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Weathering in the Lake Baikal watershed during the Kazantsevo (Eemian) interglacial: Evidence from the lacustrine clay record

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Abstract: The clay-mineralogical record of a piston core recovered on an elevated plateau in the northern basin of Lake Baikal has been investigated for the Kazantsevo interglacial period (i.e., Eemian s.s. equivalent in northern Europe). The age model (as inferred from palaeomagnetic intensity) suggests that this stage spans ca. 128 to 117 kyr BP. Relative clay mineral abundances and clay-mineral ratios are used to reconstruct the weathering conditions within the Baikal watershed at a sub-millennial resolution, and suggest that the clay record is highly variable. A bimodal clay-data distribution is in agreement with different clay sources and/or formation between the studied glacial and interglacial periods. High amounts of smectites in the Taz glacial samples (128.7-136.4 kyr BP) may be explained by an additional source of neoformed smectites during the glacial stage. In addition to the classically used smectite/illite ratio, the mineralogical results are integrated by the calculation of a hydrolysis index that takes into account the abundance of all clay species and their sensitivity to chemical weathering. A principal components analysis (PCA) of the Baikal clay minerals allows the comparison of the clay parameters with regard to weathering conditions. Clay data are further compared (i) with diatom and pollen profiles, (ii) with pollen-based quantitative reconstructions for the same core material, and (iii) with other climate reconstructions for the Lake Baikal region and Siberia. Several features of our record are highlighted here. During the early period of the Kazantsevo interglacial (128.4-125.2 kyr BP), weathering processes remain controlled by physical reworking for more than 2 kyr after the initial transition from cold to warm conditions. Inception of chemical weathering starts only after ca. 125 kyr BP, a period coincident with the warmest conditions according to both the pollen record and by the strongest chernozem development in Siberian soils. Within the interglacial interval, the hydrolysis index displays a two-step increase, punctuated by a minimum value ca. 122 kyr BP. The increasing but irregular trend persists after the transition from the Kazantsevo interglacial to the Zyryanka glacial (~ 117 kyr BP). Peak chemical weathering, as inferred by clay changes, lags the interglacial/ glacial transition by at least 2 kyr. This suggests that pedogenesis remains active after the interval of surface stabilization. Lake Baikal clay minerals trace the nature of the main weathering conditions within the watershed. We note any increase in physical weathering is rapidly recorded in sedimentary clay assemblages but the mineral imprint to chemical weathering changes is more gradual, lagging reconstructed climate conditions over the lake by ca. 2 kyr.

Keywords: Clay minerals; Weathering condition; Last interglacial; Lake Baikal; Siberia

1. Introduction

Climate reconstructions inferred from spatially diverse archives, including Greenland ice-cores (Dansgaard et al., 1993; Alley et al., 1995; Rasmussen et al., 2003), marine sediments (Cortijo et al., 1994; Keigwin et al., 1994) and lacustrine records (Thouveny et al., 1994; Rioual et al., 2001; Tzedakis et al., 2003) raise the question of the climate stability during the last interglacial, and in particular the occurrence of a mid-Eemian cooling (e.g., Karabanov et al., 2000; Sirocko et al., 2005).

In Siberia, climate reconstructions for the last inter-glacial (Eemian s.s. or Kazantsevo according to European or Siberian stratigraphy, respectively) were mainly derived from pollen analyses (e.g., Velichko, 1984; Arkhipov and Volkova, 1994; Volkova and Bakhareva, 1995), loess deposition and soil complexes (Vorobyova, 1990; Chlachula, 2003; Chlachula et al., 2004). Based on palaeofloristic reconstructions of 25 localities, the last interglacial climate over Siberia was wetter and warmer than present (Velichko, 1984). In the Lake Baikal area, annual average precipitation was estimated be greater by ca. +50 mm, January winter temperatures were ca.+4 to + 8 °C higher, although summer temperatures remained close to, or only slightly higher than present (ca.+2 to +4 °C in July). Overall, however, such reconstructions remain rather imprecise for the Baikal area due to the scarcity of data points in this region (Velichko, 1984). Lake Baikal sediments also constitute powerful archives of past climate especially over glacial and interglacial cycles (e.g., Williams et al., 1997; Grachev et al., 1998; Kuzmin et al., 2001; Kashiwaya, 2003). Palaeoproductivity during the Kazantsevo period has been investigated previously within the context of the Baikal Drilling Programme (BDP) at coring sites in both the southern (BDP-93/2) and northern basins (BDP-98, BDP96/2, VER-93/2; Grachev et al., 1997; Karabanov et al., 2000; Prokopenko et al., 2002). These studies indicated that Lake Baikal sediments recorded high diatom abundances and biogenic silica (mainly derived from diatoms) during the last interglacial interval. These studies also indicated that the interglacial spanned an interval between 127 and 115 kyr BP and was punctuated by a short mid-Eemian cooling at ca. 122 (Karabanov et al., 2000) or 120 kyr BP (Prokopenko et al., 2002). More recently, centennial diatom (Rioual and Mackay, 2005) and pollen (Granoszewski et al., 2005) records have been investigated of the Kazantsevo interglacial and associated transition periods from the northern basin of Lake Baikal. Both approaches indicate that the Kazantsevo extended between ca. 128 and ca. 117 kyr BP, with peak interglacial conditions after 124 kyr BP. More in-depth pollen-based quantitative reconstructions (Tarasov et al., 2005) suggested a sharp increase of precipitation soon after 128 kyr BP but only a gradual warming till 127 kyr BP. Temperatures and precipitation decreased after ca. 121 kyr but the moisture index remained high up until a sharp turn towards a drier climate after 118 kyr BP.

Here, we present a 500-yr-resolution clay-mineral record for the Kazantsevo interglacial (i.e., Eemian s.s. according to European stratigraphy) and associated transitions to the preceding (Taz or Saalian) and following (Zyryanka or Weichselian) glacial stages in the Lake Baikal region. Since clay minerals in sediments represent a record of weathering conditions in the watershed, they have been used as a proxy for palaeoclimate reconstruction (e.g., Chamley, 1989). In Lake Baikal, the long-term evolution of clay-mineral assemblages has been investigated for the last climate cycle at a few coring stations on Buguldeika Saddle (Melles et al., 1995; Yuretich et al., 1999) and the Academician Ridge (Fagel et al., 2003). To date, however, high-resolution studies have been performed only on the Holocene and last glacial periods (e.g., Horiuchi et al., 2000; Fagel and Boës, 2008-this issue). Increasingly, multi-proxy approaches to environmental reconstruction can help make more robust climatic interpretations (e.g., Lotter, 2003) and here we specifically compare our clay data with qualitative and quantitative reconstructions derived from biotic (diatom and pollen) proxies made on the same core material within the framework of the EU CONTINENT project (EVK2-CT2000-0057; http://continent.gfz-potsdam.de/front_content.php).

2. Materials and methods

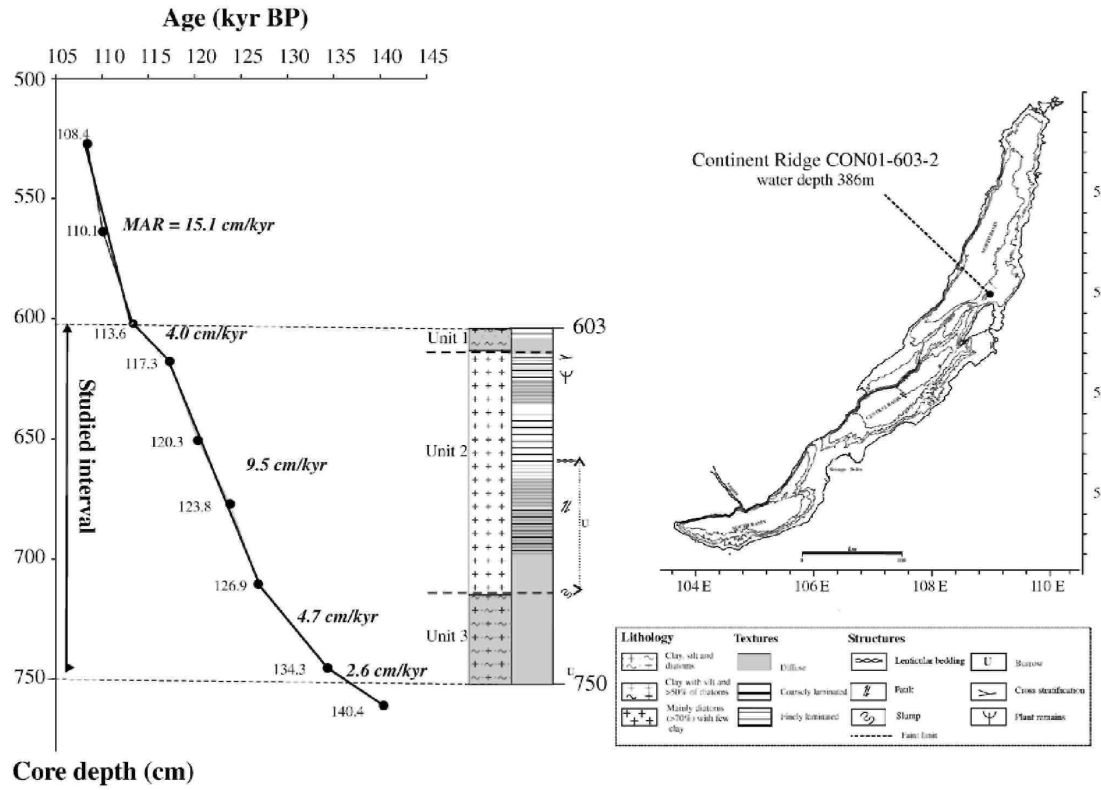
2.1. Lithology and chronology

The piston core CON01-603-2(53°57'N, 108°54' E) was recovered from a water depth of 386 m in the northern basin of Lake Baikal. This elevated plateau, selected after a seismic reconnaissance survey (Charlet et al., 2005), corresponds to a northern extension of the Academician Ridge and was called "Continent Ridge" (Fig. 1). In any palaeoecological study, core integrity is important. Core disturbance can come from a number of processes, including bioturbation, surface sediment reworking, turbidites and tectonics. While bioturbation (diffuse or punctual burrows) occurs especially in the lower part of the interval highlighted below, we know from contemporary studies that bioturbation does not have a large impact on core stratigraphy (e.g., Martin et al., 2005). Reflection seismic and side-scan sonar data both indicate that at this location surface reworking and evidence of turbidites are negligible, although there is limited evidence for tectonic features (Charlet et al., 2005; Fig. 1).

Based on smear-slide observations, we focus our study on a 147-cm interval, between 603 and 750 cm depth. This interval consists of three lithological units defined by gradual contacts (Fig. 1). The top 10 cm (unit 1: 603-613 cm) are composed of diffuse to finely laminated clayey sediments with silts, with frequent to few diatoms observed on smear slides (i.e., 10% < diatom abundance < 20%). The next core meter (unit 2: 613-712 cm) corresponds to a biogenic sedimentary interval composed of clayey silts to silty clays, and the core is finely to coarsely laminated. Smear slides indicate that diatoms are dominant to very dominant (abundance >60%) in this section. The lowest lithological unit (unit 3: 712-750 cm) is composed of diffuse silty clays with frequent diatoms (<20%).

The chronology of core CON01-603-2 is based on palaeomagnetic data tuned to a reference curve (Demory et al., 2005) from ODP Site 984 (Channell, 1999). The palaeomagnetic-derived age model was constrained by seven correlation points between 603 and 731 cm. An age model has been estimated for each sample by linear regression between adjacent reference points (Fig. 1). The applied age model suggests that the studied sediments accumulated between ~ 113.6-136.4 kyr BP (603-750 cm). The sedimentation rate, calculated by four linear regression lines (Fig. 1), averages at ca. 4 cm/kyr between 603 and 618 cm. It reaches 9.5 cm/kyr between 618 and 710 cm, then decreases to 4.7 cm/kyr between 710 and 745 cm and again to 2.6 cm/kyr between 745 and 750 cm (Fig. 1). Note that the changes in the estimated sediment-accumulation rates are in good correspondance with the observed lithological units, and are similar for the whole diatom-rich interval. The sampling resolution allows for a centennial resolution for most of the diatom-rich interval: 1 cm= ~90 years between 618-650 cm and 677-710 cm; 1 cm= ~ 130 years between 650-677 cm. The resolution is lower for the upper and lower clayey lithological units: 1 cm= ~ 250 years between 603 and 618 cm; 210 years between 710-745 cm and 390 years below 745 cm.

Fig. 1: *Core CON01+IBM-603+IBM-2: location map, sediment lithology and age model inferred from palaeomagnetic measurements (Demory et al.. 2005).*



2.2. Clay—mineralogy analyses

The clay approach defined for the complex Lake Baikal clay-mineral assemblages has been developed in a companion paper focused on the last glacial and Holocene (Fagel and Boës, 2008-this issue). Here we therefore give only a brief outline of the applied methodology. Qualitative and semi-quantitative estimations of clay-mineral assemblages are based on peak intensity measurements made on X-ray patterns processed on oriented aggregates (Moore and Reynolds, 1989). A few grams of dried bulk sediment were sieved under water at 30 µm. The <2 µm fraction was taken from the suspension after a settling time calculated according to Stake's law. The suspension was placed on a glass slide and dried overnight at room temperature. Routine XRD clay analyses included, in sequence, the recording of three X-ray patterns under air-dried or natural condition (N), after solvatation with ethylene-glycol during 24 h (EG), and after heating to 500 °C for 4 h (H). Semi-quantitative estimations (±510%; Biscaye, 1965) of the main clay species are based on the height of specific reflections measured in general on EG runs. The intensities are corrected by a weighting factor and values then summed up to 100%. Calculations of the relative abundances of the identified clay minerals are shown below in Table 1 (for full details see Fagel and Boës, 2008-this issue).

In order to take into account the whole complexity of the clay assemblage, we also calculated a hydrolysis index (HI). This parameter, first defined by Thorez (1985), integrates the abundance of each clay-mineral species and their weathering sensitivity (Jackson's stage number; Jackson, 1963), as follows: $HI=\sum_1^n$(abundance of clay species × stage number)/ $\sum_1^m$ (abundance of primary minerals) where n=number of secondary minerals, m=number of primary minerals.

Table 1: *Calculations of relative abundances of identified clay minerals*

Clay species	Relative abundance calculation
Illite	I 10 Å(EG) x l
Random 10-14 mixed-layers	I 12 Å(EG) x 2.5
Chlorite s.l.	I 14 Å(EG) x 2.5
Chlorite fresh	% Chlorite × I 14 Å(H)/I 14 Å(EG)
Weathered chlorite or vermiculite s.l	1% chlorite fresh
Al-smectite	I 17 Å(N) x 5
Swelling clay or smectites s.l	I10 Å(500)-I10 Å(EG) x l
Kaolinite	I 7Å(EG)x[I3.57 Å(N)/[I3.54 Å(N) + I3.57 Å(N)]] x 1.4
Smectite/illite ratio S/I (Biscaye, 1965)	I1 7Å(EG)/I10 Å(EG)

The calculation requires the designation of the parental minerals characterized by a stage number of 1 (i.e., fresh mineral). We retain the illite and the fresh chlorites as the two main parental minerals. We attribute a stage number to the secondary minerals according to their sensitivity to chemical weathering, according to Thorez (1985): 2 for 10-14 mixed-layer, 3 for degraded chlorite, 5 for smectite and 6 for Al-smectite. Note that we do not take into account the kaolinite abundance within the calculation as it does not vary throughout the two investigated sedimentary records. Baikal kaolinite is mainly a reworked mineral (Vogt and Larqué, 2002) probably derived from Mesozoic or Cenozoic sedimentary rocks, and as such its presence in terms of late deglacial climate changes is not relevant here.

In addition, relationships between the clay parameters themselves were investigated using the multivariate, indirect ordination technique of principal components analysis (PCA), on a correlation biplot (with centering and standardisation of the variables). This technique allows (i) dominant patterns in the dataset to be determined, and (ii) the relationships between the clay minerals through Termination 2 and the Kazantsevo interglacial to be explored. PCA was carried out using the software package CANOCO 4.5 (ter Braak and Smilauer, 2002).

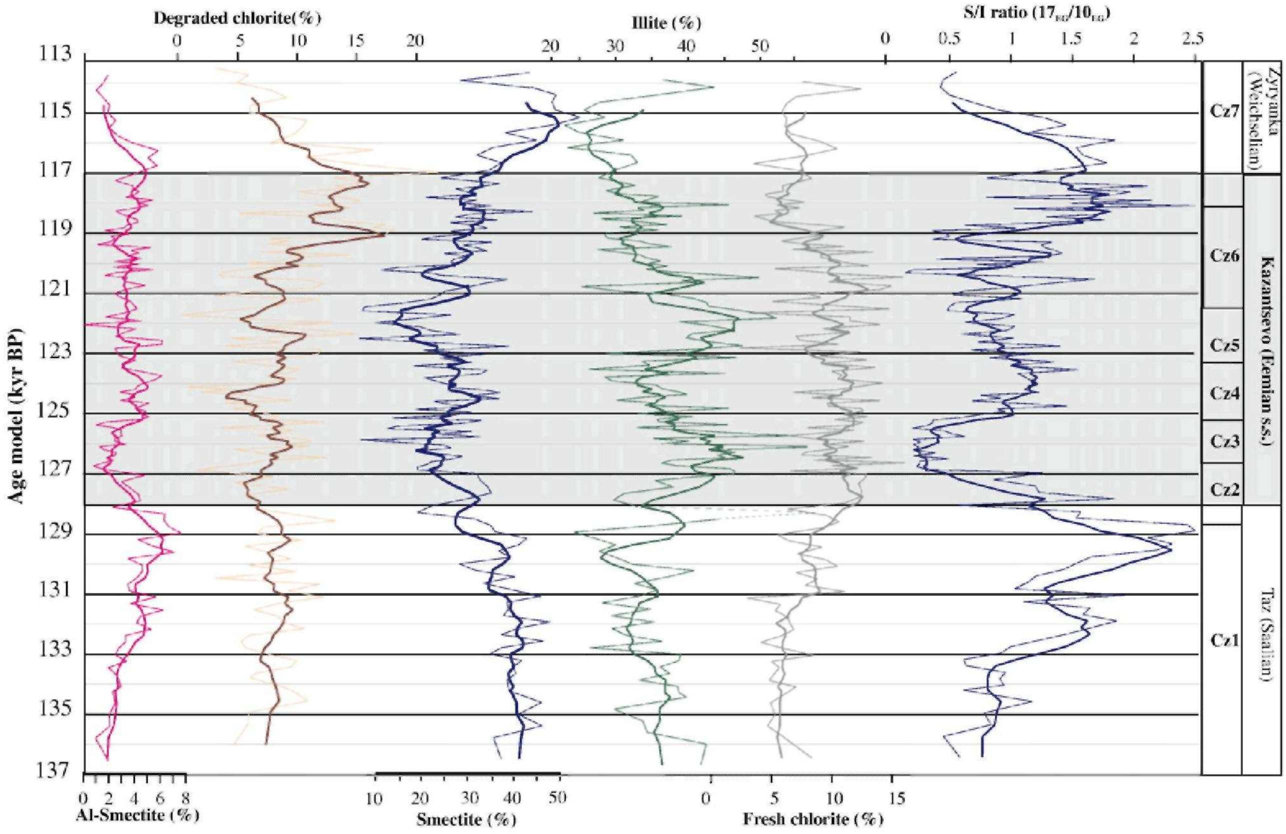
3. Clay-mineralogical results

3.1. Kazantsevo and transition clay record

The investigated clay-mineral record in CON01-603-2 spans the time period between ca. 136 and 114 kyr BP, i.e., the late part of the Taz glacial, the full Kazantsevo interglacial and the transition to the next glacial Zyryanka stage. In order to aid interpretation of the clay profiles (Fig. 2), we identify by eye 7 units or clay zones (CZ) based mainly on changes in the smectite/ illite (S/I) ratio. These zones are labelled from 1 to 7 from the base of the studied interval at 750 cm (~ 136 kyr BP) to the top at 603 cm (~ 114 kyr BP; Table 2).

Unit CZ1 (750-719 cm, ca. 136.4-128.7 kyr BP) is marked by an increase of the S/I ratio from values of around 0.5-0.8 in the lower part to a maximum of 2.5 between 722 and 719 cm (~ 129 ky BP). The abundance of smectite remains stable and high (mean of 3 8% ±6). The illite abundance is low (mean value 34% ±5 at lσ standard deviation), with an upcore decreasing trend.

Fig. 2: *Evolution of the clay-mineral assemblages of the studied interval 603-750 cm presented along model age (Demory et al., 2005), core CON01-603-2 from the northern basin of Lake Baikal. The bold curve represents the 5-point running average. The Kazantsevo interglacial is underlined by the dashed area.*



Other clays are present in similar abundances, with a mean of 8% ±3 for degraded chlorite, 7% ±2 of fresh chlorite, 7% ±2 of kaolinite. Minor Al-smectite (4% ±2) and traces of mixed-layers (2%±1) complete the assemblage. The mean hydrolysis index value (HI, Fig. 3) is high with a broad range of variation (mean 7 ± 2).

We observe a pronounced decrease in the S/I ratio to 0.2 in unit CZ2 (718-710 cm, ca. 128.4-126.8 kyr BP). It reflects a two-step increase of illite up to 45% and a concomitant decrease of smectite down to <20%. There is an opposite behaviour between smectite and illite but the relationship is not perfect (Fig. 2). The highest amount of illite (>55%) is observed at 716 cm (ca. 128 kyr BP), close to the lithological limit between clayey and diatom-rich sediments. The fresh chlorite fraction also increases whereas degraded chlorite and Al-smectite decrease. Like S/I, HI also significantly decreases (mean 4 ± 1).

Unit CZ3 (709-692 cm, 126.7-125.2 kyr BP) is defined by a stable and low S/I ratio (~0.2): the decrease of illite is counter-balanced by a slight increase of smectites from ~20 to 30%. The illite abundance is highly variable from one sample to another in this short interval (33%<illite<58%), but the 5-point running average depicts a significant decrease, evolving from 45% to 35%. This interval is marked by relatively higher values of degraded chlorite. The HI values remain low (mean 3 ± 1) during the whole interval.

The S/I ratio increases within a short period of time at the base of unit CZ4 (691-673 cm, ca. 125.1-123.2 kyr BP), after which it remains more or less stable (~ 20-25%). S/I gradually decreases in unit CZ5 (672-658 cm, ca. 123.1-121.3 kyr BP). Such evolution mimics the curve of smectite abundance. We note low mean values of smectite and the lowest individual values (<10%) in an interval between 668-658 cm (minimum ca. 122.4-122.5 kyr BP). CZ5 is characterized by the second major peak in illite (45-50%). HI depicts a similar curve than S/I, with mean value higher or lower than 4 in CZ4 and CZ5, respectively.

The next unit CZ6 (657-625 cm, 121.1 -118 kyr BP) is characterized by an irregular increase of the S/I ratio up to 2.5 (at 626 cm). The trend is interrupted by two short negative excursions at 650-649 cm (ca. 120.2 kyr BP) and 638-635 cm (ca. 119 kyr BP). The degraded chlorite reaches its highest abundance (>19%) within this interval. HI increases by a factor of 4 within CZ6. The scattering of the HI values increases again. The 5-point running average depicts three negative excursions at ca. 120, 118.5 and 117.5 kyr BP, and systematically lags (by a few samples) the lowest mean S/I values.

The last unit CZ7 (624-603 cm, 117.9-113.6 kyr BP) starts with high, stable S/I values, reflecting a final increase of smectite in our record peaking at ca. 115 kyr BP. Then the S/I ratio decreases down to 0.5, due to an increase in illite. HI reaches its highest value ca. 115 kyr BP, with an overall high mean value of 7±2.

Table 2: *Core CON0l-603-2: mean and standard deviation of clay-mineral abundance and clay parameters for the defined clay zones CZ1 to CZ7*

CON0l-603-2			**Relative clay mineral abundance (%)**												**Chlorite group (%)**				**S/I**		**HI**	
			Illite		**Chlorite**		**10-14**		**sm. s.l.**		**Al-sm.**		**Kaolinite**		**Degraded chl.**		**Fresh chl.**		**$17_{EG}/10_{EG}$**			
Unit	**Depth (cm)**	**Age (kyr BP)**	**Mean**	**1σ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**	**Mean**	**lσ**
CZ7	603-624	113.6-117.9	32	6	18	4	2	1	37	9	3	1	7	2	11	5	7	2	1.3	0.6	6.6	2.0
		min-max	23-46		11-29		1-3		21-54		1-6		5-10		3-23	3	3-12		0.4-2.1		3-11	
CZ6	625-657	118-121.1	35	6	21	3	3	1	28	7	3	1	9	2	12	4	9	3	1.1	0.5	4.8	2.0
			25-50		16-30		1-5		11-44		1-6		6-12		3-19	3	4-16		0.1-2.5		2-9	
CZ5	658-672	121.3-123.1	42.2	7	18	3	5	5	20	10	3	2	12	2	8	5	10	4	0.8	0.3	3.1	1.0
			29-52		9-30		2-22		7-39		0-6		8-18		0-16		2-15		0.5-1.4		1-6	
CZ4	673-691	123.2-125.1	36.2	6	18	3	3	1	27	7	4	1	11	2	7	3	11	2	1.1	0.3	4.1	1.0
			26-46		11-26		1-4		14-37		3-6		8-15		1-12		7-14		0.6-1.5		2-6	
CZ3	692-709	125.2-126.7	43	7	19	2	3	1	23	7	2	1	10	2	9	3	10	3	0.3	0.1	3.1	1.0
			33-58		16-23		2-5		7-33		1-3		6-13		4-13		6-16		0.20.6		1-5	
CZ2	710-718	126.8-128.4	38.2	8	18	3	3	1	28	6	4	1	9	2	6	2	11	2	1.2	0.5	4.0	1.0
			30-56		16-21		1-4		19-35		2-6		7-12		1-10		6-13		0.3-1.9		2-5	
CZ1	719-750	128.7-136.4	34	5	15	2	2	1	39	4	4	2	6	1	8	3	7	2	1.3	0.6	6.3	2.0
			24-43		11-19		1-8		28-48		1-8		3-8		3-14		3-11		0.5-2.5		4-10	

3.2. Principal components analysis of clay data

PCA summary results are shown in Table 3. Axes 1 and 2 account for 62% of variation in the clay-mineral data, whereas axis 1 alone accounts for almost 45% within the whole dataset. The first axis, or component, represents the combination of clay variables that accounts for the greatest amount of variation in the dataset. Axis 2 on the other hand is independent of axis 1 and accounts for the next largest amount of variation in the clay dataset. The PCA of the clay minerals are plotted in Fig. 4, as a correlation biplot: closed circles represent samples deposited during the Kazantsevo interglacial. The relationships between the variables can be explained by following a few biplot rules (ter Braak and Smilauer, 2002). The direction of the vector indicates increasing values, with the mean value for every variable scaled to the centre of the biplot where axis 1 and axis 2 intersect. Variables that are highly correlated with each other have vectors with small angles between them, approaching 0°. Variables that are negatively correlated with each other have angles between their vectors approaching 180°. However, variables that are uncorrelated to each other have angles between their vectors approaching 90°. Finally, the proximity of a vector or variable to one of the axes highlights its importance along that axis. In Fig. 4, axis 1 is most closely associated with HI and smectite to the left of the biplot (sediments deposited during glacial periods), and 10-14 mixed layers and illites to the right of the biplot (sediments deposited during the Kazantsevo interglacial). The small angle, approaching 0°, between degraded chlorite and smectite indicate that these clays are highly correlated. Axis 2 on the other hand is closely associated with changing concentrations of Al-smectite.

4. Discussion: origin and processes of clay formation

4.1. Glacial versus interglacial clay sources

Within the studied glacial/interglacial transitions, illite and smectite abundances depict an approximate linear and continuous trend in the CON01-603-2 clay fraction (overall linear regression r^2=0.62*;* Fig. 5). Illite values range between < slope<-1.3), except CZ2 (slope -0.6).

The data distribution is in agreement with different clay sources and/or formation between the studied glacial and interglacial period. The opposite behaviour between illite and smectites (e.g., see relationships highlighted in the PCA; Fig. 4) favors a genetic link between those clays for the interglacial samples. This suggests that most of the smectites are secondary products, formed by pedogenesis of parental illite. Variable intensities of weathering explain the continuous data distribution, indicative that the parental material is mineralogically homogeneous. According to the PCA, the clay minerals are clustered in three groups (Fig. 4): (1) Illite, fresh chlorite, 10-14 mixed layers and kaolinite; (2) smectite and degraded chlorite and, (3) Al-smectite. PCA results are in agreement with our interpretation of Baikal clays. The first group includes the primary minerals supplied to the lake by physical weathering. The second group is made by two secondary clays derived by chemical weathering during pedogenesis. There is little apparent relationship in this dataset between smectites and Al-smectite as underlined by the angles between their vectors approaching 90° (Fig. 4).

However, the absence of a clear relationship between illite and smectites in glacial CZ1 samples rather suggests that there is no genetic link between illite and smectites (Fig. 5). Perhaps surprisingly the amount of smectites is higher in the lower glacial samples (mean ca. 136.4-128 kyr BP = 38±6) than in the interglacial ones (mean ca. 128-117.4 kyr=26±8). This may perhaps be explained by the existence of an additional source of smectites during glacial periods. Indeed Vogt and Larqué (1998, 2002) already observed neoformed smectites in different periglacial conditions, and in Siberia in particular. Neoformation is explained by the confined environment under the permafrost.

CZ2 data points are scattered along two clearly defined linear trends (Fig. 5). The intermediate distribution in CZ2 mostly reflects the transitional position of this clay unit, between the Taz glacial stage and the Kazantsevo interglacial (ca. 128 kyr BP). There is no glacial-like clay distribution in CZ7. However, the Zyryanka glacial is undersampled with respect to the Taz stage (~ 9 kyr). Only the early Zyryanka is analyzed (~ 2 kyr) and it could not be representative of the average glacial stade.

4.2. Weathering type and intensity during the Kazantsevo in the northern basin of Lake Baikal

The clay record is highly variable at the centennial resolution (Fig. 2). As for the Holocene interglacial, we suspect that the intra-sampling variability partly reflects the accuracy of the XRD method on diatom-rich sediments (see Fagel and Boës, 2008-this issue). We will restrict our interpretation on weathering conditions on several running point averages rather than on individual sample data. Taking into account the estimated sedimentation rates, a 5-point running average allows for a millennium-scale order resolution in the lower and upper glacial intervals (e.g., at the base of the Kazantsevo (CZ2)), while the resolution increases to ~ 500 years during the intervening interglacial.

The mean signature of the clay unit depicts an unusual evolution through time (Table 2; Fig. 6). The smectite versus illite abundance (or S/I ratio) decreases from unit CZ1 towards unit CZ3, then increases again up to CZ7, with a reverse trend during CZ5. Such trends could a priori suggest lower chemical weathering during interglacial than during glacial stages. But this interpretation is in contradiction with regional pedogenesis conditions favouring active hydrolysis and smectite formation during warmer conditions (Chlachula et al., 2004). According to the observed different relationships between smectites and illite (see above), the high values of S/I are most probably biased by the neoformed origins of smectites during glacial confined environments, i.e., periods with extensive permafrost. An estimation of the neoformed fraction of smectites within the Baikal clay assemblages was tested by clay pre-treatments (Thorez, 1989). However K-saturated XRD runs do not present any difference at 10 A under different heating conditions (K-110 and K-330 °C). The tests were not decisive, which may be due to the complex clay assemblages present, a matrix effect (biogenic silica; Thorez, 1989) and/or low XRD resolution. Therefore, defined clay parameters could be used as chemical weathering proxies only within the interglacial interval.

Fig. 3: *Evolution of two clay parameters sensitive to chemical weathering. S/I (diamonds) is the classically smecite/illite peak-height ratio measured on the ethylene glycol XRD run. The hydrolysis index (squares) is a calculated parameter taking into account the abundance of all the clay minerals and their respective sensitivity to chemical weathering. The bold curves correspond to the 5-point running average. The Kazantsevo interglacial is underlined by the dashed area.*

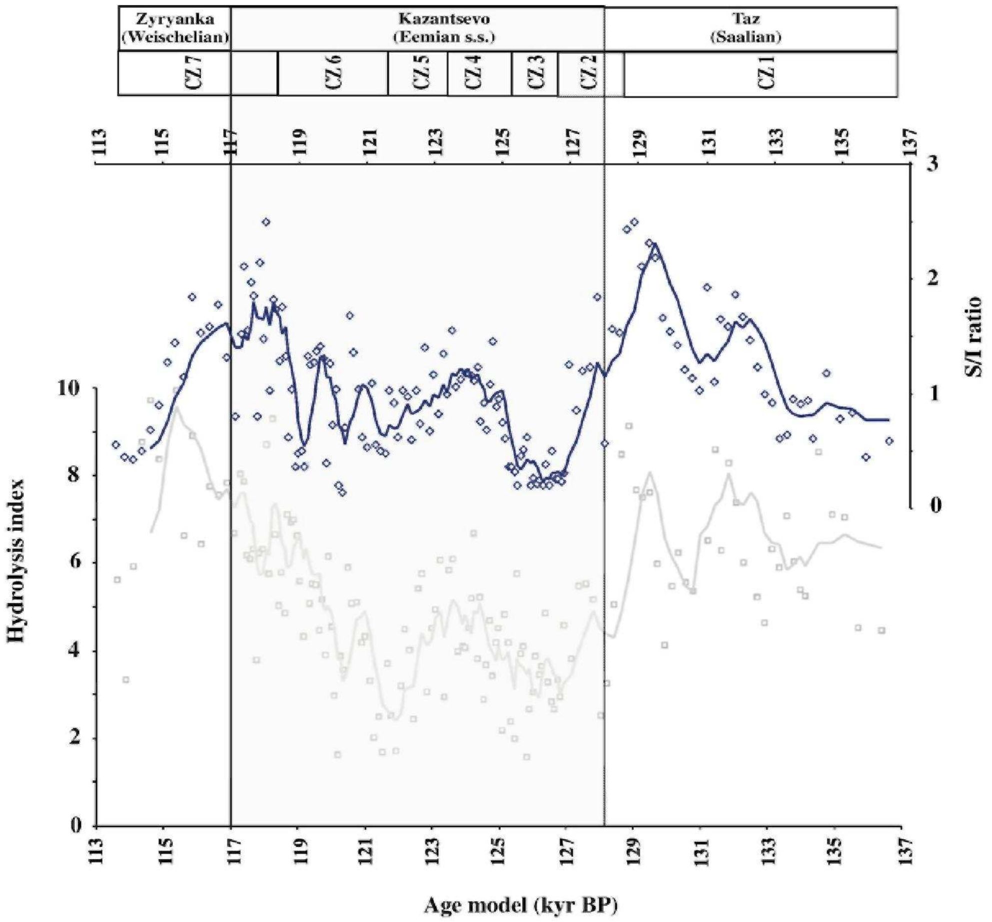


Table 3: *Summary table of principal components analysis (PCA) of the clay-mineral dataset from the Kazantsevo and associated transitions in core CON01-603-2, Continent Ridge, northern basin of Lake Baikal*

	PCA Axes				
	1	**2**	**3**	**4**	**Total variance**
Eigenvalue	0.447	0.174	0.148	0.101	1.000
P	0.01				
Cum. % variance of explanatory data	44.7	62.0	76.9	86.9	
Sum of all eigenvalues					1.000

Fig. 4: *Results of the Principal Component Analysis (PCA) of the clay database. The x-axis represents axis-1 of the ordination analysis, and the y-axis, axis 2 of the ordination. Axis 1 is by far the dominant component accounting for 45% of the variation in the clay dataset. Axis 1 is most closely associated with the hydrolysis index and smectites to the left of the correlation biplot, and illites and 10-14 mixed layers to the right of the biplot. Filled circles represent sediments deposited during the Kazantsevo interglacial, and open circles represent sediments deposited during glacial periods.*

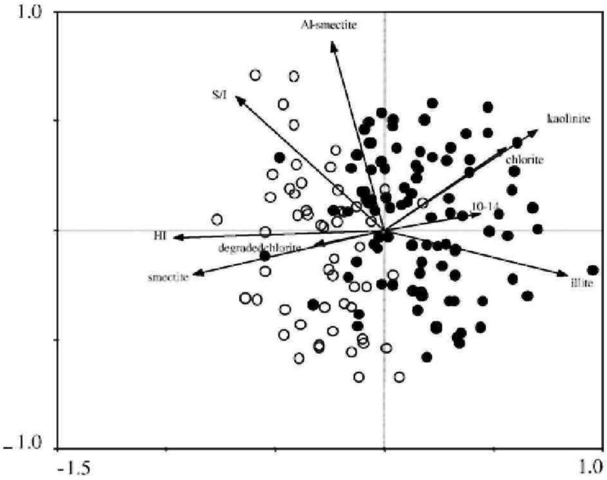
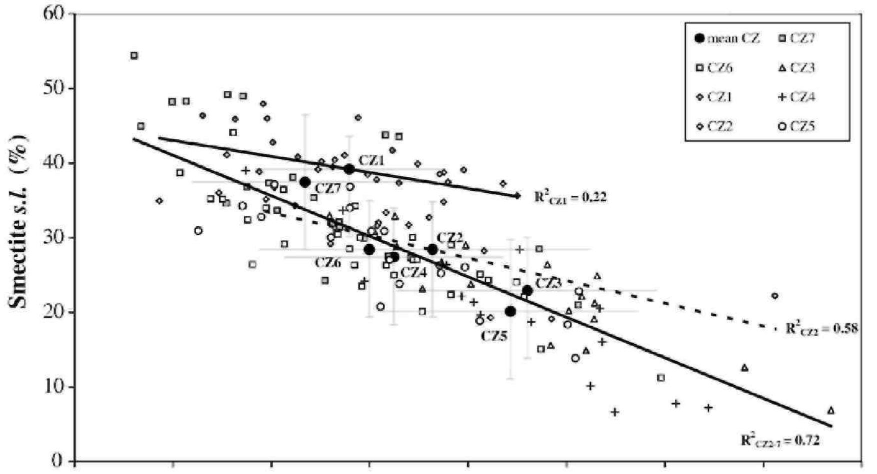


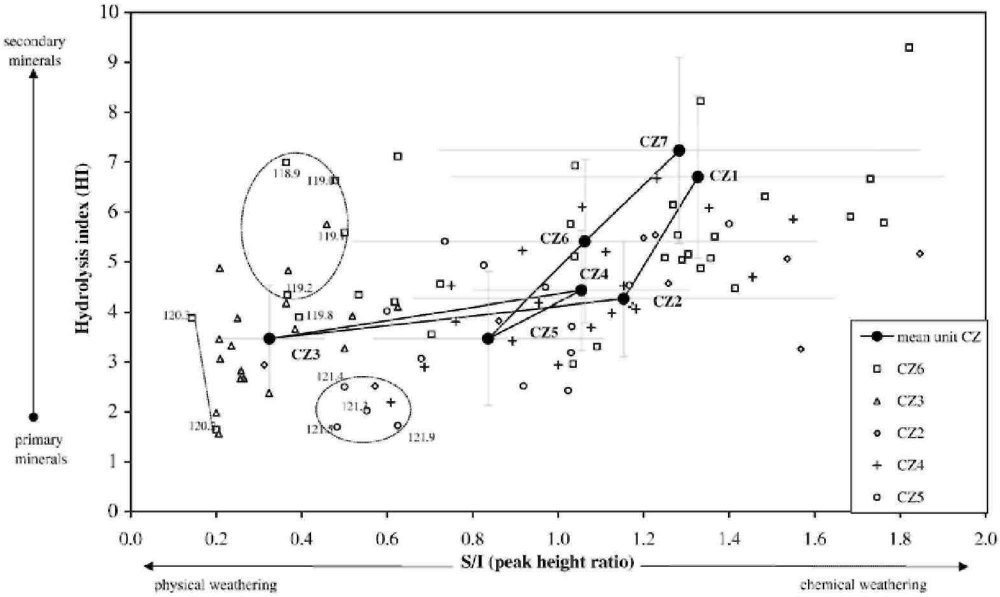
Fig. 5: *Smectites versus illite abundance (relative %). The bold lines correspond to linear regressions for clay samples. Note the location of CZ1 samples that correspond to the glacial Taz stage (older than 128 kyr BP). The CZ7 samples representing the next glacial stage fall in the field of the CZ1 sample. However, the Zyryanka stage (younger than 117 kyr BP) is undersampled in the clay database: our sampling interval ending at 603 cm does not cover the next full glacial conditions. Indeed a large decrease in diatom valve accumulation rate and increased proportions of benthic taxa were especially obvious only above 600 cm (Rioual and Mackay, 2005).*



Within the interglacial, the low S/I values (Fig. 2) underline limited secondary mineral formation in soils but active physical weathering. CZ3 and CZ5 clays delivered to Lake Baikal are enriched in primary minerals, reflecting soil destabilization by physical weathering. The clay supplies remain illite-rich for a few centuries at ca. 126 and 122 kyr BR CZ3 and CZ5 are characterized by similar, relatively low hydrolysis index (~3.5; Fig. 3). In contrast, S/I is very different for these two clay units while their illite content is similar (~ 45%). The relatively high CZ3 HI value integrates the chlorite weathering pathways: fresh chlorite is less abundant during CZ3 whereas the amount of degraded chlorite increases. It seems that a preliminary but reduced chemical weathering occurs during CZ3, in addition to an intense physical weathering. In the Baikal watershed, chlorites seem to be more sensitive to weathering than illite, in agreement with their respective weathering numbers assigned by Jackson (1963). During CZ5, illite and fresh chlorite represent up to 70% of the clay assemblage, while all the secondary clay minerals remain low. This interval corresponds to a physical reworking period.

The clay parameters are variable during CZ6, then become more stable during CZ4 (Fig. 3). These clay units are characterized by similar mean S/I values (~ 1 ; Fig. 6). In both clay units, the relatively high S/I values suggest a general reduced control of physical weathering over chemical weathering. The S/I exhibits an irregular increase during CZ6, which is perhaps reflective of instability in environmental conditions within the Lake Baikal watershed. The higher mean HI value in CZ6 records the chemical weathering of all kinds of primary minerals, chlorite and also illite. Gradual improvement in pedogenesis is punctuated by reactivation of physical erosion at ca. 120 and 119 kyr BP. At ca. 120 kyr BP, the S/I decrease is associated with a HI decrease, which suggests a physical reworking event and limited chemical weathering. At ca. 119 kyr BP, S/I decreases but not HI, reflecting a remaining chemical weathering of chlorite in addition to increased erosion.

Fig. 6: *Comparison of the measured (S/I) and calculated (HI) clay parameters.*



4.3. Climatic significance of the clay record in northern Lake Baikal

As highlighted above, the sedimentary record of clay parameters at the Continent Ridge in the northern basin of Lake Baikal may be indirectly interpreted in terms of climate conditions only during interglacial conditions, i.e., from the middle of CZ2 to CZ6. In particular, the smectite/ illite ratio and hydrolysis index are both indicative of the chemical weathering intensity within the watershed. Within the Kazantsevo interglacial, three intervals are characterized by low S/I and HI values (Fig. 6): in CZ3 ca. 125-127 kyr; in CZ5 ca. 121.5 kyr; in CZ6 ca. 120 kyr BP.

During those time intervals, clay supplies enriched in primary minerals are mainly driven by mechanical erosion processes. Such weathering conditions are favoured by cold and dry climate. It is worth noting here that pollen reconstructions (pollen zones CK3, CK5b-I and CK5b-II in Granoszewski et al., 2005) also demonstrate colder conditions atthese times. In contrast, the high values of S/I and HI parameters reflect intense chemical weathering under warm and wet conditions, leading to secondary clay-mineral formation in soils. Moderated chemical weathering starts within CZ4, i.e., between ca. 125— 124 kyr BP. The climate warms within CZ6, from ca. 121.5 kyr BP. HI reaches its maximum mean value (~ 9) at ca. 115 kyr BP, lagging the Kazantsevo/Zyryanka transition by ca. 2 kyr. Since the lacustrine clay fraction records not only weathering processes but also erosion and transport, some delay could occur between climate changes and sediment imprint, and this factor may be more relevant especially when undertaking higher-resolution climate reconstructions.

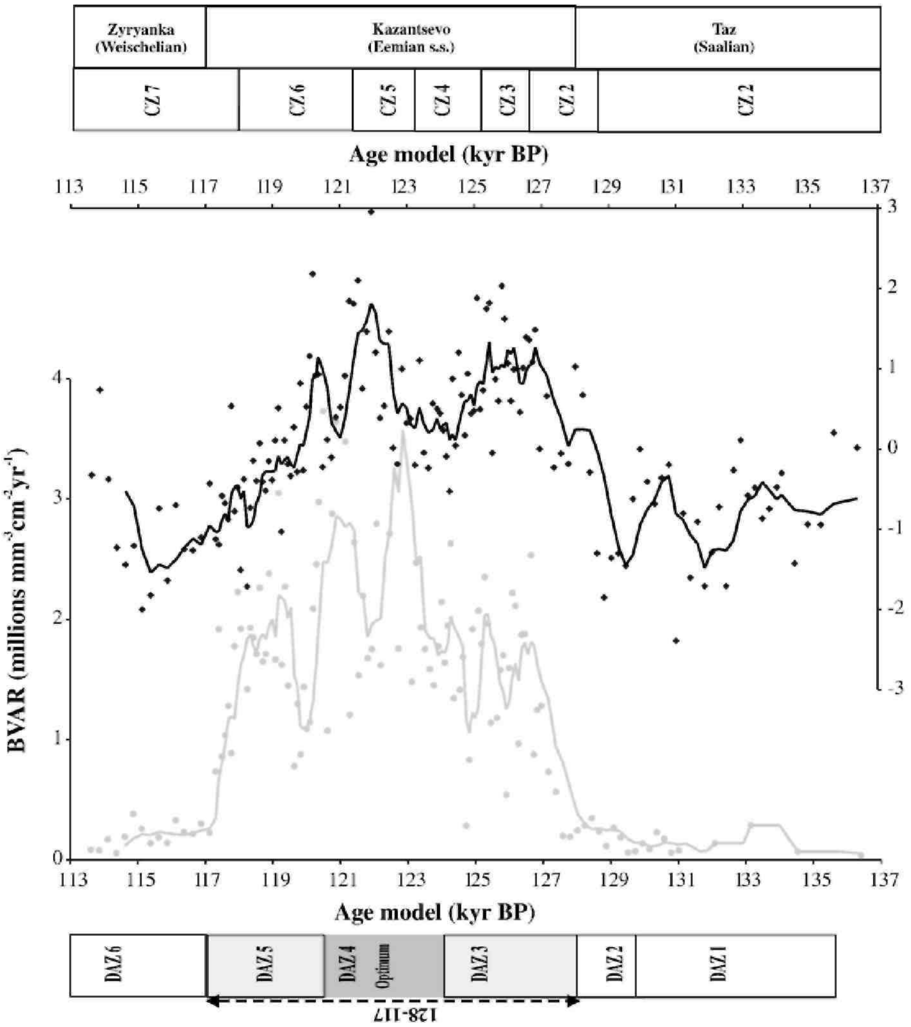
The palaeoclimate significance of the clay record at the Continent Ridge site can be more fully assessed by taking a multi-proxy approach, comparing clay-derived climate-variability interpretation with diatom- and pollen-inferred palaeo-reconstructions. The comparison allows for a validation of the clay approach and for an estimate of any time lag between weathering processes and detrital particle settling.

The major trends in the clay palaeorecord from CON01-603-2 are summarized by plotting PCA-axis-1 sample scores against diatom biovolume accumulation rates (BVAR), as previously determined by Rioual and Mackay (2005) (Fig. 7). In this figure, and as shown above, high positive values of axis-1 sample scores correspond to high amounts of detrital clay minerals, reflecting active physical weathering processes. In contrast, negative PCA-axis-1 scores are driven by high hydrolysis-index values, indicating strong chemical weathering. By focussing on selected time periods, several patterns emerge between diatom and pollen inferred data and the clay record. During the late Taz glacial up to the beginning of the Kazantsevo interglacial (136.4-128 kyr) negative values of PCA-axis-1 scores (ca. CZ1 and lower CZ2) correspond with diatom-assemblage zone (DAZ) 1, during which time diatom BVARs are very low, suggesting prevailing colder climate and reduced productivity in the lake. However, at ca. 129 kyr BP, the clay record suggests a shift to more active, physical weathering processes. This precedes major increases in productivity (associated with large increases in biovolume), but this shift does coincide with an increase in numbers of the endemic taxon *Aulacoseira skvortzowii* suggesting that at this time, productivity in the near-shore regions of the lake was increasing. These changes are also mirrored in quantitative estimates of pollen-inferred precipitation, which increase sharply at the onset of the Kazantsevo (ca. 128 kyr BP), followed by more gradual increases in pollen-inferred temperatures (Tarasov et al., 2005).

Thereafter, positive clay-record PCA-axis-1 scores throughout the Kazantsevo up to the end of the interglacial at 118 kyr BP are coincident with both diatom and pollen-inferred warm and humid conditions (Rioual and Mackay, 2005; Tarasov et al., 2005). However, there are notable and significant shifts in the PCA clay record. For example, the initiation of chemical weathering in the northern Baikal watershed is recorded in the sediment clay fraction only after 125 kyr BP (within CZ4). This period corresponds to the initiation of chernozem development in Siberian soils (dated at 125 ±5 kyr BP) because of progressive climatic warming (Chlachula et al., 2004). In Southern Siberia, the last interglacial is characterized by a succession of loess-palaeosols: the landscape stabilization is indicated by the formation of distinct soil horizons formed during warm and humid intervals separated by thin loess units accumulated during dry, cold stages. At 125 kyr BP, chernozem formation requires warmer and more humid climate conditions (Chlachula et al., 2004), such as those inferred here by pollen analyses (Granoszewski et al., 2005; Tarasov et al., 2005).

After a second period of active physical erosion within CZ5 (contemporaneous with a pollen-inferred cooling in CK 5b-I; Granoszewski et al., 2005), the clay proxies, especially the hydrolysis index, start to increase again at ca. 121.5 kyr BP (Fig. 3). The PCA-axis-1 scores decrease from +2 to -1 (Fig. 7). This improvement in chemical weathering conditions is not sustained by warmer local conditions: the relative temperature index curve decreases in value by ca. 123 kyr (Granoszewski et al., 2005); the mean reconstructed temperature gradually shifts to cooler values after ca. 121 kyr BP (Tarasov et al., 2005). However, the reconstructed moisture availability remains high until 117.4 kyr BP. The increasing but irregular trend starts after the climate optimum (i.e., DAZ4 characterized by maximum diatom productivity between c.124-120.4 kyr BP; Fig. 7). Moreover, the trend persists after the transition from the Kazantsevo interglacial to the Zyryanka glacial (until 115 kyr BP): the optimum chemical weathering inferred by clay changes lags by at least 2 kyr the interglacial/glacial transition at Continent Ridge. Those 2 kyr could reflect the time needed for re-equilibration of weathering processes in soils due to new glacial conditions, e.g., permafrost development. This suggests that pedogenesis remains active after the interval of surface stabilization. Indeed, the study of landscape development indicated a major distortion of last interglacial palaeosol documented over Siberia during the following major cooling at ca. 111 kyr BP (Chlachula et al., 2004). In the Lake Baikal area, soil stratigraphy consists in a cryoturbated unit corresponding to the interglacial maximum. The extent of cryoturbation indicates a rapid permafrost development "soon after" the interglacial (Vorobyova, 1990). The onset of the late Pleistocene glacial conditions (Zyryanka stage) occurred between ca. 117.5 and 114.8 kyr BP as indicated by cool, steppe-dominated pollen assemblages and climate reconstruction (Tarasov et al., 2005). The pronounced decrease in hydrolysis index after 115 kyr BP is therefore likely to be indicative of major changes in conditions of pedogenesis.

Fig. 7: *Comparison of CON01-603-2 clay data with diatom data. On the left: Continent Ridge diatom zonation (DAZ) and curve of biovolume accumulation rate (BVAR, Rioual and Mackay, 2005). On the right: a summary of the clay data represented by PCA-axis-1 sample scores and clay zonation. The black bold line corresponds to the 5-point running average of the clay data points (diamonds). The grey bold line corresponds to the 5-point running average of the diatom data points (circles).*



Finally, within the increasing hydrolyzing trend, a negative excursion occurs at ca. 120 kyr BP in CZ6 samples. Characterized by the lowest S/I and HI values of the entire interglacial interval (Fig. 6), this change supports a punctual physical reworking occurring within the watershed. Our clay data on Continent Ridge records then the intra-Eemian cooling event, which has also been identified at ca. 121-120 kyr BP in other Lake Baikal sediments (Karabanov et al., 2000; Prokopenko et al., 2002), and in other regions from around the world, including the Chinese loess (An and Porter, 1997), European lakes (Sirocko et al., 2005) and the sub-polar North Atlantic Ocean (e.g., McManus et al., 2002). The interpretation of the pollen profile in the Continent Ridge core also indicates a brief cooling at ca. 120 kyr BP (i.e., CK 5B-II) confirmed by the pollen-inferred quantitative reconstruction after 121 kyr BP (Tarasov et al., 2005). The diatom-accumulation rate is also marked by a pronounced drop in biovolume accumulation rate in the Continent Ridge core. However, the biotic signal is not perfectly synchronous with the clay change highlighting differences between allochthonous and autochthonous processes in and around Lake Baikal and its watershed (Fig- 7).

5. Conclusion

Our clay record from northern Lake Baikal (at the Continent Ridge site, a northern extension of Academician Ridge) allows one to follow the evolution of the weathering conditions through the Kazantsevo interglacial within the Baikal watershed. In contrast to the stable climate conditions inferred from quantitative pollen reconstructions, sub-millennium-scale clay records (ca. every 500 yr) suggest rather more variable weathering conditions throughout the interglacial.

Clay-assemblage composition is mainly controlled by physical weathering processes (supplying primary clay minerals) during three time-periods (ca. 125-127, 122 and 120 kyr BP). These three periods correspond to the three cooling events recorded in the pollen profiles from the same sediment core. They are also systematically marked by drops in lake productivity, inferred from diatom biovolume accumulation rates. The youngest clay change ca. 120 kyr BP is coincident with the so-called mid-Eemian cooling previously observed in other sediment profiles from nearby Lake Baikal. According to the measured (S/I ratio) or calculated (hydrolysis index) clay proxies, the chemical weathering conditions depict an irregular two-step increasing trend through the interglacial. The highest chemical weathering conditions lags by ca. 2 kyr the interglacial/glacial transition. Such a delay may be due to the time needed for re-equilibration of weathering processes in watershed soils due to new glacial conditions.

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References

Alley, R.B., Gow, A.J., Johnsen, S.J., Kipfstuhl, J., Messe, D.A., Thorsteinsson, T., 1995. Comparison of deep ice cores. Nature 373, 393-394.

An, Z., Porter, S.C., 1997. Millennial scale climatic oscillations during the last interglacial in central China. Geology 25, 603-606.

Arkhipov, S.A., Volkova, VS., 1994. Geological history, Pleistocene landscapes and climate in West Siberia. Russian Academy of Sciences, Siberian Branch. United Institute of Geology, Geophysics and Mineralogy, Transaction, vol. 823. Novosibirsk (in Russian).

Biscaye, P.E., 1965. Mineralogy and sedimentation of recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans. Geological Society of America Bulletin 76, 803-832.

Chamley, H., 1989. Clay Sedimentology Springer-Verlag, Berlin. 623 pp.

Channell, J.E.T., 1999. Geomagnetic intensity and directional secular variation at Ocean Drilling Program (ODP) site 984 (Bjorn Drift) since 500 ka: Comparison with ODP983 (Gardar Drift). Journal of Geophysical Research B104, 22937-22951.

Charlet, R, Fagel, N, De Batist, M., Hauregard, R, Minnebo, B., Meischner, D., SONIC Team, 2005. Sedimentary dynamics on isolated highs in Lake Baikal: evidence from detailed high-resolution geophysical data and sediment cores. Global and Planetary Change 46, 125-144.

Chlachula, J., 2003. The Siberian loess record and its significance for reconstruction of Pleistocene climate change in north-central Asia. Quaternary Science Reviews 22, 1879-1906.

Chlachula, J., Kemp, R.A., Jessen, C.A., Palmer, A.P, Toms, PS., 2004. Landscape development in response to climatic change during the Oxygen Isotope stage 5 in Southern Siberia. Boreas 33, 164-180.

Cortijo, E., Duplessy, J.C., Labeyrie, L., Leclaire, H., Duprat, J., van Weering, T.C.E., 1994. Eemian cooling in Norwegian Sea and North Atlantic Ocean preceding continental ice-sheet growth. Nature 372, 446-449.

Dansgaard, W., Johnsen, S J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.G., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P, Sveinbjorns-dottir, A.E., Jouzel, J., Bond, G., 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. Nature 364, 218-220.

Demory, R, Nowaczyk, N.R., Witt, A., Oberhänsli, H., 2005. High-resolution magnetostratigraphy of late Quaternary sediments from Lake Baikal, Siberia: timing of intracontinental paleoclimatic responses. Global and Planetary Change 46, 145-166.

Fagel, N., Boës, X., 2008. Clay-mineral record in Lake Baikal sediments: the Holocene and Late Glacial transition. Palaeogeogra-phy, Palaeoclimatology, Palaeoecology 259, 230-243 (this issue). doi:10.1016/j.palaeo .2007.10.009.

Fagel, N., Boski, T, Likhoshway, L., Oberhaensli, H, 2003. Late Quaternary clay mineral record in Central Lake Baïkal (Academician Ridge, Siberia). Palaeogeography, Palaeoclimatology, Palaeoecology 193, 159-179.

Grachev, M.A., Likhoshwai, E.V., Vorobyova, S.S., Khlystov, O.M., Bezrukova, E.V., Veinberg, E.V., Goldberg, E.L., Granina, L.Z., Kornakova, E.G., Lazo, F.I., Levina, O.V., Letunova, P.P., Otinov, P.V., Pirog, V.V., Fedotov, A.P., Iaskevich, S.A., Bobrov, V.A., Sukhorukov, F.V., Rezchikov, V.I., Fedorin, M.A., Zolotarev, K.V., Kravchinsky, V.A., 1997. Signals of the paleoclimates of upper Pleistocene in the sediments of Lake Baikal. Russian Geology and Geophysics 38, 957-980.

Grachev, M.A., Vorobieva, S.S., Likoshway, E.V., 1998. A high-resolution diatom record of the paleoclimates of East Siberia for the last 2,5 My from Lake Baikal. Quaternary Science Reviews 17, 1101-1106.

Granoszewski, W., Demske, D., Nita, M., Heumann, G, Andreev, A.A., 2005. Vegetation and climate variability during the Last interglacial evidenced in the pollen record from Lake Baikal. Global and Planetary Change 46, 187-198.

Horiuchi, K., Minoura, K., Hoshino, K., Oda, T, Nakamura, T, Kawai, T, 2000. Paleoenvironmental history of Lake Baikal during the last 23000 years. Palaeogeography, Palaeoclimatology, Palaeoecology 157, 95-108.

Jackson, M.L., 1963. Interlayering of expansible layer silicates in soils by chemical weathering. In: Ingerson, E. (Ed.), Proc. clays and clay minerals 11th Nat. Confi 1962. Pergamon Press, Ottawa-Canada, pp. 29-46.

Karabanov, E.B., Prokopenko, A.A., Williams, D.F., Khursevich, G.K., 2000. Evidence for mid-Eemian cooling in continental climatic record from Lake Baikal. Journal of Paleolimnology 23, 365-371.

Kashiwaya, K.S. (Ed.), 2003. Long continental records from Lake Baikal. Springer Verlag, Berlin. 370 pp.

Keigwin, L.D., Curry, W.B., Lehman, S.J., 1994. The role of deep ocean in North Atlantic climate change between 70 and 130 kyr ago. Nature 371, 323-326.

Kuzmin, M.I., Karabanov, E.B., Kawai, T., Williams, D., Bychinskii, V.A., Kerber, E.V., Kravchinslii, V.A., Bezrukova, E.V., Prokopienko, A.A., Geletii, V.F., Kalmychko, G.V., Goreglyad, A.V., Antipin, VS., Khomutova, M.Yu., Soshina, N.M., Ivanov, E.V, Khursevich, G.K., Tkachenko, L.L., Solotchina, E.P., Ioshida, N.,Gvozdkov, A.N., 2001. Deep drilling on Lake Baikal: main results. Russian Geology and Geophysics 42, 8-34.

Lotter, A.F., 2003. Multi-proxy climatic reconstructions. In: Mackay, A.W., Battarbee, R.W., Birks, H.J.B., Oldfield, F. (Eds.), Global Change in the Holocene. Arnold, London, pp. 373-383.

Martin, P., Boës, X., Goddeeris, B., Fagel, N., 2005. A qualitative assessment of the influences of bioturbation in Lake Baikal sediments. Global and Planetary Change 46, 87-99.

McManus, J.F., Oppo, D.W, Keigwin, L.D., Cullen, J.L., Bond, G.C., 2002. Thermohaline circulation and prolonged interglacial warmth in the North Atlantic. Quaternary Research 58, 17-21.

Melles, M., Grobe, H., Hubberten, H.W., 1995. Mineral composition of the clay fraction in the 100 m Core BDP-93-2 from Lake Baikal — preliminary results. In: Horie, S. (Ed.), IPPCC Newsletter 9. Universitätsverlag Wagner, Innsbruck, pp. 17-22.

Moore, D.M., Reynolds Jr., R.C., 1989. X-Ray Diffraction and the Identification and Analysis of Clay Minerals. Oxford Univ. Press, Oxford. 378 pp.

Prokopenko, A.A., Karabanov, E.B., Williams, D.F., Khursevich, G.K., 2002. The stability and the abrupt ending of the Last Interglaciation in Southeastern Siberia. Quaternary Research 58, 56-59.

Rasmussen, T.L., Thomsen, E., Kuijpers, A., Wastegard, S., 2003. Late warming and early cooling of the sea surface in the Nordic seas during MIS 5e (Eemian Interglacial). Quaternary Science Reviews 22, 809-921.

Rioual, P., Andrieu-Ponel, V, Rietti-Shati, M., Battarbee, R.W., de Beaulieu, J.-L., Cheddadi, R., Reille, M., Svobodova, H., Shemesh, A., 2001. High-resolution record of climate stability in France during the last interglacial period. Nature 413, 293-296.

Rioual, P., Mackay, A.W., 2005. A diatom record of centennial resolution for the Kazantsevo Interglacial stage in Lake Baikal (Siberia). Global and Planetary Change 46, 199-219.

Sirocko, R, Seelos, K, Schaber, K, Rein, B., Dreher, R, Diehl, M., Lehne, R., Jäger, K, Krbetschek, M., Degering, D., 2005. A late Eemian aridity pulse in central Europe during the last glacial inception. Nature 436, 833-836.

Tarasov, P., Granoszewski, W., Bezrukova, E., Brewer, S., Nita, M., Abzaeva, A., Oberhänsli, H., 2005. Quantitative reconstruction of the Last Interglacial climate based on the pollen record from Lake Baikal, Russia. Climate Dynamics 25, 625-637.

ter Braak, C.J.F., Smilauer, P., 2002. CANOCO Reference Manual and CanoDraw for Windows User's Guide: Software for Canonical Community Ordination (version 4.5). Ithaca, NY, USA.

Thorez, J., 1985. Argillogenesis and the hydrolysis index. Mineralogica et Petrographica Acta 29, 313-338.

Thorez, J., 1989. Argilloscopy of weathering and sedimentation. Bulletin de la Société Géologique de Belgique 98, 245-267.

Thouveny, N., de Beaulieu, J.L., Bonifay, E., Créer, K.M., Guiot, J., Icole, M., Johnsen, S., Jouzel, J., Reille, M., Williams, W., Williamson, D., 1994. Climate variations in Europe over the past 140 kyr deduced from rock magnetism. Nature 371, 503-506.

Tzedakis, PC, Frogley, M.R., Heaton, T.H.E., 2003. Last interglacial conditions in southern Europe: evidence from Ionnina, northwest Greece. Global and Planetary Change 36, 157-170.

Velichko, A.A., 1984. Late Pleistocene spatial paleoclimatic reconstructions. In: Velichko, A.A. (Ed.), Late Quaternary Environments of Soviet Union. University of Minnesota Press, Mineapolis, pp. 261-285.

Vogt, T., Larqué, P., 1998. Transformations and neoformations of clay in the cryogenic environment: examples from Transbaikalia (Siberia) and Patagonia (Argentina). European Journal of Soil Science 49, 367-376.

Vogt, T., Larqué, P., 2002. Clays and secondary minerals as permafrost indicators: examples from the circum-Baikal region. Quaternary International 95-96, 175-187.

Volkova, V.S., Bakhareva, V.A., 1995. Climate of optimum Late Quaternary (Kazantsevo) interglacial in West Siberia and climate prognosis for the nearest future. Russian Geology and Geophysics 36, 23-33.

Vorobyova, G.A., 1990. Stratigraphy and paleogeography of south-central Siberia. In: Medvedev, G.I. (Ed.), Stratigraphy, paleogeography and archeology of south-central Siberia. Irkutsk University Press, Irkutsk, pp. 10-17 (in Russian).

Yuretich, R., Melles, M., Sarata, B., Grobe, H., 1999. Clay minerals in the sediments of Lake Baikal: a useful climate proxy. Journal of Sedimentary Research 69, 588-596.

Williams, D.F., Peck, J., Karabanov, E.B., Prokopenko, A.A., Kravchinsky, V., King, J., Kuzmin, M.I., 1997. Lake Baikal record of continental climate response to orbital insolation during the past 5 million years. Science 278, 1114-1117.