Lake Baikal climatic record between 310 and 50 ky BP: Interplay between diatoms, watershed weathering and orbital forcing

Tomáš Grygar a,*, Anna Bláhová a, David Hradil a, Petr Bezdíčka a, Jaroslav Kadlec b, Petr Schnabl b, George Swannc, Hedi Oberhänslid

a Institute of Inorganic Chemistry ASCR, 250 68 Řež, Czech Republic
b Institute of Geology ASCR, Palaeomagnetic Laboratory, Rozvojová 269, 165 00 Prague 6, Czech Republic
c Department of Geography, University College London, Gower Street, London WC1E 6BT, UK
d GeoForschungsZentrum, Potsdam, D-14473 Potsdam, Germany

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Abstract

The environmental record from Lake Baikal, Russia, from 310 to 50 ky BP (MIS 9a to MIS 3) was interpreted using rock magnetic, UV–Vis spectral, mineralogical, and diatom analyses. The age model was based on a correlation of the diatom and chemical weathering records and the summer insolation curve at 55°N and checked against an age model based on the proxy of relative palaeointensity of the Earth’s magnetic field. Peaks in chemical weathering within the watershed, inferred from maximum concentration of magnetic and coloured minerals and mica, the lowest mean Fe oxidation state in silicates and highs in expandable clay minerals correlated with the Northern Hemisphere summer insolation minima at 55°N. Reconstructed changes in weathering intensity are better correlated to insolation patterns than to global ice volume records. We propose a scheme of yet missing palaeoenvironmental interpretation of the diatom assemblage, including also some extinct species. Aulacoseira baicalensis and Aulacoseira skvortzowii were abundant in the early stages of lake flora recovery immediately after deglaciation and during MIS 7e and MIS 5e; periods of more pronounced continental climate and peak chemical weathering. Stephanodiscus formosus var. minor, Cyclotella minuta and Cyclotella ornata dominated in intervals of decreased seasonality and decreased humidity at the end of most interglacial/interstadial diatom zones. Stephanodiscus grandis, Stephanodiscus carconeiformis and Stephanodiscus formosus were ubiquitous between MIS 8 and MIS 5, an interval marked by high seasonality, i.e., large differences between winter and summer insolation, and low humidity revealed by a low hydrolysis of expandable clay minerals in the watershed. Diatom concentrations peaked in the climatic optima of MIS 7e and MIS 5e and in the short periods marked by shifts to warmer conditions in the upper sections of MIS 5: MIS 5c (103–99 ky BP), MIS 5b (90–88 ky BP), and MIS 5a (84–79 ky BP) in which increased humidity resulted in enhanced hydrolysis of clay minerals. No such short similar climatic optimums were found from MIS 9a to MIS 6. Sharp climate deteriorations recorded in the diatom and clay mineral records at 107, 94, and 87 ky BP, however, occurred within 1–2 ky of cold extremes in North Atlantic sea surface temperature emphasizing the strong teleconnections between the two localities.

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1. Introduction

Lake Baikal represents a unique archive of East Eurasian continental climate. In this region the climate
reflects both global climatic changes in addition to more localised continental features (Todd and Mackay, 2003). On the one hand there is evidence of strong palaeoteleconnections between Lake Baikal and the North Atlantic region, such as the response to Heinrich events recorded in Lake Baikal (Prokopenko et al., 2001c; Swann et al., 2005). On the other, the continental character of East Eurasia significantly amplifies certain climatic fluctuations which are rather weakly represented in marine sediment records. These are similar to the difference between the continental pollen records from Europe and records from the Atlantic Ocean with the former showing substantially shorter interglacials relative to the corresponding marine isotopic stages (Tzedakis et al., 2004). The palaeoclimatic record from Lake Baikal has been intensively studied since the 1990s, particularly in the frame of two international initiatives, BDP and CONTINENT. The proxies developed and used within these initiatives for palaeoclimatic and palaeoenvironmental reconstructions have mainly been based on diatom (Prokopenko et al., 2001a; Khursevich et al., 2001; Rioual and Mackay, 2005) and pollen analyses (Tarasov et al., 2005; Granoszewski et al., 2005). At the same time though, elemental (Chebykin et al., 2004; Goldberg et al., 2005) and mineral (Sakai et al., 2005; Grygar et al., 2005) analyses have also been used to reconstruct palaeoclimatic conditions in and around Lake Baikal.

The interpretation of Lake Baikal detritus clay mineral records has developed from assuming that neof ormation of expandable clay minerals occurred during interglacials (Yuretich et al., 1999; Horiuchi et al., 2000) to an assumption that less straightforward clay mineral alterations occurred by pedogenesis (Fagel et al., 2003) to a hypothesis of complete dissolution of clay minerals in interglacials (Sakai et al., 2005). In addition, there is also a possibility that changes in clay mineralogy may have been affected by the environmentally controlled source area rather than changing neoformation (Grygar et al., 2005). Developing a clear understanding for the interpretation of clay mineral assemblages is essential due to the fact that clay minerals, and especially expandable clay minerals, have a relatively well-defined pattern in recording palaeoclimatic changes. In contrast to this, concentrations of other silicate minerals are often invariant to changes in the climate. For example Fe oxide minerals, otherwise the most sensitive palaeoenvironmental indicator, are most likely affected by post-depositional diagenesis. In our recent works (Grygar et al., 2005, 2006), we used Cation Exchange Capacity (CEC), a novel proxy recording the concentration of expandable clay minerals in the sediment, to determine the palaeoclimate and palaeoenvironmental changes in and around Lake Baikal. A significant advantage of CEC is that it can be obtained by simple and selective chemical analyses which are much less time consuming and more reliable than previously used empirical methods based on X-ray diffraction analysis.

Diatoms in Lake Baikal have been commonly used to reconstruct palaeoclimatic and palaeoenvironmental changes over glacial-interglacial cycles (e.g., Khursevich et al., 2001) as well as within interglacials (e.g., Rioual and Mackay, 2005) and glacial (e.g., Swann et al., 2005). Such work is based on the often distinct environmental characteristics of individual taxa, the strong first-order relationship which exists between insolation patterns and diatom abundance (Prokopenko et al., 2001a) and the diversity, appearance and extinction of individual species (Khursevich et al., 2001). Despite this, the interpretation of diatom records is not always clear and is often subject to uncertainties or assumptions. For example, the palaeoenvironmental interpretation of individual diatoms species is often limited by the unknown ecological characteristics of both endemic and extinct diatom taxa. This is particularly true prior to the Holocene with many of the dominant species from the Pleistocene now extinct (Khursevich et al., 2001). A further feature of the Lake Baikal diatom record is the high extent of diatom dissolution, which can occur both in the water column and at the surface sediment-interface (Ryves et al., 2003). As such, there is not always a straightforward link between the diatoms in the sediment and past levels of lake productivity, a problem which can be further confused by the differential preservation of individual species (Battarbee et al., 2005). In addition it is likely that the relative dissolution of diatoms was greater during glacial, either due to higher bacterial/biological action (Swann and Mackay, 2006) or due to the lower concentrations of dissolved silica in the lake following suppressed chemical weathering in the watershed. Consequently, due to the relative problems associated with both biological and non-biological proxies, understanding and interpreting the sediment record from Lake Baikal in terms of past environmental and climatic changes may be best achieved through a combined biogenic and non-biogenic approach (Horiuchi et al., 2000; Chebykin et al., 2004; Grygar et al., 2006).

The aim of this study is to describe the palaeoenvironmental changes in the Lake Baikal watershed between MIS 8 and MIS 4 by combining rock magnetic, mineralogical, and diatom-based proxies. This work is a continuation of our previous work describing the course of the last glacial cycle (Grygar et al., 2006). The methodologies used within this current study have been described and verified in our previous reports (Grygar et al., 2005, 2006) on the upper section of the core studied in this report. In this report
special attention is paid to the interpretation of rock magnetic measurements and the environmental interpretation of these mineralogical proxies related to Fe oxides and silicates including clay minerals. From this, we attempt to find an unequivocal proxy detailing changes in chemical weathering within the Lake Baikal watershed. Diatom assemblage records are subsequently compared to these inorganic proxies in order to establish a better understanding of the relationship between the different parameters so as to provide further insights into the palaeoclimate changes which occurred in this region over glacial–interglacial cycles. Finally, our reconstruction of climatic changes in the Lake Baikal region between MIS 5 and MIS 8 is compared to changes recorded in the North Atlantic region in order to distinguish climatic differences and similarities between the two geographical localities.

2. Methods

2.1. Sampling and magnetic measurements

Hydraulic (piston) core VER98-1-13 was obtained from the Academician Ridge, Lake Baikal, at 53.561°N, 108.011°E. The upper part of the resulting section was described in previous reports (Grygar et al., 2005, 2006). The core was sampled using plastic cubic boxes with an inner volume of 6.7 cm³ (Natsuhera Giken Co., Japan) producing a continuous series of samples with a mean vertical distance of 2.2 cm between the box centres. Magnetic measurements were done after sediment sampling in a naturally wet state. Low field volume magnetic susceptibility (MS) was measured using the Kappabridge KLY-3S (sensitivity of 1.2 × 10⁻⁸ SI units). All samples were demagnetised in 5–6 steps with a maximum alternating field of 100 mT by the LDA-3 device. The natural remanent magnetisation (NRM) components were measured with the JR-6A or JR-5A spinner magnetometers after each demagnetisation step in order to gain the primary components of the NRM. The relative magnetic field palaeointensity values were calculated as the ratio of NRM value at 20 mT in the demagnetisation field to the anhysteretic remanent magnetisation (ARM) value gained at 20 mT in the demagnetisation field combined with a 0.05 mT constant biasing field. The ARM was imparted to the samples using a AMU-1A device and measured on the spinner magnetometers. After magnetic measurements, half of the sample was air-dried at 50 °C and powdered before analysis with a Siemens D5005 diffractometer (Bruker). Diffractograms were measured in the 2Θ range 2–70° (CuKα) with 20 s counting at 0.02° steps, resulting in a total measuring time of almost 19 h. The X-ray tube was operated continuously to minimize short-term intensity variations with the actual primary beam intensity checked by regular (monthly) measurements of a reference corundum specimen (SRM 1976, NIST). The diffraction patterns were then normalized to the mean integral intensity of the (104) line of the reference. The variations of selected well-crystalline minerals were estimated from variations of integral intensities at selected diffraction lines. The content of biogenic silica was estimated from the height of the extremely broad diffraction (shoulder with width FWHM ~ 5° in 2Θ scale) centred at 2Θ 22°, as was also done by Fagel et al. (2003).

High-temperature X-ray diffraction (HT-XRD) was performed following the methodology outlined in our previous report (Grygar et al., 2005) to identify major clay minerals. Samples in a naturally wet state were suspended in a minimum volume mixture of water and ethanol (1:4) and the suspension was poured onto a heated support (Pt-foil) in an Anton Paar HTK16 high-temperature chamber. The diffraction patterns were acquired with a X’Pert PRO diffractometer (PAnalytical) with CoKα radiation and X’Celerator multichannel detector. Analyses were performed at 25–300 °C with 5 °C steps in the 2Θ range 4–40° with 0.017° steps resulting in a total measuring time of almost 23 h.

2.3. Cation-exchange capacity (CEC)

CEC, calculated using the Cu-trien method (adopted from the original methodology of Meier and Kahr, 1999), is a very convenient quantitative method for the analysis of total expandable clay structures. The method has recently been successfully tested by Ammann et al. (2005). We have previously successfully used the Cu-trien method to analyse the upper part of the VER98-1-13 section (Grygar et al., 2005). 250 mg of sample (air dried at 50 °C and then powdered) was re-suspended in 5 ml of water before the addition of 5 ml of 9 mM solution of Cu(trien)SO₄ (trien = 1,4,7,10-tetraazadecane). The suspension was then stirred for 10 min and filtered into 50 ml flasks. Under these conditions, with ~50% of the Cu-trien consumed for the exchange, Cu²⁺
ions replace exchangeable cations in the expandable structures. The difference in concentration of Cu\(^{2+}\) (\(\Delta\text{Cu}^{2+}\)) and the concentration of evolved Ca\(^{2+}\) and Mg\(^{2+}\) were determined by atomic absorption or emission spectroscopy using AAS3 (Carl-Zeiss Jena, Germany). The mean error given by a possible adsorption of Cu\(^{2+}\) or dissolution of calcium or magnesium salts expressed as a molar non-equivalency of ions involved in the reaction (\(\Delta\text{Cu}^{2+}\text{--Ca}^{2+}\text{--Mg}^{2+}\)) \(\Delta\text{Cu}^{2+}\) was 3%, i.e., the systematic error for CEC determination was insignificant.

2.4. Diffuse reflectance spectroscopy (DRS)

The UV–Vis (electron) spectra of dried and ground samples were measured using a Perkin Elmer Lambda 35 spectrometer equipped with an integrating sphere (Labsphere). The interpretation of the electron spectra of Fe-bearing clay minerals and oxides has been described in previous studies (Grygar et al., 2003; Hradil et al., 2004) and was recently applied to Lake Baikal sediments (Grygar et al., 2006). In this report we used two characteristics based on reflectance (%): \(R_{\text{UV}}\), reflectance in UV–Vis region at 270 nm (charge-transfer bands of total Fe\(^{3+}\)), and \(R_{\text{Vis}}\), mean reflectance in visible-light region (400–700 nm, total lightness). For statistical analyses, \(R_{\text{UV}}\) and \(R_{\text{Vis}}\) were recalculated to absorbance using the Kubelka–Munk formula. Additionally another parameter, B/C, was obtained based on the amount of absorption calculated from the Kubelka–Munk formula (Grygar et al., 2006) and records the absorption ratio of Fe\(^{2+}\)–Fe\(^{3+}\) (blue chromophor of aliovalent Fe minerals, 13,900 cm\(^{-1}\), ~720 nm) to Fe\(^{3+}\) absorption (green chromophor of ferric compounds, 16,000 cm\(^{-1}\), 625 nm). As detailed within Grygar et al. (2006), the B/C ratio is a good measure of Fe oxidation state within detritus minerals.

2.5. Diatom analysis

Diatom slide preparation, counting and identification were performed using a method for Lake Baikal sediments previously described in Mackay et al. (1998) and used in our previous work (Grygar et al., 2006). Sediment samples were dried at 50 °C, weighted without grinding and suspended in test tubes with a defined addition of polyvinylbenzene microspheres to calculated diatom concentrations (Battarbee and Kneen, 1982). The resulting suspension was evaporated on microscope slides and mounted using Naphrax (Brunel Miroscopes, UK). For counting an optical microscope with oil immerse objective at \(\times1000\) magnification was used. The dominant diatom flora through the analysed period of \textit{Stephanodiscus grandis} (Khurs. and Log.), \textit{Stephanodiscus carconeiformis} (Khurs. and Log.), \textit{Stephanodiscus formosus} (Khurs. and Log.), \textit{Stephanodiscus}
formosus var. minor (Khurs. and Log.), Stephanodiscus flabellatus (Khurs. and Log.), Aulacoseira baikalensis (Meyer) Simonsen, Aulacoseira skwizortzowii (Edlund, Stoermer & Taylor) (syn. Aulacoseira islandica var. helvetica), Cyclotella minuta (Skv.) Antipova, Cyclo-
tella ornata (Skv.) Flower, Cyclotella baikalensis (Meyer) Skv., Cyclotella operculata (Agardh) Kützing and Cyclotella krammeri Håkansson were identified. Benthic taxa were grouped together.

3. Results

3.1. Rock magnetic logs, UV–Vis spectral properties and their mineralogical interpretation

It is well known that the concentration of magnetic minerals decreased in sediments in Lake Baikal during warmer and/or more humid periods, with minima in magnetic susceptibility (MS) and anhysteretic remanence magnetisation (ARM) minima corresponding to interglacials or interstadials (Peck et al., 1994; Demory et al., 2005; Grygar et al., 2005, 2006). The most likely explanation for this is enhanced chemical weathering of Fe-bearing minerals in more humid climates. Fe hydroxy-oxides and Fe bearing silicates such as dark micas and amphiboles, and ferrimagnetic magnetite, are responsible for the MS of the sediments. On the other hand, ARM reflects only the concentration of ferrimagnetics. MS (reflecting the sum of ferro-, para-, and diamagnetic properties of the sediments) and ARM (reflecting mainly ferrimagnetic mineral properties) do not display the same patterns or changes over the analysed interval (Fig. 1), showing that in some parts of the section the minima of ferrimagnetics are broader than the minima influenced by higher concentration of the paramagnetics. This difference probably indicates that ferrimagnetic minerals are more easily weathered in the Lake Baikal watershed than paramagnetic minerals and/or that some extra paramagnetics are formed by the weathering of ferrimagnetics. As such, periods with lowered ferrimagnetics and high paramagnetics likely represent periods of moderate chemical weathering.

The diffuse reflectance electron spectra (DRS) produced three proxies, each defined in the methodology and shown in Figs. 1 and 2. $R_{\text{Vis}}$ is the mean reflectivity in the Vis region, i.e., the lighter samples have higher values of $R_{\text{Vis}}$. The main coloured minerals in sediments from the Academician Ridge are amphiboles and dark micas (Grygar et al., 2005, 2006), both of which are very sensitive to chemical weathering in a humid climate. There is a good correlation between $R_{\text{Vis}}$ and total diatoms or biogenic silica (Table 1), showing that white and highly light-scattering SiO$_2$ can contribute to the pattern of high $R_{\text{Vis}}$ in humid/warm climates. $R_{\text{UV}}$ is a mean reflectivity at 270 nm, which is inversely proportional to the total Fe (Fe$_{\text{TOT}}$ as obtained by

![Fig. 2. Comparison of total diatoms (light triangles marks insolation maxima with age in ky BP), CEC and B/C (dark triangles mark insolation minima with age in ky BP).]
chemical analysis) in the sediments, \(Fe_{TOT}\) decreased in interglacials or interstadials (Grygar et al., 2006). B/C ratios have been shown to be a useful proxy for the oxidation state of Fe in aluminosilicates, assuming that only a small fraction of Fe is bound in Fe\(^{3+}\) oxides (Grygar et al., 2006). This indicator is very sensitive to chemical weathering: if Fe\(^{2+}\) bearing micas are weathered to expandable clay minerals only a small fraction of Fe is mobilized while the majority is oxidized to Fe\(^{3+}\) and retained in the structure of the aluminosilicates. This process is accompanied by a decrease in the B/C ratio. Generally, if all Fe-based proxies change “in phase”, i.e., if \(R_{Vis}\) increased and \(R_{UV}\), B/C, MS and ARM simultaneously decreased, chemical weathering was highly intensive and vice versa.

### 3.2. Age model

In our previous work (Grygar et al., 2006), the upper part of the core section was dated based on a correlation of the relative palaeointensity variations in the VER98-1-13 sediment core to relative palaeointensity variations calculated from ODP site 984 (Channell, 1999). The same dating approach was used for other recent Lake Baikal studies within the framework of the CONTINET project, as detailed by Demory et al. (2005) and used, amongst others, by Swann et al. (2005), Groszewska et al. (2005) and Tarasov et al. (2005). More than 30 points were found for reliable correlation of both relative palaeointensity records. However, the magnetic assemblage in the Lake Baikal and the ODP site 984 records could not be affected with exactly comparable post-depositional processes. Despite this uncertainty reversal excursions, such as the Blake (ca. 120 ka) and the Iceland Basin (ca. 180 ka) events detected in the VER98-1-13 core section, are robust control points in the relative palaeointensity dating of the lake sediments.

An independent age model was created by correlating palaeoenvironmental proxies to the Northern Hemisphere summer insolation, calculated according to the Berger solution (Berger, 1978) (Fig. 2, left panel). The advantage of such an approach is that peaks in insolation often coincide with sedimentary peaks in diatom populations and other palaeoproductivity proxies both in interglacials and glacials. This is illustrated by the same method also being used to derive an age model for the long diatom record from Lake Baikal cores BDP96-2 (Prokopenko et al., 2001a) and BDP-69-1 and-2 (Prokopenko et al., 2006). The resulting age model for core VER98-1-13 is in good agreement with other cores from the Academician Ridge in which the age model was obtained by comparing some palaeoproductivity proxy to a marine \(^{18}\)O record (Peck et al., 1994; Fagel et al., 2003; Sakai et al., 2005).

To obtain an orbitally tuned age model during the Brunhess, Prokopenko et al. (2001a) tied Northern Hemisphere summer insolation maxima and the onset of the main interglacial (interstadial) diatom booms. The timing of the onset of interglacial and interstadial diatom peaks has been confirmed by Prokopenko et al. (2001b) and Morley et al. (2005) with the first diatom peaks in Lake Baikal appearing at 15–13 ky BP with the main diatom boom developing soon after the Younger Dryas at \(\sim 10\) ky BP, while the Northern Hemisphere summer insolation maximum occurred at 11 ky BP. According to the detailed study of the Kazantsevo interglacial (Rioual and Mackay, 2005), based on a palaeomagnetic age model which is explicitly independent of orbital forcing, the fastest increase in the size of the interglacial diatom assemblages occurred at 128–127 ky BP, coinciding with the Northern Hemisphere summer insolation maximum at 128 ky BP. A feature of the Lake Baikal record is the dramatic switches between cold and warm periods even when Northern Hemisphere insolation is relatively

### Table 1

Squares of regression coefficients of correlation between selected proxies

<table>
<thead>
<tr>
<th>Variable</th>
<th>(Fe_{TOT})</th>
<th>Diatom biovolume</th>
<th>Diatom number</th>
<th>CEC</th>
<th>B/C</th>
<th>(R_{Vis})</th>
<th>(R_{UV})</th>
<th>XRD amorphous</th>
<th>ARM</th>
</tr>
</thead>
<tbody>
<tr>
<td>MS</td>
<td>0.51</td>
<td>0.26</td>
<td>0.26</td>
<td>0.11</td>
<td>0.04</td>
<td>0.38*</td>
<td>0.36*</td>
<td>0.65</td>
<td>0.29</td>
</tr>
<tr>
<td>ARM</td>
<td>0.31</td>
<td>0.11</td>
<td>0.10</td>
<td>0.14</td>
<td>0.13</td>
<td>0.30*</td>
<td>0.24*</td>
<td>0.19</td>
<td></td>
</tr>
<tr>
<td>XRD amorphous</td>
<td>0.35</td>
<td>0.60</td>
<td>0.52</td>
<td>0.24</td>
<td>0.20</td>
<td>0.56**</td>
<td>0.66**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(R_{UV})</td>
<td>0.60*</td>
<td>0.43**</td>
<td>0.50**</td>
<td>0.49*</td>
<td>0.00*</td>
<td>0.69*</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(R_{Vis})</td>
<td>0.48*</td>
<td>0.31**</td>
<td>0.16**</td>
<td>0.37*</td>
<td>0.00*</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B/C</td>
<td>0.01</td>
<td>0.15</td>
<td>0.14</td>
<td>0.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CEC</td>
<td>0.41</td>
<td>0.21</td>
<td>0.38</td>
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<td></td>
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<tr>
<td>Diatom number</td>
<td>0.40</td>
<td>0.50</td>
<td>0.34</td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Diatom biovolume</td>
<td>0.34</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Critical values of \(R^2\) at the 95% probability level are ca. 0.06 in all correlations with \(Fe_{TOT}\) and XRD amorphous component and ca. 0.04 for other correlations.

* reflectance recalculated to Kubelka–Munk absorbances. ** original reflectance in %.
smooth. This is likely attributable to the strong climatic
teleconnections which exist between the North Atlantic
region and Central Asia/Lake Baikal (Prokopenko et al.,
2001c; Todd and Mackay, 2003; Swann et al., 2005). For
example, the Northern Hemisphere summer insolation
maxima at 105 ky BP (onset of MIS 5c) and at 84 ky BP
(onset of MIS 5a) in the North Atlantic were preceded by
abrupt climate changes (McManus et al., 1994; Chapman
and Shackleton, 1999; Shackleton et al., 2002). Similarly,
the onset of the MIS 7e, MIS 7c and MIS 5e warm
intervals was preceded by large iceberg discharges and
decreases in sea surface temperature in the North Atlantic
(McManus et al., 1994, 1999; Hiscott et al., 2001;
Kandiano and Bauch, 2003) that accelerated the onset of a
warm mode in the global ocean circulation. Due to the
strong atmospheric teleconnections between the two sites,
these fast switches in the North Atlantic permit relatively
easy identification of the onset of warm periods in the
Baikal sedimentary record.

In an analogy to the relationship between the diatom
record and insolation maxima, we suppose that there
must be a correlation between chemical weathering
minima and summer insolation minima. We tied the
chemical weathering minima, identified as maxima of
ARM, MS, or B/C, to the NH summer insolation minima
(Fig. 2, right panel). The resulting age model is shown in
Fig. 3 together with the age model developed for the
upper part of the core (roughly MIS 6 to MIS 3) using the
variations of the Earth’s magnetic field (Grygar et al.,
2006). The total mean difference of these models was ca.
1 ky with a mean standard deviation of ca. 3 ky. Because
the magnetic measurements and sampling for further
analyses resulted in each sample comprising on average
a 2.2 cm sample interval, each data point represents a
time interval of approximately 0.5 ky. This “averaging”
is consistent with the uncertainty of our age model which
is on the order of a few ky. The orbitally tuned age model
was used to re-plot the experimental results in Figs. 4
and 5. The timing of peaks in ARM, MS and B/C was not
always perfectly synchronous, but all occur within a
range of about 3 ky, indicating that their environmental
signal is not explicitly related to the same “recording”
mechanism. These chemical weathering minima were
always found in periods free of diatom valves except
during the insolation minimum at 255 ky BP in the
middle of MIS 8.

3.3. Analysis of clay minerals

The use of Cation Exchange Capacity (CEC), a mea-
sure of the concentration of expandable clay minerals, as
an environmental proxy was previously discussed by
Grygar et al. (2005). The Cu-trien method is very specific
to the target compounds, i.e., smectites, vermiculite and
inter-stratified clay minerals with expandable components
(Meier and Kahr, 1999; Ammann et al., 2005). In addition,
it is one of the simplest chemical analytical methods
applicable to large numbers of samples. CEC is directly
proportional to the total concentration of expandable clay
structures and does not need empirical calibration, as
required by P-XRD analysis of clay minerals (Yuretich
et al., 1999; Fagel et al., 2003). The pattern of decreased
clay mineral content and expandable clay minerals in a
warm/humid climate was found by Sakai et al. (2005) and
Grygar et al. (2005), respectively. This pattern is also
obvious from the comparison of environmental proxies in
Figs. 2 and 5. This pattern has only two reasonable
explanations: either hydrolysis (dissolution) of expand-
able clay minerals is occurring under humid and warm
climates in interglacials and major interstadials or some
environmentally controlled shift is occurring in the
sediment source area. In any case, CEC minima are lo-
cated in periods of increased diatom concentration, al-
though not all diatom maxima are coincident with CEC
minima.

The qualitative analysis of the clay mineral assemblage
was performed using high-temperature XRD (HT-XRD).
The results of HT-XRD were similar to our previous study
covering the last glacial cycle (Grygar et al., 2005). In
periods with enhanced chemical weathering and high
diatom productivity, the relative percentage of chlorite
increased and micas decreased, showing that the later
is preferentially consumed by chemical weathering. In
stadials and glacials, a well-ordered expandable clay mineral is present, most probably vermiculite: an intermediate stage of mica weathering. As follows from Fig. 5, the CEC pattern resembles variations in mica content estimated by P-XRD. This confirms the inter-layering of illite–smectite and vermiculite (the major expandable clay minerals found by Fagel et al., 2003; Grygar et al., 2005) with the parent mica.

3.4. Diatom analysis

Two types of information can be obtained from diatom analysis, the concentration of individual diatom species, which can be recalculated to biovolume measurements using the reference volumes given in Morley et al. (2005), Rioual and Mackay (2005), Swann and Mackay (2006) and Rioual (personal communication), and a record of changes in species diversity over time. Diatom frustules increase the content of P-XRD amorphous components and strongly scatter UV and Vis light thus increasing $R_{\text{Vis}}$ and $R_{\text{UV}}$. In Table 1, diatom biovolumes were compared to the mineralogical proxies of the opal content. The regression coefficients of correlations between diatoms and the non-biogenic proxies improve in zones dominated by smaller sized Cyclotella and Aulacoseira species. Changes in total diatom concentrations have a similar

Fig. 4. Interrelation of total diatom concentrations, concentration of ferromagnetics (ARM), UV reflectance and total content of amorphous components (from XRD). Grey rectangles indicate diatom-rich periods.
pattern to other proxies including magnetic susceptibility, $R_{UV}$, and the amount of biogenic opal estimated by P-XRD (Fig. 4 and Table 1).

Several important features in the diatom diversity are, mostly, in agreement with previous reports (Fig. 6). In general glacial diatom assemblages contain only minimal numbers of diatom valves except for MIS 8. In the upper part of MIS 8 the glacial assemblages are almost solely composed of $S. \text{grandis}$, $S. \text{carconeiformis}$, $S. \text{formosus}$, and $S. \text{formosus var. minor}$ (also reported by Khursevich et al., 2001) and in MIS 6 and Termination II mostly by $C. \text{minuta}$ and $C. \text{ornata}$ (also reported by Edlund and Stoermer, 2000; Khursevich et al., 2001; Rioual and Mackay, 2005). As expected, more diverse and larger diatom populations were found in interglacials and interstadials due to the increased

Fig. 5. Proxies of lake productivity (concentration of total diatoms), chemical weathering in the watershed (B/C and mica variations) and the CEC proxy of humidity compared to summer insolation at 55°N and eccentricity in the Earth’s orbit. Grey rectangles indicate diatom rich periods.
Fig. 6. The succession patterns of diatoms compared to proxies of chemical weathering in the watershed. Grey rectangles represent major diatom peaks (upper MIS 8 and MIS 6 diatom peaks are not shaded).
solar insolation and subsequent reduction on seasonal ice cover and snow thickness over the lake. At the beginning of MIS 8 and in particular at the onset of MIS 7 an obvious diatom succession starts with peaks in *A. baicalensis* and *A. skvortzowii*, which are immediately followed by more stable periods dominated by *S. grandis* and *S. carconeiformis*. Similar changes occur during the peaks in diatom concentrations during MIS 5. However, while the abundance and presence of *Aulacoseira* taxa vary throughout the analysed interval, large-celled *Stephanodiscus* taxa are present throughout. The ecological requirements of these extinct *Stephanodiscus* taxa remain unknown, though an important requirement is believed to be deep water mixing and turbulence in the water column (Edlund and Stoermer, 2000; Rioual and Mackay, 2005). In addition their presence, even in glacial aged samples, may indicate a wide tolerance range and an ability to adapt even to different, more inhospitable, conditions. Alternatively, it may indicate the robustness of these taxa to dissolution in the water column relative to other Lake Baikal taxa. Both *A. baicalensis* and *A. skvortzowii* are cold water taxa requiring clear winter ice permitting deep water mixing in late winter and early spring. According to Rioual and Mackay (2005) both taxa have a similar ecology developing under the ice during the spring months. However, blooms of *A. skvortzowii* start to develop in near shore waters before extending into pelagic/offshore locations during the spring months while *A. baicalensis* primarily occurs in offshore waters (Mackay et al., 2000; Richardson et al., 2000).

The seasonal distribution of solar radiation during insolation peaks in MIS 8, MIS 7 and MIS 5 was different to today with increased eccentricity leading to decreased winter insolation and increased summer insolation, producing faster spring warming (and probably also much faster autumn cooling). These seasonal continental climatic contrasts gradually reduced from MIS 8 to MIS 5. The concentrations of extant small-sized *C. minuta*, medium-sized *C. ornata*, and extinct small-sized *S. formosus* var. *minor* peak at the end of periods of high diatom abundance (Fig. 6), i.e., in periods approaching the Northern Hemisphere summer insolation minima. *S. formosus* var. *minor*, in addition to *C. minuta* and *C. ornata*, also peaked in some stadials, such as 208–200 ky BP (MIS 7b) and 94–90 ky BP (MIS 5b). High concentrations of the large *S. grandis* and *S. carconeiformis* were followed by peaks in *S. formosus* var. *minor* in the lower part of MIS 8 and the upper part of MIS 7, with a mean lag of ∼4 ky, and by peaks of *C. minuta* in the uppermost sections of MIS 7 and by both *C. minuta* and *C. ornata* in the lowermost peaks of MIS 5 with a mean lag of ∼4 ky. Similar succession trends have also been found in sediments from the Academician Ridge during the climatic optima corresponding to MIS 5c and MIS 5a (Chebykin et al., 2004, Grygar et al., 2006), MIS 7 and the Kazantsevo (Khursevich et al., 2001), and at Continent Ridge during the Kazantsevo interglacial (Rioual and Mackay, 2005).

The prevalence of *C. minuta* and *C. ornata* at the end of diatom peaks close to Northern Hemisphere summer insolation minima and in a period of rather stable Northern Hemisphere summer insolation maximum in late MIS 6, together with the abundance of these taxa in Lake Baikal in the Holocene and today, suggests that these taxa can adapt to the decreased inter-seasonal insolation contrasts which may have been more unfavourable for larger *Stephanodiscus* taxa. In the early Kazantsevo *A. skvortzowii* prevailed, while in the latter parts of the interglacial *A. baicalensis* appeared. This *Aulacoseira* succession was also found by Edlund and Stoermer (2000) at Buguldeika Saddle, Khursevich et al. (2001) at the Academician Ridge and by Rioual and Mackay (2005) at Continent Ridge, suggesting that it reflects a reduction in snow/ice cover throughout the North Basin. Following the early Kazantsevo, *A. baicalensis* then prevailed throughout the upper part of MIS 5.

Today, the Holocene seasonal insolation contrasts are the lowest of the last 300 ky, and accordingly the current Lake Baikal diatom flora is of a similar low diversity as that experienced during MIS 13-MIS 9e (Khursevich et al., 2001). The extinction of *S. grandis* and *S. carconeiformis* occurred at some point in the last glacial after MIS 5, perhaps in response to the changes in seasonal insolation described above. This kind of insolation forcing in diatom species development is in agreement with the long Pleistocene record of Khursevich et al. (2001) who found that major diatom variations/extinctions were aligned to variations in the Earth’s orbit eccentricity.

All peaks in diatom concentrations occurred in periods of enhanced chemical weathering of primary minerals as indicated by spectral, mineralogical and rock magnetic analyses. Consequently, to exclude a possible direct inter-relationship of diatom and mineralogical records in further analysis, it is important to estimate which mineralogical or rock-magnetic parameters are not explicitly related to the lake bioproduc-tivity by post-depositional processes, in other words, which mineral parameters are likely to reflect only chemical weathering in watershed. The concentration of ferrimagnetic particles, evaluated by ARM, is not safe from this point of view. Post-depositional reductive dissolution of magnetite by organic matter can decrease ARM in periods of increased lake productivity. Fe$_{TOT}$.
$R_{\text{Vis}}$ and $R_{\text{UV}}$ reflectance was also excluded from further interpretation as they are strongly related to the concentration of diatom frustules within the sediment.

4. Discussion

4.1. Interpretation of the diatom record

Chemical weathering within the Lake Baikal watershed is an essential source of silicic acid for the water column. The oxidation of Fe$^{2+}$ in aluminosilicates to Fe$^{3+}$ is explicitly related to the oxidation and leaching of excess cations during mica transformation to vermiculite and/or smectite and their possible further hydrolysis (dissolution). B/C, however, is not correlated to the diatom proxies and total Fe content (Table 1). In several zones the B/C parameter decreases before the onset of large diatom concentrations, as if certain levels of silicate weathering were required to supply sufficient silicic acid to allow the subsequent diatom boom, although after a few ky lag. That lag has recently been related to the seasonal shifts of the perihelion and perhaps different humidity and temperature optima in the interglacials (Prokopenko et al., 2006). This trend is perhaps different humidity and temperature optima in the interglacials (Prokopenko et al., 2001a), and most recently Prokopenko et al. (2006). As Khursevich et al. (2001) noticed, the diatom assemblage from MIS 9a to MIS 5 was highly stable with no dramatic taxa extinctions. This interval is marked by a well defined insolation pattern with interglacials and interstadials falling into periods of high seasonal differences caused by high eccentricity, i.e., resulting in summer insolation above and winter insolation below present levels. This increased seasonality, or climate continentality would have enhanced the vertical mixing of the water column during spring and early summer, believed to be pre-requisites for the growth of large-cell taxa $S.\ grandis$, $S.\ carconeiformis$, and $C.\ baicalensis$ (in MIS 5e), and enabled fast transportation of nutrients into the photic zone. Edlund and Stoermer (2000) and Rioual and Mackay (2005) supposed that the extent large cell diatoms required clear winter ice cover and deep water mixing. Since $S.\ grandis$ and $S.\ carconeiformis$ were present throughout the analysed section including the MIS 8 glacial, we assume that they are able to prevail even in environmentally poor condition when turbulence may not have been high. This is reiterated by their presence even in glacial aged samples. Alternatively, it is also possible that these extinct taxa are more resistant to dissolution than other taxa. Concentrations of $C.\ minuta$, $C.\ ornata$, and $S.\ formosus$ var. $minor$ peaked in periods with decreased seasonality, i.e., low Northern Hemisphere summer insolation minima (Fig. 6). Their maxima after the peak for the larger Stephanodiscus taxa can imply a decrease in the deep mixing of the water but may also reflect a reduced supply of nutrients due to climatic worsening in the later part of the interglacial or interstadial. CEC decreased in humid environment, with peaks in Aulacoseira tax in MIS 5 coincident with humidity maxima (CEC minima) at 105–100 and 85–78 ky BP (Fig. 6). Conversely, peaks in Cyclotella taxa at 99–96 and 76–72 ky BP coincided with high CEC values, i.e., a probably drier climate but with chemical weathering still relatively high as indicated by low amounts of magnetic minerals, Fe$^{3+}$/Fe$^{2+}$, and relatively low level of Fe$^{2+}$/Fe$^{3+}$ in silicates. The same pattern of high CEC and low Fe$^{2+}$/Fe$^{3+}$ is also found in MIS 7a during a shallow local maximum of $C.\ minuta$. The characteristic succession of Cyclotella species in the Kazantsevo interglacial was also accompanied by the same CEC and Fe$^{2+}$/Fe$^{3+}$ pattern with less abundant concentrations of expandable clay minerals in the early stage of MIS 5e at 128 ky BP coincident with low numbers of $C.\ minuta$. After 123–122 ky BP, however, the amount of $C.\ minuta$ and $C.\ ornata$ started to increase with a peak between 121 and 117 ky BP.

4.2. The duration and stability of the Kazantsevo interglacial (MIS 5e)

The course of MIS 5 and other climatically warm periods are shown in Fig. 6. From at least 140 ky BP the indice of chemical weathering reveals a stepwise increase toward a maxima at 128–126 ky BP. Oscillating sedimentary conditions (MS), sharp spikes of mica content (XRD), oscillations of Fe$^{2+}$/Fe$^{3+}$ content and clay mineralogy between 136 and 129 ky BP and the residues of the late glacial MIS 6 assemblage of $C.\ minuta$ and $ornata$ accompanied deglaciation. At 128 ky BP, the lower Kazantsevo stage started with a
sharp increase in *A. skvortzowii* and the omnipresent *Stephanodiscus* species including *S. formosus* var. minor and *S. grandis*. Between 128 and 122 ky BP, diatom assemblage were probably the most diverse between MIS 8 and MIS 5 with assemblages including typical MIS 5e species such as *S. flabellatus* and *C. krammeri*, stable concentrations of *A. skvortzowii* and several further *Cyclotella* species. During this lower Kazantsevo stage mica, vermiculite and total expandable clay minerals were at their lowest levels, probably due to the high humidity, which would have led to intense hydrolysis of aluminosilicates. At 122 ky BP there was a minima in the occurrence of all major diatom species, after which *S. flabellatus* and *C. krammeri* vanished from the diatom assemblage while *C. baikalensis* and *C. ornata* appeared in significant concentrations and *S. formosus* increased with respect to concentrations in the lower Kazantsevo. In this upper Kazantsevo zone between 122 and 117 ky BP, the amount of expandable clay minerals grew continuously while Fe$^{2+}$/Fe$^{3+}$ in silicates remained at low, interglacial, values. This apparent discrepancy is explained by increased levels of vermiculite (HT-XRD), an expandable clay mineral formed by mild oxidative weathering of dark micas that had probably been further hydrolysed (dissolved) in the lower Kazantsevo. *C. minuta* and *C. ornata* peaked just before the end of the diatom zone at ∼117 ky BP, reflecting the shift of the bioproductivity maximum from spring to late summer or autumn in accordance with the lowered inter-seasonal insolation contrast. This may perhaps also reflect a limited input of nutrients to the lake from 116 ky BP caused by a decrease in chemical weathering as glacial conditions became established.

The duration and stability of the last interglacial have initiated a vivid discussion in the last decade leading to similar questions about the palaeoclimatic changes recorded in the Lake Baikal record. The Kazantsevo interglacial in the sediment record analysed here lasted from 128 to about 117 ky BP. Similar ages are also reported by Rioual and Mackay (2005), Tarasov et al. (2005), Granoszewski et al. (2005) and Grygar et al. (2006). Such duration and timing are almost synchronous with estimates from marine δ$^{18}$O records (Shackleton et al., 2002) and North Atlantic sea surface temperature records (McManus et al., 1999; Kandiano and Bauch, 2003). The Kazantsevo as recorded in Lake Baikal hence ended simultaneously with an important short, but very dramatic, climate deterioration in Central Europe at 118 ky BP (Sirocko et al., 2005), which is close to the Northern Hemisphere summer insolation minimum at 116 ky BP. Elsewhere in Europe, the Eemian or MIS 5e optimum ended between 120 and 110 ky BP depending on the latitude of the analysed site with the climatic deterioration earlier in Northern Europe and later in Southern Europe (Goni et al., 2005). The sharp MIS 5d minima in sea surface temperature in the North Atlantic (cold events C24 and C23) occurred much later in the North Atlantic between 108 and 102 ky BP (Chapman and Shackleton, 1999; Shackleton et al., 2002) continuing heat transportation long after the insolation minimum (McManus et al., 2002).

A dramatic decrease in diatom opal was found by Prokopenko et al. (2001a) at 122 ky BP and was interpreted as a dramatic mid-Kazantsevo cooling. Contrariwise, a rather small-scale climatic worsening at 120 ky BP followed by very moderate cooling was inferred from diatom analyses by Rioual and Mackay (2005). Irrespective of the magnitude of the changes and its precise dating, the change was accompanied by a shift in the diatom community to rarer taxa. A decrease in diatom concentrations and biovolume between DAZ4 and DAZ5 at 120 ky BP in Rioual and Mackay (2005) shows similarities to our record with decreases in *S. flabellatus* and increases in concentrations of *S. formosus*, *S. formosus* var. minor and *C. minuta*. Khursevich et al. (2001) found similar changes at the Academician Ridge in their stratigraphical denotation of diatom zones with *A. skvortzowii* in LDAZ 6 (lower) completely replaced by *A. baikalensis* in LDAZ 5 (upper) while *S. flabellatus* was exclusively found in LDAZ 6, *C. baikalensis* exclusively in LDAZ 5 and with changes in *C. minuta* and *C. ornata* again similar to our record. *S. grandis*, *S. carconeiformis* and *S. formosus* were present throughout the MIS 5e interval. As such the climate deterioration in the last part of the Kazantsevo must be evaluated from other taxa such as *Aulacoseira* and *Cyclotella* species and non-biogenic proxies which appear to be more responsive to climatic changes. From this, trends indicating drier conditions were found at the end of MIS 5 by Grygar et al. (2006). Chebykin et al. (2004) assumed that the deficiency of nutrients in the lake was due to very low riverine input, inferred from U-series isotopes, led to the decrease in diatoms at the end of MIS 5. More specifically, they found two successions from *A. baikalensis* via *S. grandis* to *C. minuta* accompanied by a substantial decrease of the riverine input of U to the lake from 100 to ∼93 ky BP (corresponding to MIS 5c) and from ∼83 to 74.5 ky BP (MIS 5a).

The Kazantsevo (128–117 ky BP) was the climatic optimum of the studied core section. The diversity of the diatom assemblage peaked during this period with the hydrolysis of aluminosilicates also at its most intense, probably due to the high humidity also occurring in this interval. Recently Edlund (2006) noted that the diatom diversity, otherwise generally very low in Lake Baikal...
over the last 500 ky, was at its highest in the Kazantsevo, MIS 3 and in the Holocene. This may be due to the intensive chemical weathering in the watershed over this interval, resulting in the high influx of nutrients into the lake. A typical feature of the Kazantsevo diatom assemblage is the occurrence of peak diatom concentrations for most taxa relative to other periods in the analysed interval. In addition, new species emerged including *S. flabellatus*, *A. baicalensis*, and *C. baicalensis*. Furthermore, MIS 5 was the only period between MIS 8 and MIS 4 where relatively high concentrations of *Aulacoseira*, *C. minuta* and *C. ornata* prevailed alongside the more dominant *S. grandis* and *S. carconeiformis*, indicating the spread of bioproductivity to all seasons and the presence of early ice-free dates in addition to reduced spring and autumn ice/snow cover. Change in the diatom assemblages in the upper part of the Kazantsevo can therefore be attributed to decreasing seasonality, due to orbital forcing, combined with a stepwise decrease in the intensity of the chemical weathering as conditions switched towards the subsequent local glaciation (117–107 ky BP) and the expansion of ice cover/thickness over the lake.

### 4.3. Chemical weathering minima

Extremes in B/C, log(ARM) and MS were found at about 280, 230, 185, 115, and 65 ky BP, i.e., mostly in periods when summer insolation at 55°N dropped below 460 W/m (Figs. 5 and 7) and when these minima were close to low obliquity. These weathering minima did not correspond to similar peaks in SPECMAP, as shown in Fig. 7 where grey rectangles, indicating the chemical weathering minima, are always coincident with diatom barren intervals. The mean marine δ18O values in these intervals were not always extremely high with the first signs of cooling Lake Baikal during MIS 8 and MIS 6 (zones I and III, respectively) occurring before the end of the MIS 7e and MIS 5e warm intervals as indicated by the marine δ18O record (zones II and IV, respectively). In addition, the decrease in diatoms and increases in chemical weathering in Lake Baikal at the end of MIS 5e (zone V) appear to have occurred rapidly with an immediate switch from interstadial or interglacial to glacial conditions. Such fast switches are most likely caused by the development of a local glaciation in a form of river valley glacier in the Lake Baikal catchment. This sensitivity to cooling during decreased summer NH insolation is similar to the response of high-latitude N Atlantic region, and North Europe ice sheet growth soon after interglacial climatic optima.

During the mild glacial MIS 8, XRD and CEC show values indicative of “cold-regime” conditions in spite of moderate chemical weathering (B/C, ARM, MS) and two diatom maxima. The lower diatom maxima was marked at its onset by increases in *Aulacoseira* sp. before finishing with assemblages being dominated by *S. grandis*. The second peak is dominated by *S. grandis*. The lower diatom peak ends at ~290 ky BP when North Atlantic sea surface temperatures were almost as high as during MIS 5c and MIS 5a (Kandiano and Bauch, 2003) while in Lake Baikal both lake productivity and indices of chemical weathering were similar to those experienced in MIS 7e. In the Northern Hemisphere, MIS 8 was less severe than the MIS 6 and MIS 2 glacials, both in terms of global ice volume and North Atlantic sea surface temperatures (McManus et al., 1999; Hiscott et al., 2001; Kandiano and Bauch, 2003). In Lake Baikal the intensity of chemical weathering increased toward the end of these glacials. Again, lake productivity followed changes in insolation rather than global ice volume or North Atlantic temperatures, although in Lake Baikal during MIS 6 there was a small diatom peak dominated by the warmer water taxa *C. minuta* and *C. ornata* and smaller concentrations of *S. grandis* at about 150 ky BP when the North Atlantic underwent the lowest sea surface temperatures of the penultimate glaciation (Kandiano and Bauch, 2003).

### 4.4. Comparison of MIS 7 and MIS 5

Three warm periods were identified in MIS 7 (Figs. 4 and 5). Although the vast majority of the diatom species in MIS 7 are today extinct, some conclusion can be drawn. The diatom assemblage from MIS 7c to MIS 7a resembles that of the upper peak in diatom concentrations during the MIS 8 glacial while mineralogical proxies indicate much weaker chemical weathering than in the corresponding MIS 5 sub-stages. An almost identical content of biogenic SiO2 in MIS 7 and MIS 5 was reported by Prokopenko et al. (2001a). Similarly, in our record the content of P-XRD amorphous sediments in the three diatom peaks in MIS 7 was almost as high as in the MIS 5e peak. However, the hydrolysis of expandable clay minerals was much weaker, particularly in MIS 7a, indicating most likely a less humid environment. The diatom record and indices of chemical weathering clearly indicated that MIS 7e was the climatic optimum of MIS 7. There were no sharp climatic oscillations in the upper part of MIS 7 such as those found in the upper parts of MIS 5, indicating that the end of MIS 7 was a fairly stable period.

Such an evaluation of the MIS 7 climate in Lake Baikal and its watershed is similar to European continental sequences (Reille et al., 2000; Tzedakis et al., 2004) and also to other records from Lake Baikal (Prokopenko et al.,
The MIS 7e climatic optimum was rather short and ended with a very harsh cold period, MIS 7d, both in the European continent and in the North Atlantic (Reille et al., 2000; Kandiano and Bauch, 2003; Tzedakis et al., 2004). The stadial MIS 7b was much less pronounced than MIS 5b at our site, similar to the pollen and diatom records from Academician Ridge (Sakai et al., 2005). The stadial corresponding to MIS 7b in Europe was also weakly pronounced in pollen record from Central France (Reille et al., 2000) while a very shallow minimum in North Atlantic sea surface temperature was found in MIS 7b with the whole of the MIS 7c to MIS 7a interval characterised by a rather uniform warm period (Kandiano and Bauch, 2003). North Atlantic sea surface temperatures, though, were significantly lower in MIS 7 than in MIS 5e (Kandiano and Bauch, 2003).

In West European continental sequences, the MIS 7e optimum was much shorter than MIS 5e (Tzedakis et al.,
2004). A similar pattern is also present in Lake Baikal for MIS 7e and MIS 5e (Figs. 4 and 5). Although the diatom peak of MIS 7e lasted about 11 ky, it started with a relatively long, ∼4 ky, long period dominated by *Aulacoseira* taxa and without significant hydrolysis of expandable clay minerals. This was followed by a ∼5 ky minor peak in *Aulacoseira* taxa coeval with a major peak in *S. grandis* and other *Stephanodiscus* species. The latter ∼5 ky long period should hence be considered the climatic optimum of the MIS 7 interval. The relative shortness of the climatic optimum in Lake Baikal during MIS 7e may be due to two intense iceberg discharges from the Laurentide ice sheets (Hiscott et al., 2001) resulting in extremely low sea surface temperatures in the North East Atlantic Ocean region in MIS 7e compared to other MIS 7 warm intervals in the Northern Hemisphere. An alternative explanation could be that the MIS 7e insolation maximum was not as pronounced with the low eccentricity of the Earth’s orbit resulting in the onset of an interglacial which was not as intensive as the corresponding onset of MIS 5e.

5. Conclusion

Orbital forcing is very strongly pronounced in the Lake Baikal region due to its extreme continentality, high latitude and weak marine influence. Orbital forcing, namely the summer Northern Hemisphere insolation at 55°N, can be tightly related to diatom peaks (summer insolation maxima) and chemical weathering minima in the watershed (summer insolation minima, especially in periods of low obliquity) and was used to construct an age model, confirmed by the paleomagnetic dating. Neither the total number of diatom valves, total diatom bioweights nor the amount of biogenic opal can be used alone to interpret the palaeoclimate/palaeoenvironment. Instead, conditions can be evaluated from combining mineral proxies of chemical weathering and taxonomic diatom analysis to the species level. Diatoms were very abundant not only in interglacials and interstadials, but also in the insolation maxima of the MIS 8 glacial. Increased diatom diversity across all species indicates the extension of lake productivity and climatically more favourable environmental conditions during the climatic optima of MIS 7e, MIS 5e and some shorter periods from MIS 5c to MIS 5a. Two “inorganic” proxies of watershed chemical weathering not explicitly affected or controlled by lake productivity are the oxidation state of Fe in silicates and oxides, measured by diffuse reflectance spectroscopy, and the content of expandable clay minerals, obtained by determining the cation exchange capacity.

Although periods of the harshest glacial climate did not coincide with δ18O maxima in marine foraminifera records, sharp climatic changes in the North Atlantic region between the middle of MIS 5e and MIS 5a are well expressed in the Lake Baikal sediment record. The extent and magnitude of local glaciation in the Baikal watershed, however, do not correspond well with the average extent of global glaciation. The combination of diatom analysis and weathering indices indicates that within the interval from MIS 9a to MIS 4, humidity was highest in MIS 5e and in the short periods between MIS 5c and MIS 5a and moderate in MIS 7e, MIS 8e and the early MIS 8 interstadial. In contrast, the least humid periods were the interstadials MIS 8c, MIS 7c and MIS 7a. Unusually the MIS 8 glacial was very moderate compared to the early half of MIS 6, the climatically harshest period in the Lake Baikal region. Furthermore, in contrast to MIS 8 which was interrupted by two interstadials of relatively intensive chemical weathering and increases in lake productivity, MIS 6 was interrupted by only a single, very weak, increase in lake productivity.

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