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Chironomid-inferred Holocene temperature changes in the South Carpathians (Romania)

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Abstract

We present a Holocene summer air temperature reconstruction based on fossil chironomids from Lake Brazi (1740 m a.s.l.), a shallow mountain lake in the South Carpathians. Summer air temperature reconstruction was performed using transfer functions based on the Swiss (Sw-TF) and the merged Norwegian-Swiss calibration dataset (NS-TF). Our results suggest that summer air temperatures increased rapidly from the onset of the early Holocene onwards (*ca* 11,500–10,200 cal yr BP), reaching close to present July air temperatures (~11.2°C). Between *ca* 10,200–8500 cal yr BP mean reconstructed temperatures increased further by 1.5-2.0°C. Later on, from *ca* 8500 cal yr BP chironomid-based summer temperatures started to decrease, although mean values were still above present-day temperatures. The next time period (*ca* 6000–3000 cal yr BP) was cooler and with less variable temperature conditions than earlier. Afterwards (*ca* 3000–2000 cal yr BP), a sharp decrease occurred in inferred temperatures with values under present-day conditions by 1.8°C. Finally, in the last 2000 years, reconstructed temperatures showed again an increasing trend at Lake Brazi. Short-term temperature declines of 0.6–1.2°C were observed between *ca* 10,350–10,190, 9750–9500, 8700–8500, 7600–7300, 7100–6900 and 4400–4000 cal yr BP. These temperature declines are however within the estimated error of prediction of the chironomid-based inferences. Generally, our reconstructed temperatures complied with the summer insolation curve at 45°N, with other proxy records (i.e. pollen and diatoms) from the same sediment and with other records from the Carpathians and from Western Europe.

Keywords: Holocene, Chironomidae, temperature reconstruction, paleolimnology, Retezat Mountains

Introduction

The Holocene (from *ca* 11,600 cal yr BP to present; Blockley et al., 2012) is widely regarded as a warm period with relatively stable climatic conditions. Nevertheless, a series of high resolution ice, marine and terrestrial sediment records demonstrated that in Europe the climate of the Holocene was quite variable, although with smaller amplitude changes than observed for the climatic reversals of the late Pleistocene period. Several abrupt short-term warm and cold (or arid and humid) oscillations are described during the Holocene (e.g. Bond et al., 1997; Alley et al., 2003; Mayewski et al., 2004; Magny et al., 2007; Wanner et al., 2011). These short-term oscillations are believed to be related to changes in solar activity (e.g. Bond et al., 2001; Wanner et al., 2008) or variations in the North Atlantic thermohaline circulation (e.g. Alley et al., 2003; Wiersma and Renssen, 2006). In addition, the last remnants of the northern ice sheets likely had an important cooling effect also during the early Holocene, at least regionally (Renssen et al., 2009). Most of our knowledge about major Holocene climatic changes comes from Northern and Central Europe, while from the eastern part of the continent only few quantitative climate reconstructions are available (e.g. Mayewski et al., 2004). However, for a better understanding of the past climatic changes and their spatial variability across Europe, it is essential to gain more information on this less known area (Feurdean et al., 2014).

Recently quantitative climate reconstructions covering the Holocene have become available from the Carpathian Mountains. These records are based on pollen analysis (e.g. Feurdean et al., 2008), stable oxygen and carbon isotopes from stalagmite records (e.g. Onac et al., 2002; Tămaş et al., 2005; Constantin et al., 2007) and tree rings analysis (Popa and Kern, 2009). Within an intensive multi-proxy paleoecological investigation, the PROLONG project (Magyari et al., 2009a), we focused on developing proxy records from sediment sequences of mountain lakes formed during the last glaciation in the Retezat Mountains of the South Carpathians (Figure 1). First studies concentrated mainly on the late glacial and the early Holocene sediment of Lake Brazi (e.g.; Magyari et al., 2009a; Korponai et al., 2011; Braun et al., 2012; Buczkó et al., 2012; Tóth et al., 2012; Magyari et al., 2013), while sections of the sequences covering the middle and late Holocene are currently being analyzed (e.g. Buczkó et al., 2013; Magyari et al., 2013; Finsinger et al., 2014; Pál et al., *in press*). As a part of this project, subfossil chironomid (Diptera: Chironomidae) assemblages have already been used for temperature reconstructions of July air temperature during the late glacial (Tóth et al., 2012).

Based on the strong relationship between chironomid distribution and temperature (e.g. Brooks, 2006; Eggermont and Heiri, 2012), chironomid temperature transfer functions have been developed for several geographical regions (e.g. Larocque et al., 2001; Heiri et al., 2011; Self et al., 2011), and have been used successfully to reconstruct past temperature changes during the late glacial and the Holocene (e.g. Brooks and Birks, 2000; Heiri et al., 2003; Ilyashuk B et al., 2009; Larocque-Tobler et al., 2010; Ilyashuk EA et al., 2011; Płóciennik et al., 2011; Heiri et al., 2014). Nevertheless, since temperature changes during the Holocene are characterized by smaller amplitude than during the late glacial, other environmental factors (for example pH, trophic conditions, water depth, etc.) may influence a temperature reconstruction within the Holocene (e.g. Brodersen and Anderson, 2002; Heiri and Lotter, 2005; Velle et al., 2005; Brooks, 2006; Velle et al., 2010).

Here we present a high-resolution Holocene chironomid record from the Retezat Mountains; which together with the already described late glacial part of the same sediment sequence (Tóth et al., 2012) provides the first complete (late glacial and Holocene) chironomid record from the South Carpathians. The objectives of the present paper are (1) to describe the main compositional trends of subfossil chironomid assemblages during the Holocene; (2) to reconstruct Holocene mean July air temperature trends using the Swiss and the merged Norwegian-Swiss transfer function; and finally (3) to compare the inferred temperature trends with those obtained in other regional and European records. Our results provide a valuable addition to the available temperature records from Eastern Europe that can be used to evaluate climate model results for the Holocene and to understand ecosystem responses to abrupt climate changes.

Study site

The Retezat Mountains, located in the South Carpathians, are among the wettest massifs (annual rainfall 1400 mm yr^{-1} at 1500–1600 m a.s.l.) in Romania due to vapor supply by both Mediterranean and Atlantic air masses (Jancsik, 2001; Magyari et al., 2009a). The climate of the Retezat is temperate continental. The mean annual temperature is around +6°C in the foothill zone and -2°C at the top of the mountain (2500 m a.s.l.). At present, January is the coldest and July is the warmest month characterized with mean temperatures of 6.6°C and +11.2°C at 1740 m a.s.l., at the elevation of Lake Brazi (estimated by linear interpolation from the five nearest meteorological stations; Bogdan, 2008; Magyari et al., 2013).

In this study, a glacial lake called Lake Brazi (Tăul dintre Brazi, 45°23'47"N, 22°54'06"E; Figure 1), was investigated. The lake is situated in the subalpine belt at 1740 m a.s.l. on the western marginal side of the Galeş glacial valley, in a mixed Norway spruce (*Picea abies)* - stone pine *(Pinus cembra)* forest. Lake Brazi is a small shallow lake with maximum water depth of 1.1 m and a surface area of 0.4 ha (Magyari et al., 2009a).

Pollen data from Lake Brazi (E. Magyari, *unpublished data*) suggest that human impact likely occurred in the vicinity of the lakes in the last 1500 years, with occasional cutting of the nearby spruce forests. This is similar to other high mountain pollen records in the Carpathians that suggested also significant human activities in other parts of the Romanian Carpathians in the last 1500 years (Feurdean and Astaloş, 2005). Additionally, minor impact by grazing animals (sheep, cattle, horse) that were shepherded to the summer mountain pastures in the nearby Galeş river valley is also reported (Maderspach, 1986).

[insert Figure 1]

Chronology

A chronological framework of the sediment was established using twenty-one AMS ^{14}C dates (Table 1). Radiocarbon dates suggest that sediment accumulation started at around 15,750 cal yr BP and was continuous throughout the late glacial and the Holocene. The age-depth relationship of the sediment was assessed using two models: (1) a weighted non-linear polynomial regression model below 502 cm sediment depth, for the late glacial and early Holocene (see more details in Magyari et al., 2009a), and (2) a smooth spline function in CLAM v2.1 (Blaauw, 2010) between 502–111 cm, for the Holocene sediment (Figure 2). This latter section had considerably larger sediment accumulation rates, which justifies the application of separate age-depth models for the two sections.

Overall, we excluded five dates from the age-depth modelling, because they were stratigraphically inconsistent with the majority of the ^{14}C dates (shown in grey bands in Table 1).

[insert Figure 2 and Table 1]

Methods

Fieldwork and laboratory analyses

A 490 cm long sediment core (TDB-1; 111–600 cm) was taken from the central part of the lake (water depth 111 cm) in August 2007 with a modified Livingstone piston corer (diameter 7 cm). Here we concentrate on the upper Holocene part (111–552 cm) of this core.

The sediment stratigraphy and organic content record were described in detail in Magyari et al. (2009a) and Buczkó et al (2013).

For chironomid analysis 0.56–1.5 cm^3 sediment was investigated at 2 cm intervals between 552–502 cm and at 4 cm intervals between 500–111 cm. Sub-samples were deflocculated in 10% KOH and heated at 60°C for 20 min. Afterwards the sediment was sieved with a 100 µm mesh. Chironomid larval head capsules were picked from a Bogorov-counting tray (Gannon, 1971) under stereomicroscope at 40x magnification. Larval head capsules were mounted on microscope slides in Euparal® mounting medium for microscopic identification. At least 47-50 head capsules were identified from each sub-sample, so they provided a representative count for quantitative analysis (Heiri and Lotter, 2001). Identification of chironomid head capsules followed Wiederholm (1983), Rieradevall & Brooks (2001), and Brooks et al. (2007).

Plotting, numerical analyses and temperature reconstruction

The chironomid relative abundance diagram was plotted using the program Psimpoll 4.27 and zonation was based on optimal splitting by information content (Bennett, 2007). To summarize major changes in subfossil chironomid assemblages Detrended Correspondence Analysis (DCA) was performed using CANOCO version 4.5 (ter Braak and Šmilauer, 1998). Before the ordination, percentage chironomid species data were square-root transformed and rare taxa were down-weighted. The gradient length of the longest DCA axis (axis 1) was 2.15 SD units.

Since calibration data-sets are not available from the Carpathian region, we used calibration sets from other regions for summer air temperature reconstruction. Chironomid-inferred mean July air temperature (T_{VII}) reconstructions were performed using weighted averaging partial least-squares regression (WA-PLS; ter Braak and Juggins, 1993) based on both the Swiss and the merged Norwegian-Swiss chironomid-temperature calibration training sets. The Swiss training set includes altogether 117 lakes situated in the Jura Mountains, the Swiss Plateau and the Swiss Alps (Lotter et al., 1997; Heiri et al., 2003; Bigler et al., 2006). The composition of the subfossil chironomid assemblages from Lake Brazi is very similar to the ones known from the Alps; therefore the application of the Swiss transfer function seemed to be reasonable. On the other hand, the merged Norwegian-Swiss training set – based on surface sediment samples from 274 lakes (including the Swiss training set and data from Norway and Svalbard) – covered wider altitudinal, latitudinal and lake water pH ranges, and wider temperature gradients than the Swiss or the Norway models individually (Heiri et al., 2011; Brooks and Birks 2000, 2001). Furthermore, it has been used before to reconstruct temperatures based on chironomid assemblages in the late glacial part of the Lake Brazi record (Tóth et al., 2012). Before the analyses, 15 lakes from the Swiss data-set and altogether 19 lakes from the Norwegian-Swiss data set were excluded as outliers. These lakes are characterized by unusual hydrological conditions; their chironomid assemblages were influenced by running waters, glacier meltwater or extensive snow meltwater (Norwegian lakes), or they were distinctly larger than the remaining training set lakes (for more details see Heiri et al., 2011). Prior to temperature reconstruction, percentage chironomid data were square-root transformed. The summer air temperature reconstructions and sample-specific errors of prediction (SSPEs) based on bootstrapping (999 bootstrap cycles) were calculated using the program C2 (Juggins, 2007).

Reconstruction diagnostic statistics

In order to evaluate the reliability of the chironomid-inferred temperature reconstructions, we estimated the cross-validated root mean square error of prediction (RMSEP), the chi-square distance to the closest modern analogue, the percentage of rare taxa in the training set with the C2 (Juggins, 2007) and goodness-of-fit measures using CANOCO version 4.5 (ter Braak and Šmilauer, 1998). All calculations were based on square-root transformed percentage abundances.

Fossil assemblages with a chi-square distance to the most similar assemblage in the modern calibration dataset larger than the 2nd and the 5th percentile of all squared chi-square distances in the modern data were identified as samples with "no close" and "no good" analogue, respectively (Birks et al., 1990; Heiri et al., 2003). Fossil samples with a residual distance to the first CCA axis larger than the 90th and 95th percentile of the residual distances of all the modern samples were identified as samples with "poor fit" and "very poor fit" with temperature, respectively (Birks et al., 1990). Chironomid taxa with a Hill's N2 (Hill, 1973) below 5 in the calibration data were considered to be rare in the modern dataset (Heiri et al., 2003).

Results

Chironomid assemblages

Altogether 22 chironomid taxa (overall 10,960 head capsules) were identified from the sediment and out of these 17 taxa were present in the late glacial sediment section as well (Tóth et al., 2012). Based on the relative abundances, six chironomid assemblage zones were distinguished (Figure 3). Zone boundaries are mainly associated with abundance shifts of the most dominant taxa *Tanytarsus mendax*-type, *Psectrocladius sordidellus*-type, *Zavrelimyia*

type A and *Tanytarsus lugens*-type. The first zone (Zone-1; 552–542 cm; *ca* 11,500–10,900 cal yr BP) was dominated by *T. lugens*-type and *Micropsectra insignilobus*-type. Both taxa disappeared by *ca* 11,000–10,900 cal yr BP, and the abundance of *Zavrelimyia* type A and *Paratanytarsus austriacus*-type started to increase in the second part of the zone. At the onset of the second zone (Zone-2; 542–518 cm; *ca* 10,900–10,200 cal yr BP) relative abundances of *C. anthracinus*-type (23–38%), *Tanytarsus pallidicornis*-type 2 and *Procladius* reached maximum values. Later on, from *ca* 10,400 cal yr BP, *P. austriacus*-type and *P. sordidellus*-type became dominant. The third zone (Zone-3; 518–330 cm; *ca* 10,200–6300 cal yr BP) was dominated by *T. mendax*-type with a broad thermal tolerance. In the fourth zone (Zone-4; 330–230 cm; *ca* 6300–3200 cal yr BP) *P. sordidellus*-type became dominant, while the abundance of *T. mendax*-type decreased sharply. The fifth zone (Zone-5; 230–160 cm; *ca* 3200–1550 cal yr BP) was marked by the dominance of *T. lugens*-type, while relative abundances of *P. sordidellus*-type and *T. mendax*-type decreased further. Finally, in the sixth zone (Zone-6; 160–111 cm; *ca* 1550 cal yr BP–present) *T. lugens*-type started to decrease, but still dominated together with the increasing *T. mendax*-type and *Zavrelimyia* type A.
The chironomid relative abundance diagram, presenting all of the chironomid taxa as percentage abundances, is shown in Figure 3.
[insert Figure 3]

Ordination of the chironomid record

The first two DCA axes explained 51.7% (31.5% and 20.2%, respectively) of the variance in the chironomid dataset. Along the first DCA axis, notable changes of about 1.8 and 0.7 SD units were observed at the transition from Zone-1 to Zone-2 (at *ca* 10,900 cal yr BP) and from Zone-4 to Zone-5 (*ca* 3200 cal yr BP), respectively. Generally, the first DCA axis separated Zone-1 and Zone-5 based on the higher relative abundance of *T. lugens*-type and

Micropsectra insignilobus-type. These taxa had negative values on the first DCA axis, while others (i.e. *Tanytarsus pallidicornis*-type2, *T. mendax*-type, *Chironomus anthracinus*-type and *Endochironomus impar*-type) had strong positive values on the same axis (Figure 3).
On the second DCA axis, Zone-2 was separated clearly from the other sediment layers. This axis represented a gradient mostly characterized with relative abundance changes of *T. pallidicornis*-type 2 and *C. anthracinus*-type (Figure 3).

Summer air temperature reconstructions

The reconstructed July air temperatures (T_{VII}) ranged from 8.1 to 14.2°C with the merged Norway-Swiss (NS-TF) and from 8.3 to 15.2°C with the Swiss transfer function (Sw-TF). For samples older than *ca* 5000 cal yr BP, the reconstructed temperature values were consequently higher with the Sw-TF than with the NS-TF. At the same time, the reconstructed summer air temperature trends based on the two transfer functions were highly consistent (Figure 4).
Between *ca* 11,500–10,980 cal yr BP (552–543 cm), inferred temperatures increased rapidly by about 1.2°C and 0.8°C (until *ca* 9.3–9.7°C) based on the NS-TF and Sw-TF, respectively. It was followed by a further temperature increase by *ca* 2.2–2.5°C until 10.8–12.3°C (NS-TF) and 11.5–12.9°C (Sw-TF) between *ca* 10,980–10,220 cal yr BP (543–518 cm). Later on, from *ca* 10,220 cal yr BP (518–330 cm), inferred temperatures fluctuated strongly above present-day T_{VII} (~11.2°C) by *ca* 1.4°C (NS-TF) and by *ca* 2.5°C (Sw-TF). Then, T_{VII} started to decrease by *ca* 1.6–1.7°C between *ca* 6300–3300 cal yr BP (330–235 cm). It was followed by a further decrease until chironomid-inferred T_{VII} fell below present-day values by *ca* 1.6°C between *ca* 3300–2000 cal yr BP (235–170 cm). Finally, in samples younger than *ca* 1500 cal yr BP (170–111 cm), reconstructed temperatures fluctuated strongly between *ca* 9.7–11.6°C, close to the modern T_{VII} (Figure 4).

Besides numerous single-sample temperature drops and rises, we found six short periods (covering at least 2–3 samples and at least 150 years based on both of the TFs applied) with temperature declines of 0.6–1.2°C: between *ca* 10,350–10,190, 9750–9500, 8700–8500, 7600–7300, 7100–6900 and 4400–4000 cal yr BP (Figure 4). All of these declines were however within the estimated error of prediction of the chironomid-based inferences.

[insert Figure 4]

Reliability of the inferred temperatures

The chironomid-inferred July air temperature reconstruction from Lake Brazi revealed a RMSEP of 1.39°C and 1.40°C based on the NS-TF and the Sw-TF, respectively. We found "no close" analogue situation in 22% and 41% of the samples based on NS-TF and Sw-TF, respectively. Additionally, 10% of all samples had "no good" analogue in the modern data based on Sw-TF. Generally, the samples older than 7000 cal yr BP are affected by analogue problems based on both transfer functions (Figure 5).

Goodness-of-fit statistic showed that only 0.9% and 3.4% of the samples have "poor fit" with temperature based on NS-TF and Sw-TF, respectively. Further 2.5% of the samples have "very poor fit" with temperature based on the Sw-TF (Figure 5).

All of the chironomid taxa encountered in the sediment of Lake Brazi occurred in both modern training sets. Furthermore, the taxa that are considered to be rare (Hill's N2< 5) in the merged Norwegian-Swiss (*Metriocnemus fuscipes*-type) and in the Swiss (*Cricotopus sylvestris*-type, *M. fuscipes*-type) training sets were present with maximum relative abundances less than 2% (Figure 5).

[insert Figure 5]

Discussion

Ecological interpretation of changes in the subfossil chironomid assemblages

During the early Holocene (*ca* 11,500–11,000 cal yr BP), dominant chironomid taxa (*T. lugens*-type, *M. insignilobus*-type; *Zavrelimyia* type A and *Paratanytarsus austriacus*-type) indicate relatively cold and oligotrophic conditions (Brodersen and Anderson, 2002; Boggero et al., 2006; Tátosová et al., 2006). At the same time, increasing number of *C. anthracinus*-type and increasing loss-on-ignition (LOI) values support the onset of warming in the early Holocene (Figure 3). Later on, from *ca* 11,000 cal yr BP, Chironomini taxa (*C. anthracinus*-type and *E. impar*-type) reach their maximum abundance and indicate generally higher temperatures than before (e.g. Larocque et al., 2001; Velle et al., 2005; Heiri et al., 2011). Moreover, *C. anthracinus*-type indicates slightly eutrophic conditions as well (Kansanen, 1986; Velle et al., 2005). Next to Chironomini, *T. pallidicornis*-type 2, typical for littoral sediments of mesotrophic, temperate or warmer lakes (Sæther, 1979; Heiri et al., 2003; Luoto, 2010; Heiri et al., 2011), occurs in high relative abundance. The dominant taxa suggest warming summer air temperatures with meso- to eutrophic lake conditions at Lake Brazi in the second part of the early Holocene (Figure 3 and Table 3).

Between *ca* 10,200–6300 cal yr BP, *T. mendax*-type became the most dominant taxon, including several chironomid species with different ecological optima (Brooks et al., 2007). Generally, they occur under relatively warm climatic conditions (Engels and Cwynar, 2011). Based on Fjelheim et al. (2009), *Tanytarsus gregarius* Kieffer 1909 is the only known species belonging to the *T. mendax*-type from the Retezat Mountains (the species was found in Taul Negru and Taul Gemenele). In the Northern Carpathians (Tatra Mountains), *T. gregarius* occurs mainly in lakes with relatively high productivity and sometimes with acidic conditions (Bitušík et al., 2006, 2010; Kubovčík and Bitušík, 2006). Therefore, between *ca* 10,200–6300 cal yr BP, chironomid assemblages suggest warm summer temperatures and moderately high nutrient levels, which are also supported by the increasing LOI values (Figure 3 and Table 3). Later on, between *ca* 6300–3300 cal yr BP, relative abundance of *T. mendax*-type decreases (Figure 3), and the dominant taxa (*P. sordidellus*-type and *P. austriacus*-type) are characterized by lower temperature optima than in the previous time period (e.g. Velle et al., 2005; Heiri et al., 2011), even though *P. sordidellus*-type occurs in temperate climatic conditions in European lowlands as well (e.g. Brooks and Birks, 2000). Additionally, both taxa tolerate (but not necessary indicate) periodic acidification (Velle et al., 2005; Tátosová et al., 2006; Brooks et al., 2007) and are frequently associated with macrophytes (e.g. Brodersen et al., 2001; Luoto, 2010; Engels and Cwynar, 2011). Therefore, chironomid assemblages suggest lower temperatures, but still warm summers at Lake Brazi between *ca* 6300 and 3300 cal yr BP (Table 3).

Between *ca* 3300–1500 cal yr BP, *T. lugens*-type became the most dominant taxon, and together with the reappearing *M. insignilobus*-type (Figure 3) indicates cold and mainly oligotrophic conditions (e.g. Larocque et al., 2001; Brodersen and Anderson, 2002; Velle et al., 2005; Tátosová et al., 2006; Heiri et al., 2011). Additionally, *T. lugens*-type is described from the deeper parts of shallow lakes or from the profundal zone of deep lakes (Luoto, 2010; Engels and Cwynar, 2011). In summary, from *ca* 3300 cal yr BP the chironomid fauna suggests increasing water level and decreasing nutrient status with summer cooling at Lake Brazi (Table 3).

Finally, during the last *ca* 1500 years the previously dominant taxa (*T. lugens*-type, *P. sordidellus*-type, *T. mendax*-type, *Zavrelimyia* type A), formed a diverse assemblage that also included Chironomini taxa in very low number (Figure 3 and Table 3). As discussed above, these taxa indicate very diverse environmental conditions, since they occur in a wide range of summer air temperatures, trophic states and water depth gradients. Furthermore, the last 2000 years are characterized by an abrupt increase in diatom-inferred total epilimnetic phosphorous (DI-TP) values that suggests increased eutrophication in Lake Brazi (Buczkó et al., 2013). Additionally, pollen data from Lake Brazi show a clear decrease in the relative abundance of trees (by *ca* 14-15%), primarily driven by a decline in relative abundances and accumulation rates of *Picea abies*, and an increase of relative abundance (by about 6%) of Poaceae from *ca* 1600cal yr BP (E. Magyari, *unpublished data*). These data suggest intensifying human impact (most likely forest management) in the South Carpathians, similar to other parts of the Romanian Carpathians from about 1500 cal yr BP (Feurdean and Astaloş, 2005) and from *ca* 2000 cal yr BP in the South Balkans (e.g. Panagiotopoulos et al., 2013). Therefore it is challenging to assess whether climate or increasing human impact were driving changes in the chironomid assemblages in the last 1500 years.

At the same time, corresponding to the ordination results, we assume that the most influential factor on observed assemblage changes is summer air temperature (Table 3). The first DCA axis clearly separated two time periods (between *ca* 10,900–10,200 and 3200–1550 cal yr BP) in the Brazi chironomid stratigraphy. These periods are characterized by chironomid taxa (i.e. *T. lugens*-type and *M. insignilobus*-type) living in subalpine and alpine regions and mainly in cold lakes (Brooks et al., 2007; Heiri et al., 2011). These taxa have negative values on the first DCA axis, while others present in the sequence (i.e. *T. pallidicornis*-type2, *T. mendax*-type, *C. anthracinus*-type and *E. impar*-type) are typical for warmer climatic conditions (Velle et al., 2005; Samartin et al., 2012) and have strongly positive values on the same axis. Therefore, it is very likely that the first DCA axis represents a temperature gradient and thus supports the marked influence of temperature changes on the Holocene chironomid assemblages.

Comparison of chironomid-inferred temperatures with other proxies from the region and Western Europe

At the onset of the Holocene (at *ca* 11,500 cal yr BP), chironomid-based reconstructions show a marked two-step increase in July air temperatures by *ca* 3.8°C in Lake Brazi (Tóth et al., 2012). Rising temperatures are inferred also by pollen and plant macrofossil data from the same sediment sequence (Magyari et al., 2012), by pollen data from the Eastern Carpathians (Feurdean et al., 2008) and by increasing $\delta^{18}O$ values of stalagmites from both the Western (Tămaş et al., 2005) and the Southern Carpathians (Constantin et al., 2007). The amplitude of the reconstructed increase at Lake Brazi is similar to other chironomid records from the Southern and Central Swiss Alps, as well as from the Northern Apennines (*ca* 3.8–4°C; Samartin, 2011; Samartin et al., 2012; Ilyashuk B et al., 2009). At the same time, smaller chironomid-based temperature changes are reported from Northern Italy by *ca* 2.5°C (Larocque and Finsinger, 2008), from the Northern Swiss Alps and the French Jura Mountains by *ca* 1.5°C (Heiri et al., 2003; Heiri and Millett, 2005). A short-term temperature drop is indicated by chironomids at Lake Brazi between *ca* 10,350–10,190 cal yr BP. This cool period may coincide with the 10.2 ka cold event described from the same sediment sequence (Buczkó et al., 2009; Magyari et al., 2012) and other records of the Carpathians (e.g. Tămaş et al., 2005; Feurdean et al., 2008).

Later on, between *ca* 10,200 and 8500 cal yr BP, generally high summer air temperatures are detected at Lake Brazi with mean reconstructed values above the present-day temperature (~11.2°C) by about 1.5–2.5°C. These results coincide with decreasing lake levels associated with higher than present summer temperatures as indicated by diatoms and pollen from the same sediment (Buczkó et al., 2013; Pál et al., *in press*), and probably with higher summer insolation (Figure 4) from *ca* 9500 cal yr BP. A similarly warm climate and shallow lake conditions were reported from the Eastern Carpathians (Feurdean et al., 2008; Magyari et al., 2009b) based on pollen (from *ca* 10,200 cal yr BP) and multi-proxy data (between *ca* 9300–8900 cal yr BP), as well based on stalagmite records from the Western and Southern

Carpathians (Tămaş et al., 2005; Constantin et al., 2007). Moreover, from *ca* 10,200 cal yr BP our chironomid inferences comply with summer insolation at 45° N (Laskar et al., 2004; Figure 4).

Largely associated with the summer insolation maximum, a relatively warm climate (Holocene Thermal Maximum; HTM) during the early and mid-Holocene is described in middle and high latitudes of the Northern Hemisphere (Wanner et al., 2008; Renssen et al., 2009). Model simulations suggest that the HTM took place between *ca* 8000 and 6000 cal yr BP in most parts of Europe, followed by a marked cooling in the late Holocene (Renssen et al., 2009). The warmest period of our temperature reconstructions at Lake Brazi (45°N, 22°E) dates between *ca* 9400 and 8900 cal yr BP when summer air temperatures were 2–3°C higher than at present. This temperature pattern coincides with the summer insolation maximum at 45° N (Laskar et al., 2004; Figure 4) and a recently published charcoal record from the Transylvanian Plain (46°N, 23°E; Feurdean et al., 2013) where the highest mean fire interval is detected between 10,100 and 7100 cal yr BP. The authors suggested that the observed trends in fire activity are associated strongly with high summer insolation, which led to higher temperatures (of about 4°C) and lower available moisture in this time interval than at present (Feurdean et al., 2013). Similarly, an earlier HTM was reported from the Central Eastern Alps (46° N, 10° E) between *ca* 10,000–8600 cal yr BP (Ilyashuk EA et al., 2011) than suggested by the climate model simulations of Renssen et al. (2009). The HTM is mainly associated with the orbitally forced summer insolation maximum; however, the presence of the last remnants of the northern ice sheets may have produced cold surface ocean conditions in the North Atlantic, which in turn influenced early Holocene temperatures in NW Europe (Wanner et al., 2008; Renssen et al., 2009; Wanner et al., 2011). Lake Brazi is in sheltered position by the Eastern and Northern Carpathians against westerly air masses, and this continental interior region was less affected by cooling events of the North Atlantic, which could explain why the temperatures at Lake Brazi closely followed the Holocene insolation changes (Feurdean et al., 2014).

Between *ca* 10,200 and 8500 cal yr BP, two short-term temperature declines were indicated at Lake Brazi. The first decline appeared between *ca* 9700–9500 cal yr BP and coincided with episodic high lake level and decreasing DI-TP values in the diatom record (Buczkó et al., 2013), and with decreasing LOI values (Figure 4). Additionally, decreased $\delta^{18}O$ values of a stalagmite denote a temperature decline also at around 9300 cal yr BP in the Western Carpathians (Tămaş et al., 2005). This temperature drop is synchronous with a temperature decrease observed in the chironomid-based temperature reconstructions from Northern and Southern Europe at *ca* 9350–9200 cal yr BP as well (Korhola et al., 2002; Samartin, 2011). The second decline was dated between *ca* 8700–8500 cal yr BP. A diatom-based $\delta^{18}O$-record form Lake Brazi showed a distinct decline also between *ca* 9000–8500 cal yr BP, likely associated with increased winter precipitation in the South Carpathians (Magyari et al., 2013). During the mid-Holocene, chironomid-inferred T_{VII} started to decrease slightly at Lake Brazi from *ca* 8500 cal yr BP but mean summer temperatures were still above present-day values by 0.5–1.3°C. Based on diatoms from the same sediment sequence a distinct lake-level rise occurred after *ca* 8400 cal yr BP (Buczkó et al., 2013), most likely related to the well-defined 8.2 ka event (Alley et al., 2003). Otherwise, pollen data and increased microcharcoal accumulation rates indicated still high summer temperatures between *ca* 8300 to 8100 cal yr BP, likely associated with summer droughts, but overall macrocharcoal inferred local fire frequencies decreased in this period (Finsinger et al., 2014; Pál et al., *in press*). The 8.2 ka cooling event is well-documented in the Carpathian and Balkan region (e.g. Feurdean et al., 2013; Buczkó et al., 2013; Panagiotopoulos et al., 2013), however, in this region it had the greatest influence on the winter and spring climate, while summer temperatures remained high (Feurdean et al., 2008, 2014; Pál et al., *in press*). This latter assumption coincides with the lack of a clear 8.2 ka cooling in our chironomid record. On the other hand, this cooling is also well detected in chironomid-inferred summer temperature records in the Alps (Heiri et al., 2003, 2004; Ilyashuk EA et al., 2011; Samartin, 2011) and in Fennoscandia (Korhola et al., 2002).

Between *ca* 8500 and 6500 cal yr BP, two distinct short-term temperature declines are noted at Lake Brazi. The first is dated between *ca* 7600–7300 cal yr BP and coincides with a distinct decline in the diatom-based $\delta^{18}O$ values measured from the same sediment sequence (*ca* 7800–7300 cal yr BP) and is likely associated with increased winter precipitation (Magyari et al., 2013) as well as with a distinct decline in LOI (Figure 4). Similarly cooler and wetter climatic conditions are noted based on a stalagmite $\delta^{18}O$-record from the Western Carpathians (Tămaş et al., 2005) and based on pollen data from the Eastern Carpathians (Magyari et al., 2009b). The second cooling occurred between *ca* 7100–6900 cal yr BP. Decreasing $\delta^{18}O$-values from the Western and the Southern Carpathian speleotherms (Onac et al., 2002; Constantin et al., 2007) also suggest a temperature decline at around 7100–7000 cal yr BP. From *ca* 6500 cal yr BP onwards, the chironomid record suggests a further decrease in T_{VII} until reconstructed temperatures fluctuate close to modern summer air temperature value at Lake Brazi. Moreover, LOI showed a clear decline at the same time (Figure 4). Similarly, a stalagmite record from the Western Carpathians (Onac et al., 2002) indicated lower annual temperatures reaching modern values as well from *ca* 6800 cal yr BP. Diatoms suggest higher lake levels at Lake Brazi from *ca* 6000 cal yr BP (Buczkó et al., 2013), which seems to comply with summer cooling and increasing precipitation (Magyari et al., 2009a). Similarly increasing lake levels and cooling climate are noted from the Eastern, the Western and from other parts of the Southern Carpathians (Onac et al., 2002; Constantin et al., 2007; Magyari et al., 2009b) between *ca* 5500–4200 cal yr BP. A short-term temperature decline was noted between *ca* 4500–4000 cal yr BP, followed by a slight temperature rise. This temperature drop is concomitant with a decline in the diatom $\delta^{18}O$-record and in LOI values from Lake Brazi (Figure 4), and is most likely associated with increased winter precipitation (Magyari et al., 2013). Similarly, a stalagmite record from the Southern Carpathians and pollen-inferred temperatures from the Eastern Carpathians show a temperature decline around *ca* 4000 cal yr BP (Constantin et al., 2007; Feurdean et al., 2008). A distinct temperature drop is also reconstructed based on chironomids from the Northern Apennines between *ca* 4900–4050 cal yr BP (Samartin, 2011) and from the Finnish Lapland around 4200 cal yr BP (Korhola et al., 2002).

Generally, chironomid-inferred summer air temperature reconstructions from Western Europe suggest warmer summers during the early and mid-Holocene than during the late Holocene (Wanner et al., 2008). According to most studies, this marked temperature decline took place around *ca* 4900–3900 cal yr BP (e.g. Heiri et al., 2003; Larocque-Tobler et al., 2010; Ilyashuk EA et al., 2011), while our chironomid record shows this temperature drop with a delay, from *ca* 3300 cal yr BP onwards, and in parallel with a marked decrease in organic content (Figure 4). This seems to be in agreement with a stalagmite record published also from the Southern Carpathians, which suggests a “mid-Holocene peak” at *ca* 3300 cal yr BP followed by a temperature decrease (Constantin et al., 2007).

From *ca* 3000 cal yr BP onwards, chironomid-inferred T_{VII} remained under modern values by *ca* 1.8–1.9°C and fluctuated strongly at Lake Brazi. This time period is characterized by the highest lake level during the Holocene, which are inferred from the Brazi diatom-record at *ca* 2800 cal yr BP (Buczkó et al., 2013). Similarly, a cool climate and lake level rise with maximum precipitation at *ca* 2800 cal yr BP was indicated by multi-proxy data (Magyari et al., 2009b) and a testate amoebae record (Schnitchen et al., 2006) from the Eastern Carpathians, and from the Carpathian Basin at around 3000 cal yr BP (Jakab and Sümegi, 2007). This late Holocene period, characterized by gradual cooling of summer air temperatures, is also indicated in Western European chironomid records, but differs in its amplitude among sites. At Lake Brazi reconstructed summer air temperatures declined by 1.8–1.9°C (at *ca* 3300 cal yr BP), which is in accordance with the 1.8–2°C decrease published from the northern Swiss Prealps (Larocque-Tobler et al., 2010) and from the Central Eastern Alps (Ilyashuk EA et al., 2011) starting at *ca* 4900–3900 cal yr BP. At the same time, a smaller decrease (~0.4–1.0°C) was recorded in the Bernese Alps (Heiri et al., 2003) and in the Northern Apennines (Samartin, 2011).

Finally, chironomid-inferred T_{VII} showed an increasing trend again until present-day values in the last 2000 years at Lake Brazi. The diatom record from the same sediment sequence suggests a gradual increase in the trophic status of the lake associated with gradual shallowing (Buczkó et al., 2013). During the last 2400 years, the pollen records suggest increased precipitation and cooler summers from the Southern Carpathians, while trends are less consistent in the last 1500 years because of intensified human-impact (e.g. Feurdean et al., 2008; Magyari et al., 2009b).

Overall, the compliance of our temperature reconstruction with other proxy-records (i.e. pollen, macrofossil, stalagmite, charcoal, diatoms) from the region corroborates the reliability of the long term summer air temperature trends (Table 3) during the Holocene at Lake Brazi. The reconstruction diagnostic statistics, however, showed better reliability of inferred temperatures with NS-TF than with Sw-TF. The latter calibration dataset was characterized by a weaker analogue situation and weaker fit with reconstructed temperature during the early to mid-Holocene (Figure 5). Therefore, for samples older than 5000 cal yr BP, inferred temperatures are probably overestimated (*ca* 0.5–1°C) by the Sw-TF.

Conclusions

We present the first chironomid-based summer air temperature reconstruction from the South Carpathians (Lake Brazi, Retezat Mts.) for the entire Holocene. Our results denote that temperature is likely the most important driver of the chironomid assemblage changes during the time period investigated.

Summer air temperature inferences (using the Swiss and the merged Norwegian-Swiss transfer function) suggest a two-steps rise at the onset of the early Holocene: an initial rise from *ca* 11,500 cal yr BP, followed by a further increase between *ca* 10,200–8500 cal yr BP until reconstructed temperatures reached higher than present values. During the mid-Holocene, from *ca* 8500 cal yr BP, chironomid-inferred temperatures decreased slightly, however mean values were still above present day temperatures. Between *ca* 6000–3000 cal yr BP, cooler summers were inferred. Although, during the late Holocene, between *ca* 3000 and 2000 cal yr BP, a distinct decrease in July air temperatures occurred with reconstructed summer air temperatures under modern values. Finally, in the last 2000 years reconstructed temperatures showed an increasing trend at Lake Brazi.

The most distinct climatic changes in our reconstructions are the Holocene Thermal Maximum (HTM), which occurred earlier (between *ca* 9400–8900 cal yr BP), and the late Holocene summer temperature decline, which occurred with a delay (between *ca* 3000–2000 cal yr BP) at Lake Brazi (South Carpathians) than in the Western European chironomid records. We conclude that Holocene summer temperatures at Lake Brazi closely followed summer insolation changes during the Holocene, which might be explained by the high altitude of the study site and its distance from the North Atlantic region, where short-term climatic perturbations strongly influence the trend of the chironomid-inferred Holocene summer temperature changes.

Short-term declines (within the estimated error of prediction of the chironomid-based inferences) of *ca* 0.6–1.4°C in reconstructed temperatures were detected between *ca* 10,350–

10,190; 9750–9500; 8700–8500; 7600–7300; 7100–6900 and 4400–4000 cal yr BP. These intervals agree well with other proxy-records (diatom, pollen and stalagmite) from the Carpathian region and with chironomid records from Central and Western Europe. Generally, our reconstructed temperatures complied with other proxy records from the same sediment sequence and with other records from the Carpathian region and Western Europe.

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Figures and Table:

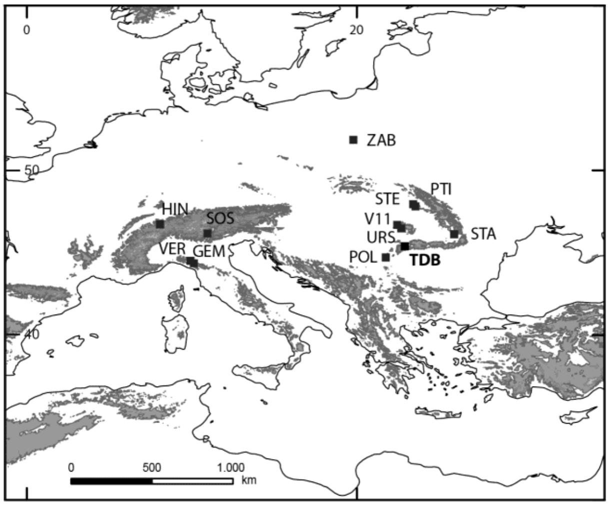

Figure 1. Location of Lake Brazi (Tăul dintre Brazi) in the South Carpathians and the selected records used for comparisons. Romania: TDB=Lake Brazi (this study); STA=Lake Saint Ana (Magyari et al., 2009b); STE= Steregoiu and PTI=Precula Tiganului (Feurdean et al., 2008); V11=V11 Cave (Tămaş et al., 2005); URS=Usilor Cave (Onac et al., 2002); POL=Poleva Cave (Constantin et al., 2007). Poland: ZAB=Zabienic bog (Płóciennik et al., 2011). Austria: SOS=Schwarzsee ob Sölden (Ilyashuk EA et al., 2011). Italy: VER=Lago Verdarolo and GEM=Lago Gemini, Italy (Samartin, 2011). Switzerland: HIN=Hinterburgsee (Heiri et al., 2003).

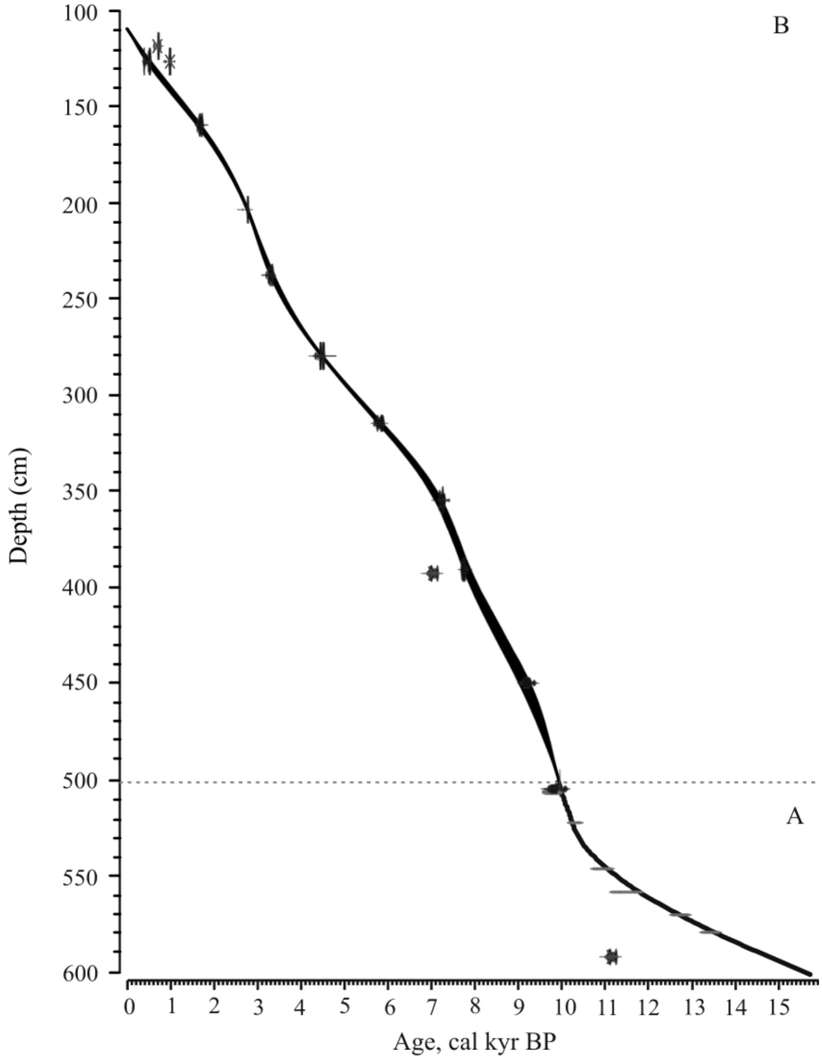

Figure 2. Age-depth models: (A) weighted non-linear polynominal regression for modelling the age-depth relationship in the late glacial and early Holocene part of the sediment record (600–502cm) and (B) a smoothing spline function in the Holocene part (505–111.14 cm) of the sediment record of Lake Brazi (Retezat Mts., South Carpathians).

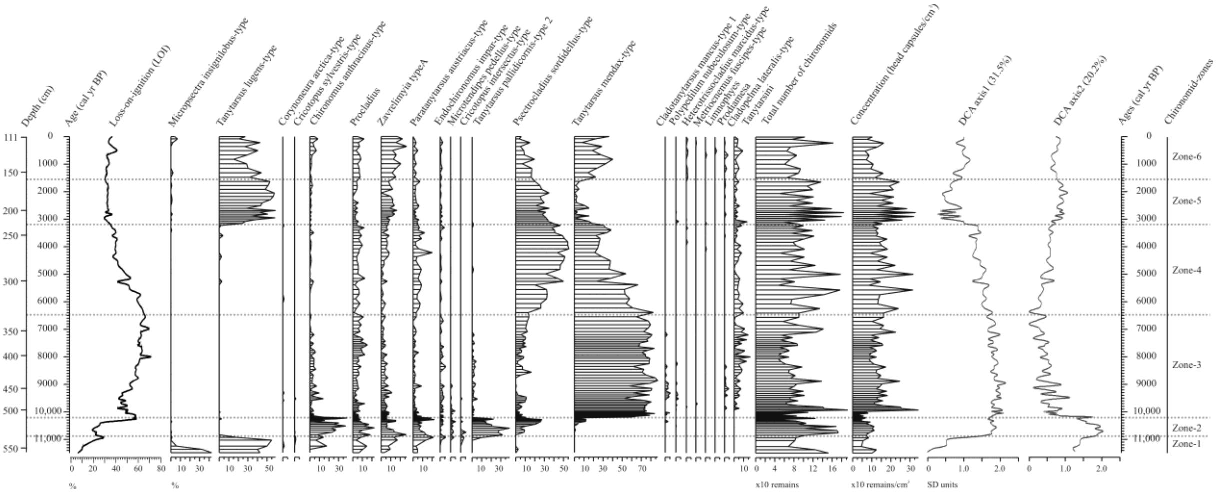

Figure 3. Chironomid relative abundance diagram including all chironomid taxa with Detrended Correspondence Analysis axes (DCA axis 1 and 2, with percentage variance explained by axes), loss-on-ignition (LOI) changes, and zones for the chironomid stratigraphy from Lake Brazi (Retezat Mts., South Carpathians).

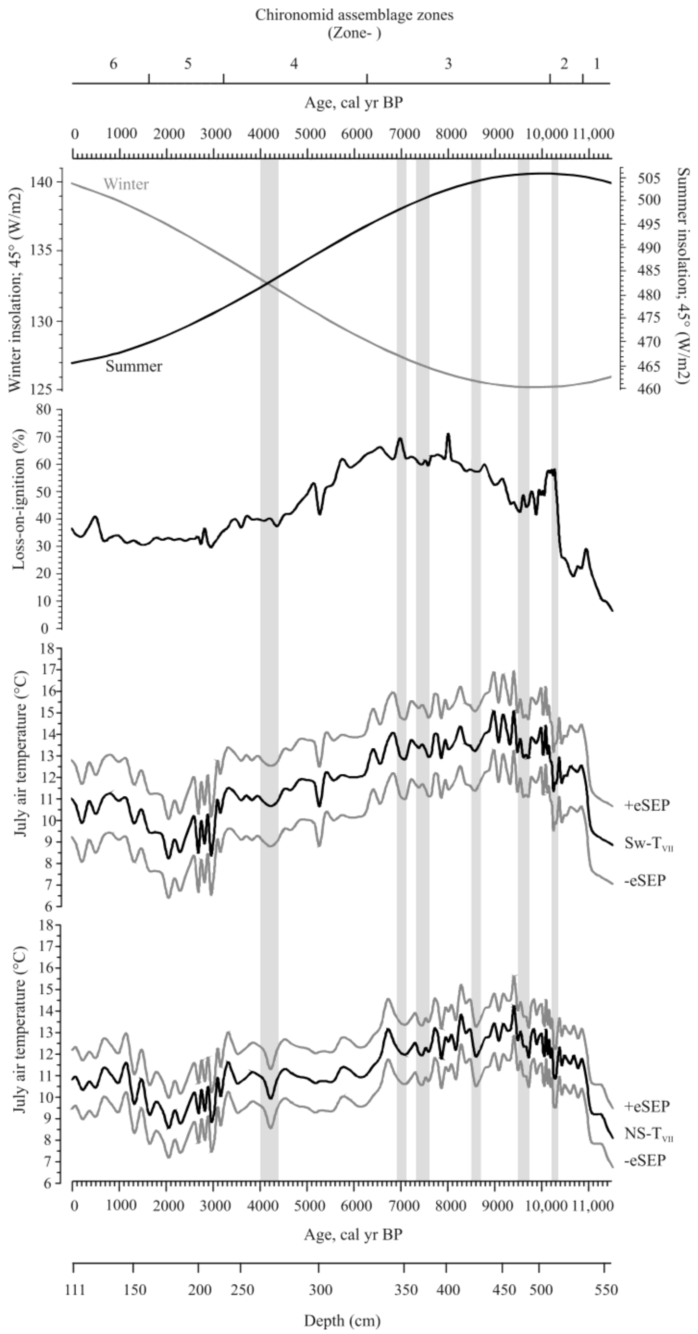

Figure 4. Chironomid-inferred July temperature estimates (black lines) based on the merged Norwegian-Swiss (NS-T$_{VII}$) and on the Swiss (Sw-T$_{VII}$) calibration datasets; with their sample-specific standard errors (eSEP; grey lines), loss-on-ignition values (%) from Lake Brazi (Retezat Mts., South Carpathians) as well as summer and winter insolation at 45°N (W/m^2; Laskar et al., 2004) and zones for the chironomid stratigraphy (Zone-). Grey bands correspond to the discussed temperature declines during the Holocene.

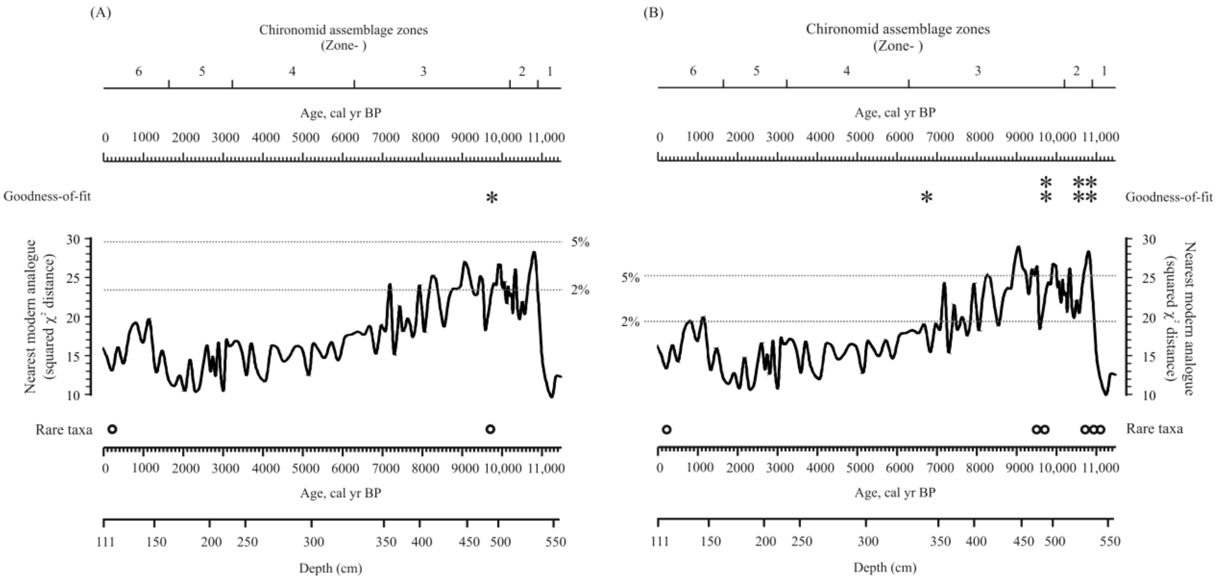

Figure 5. Reconstruction diagnostic statistics of the chironomid-inferred July air temperature reconstructions from Lake Brazi (Retezat Mts., South Carpathians) based on the (A) the merged Norwegian-Swiss (NS-T_{VII}) and on the (B) Swiss (SW- T_{VII}) transfer functions, plotted together with zones for the chironomid stratigraphy (Zone-). Diagnostic statistics include nearest modern analogues for the fossil samples in the calibration data set, goodness-of-fit statistics of the fossil samples with temperatures (one star indicate "poor fit" and double stars "very poor fit" with temperature) and samples with rare taxa (Hill's N2 <5; open circles).

Table 1. Radiocarbon dates from Lake Brazi. AMS ^{14}C dates were obtained from the Poznan Radiocarbon Laboratory, Poland (Poz-) and from the Hertelendi Laboratory of Environmental Studies at ATOMKI in Hungary (DeA-). Depths values are corrected for compression.

Core	**Laboratory code**	**Dated material**	**Depth (cm)**	**^{14}C age years BP**	**Calibrated range years BP**	**Error of the average years BP**	**Remarks**
TDB-1	Poz-26103	*Picea abies* needles	119	725±30	652-723		outlier
TDB-1	DeA-1237	>180 µm fraction, plant macrofossil	127	375±25	319-503	411 ± 92	
TDB-1	DeA-1238.1.2	>180 µm fraction, particular organic matter	127	1018±23	913-970		outlier
TDB-1	Poz-26104	*Pinus mugo* cone scale	160	1735±30	1562-1712	1637±75	
TDB-1	DeA-1239	*Pinus mugo* shoot	204	2611±23	2724-2763	2743,5±19,5	
TDB-1	Poz-206106	*Pinus mugo* cone	238	3045±30	3205-3356	3280,5±75,5	
TDB-1	DeA-1240	>180 µm fraction, plant macrofossil	280	3962±30	4381-4520		outlier
TDB-1	DeA-1241	>180 µm fraction, particular organic matter	280	3987±26	4416-4521	4468,5±52,5	
TDB-1	Poz-26107	*Pinus* twig	315	5040±40	5708-5902	5805±97	
TDB-1	Poz-26108	*Picea abies* needles	355	6320±40	7163-7324	7243,5±80,5	
TDB-1	DeA-1242	>180 µm fraction, plant macrofossil	391	6925±30	7683-7828	7755,5±72,5	
TDB-1	Poz-26109	*Picea abies* needles	393	6130±40	6926-7160		outlier
TDB-1	Poz-26110	*Picea abies* needles and seed	450	8240±50	9072-9326	9199±127	
TDB-1	Poz-26111	*Picea abies* needles	505	8810±50	9670-10 155	9912,5±245,5	
TDB-1	Poz-31714	*Pinus mugo* needles	521	9150±50	10 226-10 433	10 329,5±103,5	
TDB-1	Poz-26112	*Picea abies* cone	545	9610±50	10 766-11 167	10 966,5±200,5	
TDB-1	Poz-31715	*Pinus mugo* needles	557	9980±100	11 216-11 826	11 521±305	
TDB-1	Poz-31716	charcoal	569	10 870±70	12 598-12 925	12 761,5±163,5	
TDB-1	Poz-27305	*Pinus sp.* needles (2)	578	11 590±60	13 287-13 620	13 453,5±166,5	
TDB-1	Poz-26113	*Picea abies* cone scales	591	9690±50	11 067-11 225		outlier

Table 2. Main characteristics of the Swiss and the merged Norwegian-Swiss training sets and Lake Brazi (Retezat Mts., South Carpathians). July air temperature value at Lake Brazi means the present day July air temperature, estimated by linear interpolation from the five nearest meteorological stations (Bogdan, 2008). Model statistics based on bootstrapped cross-validation of chironomid-based temperature inference models with two WA-PLS components, where RMSEP is a root mean square error of prediction and r^2 is a coefficient of determination.

	Swiss calibration data-set	**Norwegian-Swiss calibration data-set**	**Lake Brazi**
Altitude (m a.s.l.)	418–2815	5–2815	1740
July air temperature (°C)	5.0–18.4	3.5–18.4	11.2
Water depth (m)	2.1–85	0.9–85	1.11
Model statistics			
No. of lakes included	117	274	
No. of outliers deleted	15	19	
Cross-validated			
RMSEP (°C)	1.4	1.39	
average bias (°C)	-0.02	-0.09	
maximum bias (°C)	0.97	1.44	
r^2	0.91	0.89	

Table 3. The chironomid inferred temperature (T_{VII} = July air temperature) and assemblage changes from Lake Brazi (Retezat Mts., South Carpathians) in comparison with changes in proxy-records from the Carpathians.