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Reconstructing hydrological variability in Lake Baikal during MIS 11: an application of oxygen isotope analysis of diatom silica

Short title: Reconstructing hydrological variability in Lake Baikal during MIS 11

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Abstract

In this paper we reconstruct hydrological variability in Lake Baikal during MIS 11 (427 – 362 ka BP) from oxygen isotope analysis of diatom silica. Highest $\delta^{18}O_{diatom}$ values are found during MIS11.3, highlighting the dominance of hydrological input from rivers flowing into the south and central basins of Lake Baikal, especially the Selenga River. Hydrological input from south basin rivers dominated for over 30 ka. However, there is evidence from both biogenic silica and $\delta^{18}O_{diatom}$ records for an abrupt cooling event at c. 390 ka BP. Stadial conditions at this time are coincident with an iceberg discharge event into the North Atlantic. The decline in $\delta^{18}O_{diatom}$ values suggests increasing proportion of hydrological input from rivers to the north of Lake Baikal, due to greater influence of winter precipitation and snow-melt. After a period of interstadial conditions during the early stages of MIS11.1, biogenic silica and $\delta^{18}O_{diatom}$ values decline, mirroring the slow growth in northern hemisphere ice sheets. Despite rigorous cleaning procedures, palaeoclimatic inferences need to be treated with caution due to contamination of the $\delta^{18}O_{diatom}$ record; during stadial and glacial periods, contamination of the $\delta^{18}O_{diatom}$ record is even more significant.

Keywords

MIS11; palaeoclimate; Lake Baikal; $\delta^{18}O_{diatom}$; biogenic silica

1. Introduction

Marine Isotope Stage 11 (MIS11) is increasingly being investigated to provide clues as to the duration of the current interglacial (Forsström, 2001), and if we can expect any surprises in the form of increasing instability in the future climate system. This is largely because astronomically-driven insolation and low amplitude of eccentricity between c. 423 ka – 362 ka are similar to variation during the latter part of the Holocene (Howard 1997) and to variation modelled for the near future (Loutre & Berger, 2003). MIS11 encompasses c. 2.5 precessional cycles, similar in length to subsequent isotope stages MIS9, MIS7 and MIS5. However, full interglacial conditions during MIS11 persisted for a significantly longer period of time than subsequent interglacials (i.e. MIS9.3, MIS7.5 and MIS5.5) (e.g. McManus et al., 1999; Droxler & Farrell, 2000; de Abreu et al., 2005). The extent to which climates during MIS11 interglacial were warmer than the Holocene is regionally complex. For example, sub-polar marine records of SSTs off the coast of Ireland were at their highest during MIS11 (McManus et al., 1999; Droxler & Farrell, 2000). Further north in the Nordic Seas however, foraminifera evidence indicates that SSTs were cooler than the Holocene (Bauch et al., 2000). Recent ice–core data obtained by the EPICA community members from Dome C in Antarctica (Spahni et al., 2005) highlight that δD values (indicative of Antarctic temperatures) were highest between 410 – 402 ka BP.

Many European terrestrial records indicate that the MIS11 interglacial was very humid (see Rousseau, 2003 for a review). This suggests that influence from the Westerlies was significant and at least as important as the present. Quaternary loess-soil records in China indicate that paleosol S4 (which is equivalent to MIS11) was one of the best developed throughout the last 1.2 Ma, linked to strengthened summer East Asian Monsoon (Guo et al. 2000). In central Asia, biogenic silica records from Lake Baikal (which act as a proxy for lake primary productivity) mirrored the rapid decline and slow growth of northern hemisphere ice sheets (Karabanov et al., 2003). Further, the proportion of biogenic silica in Lake Baikal sediments at this time reached some of their highest values for any interglacial during the last c. 450 ka, again highlighting that prevailing conditions were comparatively warm and humid.

Lake Baikal is an important location for studies of past climate change because it lies in a region that is sensitive to changes in precessional forcing, which has brought about major changes in the lake's ecosystem (Khursevich et al. 2000). Aspects of these past ecosystems are subsequently stored in its long, continuous sedimentary record (Williams et al. 1997; Prokopenko et al. 2002). Diatoms and associated biogenic silica (BioSi) are the most extensively utilized palaeoclimatic proxies found in Lake Baikal sediments, and arguably the most important (see Mackay 2007 for a review). Other sources of biogenic silica do exist (e.g. from chrysophytes and sponge spicules) but they form a relatively minor component (less than c. 2%; Granina et al. 1992). Morley et al. (2005) and Kalmychkov et al. (2007) sought to exploit the technique of oxygen isotope analysis of diatom silica, thereby providing a complementary, yet independent proxy climate signal to either diatoms or BioSi. This technique is proving valuable in the context of Lake Baikal palaeoclimate studies because carbonates and organic matter are rare or preservation potential is low, making oxygen isotope analysis of these sedimentary components unviable.

The aim of this study is to characterise hydrological variability in Lake Baikal during MIS11 through the application of $\delta^{18}O_{diatom}$ analysis at a millennial-scale resolution, thereby providing an independent proxy climate signal. Our data suggest distinct hydrological variability in central Asia during MIS11, against a backdrop of prevailing warmer temperatures that persisted for at least 30 ka.

Regional setting

The region of central Asia that includes Lake Baikal is characterized by the world's highest degree of continentality (Lydolph, 1977). Throughout spring, as zonal circulation intensifies, there is a strong westerly progression of cyclones moving through west Siberia to the Lake Baikal region. In summer, low-pressure systems form along the Asiatic polar front and as the strength of the westerly transport weakens cyclonic activity and rainfall increases (Fig 1). During July, average daily air temperatures are c. +19°C. In autumn, deep intrusions of cold arctic air progress from the Kara Sea to the Lake Baikal region, bringing widespread cooling throughout eastern Siberia. This marks the beginning of the growth of the Siberian High, a high-pressure cell which dominates this region of central Asia during winter (Gong & Ho 2002; Panagiotopoulos et al. 2005) (Fig 1). In January, air temperatures fall to an average of c. –25 °C. Characteristically, summers are short, warm and wet, while winters are long, cold and dry.

Although Quaternary glaciations have had major impacts on Lake Baikal's hydrology, sedimentology, ecosystem and shoreline, the bottom sediments have never been directly glaciated (Grosswald & Kuhle, 1994). Lake Baikal contains therefore, a potentially uninterrupted palaeoclimate archive consisting of over 7,500 m of sedimentary deposits, extending back more than 20 million years (Williams et al. 2001). To the east and west of

Lake Baikal itself are large, steeply sloping mountain ranges (e.g. Khamar-Daban and the Primorsky mountain ranges), broken only occasionally by valleys and deltas belonging to some of the larger rivers. The physical catchment of Lake Baikal occupies an area of c. 540,000 km^2, spanning south-eastern Siberia and northern Mongolia. Within this catchment, more than 300 rivers flow into Lake Baikal. The largest and most significant rivers are the Selenga, Upper Angara, and Barguzin Rivers. The Selenga is the largest river and delivers just under 50% inflow to the lake. The Upper Angara and Barguzin rivers contribute to c. 13% and 6% of the total annual river inflow respectively (Granina, 1997). The Angara River in the south basin accounts for c. 81% of outflow from the lake, while remaining water loss is through evaporation.

2. Methodology

The material used in this study was taken from the Baikal Drilling Program (BDP) core BDP-96-2, located on the Academician Ridge, an underwater mountain range which separates the central and northern basins of Lake Baikal (Fig 2). The 5 m sediment section presented here spans the latter part of the MIS12 (21.0-19.92 m), the complete MIS11, and the first section of the glacial period MIS10 (16.0-17.0 m) (Fig 3). Karabanov et al. (2003) have previously described methodologies for the construction of the age model, together with biogenic silica and diatom concentrations. Briefly, the age model was determined by correlation with the marine isotope curve based on palaeomagnetic reversals (Williams et al., 1997) with refinement through tuning to 65^{o} N insolation record (Prokopenko et al., 2001). After tuning to the insolation record, every interstadial precessional peak in BioSi could be recognised in the Brunhes sequence

(Prokopenko et al. 2001) – we are thus confident that the sequence presented here is that of MIS11.

Biogenic silica analyses were undertaken using a wet-alkaline extraction method (Mortlock & Fröhlich, 1989), with the blue silica molybdate complex determined colorimetrically. BioSi results are presented as % dry sediment weight. By tuning biogenic silica (BioSi) records to insolation at 65 ºN, the Lake Baikal MIS11 record spans approximately 65 ka years (427 - 362 ka BP) (Karabanov et al., 2003). Diatoms were counted in transects across smear slides (sampled every 2 cm, at a resolution of c. 500 yrs) using light microscopy at x 1000 magnification (Karabanov et al., 2003). Counts are presented as x 10^6 valves per gram sediment. The abundance of rare taxa has been exaggerated by a factor of 50 in Fig. 3.

Here we focus on the methodology used to prepare and clean sediment samples from BDP-96-2 for $\delta^{18}O_{diatom}$ analysis because contamination from silt and clay particles can significantly influence the $\delta^{18}O_{diatom}$ signal (e.g. Morley et al. 2004; Brewer et al. this volume). The sample cleaning method mainly follows stages developed by Morley et al. (2004), although separation of diatoms and silt using the heavy liquid sodium polytungstate did not improve the amount of diatom material obtained from these core samples. Stage 1: pairs of neighbouring samples were amalgamated from BDP-96-2 to provide enough diatom silica for $\delta^{18}O_{diatom}$ analysis (resolution of c. 1800 yrs). Organic material was removed from each sample by heating in 30% H_2O_2 at 90^{o}C for c. 2 hours. After washing, carbonates were removed by leaving samples overnight in c. 40 ml of 5% HCl. Stage 2 involved the removal of clay particles by sieving samples through a 10 μm sieve cloth, while most silt material was removed using a 75 μm sieve cloth. The 10 μm –

75 µm fraction was retained as it was found to maintain the highest proportion of diatoms to silt particles after trial with different sieve sizes. Samples were then dried at 40^{o}C for 24 hours ready for oxygen isotope analysis. In order to remove the unstable, diatom hydrous silica layer before full reaction with BrF_5, purified diatom samples were first subjected to a prefluorination step. Liberated oxygen was converted to CO_2 and measured alongside standard laboratory quartz and a diatom control sample. A dual inlet mass spectrometer was used to measure $^{18}O/^{16}O$ ratios; $\delta^{18}O$ values were normalised through international standards. The data are presented as per mil (‰) deviations from SMOW, and sample standard deviations were of the order of ±0.3‰ (1 σ).

We used a mass balancing method ((%diatom x isotope value) + (%silt x isotope value)) to compensate for the effect of silt contamination (Morley et al., 2005). To do this, a small amount of cleaned sediment was prepared for microscopy prior to final drying as above, so that the degree of silt remaining in each sample could be assessed at x1000 magnification using a 10 x 10 grid graticule (Morley et al. 2004). For each sample, diatom and contaminant content were estimated semi-quantitatively as a proportion of the non-empty graticule grid squares counted. The $\delta^{18}O$ value of Lake Baikal silt was taken to be 12.3‰, calculated as an average of rock fragments and silt after all diatoms were removed from sediment samples (ibid.). Bulk $\delta^{18}O$ values were taken as a linear mixture of oxygen from the silt and diatoms, so that a value for pure diatoms could be calculated using the estimated percentage content of diatoms and silt. We acknowledge that this method may underestimate the effect of silt $\delta^{18}O$ on overall oxygen isotope composition because volumes of diatoms and silt particles are not taken into account (Leng & Barker, 2006). More recently, geochemical methods have been employed to remove the effects of contaminants from diatom oxygen isotope records (e.g. Lamb et al. 2007). While these techniques show great promise (Brewer et al. this volume), they were not developed for our purpose at the time of sampling processing presented here.

3. Results

The glacial stages in this sequence (MIS12 and MIS10) are characterised by fine silty clays with inclusions of coarse silt with sand (lenses and pockets) and pebbles (Karabanov et al. 2003) (Fig 3). Such coarse sediments result from valley glaciers reaching the lake itself, and associated production of erosional material from glaciers (Karabanov et al., 1998). BioSi concentration during MIS12 is low, c. 5%. Diatoms are almost completely absent, with rare occurrence of the extinct endemic, *Stephanodiscus binderanoides* (which does not appear again in the this sequence), and some benthic taxa, between c. 440 ka - 435 ka BP. There is no concurrent increase in BioSi at this time, most likely due to dissolution processes, which can impact on the integrity of the BioSi signal in slowly accumulating glacial sediments (Swann & Mackay 2006). Where preserved, $\delta^{18}O_{diatom}$values approximate $\delta^{18}O$ values for silt without any diatoms, and the proportion of silt contamination to diatom counts is at its highest level for the whole profile (Fig 3).

MIS11 can be delimited into 3 stages: MIS11.3 (427 ka -414 ka), MIS11.2 (414 ka - 394 ka) and MIS11.1 (394 ka – 362 ka) . MIS11.3 lasts for c. 13 ka. At the base of the sequence there is a sharp increase in both BioSi and diatom concentration. The diatom flora is dominated by the extinct endemic diatom *Stephanodiscus distinctus* v. *distinctus*. Peak abundance of this species occur between 422 – 418 ka BP. During this time, extant endemic species, e.g. *Cyclotella minuta*, *Synedra* spp., *A. skvortzowii* and benthic taxa are also present at low abundance. The first measured $\delta^{18}O_{diatom}$ values are at c. 423 ka BP of +23.3‰, which exhibit a rising trend thereafter to peak values of +27.7‰, the maximum value for the sequence at c. 420 ka BP. Corresponding silt contamination is very low, at < 5%. Towards the end of MIS11.3, there is a shift in diatom composition, as *S. distinctus* is replaced in dominance by another endemic species, *Stephanodiscus exiguus*. The abundance of *Synedra* and benthic taxa also show small increases. These increases are concomitant with some of the highest diatom concentrations in the sequence, while BioSi fluctuates between 23-35%. $\delta^{18}O_{diatom}$ values however, decline to c. +22‰. This decline of c. 6‰ units is substantially larger than the standard deviation of the samples (Table 1).

MIS11.2 lasts for approximately 20 ka, between 414 – 394 ka BP. The start of the stage is marked by a rapid, terminal decline in *S. exiguus*. This decline is associated with a decline in overall diatom concentration back to < 240x10^6 valves / g dry wt. Also of note is a rapid increase in abundance of *Stephanodiscus binderanus*, which is likely to be conspecific with the extant endemic *S. meyerii*. In this sequence this species only occurs during MIS11.2. From c. 410 ka BP, there is an increasing trend towards higher abundance of *S. distinctus v. distinctus* and *C. minuta*. The associated increase in diatom concentration and BioSi reaches a peak for the sequence, of c. 570 x 10^6 valves and 45% respectively. $\delta^{18}O_{diatom}$ values also increase to +26.4‰ at c. 407 ka BP, coincident with increases in BioSi. Between c. 407 ka BP to c. 398 ka BP, $\delta^{18}O_{diatom}$ values decline to 22.0‰, concomitant with an increase in contamination between c. 5% - 32%. However, during the very latter stages of MIS11.2, $\delta^{18}O_{diatom}$ values increase once more to +26.1‰, with very low levels of contaminating silt (<5%).

The boundary between MIS11.2 and MIS11.1 occurs at c. 394 ka BP, and is marked by sharp declines in both BioSi values and diatom concentration. These declines are concomitant with declines in *S. distinctus*, although *C. minuta* actually increases to its highest abundance of just over 100 $x10^6$ valves / g dry wt sediment at c. 392 ka BP. The date of 390 ka BP is significant in this profile, as it marks the time when BioSi levels were at their lowest throughout the whole profile. There is also a marked decline in *C. minuta*, and total diatom concentration is very low. Between c. 394 ka BP and 390 ka BP, there is a significant decline in $\delta^{18}O_{diatom}$ values, reaching a low of +16.2‰ at 390 ka BP. This is accompanied by an increase in contaminating silt to c. 20%.

The latter stages of MIS11 are characterized by a gradual transition into the glacial conditions of MIS10. After the stadial period at c. 390 ka BP, BioSi-inferred productivity during MIS11.1 exhibits a small but temporary increase between 390 ka BP – 380 ka BP. During this period, valves of *S. distinctus* dominate the assemblage, and are gradually replaced by valves of *C. minuta*. $\delta^{18}O_{diatom}$ values increase again to +24.7‰ at c. 384 ka BP, before gradually declining to +14.5‰ at the top of this stage. This decline is accompanied by increases in the level of contamination found in the samples, which fluctuate markedly at the end of MIS11.1 and during MIS10.

4. Discussion

In this discussion we focus on the contribution to knowledge on hydrological stability in Lake Baikal from the $\delta^{18}O_{diatom}$ records. Before interpreting the $\delta^{18}O_{diatom}$ records in a palaeoclimatic context however, issues over $\delta^{18}O_{diatom}$ representivity need first to be considered. Interpretations of changing $\delta^{18}O_{diatom}$ values determined from Lake Baikal sediments rely on knowledge of the contemporary system and potential confounding factors with the methodology for the technique (Morley et al., 2004). The principal controls on $\delta^{18}O$ in the waters of Lake Baikal include relative inputs from northern basin and southern basin rivers, changes in seasonal precipitation, atmospheric circulation changes, temperature dependent $\delta^{18}O$ in precipitation, evaporative enrichment, and glacier retreat resulting in increased amounts of depleted meltwater (Seal & Shanks, 1998; Morley et al., 2005). Mean $\delta^{18}O$ and δD values in each of the three main basins and their associated rivers is summarised in Table 2. Lake water isotope values are extremely consistent both across the length of the lake and at depth: $\delta^{18}O = -15.8 \pm 0.2‰$ (2σ); $\delta D = -123 \pm 2‰$ (2σ) (Seal & Shanks 1998; Morley et al. 2005), confirming previous findings that despite its large size, the lake is well mixed (Weiss et al. 1991). These findings have important implications for palaeo studies because they indicate that $\delta^{18}O$ values will be a weighted average of all input sources minus outputs (Morley et al. 2005). River inflow accounts for c. 83% of annual hydrological input into lake, with precipitation accounting for c. 16% (Gronskaya & Litova 1991). As highlighted above, the Selenga River which flows into the south and central basins is by far the dominant source of water into the lake, and therefore changes in relative input between the Selenga and rivers which flow into the north basin (c. 20% of total fluvial input) is likely to be one of the dominant controls on varying $\delta^{18}O$ values through time (Morley et al. 2005). The remaining rivers which flow into the south and middle basin account for c. 7% and 11% of total fluvial input respectively.

Confounding factors which can affect $\delta^{18}O_{diatom}$ values

Three factors which can have important impacts on the $\delta^{18}O_{diatom}$ values, as applied to Lake Baikal sediments include: (i) contamination from other oxygen-bearing minerals in the sediment; (ii) diatom dissolution; and (iii) vital effects from changing diatom communities. As highlighted above, contamination is probably the major problem in palaeo studies exploiting the $\delta^{18}O_{diatom}$ technique. In our methodology we have tried to minimise contamination through careful development of the diatom cleaning procedure, and the application of a mass-balance approach to take account of the relative abundance of potentially contaminating grains in the sediment samples. In our interpretation, we take account of contamination by plotting up the record of % contamination alongside the $\delta^{18}O_{diatom}$ record (Fig 3). It is fair to say that when levels of contamination are high, we have less confidence in the accuracy of the $\delta^{18}O_{diatom}$ record, and this is borne out by recent work published in this volume by Brewer et al.. They found that a decline in $\delta^{18}O_{diatom}$ values (associated with high abundances of contaminating grains) during the Younger Dryas in Lake Baikal was significantly over-estimated, and could be compensated for using elemental geochemistry. These recent findings, together with the observation that all $\delta^{18}O_{diatom}$ peaks coincide to some extent with minima in contaminating silt, are borne in mind when interpreting the profiles presented here.

In Lake Baikal, although diatoms dominate sedimentary assemblages during interglacials, diatom dissolution at the sediment-surface water interface is still a significant process. Only c. 1% of valves present in the lake's water column are preserved in the sedimentary record (Ryves et al. 2003). Once incorporated into the sediments, dissolution of diatom silica continues into the pore-waters, until saturation is reached, thereby buffering against further dissolution (Conley & Schelske 1989; Carter & Colman 1994). There is the possibility therefore that dissolution of diatom silica may also influence $\delta^{18}O_{diatom}$ values. For example, dissolution preferentially removes certain species from the sedimentary assemblage, and *Synedra* spp have been shown to be more susceptible to dissolution than *Cyclotella* spp. (Battarbee et al. 2005). Theoretically this could influence the isotope signal by biasing potential vital effects, although such vital effects are likely to be smaller than overall analytical error (Leng & Barker (2006). A recent experimental study on the direct impacts of diatom valve dissolution on $\delta^{18}O_{diatom}$ found that while enrichment of $\delta^{18}O_{diatom}$ values occurred when dissolution took place at high pH (pH 9.0), there was no significant change on $\delta^{18}O_{diatom}$ values at near neutral pH (Moschen et al. 2006). The pH of Lake Baikal is generally 7.1-7.2 (Votintsev, 1961). It is likely therefore that dissolution will have had at most a minor effect on the $\delta^{18}O_{diatom}$ values observed, although direct evidence is still needed from experimental studies.

The impact of vital effects on $\delta^{18}O$ values is rather less well known for diatoms than some other organisms. Increasingly however, studies that have been carried out suggest the impact of differential species fractionation is probably 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). Recently, Swann et al. (2007) suggest that size of diatom may be an important factor in influencing $\delta^{18}O_{diatom}$ values. In the MIS11 sequence from Lake Baikal, the sizes of dominating diatoms in the core are very similar (unlike for example, the MIS5e sequence, which is dominated by the very large *Stephanodiscus grandis* (Rioual & Mackay 2005)). Also relevant to this study is the fact that many of the species represented in the core are now extinct, limiting any further interpretation on the possible role of vital effects, at least for this particular sequence.

Reconstructing hydrological variability in Lake Baikal during MIS11

Because of low eccentricity and dampened precessional amplitude MIS11.3 (c. 427 - 414 ka BP) does not end abruptly in a glacial stage (Petit et al., 1999). Instead, high levels of BioSi and $\delta^{18}O_{diatom}$ values extend through MIS11.2 until c. 394 ka BP, i.e. warm, interglacial conditions persisted in central Asia for c. 32 ka (Karabanov et al., 2003) (Fig 3). This is much longer than subsequent Lake Baikal interglacials (Williams et al., 1997) and is comparable to, for example, the longest interval of reduced millennial-scale SST variability determined in North Atlantic ocean sediments over the last 0.5 million years (McManus et al. 1999).

Highest $\delta^{18}O_{diatom}$ value of +27.7‰ for the whole sequence is found during MIS11.3 at c. 420 ka BP, suggesting that hydrological input into Lake Baikal was dominated by the Selenga and other rivers flowing into the south basin. At this time, diatom assemblages are dominated by *Stephanodiscus distinctus* and its varieties (Fig 3). *Stephanodiscus* species normally prefer low Si:P ratios. This will occur when nutrient dynamics are dominated by circulation, promoting a long period of spring overturn and large release of nutrients from bottom sediments / or the hypolimnion, i.e. as brought about when the Siberian High is weak, together with strengthened influence of the Westerlies. All these data suggest a period of prevailing humid climate. However, the decline in $\delta^{18}O_{diatom}$ values also coincides with small, but distinct increases in *Synedra* spp, benthic taxa and a small increase in contamination. Previous palaeo studies have suggested that increases in *Synedra* can be interpreted as being associated with increased concentrations of silica, related to increased catchment run-off and river discharge (Bradbury et al. 1994). Moreover, as the Academician Ridge is an isolated high, for benthic taxa to increase would have needed a substantial increase in lake circulation, perhaps enhanced by increased storminess, to bring these littoral taxa into this pelagic region. This process may also account for the increase in silt particles. A more rigorous approach to the assessment of contamination using geochemical techniques, as outlined by Brewer et al. (this volume), could help to address the pattern of observed $\delta^{18}O_{diatom}$ values here. The $\delta^{18}O_{diatom}$ record provides some evidence that prevailing warm, humid climates dominated in this region of central Asia. However, possible impacts from increased contamination brought about by these conditions needs further work.

At the transition between MIS11.3 and MIS11.2 (at c. 414 ka BP), the decline in $\delta^{18}O_{diatom}$ values of about 5‰ is concurrent with a small decline in BioSi and shift in species dominance from the extinct endemic *S. exiguus* to the extant species *S. binderanus*. The transition between MIS11.3 and MIS11.2 therefore is marked by a relative increase in hydrological input into Lake Baikal from rivers flowing into the north basin, such as the Upper Angara, and a concomitant reduction in south basin river discharge, especially the Selenga. This is associated with the onset of cooler temperatures and a small decline in diatom productivity in Lake Baikal. This pattern is repeated at c. 398 ka BP. These small but distinct fluctuations are similar in scale to those determined in Holocene $\delta^{18}O_{diatom}$ records (Morley et al. 2005; Mackay 2007). Even using improved geochemical techniques, when contamination in the Holocene core is small, shifts in $\delta^{18}O_{diatom}$ values are preserved (Brewer et al. this volume), suggesting that these are real responses to changes in climate.

390 Throughout MIS11.2 (c. 414 ka BP – 394 ka BP), $\delta^{18}O_{diatom}$ values fluctuate between c. +22 to +27‰. At this time, *S. binderanus* dominates the diatom assemblage, which grows best in conditions that promote strengthened water circulation and nutrient regeneration in the lake (Mackay et al., 2006). The strength of the Westerlies must have been very significant during this period because abundances of *S. distinctus* also increase during this time. Furthermore, an increasing abundance of *C. minuta* (which blooms in the autumn) indicates strengthened circulation later in the year. BioSi values are at their highest during MIS11.2, which Karabanov et al. (2003) suggest indicate a slightly warmer climate than MIS11.3. Mean $\delta^{18}O_{diatom}$ values for both MIS11.3 and MIS11.2 are very similar (Table 1), but given that these values are controlled by hydrological inputs, we would not necessarily expect them to differ with respect to relatively minor changes in temperature variability.

Isotopic evidence for an abrupt cooling event associated with millennial-scale variability

After c. 394 ka BP, values of BioSi, total diatoms and $\delta^{18}O_{diatom}$ decline markedly, delimiting the boundary between MIS11.2 and MIS11.1. Of note, the decline in $\delta^{18}O_{diatom}$ values lags the BioSi decline by c. 3 ka. This is likely related to the fact that biological productivity in the lake shows a much more rapid response to changing climate than the responses of changing hydrological sources and catchment processes, which drive declines in $\delta^{18}O_{diatom}$ values. This transition is marked therefore by the onset of cooler temperatures and declining productivity in Lake Baikal. This is also reflected in the diatom assemblage composition, as now only the autumnal blooming taxon *C. minuta* is dominant between 393 – 390 ka BP, followed by a decline in abundance of all diatoms. The extent of the $\delta^{18}O_{diatom}$ decline between c. 396 ka BP and 390 ka BP is significant,

+26‰ to +16‰, although it should be noted that the lowest value of +16‰ is based on a single sample. This decline is similar in scale to that which occurred during the Younger Dryas stadial, when values fell from c. +26‰ to +14‰ (Morley et al., 2005). Brewer et al. (this volume) have significantly reduced the extent of the range of the decline in $\delta^{18}O_{diatom}$ values during the Younger Dryas by compensating for high levels of contamination in the samples of up to 45% (Morley et al. 2005). In this study however, the extent of contamination is much less (Fig 3) and therefore any modification of the $\delta^{18}O_{diatom}$ record at this time will likely be reduced as well. The occurrence of stadial conditions in central Asia is supported by the core lithology, which documents significant changes in the Lake Baikal catchment (Karabanov et al. 2003). Minerogenic-rich sediments dominate the sequence, although the absence of ice-rafted debris material at this time suggests that this glaciation was limited to the mountain ranges surrounding Lake Baikal (in contrast to sediments deposited during MIS12 and MIS10).

The occurrence of stadial conditions during MIS11 has been found in many archives around the world. For example, Antarctic temperatures fell between c. 402 ka BP – 388 ka BP, together with marked declines in greenhouse gas concentrations at both Vostok (Petit et al., 1999) and Dome C (Spahni et al., 2005). In the North Atlantic, there is a substantial increase in ice-rafted debris (IRD), the first since at least 420 ka BP. This is indicative of a major increase in iceberg discharge, resulting in a slowdown in North Atlantic thermohaline circulation and southwards penetration of cooler, polar waters (McManus et al., 1999). Such THC responses are likely to have been responsible for the very large increase in polar foraminifera assemblages found off the Iberian Peninsula (de Abreu et al., 2005). An abrupt cooling episode is also expressed as a sandy loess layer intercalated within the S4 soil sequence from Shimao, at the northern edge of the Chinese Loess plateau on the border of the Mu Us desert (Sun et al., 1999).

The decline in $\delta^{18}O_{diatom}$ values to a minimum at c. 390 ka BP can be interpreted as a relative increase in northern basin river discharge (which is fed predominately by isotopically-low winter precipitation) and a concomitant reduction in south basin river discharge, especially the Selenga. These shifts in fluvial patterns are consistent with shifts in atmospheric circulation over central Asia during periods of intense cold resulting from the slowing down in the North Atlantic THC. Subsequent, extended Eurasian spring snow cover extent results in the weakening of the East Asian monsoon and a strengthening of the Siberian High, which causes increased aridity across the catchment of rivers to the south of Lake Baikal, including the Selenga (Bartlett et al., 1988; Lui & Yanai, 2002). Future work ought to be able to expand on this hypothesis. For example, the expansion of steppe-like vegetation communities in this region is indicative of cooler and drier conditions (e.g. Demske et al. 2005). Such changes ought to be reflected in the pollen record, which for MIS11in Lake Baikal is currently being investigated.

Slow growth of northern hemisphere ice sheets

Previous interpretations of Lake Baikal palaeoclimate during the latter stages of MIS11 concluded that biogenic silica concentrations mirrored the slow growth of northern hemisphere ice sheets (Karabanov et al., 2003). However, while BioSi content in Lake Baikal sediments was closely associated with changes in insolation up to c. 390 ka BP, for the remainder of MIS11 this association is much weaker. We are unable to state at this stage if this is a decoupling of the Lake Baikal's BioSi production response to insolation, but a recent study has shown that in pelagic regions of Lake Baikal, dissolution has a relatively greater impact on BioSi records during cooler, interstadial conditions (Swann & Mackay 2006). The diatom assemblage at this time is dominated by the autumnal *C. minuta*, highlighting the potential suppression of spring-blooming diatoms by extended ice and snow cover on Lake Baikal during periods of cooler climate (Mackay et al., 2005).

From c. 380 ka BP to the MIS11-10 boundary, BioSi values decline to almost glacial levels. However, $\delta^{18}O_{diatom}$ values show a markedly different response with a return to relatively high values of c. +26‰ at 386 ka BP, but then values gradually decline to c. + 15‰ at the end of MIS11 (Fig 3). Other records also show evidence of gradual cooling in the latter half of MIS11, such as increasing relative abundance of polar foraminifera in North Atlantic marine sediments (de Abreu et al., 2005), that are linked to slow ice sheet growth in the northern hemisphere.

During the latter stages of MIS11 in Lake Baikal, notable minima in $\delta^{18}O_{diatom}$ values occur. However, during this period the mean proportion of silt contamination is also much higher (Table 1). The amplitude of the changes in $\delta^{18}O_{diatom}$ values during MIS11.1 are greater than those measured during peak interglacial conditions MIS11 3/2, perhaps suggesting a higher degree of instability in the climate system (Karabanov et al., 2003). However, the extent of contamination is also greater during MIS11.1 than during the previous interglacial (Table 1), thus highlighting the need for further improvements to be made to obtaining $\delta^{18}O_{diatom}$ values which accurately take account of levels of persistent contamination (Brewer et al. this volume).

488 **4. Conclusions**

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490 In this study we report on additional analyses performed on BDP-96 material from the
491 Academician Ridge. Previously, biogenic silica and diatom analyses demonstrated that
492 interglacial glacial climate persisted in central Asia for over 32 ka, longer than anytime in
493 the last 450 ka (Karabanov et al. 2003). $\delta^{18}O_{diatom}$ analyses of the same sediments
494 highlight that over this period, hydrological input into the lake was dominated by rivers
495 flowing into the south basin, especially the Selenga. There is distinct variability in the
496 $\delta^{18}O_{diatom}$ record at this time. However, all fluctuations in $\delta^{18}O_{diatom}$ values are consistent
497 with small fluctuations in relative contamination, limiting further palaeoclimatic
498 interpretation. At c. 390 ka BP, the $\delta^{18}O_{diatom}$ record highlights that during this stadial
499 period, river inflow from the north of Lake Baikal became more important. This was
500 presumably due to intensification of the Siberian High, linked to iceberg discharge into
501 the North Atlantic ocean. Biogenic silica, diatom and $\delta^{18}O_{diatom}$ records all show a shift to
502 interstadial conditions during the early stages of MIS11.1, followed by slow decline in
503 values, mirroring the slow increase in extent of northern hemisphere ice sheets.
504 Contamination of the $\delta^{18}O_{diatom}$ record occurs throughout the whole profile, and becomes
505 especially significant during the latter stages of MIS11.1 and at the MIS11 / MIS10
506 transition.

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Figure legends:

Fig 1

Mean sea level pressure (hPa) across central Asia in (a) July and (b) January. Lake Baikal is indicated and the coring location of BDP-96-2 highlighted (τ). Maps adapted from Lydolph (1977).

Fig 2

Map of Lake Baikal, showing sites BDP-96 on the Academician Ridge and CON-01-605-3 on the Vydrino Shoulder in the south basin.

Fig 3

Sediment archive BDP-96-2 from Lake Baikal showing (in order): total diatom concentration; diatom concentrations for selected major species (*Stephanodiscus distinctus*, *Stephanodiscus binderanus*, *Stephanodiscus binderanoides Stephanodiscus exiguus*, *Cyclotella minuta, Aulacoseira skvortzowii, Synedra* spp, total benthic taxa); % biogenic silica; BDP-96-2 mass-balanced $\delta^{18}O_{diatom}$ record; contamination record of clays and silt (used in mass-balance calculation; lithology (from Karabanov et al. 2003). MIS11 boundary is delimited by black, solid line at 427 ka BP and 362 ka BP. MIS11.3/2 interglacial boundary and MIS11.2/1 boundary are delimited by dashed, black line. The lithology during MIS12 is characterized by glacial clays (vertical lines), MIS11 characterized by diatomaceous sediments (short vertical dashes), and MIS10 characterized by clays with diatoms present. Note that the dashed lines for *A. skvortzowii, Synedra* spp and total benthic taxa are exaggerated by x 50.

Table 1:

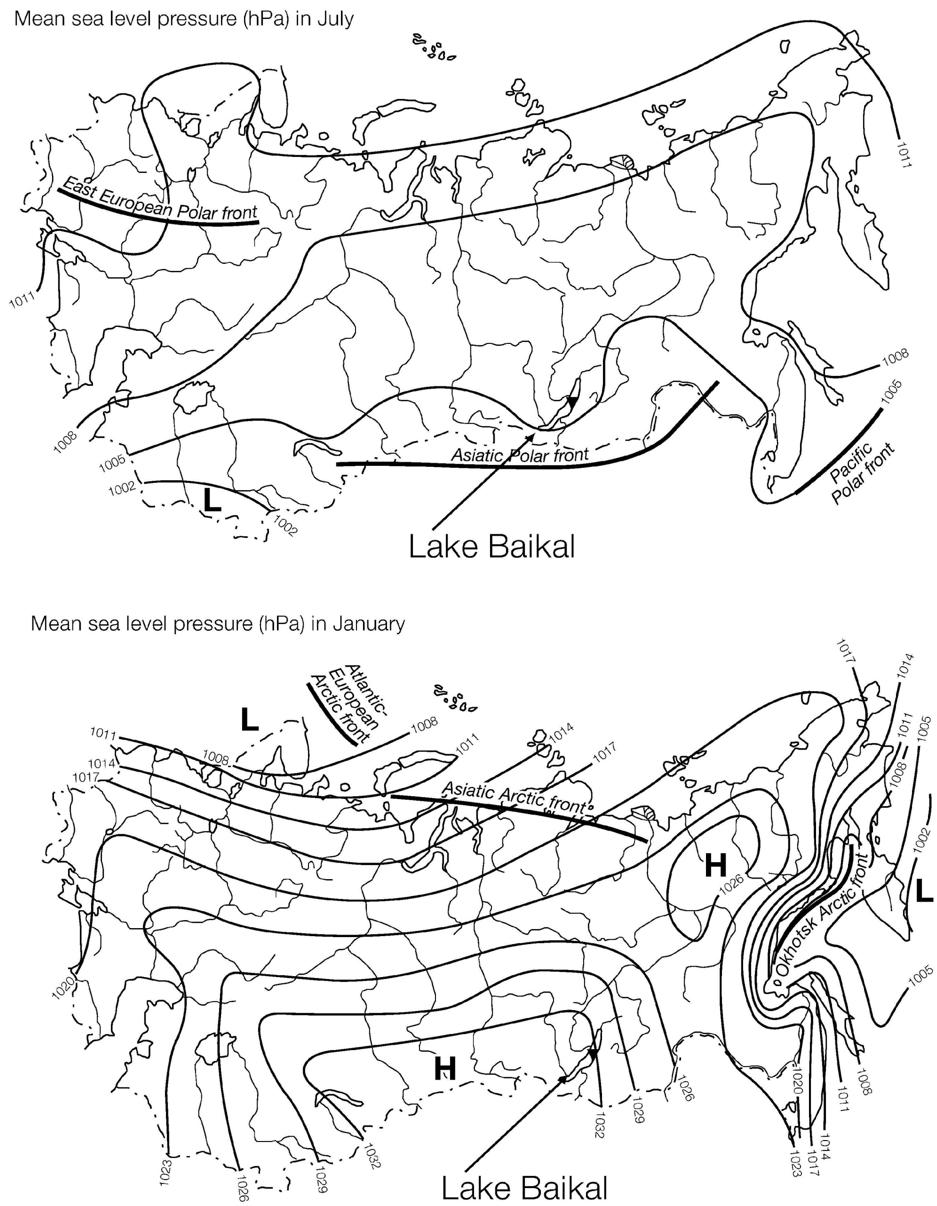
Mean and standard deviations of oxygen isotope values and silt contamination for marine isotope stages highlighted in the text

Stage	Mean $\delta^{18}O_{diatom}$ values (‰)	1 σ $\delta^{18}O_{diatom}$ values (‰)	Mean % silt	1 σ % silt
MIS10 (n = 5)	16.2	3.6	48.8	33.8
MIS11.1 (n = 18)	20.3	3.6	26.0	20.4
MIS11.2 (n = 12)	24.3	1.4	13.1	7.6
MIS11.3 (n = 6)	24.1	2.2	10.1	8.9
MIS11.2+3 (n = 18)	24.2	1.6	12.6	7.9
MIS1 (n = 81)	25.5	2.5	6.6	5.4

Table 2:

Summary table showing mean $\delta^{18}O$ and $\delta D \pm 2\sigma$ (‰ vs. SMOW) values for the north, central and south basin lake water and associated river input. Values come from two main sources: Seal and Shanks (1998)[1] and Morley et al. (2005)[2].

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



Mean sea level pressure (hPa) across central Asia in (a) July and (b) January. Lake Baikal is indicated and the coring location of BDP-96-2 highlighted (τ). Maps adapted from Lydolph (1977).
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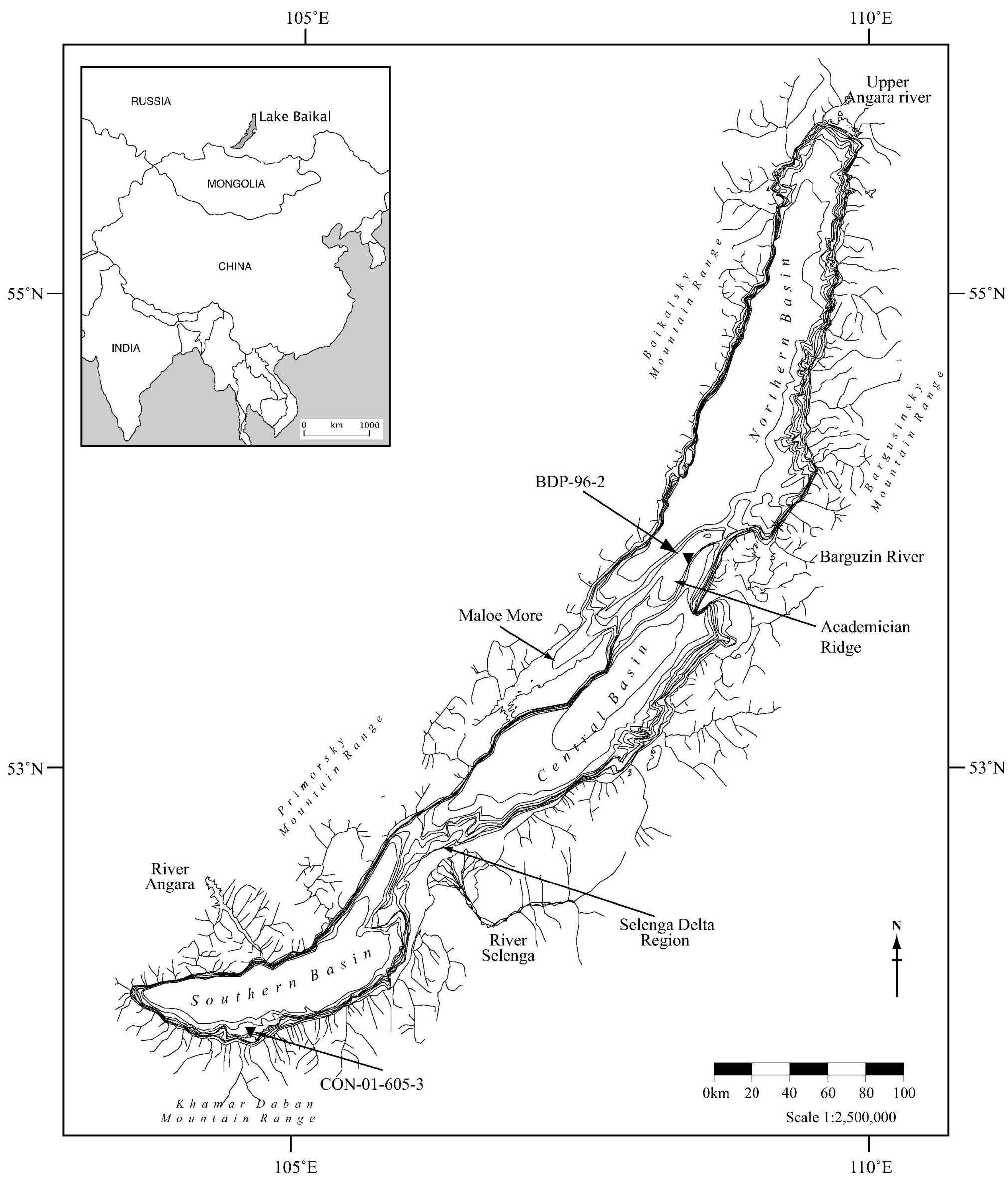
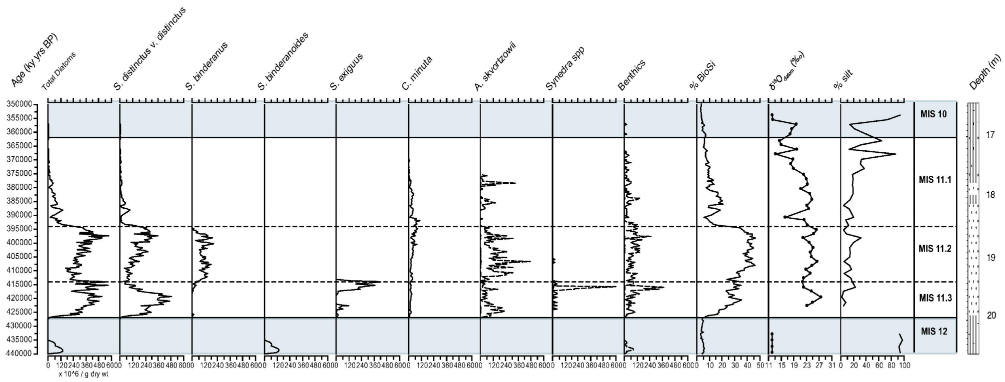


Map of Lake Baikal, showing sites BDP-96 on the Academician Ridge and CON-01-605-3 on the Vydrino Shoulder in the south basin.
196x229mm (600 x 600 DPI)



Sediment archive BDP-96-2 from Lake Baikal showing (in order): total diatom concentration; diatom concentrations for selected major species (Stephanodiscus distinctus, Stephanodiscus binderanus, Stephanodiscus binderanoides Stephanodiscus exiguus, Cyclotella minuta, Aulacoseira skvortzowii, Synedra spp, total benthic taxa); % biogenic silica; BDP-96-2 mass-balanced δ18Odiatom record; contamination record of clays and silt (used in mass-balance calculation; lithology (from Karabanov et al. 2003). MIS11 boundary is delimited by black, solid line at 427 ka BP and 362 ka BP. MIS11.3/2 interglacial boundary and MIS11.2/1 boundary are delimited by dashed, black line. The lithology during MIS12 is characterized by glacial clays (vertical lines), MIS11 characterized by diatomaceous sediments (short vertical dashes), and MIS10 characterized by clays with diatoms present. Note that the dashed lines for A. skvortzowii, Synedra spp and total benthic taxa are exaggerated by x 50.
286x107mm (600 x 600 DPI)