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Hydrological instability during the Last Interglacial in central Asia: a new diatom oxygen isotope record from Lake Baikal

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ABSTRACT

Last Interglacial variability is commonly used as an analogue for variability in a future, warmer world. Pervasive cycles are increasingly apparent in Last Interglacial archives, although studies in continental regions are under-represented. Here we provide a new isotopic record of diatom silica ($\delta^{18}O_{diatom}$) spanning c. 127.5–115 ka BP from Lake Baikal in central Asia. Peak rain-fed discharge occurred c. 125.4 ka BP, shortly after July insolation maximum and initiation of Siberian soil development. Between 127 and 119.7 ka BP there are six marked fluctuations in $\delta^{18}O_{diatom}$ values, with a pacing of approximately 1.26 ± 0.3 ka, similar to fluctuations of within lake productivity. Fluctuations in $\delta^{18}O_{diatom}$ values show good agreement with patterns in Atlantic meridional overturning circulation (AMOC), supporting hypothesis of strong teleconnections via the Westerlies between the North Atlantic and central Asia. Two periods of low $\delta^{18}O_{diatom}$ values are especially notable. The earliest between c. 126.5 and 126 ka BP is concurrent with the final stages of the Heinrich 11. The second between 120.5 and 119.7 ka BP is also concurrent with an increase in ice-rafted debris in the North Atlantic. Aquatic productivity in Lake Baikal increased between 119.7 and 117.4 ka BP before declining to the top of the record (115 ka BP) concomitant with a shift to predominately cool steppe catchment vegetation. However, isotopic composition of discharge into Lake Baikal provides evidence for strong penetration of Westerlies into central Asia during the latter stages of the Last Interglacial. Variability in $\delta^{18}O_{diatom}$ values was compared between the Last Interglacial and the Holocene. Millennial-scale variability was significantly more stable during the Last Interglacial, possibly linked to diminished influence of freshwater discharge on AMOC during periods of higher, global mean temperatures.

1. Introduction

Climate variability and its potential impact on humanity is one of society's greatest concerns, yet substantial uncertainty exists as to how the climate will vary in the future, especially within the context of a warmer world. Considerable work has been undertaken to address this uncertainty by investigating previous interglacials which were either similar to the Holocene in terms of orbital geometry e.g. Marine Isotope Stage (MIS) 19 (Tzedakis et al., 2009); MIS 11 (Berger and Loutre, 2003) or were warmer e.g. MIS 5e (Velichko et al., 1991; Kukla et al., 2002; CAPE Last Interglacial Project Members, 2006). Of key interest is the occurrence and tempo of instabilities in the climate system, such as pervasive 1.5 ka cycles that have been determined during the Holocene (Bond et al., 1997, 2001) the Last Glacial period (Bond et al., 1999), and the Last Interglacial (c. 130–116 ka BP) (e.g. Bond et al., 2001). If models are to accurately predict future climate, they need to incorporate these pervasive cycles, and validation is best done on previous interglacials where there has been no determinable anthropogenic activity.

Our knowledge of climate instability during previous interglacials is poor because of the paucity of highly resolved and well

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constrained records (Tzedakis et al., 2009). For example, whilst the Last Interglacial was previously considered a period of relative stability (e.g. Grootes et al., 1993; McManus et al., 1994; Rioual et al., 2001), many archives have since been shown to contain millennial scale variability (Bond et al., 2001) with cool (Müller et al., 2005; Couchoud et al., 2009) and sometimes abrupt events (Seidenkrantz et al., 1995; Linsley, 1996; Maslin and Tzedakis, 1996; Rioual and Mackay, 2005; Sirocko et al., 2005) linked to variability in Atlantic meridional overturning circulation (AMOC) (Bond et al., 1999, 2001).

The Last Interglacial can be broadly defined as the penultimate period when sea-levels were at, or above, those of the present (CAPE Last Interglacial Project Members, 2006). The majority of the Last Interglacial reconstructions have been mainly determined for Atlantic regions, and a key challenge is to determine whether AMOC variability had global influence, e.g. impacting continental regions far from oceanic influences. Lake Baikal in central Asia is therefore of great importance because it holds an uninterrupted sedimentary archive spanning millions of years, which has been shown to be very responsive to global changes in climate (see Mackay, 2007 for a review).

Where preservation of carbonates is poor in fresh, lacustrine systems, oxygen isotope analysis of diatom silica ($\delta^{18}O_{diatom}$) has been used as a direct replacement for $\delta^{18}O_{carbonate}$ e.g. tropical Africa (Barker et al., 2011), Alaska (Hu and Shemesh, 2003), Europe (Rioual et al., 2001; Shemesh et al., 2001; Rosqvist et al., 2004), South America (Polissar et al., 2006; Hernández et al., 2008) and in different regions of Russia (northwest: Jones et al., 2004; northeast: Swann et al., 2010; and southeast: Morley et al., 2005; Mackay et al., 2008, 2011). The majority of these studies span the Holocene period, although older records exist for MIS 5e in France (Rioual et al., 2001), the past 250 ka in NE Siberia (Chapligin et al., 2012), and MIS 11 in Lake Baikal (Mackay et al., 2008).

The $\delta^{18}O_{diatom}$ record from Lake Baikal has been shown to be an important proxy of hydrological variability in central Asia, linked to AMOC and intensity of the Siberian High during the Lateglacial–Holocene (Mackay et al., 2011). Using older sediment records from Lake Baikal, this study has two aims: to characterise hydrological instability during the Last Interglacial in central Asia, and to evaluate whether hydrological variability during the Last Interglacial was greater than Holocene variability.

2. Study area

The region of central Asia that includes Lake Baikal (Fig. 1) is characterized by the world's highest degree of continentality

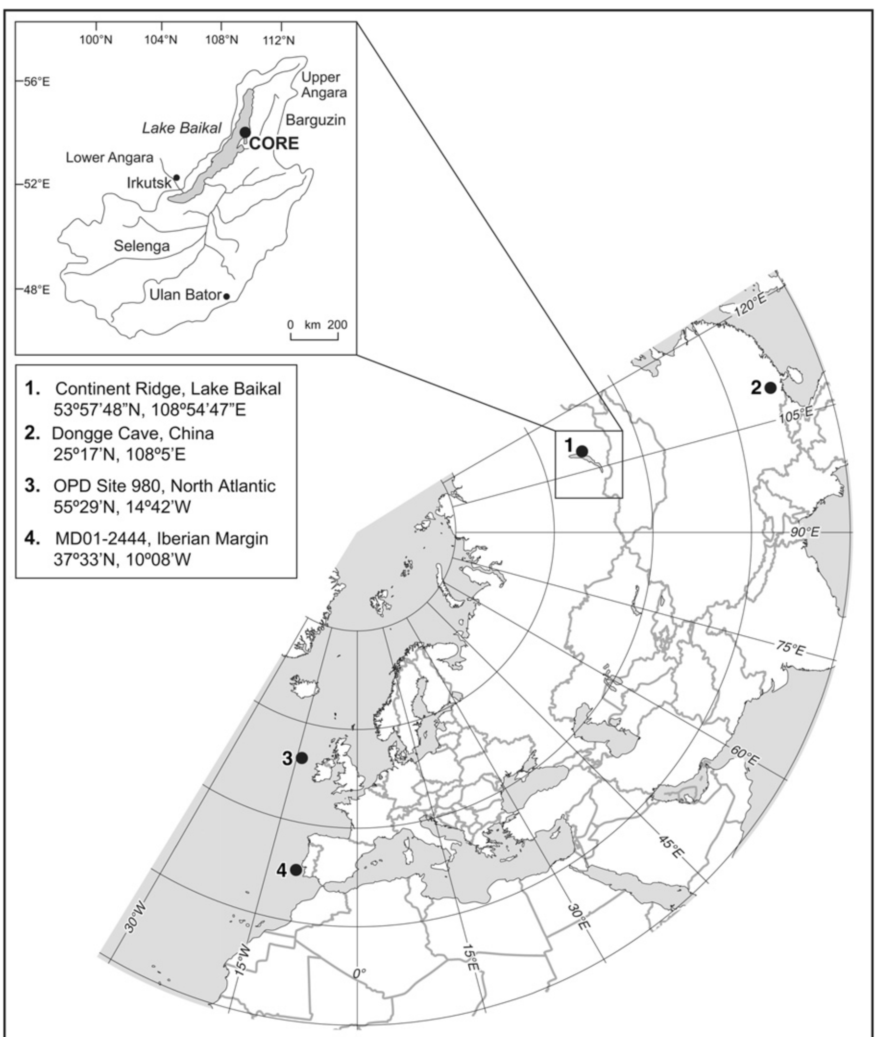

Fig. 1. Map showing location of Core CON-01-603-2 from Lake Baikal, and of other principal sites mentioned in the text (core 980 from North Atlantic; core MD01-2444 from the Iberian Margin; Dongge Cave, south China).

(Lydolph, 1977). Summers are short, warm and wet, while winters are long, cold and relatively dry. Rainfall and precipitation data are given for the nearby city of Irkutsk (Fig. 2). Lake Baikal itself moderates climate within the Baikal depression, resulting in cooler summer temperatures by between 6 and 8 °C and warmer temperatures during autumn and early winter by approximately 10 and 15 °C (Shimaraev et al., 1994). Water from the lake is mainly lost via the Angara River outflow, with evaporation only accounting for less than 20% water loss. Evaporation occurs mainly during the autumn and early winter (Sept–Dec), whilst during summer months evaporation is insignificant (Shimaraev et al., 1994). A strong westerly progression of cyclones moves through west Siberia to the Lake Baikal region during spring because of the intensification of zonal circulation. Low-pressure systems form along the Asiatic polar front in summer and as the strength of the westerly transport weakens, cyclonic activity and rainfall increases. Deep intrusions of cold Arctic air from the Kara Sea to the Lake Baikal region during autumn bring widespread cooling throughout eastern Siberia, which marks the beginning of the growth of the Siberian High. This high pressure cell remains strong during winter until the following April (Panagiotopoulos et al., 2005) and is

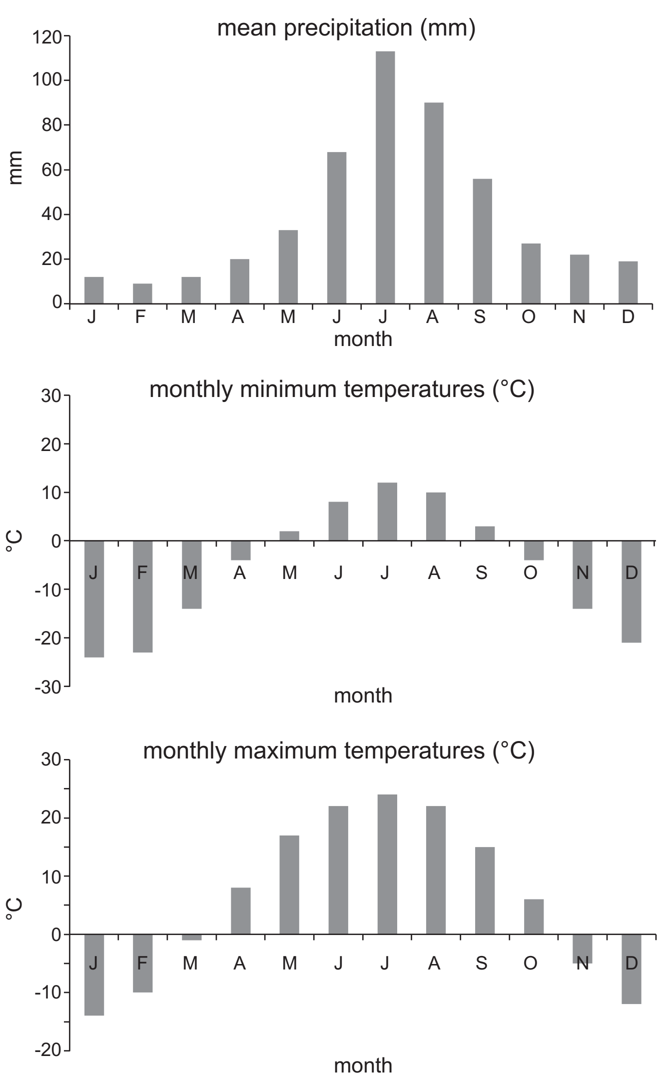

Fig. 2. Summary climate data for the city of Irkutsk, which lies on the Angara outflow, 70 km from Lake Baikal. (i) Monthly precipitation (mm); (ii) monthly temperature minima (°C); (iii) monthly temperature maxima (°C).

responsible for the movement of cold winter air flowing over Asia (Gong and Ho, 2002) resulting in a strong East Asian Winter Monsoon (EAWM) and attenuated East Asian Summer Monsoon (EASM) (D'Arrigo et al., 2005).

Lake Baikal is the world's most voluminous lake, containing approximately one fifth of global resources of surface freshwater (Shimaraev et al., 1993). It has over 330 tributaries, the most significant being the Selenga, Upper Angara and Barguzin Rivers, contributing c. 50%, 14% and 7% of total annual river inflow respectively (Fig. 1) (Shimaraev et al., 1994). The catchment of Lake Baikal is over 540,000 km^2, of which about half belongs to the Selenga River basin, that drains much of northern Mongolia and Buryatiya between 46–52° N and 96–109° E (Fig. 1) (Ma et al., 2003). The Barguzin River has a catchment east of the lake, while the Upper Angara drains a catchment to the northeast (Seal and Shanks, 1998). Due to their latitudinal differences, rivers are fed by varying amounts of precipitation and snowmelt.

3. Materials and methods

3.1. Site location, coring and dating

The core spanning the Last Interglacial was located in the north basin of Lake Baikal (53°57′ N, 108°54′ E) on an isolated high called the Continent Ridge (Fig. 1). Side-scan sonar and reflection seismic data show a flat, featureless morphology that suggests low tectonic activity and an undisturbed fine-grained sedimentation (Charlet et al., 2005). Core CON-01-603-2 was recovered in July 2001 in 386 m water depth using a piston corer equipped with 12 cm diameter aluminium liners. An independent chronology for the section was constructed by Demory et al. (2005), based on geomagnetic palaeointensities tuned to a well-constrained reference curve (ODP Site 984, Channell, 1999). Anchored by a geomagnetic excursion (the Iceland basin event, dated at 186–189 ka) this age model is constrained by 55 correlation points for a time span of ~200 ka. For the Last Interglacial period presented here, the palaeomagnetic-derived age model was constrained by five of these correlation points. Sediment accumulation rates varied through the Last Interglacial period; between c. 127.5 and 117.3 ka BP the rate was approximately 9.5 cm/ka, but declined to 4.0 cm/ka between c. 117.3 and 115 ka BP. Full details are given in Demory et al. (2005).

3.2. $\delta^{18}O_{diatom}$ analysis

Various methodologies for ensuring pure diatom samples for $\delta^{18}O_{diatom}$ analysis have been reported (Morley et al., 2004; Leng and Barker, 2006; Brewer et al., 2008; Mackay et al., 2011), and involve the step-wise removal of contaminants using chemical and physical separation techniques. Cleaned, dried samples were then subjected to a pre-fluorination process to remove the unstable hydrous silica layer from the diatom valves, before full reaction with BrF_5 (Leng and Sloane, 2008). Liberated oxygen was converted to CO_2 and measured alongside BFC_{mod} the NIGL diatom standard with $\delta^{18}O$ analysis performed using an Optima dual inlet mass spectrometer. The data are presented as per mil (‰) deviations from VSMOW with replicate analysis of sample material indicating an analytical reproducibility of ±0.34‰ (1 SD). The method has been verified through an inter-laboratory calibration exercise (Chapligin et al., 2011). Non-diatom components remaining after sample preparation were compensated for using a geochemical mass balance approach in which residual contaminants were calculated using XRF from the amount of Al_2O_3 in individual samples (Brewer et al., 2008; Mackay et al., 2011). The average isotope value of silt from Lake Baikal is +11.7 ± 0.3‰ (Brewer et al.,

2008) while Al concentrations in fully purified Last Interglacial diatom samples from Lake Baikal are 0.08% [1 SD = 0.02] (Swann, 2010). Although this level of diatom- bound Al typically only alters modelled $\delta^{18}O_{diatom}$ within the limits of analytical reproducibility (0.34‰) it nevertheless remains important to account for this contribution. Full details of mass-balance calculations are given in Mackay et al. (2011).

4. Results and discussion

The main controls on $\delta^{18}O$ in Lake Baikal waters include relative inputs from northern and southern basin rivers, atmospheric circulation changes, changes in seasonal precipitation, temperature dependent $\delta^{18}O$ in precipitation and evaporative enrichment (Seal and Shanks, 1998; Morley et al., 2005). Interpretation of the $\delta^{18}O_{diatom}$ record requires extensive knowledge of isotopic variability of the different hydrological inputs into Lake Baikal (Table 1). The $\delta^{18}O$ composition of lake water is the same across the length of the lake, highlighting that the waters are very well mixed ($\delta^{18}O = -15.9$ to -15.7‰) (Weiss et al., 1991). However, the $\delta^{18}O$ composition of rivers flowing into the lake is considerably higher for those with southern catchments, such as the Selenga (Seal and Shanks, 1998) because they are fed by a lower proportion of isotopically-lower snowmelt (Table 1). During periods of prolonged winters, increased snow cover extent results in a reduction in summer precipitation (especially over the Selenga catchment) through increased anticyclonic activity and the strength of the Siberian High (Lui and Yanai, 2002). Thus discharge from rivers with higher $\delta^{18}O$ values to the south of Lake Baikal decline in volume, while relative discharge from northern rivers increases (Morley et al., 2005). These findings indicate that $\delta^{18}O$ values are a weighted average of all input sources minus outputs (Morley et al., 2005), and therefore have important implications in terms of palaeoclimatic interpretations.

4.1. Confounding factors

Mackay et al. (2008) discuss the three main factors which can have important impacts on the $\delta^{18}O_{diatom}$ values, as applied to Lake Baikal sediments: (i) contamination from other oxygen-bearing minerals in the sediment; (ii) diatom dissolution; and (iii) vital effects from changing diatom communities. Brewer et al. (2008) and Mackay et al. (2011) discuss in detail the use of sample Al_2O_3 concentrations as an estimator of contamination. The vast majority of samples spanning the Last Interglacial period were extremely pure (containing between only 1–3% Al_2O_3). We are very confident therefore that any uncertainties around mass-balancing are minimal. Levels of non-diatom material prior to 128 ka BP are high (up to 75%), making robust interpretations uncertain (Fig. 3). We therefore focus our discussion on hydrological variability during the period between c. 127.5 and 115 ka BP. Dissolution of Lake Baikal diatoms at the surface sediment–water interface is a significant process, and only c. 1% of valves are eventually preserved in the sedimentary record (Ryves et al., 2003). Dissolution of diatom silica continues into the sedimentary pore-waters, until saturation is reached (Conley and Schelske, 1989; Carter and Colman, 1994). Despite diatom dissolution being a common process in freshwater ecosystems, very little is known about its impact on $\delta^{18}O_{diatom}$ values. Through experiments, Moschen et al. (2006) determined that $\delta^{18}O_{diatom}$ values could become enriched if dissolution occurred at high pH (pH 9.0) only, i.e. there was no significant change on $\delta^{18}O_{diatom}$ values at near neutral pH. As the pH of Lake Baikal is generally 7.1–7.2 (Votintsev, 1961), it is unlikely that dissolution will have had an effect on the $\delta^{18}O_{diatom}$ values observed. In comparison to some other organisms, the impact of vital effects of diatoms on $\delta^{18}O$ values is not well known. Of the few studies that have been carried out, the impact of differential species fractionation appears to be very limited (e.g. Shemesh et al., 1995; Swann et al., 2006), and within the range of reproducibility that can be achieved using fluorination techniques (Swann et al., 2007).

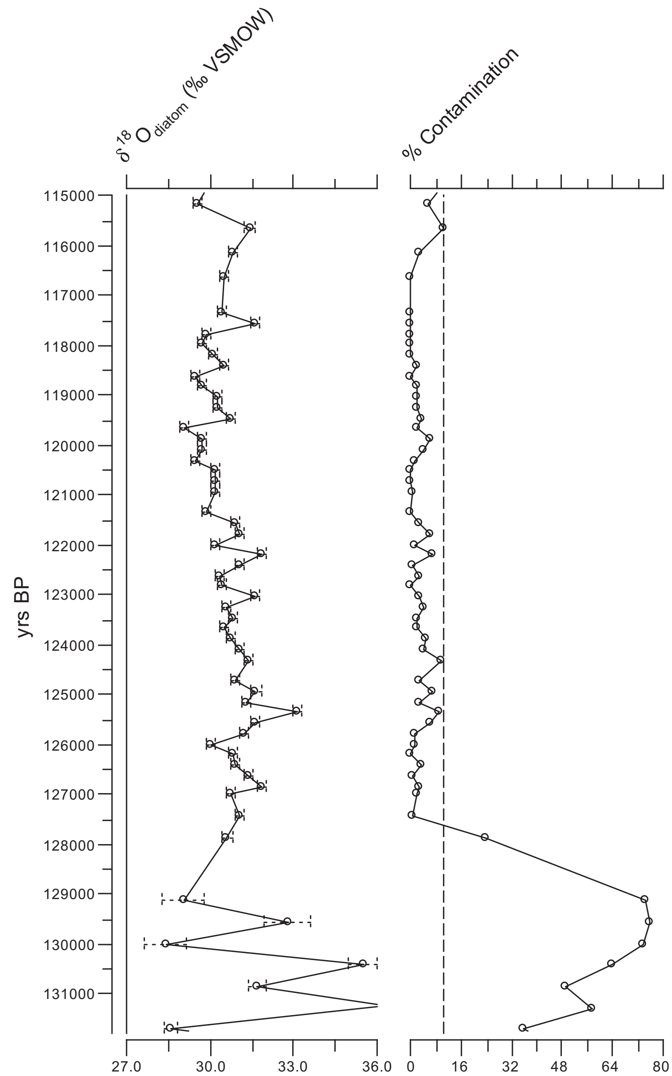

Fig. 3. Comparative plots for period 131.5–115 ka BP showing $\delta^{18}O_{diatom}$ and % silt contamination calculated from the amount of Al_2O_3 in individual samples. See text for details.

Table 1
Mean $\delta^{18}O$ and $\delta D \pm 2$ SD (‰ versus SMOW) values for rivers flowing into the south, central and north basins of Lake Baikal, and of lake water within those basins (Seal and Shanks, 1998[a]; Morley et al., 2005[b]). [1]Water samples taken from underneath the ice of frozen Lake Baikal.

Source	Date measured	$\delta^{18}O$	δD
South basin rivers[a]	1991–1992	−15.9 ± 4.9	−120 ± 31
Central basin rivers[a]	1991–1992	−17.6 ± 3.7	−132 ± 21
North basin rivers[a]	1991–1992	−20.4 ± 2.2	−151 ± 13
South basin[a]	June 1992	−15.8 ± 0.1	−123 ± 2
South basin[b]	April 2000[1]	−15.9 ± 0.1	−123 ± 2
South basin[b]	July 2000	−15.7 ± 0.1	−122 ± 1
South basin[b]	March 2001[1]	−15.8 ± 0.1	−125 ± 1
Central basin[a]	June 1992	−15.8 ± 0.9	−123 ± 3
Central basin[a]	June 1992	−15.9 ± 0.1	−123 ± 2
North basin[a]	June 1992	−15.8 ± 0.1	−124 ± 1

Although the diatom flora during the Last Interglacial is diverse, diatom biovolume is dominated throughout the period by large *Stephanodiscus grandis* valves (>100,000 μm^3) (Rioual and Mackay, 2005). Biogenic silica from other species is therefore relatively low; thus possible effects of changing assemblage composition on $\delta^{18}O_{diatom}$ values are likely to be low too.

4.2. Intra-interglacial variability

During the early stages of the Last Interglacial (127.5 − c. 124 ka BP), peak $\delta^{18}O_{diatom}$ values of +33.1‰ at 125.4 ka BP, are concurrent with peak July insolation in central Asia at 65 °N (Berger and Loutre, 1991), marked increase in North Atlantic SSTs and lowest global ice volume (Fig. 5) (Kukla et al., 2002). Boreal taiga forest rapidly expanded in the Lake Baikal region (Granoszewski et al., 2005), linked to significant increase in pollen-inferred annual precipitation, followed by increase in temperature of the coldest month (Fig. 4) (Tarasov et al., 2005, 2007). Chemical weathering was only recorded after 125 ka BP (Fagel and Mackay, 2008), corresponding to the initiation of Siberian soil chernozem development (Chlachula et al., 2004). High $\delta^{18}O_{diatom}$ and increasing C/N indicate a significant increase in discharge and allochthonous transport into Lake Baikal from the Selenga River, linked to a weak Siberian High (Velichko et al., 1991).

$\delta^{18}O_{diatom}$ gradually declined from the onset of the Last Interglacial to lowest level of +29.0‰ at 119.7 ka BP. Deviations in $\delta^{18}O_{diatom}$ from the mean progressively declined after 125.4 ka BP (Fig. 4). However, superimposed upon this trend are several marked fluctuations, similar to fluctuations in lake productivity (Fig. 4) (Fietz et al., 2007; Mackay, 2007), and further afield to variations in deep ocean circulation (Fig. 5) (Oppo et al., 2006). Notable minima occurred at 126.0 ka BP (+30.0‰), 124.7 ka BP (+30.9‰), 123.7 ka BP (+30.5‰), 122.6 ka BP (+30.4‰), 121.4 ka BP (+29.9‰). Pacing between these minima was approximately 1.26 ± 0.3 ka, which was less frequent than Last Interglacial IRD events in the North Atlantic (these occurred approximately every 1.48 ka; Bond et al., 2001), although values are within uncertainty limits. Millennial-scale variability has also been found to occur during the Last Interglacial in the form of European pollen-inferred cold events (Müller et al., 2005; Binka and Nitychoruk, 2011) and diatom variability in Ribains maar in the French Massif Central (Rioual et al., 2007).

4.3. Palaeoclimatic interpretation

The minimum in $\delta^{18}O_{diatom}$ values at c. 126 ka BP occurs at the same time as marked declines in pollen-inferred temperatures of the warmest and coldest months in the Lake Baikal region. Cooler temperatures are likely responsible for decline in diatom BVAR, and also for lowest C/N values for the complete sequence, which highlights the importance of autochthonous sources of carbon to the lake (Meyers, 1994). Low $\delta^{18}O_{diatom}$ values at c. 126 ka BP are also concurrent with an increase in IRD associated with the final stages of Heinrich 11 (Skinner and Shackleton, 2006), a decline in North Atlantic SST and AMOC (Oppo et al., 2006) (Fig. 5) and a proposed southern shift of the North Atlantic drift which resulted in a period of reduced rainfall in south-western France (Couchoud et al., 2009). Elsewhere, the $\delta^{18}O$ from Dongge Cave indicates a decline in precipitation linked to reduced summer monsoon intensity in southern China (Yuan et al., 2004), while a shift to more arid climate has been observed from the Greek Ioannina sequence between c. 127 and 126 ka BP (Tzedakis et al., 2003). Sun et al. (2012) determined that strengthened winter monsoon (i.e. increase in strength of the Siberian High) was caused by a slowdown of AMOC during the Last Glacial period, influencing the northern Westerlies. Given that cyclicity in monsoon variability spanned glacial–interglacial cycles, the decline in $\delta^{18}O_{diatom}$ values during the Last Interglacial was likely linked to the decline in AMOC influencing the Westerlies, which resulted in increased intensity of the Siberian High, and greater proportion of isotopically depleted snowmelt discharge into Lake Baikal.

The shift to lowest $\delta^{18}O_{diatom}$ values between 120.5 and 119.7 ka BP occurs at the same time as an increase in North Atlantic IRD, lower SSTs (Oppo et al., 2006) and a small increase in global ice volume (McManus et al., 2002) (Fig. 5), Shifting AMOC and concomitant increase in intensity of the Siberian High likely led to prevailing cooler regional temperatures (Tarasov et al., 2005) and the observed increase in proportion of snowmelt inflow to Lake Baikal, together with depressed primary productivity in the lake (Karabanov et al., 2000; Prokopenko et al., 2002; Rioual and Mackay, 2005; Fietz et al., 2007) (Fig. 4). This cool event has been observed elsewhere in many other records, including European lakes (Field et al., 1994; Rioual et al., 2001; Tzedakis et al., 2003; Sirocko et al., 2005) and Chinese archives which highlight a marked decline in summer monsoon intensity (Fig. 5) (Zhisheng and Porter, 1997; Yuan et al., 2004).

Cool steppe and tundra vegetation expanded in the catchment around Lake Baikal between 119.7 and 117.4 ka BP (Tarasov et al., 2005), although taiga biome still dominated the landscape. After the 120.5–119.7 ka BP cool event, diatom productivity in Lake Baikal increased, although at lower rates, until 117.4 ka BP (Fig. 4). During this period, $\delta^{18}O_{diatom}$ values increased overall from +29.0‰ at 119.7 ka BP to +31.6‰ by 117.6 ka BP, indicative of sustained isotopically higher hydrological input into the lake, concurrent with elevated allochthonous carbon (Fietz et al., 2007) and persistent high effective moisture (Tarasov et al., 2005). A shift in diatom assemblage composition occurred from large extinct *Stephanodiscus* species, which likely required strong spring mixing, to extant *Cyclotella* species that today are most abundant during autumn overturn (Rioual and Mackay, 2005). There were also notable $\delta^{18}O_{diatom}$ minima at 118.7 ka BP (+29.5‰), 118.0 ka BP (+29.7‰) and 117.4 ka BP (+30.4‰), which occurred more frequently than the sequence of minima earlier in the record. Proxy evidence for deep-water circulation is very poor at this time (Oppo et al., 2006) although in general periods of elevated rain-fed discharge into Lake Baikal occurred when AMOC was higher (Fig. 5).

The final stage of the Last Interglacial, leading to the MIS 5e/5d transition (117.4–115 ka BP) (Kukla et al., 2002; McManus et al., 2002), was complex with growth of northern ice sheets linked to declining insolation. Continental interior regions are especially sensitive to changes in declining insolation, with major falls in summer temperature predicted by energy balance models (Short et al., 1991). Around Lake Baikal, cool steppe vegetation dominated the landscape, with minor contributions from boreal trees and shrubs (Tarasov et al., 2005). Primary productivity within the lake also declined (Fig. 4), concurrent with a marked decline in North Atlantic SSTs (Oppo et al., 2006) (Fig. 5). Reconstructed temperatures showed that mean temperature of the warmest and coldest month remained low (Tarasov et al., 2005, 2007).

During the Last Interglacial, Gulf Stream currents extended past the Bering Sea, along the coast of northern Siberia as far as 140 °E (Velichko, 1984). Despite low insolation, full interglacial conditions took place in Nordic seas and high Arctic only after 118 ka BP (Van Nieuwenhove et al., 2011). In the North Atlantic at this time, AMOC strengthened considerably (Fig. 5) (McManus et al., 2002), concurrent with sustained input of rain-fed discharge into Lake Baikal, increased inputs of allochthonous carbon and very active catchment pedogenesis (Fagel and Mackay, 2008). Strong AMOC

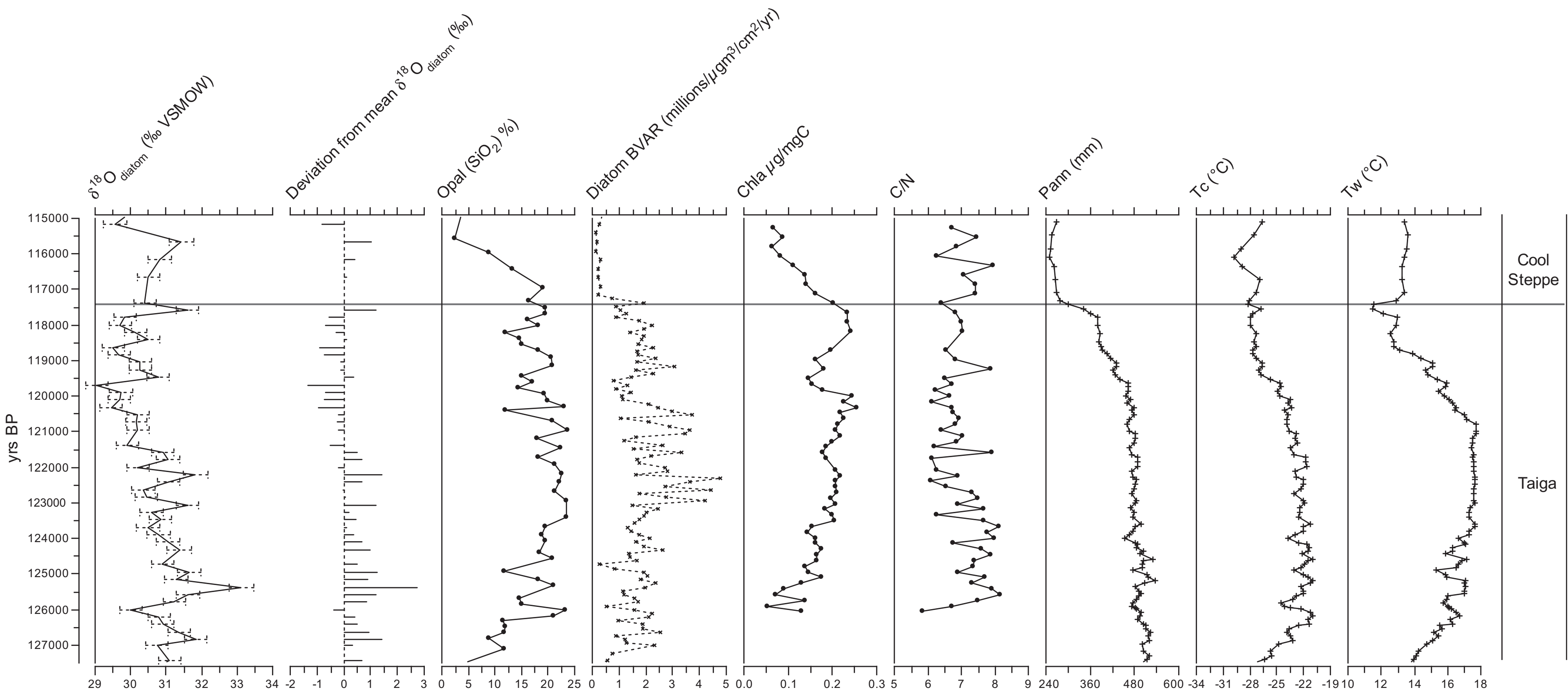

Fig. 4. Profiles of Lake Baikal indicators highlighted in the text: $\delta^{18}O_{diatom}$ profile with associated errors linked to mass-balancing isotope measurements; deviation from mean $\delta^{18}O_{diatom}$ values; % biogenic silica; diatom biovolume accumulation rates (BVAR) (millions valves/$\mu m^3/cm^2$/yr) (Rioual and Mackay, 2005) smoothed with locally weighted lowess regression; Chla/C and C/N ratios (Fietz et al., 2007); three-point moving averages of selected climate variables mean annual precipitation (Pann (mm)), mean temperature of the coldest month (Tc (°C)), mean temperature of the warmest month (Tw (°C)). Vegetation biomes follow Tarasov et al. (2007).

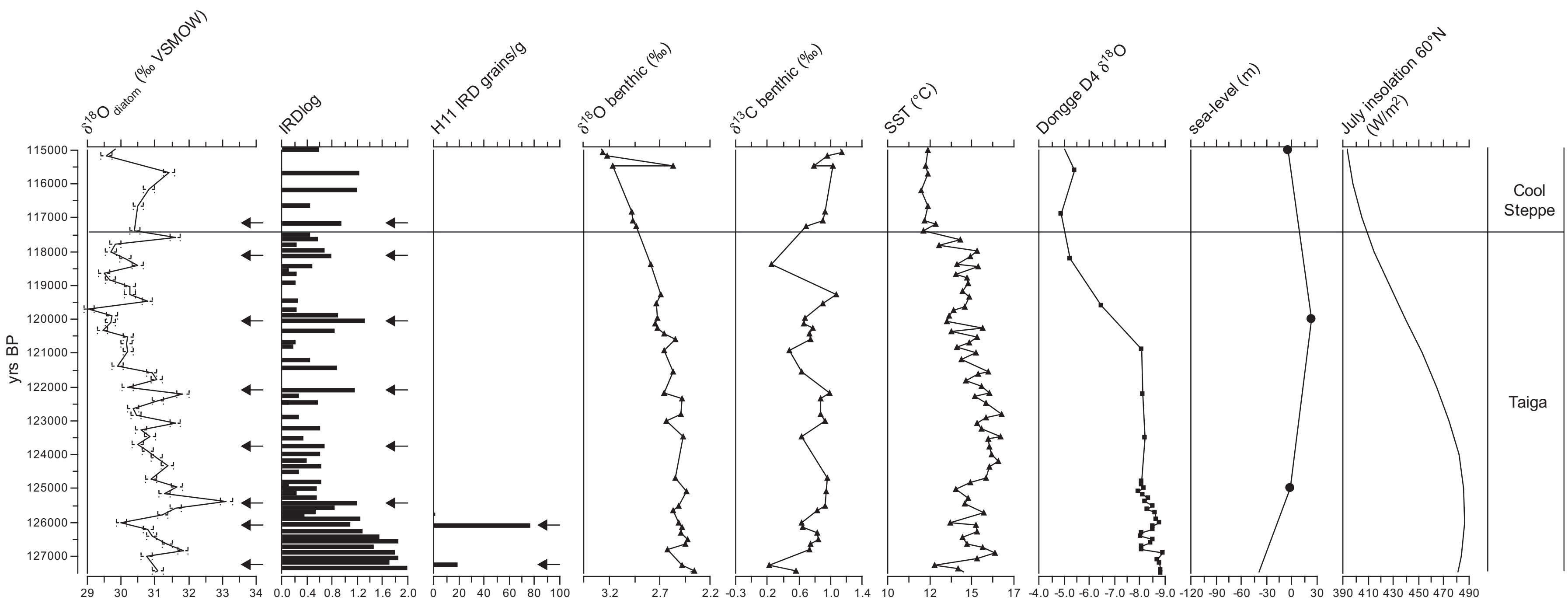

Fig. 5. $\delta^{18}O_{diatom}$ profile with associated errors linked to mass-balancing isotope measurements; lithic abundance data >150 μm, plotted on a logarithmic scale, from ODP Site 980 (Oppo et al., 2007); H11 event from IRD from core MD01-2444, retrieved from the Iberian Margin (Skinner and Shackleton, 2006); benthic $\delta^{18}O$ and $\delta^{13}C$ data from specimens of *Cibiciduides wuellerstorfi* >150 μm from ODP Site 980 (Oppo et al., 2007); SST estimates derived from modern analog technique from ODP Site 980 (Oppo et al., 2007); $\delta^{18}O$ Dongge Cave (Yuan et al., 2004); Sea level estimates (Miller et al., 2005); July insolation at 60 °N (W/m^2) (Berger and Loutre, 1991).

also resulted in strengthened summer monsoon intensity (Yuan et al., 2004; Fig. 5). At the MIS 5e/5d transition, glacial inception commenced in northwest Siberia because of the deep penetration of warm Atlantic currents across the northern coastline of Siberia and to declining temperatures linked to the MIS 5d insolation minimum (Karabanov et al., 1998). Thus while summer temperatures were very cold, leading to changes in catchment vegetation, moisture was still being transported to the region.

4.4. Comparison of interglacial hydrological variability

The Last Interglacial is not a terribly good analogue for Holocene climate because of differences in its orbital setting and insolation trends in particular (Berger and Loutre, 2003). In the Lake Baikal region, these differences contributed to consistently warmer pollen-inferred temperatures and elevated precipitation in the Lake Baikal region between 128 and 119 ka BP in comparison to the Holocene (Tarasov et al., 2007). A key question to ask therefore is whether differences in hydrological variability also existed between these two time periods.

Bond et al. (1999) proposed that increased freshwater discharge to the North Atlantic from glacier melt during interglacials were not enough to push circulation over a threshold into a new state. Thus pervasive cycles during the Holocene, and other long interglacials, were diminished in comparison to glacial periods. Here we extend this analogy and test the hypothesis that isotopic variability during the Last Interglacial in Lake Baikal was also diminished in comparison to the Holocene, because of higher global mean surface temperatures (Otto-Bliesner et al., 2006). Isotopic records were initially synchronized by aligning precession maxima (10.1 ka BP Holocene; 126.0 ka Last Interglacial). The Holocene record between 10.1 ka BP and AD 1910 (34 samples) was compared with the Last Interglacial period between 126.0 and c. 115.9 ka BP (43 samples) (Fig. 6). $\delta^{18}O_{diatom}$ values during the Last Interglacial in Lake Baikal were significantly higher, with less spread and uncertainty than during the Holocene (Fig. 6; Table 2). Both records show distinct millennial-scale variability. However, while fluctuations during the Holocene frequently exceeded 1‰, this was rarely the case during the Last Interglacial, except prior to 129 ka BP and after 118 ka BP. These data clearly demonstrate that Lake Baikal hydrology was much more stable during peak Last Interglacial than during the Holocene. Furthermore, actual minima in $\delta^{18}O_{diatom}$ during the Last Interglacial were all higher than Holocene $\delta^{18}O_{diatom}$, indicative of sustained isotopically-higher discharge into Lake Baikal. This concurs with persistent high regional precipitation and effective moisture in comparison to the Holocene (Tarasov et al., 2007).

Hydrological variability has previously been investigated for an MIS 11 sequence from Lake Baikal (Mackay et al., 2008). Unfortunately MIS 11 isotope data cannot be directly compared to the data from MIS 5e and MIS 1 as improvements in the handling of the effects of contamination were only made during the work on the latter two sections. Considering MIS 11 by itself, relative changes highlighted at least one major decline in $\delta^{18}O_{diatom}$ values between 396 and 390 ka BP, which was shown to corroborate decreases in primary productivity (Prokopenko et al., 2010) as well as reductions in greenhouse gas concentrations (Spahni et al., 2005), major increases in North Atlantic IRD (McManus et al., 1999) and the penetration of polar foraminifera assemblages off the Iberian Peninsula (de Abreu et al., 2005). Thus abrupt events in AMOC during previous interglacials also likely influenced Lake Baikal hydrology. Although $\delta^{18}O_{diatom}$ analyses have yet to be undertaken on other Lake Baikal interglacial sequences, during the Last Glacial period Heinrich Events were shown to have impacted lake productivity during MIS 3 (Prokopenko et al., 2001; Swann et al., 2005). Because Heinrich events are closely coupled with Dansgaard–Oeschger cycles they are also an integral component of the pervasive 1–2 ka cycle (Bond et al., 1999). We believe therefore that there is substantial evidence for variations in AMOC impacting on central Asian environments during both glacial and interglacial states, and that these cycles may be a persistent feature of Quaternary climates in general. This supports the idea that modern-day strong climatological teleconnections between North Atlantic/Europe and Siberia are a permanent feature of the Earth's climate state.

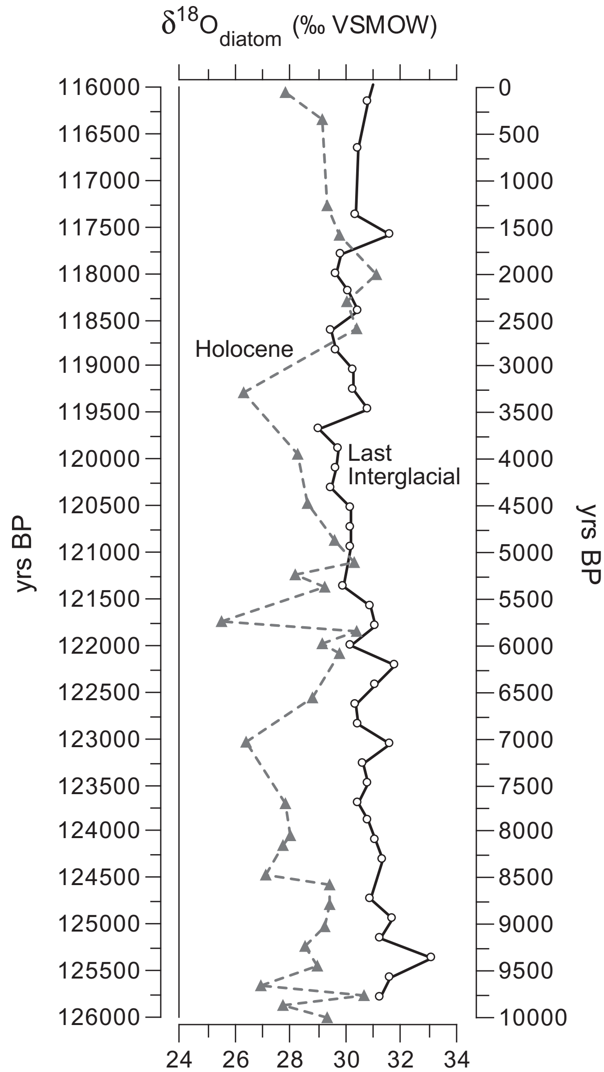

Fig. 6. Aligned $\delta^{18}O_{diatom}$ isotopic reconstructions of the Last Interglacial (this study; solid line) and the Holocene (Mackay et al., 2011; dashed line). Profiles are only shown for c. 10 ka, and synchronized by aligning Holocene and Last Interglacial precession maxima at 10.1 ka BP and 126.0 ka BP, respectively.

Table 2
Measures of central tendency and spread of $\delta^{18}O_{diatom}$ data from Last Interglacial (this study) and the Holocene (Mackay et al., 2011). Differences between the dataset were tested for significance using the non-parametric Kruskal–Wallis test (chi-square = 34.086; $p = 0.000$).

	LIG full	LIG 10 ka	Holocene 10 ka
Mean	30.9	30.4	28.7
SD	1.4	0.65	1.34
Range	7.9	2.8	5.6

5. Conclusions

Isotopic evidence for hydrological input into Lake Baikal shows that discharge was sensitive to variations in AMOC during the Last Interglacial. Increased influence of isotopically lower snowmelt occurs with a periodicity similar to Bond events, with an early event linked to the final stages of Heinrich Event 11. Therefore it is apparent that the sediments of Lake Baikal are sensitive indicators of millennial-scale climate variability during interglacial periods. In general, there is very good concordance between hydrological variability and aquatic productivity in the lake, until the latter stages of the interglacial, after 117.4 ka BP. Prevailing cooler climate resulted in reduced productivity, but persistent strong penetration of Westerlies into central Asia resulted in sustained input into Lake Baikal from rivers such as the Selenga. Hydrological variability was statistically lower during much the Last Interglacial than the Holocene, which was likely linked to globally warmer temperatures. Tentative evidence exists for millennial-scale variability during earlier interglacials, and future work should focus on those that have similar orbital configurations to the Holocene and the near future so that potential impacts of future climate change can be robustly modelled.

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