 2014), and even nearby the Eifel (Bunkercave, D. Scholz, pers.comm.) indicate that stalagmites grew in central Europe from 60(55)000 b2k to 46,000 b2k, which we use to anchor the last occurrence of *Picea* in the ice core tuned AU2 record. The section below with abundant *Picea* is then tuned to the GI-12 to GI-17 succession (Fig. 19). This tuning places the beginning of the *Picea* Zone at 60,000 b2k, almost exactly at the MIS 3–4 transition.

4.2. The ELSA vegetation stack

Based on the above stratigraphy and proxies the Pleistocene record of AU2 (Auel) and DE3 (Dehner) are stacked together with the Holocene records of SMfreeze (Schalkenmehren) and HM1 (Holzmaar) in order to construct the ELSA vegetation stack spanning the entire last 60,000 years.

4.2.1. Late Holocene (0–6000 b2k)

The uppermost part of the Holocene pollen record shows the spruce maximum of the Prussian afforestation after 1820 CE (well visible in the plot of *Picea*) and the medieval forest clearings back to the early

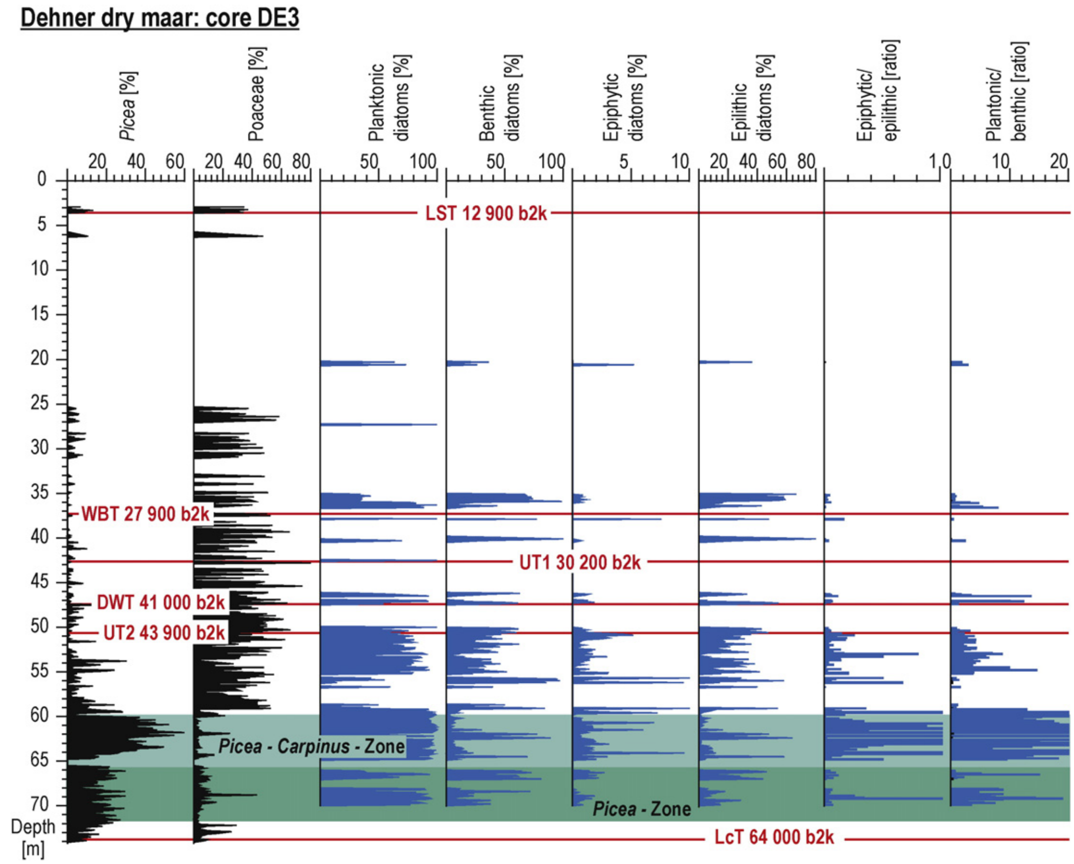

Fig. 13. Diatoms of core DE3 from Dehner dry maar versus depth.

medieval dense hornbeam and beech forest that re-established after the retreat of the Romans (Fig. 5). The vegetation cover in the Eifel must have been similar during the Roman times and earlier Celtic times starting at the beginning of the Iron Age at 800 BCE when a pronounced increase of cereals and grass pollen indicating further opening of the Bronze Age landscape. The Bronze Age landscape change had started in the Eifel at 1800 BCE (Kubitz, 2000) indicated by the spread of beech forests on used and abandoned farmland (Kalis et al., 2003).

The beginning of the Neolithic in the Eifel is marked by an increase in cereal pollen (Fig. 5), which are barely visible at Holzmaar, but clearly visible in the Ulmener Maar record, where farming had started already by 3700 BCE (Gronenborn and Sirocko, 2009). Prior to the spread of cereals and other grasses, the HM1 record also shows the well-documented elm decline at 4300 to 4000 BCE, synchronous with the spread of alder (Figs. 5, 20). This feature is visible in both the pollen and the plant macroremain records and is also characteristic of many other parts of Europe (Kalis et al., 2003). This transition sees the introduction of cattle into the Eifel forest where elm, lime-tree, and ash-tree declined in abundance because the leaves of these trees were used for fodder production. Thus, from this time on forests were managed by humans for domestic purposes so that the mid-to-late Holocene vegetation is an anthropogenic-modified broadleaf forest within an open landscape dominated by cereal cultivation and cattle farming.

4.2.2. Early Holocene (6000–10,500 b2k)

The lower elevations of the Eifel region prior to 4300 BCE were covered with oak-lime woods, whilst higher altitudes were covered with oak-elm woods and hazel (Kalis et al., 2003). Neolithic settlers did not inhabit the Eifel at this time, and the plant macroremain assemblage is dominated by fruits and seeds of aquatic plants and taxa specific of flood-plain forests (Herbig and Sirocko, 2012). The representative taxa from reed/littoral vegetation are *Scirpus lacustris*, *Typha* sp., *Eupatorium cannabinum* and *Lycopus europaeus* (Fig. 10).

A cooling pulse from the latest meltwater discharge into the North Atlantic occurred at 6200 BCE (8200 b2k). This cold anomaly lasted for about 100 years and is visible in the sediments of Lake Holzmaar and Meerfelder Maar (Prasad and Baier, 2014). Related changes in absolute temperature and precipitation cannot have been very large, because we do not see a corresponding change in the forest vegetation (Figs. 5, 20), just a slight increase in grass and pine pollen. Accordingly, a mixed oak forest with significant hazel dominated the Eifel landscape during the first millennia of the Holocene.

Hazel was the first temperate tree to spread in the late glacial landscape and constituted up to 80% of LEZ 2 pollen as well as found in the early Holocene plant macrofossil records (Fig. 10). The foliage of the birch and hazel trees must have been an important nutrient source for the brown earth soils that started to develop during this period. Increasing soil fertility and clay mineral content provided ideal conditions for the spread of oak, elm, and other deciduous trees during the millennia encompassed by LEZ 2.

4.2.3. Late glacial/early Holocene boreal forest (10,500–14,700 b2k)

The vegetation after the end of the Younger Dryas at 11,700 b2k was still dominated by pine and birch (Fig. 20), but reveal a significant presence of aquatic plant macrofossils, in particular *Nymphaea alba*, *Najas marina* and *Potamogeton* sp. (Figs. 10, 11, 12).

The Younger Dryas cold spell is characterized by subarctic steppe tundra with heliophytes (*Artemisia*, *Helianthemum*), shrubs including *Juniperus*, *Betula nana* and sporadic tree birches. Birch and pine are present (Figs. 5, 20), but most likely as dwarf forms and in patches on favorable stands such as south exposed slopes, river valleys etc. The occurrence of *Nymphaea alba* indicates a shift to more eutrophic condition in the Holzmaar during the late Glacial/Holocene transition.

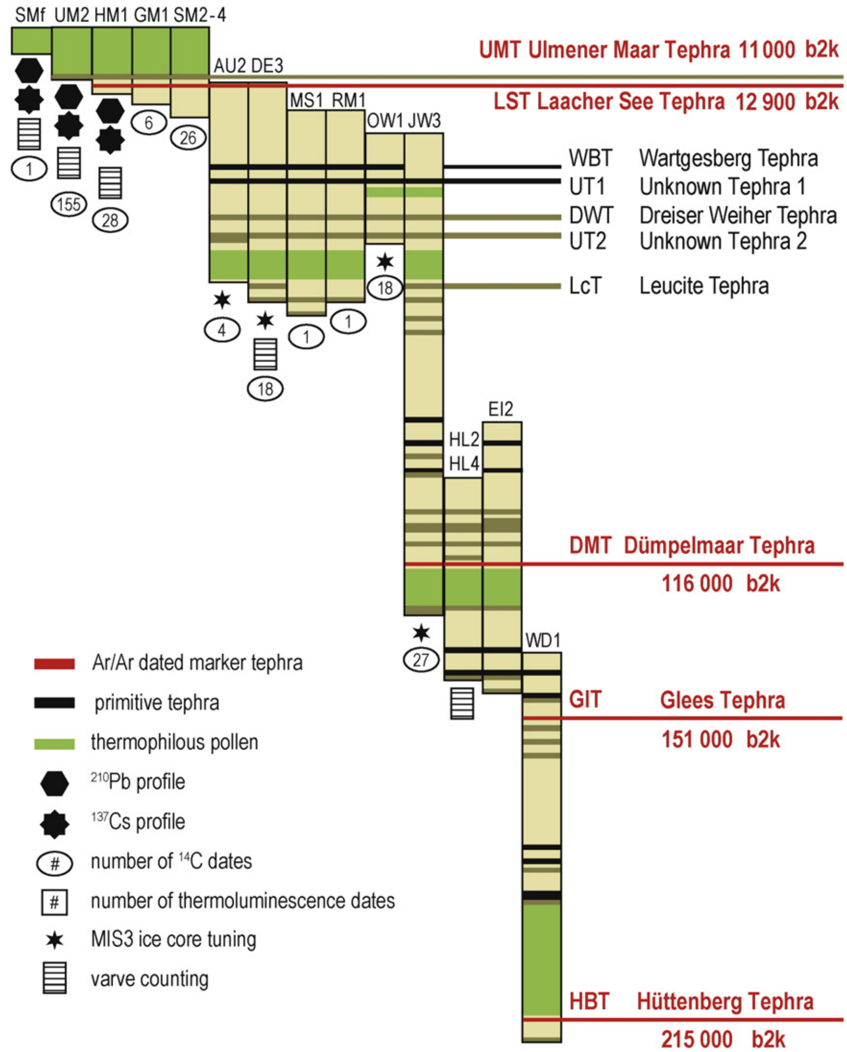

Fig. 14. Overview of main ELSA cores covering MIS 1–7. Marker layers from the ELSA-Tephra Stack are used to correlate the ^{14}C dated sediment cores and evolved marker tephra are highlighted in red with Ar/Ar ages taken from van den Bogaard (1995) and van den Bogaard and Schmincke (1985). The detailed description of the ELSA Tephra Stack over the entire last 500,000 years is given by Förster and Sirocko, 2016–in this volume.

Remains of aquatic plants like *Ranunculus aquatilis* or *Najas marina*, as well as single diaspores from herbaceous plants and *Typha* sp., are typical around the time of Laacher See ash emplacement: a 10-cm thick tephra layer in core HM1/DE3/AU2 deposited during late Alleröd times (Figs. 2, 3 and 4). Steppic conditions with scattered birch and pine covered the landscape during the preceeding Bölling (Fig. 20). Tundra dominated by grass was established in the very early Bölling when a few trees, i.e. *Salix* and *Populus*, grew in small stands around the maar. Macroremains of the Bölling reveal abundant oospores from Characeae (*Chara aspera*, *Chara globularis*, *Chara contraria*) as underwater vegetation in particular during the deglacial tundra (Figs. 10–12), when ostracods were most abundant in the Eifel maar lakes.

The plant macrofossil record points to birch trees scattered in grassland with abundant heliophytes (Figs. 10–12, 20). The lake water must have been strongly oligotrophic with *Menyanthes trifoliata*, *Najas marina*, Potamogetaceae and Characeae, while the swamp vegetation in the vicinity comprise *Filipendula ulmaria*, *Isolepis secaceae*, *Juncus* sp., *Carex* sp. *Eupatorium cannabinum* and *Typha* cf. *latifolia* (Figs. 10–12).

The drastic increase in ostracods and oospores from Characeae during a time of already existing shrubs and pine/birch dwarf vegetation most likely parallels the abrupt warming at around 14,700 b2k when the Gulf Stream/North Atlantic drift system jumped within several decades into an interglacial mode. Accordingly, the lake water was warm, but nutrients and biomass were still low. The development of vegetation during this part of Termination I was studied at high resolution by Stebich (1999); Litt and Stebich (1999) and Litt et al. (2001, 2003) and we refer the reader to these papers for a detailed description of the classical pollen zonation.

Seven *Chara* taxa have been identified in the ELSA maar lake records during the Termination I; *Chara aspera*, *C. globularis*, *C. contraria*, *C. vulgaris*, *Nitella capi llaris*, *Nomada opaca* and *Tolypella glomerata* indicate a continuous warming of the lake, but without eutrophication.

4.2.4. The polar desert of the Last Glacial Maximum (23,000–14,700 years b2k)

The maar lake sediments spanning 23,000 to 14,700 b2k are in general problematic for paleoecological reconstructions because they show poor pollen preservation in all cores. The only core with glacial stage pollen preserved is from Schalkenmehrener Maar in which we observe a few grass pollen after ca. 16,000 b2k (Sirocko et al., 2013). Accordingly, we have inferred for that time the existence of a landscape with permafrost, but some grassland. Characea oogones and ostracod give even further evidence for more biomass after 16,000 b2k, which might have been important for the late glacial megafauna like the mammoths and the first hunters moving from the glacial refuge areas in southern France back into central Europe.

Pollen is completely absent during the LGM from 21,000–17,000 years b2k but this might also be a result of poor pollen

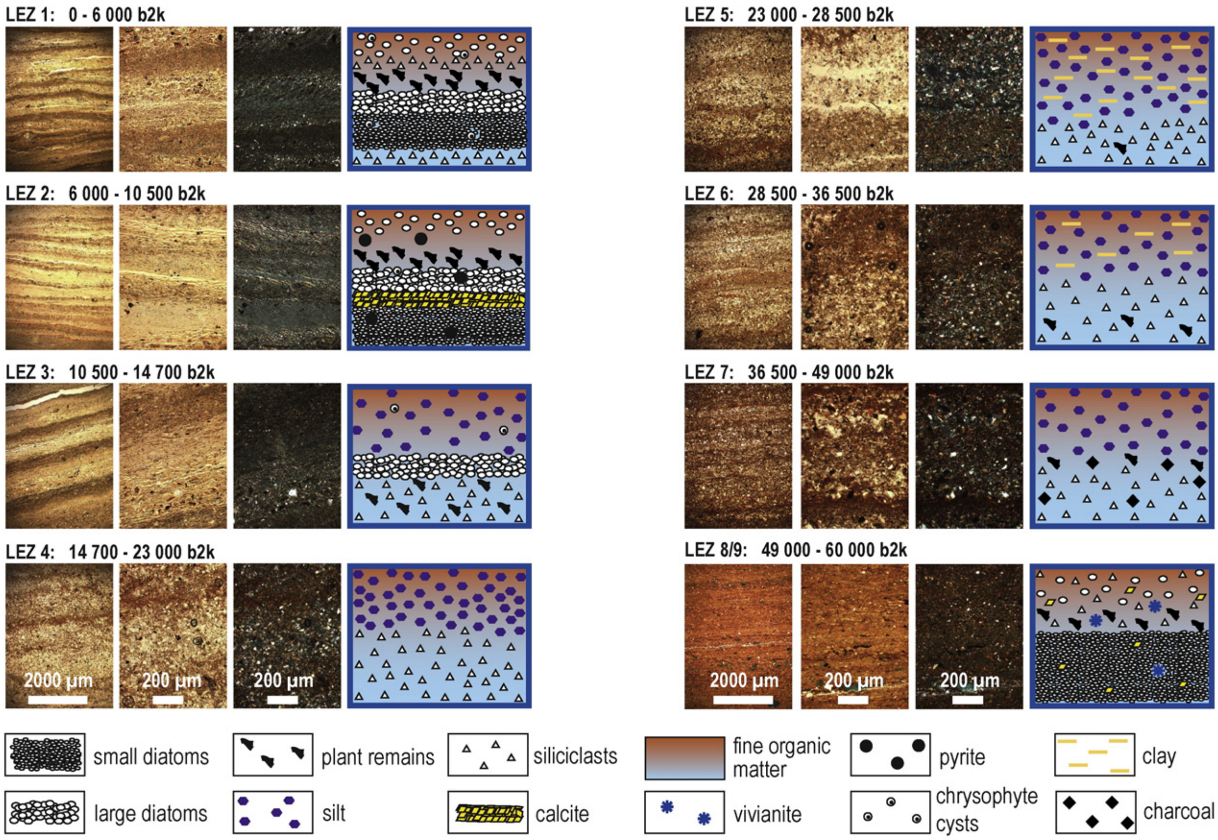

Fig. 15. Schematic diagrams of annual varve composition during the last 60,000 years as visible on petrographic thin sections.

preservation. Oxic conditions must have reached the deepest part of the maar lake even on cold summer nights whereby only coarse plant macroremains would be capable of surviving the decay processes in the well-oxygenated glacial stage water columns (Veres et al., 2008, 2009). Accordingly, the LGM plant macroremains consist only of a few mosses, oogonia, ostracods, as well as some single seeds mainly from Ranunculaceae (Figs. 10–12). The LGM polar desert was thus likely not completely abiotic with at least some biomass present during the summer.

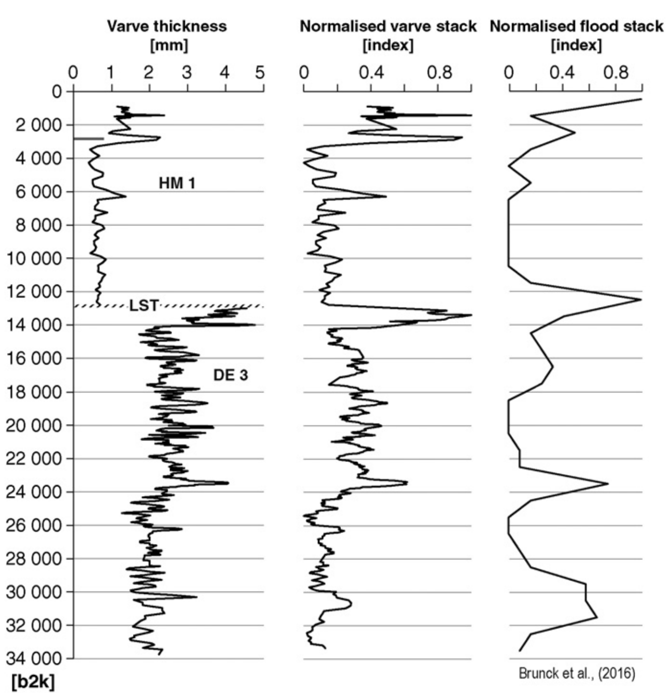

Fig. 16. Varve thickness and normalised varve stack in comparison with normalised flood stack (modified after Brunck et al., 2016–in this volume).

The occurrence of Ranunculaceae seeds and a diatoms spike at 21,000 b2k provides evidence that the LGM polar desert was punctuated with short-lived returns to tundra conditions (Figs. 20, 21).

4.2.5. The tundra of the early MIS2 (28,500–23,000 b2k)

Moss, fungi sclerotia (*Coenococcum geophilum*) and insect remains are typical for the millennia before the LGM (Fig. 20) and indicate a tundra vegetation. The ^{14}C dating of this time is complicated because there is generally very little organic material, except for some Ranunculaceae, which reach from the steppe into the tundra phase.

The other typical plant macrofossils of the tundra reflect riparian and damp ground taxa i.e. *Stellaria aquatica*, *Silene flos-cuculi*, *Juncus* sp. and Cyperaceae as well as Caryophyllaceae, Poaceae, Brassicaceae and Asteraceae. Insects and sclerotia from *Coenococcum geophilum* (indicating the presence of fungi) (Figs. 20, 21). The peak of *Coenococcum geophilum* might indicate higher sediment/soil erosion and runoff into the lake (Drescher-Schneider, 2008).

4.2.6. The steppe of the middle and late MIS 3 (36,500–28,500 years b2k)

Many sediment samples spanning the middle and late MIS 3 did not reach the minimum threshold of 50 pollen grains so that calculation of pollen percentage values was not possible and these data points were omitted from the pollen diagrams (Fig. 20). The pollen concentration curves show no structure during all of MIS 3, but the interstadials of MIS 3 become clearly visible in the pollen count numbers (Figs. 7, 18, 20, 21). The same taxa observed during the cold stadials were apparently also present during the warm interstadials, however with higher pollen numbers. This could be an effect from better pollen preservation in the carbon-rich interstadial sediments, but could alternatively be also

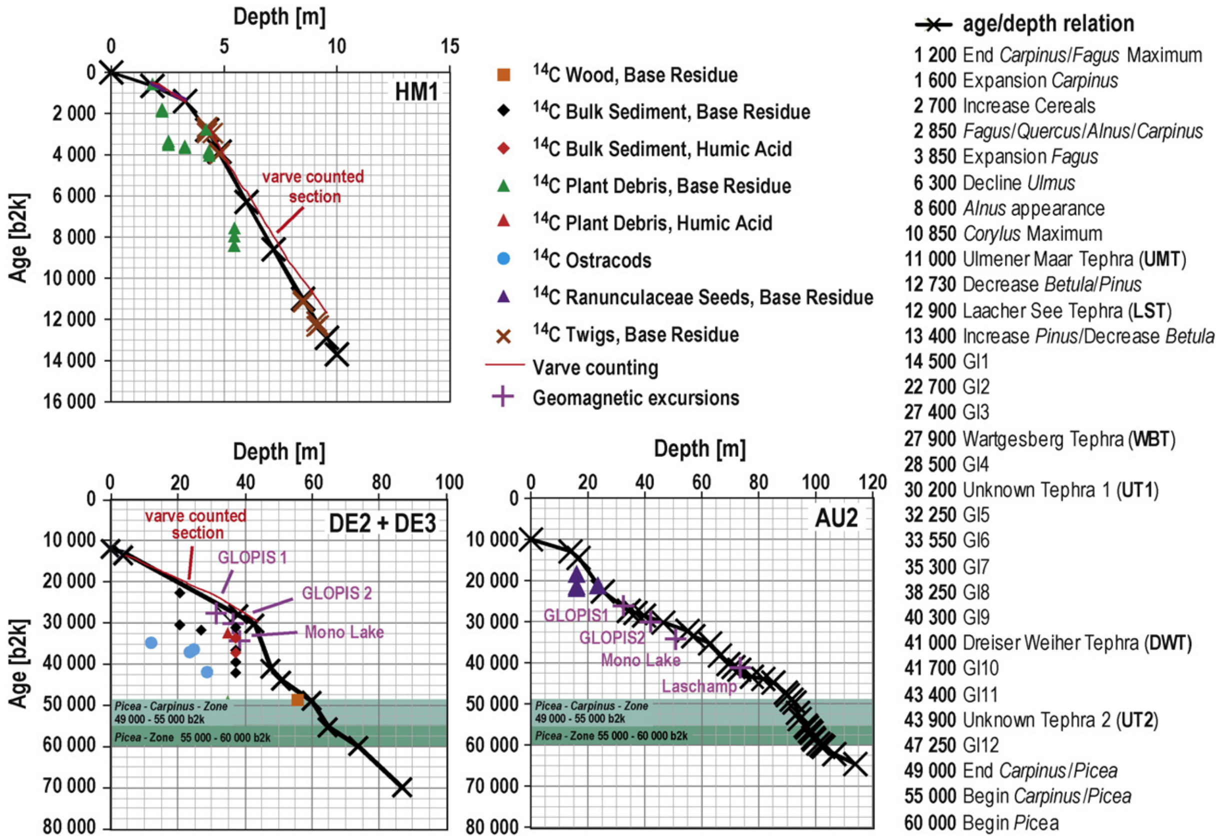

Fig. 17. Age/depth relations for cores AU2 (Auel dry maar), DE2/DE3 (Dehner dry maar) and HM1 (Holzmaar lake) with ^{14}C dates and the *Picea-Carpinus*-pollen-zones discussed in the main text.

explained by general higher plant numbers in the interstadial times (Fig. 21). In this case, the steppe persisted with the same taxa during all of the middle and late MIS3, but the abundance of trees and diatoms increased significantly during the interstadials.

Plant macrofossils are generally scarce and can only be found in Auel Maar Lake where even the oospores of Characeae that are indicative of in-lake vegetation were often missing. Biomass was generally low — with the exception of *Coenococcum* sclerotia, especially during the late MIS 3 (Figs. 20, 21). The sclerotia from *Coenococcum geophilum* coincide with occurrence of *Betula* seeds and may be indicative of *Alnus* and *Betula* peat along the banks of the lakes (Grosse-Brauckmann, 1974). Beside the remains of *Betula* several wetland taxa were found, including *Stellaria aquatica*, *Silene flos-cuculi*, *Lysimachia* sp., Cyperaceae, *Carex* sp., *Potamogeton* sp., and *Typha* sp. (Fig. 11, Supplements 4, 5). It must be emphasized that *Coenococcum geophilum* is not a climate indicator as it is connected via mycorrhiza symbiosis with several plant taxa and thus not suitable for vegetation reconstruction (Tinner et al., 2005). However, during this time of low biomass we suppose that it indicates cold, moist conditions with abundant *Cristatella mucedo*. The landscape from 36,500 to 28,500 b2k is thus a steppe with scattered trees. Charcoal is almost absent suggesting that it was moist enough to prevent intensive burning of the grassland.

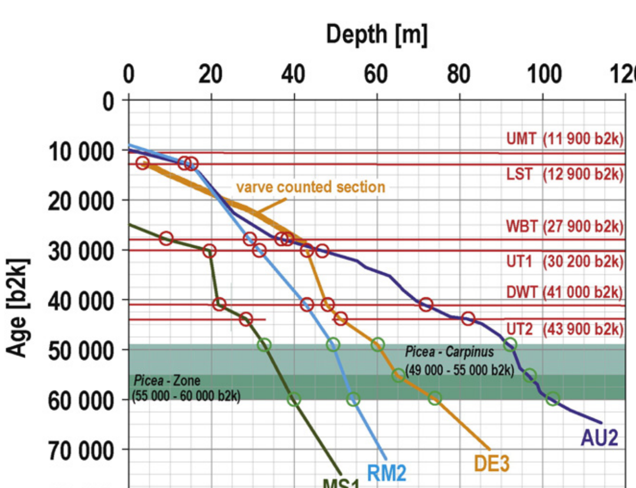

Fig. 18. Age/depth relations for cores MS1 (Merscheider dry maar), RM2 (Rother Maar), DE3 (Dehner dry maar) and AU2 (Auel dry maar) with detected tephra marker layers (after Förster and Sirocko, 2016–in this volume) and the *Picea* dominated pollen zones. Cores RM2 and MS1 have only low resolution pollen profiles and are not included in the detailed paleobotanic analysis discussed in this paper.

The dominant pollen are grass associated with abundant heliophyte macroremains such as *Papaver rhoeas*, *Papaver argemone*, *Cerastium* sp., *Chenopodium* sp. and *Erica* sp. (Figs. 6, 11, 12). Alder must have been growing close to the maar lake because their macroremains were found in the sediments. Fish remains and water plants were recorded for the last time during GI-8.

4.2.7. The boreal forest of the middle MIS 3 — (46,000–36,500 years b2k)

The period from 46,000 to 40,000 years b2k is marked by birch, pine and a strong increase in charcoal, which indicates that the boreal forest burned regularly, pointing to at least seasonal arid conditions during that interval (Figs. 20-22).

Three charcoal maxima (Fig. 20) are parallelized by synchronous rises in Characeae oogonia (indicative of shallow water) and Cyperaceae (shoreline vegetation) remains. In addition, several maxima in wood remains are visible, but not during the time of the charcoal maxima (Fig. 21). We cannot address these maxima of the forest remains to a

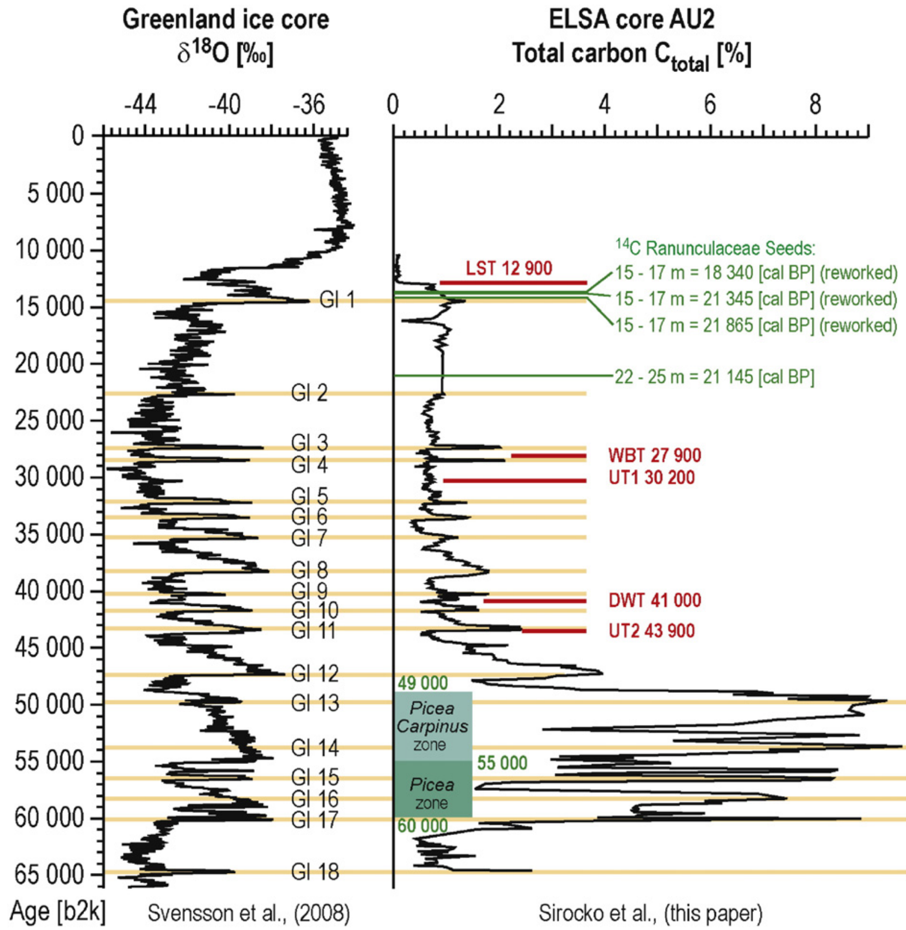

Fig. 19. AU2 ^{14}C dates on Ranunculaceae seeds and time series of total carbon concentrations tuned to the Greenland ice core stadial/interstadial events as evidenced by variations in oxygen isotopes (Svensson et al., 2008).

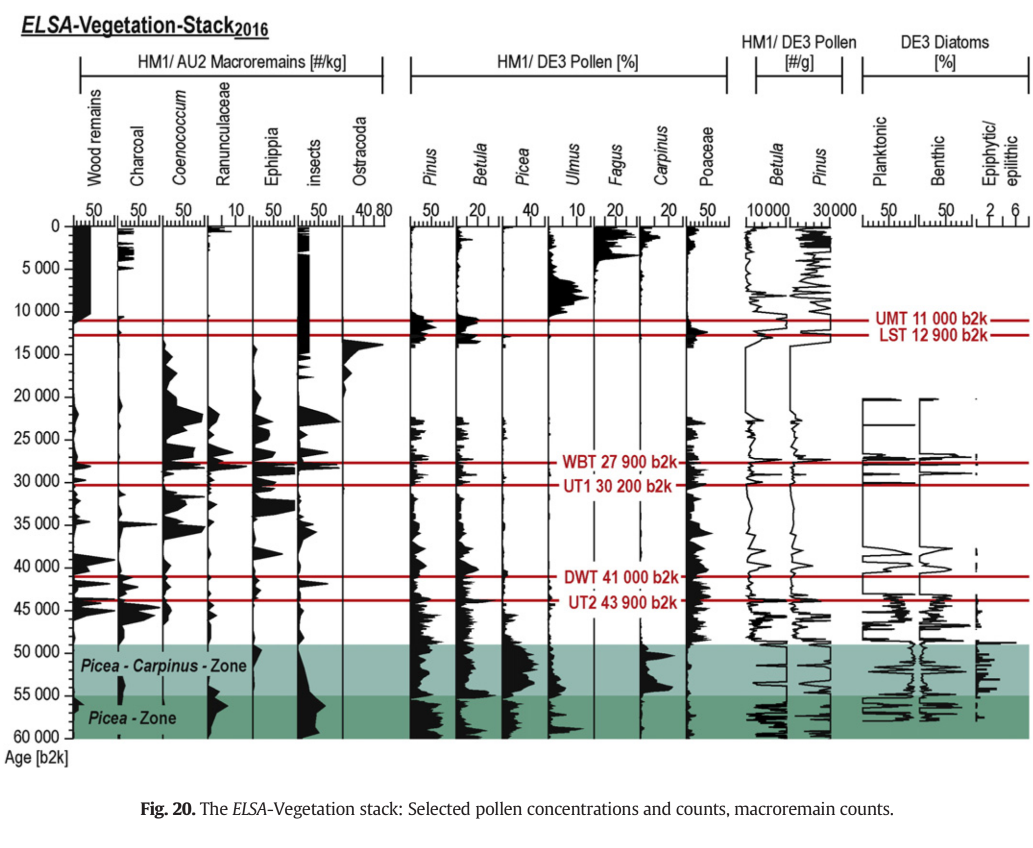

Fig. 20. The *ELSA*-Vegetation stack: Selected pollen concentrations and counts, macroremain counts.

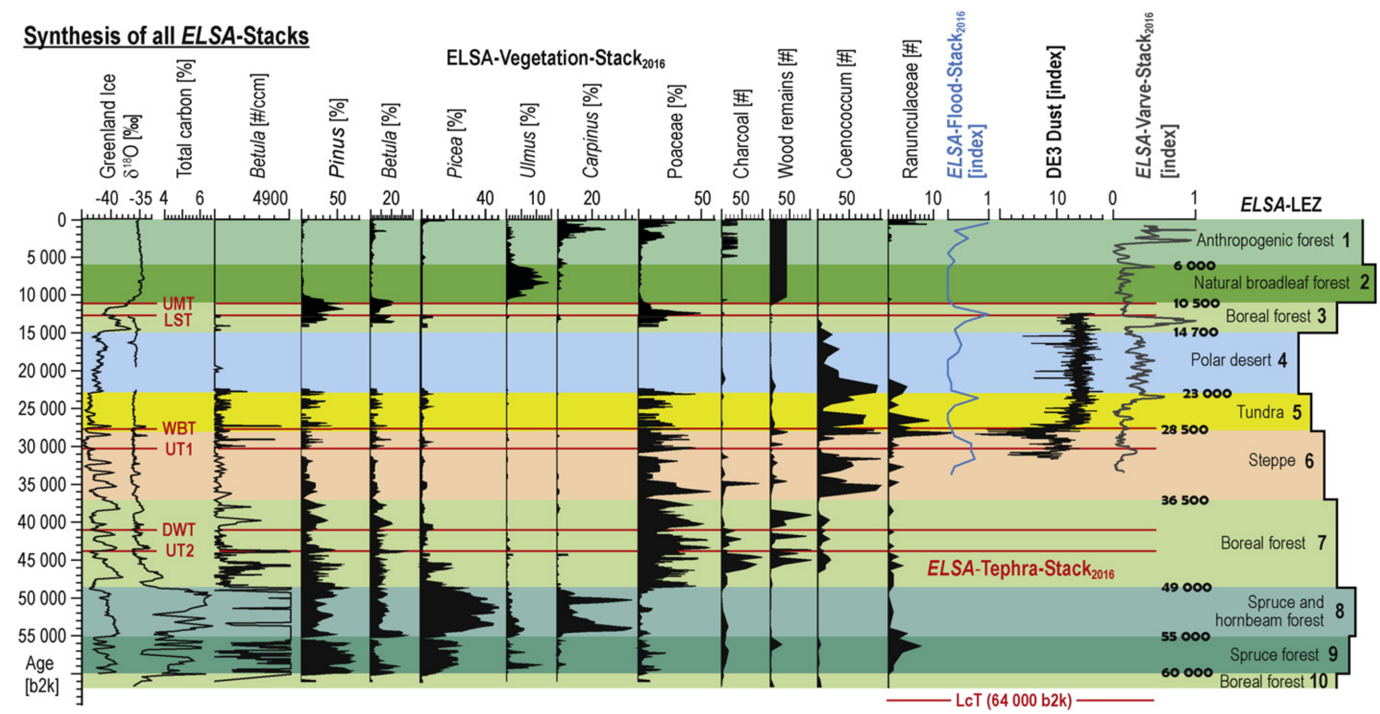

Fig. 21. Synthesis of all *ELSA* stacks: Total carbon of core AU2, selected pollen from core DE3, varve thickness stack for cores HM1 and DE3, updated part of the DE3 dust record (Seelos et al., 2009) and the ELSA-Tephra-Stack (Förster and Sirocko, 2016–in this volume) together with Greenland Ice isotope temperature index (Svensson et al., 2008) and with LEZ numbers.

specific climatic phase because the resolution of the macroremain sampling of AU2 is for entire core meters, avering about 500 years, whereas the C_{total} record is of 20 cm resolution, representing 100 years. It will have to wait for a future analysis of thin sections for this time interval to arrive at clear inferences on the nature and timing of the charcoal and wood remain maxima relative to the stadials and interstadials.

The occurrence of *Ceratophyllum demersum*, *Schoenoplectus lacustris*, *Thypha* sp. and *Najas marina* in the time of the early charcoal maxima (Fig. 21 and Supplements 2, 4) point to mean July temperature between 15 and 18 °C during late interstadial phase of GI-12 at 46,000–45,000 years b2k (Fig. 21). Moreover, the presence of *Urtica dioica* indicates nutrient rich soil/sediment conditions between 46,000 and 43,000 years b2k.

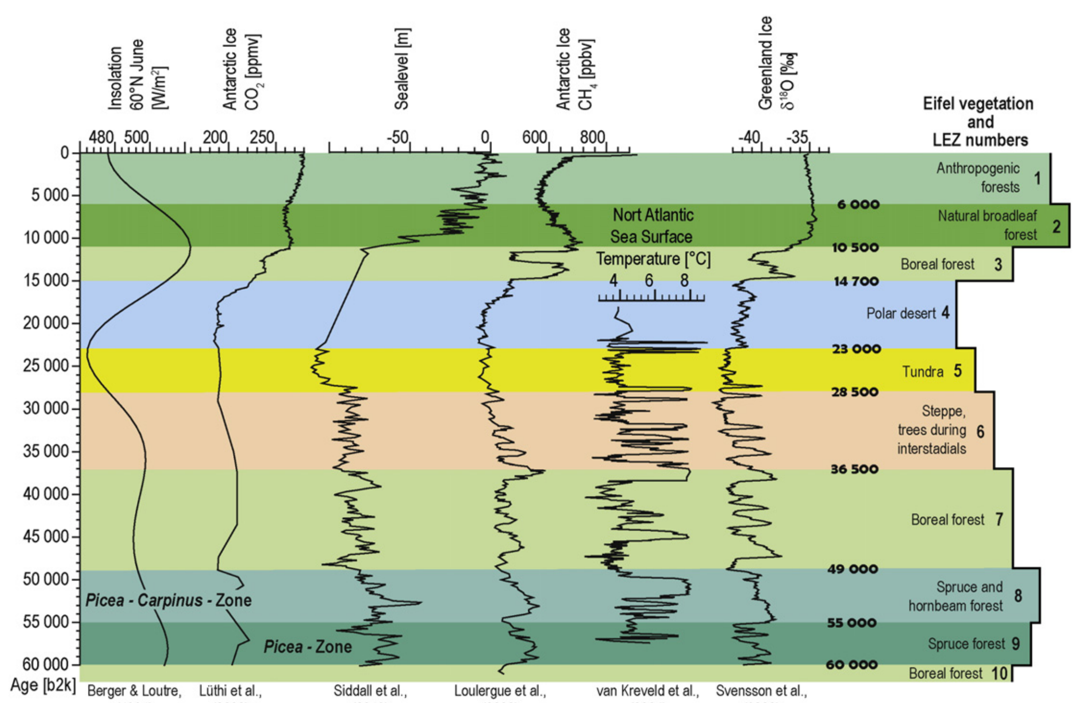

Fig. 22. Time series of the major forcings for the central European climate during the last 60,000 years: summer insolation at 65°N (Berger and Loutre, 1991), Greenland Ice isotope temperature index (Svensson et al., 2008), the ice-core derived global records of CO_2 (Lüthi et al., 2008) and CH_4 (Loulergue et al., 2008) on the AICC2012 chronology (Veres et al., 2013), sea surface temperatures (van Kreveld et al., 2004) from the North Atlantic Drift.

The first traces of the subsequent steppic conditions become visible in the boreal forest accompanied by Cyperaceae, insects, Ranunculaceae, *Papaver argemone* and the first appearance of *Coenococcum sclerotia* (Figs. 20, 21). Apparently, the landscape was much more open at this time than during the very early part of MIS 3.

4.2.8. The spruce forest of the early MIS 3 — (60,000–49,000 years b2k)

All maar records studied in this synthesis on the MIS 3 vegetation in the Eifel reveal a phase of several thousand years during the early MIS 3 with abundant tree pollen. Both pollen and plant macroremains show the omnipresent *Pinus* and *Betula* (Figs. 20, 21), but with the dominance of a long-lasting peak of *Picea* that is just beyond the limit of ^{14}C dating and thus assigned to GI-17–GI-13 (60,000–49,000 years b2k). Other thermophilous trees including *Ulmus*, *Quercus*, *Tilia* are also represented in the pollen assemblages. Climatologically, these broadleaf taxa are indicative of summer temperatures slightly lower than modern temperatures, even if present only in small quantities. The warm phase had a duration of about 7000 years and favoured soil formation. The pronounced *Picea* pollen maximum is observed not only in AU2 and DE3 (Figs. 20, 21), but in five other ELSA records not included in this study, but shown in the age-depth relations of Fig. 18.

The macrofossil evidence corroborates the findings from the pollen record, because *Picea* sp. and *Abies alba* needles have been found at the same depth as the respective pollen maxima. The macroremain record is further supported by the presence of seeds and fruits of *Rubus idaeus*, *Rubus* sp., *Alnus* sp. and *Urtica dioica*, which may point to nutrient rich soil/sediment.

Small amounts of lime tree pollen are present that might be explained by long distance transport from the Mosel-valley (see also discussion) where average summer temperatures are today 2° warmer than in the Eifel. We observed seeds of *Ceratophyllum demersum* at 86.5 m depth in core AU2 (Fig. 12, Supplement 4). This plant needs summer temperatures of at least 17 °C so that it appears that the early MIS 3 growing season temperatures in the Eifel area were little lower than today. The occurrence of *Typha* sp. (13–15 °C) and several seeds of *Najas* support this paleoclimatologic inference.

5. Discussion

5.1. Landscape evolution zones (LEZ) during the last 60,000 years

Here we present LEZ-based summaries of paleoclimatic and -ecological evolution of the Eifel region during the last 60,000 years (Fig. 21). The ordering starts at the present with LEZ 1 and finishes at LEZ 10 in the very early MIS 3 (Table 1).

5.1.1. LEZ 1 0–6000 years b2k

LEZ 1 is the time when humans altered the early Holocene forests containing elements from cereal and cattle farming. The timing of such environment changes is regionally very different depending on the timing of the introduction of foliage fodder production, soil dependent intensity of cereal farming, and urbanization.

Consequently, the varved lake sediment of the last 6000 years have very different character to that produced in earlier LEZ's, mostly consisting of a dark minerogenic autumn and winter deposition and a bright spring and summer diatom influx layer (Fig. 15). Eolian quartz particles are uncommon, but occurred sporadically during the last few centuries. The varves in the maar sediments revealed abundant spring and summer diatom blooms, sometimes with calcite, and plant leaves are common. The winter layers consist of minerogenic detritus with high proportions of chrysophyte cysts. LEZ 1 flood layers during are most common during Medieval and Roman times (Fig. 21), with the major events at 1342 CE, 800 BCE, and 2100 BCE, respectively (Brunck et al., 2016–in this volume). Volcanic activity was absent from the Eifel during LEZ 1.

5.1.2. LEZ 2 6000–10,500 b2k

The late glacial boreal forests transformed quickly into an early Holocene forest, which was soon dominated by hazel, oak, elm and lime (Fig. 20). The cause of this typical forest succession was not only the higher insolation during the early Holocene, but also the development of the soil profile. This can be attributed to the 6 °C inferred increase of North Atlantic sea surface temperatures when the Gulf Stream/ North Atlantic drift system warmed abruptly at 11,600 b2k (e.g. Alley et al., 1995; Knorr and Lohmann, 2007).

The temperature in central Europe was high from 11,500–6000 b2k, which is best indicated by very high snowlines in the Alps (Nicolussi and Patzelt, 2006). The middle part of LEZ 2 displays a cold spell associated with the 8.2 ka event, a cool pulse from the final meltwater discharges from the late North American ice sheet into the North Atlantic. This cold anomaly affected all of central Europe (von Grafenstein et al., 1998; Hošek et al., 2014) and caused a 120-year long cool and dry signal in the sediments of the Holzmaar (Prasad and Baier, 2014). However, the 8.2 ka event did not change the Eifel landscape ecology in the long term (Fig. 5) as mixed oak forest with hazel covered the Eifel landscape continuously from ca. 10,500 to 6000 b2k.

The varves corresponding to the LEZ 2 sediments are composed of three layers. Spring layers consist mainly of calcite, followed by the summer layer of diatom frustules. The autumn/winter layers consist mainly of detritus and chrysophyte cysts (Fig. 15). Flood layers are not as clearly visible in the Holzmaar or in the Schalkenmehrener Maar as during the later LEZ 1 (after initiation of crop cultivation), but visible flood layers are regularly observed in the LEZ 2 record from Ulmen (Sirocko, 2009; Gronenborn and Sirocko, 2009). The eruption of the Ulmener Maar occurred at the LEZ 3–2 boundary, but we regard the youngest eruption of the Eifel volcanic fields as part of LEZ 3.

5.1.3. LEZ 3 15,000–10,500 b2k late glacial/early Holocene boreal forest

Summer insolation over the Northern Hemisphere increased at 17,000 b2k, when global sea level started to rise coeval with the initiation of post-LGM temperature increase around the Antarctic (Jouzel

Table 1
Top and base for LEZ 1–10 in all ELSA cores.

	Age	HM1		SM3		AU2		RM2		DE3		MS1	
	[b2k]	Top [m]	Base[m]	Top [m]	Base [m]	Top [m]	Base [m]	Top [m]	Base [m]	Top [m]	Base [m]	Top [m]	Base [m]
LEZ 1	6000	0.00	5.85	0.00	4.51								
LEZ 2	10,500		8.24		5.78	0.00	2.39	0.00	5.77				
LEZ 3	14,700					2.39	16.96	5.77	16.67	0.00	7.60		
LEZ 4	23,000					16.96	23.49	16.67	22.52	7.60	22.04		
LEZ 5	28,500					23.49	39.75	22.52	29.56	22.04	39.16	0.00	11.54
LEZ 6	36,500					39.75	64.33	29.56	37.97	39.16	45.65	11.54	20.61
LEZ 7	49,000					64.33	91.50	37.97	49.10	45.65	59.90	20.61	32.50
LEZ 8	55,000					91.50	96.30	49.10	51.80	59.90	64.65	32.50	36.50
LEZ 9	60,000					96.30	101.95	51.80	53.90	64.65	73.90	36.50	39.50
LEZ 10						101.95		53.90		73.90		39.50	

et al., 1994) (Fig. 22). The AU2 and DE3 cores reveal thick layers of laminated, but disturbed sediment from 16,000–14,000 b2k, which we associate with first melting of the LGM permafrost landscape. This initial summer warming resulted in the first deep thawing of the soils, which were soliflucted even on very gentle slopes at several cm/m per year. However, the summers were warm enough so that grasses spread, followed by birch and pine after 14,700 b2k.

Eolian activity was significant well into the deglacial phase, but not as strong as during the preceding LGM (Fig. 21). The drastic increase in ostracods and oospores from Characeae during a time of dominant shrub and pine/birch dwarf vegetation, most likely parallels the abrupt warming at around 14,700 b2k, when the Gulf Stream/North Atlantic drift system jumped to an interglacial mode over a few decades. Accordingly, the lake water columns were warm, but nutrient availability and biomass was still low.

The subsequent Alleröd interval saw animals like red deer move into the expanding forests. This environmental change is also reflected in changing human behavior, because the use of the bow and arrow are first documented during this period, thus indicating close range hunting of smaller animals in a closed/open pine and birch forest.

The Netherlands and northern Germany were strongly affected by large-scale cover-sand deposition during the Younger Dryas (Vandenberghe, 1992a, 1992b; Kasse et al., 1995, 2003; van Huissteden et al., 2003). The end of the Younger Dryas came abruptly within several years at around 11,600 b2k and ended the dominance of grass, pine and birch in the late glacial landscape mantled with weakly developed soils (Fig. 21). Subsequently, loess and cover-sand deposits were transformed into fertile brown earth soils during the warmth of the subsequent LEZ 2.

The varves of the late glacial LEZ 3 are composed of three layers (Fig. 15): (1) spring layers consist of minerogenic components and plant residues; (2) diatom frustules form the summer layers, whereas (3) the autumn/winter layer consists of detritus, chrysophyte cysts and diatoms. The plant residues deposited in spring are interpreted as inwash by intense late winter snowmelt in a climate when plant biomass was present, but no consistent leaf fall in autumn existed yet, as indicated by the absence of autumn layers from leaf fall.

LEZ 3 experienced an average of 4.8 flood events per 1000 years and the largest events were documented at ca. 11,700, 12,500 and 13,800 b2k respectively (Brunck et al., 2016–in this volume). 12 strong flood events occurred in the time window 12,000–13,000 b2k (representing most of the Younger Dryas) and reveal the highest number of flood events per 1000 years in the whole of AU2, a situation likely caused by intense winter snow fall and early spring meltwater events.

The highly explosive eruption of the Laacher See occurred at 12,880 b2k in the middle of LEZ 3 followed by the Ulmener Maar eruption at 11,000 b2k. These are the only Eifel-derived tephra identified during the last 27,000 years.

5.1.4. LEZ 4 The polar desert of the Last Glacial Maximum (23,000–15,000 years b2k)

Landscape Evolution Zone 4 denotes the coldest time of the last glaciation, equivalent to the Last Glacial Maximum (LGM). Pollen grains are completely absent and identifiable organic macroremains consist only of few mosses, oogonia and ostracods (Fig. 21). A few Ranunculaceae seeds were found in the sediment, but the ^{14}C ages show that they are reworked from a Ranunculaceae maximum at 21,000 b2k (Fig. 19). Two short warming intervals are apparent in the diatom record (Figs. 13, 20) and by the occurrence of *Thypa* sp. (Figs. 10, 11, 13), which today requires mean July temperatures above 13–15 °C. Oogonia are other indicators of submerged vegetation and reveal that Characeae were abundant even during the LGM, however, we cannot exclude that they are also reworked like the ^{14}C dated Ranunculacea seeds (Fig. 19).

Dust activity was at a maximum during the LGM. The ELSA dust stack shows that silt deflation became strong near 28,000 b2k (Fig. 21) and dominated the sedimentation up to 14,000 b2k, when the vegetation cover slowly increased. The peak LGM dust deposition is even clearer in the varve thickness records, whereby they increase markedly at 23,000 b2k (Fig. 21).

The frontal margin of the Scandinavian ice sheet reached only up to the north of the river Elbe even during the maximum ice advance and thus the ice margin was about 500 km of the Eifel. The glacial maar lake sediments are laminated and yellowish, and consist of almost pure silt and fine sand (no clay or gravel). There is no indication of a permanent ice cover or drying of the lakes (Figs. 3, 4). There are, however, subaqueous terraces at around 14 m below modern lake surface in Holzmaar and Schalkenmehrener Maar, indicating substantial lowering of the ground water and lake level during the LGM. The lakes were open during summer, however on a lower lake level, but we see no indication of strong weather anomalies. The climate of the LGM in the Eifel was apparently uniformly cold with clear air, with most active dust transport during late spring or summer.

The LGM section at Dehner Maar is about 8 m-thick and consists of countable silt laminae. The number of eolian laminae counts matches with a 4% error the number of years for this section as derived from the ice core tuned age model. The nature of the layers are not fully understood yet and could represent continuous background dust activity in all summer months or just one dust storm in the spring of each LGM year. Presently, dust in desert storms often leaves layers of 1 mm to 1 cm thickness on the ground (Sirocko and Raschke, 1993). Accordingly, one spring dust storm would be enough to explain the observed glacial stage eolian lamination record.

Unexpectedly, the maar lake sediments do not document any clear response to North Atlantic Heinrich Event 1 (Fig. 21), which is thought to have caused extreme cold in central Europe, because of the massive iceberg presence in the entire North Atlantic and the resulting complete shutdown of the Meridional Overturning Circulation/Gulf Stream system (Thornalley et al., 2011). In contrast, the LGM conditions in the later part of LEZ 4 must have been quite stable until 14,700 b2k, when the Atlantic warmed abruptly. At this stage we cannot firmly assess whether the first grass pollen spread happened before or during that dramatic climatic change. This time interval is varve counted in DE3 but contains no pollen, whereas the core from Schalkenmehrener Maar clearly shows grass pollen in the glacial sediments, although the stratigraphy is problematic for this core. Archeological evidence from the middle Rhine valley site of Gönnersdorf suggests that grasses grew even before 14,700 b2k (Terberger and Street, 2003; Street et al., 2012), which could place the first grass pollen in core SM3 from the Schalkenmehrener Maar (Sirocko et al., 2013) at 16,000 b2k in the polar desert landscape.

The laminations observed during LEZ 4 consist of only two clear components: bright coarse silt quartz layers and brown layers of fine silt (Fig. 15). Both layer types consist of well sorted quartz silt indicating that both layer types are of eolian origin. Terrestrial plant remains are missing so that there is no clear seasonal variability apparent and these laminations cannot regarded as classical varves, but still they will most certainly represent annual layers because the number of these dust layers match perfectly the time between the Laacher See Tephra and the GI 3, 4. It is most likely that the fine dust layer presents continuous slow deflation during the snow-free summer season, and the coarser thick layer a large spring/summer dust storm.

The late LEZ 4 revealed an average of 1.8 flood events per 1000 years and the biggest events were documented at about 15,300, 16,200 and 17,400 b2k, most probably associated with sporadic snow fall/melt events during the Heinrich 1 interval (Brunck et al., 2016–in this volume). The Last Glacial Maximum itself shows no recognizable flood events (Fig. 21) so that it must have been extremely dry during all seasons of peak glacial conditions.

5.1.5. LEZ 5 the tundra of the early MIS 2 — (28,500–23,000 years b2k)

LEZ 5 represents the transition from the steppe/temperate environments of MIS 3 to the MIS 2 glacial conditions. Pine and birch pollen has occasionally been identified, but only sporadically and at low abundance (Fig. 21). In contrast, Ranunculaceae seeds, moss fragments, fungi sclerotia (*Coenococcum geophilum*) and insect remains are the typical organic macroremains (Fig. 9) and indicate a tundra environment persisted during LEZ 5 (Fig. 21).

The sediment in AU2 shows yellowish eolian silt laminae (typical of the LGM, LEZ4), but in LEZ5 the silt is intercalated with grey clay layers (Fig. 15). Thus, compared to the LGM (LEZ4), runoff was active during LEZ 5; either caused by summer rain, or spring snowmelt. A similar depositional environment with alternation of clay and silt layers is also visible in DE3, which has no river inlet so that runoff caused erosion of the poorly vegetated tuff ring and influx of clay into the maar. LEZ 5 is a time of tundra vegetation and spanned the transition from the forested LEZ 6 to the sparse vegetation cover that dominated LEZ 4. Archaeologically, this time interval corresponds to the later Gravettien, when humans hunted reindeer and mammoth, an observation in accordance with the reconstruction of a tundra environment over the Eifel region.

The sediment laminations produced during LEZ 5 consist of two main components (Fig. 15). Well sorted layers with sand sized minerogenic components are certainly of eolian origin (spring or summer), but the occurrence of clay layers indicates strong snowmelt events. Apparently, winter snowfall was at a maximum during the millennia when the northern continental ice sheet grew larger, but the North Atlantic sea surface temperatures where still high during GI3 and 4 (Fig. 22).

LEZ 5 contains an average of 1.8 flood events per 1000 years with the biggest events documented at 23,500, 23,600 and 24,000 years b2k, i.e. around the time of the Heinrich Event 2 (Fig. 22). The massive tephra layers produced by the Wartgesberg eruption at 27,500 b2k characterizes the LEZ5/6 boundary (Fig. 21).

5.1.6. LEZ 6 the steppe of the middle and late MIS 3 — (36,500–28,500 years b2k)

Eolian dust deposition was active at least during the latter part of LEZ 6, but not at the level of the LGM. The beginning of LEZ6 is clearly associated with the spread of *Coenococcum* at around 36,500 b2k at a time when wood remains became scarce in the cores (Fig. 21). Charcoal is almost absent suggesting that trees were very scattered in the landscape and it was moist enough to prevent the grassland from intensive burning. *Ephippa* and *Coenococcum* sclerotia are typical of the earlier part of LEZ 6, whereas Ranunculaceae and insects are characteristic for the later part of LEZ 6, when trees were almost absent. The landscape at this time must thus have been a steppe with scattered trees and the occasional dust storm during a dry season.

Early LEZ 6 following GI-8 (38,000 b2k) is when the early modern humans spread into central Europe, even if the oldest artifact (Venus statuette from Hohle Stein) is now dated to 43,000 b2k (Conard et al., 2006, Conard and Bolus, 2008; Nigst et al., 2014). These early modern humans hunted horse, reindeer and mammoth with the former taxa indicative of steppe vegetation and mammoth/reindeer of a tundra environment. This hunting practice, documented archaeologically, matches the landscape almost exactly as reconstructed from the ELSA pollen, plant macroremains, and dust activity.

An important marker of the late LEZ 6 is the onset of continuous eolian activity after 30,000 years b2k (Fig. 21), which corresponds to Heinrich Event H3. At this time, the Scandinavian ice sheet must have moved southward to an extent that it affected atmospheric circulation over central Europe (Mangerud et al., 2011). Dietrich and Seelos (2010) presented scenarios of changing wind speed and direction over the Eifel, most probably associated with the position/height of the ice sheet and sea surface temperatures over the North Atlantic.

The classical Dutch MIS 3 records call the GI 3–5 time as the Denekamp interstadial, which saw the last phase of soil formation before the LGM in central Europe (Whittington and Hall, 2002). The GI 3, 4 forest is best observed in core DE2 pollen count data at 33 m (Fig. 7), where pine and birch reach a short-lived maximum, indicating the last forest phase before the climate deterioration toward the tundra environment that dominated LEZ5.

The LEZ 6 steppe environment has seen an average of 4 flood events per 1000 years and the biggest events are at 30,200, 30,800, 31,100 y b2k, i.e. during the transition from GI-5 to GI-4 (Fig. 21). These may reflect intense winter snowmelt events, but also infrequent summer rains cannot be excluded. The large Wartgesberg eruption and another eruption from an unknown site also characterize the late LEZ 6. Thick tephra layers (20 cm) must have covered the sparse steppe vegetation and probably impacted on the development of the vegetation as it changed to a tundra environment after the Wartgesberg eruption (Fig. 21).

The sediment laminations produced during LEZ 6 consist of two or three layers, including minerogenic spring layers of mainly quartz, and autumn/winter layers of fine silt and clay. Some laminations are characterized by orange-brown silt layers, which probably reflect soil formation processes. This observation could match the evidence from the soil profile at Schwalbenberg in the Rhine valley, which revealed for the first time that loess-paleosoils developed during all MIS3 interstadials, especially during early MIS3 (Schirmer, 2012).

5.1.7. LEZ 7 the boreal forest of the middle MIS 3 — (49,000–36,500 years b2k)

LEZ 7 is dominated by grass, but with presence of pine and birch (Fig. 21). Spikes of high charcoal abundance are prominent so that vegetation burning must have impacted on the landscape occasionally. The pine forest must have been still dense enough to burn during times of inferred severe drought. The laminations of LEZ 7 consist of two layers: spring/summer layers of quartz and dark autumn/winter layer of fine detritus (Fig. 15). Charcoal is found in both layers indicating that burning was a common process in the taiga forests.

LEZ 7 displays an average of 0.8 flood event per 1000 years with the biggest events at 44,100, 44,300 and 44,500 b2k (Fig. 21). The flood events are concentrated over a short period of time and must have occurred when the boreal forest deteriorated, i.e. during a time of vegetation and soil change during the first pronounced stadial/interstadial changes.

LEZ 7 experienced the very large eruption of the Dreiser Weiher (DWT) and the deposition of an unknown tephra (UT2) (Figs. 16, 21), both of which comprise of >10 cm-thickness in all ELSA cores. These eruptions occurred in a boreal forest environment and mark the end of wood remains in the maar records. Again, the volcanic activity probably had an impact into the vegetation and can be responsible also for the abundant charcoal in LEZ7.

5.1.8. LEZ 8 The spruce–hornbeam forest of early MIS 3 — (55,000–49,000 years b2k)

The period from 60,000–49,000 b2k is characterized by abundant spruce (*Picea*) (Fig. 21), but this *Picea*-Zone is divided into two phases, the upper one with two maxima of *Carpinus* (*Picea*-*Carpinus*-Zone). *Corylus*, *Ulmus*, *Quercus*, *Fraxinus*, *Tilia*, *Abies* and *Alnus* were also present during this warmest interval of MIS 3. The ice core tuning has assigned the two maxima of *Carpinus* to GI-13 and GI-14 respectively, and indeed, GI 14 is the longest interstadial event of the early MIS 3.

Two pieces of *Picea* wood were ^{14}C dated to ca. 46,000 BP (uncalibrated) (Sirocko et al., 2013) and mark the very end of the zone with abundant *Picea*. A reliable calibration for such an old ^{14}C age is impossible; the ice core tuning suggests an age of 49,000 b2k. The lake level must have lowered significantly at the end of LEZ8 leading to reworking of coarse-grained sediment with wood fragments and their influx into the maar lakes. This strong climatic change is marked by a marked decrease in the C_{total} (Fig. 19), which can at least partly be explained by extremely low sedimentation rates during the entire spruce dominated interval. The very small layer thickness could be a climatic signal, but it could also be caused by a principle change in the drainage pattern, if the maar was not connected to a stream at that time, but a solitary deep crater with anoxic bottom water. The occurrence of the thermophilous pollen however, cannot be explained by the drainage processes, in particular because exactly the same pollen profile is observed in the Dehner Maar, which has not fluvial inflow at all.

The tuning places the initial development of the hornbeam forest at 55,000 b2k, i.e. at the beginning of GI14 (Fig. 21). The recovery of needle leaf fragments in the sediments of the *Picea* zone indicates that also *Abies alba* grew in the catchment of the maar. Consequently, the occurrence of the thermophilous taxa cannot be explained only by long distance transport. However, we cannot exclude the possibility that some of the pollen was derived from the nearby Mosel and Rhone valley (Fig. 1), which are today about 2–3 °C warmer in summer than at the elevation of the Eifel. If the stands in the Mosel, where wine is cultivated today, acted as local refugia during the MIS 4 cold phase, it appears likely that the reappearance of such taxa could have occurred rapidly as soon as temperature and moisture were sufficiently high. Accordingly, the Eifel environment could have been different from other sites in northern and southern Germany where MIS 4 refugia were more distant. In addition, the Eifel is 400 km further south of Oerel in Northern Germany, where *Picea* obviously did not grew, but only *Betula* and *Pinus* (Behre and Lade, 1986). Accordingly, the Eifel region was probably beyond the periglacial rim surrounding the MIS 4 ice sheet and under the influence of warm and moist winds from the southwest (Seelos et al., 2009) — whereas the vegetation at Oerel must have been under the immediate influence of the nearby continental ice sheet.

5.1.9. LEZ 9 the spruce forest of the very early MIS 3 (60,000–55,000 b2k)

GI-18 to GI-15 period in the maar lake records are also dominated by *Picea* with some *Ulmus* and *Quercus*, but in lower abundance (Figs. 20, 21). The vegetation is thus similar to LEZ 8, but the absence of *Carpinus* is significant enough to delimit this interval as an individual LEZ. The begin of LEZ 9 at 60,000 b2k matches when stalagmites started to grow again in the Spannagel cave in the Alps, indicating high mean annual air temperatures and significant precipitation and water availability during the very early MIS 3 warming (Moseley et al., 2014). Summer insolation was as high as today in the Northern Hemisphere, and the North Atlantic warming comparable to today (van Kreveld et al., 2004), and CO_2 levels only 10 ppm lower than during interglacial times (Fig. 22). Consequently, it is not surprising that we see a rather warm climate distal of the northern ice sheets (see further discussion below).

The laminations produced during both LEZ 8 and 9 are all extremely thin, consisting mainly of diatoms and almost no coarse-grained detrital components. Flood events are absent from the entire early part of core AU2 with high C_{total} content (Figs. 19, 21). The thin laminations of LEZ 9 and 8 could be related to climate if it was indeed almost as warm as today. However, we cannot exclude that a fundamental change in the landscape structure appeared near to the maar, for example if the creek that feeds the maar still today got in contact with the maar basin only at the end of *Picea* Zones.

5.1.10. LEZ 10 The MIS 4–MIS 3 transition

Sediments below the *Picea* Zones have low organic carbon content and show only birch, pine and grass pollen, which indicate a boreal forest environment during the time when the Auel and Dehner Maar erupted. Apparently, the Dehner Maar is a little older than Auel, because it reveals the Leucite Tephra (LcT) in the lowermost lake sediments (Förster and Sirocko, 2016–in this volume).

5.2. Evidence for a warm early MIS 3 (49,000–60,000 b2k)

Terrestrial records of climate and vegetation history in central Europe during MIS 3 and MIS 2 are few because of a lack of suitable lakes to provide the necessary sedimentary basins (Helmens, 2014). Lakes in northern Europe and the Alps are generally formed by glacial processes either at the end of MIS 6 or MIS 2 and consequently mainly contain sediments deposited during the subsequent interglacials MIS 5e or MIS 1. Accordingly, Eemian and Holocene paleoclimate records are numerous across Europe. In addition, such glacial lakes are mostly not very deep and rapidly infilled with sediments. Lakes of other origin that might have contained MIS 3 records were overridden by the maximum glacier extent during MIS 2 and therefore not preserved.

A few sedimentary basins survived the MIS 2 glacier advance in northern Germany, however only inside the area glaciated during MIS 6, but outside the glaciated area during MIS 2. Best known are the classical sites of Oerel (Behre and Lade, 1986), Rederstall (Menke and Tynni, 1984) or Gröbern (Litt et al., 1996). Gross Todtshorn may be the only site in northern Germany to cover most of MIS 3 (Caspers and Freund, 2001), but the lake was situated in a colder environment than the Eifel.

In particular, the site at Oerel is known for the temperate Oerel interstadial with birch and pine trees at around 55,000 b2k (Behre and Lade, 1986). Even some pollen of *Quercus* were reported, but they are regarded as representing long distance transport from the south. A number of other dated MIS 3 records come from the Netherlands (e.g. Kasse et al., 1995; Vandenberghe, 1992a, 1992b; Zagwijn, 1974, 1996) or East Germany (e.g. Mania and Stechemesser, 1970; Eissmann, 1981; Litt et al., 1996; Mol, 1997). Some of these are fluvial deposits, which contain sections of interstadials as peat deposits or in-fills of small local lakes (often oxbows) in the river alluvial plain.

The general paleo-temperatures and overall climate dynamics of MIS 3 cold and warm phases have also already been reconstructed (Huijzer and Vandenberghe, 1998; Fletcher et al., 2010; van Meerbeck et al., 2011; Müller et al., 2011; Helmens, 2014). Profiles with several MIS3 interstadials were reported for Aschersleben in the Harz area (Mania and Stechemesser, 1970), Nussloch east of the Upper Rhine (Antoine et al., 2001) and Schwalbenberg in the Middle Rhine area (Schirmer, 2012).

The correlation of these records from these different sites across central Europe is however difficult, because they all show pine and birch if they contain pollen, but there is no specific assemblage characteristic for a distinct interstadial (Fletcher et al., 2010). An interpretation for these observations becomes plausible when the pollen in the DE3 core are evaluated on the basis of pollen counts (Fig. 7). Spikes become visible, that could be interpreted in a way that the community of trees did not change between stadials and interstadials, but only the total abundance of trees. *Betula*, *Pinus*, and to some extent *Ulmus* and *Picea*, show maxima in their absolute numbers during the interstadials (Figs. 7, 21). However, these maxima cannot be correlated to the classical Dutch interstadial succession, because the exact age and the succession of the interstadials Oerel, Glinde, Hengelo, Moorshöft and Denekamp are still not fully consistent between different authors. Some studies even shift the Brörup interstadial, which is in most papers attributed to MIS 5a/GI-21 (Caspers and Freund, 2001; Reille and de Beaulieu, 1990; Woillard and Mook, 1982) or the Dürnten interstadial (mostly attributed to GI-20 or 19 (Welten, 1982; Müller, 2001) into early MIS 3 (e.g. Leroy et al., 1996).

Fortunately, the order of interstadials is given in some French records, namely Grande Pile (Woillard, 1978), Velay (Kukla et al., 2002), but best resolved in the lake sediment records of Les Echets (De Beaulieu and Reille, 1989), which was reinvestigated following a multi-proxy approach with an updated chronology (Wohlfahrt et al., 2008). For Les Echets, it has been shown that the lake responded sensitively to past environmental forcings, with both MIS 3 interstadial-stadial events clearly recognizable in the trends of organic content (Veres et al., 2008, 2009) that in turn matched the lake internal productivity and arboreal/non-arboreal pollen trends (Wohlfahrt et al., 2008). Moreover, two intervals of high organic content dated to early MIS 3 (Veres et al., 2008, 2009) also document the presence of spruce pollen (De Beaulieu and Reille, 1989), further supporting assertions that the early MIS 3 LEZ in the Eifel sensitively capture a regional signal of environmental change that also characterize the Alpine forelands (see also Heiri et al., 2014).

Other high-resolution records of MIS 3 vegetation come from Italy, e.g., Lago Vico (Leroy et al., 1996) and Monticchio (Allen et al., 2000) but these sites near Rome and Naples respectively present the Mediterranean plant assemblages, which cannot be easily compared to the floral zones north of the Alps.

The general hemispheric boundary conditions during the early MIS3 favour a warm summer climate in central Europe. Summer insolation was higher than today, whilst atmospheric CO_2 and CH_4 were on the level of the early Holocene (Fig. 22). The Gulf Stream - North Atlantic drift reached to the southern tip of Norway (Kuijpers et al., 1998) and sea surface temperatures in the North Atlantic where only 2 °C lower than today (van Kreveld et al., 2004). Ice sheet reconstructions for the middle MIS 3 show a small ice sheet covering only little parts of the Norwegian mountain ranges (Mangerud et al., 2011; Wohlfahrt et al., 2008). Sea level was about 60 m lower than today and parts of the current North Sea and Baltic Sea were not submerged (Siddall et al., 2003). Accordingly, the general climate forcing conditions during the early MIS 3 were not so very different from parts of the Holocene (Fig. 22). The mean summer temperatures today in the Eifel are at 17 °C, and 19 °C in the Mosel and Rhine valleys. The plant macrofossils indicate that the early MIS 3 summer temperatures during LEZ 8 where near 15–16 °C, thus 2–3 °C lower than today.

Two grey forest soils in the loess around the Upper Rhine-Graben were described and named Gräselberger Boden 1 and 2 (Antoine et al., 2001; Bibus et al., 2007). Based on luminescence dating and relative paleosol stratigraphy they were assigned to early MIS 3. These soils are below the younger Lohne soil, which indicated a tundra environment with sparse trees, the only indicators of a pine-dominated forest in the loess landscape of Western Europe (Zech, 2012).

Consequently, it is possible that such conditions can allow the development of a *Picea* forest with abundant thermophile broadleaf taxa, in particular during the long interstadials GI-13 and GI-14, when temperatures in the Greenland NEEM record where up to 15 °C warmer than during the MIS 3 stadial phases (Rasmussen et al., 2014).

Other *Picea* forests have been dated to this time-interval on the Switzerland site of Gossau (Preusser, 2004). It appears that quite a number of records in France, Switzerland, the Eifel and even England show the environmental impact of a generally warm phase spanning GI-17 to GI-13, with duration of about 14,000 years.

The concept of a warm early MIS 3 in NW Europe is not new. Coope et al. (1998) investigated fossil beetle assemblages in English peat bogs and suggested that temperatures during this time-interval were only a few degrees below modern values, although there was no forest at this time in England. Quite likely, the forests from which the beetles spread were further to the east, most probably the Atlantic forests along the coast of France (Sanchez Goñi et al., 2013), where the forest community became increasingly dominated by *Picea* — similar to the forest in the Eifel during early MIS 3 (GI-17 to GI-12).

The ELSA core from the Oberwinkler Maar presents another line of evidence for a warm middle MIS 3. Here the sediments contain abundant chironomid remains, which included species that live in southern Scandinavia where summer temperatures are today ~14 °C (Engels et al., 2008).

The most compelling evidence for a warm early MIS 3 is the growth of speleothems in the Alps and in a cave of the Sauerland (D. Scholz, pers. Comm.), just 150 km away from the Eifel. Speleothems grew from 56,000 to 47,000 b2k even at the elevation of the modern snowline in the Alps (Spötl and Mangini, 2007; Moseley et al., 2014). Accordingly, the early MIS3, which we call LEZ8 in the ELSA records, has a high summer insolation forcing at a time of high North Atlantic sea surface temperatures and small Skandinavian ice sheet extent; a climatic situation which favours the development of a spruce forest and allows even for thermophilous trees in well suited stands as the Moselle and Rhine from where the trees can spread fast into the nearby Eifel.

6. Summary and conclusions

The ELSA core repository at Mainz now hosts 2700 m of laminated lake sediment, dated by 8 different methods, including 369 ^{14}C ages (Sirocko et al., 2013). The core HM1 from the Holzmaar (0–12,900 b2k) and core DE3 from the Dehner Maar (12,900–32,000 b2k) have annual-resolution chronologies developed from varve counting. Core AU2 from the Auel Maar is event-layer-laminated, because it is fed by a large stream and has the highest sedimentation rate of all extant and infilled Eifel maar lakes. It is tied tephrachronologically to the other MIS3 ELSA cores by five tephra marker layers (Förster and SIrocko, 2016–in this volume). Pollen and plant macrofossils from Dehner and Auel Maar are used to build the ELSA paleovegetation stack.

Core AU2 was sampled at 500 year intervals for pollen and macrofossils and at 100 year resolution for total carbon. The C_{total} record reveals the succession of all MIS3 interstadials from the Laacher See Tephra back to 60,000 b2k allowing precise tuning to the Greenland ice core chronology. This age model was used to define the boundaries of Landscape Evolution Zone (LEZ) for the last 60,000 years in central Europe.

LEZ 1 (0–6000 b2k): Forest with settlements and regional clearings for cattle and cereal farming.
LEZ 2 (6000–10,500 b2k): Natural broadleaf forest of the early Holocene climatic optimum with widespread thermophilous trees.
LEZ 3 (10,500–14,700 b2k): Boreal forest with cooler and warmer phases, volcanic activity.
LEZ 4 (14,700–23,000 b2k): Glacial desert with moss, some grass and submerged characea; annual dust activity, interpunctuated by short phases of tundra most likely during GI2.
LEZ 5 (23,000–28,500 b2k): Tundra with abundant Ranunculacaea seeds, insect remains and fungal spores (Coenococcum). Dust activity almost every year.
LEZ 6 28,500–36,500 b2k): Steppe with grass, pine, birch and fungal spores. Increases in the amount of pollen from pine and birch during interstadials, the last dense boreal forest around 30,000 b2k. Complete disappearance of all vegetation during stadials. Dust storms in 2 of 3 years. Volcanic activity during the final millennia of this LEZ. Spread of anatomically modern humans in the increasingly open landscape, where horse, reindeer and mammoth, the favoured hunting preys, must have been abundant.
LEZ 7 (36,500–49,000 b2k): Boreal forest of pine, birch and few spruce, little dust activity. Charcoal indicates drought stress and frequent forest fires, volcanic activity.
LEZ 8 (49,000–55,000 b2k): Spruce and hornbeam forest with thermophilous trees.
LEZ 9 (55,000–60,000 b2k): Spruce forest.
LEZ 10 (older than 60,000 b2k): Intense volcanic activity at the end of MIS4 in a fast changing environment of boreal forest.

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Christoph Herbig did the already published plant macroremain counting of HM1 and provided the unpublished data of core DE3. Hannes Knapp did the macroremain analysis of core AU2; photographs of macroremains were done by Christel Adams. Pollen for all cores were counted by Frank Dreher; Michael Zech provided the total carbon results. Michael Förster, Johannes Albert, Heiko Brunck and Marieke Röhner worked on the stratigraphy. Simone Illig contributed the diatom analysis and the Characeae identification was done by Michael Dilger. Stephan Dietrich, Ulrich Hambach and Daniel Veres set the above

ELSA results into the context of the stadial/interstadial succession of MIS3. Core photographs were taken by Carsten Costard and processed by Klaus Schwibus. Saskia Rudert organized the data handling in the ELSA*interactive* system, programmed by Martin Seelge. Petra Sigl finally elaborated the figures with great care.

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The data for all ELSA cores are available from the ELSA website at http://www.klimaundsedimente.geowissenschaften.uni-mainz.de.

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