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Paleomagnetic dating of climatic events in Late Quaternary sediments of Lake Baikal (Siberia)

by
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Acknowledgments

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ABSTRACT

The intracontinental Lake Baikal is located far from the marine influences, providing an excellent climatic archive for Central Eurasia as global climatic variations are continuously depicted in its sediments. We performed continuous rock magnetic and paleomagnetic analyses (2743 samples) on hemipelagic sequences retrieved from 4 underwater highs reaching back 300 ka. The most complete sedimentary sequence covering 150 ka, located in the North Basin, was subject to continuous grain size analysis (448 samples).

The rock magnetic study combined with TEM, XRD, XRF and geochemical analyses evidenced that a magnetite of detrital origin dominates the magnetic signal in glacial sediments whereas interglacial sediments are affected by early diagenesis. They show biomagnetite-greigite in the case of fast burial and dissolution of magnetite in the case of slow burial. The samples with magnetite dissolution and greigite were skipped and excluded from paleomagnetic interpretations. HIRM roughly quantifies the hematite and goethite contributions and remains the best proxy for estimating the detrital input in Lake Baikal.

Relative paleointensity records of the earth's magnetic field show a reproducible pattern, which allows for correlation with well-dated reference curves and thus provides an alternative age model for Lake Baikal sediments. This age model is anchored by *AMS* ^{14}C dating for the first 15 ka as well as by paleomagnetic excursions like Laschamp (at about 42 ka) and Iceland Basin events (at about 185 ka), which preserved as a strong geomagnetic deviation and a full reversal respectively. According to the age model, paleomagnetic records have different resolution (120 years to 800 years) depending on the sedimentation rates. The average sedimentation rates ranging from 2.9 cm ka^{-1} to 19.4 cm ka^{-1}, largely dependent on the vicinity of the shore and, therefore, on detrital input. The age model shows that hiatuses occur. However, these are local since they cannot be correlated from one site to another.

Using the paleomagnetic age model we observed that cooling in the Lake Baikal region and cooling of the sea surface water in the North Atlantic, as recorded in planktonic foraminifera δ^{18}O, are coeval. On the other hand, benthic δ^{18}O curves record mainly the global ice volume change, which occurs later than the sea surface temperature change. This proves that a dating bias results from an age model based on the correlation of Lake Baikal sedimentary records with benthic δ^{18}O curves.

The compilation of paleomagnetic curves provides a new relative paleointensity curve, "Baikal 200". This stack, covering the last 200 ka, is a new paleomagnetic curve of reference for Central Eurasia.

With a laser-assisted grain size analysis of the detrital input, three facies types, reflecting different sedimentary dynamics can be distinguished. (1) Glacial periods are characterised by a high clay content mostly due to wind activity and by occurrence of a coarse fraction (sand) transported over the ice by local winds. This fraction gives evidence for aridity and little or no snow in the hinterland. (2) At glacial/interglacial transitions, the quantity of silt increases as the moisture increases, reflecting increased sedimentary dynamics. Besides mountain glacial melting and growing river input, wind transport and snow trapping are the dominant process bringing silt to a hemipelagic site (3) During the climatic optimum of the Eemian, the silt size and quantity are minimal due to blanketing of the detrital sources by the vegetal cover.

North Hemispheric teleconnections are evidenced. The sand and hematite (evidenced by rock magnetism) input events are synchronous with dust input in Greenland Ice during cold periods. This marks cold and arid events affecting both Siberia and Central Asia, the latter being one dust source for Greenland Ice. Teleconnections are also observed in Terminations I and II. The increased moisture around Lake Baikal during these periods results from more active Westerlies. This is probably due to new pathways linked to reorganisation of atmospheric pressure cells, when insolation increases.

KURZFASSUNG

Der Baikalsee liegt weit ab mariner Einflüsse und ist ein ideales Klimaarchiv für die Mitte Eurasiens. Klimatische Vartiationen sind kontinuierlich in den Sedimenten aufgezeichnet. In dieser Arbeit wurde gesteinsmagnetische und plaeomagnetische Analysen (2743 Proben) an hemipelagischen Sequenzen von vier Lokationen analysiert. Die Kerne erreichen ein Alter von maximal 300 ky. Die vollständigste sedimentäre Sequenz umfasst 150 ka und stammt aus dem N Becken. Diese wurde zusätzlich einer detaillierten Korngrössenanalyse unterworfen.

In Kombination mit TEM, XRD, XRF und weiteren geochemischen Analysen zeigt die gesteinsmagnetische Studie, dass detritischer Magnetit das magnetische Signal der glazialen Sedimente dominiert. Die magnetischen Signale der interglazialen Sedimente wurden durch diagenetische Prozesse verändert. Bei hoher Sedimentationsrate zeigen die interglazialen Sedimente Biomagnetit-Greigit. Bei niedriger Sedimentationsrate beobachtet man Lösung der Magnetite. Proben in denen gelöster Magnetit sowie auch Greigit auftraten, wurden von weiteren Betrachtungen ausgeschlossen. Mittels HIRM können Hämatit und Goethit quantifiziert werden. Diese Methode eignet sich, den detritischen Eintrag in den Baikalsee abzuschätzen. Relative Paleointenistäten des Erdmagnetfeldes ergaben reproduizerbare Muster, welche in Korrelation mit gutdatierten Referenzproben die Ableitung eines alternativen Altersmodells für die Datierung der Baikalsedimente ermöglichten. Dieses Altersmodell wurde für die ersten 15 ka mit der AMS-Radicarbon–Methode kalibriert. Zudem wurden paleomagnetische Exkursionen, wie das Laschamp (ca. 42 ka) und das Island Basin (ca 185 ka) Ereignis zur Kalibirierung verwendet. Dieses Altersmodell berücksichtigt unterschiedliche Auflösung der paleomagnetischen Daten als Funktion der Sedimentationsraten. In Abhängigkeit der Distanz zum Ufer und des detritischen Eintrags variiert die durchschnittliche Sedimentationsrate von 2,9 cm/ka^{-1} bis 19,4 cm/ka^{-1}. Zudem zeigt das Altermodell, dass Hiaten auftreten. Allerdings sind diese lokal und können nicht zwischen den Kernlokationen korreliert werden. Bei Anwendung des paleomagnetischen Altersmodells beobachtet man, dass die Abkühlung im Baikalgebiet und im Oberflächenwasser des Nordatlantiks wie sie aus den ?^{18}O-Werten planktonischer Foraminiferen abgeleitet werden kann, zeitgleich ist. Wird das aus benthischen ?^{18}O-Werten abgeleitete Altermdodell auf den Baikalsee angewandt, ergibt sich eine deutliche Zeitverschiebung. Das benthische Altersmodell repräsentiert die globale Veränderung des Eisvolumens, welche später als die Vänderung der Oberflächenwassertemperatur auftritt.

Die Kompilation plaeomagnetischer Kurven ergab eine neue relative Paleointensitätskurve "Baikal 200". Diese deckt die letzten 200 ka ab. und stellt eine neue paleomagnetische Referenzkurve für Zentraleurasien dar.

Mittels Korngrössenanalyse des Detritus konnten drei Faziestypen mit unterschiedlicher Sedimentationsdynamik unterschieden werden: 1) Glaziale Peroiden werden durch hohe Tongehalte infolge von Windeintrag und durch grobe Sandfraktion mittels Transport durch lokale Winde über das Eis charakterisiert. Dieser Faziestyp deutet auf arides Klima mit wenig oder kein Schnee im Hinterland. 2) Während der Glazial/Interglazial-Übergänge steigt die Siltfraktion an. Dies deutet auf erhöhte Feuchtigkeit und damit verbunden erhöhte Sedimentdynamik. Neben dem Abschmelzen von Gletschern und zunehmendem Flusseintrag, sind Windtransport und in den Schnee der Eisdecke eingetragener Staub die vorherrschenden Prozesse, welche den Silt in hemipelagischer Position zur Ablagerung bringen. 3) Während des klimatischen Optimum des Eeemian werden Grösse und Quantität des Silts minimal, was auf eine geschlossene Vegetationsdecke im Hinterland deutet.

Fernwirkungen in der Nordhemisphäre sind evident. Während kalter Perioden sind Sand und Hämatit Eintragsereignisse synchron mit dem Staubeintrag im Grönlandeis. Dies zeigt, dass kalte und aride Ereignisse sowohl Sibirien als auch Zentralasien beeinflussen. Zentralasien ist eine der Staubquellen für das Grönlandeis. Fernwirkungen können auch in Termination I und II beobachtet werden. Die Zunahme der Feuchtigkeit im Baikalgebiet deutet auf aktivere Westwinde hin. Dies wird wahrscheinlich durch erhöhte Insolation bedingt, was zur Reorganisation der atmosphärischen Druckzellen und damit verbunden, neuen Strömungstrajektorien führt.

RESUME

Le lac intracontinental Baïkal, éloigné des influences maritimes, constitue une excellente archive des variations climatiques globales. Nous avons effectué une étude magnétique et paléomagnétique continue sur des séquences hémipélagiques issues de quatre hauts géomorphologiques ennoyés et couvrant les derniers 300 ka, et une analyse granulométrique continue (448 échantillons) sur la séquence sédimentaire la plus complète couvrant les derniers 150 ka.
L'étude magnétique combinée à des observations au MET, à de la diffraction X et à de la fluorescence X ainsi qu'à des analyses géochimiques montre qu'une magnétite d'origine détritique domine le signal dans les sédiments glaciaires, alors que les sédiments interglaciaires sont affecté par de la diagenèse précoce. Dans ces derniers, un assemblage biomagnétite-greigite domine lorsque l'enfouissement est rapide tandis que la magnétite est dissoute lorsque l'enfouissement est lent. Les échantillons affectés par la dissolution de magnétite et ceux contenant de la greigite sont exclus des interprétations paléomagnétiques. L'HIRM, qui quantifie l'hématite et la goethite, reste le meilleur estimateur de l'apport détritique.
L'enregistrement des paléointensités relatives du champ magnétique terrestre montre un schéma reproductible qui permet des corrélations avec les courbes de références bien datées. De fait, il fournit un modèle d'age pour les sédiments du Lac Baïkal, ancré par datations ^{14}C pour les derniers 15 ka et par des excursions paléomagnétiques : Laschamp à ~ 42 ka et Iceland Basin à ~ 185 ka. Celles-ci sont respectivement enregistrées par une importante déviation géomagnétique et une inversion complète. Selon notre model d'age, les enregistrements paléomagnétiques ont des résolutions de 120 à 800 ans dépendantes des taux de sédimentation de 2,9 cm ka^{-1} et 19,4 cm ka^{-1} en fonction de leur proximité au rivage, et par conséquent, de l'apport détritique. Le modèle d'age révèle des hiatus de sédimentation locaux non corrélés d'un site à l'autre.
Sur la base du modèle d'age paléomagnétique, le refroidissement dans la région du lac Baïkal et le refroidissement de la surface de l'océan Atlantique Nord (révélé par le $\delta^{18}O$ des foraminifères planctoniques) sont simultanés. Comparé au $\delta^{18}O$ planctonique, le $\delta^{18}O$ benthique enregistre préférentiellement les changements de volume de glace qui surviennent plus tard que le refroidissement des eaux de surface. Ceci montre l'erreur de datation qu'implique une corrélation directe entre des enregistrements sédimentaires du lac Baïkal et les courbes de variation du $\delta^{18}O$ benthique.
La compilation des courbes paléomagnétiques issue des différents sites étudiés permet de proposer une nouvelle courbe de paléointensité relative « Baïkal 200 » couvrant les derniers 200 ka et constituant une nouvelle courbe de référence pour l'Asie Centrale.
L'analyse granulométrique au laser de la fraction détritique a fait apparaître trois faciès types : (1) un faciès caractéristique des périodes glaciaires, riche en argiles probablement éoliennes, avec présence de sables transportés par les vents locaux sur la glace couvrant le lac Baïkal Cette fraction indique un climat aride avec peu ou pas de neige; (2) un faciès riche en silts, corrélé à l'augmentation de l'humidité et à une dynamique sédimentaire plus forte. Cette fraction résulte partiellement de la fonte des glaciers et de l'apport accru par les rivières mais aussi du transport éolien et d'un piégeage des particules éoliennes dans la neige présente sur le lac Baïkal; (3) un faciès caractérisé par la taille et la quantité minimales de silt est typique de l'optimum climatique de l'Eemien et résulte d'un recouvrement des sources sédimentaires par la végétation.
Les téléconnections à l'échelle de l'hémisphère nord sont évidentes. Des décharges de sable et d'hématite sont contemporaines des poussières piégées dans les glaces groenlandaises. Elles correspondent à des épisodes froids et arides en Sibérie et en Asie Centrale. Les téléconnections sont également visibles aux Terminaisons I et II, lorsque les vents d'ouest apportent plus l'humidité dans la région du Lac Baïkal. Ceci indique probablement de nouvelles circulations atmosphériques liées à une réorganisation des pressions atmosphériques lorsque l'insolation augmente.

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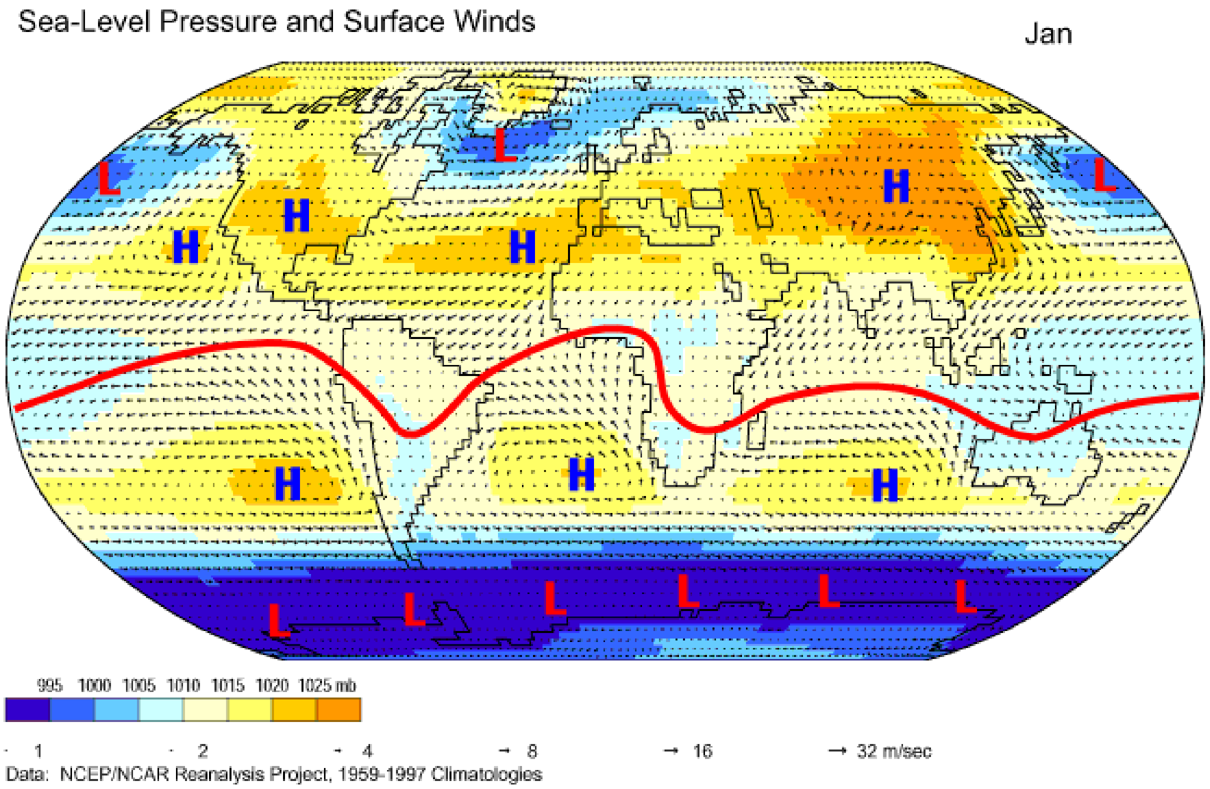
I. Introduction

Understanding the past earth's climate natural variability is extremely important for quantifying the anthropogenic impact on the present and future climates. The climatic history of the last 150 ka is of great interest because it includes two interglacials (Eemian and Holocene), separated by a glacial period (Weichselian). The Eemian, warmer than the present interglacial, could be an analogue scenario for better understanding processes and feedback mechanisms related to future warming. The last 150 ka show stadials and interstadials as a response to changes in orbital parameters. Furthermore higher frequency, sub-Milankovitch cycles, like the Dansgaard-Oeschger Events oscillating at a millennium scale (Dansgaard et al., 1984) are well known. The climate record of the last 150 ka is further punctuated by sudden and short cooling events, the Heinrich events (Heinrich, 1988). These brief events were first recognised in North Atlantic records and have recently been documented in Siberian loess records (Evans et al., 2003). This indicates atmospheric connections at a northern hemispheric scale. The present-day earth's atmospheric links are well described and the distribution of air masses and their connections are relatively well known. Notably, the Siberian High Pressure Cell builds up during winter over Central Siberia while in summer a low pressure cell is established. The resulting wind patterns are well constrained (Fig. I.1). However, this only holds for the present day situation with high resolution instrumental data sets. Past atmospheric circulations are quite difficult to infer. They must be reconstructed from climatic proxies or from climate modelling. During the last 150 ka climatic variability was quite significant, thus it is quite a challenge to reconstruct atmospheric circulation scenarios for this time window.
Late Quaternary high resolution climatic variations are well documented from many marine and lacustrine records. However, it is still a challenge to investigate Late Quaternary climatic changes as recorded in lakes because the interpretation of proxy records for most lacustrine environments is complex as interpretations are less straightforward than in marine environments. One reason is the difficulty in establishing reliable age models for lake sediments and other continental sequences. Furthermore, each continental system has its specific critical points. The biology is often endemic or the biological system is poorly understood due to various local influences which may overprint the regional and global climate signals.

I.1. Lake Baikal: an excellent climatic archive

Many scientific groups have focused on Lake Baikal since the early 1990s. At the time of writing many debates and controversies remain, particularly concerning dating (Prokopenko et al., 2002; Kashiwaya et al., 2002).
As Lake Baikal is located in the middle of the Eurasian Continent it is far from oceanic influences. Hence, Lake Baikal is a key area for understanding paleoclimate variations in an intracontinental region. Studies have clearly established that biogenic productivity, which builds up diatomaceous layers during interglacials, is related to the insolation pattern (e.g. Khursevich et al., 2001) while cold periods are marked by a clayey sedimentation, which according to Karabanov et al. (2004), results from erosion by glaciers of the mountains surrounding Lake Baikal. These diatomite-clay alternations are controlled by the precession and eccentricity cycles (e.g. Goldberg et al., 2000; Kashiwaya et al., 2001). This rather monotonous sedimentation lasts for at least 12 Ma (Kashiwaya et al., 2001) making Lake Baikal an excellent paleoclimatic archive. Today the Siberian High Pressure Cell dominates weather conditions over Lake Baikal during winter time. Moisture is carried by Westerlies to Central Siberia during spring and summer. Changing atmospheric conditions make Lake

Baikal extremely sensitive to climatic changes in the past too. In this area the system response is instantaneous and controlled by changes occurring at a regional or even global level. A proof might be the high frequency events, which Prokopenko et al. (2001a) have related to Heinrich events and Bond cycles, both being well known from North Atlantic sedimentary records (Heinrich, 1988; Bond et al., 1993). This coeval climatic variability may highlight the climatic affinity or teleconnections between Central Siberia and the N. Atlantic.



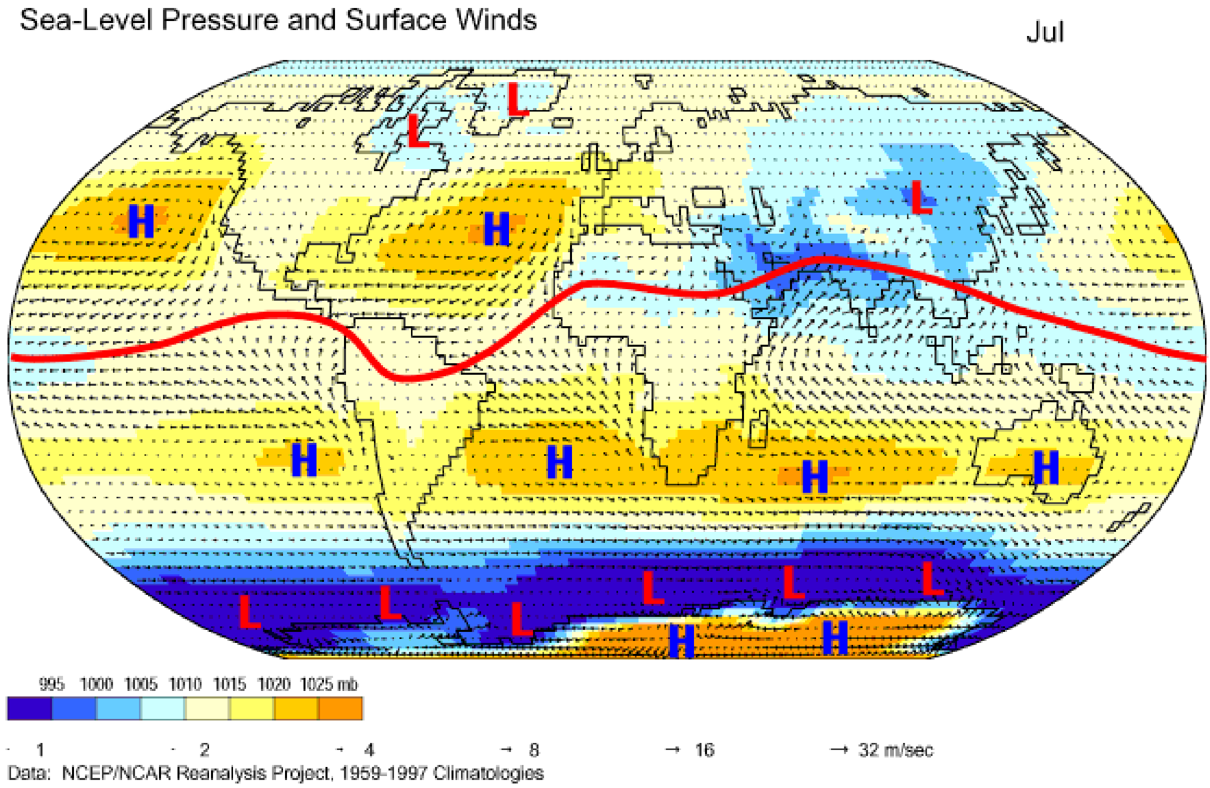


Figure I.1: Mean January (up) and July (down) prevailing surface winds and centres of atmospheric pressure, 1959-1997. The red line on this image represents the intertropical convergence zone (ITCZ). Centres of high and low pressure have also been labelled. (Source of Original Modified Image: Climate Lab Section of the Environmental Change Research Group, Department of Geography, University of Oregon - Global Climate Animations).

I.2. Dating: the crucial point

When correlating paleoclimatic events between near and distant sediments, a major question has to be addressed: do we have appropriate dating tools and how accurate is the dating? Indeed, when correlating particular events, mostly tuning and wiggle-matching techniques to standardised marine oxygen isotope curves were used. This often gives the impression that climatic variations recorded in Lake Baikal and in marine sediments are synchronous. In detailed studies, however, delays are often indicated between marine and intracontinental sedimentary reactions as recorded in Lake Baikal sediments (Peck et al., 1996; Oda et al., 2002) but not yet quantified. Leads and lags can be quantified in sub-recent and older sediments using absolute dating. However, *AMS* ^{14}C dates from Lake Baikal sediments with reasonable error bars are only available for the last 15 ka. In older sediments, the content of organic matter is too low to apply the *AMS* ^{14}C dating method. Dating sediments back to 50 ka, which has been attempted in Lake Baikal (Colman et al., 1996), was therefore a rather speculative approach. Until recently the age models used to date Lake Baikal sediments derived from tuning of lithological variations to the SPECMAP curve as the warm cold cycles are outlined by magnetic susceptibility, and opal or water contents. However, such an age model may be biased by changing preservation of magnetic minerals and opal and, therefore, can not be used for addressing possible leads and lags as responses to regional and global climatic change. This has provoked fundamental debates on the significance of climatic signals as recorded in Lake Baikal sediments (Prokopenko et al., 1999; Grachev, 2000 and Prokopenko, 2000).

I.3. Paleomagnetism: a powerful dating tool

In this thesis, we present new paleomagnetic records which may present a reference curve for Central Siberia and can be added to existing data sets. Nowadays, a worldwide paleomagnetic database is available. The history of the geodynamo can be reconstructed from this paleomagnetic database fusing intensity variations and short reversal events (geomagnetic excursions) documented for the Late Quaternary. This allows comparison and correlation of paleomagnetic records and consequently should lead towards a better dating of sediment sequences for Lake Baikal and similar systems.
Prior to paleomagnetic investigations, magnetic mineralogy has to be determined in order to test the suitability of the sediments. The sediment has to be undisturbed and the carrier of the magnetic signal has to be constant throughout the sedimentation. Moreover, early diagenesis influencing the magnetic mineralogy has to be estimated in order to understand signals, which may bias the primary paleomagnetic information.
In addition, rock magnetic parameters describing the magnetic mineralogy have a broad interest since they inform the redox state of the sediments (e.g. Robinson, 2001), which may reflect changes in the climate regime (e.g. Timothy et al., 2004)…

Once paleomagnetic records are established, correlations have to be performed. However, far-distant correlation can be difficult due to non-dipolar behaviour of the geomagnetic field which last for 20% of the Brunhes Chron (Merrill and McFadden, 1994).
Concerning previous paleomagnetic investigations in Lake Baikal, an 84 ka-long record of high quality has been performed by (Peck et al., 1996). This record is anchored by many *AMS* ^{14}C dating on bulk organic carbon samples in its topmost part. However, older sediments were dated using conventional age model based on paleoclimatic correlations with marine records. This dating procedure was also used by (Oda et al., 2002) in order to date a paleomagnetic excursion, the Iceland Basin event. Surprisingly, dating of this excursion established by Oda et al. (2002) is younger than commonly admitted (Channell et al., 1997). Oda et al. (2002) suggested that a dating bias could exist because of the age model.

I.4. Detrital input: a promising climatic proxy

Once a detailed age model had been established for all the sedimentary sequences (6 in total), the following task of this thesis was to investigate the detrital input using laser-assisted grain size analyses. Indeed, up-to-date climatic studies in Lake Baikal deal with biogenic proxies such as diatoms (e.g. Mackay et al., 1997; Karabanov et al., 2000b), pollen (Demske et al.; Granoszewski et al. 2004) or photosynthetic pigments (Tani et al., 2002). Only a few studies focused on the detrital input. They consisted in clay mineralogy (Solotchina et al., 2002; Fagel e al. 2003), and grain size distribution based on image analyses (Francus, 1998; Francus and Karabanov, 2000). Some studies were focusing on laser-assisted grain size analysis (Ochiai and Kashiwaya, 2003). The originality of our approach was to separate different fraction such as clay and opal, and to performed grain-size analysis on the remaining sediments, i.e. the detrital fraction > 2µm. Compared to a study performed on bulk sediment, this approach allows to be free from the biogenic productivity but also to separate the signal carried by clays from the signal carried by coarser particles. Indeed, these fractions could carry different climatic signals.

I.5. Objectives of the study

When the European project CONTINENT was proposed, the investigators of the project wanted to overcome the dating problem and the group planned to use several dating methods to finally evaluate the most suitable approach. Indeed, when working at very high resolution, dating errors ensuing from conventional orbital tuning can lead to strong misinterpretation. This is where paleomagnetism came into play. The group planned to create a high resolution record for establishing an age model independent of lithological variations. With that purpose in mind, and considering the above open questions, the first goal of the present study was clearly identified. It was my task to establish an age model using paleomagnetism.
For 6 sedimentary sequences from Lake Baikal, a rock magnetic study is carried out in order to estimate the different processes influencing the magnetic mineralogy. Notably, the characterisation of the eventual effect of early diagenesis on the carrier of the magnetic signal could help to test the suitability of sediments for paleomagnetic investigations.
On the same sequences, a paleomagnetic study is carried out in order to reveal the variations of the geomagnetic field recorded in the sediment. Curves of paleomagnetic inclination and declination are established in order to trace eventual relics of paleomagnetic excursions, which could be anchor points for correlation. Curves of relative paleointensity are also established and correlated with dated relative paleointensity curves of reference in order to establish an age model for the different sedimentary sequences. A new reference curve of

paleomagnetism for central Asia is established by compiling the different relative paleointensity curves.
The second goal of the study is to perform, on the best sedimentary sequence, a laser assisted grain size analysis on the detrital fraction in order to reveal the dynamics of the detrital input in Lake Baikal. Effort is put on the recognition of the processes responsible for the detrital input, their relative influence under different climatic regimes and, if any, recognition of North Hemispheric teleconnections.

The thesis consists of 6 other chapters. Chapter II, “Materials”, briefly presents where the material of this study has been retrieved from and illustrates the lithology of the sedimentary sequences. Chapter III, “Background and Methods” deals with the type of measurement we performed and their implications. This chapter is documented with results for the present work. Chapter IV, “Detrital input and early diagenesis in sediments from Lake Baikal revealed by rock magnetism” is a publication submitted to Global and Planetary Change and mainly deals with the different factors (detrital input, early diagenesis) influencing the magnetic mineralogy. Chapter V, “High-resolution magnetostratigraphy of late Quaternary sediments from Lake Baikal, Siberia: timing of intracontinental paleoclimatic responses”, is another publication submitted to Global and Planetary Change and mainly deals with the setup for an age model using paleomagnetism and its implications. Chapter VI, “Detrital input traced by grain size distribution and rock magnetism in Late Quaternary sediments from Lake Baikal”, deals with detrital input estimated by laser-assisted grain size analysis and rock magnetism. Chapter VII is the conclusion and outlook on the overall work presented in this study.

II. Material

II.1. Coring sites

Sediment cores have been retrieved from different underwater highs separated from the shore by deep basins. These sites have been chosen in order to obtain hemipelagic sediments and to avoid turbidites (Fig. II.1). Two types of coring technique were used: piston coring, and pilot coring (Meischner and Rumohr, 1974). The piston coring allows us to retrieve cores of about 10 m length; the disadvantage of this method is that the topmost sediments are missing. To complete the sedimentary sequence, pilot coring was performed, which allows to obtain up to 2 m long cores with preserved sediment tops. With a composite core we can complete the sedimentary column. More details about the coring location and core recovery are described in the sections, "Material and Methods", Chapters IV and V. This thesis includes results obtained from analyses of six sediment cores, three from Academician Ridge, one from Continent Ridge, Posolsky Bank and Vydrino Shoulder (Fig. II.1).

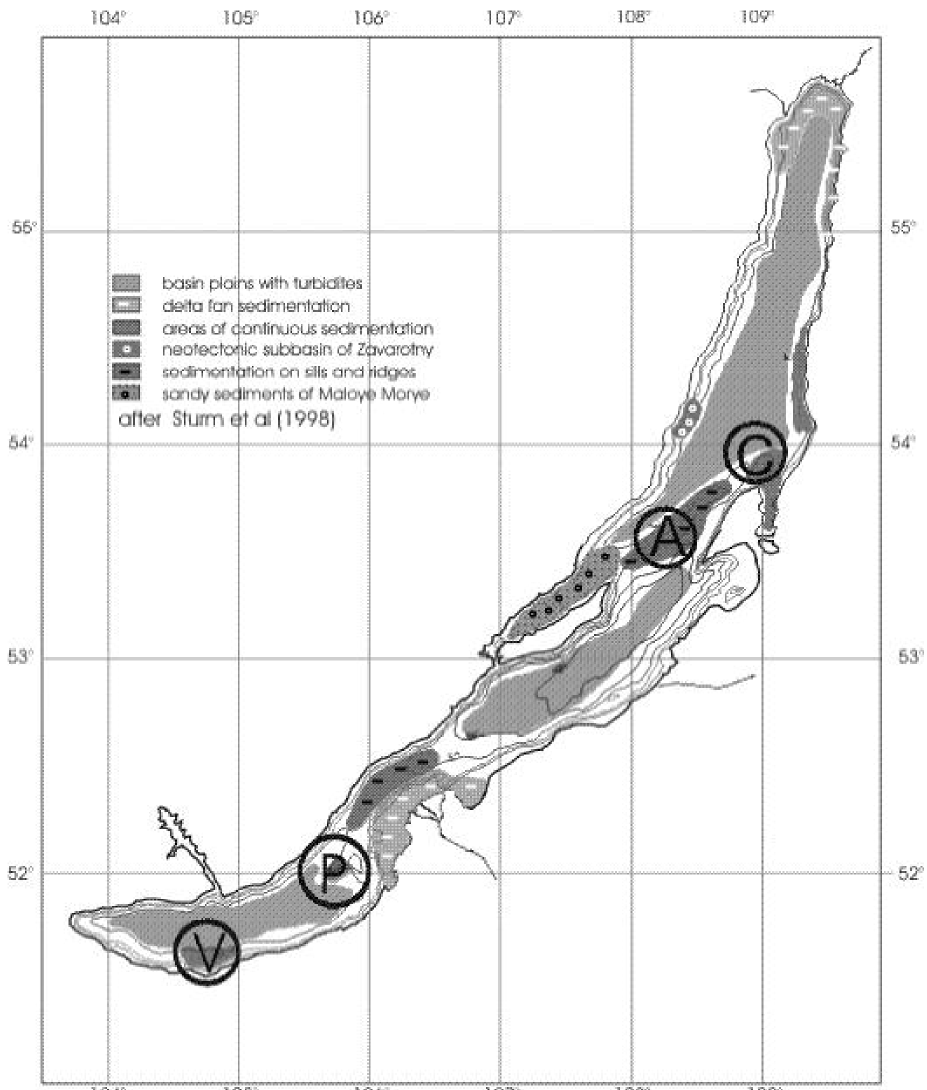


Figure II.1: Sedimentary environments reconstructed from analyses of short cores. A: Academician Ridge, cores VER 98-1-1, VER 98-1-3 and VER 98-1-14; C: Continent Ridge, core CON O1-603-2; P: Posolsky Bank, core CON 01-604-2; V: Vydrino Shoulder, core CON 01-605-3 (after Sturm et al., 1998)

II.2. Quality of the cores

To estimate the quality of the retrieved cores, radiographs have been acquired on sediments slabs. Wherever structures were not visible on radiographs, the anisotropy of magnetic susceptibility (AMS) has been used for estimating the orientation of the sedimentary horizons Indeed, magnetic anisotropy generally reflects the orientation of the magnetic grains and, therefore, the sedimentary fabric (see Chapter III, “Background and Methods”). Good quality coring is emphasized by sedimentary horizons perpendicular to the liner. In Figure II.2A, the AMS ellipsoid is quite well oriented with a compaction axis (κ_{min}) close to vertical for the topmost 6 meters. Below this depth, the AMS ellipsoid is reoriented and develops a prolate shape (Fig. II.2B). This elongated ellipsoid parallel to the liner is due to suction of sediments during coring. The sediments below 6 meters for this core have not been subject to further investigations. The deviation of κ_{min} from the vertical (i.e. from the centre of the stereogram) in Figure II.2C shows that the pilot cores CON 01-605-3a and CON 01-604-2a have been cored slightly obliquely. Core tilting is restricted to some pilot cores as usually the quality of the recovered sediments is good (see Chapter IV “rock magnetism”).

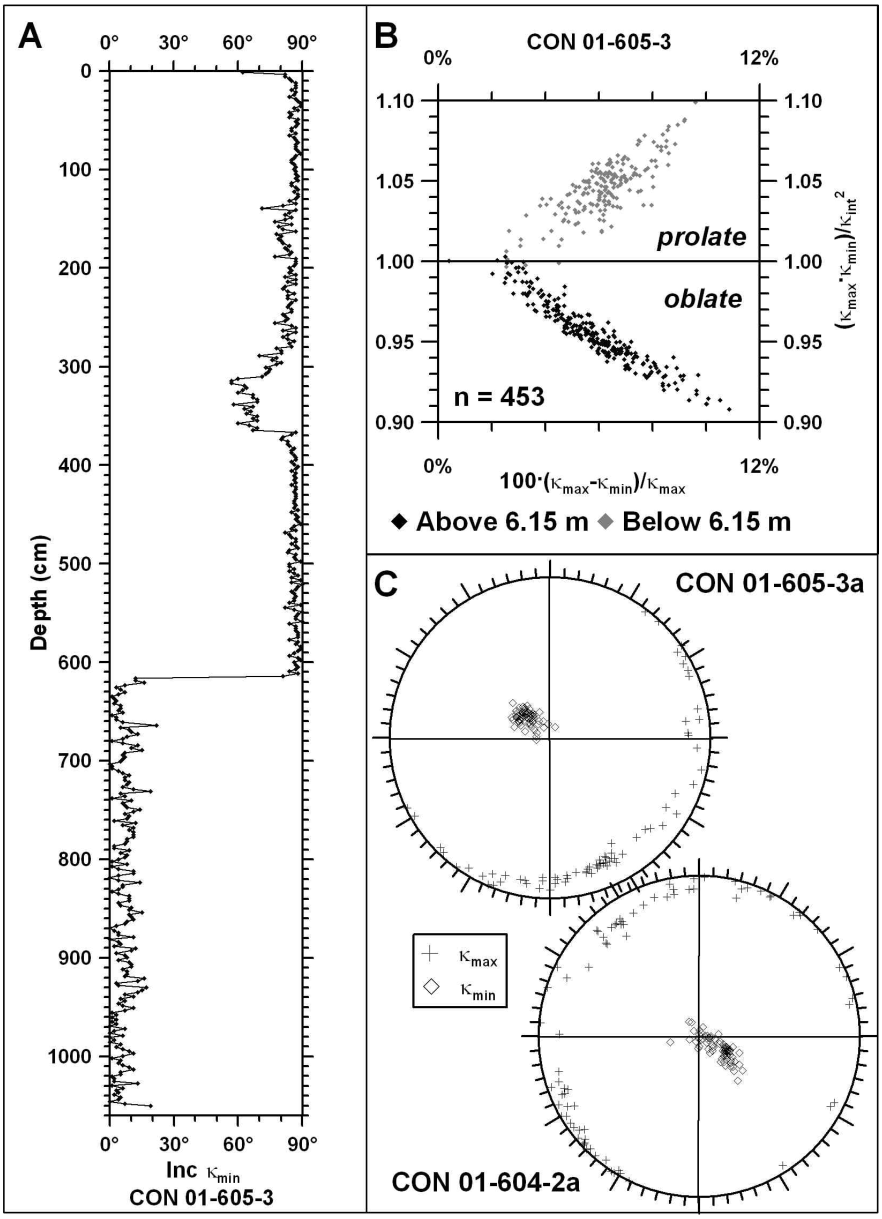


Figure II.2: A. Downcore variations of the inclination of the short axis (k_{min}) of the AMS ellipsoid for piston core CON 01-605-3. B Diagram showing the shape of the ellipsoid versus the degree of anisotropy for the same piston core. Above 6 m, the short axis of AMS (k_{min}) is vertical and the AMS ellipsoid is oblate, which is typical for compacted sediments. Below 6 m, the ellipsoid is elongated (prolate shape) parallel to the core liner, reflecting suction of the sediments during coring. C. Stereographic projection of the short (k_{min}) and long (k_{max}) axes of AMS ellipsoid for pilot cores CON 01-605-3a and CON 01-604-2a. The slight deviation of k_{min} from the centre of the stereogram results from an oblique coring.

II.3. Sediment composition

Sediments have been described in details on 2 of the piston cores retrieved from the Academician Ridge in 1998 (Fig. II.3) and on 3 piston cores, CON 01-605-3, CON 01-604-2 and CON 01-603-2 retrieved in 2001 from Vydrino Shoulder, Posolsky Bank and Continent Ridge, respectively (Fig. II.4). At the Academician Ridge, sediment cores consist of olive-coloured diatomite and greyish clay-rich layers. This alternation also exists for the other cores. Small differences exist from one site to another: Vydrino core (CON 01-605-3) shows more detrital mud, Posolsky core (CON 01-604-2) shows biogenic diatom-bearing intervals, mixed biogenic/detrital layers and pure clays, while Continent core (CON 01-603-2) ranges from pure diatom-rich layers to pure detrital clay-rich layers, with transitional biogenic and detrital lithologies. The detailed description shows presence of sand layers in clay-rich sediments at certain sites and the occurrence of more silty parts. This illustrates that grain size analyses are valuable proxies recording climatic variability. Moreover, the description shows the occurrence of scarcely distributed but existing bioturbations, as well as faults and slumps which could potentially disturb the sedimentary fabric. To summarise, whatever the coring location, sedimentary sequences roughly display an alternation of diatomaceous layers with clay-rich layers. This alternation marks the climatic variations as recorded in Lake Baikal lithology with high diatom productivity during warm periods and clay-rich sedimentation during cold periods. In the following chapters, we used simplified sedimentary descriptions showing only the diatomaceous/clay-rich alternation.

This lithological alternation is often used for roughly dating Lake Baikal sediments. Indeed, many authors correlate the sedimentary alternation of Lake Baikal with the marine isotope stages (MIS) defined in marine environments (see e.g. Prokopenko et al., 2001b) and references therein). Diatomaceous layers are therefore considered as equivalent to marine interglacials and interstadials while clay-rich layers are considered as equivalent to marine glacials and stadials. This approach can be a first step establishing an age model in the present study. As a first approximation, the different thicknesses of deposits from on site to another show various sedimentation rates for these underwater highs.

II.4. Sampling protocol

Sedimentary sequences from 6 sites have been continuously sub-sampled for rock- and paleo-magnetism. After splitting of the sediment cores in two halves, plastic boxes were pushed into the sediment in the centre of the liner. This has been done throughout the length of the core, and given the diameter of a box (2.2 cm), the step of sampling is of about 2.5 cm. The plastic box and sediment fill have been enclosed in air-free plastic bags; moisture being maintained with wet sponges. The bags were stored in a cool room at 4°C, with measurements being made a few days later.

For the grain-size analysis, we used the freeze dried samples from Core CON 01-603-2 with a sampling resolution of 2 cm.

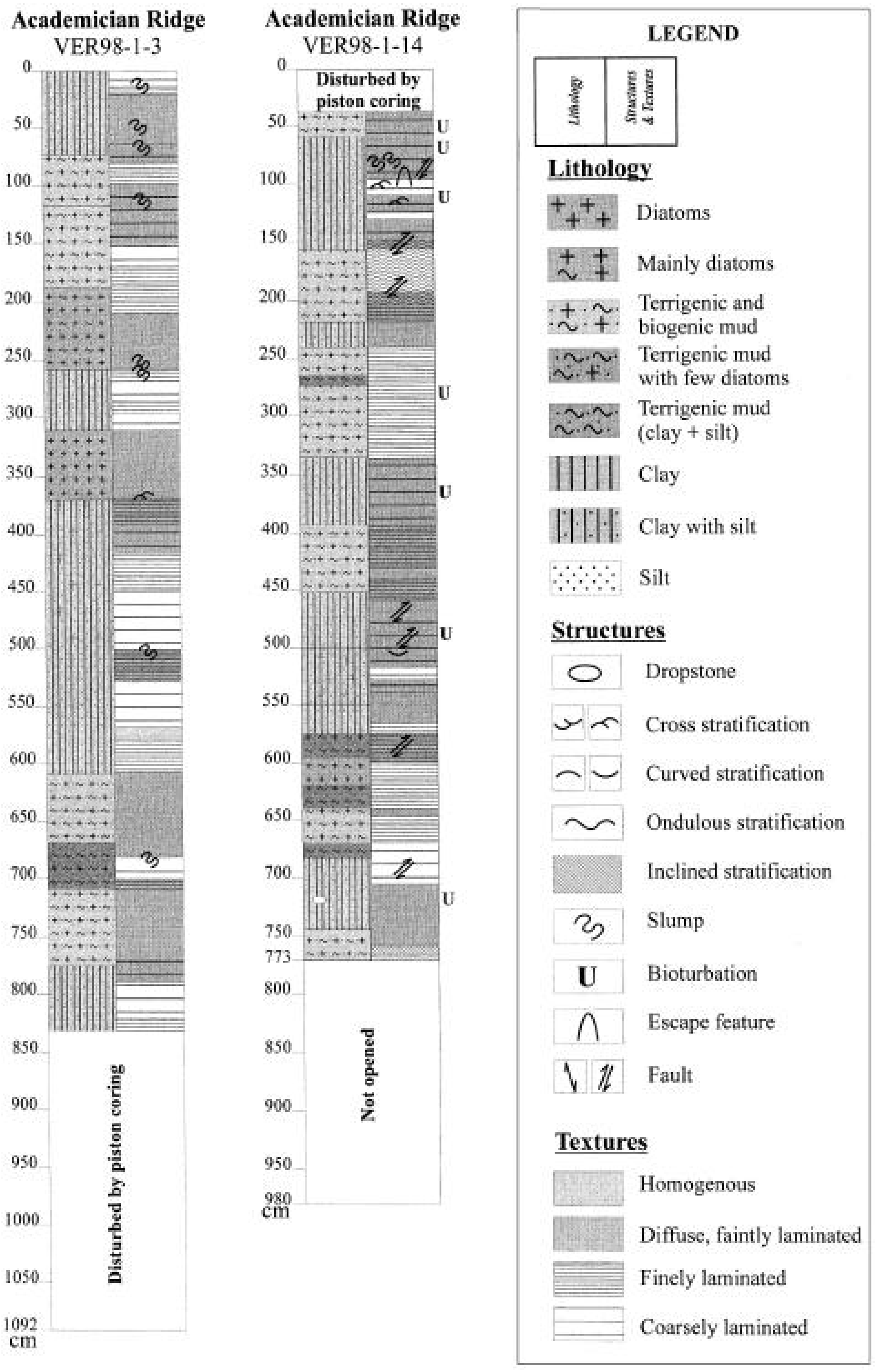


Figure II.3: Lithology of sediments cores VER 98-1-3 and VER 98-1-14 Academician Ridge (after Fagel et al., 2003).

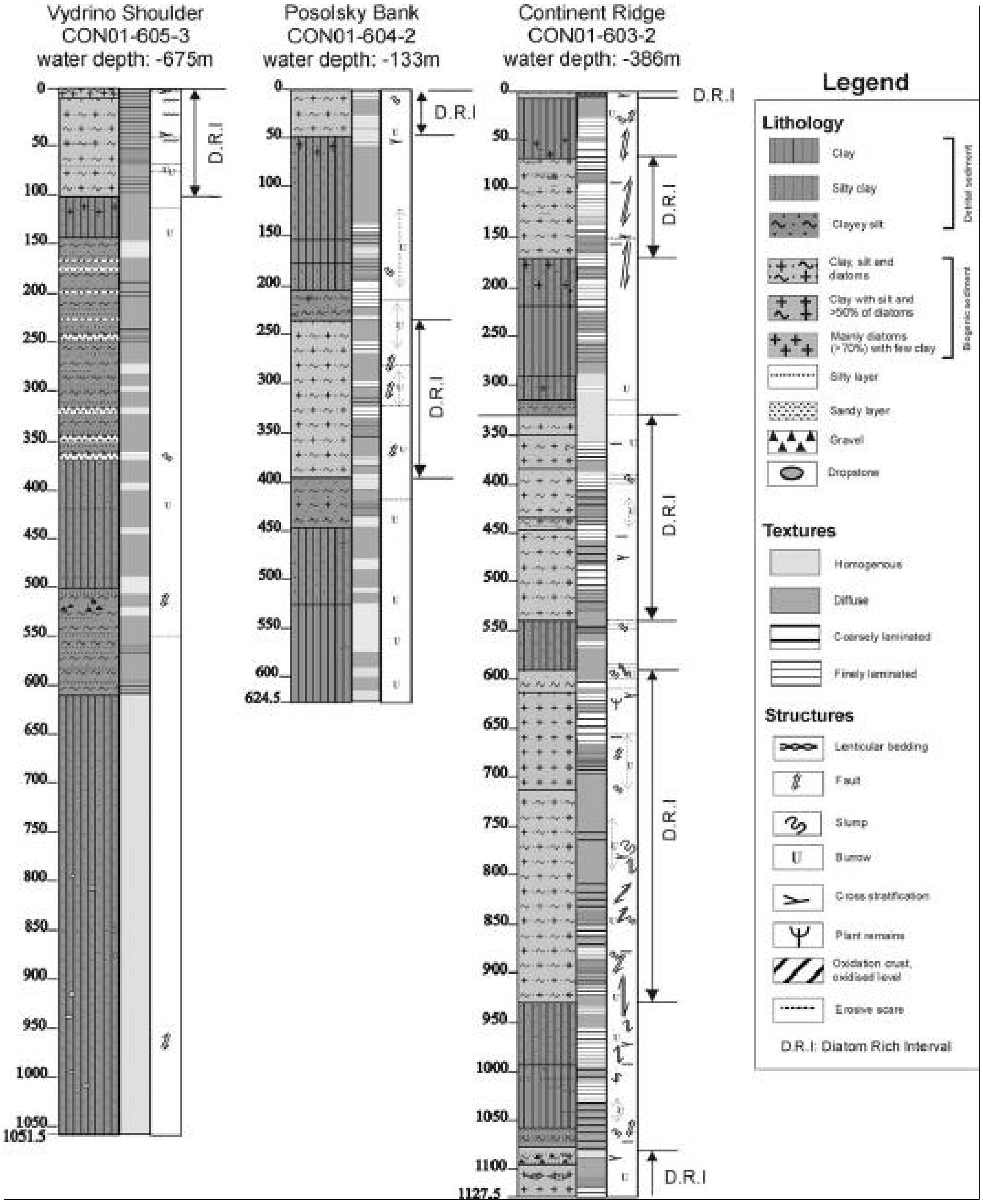


Figure II.4: Lithologs of the three piston cores from Vydrino Shoulder (CON 01-605-3), Posolsky Bank (CON 01-604-2) and Continent Ridge (CON 01-603-2), (after Charlet et al., submitted).

III. Background and methods

This chapter is divided into 3 sections. The first two sections provide background information on rock magnetism and paleomagnetism methods used in the following chapters. The third section is an introduction to the laser-assisted grain size analysis and sample preparation.

III.1. Rock magnetism

A rock magnetic study consists of a set of magnetic measurements, which help to constrain the magnetic mineralogy. For a better characterisation of the magnetic mineralogy, the study can be complemented by other methods such as X-ray diffraction (XRD) and Transmission Electronic Microscopy (TEM) and compared with geochemical proxies, as performed in the present work (see chapter IV). This approach has two goals: (i) there is an environmental interest since magnetic mineral assemblages are influenced by sedimentary sources, transport processes and diagenesis; (ii) a characterisation of the magnetic mineralogy is necessary for reinforcing a paleomagnetic study since good quality paleomagnetic records require a carrier of the magnetic signal with constant characteristics.

III.1.1. Magnetic properties of sediments

In general there are three different types of magnetic behaviour: (i) diamagnetism, (ii) paramagnetism and (iii) ferromagnetism (Table III.1).

(i) *Diamagnetism* is a characteristic of minerals (or substances) consisting of atoms with paired electrons. When an external magnetic field is applied the electron orbits of the atoms in the crystal lattice become aligned and opposed to the applied field. A weak and negative magnetic susceptibility is typical for this type of magnetic behaviour (Table III.1). The magnetic moment of a diamagnet is lost when the external magnetic field is removed.

(ii) In the case of *paramagnetism*, the crystal lattice has some atoms with unpaired electrons. In an external magnetic field the atoms acquire a magnetic moment parallel to the applied field, i.e. magnetic susceptibility > 0. However, due to large distances between the atoms with unpaired electrons, the magnetic moment is lost when the magnetic field is removed.

Type of Magnetism	General behaviour	example of substances
diamagnetic	Weak negative susceptibility	water, organic matter, plastics, quartz, feldspars, calcium carbonate
paramagnetic	Weak positive susceptibility	many Fe-containing minerals and salts e.g. biotite, olivine, ferrous sulphates
ferromagnetic	Strong positive susceptibility	elemental iron, nickel, chromium
ferrimagnetic	Strong positive susceptibility	some iron oxides and sulphides e.g. magnetite, maghaemite, pyrrhotite,greigite
canted antiferromagnetic	Moderate positve susceptibility	some iron oxides, e.g. hematite, goethite

Table III.1: Type and general behaviour of magnetic minerals and elements (modified after Dearing, 1998).

(iii) The *ferromagnetism (s.l.)* is characteristic for regularly spaced atoms with unpaired electrons. Due to interactions a magnetic moment is retained even without an external magnetic field. Ferromagnetic behaviour is a characteristic of transitional metals (Table III.1).

Depending on the organisation of the atoms in the lattices, the ferromagnetic behaviour can be subdivided in *ferromagnetism s.s., antiferromagnetism, ferrimagnetism, and canted antiferromagnetism* (Fig. III.1a-d). The latter three magnetic responses are characteristic of Fe-oxides, which are the most common magnetic carriers in natural rocks.

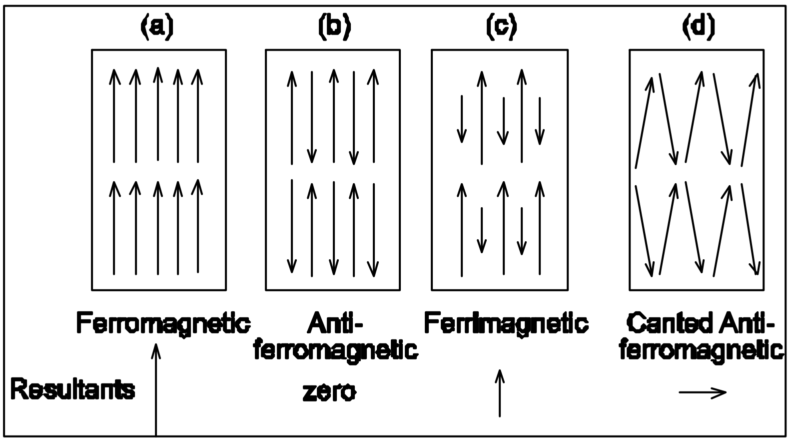


Figure III.1: Schematic representation of the distribution of magnetic moments in crystal lattices, showing the resultant spontaneous magnetisation (after McElhinny, 1973).

The high energy of the magnetic free poles on the surface of magnetised particles can be reduced when the grain is subdivided into smaller regions (domains) with individual spontaneous magnetisations. Then, magnetic moments of the single domains compensate each other (Fig. III.2). Particles with one domain are called single domain (SD) and particles with more than one domain are called multi-domain (MD). The existence and the behaviour of magnetic domains depend on the size and shape of the magnetic grains. In the case of a single domain, the magnetic spin moments are parallel to each other and oriented along an easy direction of magnetisation. When an external field is applied, these magnetic moments rotate in order to become aligned to the external field. In multi-domain particles the volume of the domains with moments parallel to the external field first increases (Fig. III.2b). When the magnetic field increases further the magnetic moments start to rotate as for SD.

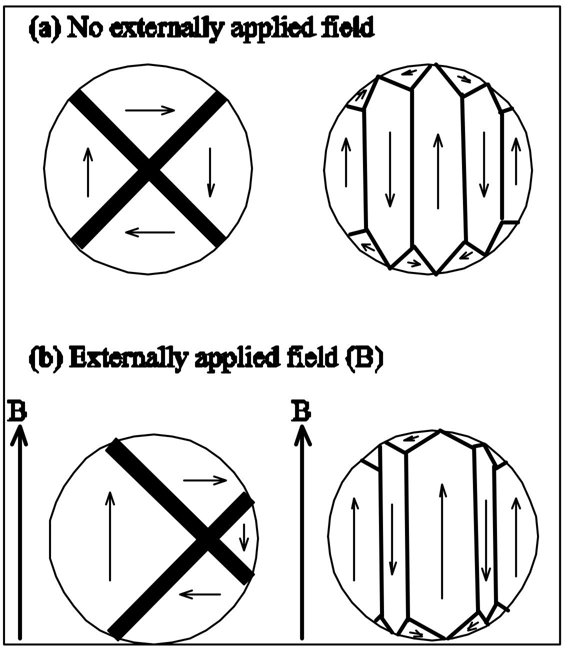


Figure III.2: Schematic diagram of possible domain configurations for a spherical multidomain grain (after Dankers, 1978).

A peculiar behaviour is *superparamagnetism*. This behaviour is characteristic of very small ferromagnetic particles (below 300 µm for magnetite), which do not retain a remanence but behave like paramagnets with a high magnetic susceptibility.

III.1.2. Magnetic hysteresis loops

The relationship between the applied external magnetic field and the induced magnetisation differs according to the magnetic behaviour (dia-, para-, or ferromagnetic).
The relationship is linear for dia- and paramagnetic responses. However, the relation is more complicated for ferromagnetic behaviour. Magnetisation increases non-linearly with increasing field strength (Fig. III.3), until saturation magnetisation (Ms) is reached. The magnetic field necessary for saturation is called a saturating field (Bs). Decreasing the magnetic field down to 0 does not reduce the remanent magnetisation completely i.e. a part of the saturation remanent magnetisation (Mrs) is retained by the sample. The field needed to demagnetise the sample is called coercive force (Bc). The same parameters can be determined in a magnetic field with an opposite (negative) direction. The loop resulting from a full cycle is called a hysteresis loop (Fig. III.3). The measurement of the hysteresis loop is complemented by a back-field measurement, which allows for a determination of the coercive force of remanence (Bcr). Bcr is the field strength necessary to completely demagnetise the remanent magnetisation. According to the above section, the shape of the loop (and all its characteristics Bc, Bcr, Mrs, Ms) depends on the number of domains and hence on the magnetic grain size.

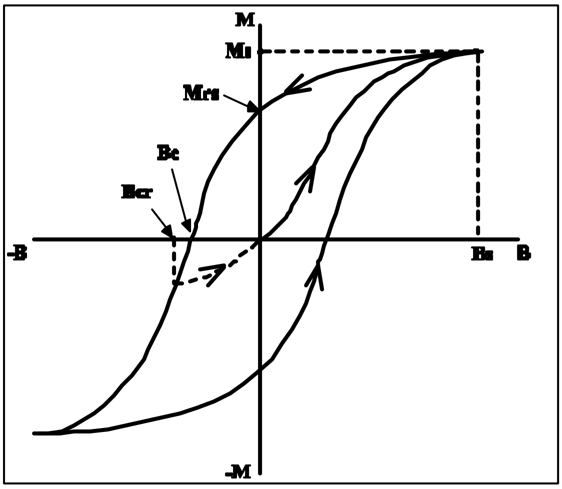


Figure III.3: Hysteresis loop of a ferromagnetic (s.l.) response with characteristic parameters (simplified after Smith, 1998).

III.1.3. Acquisition of magnetisations

For the rock magnetic study, the induced or remanent magnetisations described below have been systematically measured. The equipment used for acquisitions and measurements of rock magnetic parameters is described in details in sections "Material and methods" of chapters IV and V.

Here we summarise briefly:
-The low field magnetic susceptibility (κ_{LF}), which is non destructive and reversible, consists of measuring the induced magnetisation under a weak field. This measurement includes the contribution of dia-, para- and ferromagnetic particles.

-The anhysteretic remanent magnetisation (ARM) is an artificial magnetisation gained in an alternating magnetic field (with a maximum amplitude of 100 mT) superimposed on a steady field of 50 µT, which is comparable to the strength of the geomagnetic field. The ARM is afterward subject to stepwise demagnetisation.
-Isothermal remanent magnetisation (IRM) is imprinted in a steady magnetic field. The field must be strong (1 to 2 T) when saturation of the sample is desired. In this case, the induced remanent magnetisation is called saturation isothermal remanent magnetisation (SIRM). Another isothermal remanent magnetisation is acquired in the opposite direction in order to remagnetise the soft magnetic minerals present in the sample, the field used in this case is generally 0.3 T.

III.1.4. Rock magnetic parameters

Concentration-dependant parameters
All the magnetisations described above provide a rough estimation of concentration in ferromagnetic particles. Nevertheless, SIRM is the only parameter independent from the magnetic grain size. For ARM and κ_{LF} parameters are influenced by the magnetic response of each particle, which differs according to their magnetic grain size.

Curie temperature:
Each ferromagnetic mineral is characterised by a specific Curie temperature (Tc) i.e. the temperature at which the magnetic ordering is lost (Table III.2). Tc is usually determined from the temperature-dependent magnetic susceptibility or the induced magnetisation.

Mineral	Formula	T_c
Magnetite	Fe_3O_4	578°C
Hematite	αFe_2O_3	675°C
Maghemite	γFe_2O_3	645°C
Goethite	$\alpha FeOOH$	120°C
Greigite	Fe_3S_4	350°C

Table III.2: Curie temperatures (T_c) for some natural magnetic minerals (simplified after Dekkers (1997) and the reference therein)

Coercivity-dependant parameters:
The coercivity spectrum of a natural sample (typically a mixture of different minerals and grain sizes) depends on the type and size of its magnetic particles. Low coercivity minerals like magnetite reach saturation magnetisation (SIRM) in a weak external field (less than 300 ka, Fig. III.4A). Antiferromagnetic minerals like hematite or goethite require a much stronger field (above 2 T). Hence the ratio:

$$\text{S-ratio} = \frac{1}{2}\left(1 - \frac{\text{IRM}_{-0.3\text{T}}}{\text{SIRM}_{2\text{T}}}\right)$$

S-ratio can be used as a measure of the relative abundance of high coercivity minerals and ranges from 1 for pure magnetite toward 0 for pure hematite (Bloemendal et al., 1992).

The high isothermal remanent magnetisation (HIRM):

$\text{HIRM} = (\text{SIRM}-\text{S}_{-0.3\text{T}})/2$

HIRM estimates absolute concentrations of high coercivity minerals (Maher & Dennis, 2001).

Additionally, a stepwise acquisition of IRM supplies more information about the relative contribution of high and low coercivity phases in a single sample (Fig. III.4B).

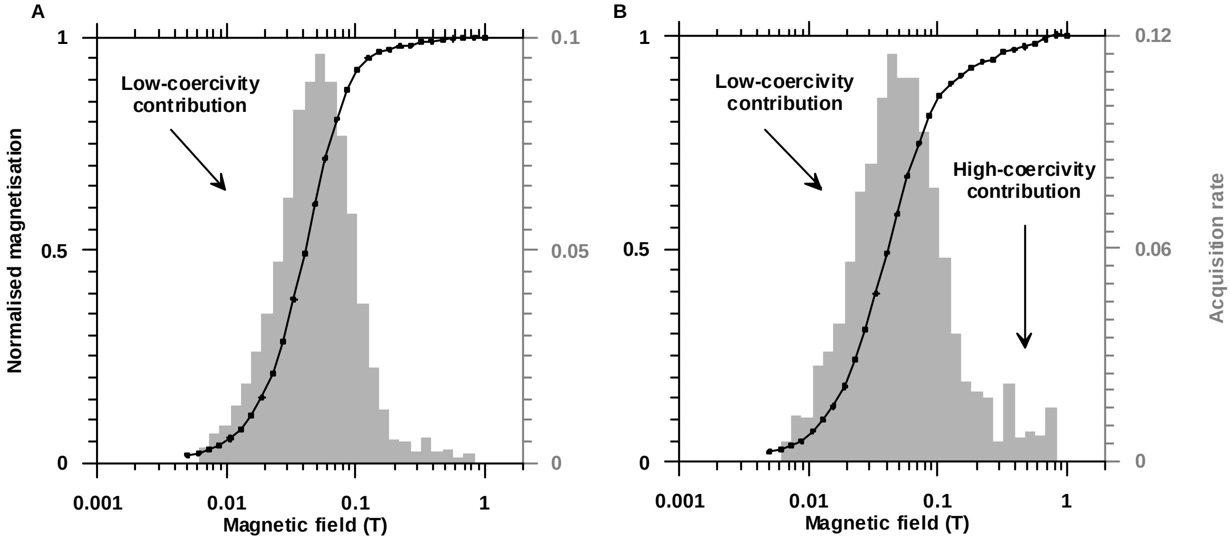


Figure III.4: Stepwise acquisition of IRM (A) for magnetite only (sample from depth 110.1 cm of core CON 01-603-2); (B) for a mixed mineralogy (sample from depth 61.5 cm of core CON 01-603-2).

Magnetic grain size sensitive parameters:

Since the shape and characteristic parameters of a hysteresis loop are linked to the domain status of magnetite, they can be used to approximately estimate the grain-size distribution of this most common magnetic carrier. On a plot of Mrs/Ms versus Bcr/Bc (Day diagram), fixed boundaries separate areas for both the single domain (SD), pseudo-single domain (PSD) and multidomain (MD) status (Day et al., 1977; Dunlop, 2002a). Data for mixtures of various size are found to scatter across specific paths on the Day plot depending on the relative concentrations of the single phases (Dunlop, 2002b). Similarly in Lake Baikal, the scattering of data in the magnetite dominated sediment intervals is assumed to result from varying grain size mixtures (Fig. III.5A). Since the Day diagram does not image the depth- or the time-related variations, for stratigraphic interpretations, the Mrs/Ms and Bcr/Bc parameters must be plotted versus depth (Fig. III.5B). In this present work, the down-core variations of the Mrs/Ms and Bcr/Bc parameters show that besides early diagenesis, which locally changes the magnetic grain size due to bacterial greigite and magnetite mineralisation, magnetite dissolution (see Chapter IV "Rock magnetism") and the coarsening of the magnetic grains at transition MIS 2 to the Holocene could be interpreted as a detrital artefact. Indeed, it would corroborate the grain size analysis, which shows a decrease of clay and an increase of silts in this interval (see Chapter VI).

Besides the Day plot, there are other magnetic grain size estimators. Simple rock magnetic ratios like ARM/SIRM or ARM/κ_{LF} can be used for a rough estimation of the relative magnetic grain size as well. Since ARM affects preferably finer grains, ARM/SIRM decreases with the coarsening of the magnetic carriers (Maher, 1988). The use of rock magnetic ratios, however, is limited to the case of magnetite as a unique carrier of the magnetic signal. Indeed in case of a high coercivity phase contributing to the signal, ARM/SIRM decreases solely due to the inefficient ARM acquisition in high coercivity minerals.

The frequency dependant magnetic susceptibility (κ_{FD}) estimates the abundance of the smallest superparamagnetic particles in a magnetic assemblage (Mullins, 1973). This parameter is mostly used in soil science.

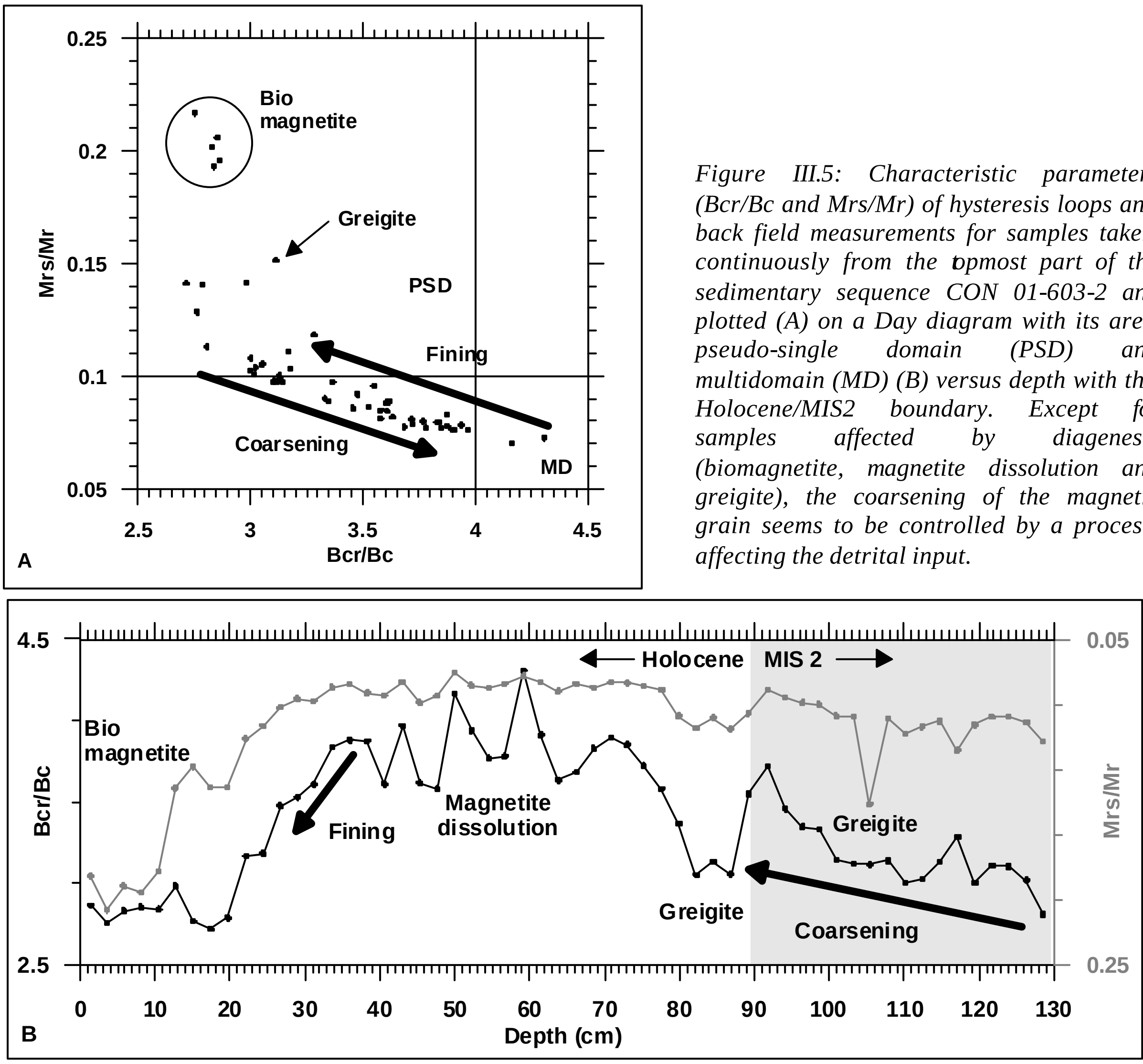


Figure III.5: Characteristic parameters (Bcr/Bc and Mrs/Mr) of hysteresis loops and back field measurements for samples taken continuously from the topmost part of the sedimentary sequence CON 01-603-2 and plotted (A) on a Day diagram with its area pseudo-single domain (PSD) and multidomain (MD) (B) versus depth with the Holocene/MIS2 boundary. Except for samples affected by diagenesis (biomagnetite, magnetite dissolution and greigite), the coarsening of the magnetic grain seems to be controlled by a process affecting the detrital input.

Anisotropy of magnetic susceptibility (AMS):

The anisotropy of magnetic susceptibility (AMS) is the directional variability of κ_{LF}. This can be best represented by the principal axes (κ_{max}, κ_{int} and κ_{min}) of an ellipsoid. AMS is controlled by the shape and crystallinity of the magnetic grains (Daly, 1970). For magnetite, the elongation of the grain is considered to be dominant whereas Fe-sulphides like greigite or pyrrhotite have a strong crystalline anisotropy. AMS reflects the fabric of magnetic grains and, per extrapolation, the sedimentary fabric (Rees, 1965). In paleomagnetic studies, AMS detects sedimentary disturbances induced by coring e.g. suction and oblique coring, and/or of natural origin e.g. slumping (see Fig. II.2).

III.1.5. Different influences on the magnetic components of the sediments

The magnetic assemblage in sediments is typically composed of particles originating from detrital input and/or biogenic production as well as particles of post-depositional (authigenic) origin (diagenesis). The conditions of sedimentation and productivity that influence the rock magnetic assemblages are plotted in Figure III.6. In Lake Baikal, the chosen sites present low sedimentation rates and the biogenic productivity is high during interglacials. Comparing Figures III.6A and III.6B, we observe that the magnetic mineral assemblages we want to characterise are potentially influenced by different means of sedimentary transport as well as by bacterial activity. The chemical conditions relevant for the stability of magnetic minerals are also shown in Figure III.7. In our study, the conditions for magnetic particle stability as described in Figure III.7A are characteristics for the catchment area whereas Figure III.7B illustrates the conditions of stability that could be relevant at sites of deposition or even conditions occurring after deposition, i.e. during burial. In Lake Baikal, we could separate between the detrital and the diagenetic information (see chapter IV) using different controls such as the redox state in the sediment, the rate of burial and the bacterial mediation.

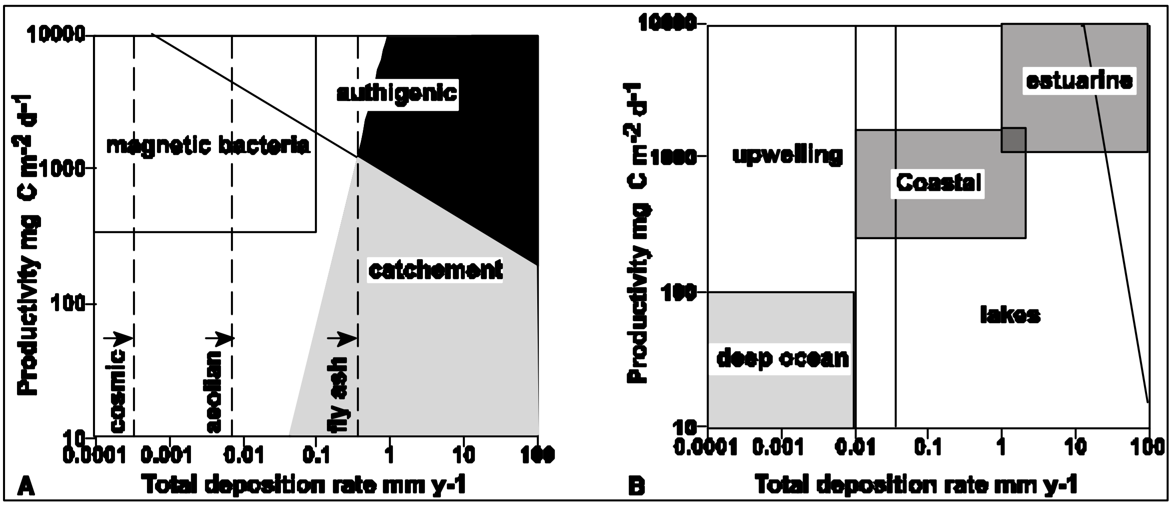


Figure III.6: Magnetic minerals in aquatic environment from (Hilton, 1987); (A) regions influenced by different magnetic mineral sources; (B) combinations of productivity and accumulation rate for different aquatic environments.

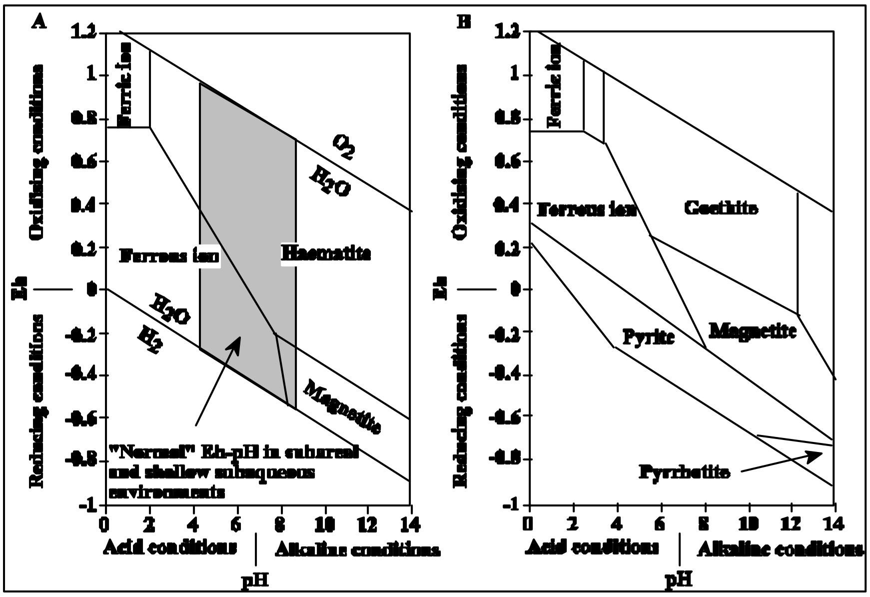


Figure III.7: Eh-pH diagrams showing stability fields of some iron oxides, hydroxides and sulphides; (A) in subaerial and shallow subaqueous environments; (B) in intermediate and deep-marine conditions (from Tarling and Hrouda, 1993).

As we have seen in this section and, as will be shown in the following chapters, rock magnetism could have climatic and diagenetic implications. Rock magnetic parameters are quite often used, as in Lake Baikal, as correlation tool between sediment cores (e.g. Sakai et al., 2000; Kravchinsky et al., 2003). We will show in this thesis that the effects of early diagenesis and hiatuses of sedimentation can lead to miscorrelations. In addition, rock magnetism is crucial for paleomagnetic studies since it is important to know about the quality of the sediment. A constant carrier of the magnetic signal is necessary to establish good paleomagnetic records (King et al., 1983; Tauxe, 1993). In this respect, our rock magnetic study shows that at least for most of the intervals, a low coercive magnetite carries the paleomagnetic signal in Lake Baikal sediments. Nevertheless, we will show in this thesis that confined to certain intervals, dissolution of magnetite and greigite mineralisation during diagenesis can alter the paleomagnetic signal. However, using rock magnetism *s.s.*, these features can be identified and where necessary, they have to be ruled out (Fig. III.8).

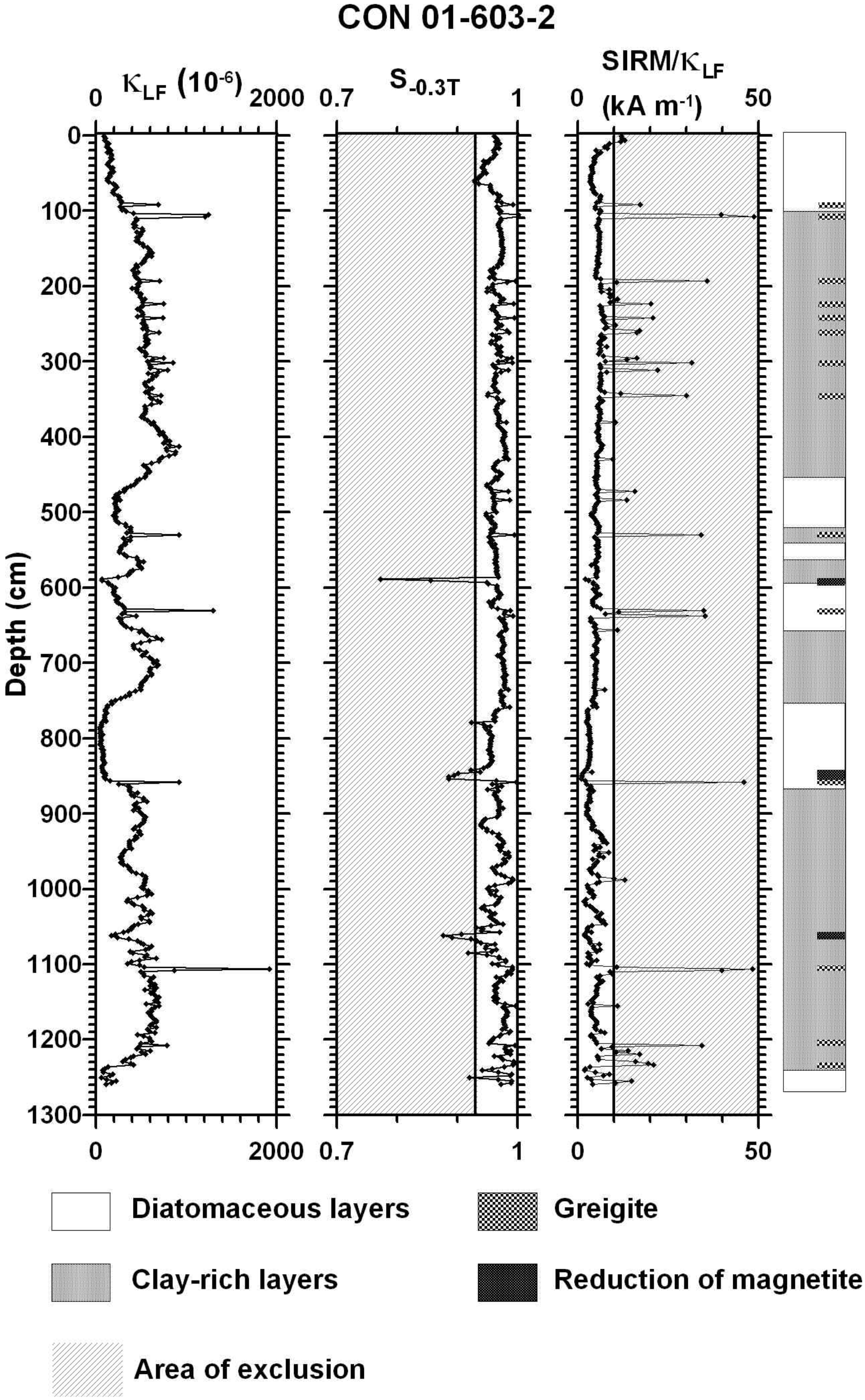


Figure III.8: Down core variation of magnetic susceptibility, S-ratio and SIRM/k_{LF} and simplified lithological description for sedimentary sequence CON 01-603-2. Samples presenting S-ratio < 0.93 (i.e. magnetite dissolution) and SIRM/k_{LF} > 10 kA m-1 (i.e. greigite mineralisation) have been excluded from paleomagnetic investigations.

III.2. Paleomagnetism

III.2.1. How to reveal the paleomagnetic signal?

The *natural remanent magnetisation* (NRM) carried by the samples is measured prior to any destructive investigation (i.e. artificial magnetisation acquisition) that could be done for magnetic mineralogy identification (see section on rock magnetism). This magnetisation contains the paleomagnetic information, i.e. the earth's magnetic behaviour at the time of the particle deposition. The NRM is a sum of vectors for all remanent magnetisations carried by the sediment.
The *detrital remanent magnetisation* (DRM), as a first approximation, is the magnetisation that sediments acquire during their deposition (Fig. III.4). This magnetisation carries the paleomagnetic information. The detrital particles are able to carry a remanence (ferromagnetics) record in the direction and intensity of the ambient (geo)-magnetic field at the time of deposition. The DRM can be decomposed in depositional detrital remanent magnetisation, which is acquired during deposition and in post-depositional detrital remanent magnetisation. This is acquired after deposition and before consolidation. This implies a lock-in depth for the paleomagnetic information of ~ 20 cm (e.g. Butler, 1992; Carcaillet et al., 2004), which depends on the magnetic grain size more than on the sedimentation rates. Lock-in time can last as long as 2 ky (Nourgaliev, pers. communication 2003).
The primer DRM can be overprinted by a *chemical remanent magnetisation* (CRM) acquired during diagenesis when new magnetic mineral form in the sediment. The *Viscous Remanent Magnetisation* (VRM) is acquired without any mineralogical change, just by "sitting" in the magnetic field of the sediment. Demagnetising the samples i.e. a cleaning of the NRM from the CRM and VRM, is attempted in order to reveal the DRM that recorded the ambient magnetic field during the sediment deposit (Fig. III.9). When the DRM is reached, the stable direction is measured. This stable magnetisation is called characteristic remanent magnetisation (ChRM).

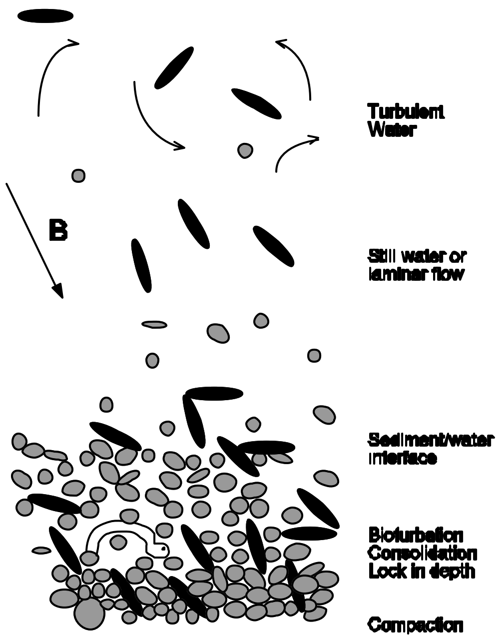


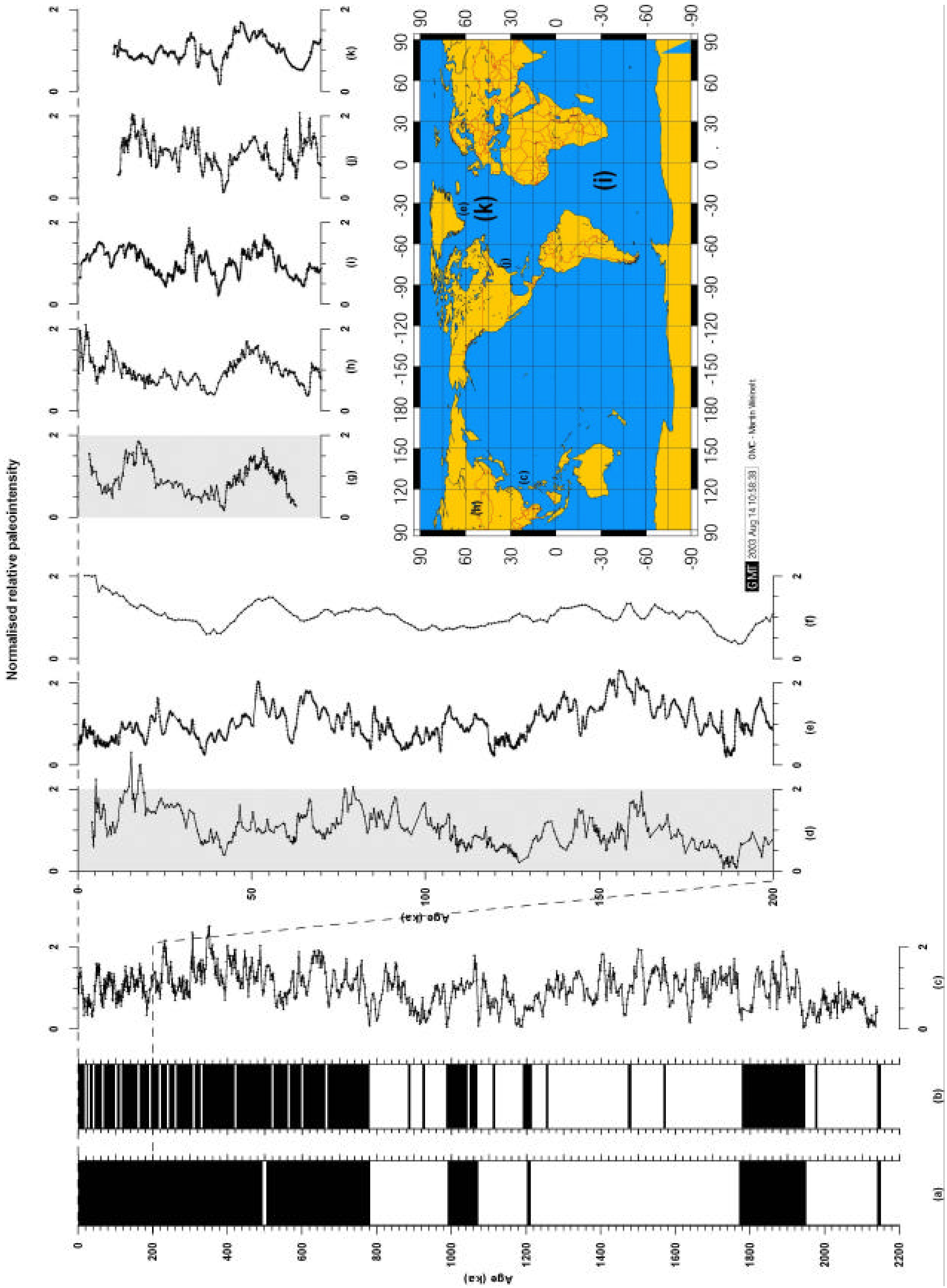
*Figure III.9: Schematic drawing of the journey of magnetic particles from the water column to burial (after Tauxe, 1993). The vector **B** represents the geomagnetic field.*

Afterward, the DRM, i.e. the NRM cleaned from post depositional processes, is normalised by the concentration of ferromagnetic particles, and provides a good estimation of the relative intensity of the magnetic field at the time of particle deposition. The ChRM gives the direction of the magnetic field. These parameters, measured continuously on a sediment sequence, depict a pattern corresponding to the geomagnetic field behaviour variations through the time.

III.2.2. Geomagnetic field variations

The Earth's magnetic field is generated by the geodynamo, which consists of iron-rich fluid convection in the outer core. On the Earth's surface, the geomagnetic field is roughly dipolar with a present day North Pole located in North Canada (11° far from the geographic North Pole). However, due to changes in the outer core fluid dynamics, the magnetic poles can move, or even reverse (Fig. III.10a & b). The strength (intensity) of the geomagnetic field changes as well (Fig. III.10c). During the last 780 ka, the geomagnetic field is characterised by a relative stable normal polarity (Brunhes chron; Fig. III.10a & b). In the last decades of paleomagnetic investigation, several short reversal (excursions) superimposed on the prevailing normal polarity of the Brunhes chron have been documented by Bonhommet and Babkine (1967), and also in recent syntheses (Nowaczyk et al., 1994; Langereis et al., 1997 and Thouveny et al., 2004). Excursions of the geomagnetic field are generally related to periods of low field intensity. Low intensities of the natural magnetisation of sediments coincides further with the enhanced production of cosmonucleids (Wagner et al., 2000; Carcaillet et al., 2004), which is also modulated by the intensity of the earth's magnetic field. During sedimentation, magnetic particles able to carry a remanence record the ambient magnetic field (see DRM, Fig. III.9). In addition to the excursions, the geomagnetic field is characterised by small deviations from its expected dipolar direction, i.e. paleosecular variations. These paleosecular variations have a period up to 10^5 years (Butler, 1992).
In Lake Baikal, no paleosecular variations could be reconstructed since the complex dynamic of sedimentation and the active tectonic context have generated slight disturbances. We could misinterpret these disturbances as paleosecular variations. The direction of the stable magnetisation measured in the sediment (ChRM, see section III.2.1) is thought to represent the primary DRM. ChRM is characterised by its inclination, i.e. the angle between the measured vector and a horizontal plane and its declination, i.e. its horizontal deviation from the geographic North Pole.

Figure III.10 (next page): Paleomagnetic records from various sites and their location on a world map. Synthetic polartiy scale after (Cande and Kent, 1995)(a) and a compilation by Nowaczyk using (Langereis et al., 1997);(Nowaczyk and Frederichs, 1999);(Singer et al., 1999);(Nowaczyk and Knies, 2000);(Lund et al., 2001);(Channell et al., 2002);(Singer et al., 2002)(b). Relative paleointensity records from Philippines Sea (Horng et al., 2003)(c); CON 01-603-2, Lake Baikal (present study)(d); ODP Site 983, North Atlantic (Channell et al., 1997)(e); synthetic curve SINT 200 (Guyodo and Valet, 1996)(f); CON 01-605-3, Lake Baikal (present study)(g); Lake Baikal (Peck et al., 1996)(h); Synthetic curve SAPIS (Stoner et al., 2002)(i); Blake Outer Ridge (Schwartz, 1998)(j); Synthetic curve NAPIS (Laj et al., 2000)(k).


Normalised relative paleointensity
Age (ka)
(a)
(b)
(c)
(d)
(e)
(f)
(g)
(h)
(i)
(j)
(k)

III.2.3. Records of the earth's magnetic field

Concerning the present day situation, modern techniques (mainly satellites) are able to detect instantaneous changes of the geomagnetic field (Holme et al., 2003). Concerning past field variations, only an approximate estimation of the direction and intensity is possible. However, the sensitivity and the capacity of measurements from the newest magnetometers allow determination of the remanent magnetisation with high precision. Since sediments can record the ambient geomagnetic field at the time of their deposition (see section III.2.1) their natural remanent magnetisation can be used to reconstruct the past changes of the field. Generally, paleomagnetism has two main implications in quaternary studies. First, paleomagnetic records dated by radio-chronology or by Optical Stimulating Luminescence (OSL) improve our fundamental knowledge about variations of the geomagnetic field in terms of intensity and direction. These databases can be used for feeding geomagnetic models (e.g. Korte and Constable, 2003). Second, well-dated geomagnetic records comprise a useful tool for establishing an age model. Correlation of paleointensity records with published curves provides a high-resolution age model beyond the physical limits of *AMS* ^{14}C dating method. The time resolution of paleomagnetic investigations on sediments depends on two further parameters, i.e. the sampling step and the sedimentation rate. In the present study, the sampling step is a fixed parameter, which approximates to 2.2 cm. Thus, the resolution depends solely on the sedimentation rate. The resolution of the paleomagnetic records from Lake Baikal ranges from 800 years for the lowest sedimentation rate (Fig. III.10d) to 120 years for the highest sedimentation rate (Fig. III.10g). In the studied sediment profiles, one full reversal was found and dated at about 185 ka (Iceland Basin event, Fig. III.11). The age model of CON 01-603-2 is based on correlation to ODP site 984 (Channell, 1999). Since this event was also recorded as a complete reversal at another site in North Atlantic (Channell et al., 1997), it might be an indication of the dipolar character of this event. However, many questions about the appearance and distribution of geomagnetic excursions are still in debate because such polarity events are not always recorded directly in the sediment. This might be due to non-dipolar behaviour of the local geomagnetic field at times of the excursion. Another crucial point is the post-depositional mineralogical changes, which might affect the primary geomagnetic signal recorded in the sediment. As has already been mentioned, to ensure paleomagnetic curves quality criteria have been established in the present study to avoid misleading interpretations (King et al., 1983; Tauxe, 1993) (see section III.1.5 and Chapter V).

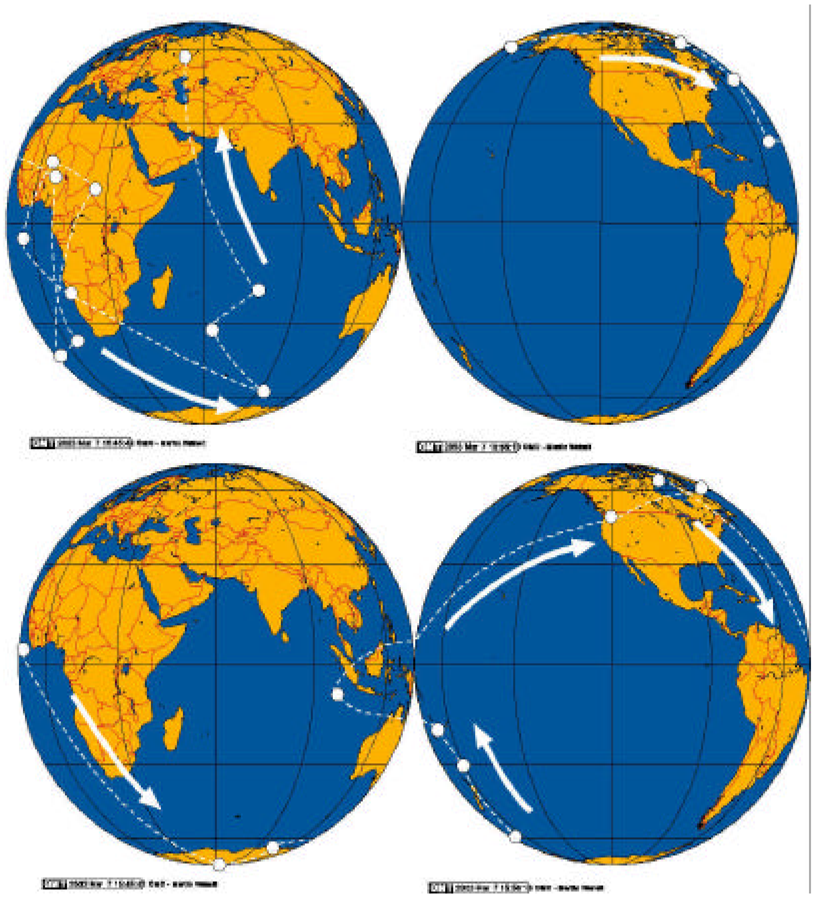


Figure III.11: Virtual Geomagnetic Pole (VGP) path of the Iceland Basin event recorded in two different cores from Lake Baikal. TOP at a site of the Academician Ridge (Oda et al., 2002) BOTTOM at a site of the Continent Ridge (Demory et al., 2003).

III.3. Laser-assisted grain size analysis

Besides the rock magnetic study, which has many implications, notably climatic (Chapter IV) and the paleomagnetic study offering a high-resolution age model (Chapter V) we gathered other climatic proxies. The study was complemented by a laser-assisted grain size analysis. The results are presented in chapter VI.

III.3.1. Principle of laser-assisted grain size analysis

The size distribution of particles in sediments supplies information about transport pathways e.g. wind, rivers and turbiditic currents in the depositional environment. Additionally, the particles that formed in situ as a result of bioproductivity and/or diagenesis contribute to the overall grain size distribution of the sediment. We prepared a tedious protocol for removing the biogenic fraction (commonly diatoms and spicules of sponges), the possible diagenetic particles and the clay fraction.

Laser diffraction analysis is based on the principle that particles of a given size diffract light through a given angle, which increases with decreasing particle size. A parallel beam of monochromatic light passes through a suspension, and the diffracted light is focused onto a multi-element ring detector. The detector measures the angular distribution of scattered light intensity (Fig. III.12). Laser-assisted grain size analysis has been performed on the detrital fraction > 2 µm. Grain size distribution has been calculated using the refractive index of quartz (1.56) for the sediment and 1.0 for the water. Fixed to the measuring equipment, we used the manual dispersion unit Hydro SM for wet sample analysis because very small quantity of sediment suspension can be measured (50 to 120 ml). The dispersion mechanism is a continuously variable single shaft pump and stirrer. The filling, draining and cleaning are manual.

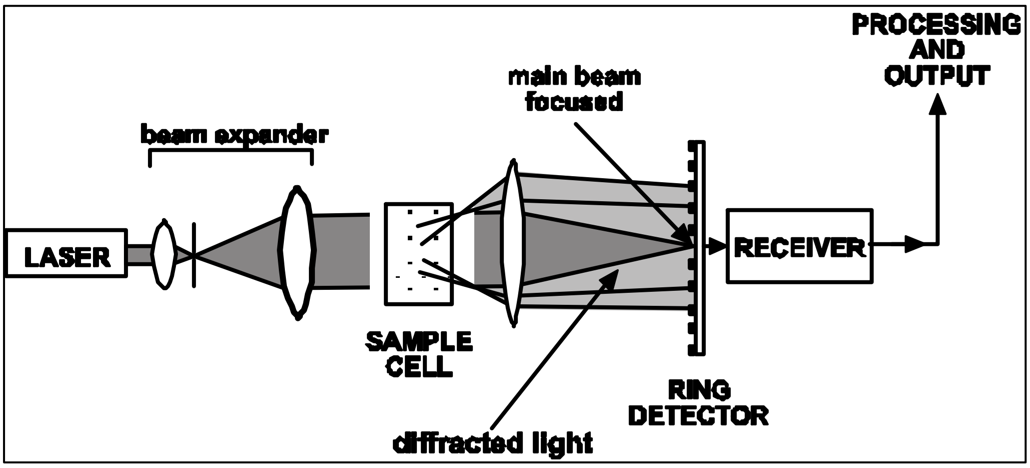


Figure III.12: Schematic diagram of a laser particle size analyser showing the primary components of light source, sample, focusing lens, detector, and processing system (from McCave et al., 1986).

III.3.2. Preparation of samples

448 samples have been investigated. They were continuously sampled from an 8.8 m-long sediment core retrieved at the Continent Site. This site has continuous sedimentation and sedimentation rate is rather low (~ 6.5 cm ka^{-1}). A low sedimentation rate could amplify the information on the eolian input related to the atmospheric circulation in Northern hemisphere. The sedimentary section covers the last 140 ka and, therefore, the last two interglacial periods (the Holocene and the Eemian).

Dissolution of the organic carbon:
Freeze dried sediments (weight from 1.5 to 2 g) were put into a H_2O_2 solution (5%). The solution was agitated for a few seconds in an ultrasonic bath. Between 3 and 5 cm^3 of H_2O_2 were necessary to dissolve the organic carbon. During this dissolution, the samples were shaken. The dissolution takes 4 hours for an organic carbon-poor sample (in Lake Baikal, the clay-rich layers from the glacial periods) and about 2 weeks for an organic carbon-rich sample (in Lake Baikal, the diatomaceous layers from the interglacial periods).

Separation and quantification of the clay fraction:
The clay fraction was separated by centrifugation at 1000 rev/min for one minute. The clay fraction stayed in the suspension which was poured in a container. The centrifugation was repeated until the suspension was perfectly clean (for some samples, it took up to 30 times). The clay fraction settling took about four weeks. Afterwards, the excess water was pumped and the clay fraction was freeze dried and weighed.

Dissolution and quantification of the biogenic silica:
The samples were heated for 5 hours at 90°C in Na_2CO_3 solution (2M) in a shaking bath. The volume of the solution needed ranges from 5 cm^3, for biogenic silica-poor samples, to 12 cm^3 for biogenic silica -rich samples. The volume of the created suspension was measured as well

as its silicium concentration (measured with VARIAN Vista MPX at Potsdam University). The remaining sediment was rinsed. Often clays are still present and can be added to the vessels containing the clay fraction.

Dissolution of carbonates and some authigenic minerals:
The remaining sediment was put in a HCl solution (1.1N) for 1 hour in an ultrasonic bath. The samples were rinsed again afterwards.

Density separation of the biogenic silica:
The efficiency of the alkaline dissolution of the opal was estimated from smear slides of the remaining sediments. As already known from previous studies (Olivarez Lyle and Lyle, 2002), opal dissolution was not always 100 % successful. Indeed, some diatoms (especially Aulacoseira) as well as spicules of sponges are resistant to Na_2CO_3. A density separation has been performed on 56 samples using a solution of poly-tungsten of sodium of a density of 2.32 g cm^{-3}. Indeed, the difference of density, ~ 1.9 g cm^{-3} for the opal, and ~ 2.65 g cm^{-3} for the quartz (main contributor of the detrital fraction), makes the separation possible. The effects of diatom relics are observed on grain size distributions (Fig. III.13) as well as on downcore variation of grain size parameters (Fig. III.14).

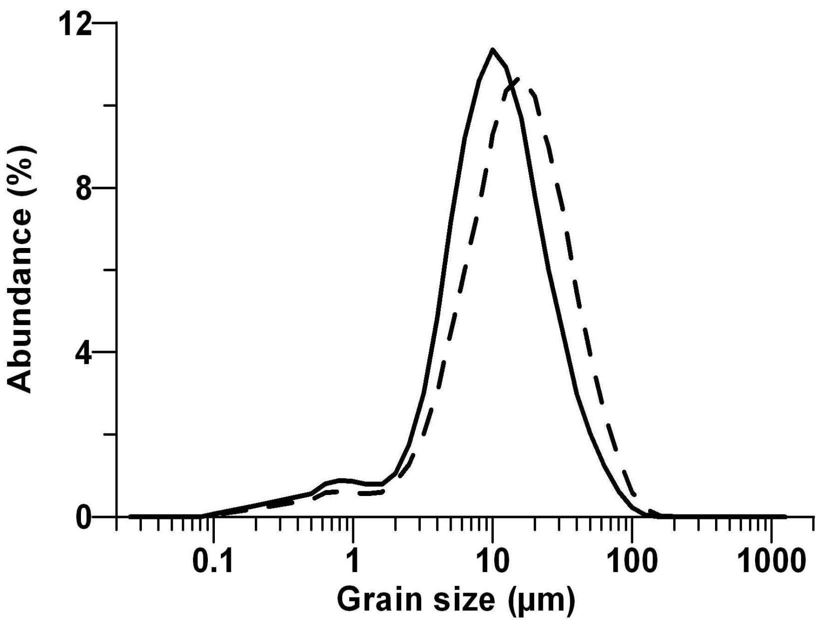


Figure III.13: Grain size distribution of the sample taken at depth 59.7 cm of core CON 01-603-2. Dashed line, with opal dissolution, continuous line with opal dissolution combined with density separation. The difference in the distribution is due to relics of diatom (especially Aulacoseira), which were still preserved after the NA_2CO_3 treatment.

Quantification of the fraction > than 2 µm:
For each sample, the fraction > 2 µm, was divided in two aliquots. One was used for the grain size analysis and the other was freeze dried and weighted. The laser assisted grain size analysis has been performed at the University of Lille (France) using the Malvern Mastersizer 2000; the measurement is automatically done twice and both results are averaged.

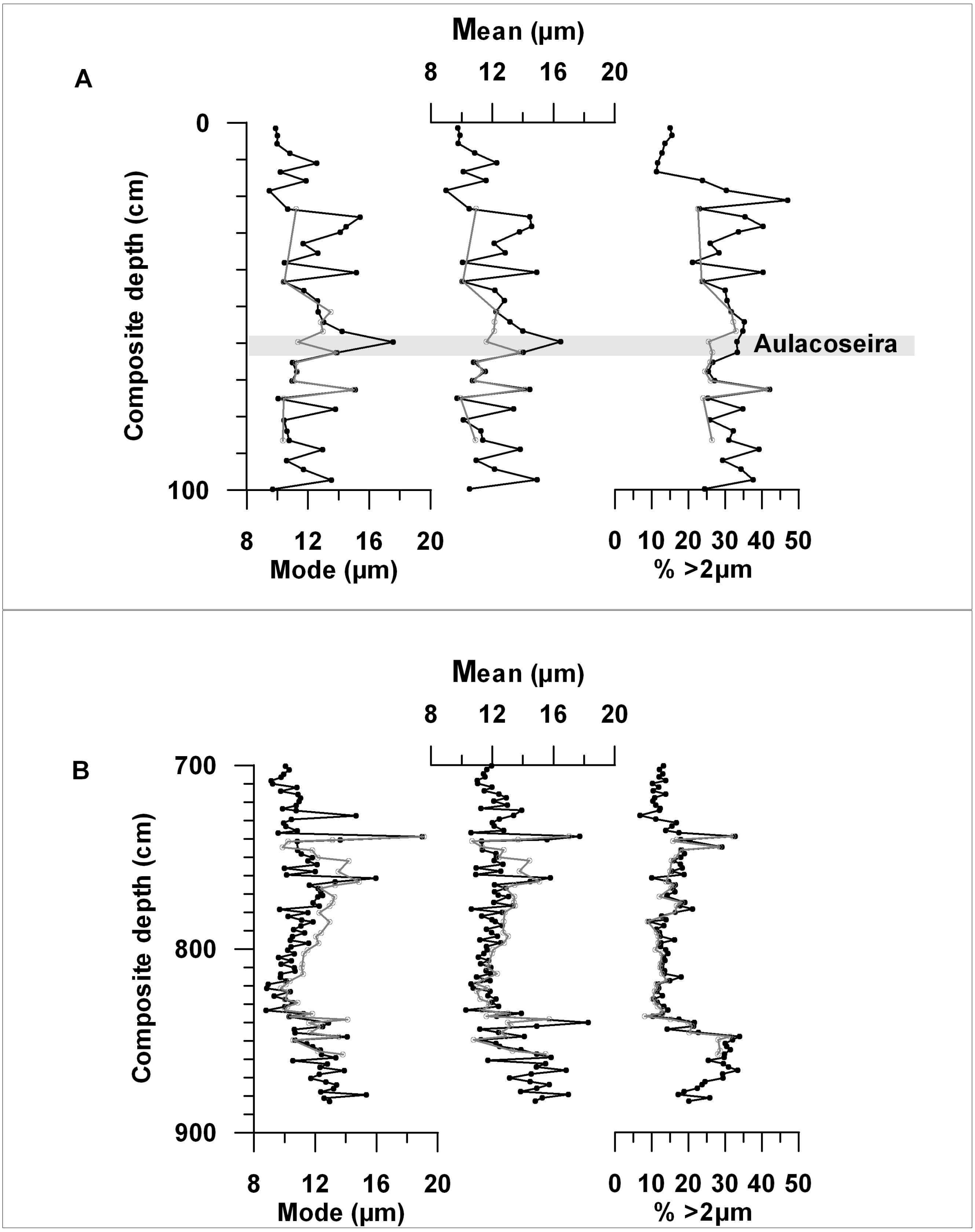


Figure III.14: Downcore variations of some grain size parameters such as mode, median and % of the detrital fraction >2 µm for sediment sequence CON 01-603-2. Grey curves denote the results after density separation of opal. A. Top of the section covering the Holocene and the termination 1, a zone with a high concentration of Aulacoseira is also highlighted in grey; B. Section covering the Kazantsevo interglacial (equivalent to Eemian in Europe).

III.3.3. Estimation of an error bar

The error bar reported by the manufacturer (Malvern) on the measurement device is very low (< 1%) but increases with the decrease of the size measured: it reaches 2 % between 900 and 5 µm and 6 % for the finer fractions.
All the data presented in this thesis result from the average of two runs of measurement, the data were accepted only when both measurements were visually similar. The error due the manipulation of the samples during the sample preparation is probably higher but difficult to quantify. The most interesting test of reproducibility, which could help to quantify the error, was done when density separation was needed. For 56 samples already measured, the grain size had to be re-measured after density separation (Fig. III.15). Apart from the samples that presented strong differences before and after analyses due to presence of diatoms, most of the samples show a second measurement very close to the first one. From this double analysis we estimated an error due to laboratory manipulation of ± 1 µm on the mode of the grain size distribution.
Concerning the quantification of the different fractions, we estimated an error of ± 5 %.

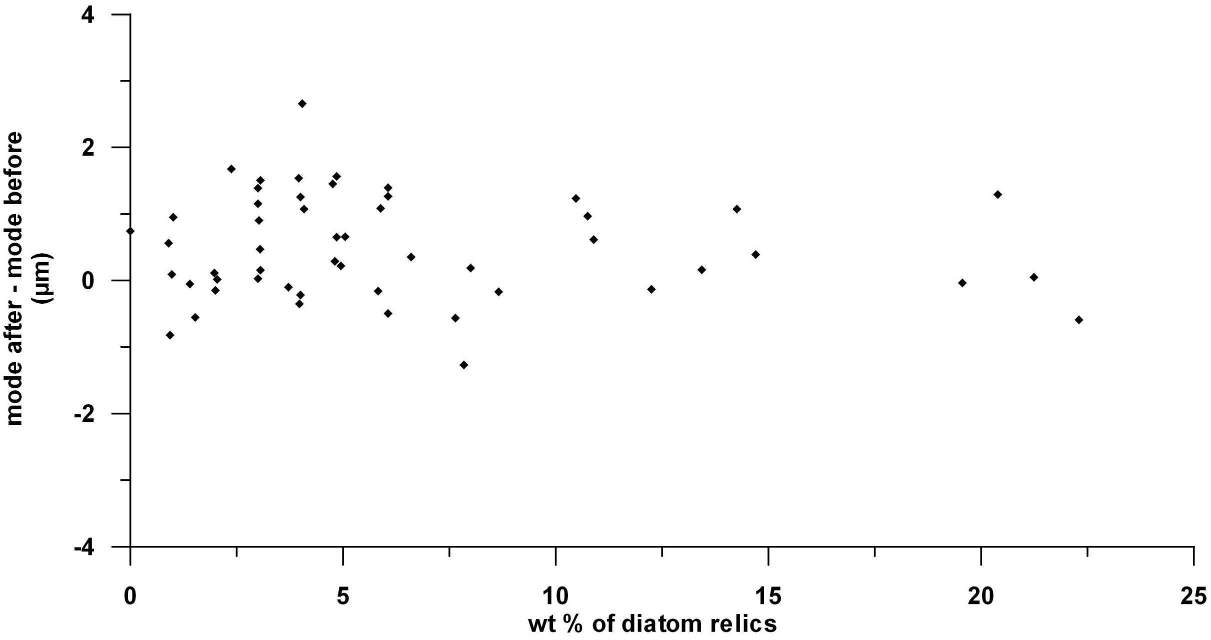


Figure III.15: weight fraction of diatom relics versus difference between modes after and before density separation, here plotted for 52 of the 56 samples (4 samples are out of the window).

IV. Detrital input and early diagenesis in sediments from Lake Baikal revealed by rock magnetism

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Abstract

A rock magnetic study was performed on sediment cores from 6 locations in Lake Baikal. For a comprehensive approach of the processes influencing the rock magnetic signal, additional data are presented such as total organic carbon (TOC), total sulphur (TS), opal, water content and relative variations in iron and titanium measured on selected intervals. In glacial sediments the magnetic signal is dominated by magnetite, which is considered to be of detrital origin. This predominance of magnetite is interrupted by distinct horizons of authigenic greigite, probably confined to reductive microenvironments. In interglacial stages, besides dilution by biogenic silica and a decreasing detrital input, the weakness of the rock magnetic signal is also due to a reductive dissolution of magnetic particles. The magnetic assemblage is strongly linked to the redox history of interglacial sediment. In the oxidised bottom sediments of Lake Baikal, a biogenic magnetite is observed (Peck and King, 1996). After burial under the redox front the magnetite is preferentially dissolved and detrital hematite remains dominant when the sedimentation rate is low and when the residence time of the magnetite close to the redox boundary is long. During these low sedimentation rate conditions, the redox front is preserved (Granina et al., 2004). At constant sedimentation rate and fast burial, the magnetite is preserved or transformed into greigite when sulphate reducing conditions are reached in the sediment. In interglacial sediments, the magnetic assemblages depict changes in the sedimentation rate, which are traced using the ratio of magnetite over hematite (S-ratio). At the beginning of interglacials, the sedimentation rate is constant with an assemblage magnetite + greigite (high S-ratio) and at the end of some interglacials the sedimentation rate decreases with a predominance of hematite (low S-ratio).

Keywords: Lake Baikal - Late Quaternary - Rock magnetism – Diagenesis - Detrital input

IV.1. Introduction

Using rock magnetism, we can reconstruct the entire depositional history of magnetic particles in lake sediments. The magnetic minerals assemblage in sediments is first of all influenced by weathering and transport processes occurring in the catchement areas. After deposition, magnetic minerals can be altered, depending on chemical and biogenic processes occurring at the water/sediment interface and during diagenesis (e.g. Verosub, 1977; Snowball, 1993). In addition, during laboratory storage of the cores, the magnetic signal may also be altered (Thompson et al., 1980; Hilton, 1987). If diagenetic features such as dissolution and neoformation of magnetic particles can be identified and ruled out, rock magnetic properties become a powerful tool for tracing climate dynamics (e.g. Thouveny et al., 1994).
Since Lake Baikal sedimentation sensitively traces climate changes (Colman et al., 1995), it has been a natural laboratory for rock magnetic studies. However, rock magnetism was mainly used as a

correlation tool since the concentration of magnetic minerals reflects the lithological variations in the sediment (e.g. Sakai et al., 2000; Kravchinsky et al., 2003). Specific studies identified fine-grained biogenic magnetite as one carrier of the magnetic signal in bottom sediments (Peck and King, 1996), which may however be susceptible to reductive diagenesis (Snowball, 1994). Peck et al. (1994) have also shown that magnetic properties are linked to climatic variations with biogenic dilution during interglacial periods and with the increase of terrigenous high coercivity minerals during glacial periods. For post-depositional processes, Dearing et al. (1998) have shown a weak reductive diagenesis.

The goal of this study is to better understand the links between detrital input, biogenic productivity and early diagenesis, which influence the rock-magnetic signal. This in turn allows us to validate the paleomagnetic variations recorded in Lake Baikal sediments (Demory et al., submitted-b), while ensuring a more critical interpretation of the signal carried by rock magnetic parameters. For this purpose, we carefully defined the magnetic mineralogy of different sediment cores of Lake Baikal and compared them to other sediment characteristics such as organic carbon, water, opal, iron and titanium contents.

IV.2. Material and methods

Lake Baikal (eastern Siberia) is a rift valley lake located at 51°N-56°N and 104°E-110°E. It consists of three deep basins separated by two ridges (Fig. IV.1) with a 7.5 km thick sediment cover accumulated since the Miocene (Hutchinson et al., 1992). The major catchment area of Lake Baikal is located to the east-southeast. The hinterland is mainly composed of batholiths largely of granitic composition (e.g. Antipin, 1997), which provide many kinds of magnetic minerals. The Lake Baikal sediments are built up from detrital input and biogenic productivity, which are modulated by climatic conditions. The biogenic silica production is very low during glacial/stadial periods and high during interglacial/interstadial periods and outlines, therefore, the climatic variations associated with the expansion and retrait of northern hemisphere ice sheets (BDP-93 and Members, 1994; Colman et al., 1995; Grachev et al., 1997; Prokopenko et al., 2001).

Six sites were cored on geomorphologic highs on the margins of the basins (Fig. IV.1), to be far from turbiditic and river influences. Hence, the detrital input is mainly influenced by westerly and local winds, which are also likely to be the main source of transport for eolian sediment input into Baikal. Three piston and two pilot cores retrieved from three sites at Academician Ridge during a coring campaign in 1998 and three piston and three pilot cores retrieved during a coring campaign in 2001 were investigated (Table IV.1). Coring locations of 2001 are the Continent Ridge close to the northern basin (site CON 01-603), the Posolsky Bank between the central and the southern basin (site CON 01-604) and the Vydrino Shoulder on the southern flank of the southern basin (site CON 01-605). Due to the coring method, the uppermost sediments of the piston cores were lost. Where possible, the sedimentary column was completed by a combination of pilot and piston cores (see Demory et al., submitted-b).

Rock magnetic samples were continuously taken using 6.2 cm^3 cubic plastic boxes (Table IV.1). The anisotropy of magnetic susceptibility (AMS) was measured using a Kappabridge KLY-3S (AGICO Brno, Czech Rep.). The AMS ellipsoid is characterised by its three principal axes k_{max}, k_{int}, k_{min}, and their dip and azimuth. The AMS orientation gives, as a first approximation, the preferred orientation of the magnetic grains and therefore the orientation of the sedimentary fabric (e.g. Rees, 1965; Frank et al., 2002). In unlaminated sediments, the orientation of the AMS axes indicates whether the coring is perpendicular to the sedimentary deposits, thereby allowing the determination as to whether there is any evidence for natural deformations of the sedimentary fabric. Furthermore the shape of the AMS ellipsoid is informative as an oblate shape is typical for compacted sediments whereas a prolate shape could reflect any deformation of the sedimentary fabric.

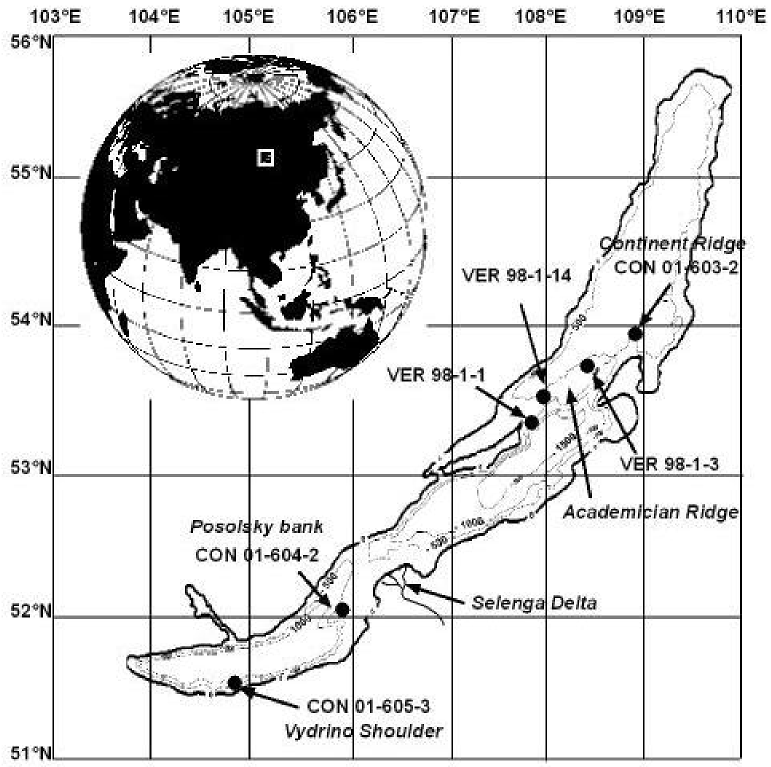


Figure IV.1: Simplified bathymetric map of Lake Baikal. The 6 core locations are marked by black dots.

Core name and type	Location	Water depth (m)	Latitude	Longitude	Length (cm)	Number of samples
VER 98-1-14 Piston	Academician Ridge	412	53°31'23''N	107°58'10'' E	983	383
VER 98-1-14a Pilot	Academician Ridge	412	53°31'23''N	107°58'10'' E	192	80
VER 98-1-3 Piston	Academician Ridge	373	53°44'56''N	108°19'02'' E	831	331
VER 98-1-1 Piston	Academician Ridge	245	53°23'36''N	107°55'22'' E	1120	470
VER 98-1-1a Pilot	Academician Ridge	245	53°23'36''N	107°55'22'' E	100	41
CON 01-603-2 Piston	Continent Ridge	386	53°57'48''N	108°54'47'' E	1127	479
CON 01-603-2a Pilot	Continent Ridge	386	53°57'48''N	108°54'47'' E	190	82
CON 01-604-2 Piston	Posolsky Bank	133	52°04'46''N	105°51'27'' E	622	268
CON 01-604-2a Pilot	Posolsky Bank	133	52°04'46''N	105°51'27'' E	188	82
CON 01-605-3 Piston	Vydrino Shoulder	675	51°35'06''N	104°51'17'' E	1052	453
CON 01-605-3a Pilot	Vydrino Shoulder	675	51°35'06''N	104°51'17'' E	173	74
					Total:	2743

Table IV.1: Name, type, length of sediment cores presented in this study as well as their location, water depth and number of discrete rock magnetic samples.

Low field magnetic susceptibility (κ_{LF}) was measured in order to estimate the contribution of ferrimagnetic particles in the sediments.
Anhysteretic remanent magnetisation (ARM) measurements, free of the effect of the dia-, para- and superparamagnetic components, provide an estimation of the magnetic remanence carriers, mainly of single domain (SD) grain size.
The isothermal remanent magnetisation (IRM) acquired in a 2 T direct current (DC) field using a pulse magnetiser, is defined as the saturation isothermal remanent magnetisation (SIRM). After applying this maximum field, all the samples were magnetised in a field of 300 mT in the opposite direction. The fraction of high-coercivity magnetic minerals was estimated by calculation of the ratio (Bloemendal et al., 1992):

$$\text{S-ratio} = \frac{1}{2}\left(1 - \frac{\text{IRM}_{-0.3\text{T}}}{\text{SIRM}_{2\text{T}}}\right)$$

S-ratio values range from 0 to 1. The value 1 is obtained for pure magnetite or greigite and decreases with increasing proportion of antiferromagnetic particles such as hematite.
The quantity of high coercivity minerals (hematite or goethite) was estimated by the High Isothermal Remnant Magnetisation HIRM, which results from the following relationship:
$\text{HIRM} = (\text{IRM}_{2\text{T}}\text{-IRM}_{0.3\text{T}})/2$
This parameter has been used to estimate the detrital input, e.g. eolian input in ice records (Maher and Dennis, 2001) or in Lake Baikal sediments (Peck et al., 1994) despite discussions about its accuracy (Liu et al., 2002).
Hysteresis loops and backfield curves were determined for a representative population of samples using an Alterning Gradient Force Magnetometer (AGFM, Princeton Measurements Corp. model 2900 MicroMag). For this purpose, samples were freeze-dried and gently mortared, and mixed with an Agar jelly agent suspension and solidified in a 3×3 mm cylindrical mould. The coercive force (B_c) and the ratio of saturated remanent magnetisation (M_{rs}) and saturation magnetisation (M_s) were determined from the hysteresis loop measurements. The coercivity remanence (B_{cr}) was deduced from the backfield curves of the saturated samples. The hysteresis parameters indicate the domain status of ferromagnetic particles (e.g. Day et al., 1977), which is linked to the absolute grain size and shape when magnetite dominates the signal (Dunlop, 2002a; Dunlop, 2002b).
High temperature measurements of magnetic susceptibility were performed on 13 representative samples using the same equipment as for κ_{LF} and AMS. Samples showing occurrence for greigite were especially targeted. Indeed, thermomagnetic measurements are generally performed on previously dried sediments whereas it is known that greigite can be unstable on contact with air (Oldfield et al., 1992). Therefore, relevant thermomagnetic curves were obtained only on wet sediments from the intervals where greigite was suspected. Thermomagnetic curves for these wet samples were determined using a “Variable Field Translation Balance” (VFTB), which is a combination of a Curie balance with a vibrating sample magnetometer (Nowaczyk et al., 2000). The measurement was performed at saturation magnetisation and in an argon atmosphere.
Magnetic extracts were analysed by X-ray diffraction and transmission electronic microscopy (TEM) using a STOE Stadi P diffractometer and a Philips CM200 transmission electron microscope respectively. Magnetic minerals were extracted from an interval dominated by magnetite at depth 527.5 to 529.9 cm (clay-rich layer) and one from an interval dominated by hematite at depth 583 to 590 cm (diatomaceous layer) in core VER 98-1-1. The extraction was done after removal of the organic carbon using a magnetic needle plunged in the sediment suspended in water in an ultrasonic

bath. The goal was to better constrain the rock magnetic results where the X-ray analysis gives a semi-quantitative characterisation the components of the magnetic extract (Rietveld structure refinement method using the program GSAS, see Larson and Von Dreele, 1987) and the TEM allows magnetic grain size to be directly observe and to proceed to elemental dispersive spectra on specific minerals.

We measured on the same cores water, opal, total organic carbon (TOC), total sulphur (TS), (relative) iron and (relative) titanium contents. The water content was determined from samples retrieved with cylindrical syringe samplers, weighed wet and weighed again after freeze-drying. Opal content was obtained by measuring the $NaCO_3$ leached solution by inductive coupled plasma with an optical emission spectrometer (ICP-OES). Total organic carbon (TOC) and total sulphur (TS) were determined using a LECO CS-225-Analyser. A semi-quantitative estimate of titanium, iron and sulphur contents was obtained on selected intervals using X-ray fluorescence (XRF) logging at 2 mm steps (Jansen et al., 1998).

IV.3. Sedimentary fabric

Visual screening of sediments to determine disturbances was successful whenever laminations occurred. As sediments of Lake Baikal are often unlaminated, we estimated the bedding direction by measuring the orientation of magnetic minerals using AMS. The AMS data are plotted in a stereographic projection (Fig. IV.2). As most AMS data plot with long axes at 0° and short axes at 90°, an overall horizontal stratification is evident. However for a few samples the short axis of AMS deviates from the vertical orientation indicating oblique laminations which are either due to coring artefacts as observed at the very top or bottom of the cores, or due to earthquake-induced slumps (Oberhänsli, 2000), with earthquakes being common in the Baikal rift valley (Sherman and Gladkov, 1999).

Furthermore, a slight preferential orientation parallel to the 0° direction of the stereographic projection is observed for the long axis of magnetic anisotropy (Fig. IV.2). As the samples were not geographically oriented, we interpret this preferred orientation to be a sampling artefact resulting from a slight shearing when the plastic box penetrates the sediment during subsampling rather than indicating paleocurrent directions (Oda et al., 2002).

In conclusion, the AMS results document that the sediments under investigations are mostly undisturbed and given the oblate shape of the AMS ellipsoids (Fig. IV.2), compaction has been effective.

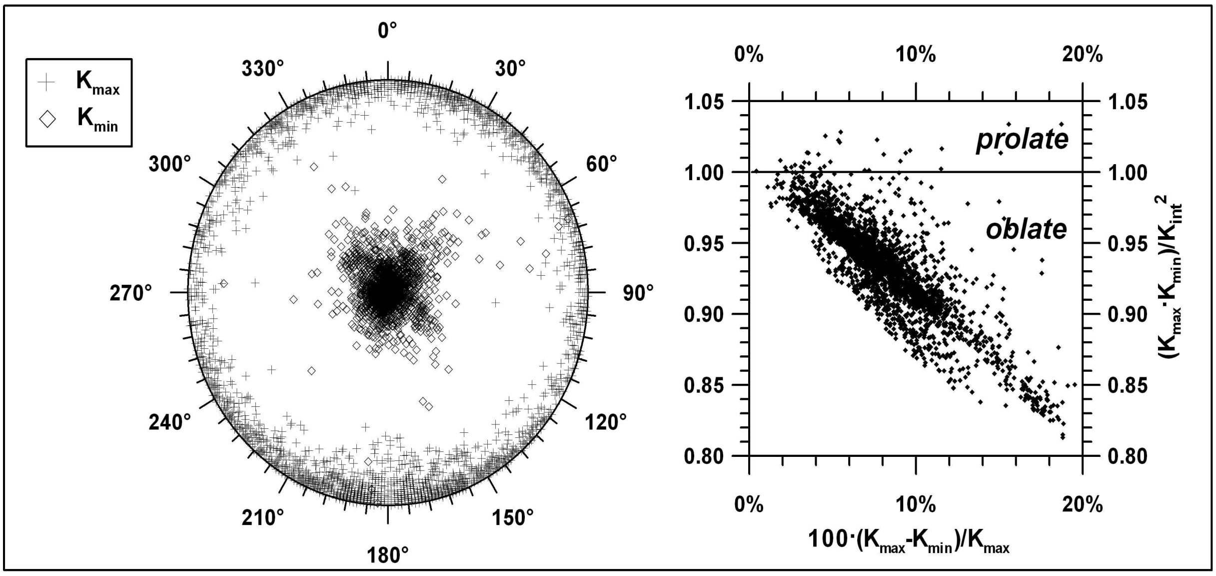


Figure IV.2: Stereographic projection of the minimum (k_{min}) and maximum (k_{max}) axes of the AMS ellipsoid and diagram of the shape of the AMS ellipsoid (Stacey et al., 1960) versus the degree of magnetic anisotropy (Graham, 1966) for all cores.

IV.4. Lithological variations and ARM

The typical Baikal sedimentary succession consists of clay-rich layers alternating with diatomaceous layers, indicating glacial (cold) and interglacial (warm) periods, respectively (BDP-93 and Members, 1994; Colman et al., 1995; Grachev et al., 1997) (Fig. IV.3). The clay-rich layers are characterised by a high ARM and low levels of TOC, TS, opal and water. The diatomaceous layers, indicating high biogenic productivity, show low ARM and high levels of TOC, TS, opal and pore water. Instead of using the magnetic susceptibility (κ_{LF}) for monitoring the varying concentration in magnetic minerals (e.g. Sakai et al., 2000), we used ARM to counter smoothing of the magnetic susceptibility signal by high contents in diamagnetic opal. Correlating the ARM from Lake Baikal cores with the orbitally tuned δ^{18}O curve from ODP 677 (Shackleton et al., 1990), we identified marine isotope stages (MIS) (Fig. IV.3) and estimated the time span recorded in sediments.

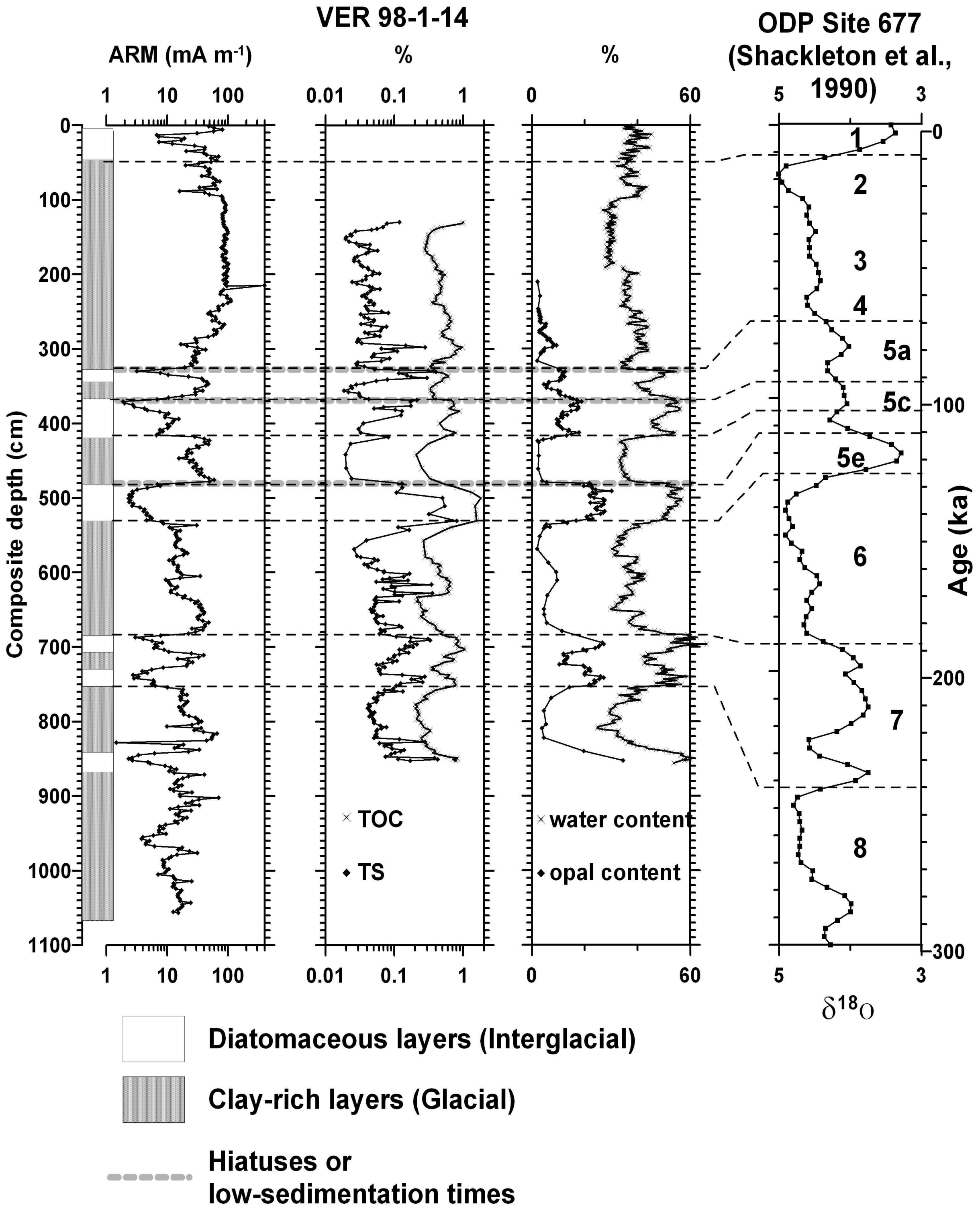


Figure IV.3: Downcore variations of the ARM, TOC, water and opal content and simplified lithology for the sedimentary sequence of VER 98-1-14. Correlation to MIS using the $\delta^{18}O$ variations measured in ODP Site 677 by (Shackleton et al., 1990).

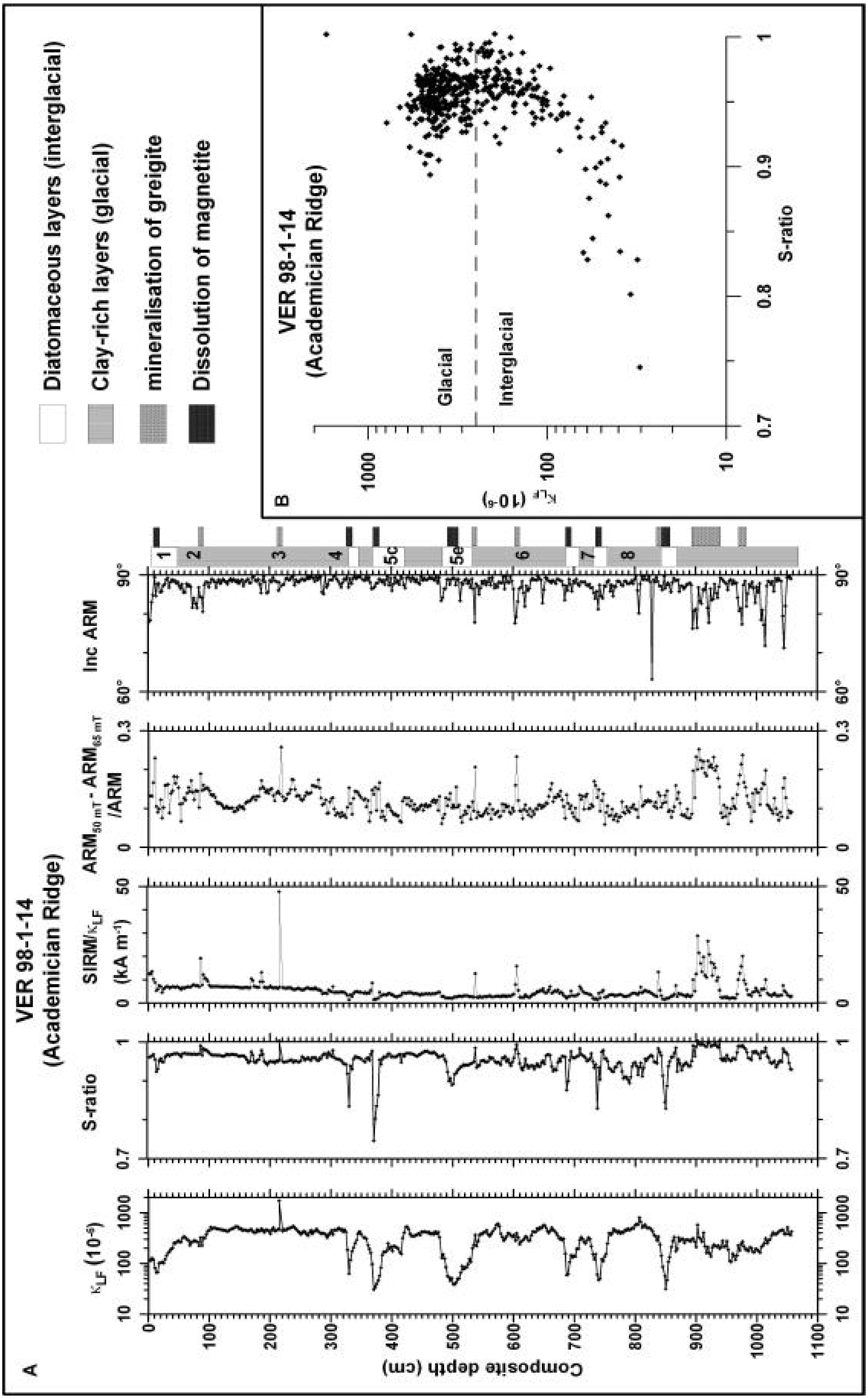
The ARM records also allow cores from different sites across Lake Baikal to be correlated (Fig. IV.4). Sediment discontinuities as suggested by lithological descriptions at several transitions from interstadial to stadials are marked by abrupt changes in ARM. We show later in this paper that these periods of low or no sedimentation strongly influence the magnetic mineralogy.

IV.5. Magnetic mineralogy - Results

Figure IV.4A shows variations of different rock magnetic parameters in one core in order to assess the magnetic mineralogy as described hereafter. The magnetic susceptibility does generally not exceed 500 $\times 10^{-6}$ SI, except at certain depths where peak values are about one order of magnitude higher. The S-ratio is rather constant with values around 0.95 over the whole depth range. Still, some lower values of about 0.75 to 0.85 are observed, coinciding with low susceptibility values. At certain depths the S-ratio is close to 1, e.g. between 900 and 940 cm. SIRM/κ_{LF} is constantly low, apart from single peaks which correlate with the highest S-ratios. The parameter indicating an ARM loss between 50 and 65 mT has a similar pattern than SIRM/κ_{LF}, but it is much more influenced by noise. The inclination of the acquired ARM is ~ 90°, apart from some exceptions. Lower inclination occurs sporadically and correlates partly (e.g. at 890-940 cm) with S-ratio, SIRM/ κ_{LF} and ARM_{50mT}-ARM_{65mT}/ARM maxima. Figure IV.4B shows that magnetic susceptibility is high in glacial and low in interglacial intervals. In addition it shows that the S-Ratio varies less in glacial sediments, ranging from 0.9 to 1, than in interglacial sediments where the S-ratio is scattered from 0.7 to 1.

In order to understand the previously described stratigraphic variations, we performed thermomagnetic analyses (susceptibility and induced magnetisation) from characteristic depths. Figure IV.5A shows the temperature dependence of magnetic low-field susceptibility for a sample from a clay-rich layer with a S-ratio of ~ 0.95, a κ_{LF} of ~ 500?× 10^6 SI, low value SIRM/κ_{LF} and no deviation of the ARM inclination. The susceptibility remains constant until 400°C, then increases until a peak value at ~ 560°C and decreases rapidly until 610°C. The sample in Figure IV.5B originates from a diatomaceous layer (low susceptibility and S-ratio). The susceptibility decreases steadily until 670°C, when the signal is lost. The samples in Figures IV.5C and IV.5D come from depths where almost all magnetic parameters have peak values (see Figs. IV.4 and IV.12). Both graphs indicate a clear maximum at 300°C. The magnetisation then decreases until 600°C, above this temperature the magnetisation is zero. It should be noted that all curves are irreversible, indicating the thermal instability of the magnetic minerals present in the sediments.

Figure IV.4: A. Downcore variations of rock magnetic parameters and simplified lithological description for the sedimentary sequence VER 98-1-14. Here MIS are denoted by numbers in the lithological column. The black squares filled intervals mark occurrences of greigite characterised by a high magnetic susceptibility (k_{LF}) in parts with a low coercive mineral dominating the magnetic signal (S-ratio close to 1), a high SIRM/k_{LF}, a strong loss of ARM intensity between the demagnetisations steps 50 and 65 mT and finally a deviation of the inclination of ARM. The dark grey intervals mark occurrences of magnetite dissolution with a low S-ratio resulting from relative higher hematite content in the ferromagnetic components. The assignment of greigite and dissolved magnetite is based on subsequent interpretation; B. Magnetic susceptibility (k_{LF}) versus S-ratio for the sedimentary sequence VER 98-1-14 showing S-ratio gathered around 0.95 in glacial sediments and scattered from 0.7 to 1 in interglacial sediments.



A
VER 98-1-14
(Academician Ridge)
κ_{LF} (10^{-6})
S-ratio
SIRM/κ_{LF}
(kA m^{-1})
ARM$_{50\ mT}$ - ARM$_{65\ mT}$
/ARM
Inc ARM
Composite depth (cm)
B
VER 98-1-14
(Academician Ridge)
Glacial
Interglacial
κ_{LF} (10^{-6})
S-ratio
Diatomaceous layers (interglacial)
Clay-rich layers (glacial)
mineralisation of greigite
Dissolution of magnetite

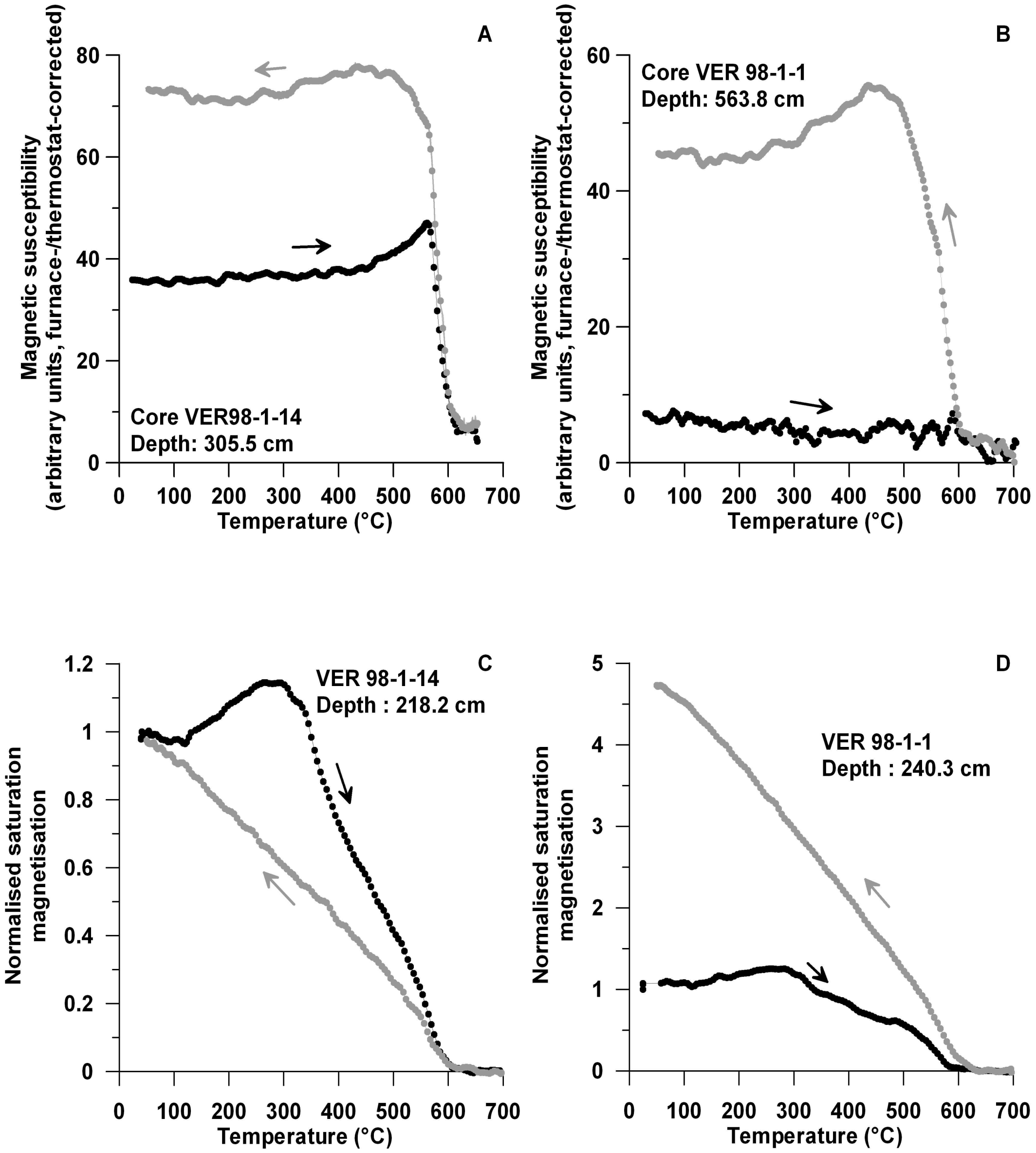


Figure IV.5: A. High temperature measurements of magnetic susceptibility in clay-rich layers: magnetite characterised by a Curie temperature of 590°c dominates the signal; B. High temperature measurements of magnetic susceptibility in diatomaceous layers: most of the weak susceptibility is still present up to the temperature of 670°C and carried by hematite; C. and D. High temperature measurements of the saturation magnetisation, both samples show a loss of a part of the signal at temperature between 350 and 400°C, typical disintegration temperature for greigite. The remaining signal disappears above a temperature of 590°C, typical for magnetite.

Figure IV.6A shows hysteresis measurements of three samples with different S-ratios. The highest saturation magnetisation (M_s) and coercive force (B_c) was observed in the sample with a S-ratio > 0.95, the lowest M_s and B_c in the sample with a S-ratio < 0.9. In all cases the hysteresis curve is influenced by paramagnetism, the effect being largest in the sample with low S-ratio. The day plot (Fig IV.6B) indicates that samples with S-ratios > 0.95 plot rather in the single-domain (SD) to pseudo-single-domain (PSD) range. Samples with S-ratios between 0.9 and 0.95 instead, plot rather in the PSD to the multi-domain (MD) range. Low S-ratio samples are not greatly dispersed and have ratios of B_{cr}/B_c and M_{rs}/M_s of 3.5 and 0.1 respectively.

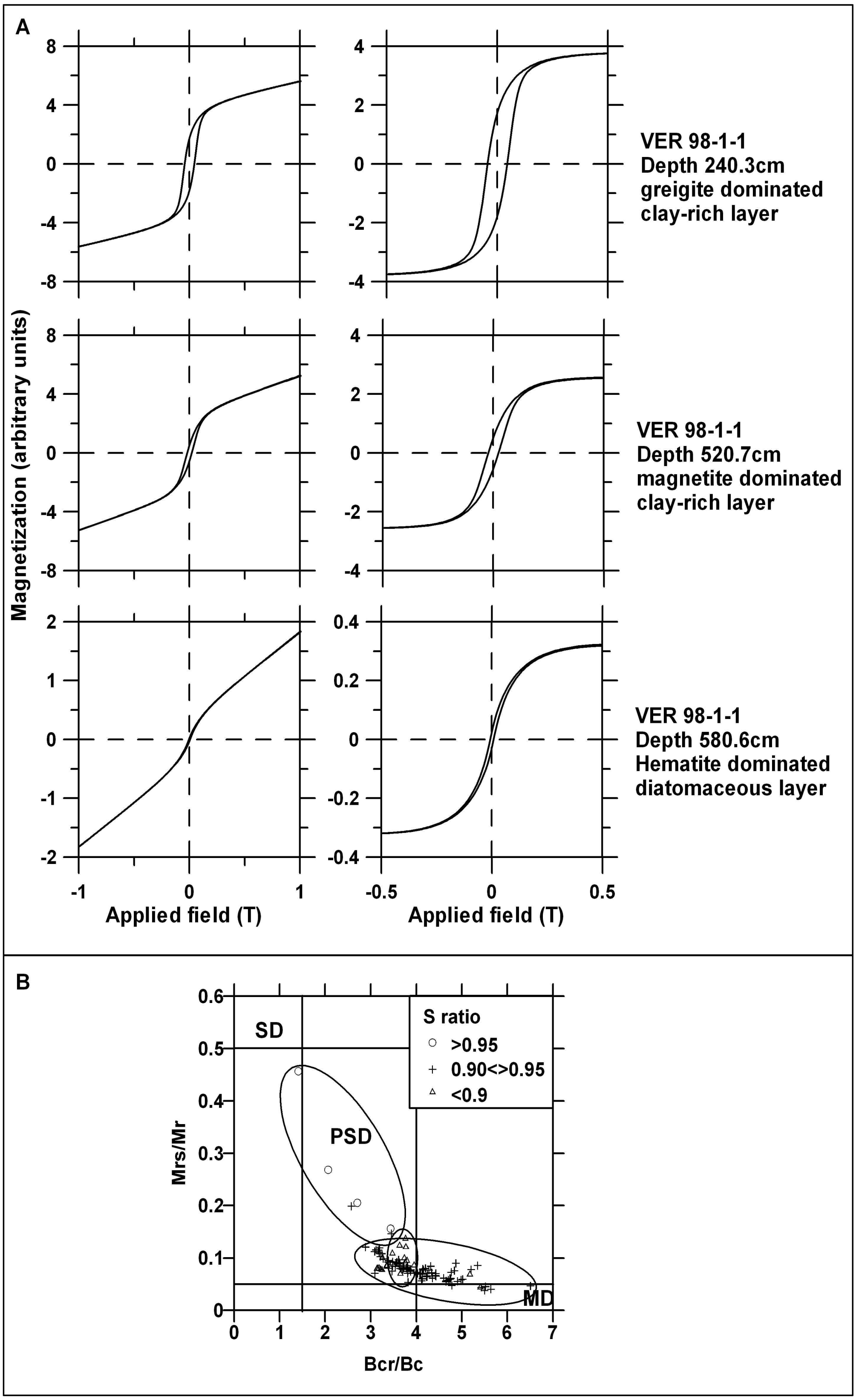


Figure IV.6: A. Three characteristic hysteresis loops uncorrected (to the left) and corrected for the paramagnetic influence (to the right). On top a single domain behaviour hysteresis loop for a sample with occurrence of greigite, in the middle a pseudo-single domain behaviour of the hysteresis loop from a clay-rich layer and below a pseudo-single to multi-domain behaviour of the hysteresis loop from a diatomaceous layer. B. Day plot (Day et al., 1977) of hysteresis parameters of representative samples with the boundaries between single-domain (SD), pseudo-single-domain (PSD) and multi-domain (MD) area. We classified the samples according to their S-ratio (see the three ellipses).

IV.6. Magnetic mineralogy – Interpretation

Three different magnetic mineral populations seem to be present in the sediment cores from Lake Baikal. The temperature dependence of magnetic susceptibility indicates a magnetite/maghemite population for S-ratios between 0.9 and 0.95, which is probably of detrital origin because it occurs in clay-rich sediments deposited during glacial-periods (Fig. IV.3). Energy dispersive X-ray spectroscopy (EDS) on transmission electronic microscopy (TEM) indicates an enhanced iron and oxygen content in analysed grains (Fig. IV.7), which are suggested to consist of iron oxides. The X-ray diffraction (XRD) spectra from magnetic extracts of clay-rich layers (Fig. IV.8A) support the previous suggestion, indicating magnetite as dominant mineral. S-ratios close to 1 indicate the absence of a high coercivity mineral component. The Curie temperature measurements (Fig. IV.5C and IV.5D) show a loss of remanence above 300°C, which is typical for the mineral greigite (Roberts, 1995). The hysteresis loop indicates a coercive force of ~ 45 mT, which is in agreement with greigite data from Roberts (1995), in line with other hysteresis parameters. Deflections from the ARM axis can be caused by a gyroremanent magnetisation acquired during static AF demagnetisation in anisotropic samples (Stephenson, 1980). Such gyroremanence has been identified in natural greigite samples (Hu et al., 1998; Stephenson and Snowball, 2001; Hu et al., 2002). According to Roberts (1995) and Oldfield (1998), both, a high SIRM/ κ_{LF} and the absence of a hard remanence component in combination with strong gyroremanences, are typical for authigenic greigite.

The interpretation of samples with low S-ratios is rather difficult. Low S-ratios indicate the presence of a high coercivity component, i.e. grains with remanent coercive forces > 300mT. We tend to interpret it as a signal of hematite, based on thermomagnetic curves (Fig. IV.5B), while the hysteresis loop contradicts the S-ratio. The measured coercive force (B_c) in Figure IV.6A (lowermost loop) is about 11 mT only. According to the ratio B_{cr}/B_c the remanent coercive force would be around 40 mT. Hematite with such small B_c is of multi-domain size and about 1 mm in diameter (Kletetschka and Wasilewski, 2002) and should be remagnetised when applying an opposite field during the S-ratio procedure. Nevertheless, the S-ratio is low and we suggest that the presence of a certain amount of superparamagnetic magnetite/maghemite particles is lowering the coercive force. These superparamagnetic grains would not affect the S-ratio as they are not able to carry a remanent magnetisation. Indeed, the TEM microphotographs in Figure IV.7B, show that grains in the superparamagnetic grain size range are present in the diatomaceous layers. They are probably relics of biogenic magnetite, which has formed at the water/sediment interface, as has been described by Peck and King (1996). In addition, there are also grains containing titanium (Fig. IV.7C), indicating substitution of titanium for iron oxide. This is confirmed by XRD analysis (Fig. IV.8B), which shows a relative abundance of titanite and titanomagnetite while the magnetite content is much less.

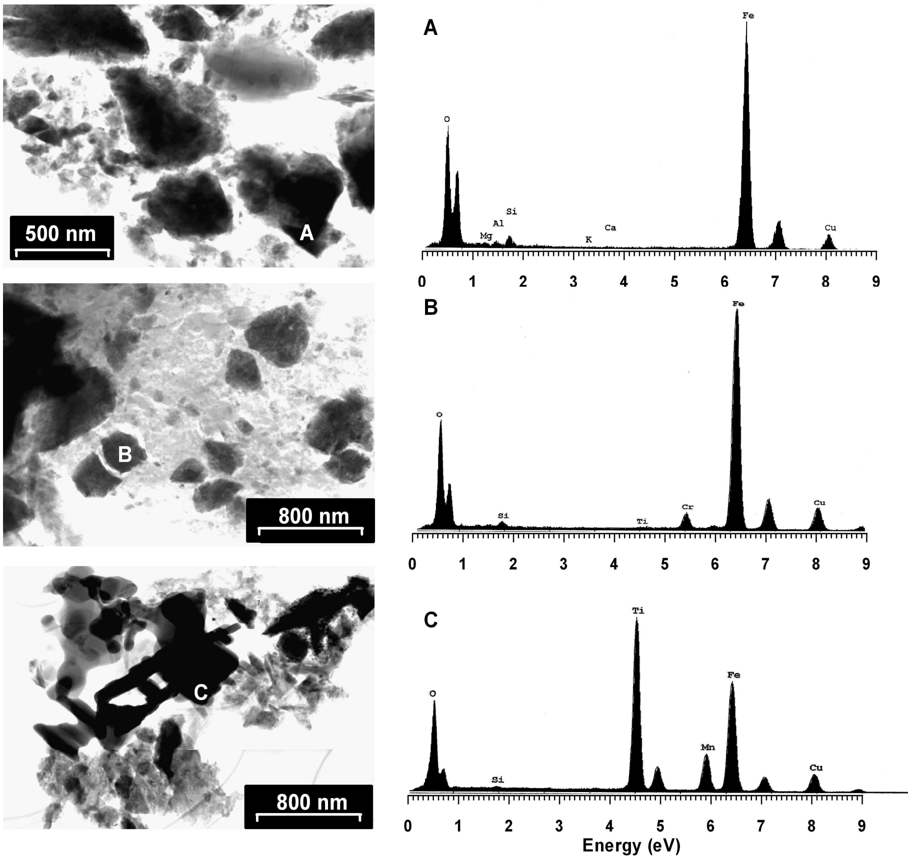


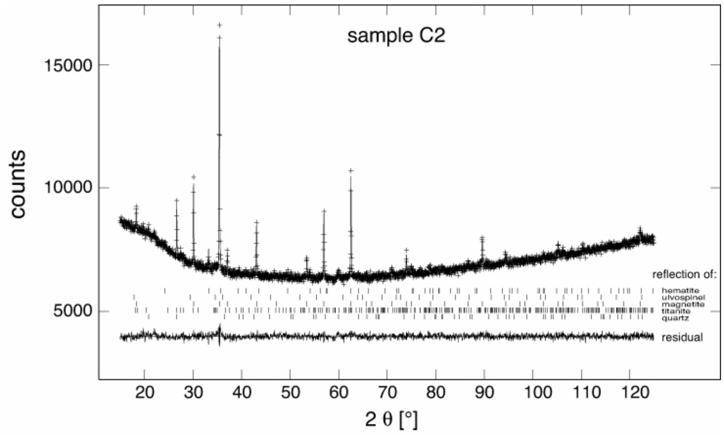
Figure IV.7: Representative transmission electron microphotographs and elemental dispersive spectra of magnetic extract. Oxides of different size (from submicron to micron) are visible on the microphotographs; A. in the clay-rich layer (interval 527.5 to 529.9 cm, core VER 98-1-1), the extract is dominated by pure iron oxides (magnetite); B and C., in the diatomaceous layer (interval 583 to 590 cm, core VER 98-1-1), there is a relative enrichment in titanium and chromium of the oxides.

IV.7. Discussion

The magnetic mineralogy demonstrates that magnetite, probably of detrital origin, dominates the signal in most of the glacial sediments, except in a few intervals characterised by the presence of greigite. The interglacial sediments present different magnetic mineral assemblages with predominance of either magnetite (sometimes with greigite) or hematite.

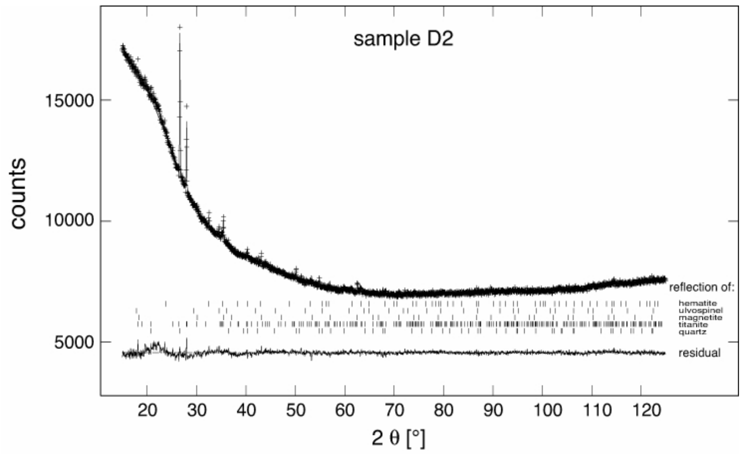
In this section, we compare the magnetic mineralogy with non-magnetic proxies such as water, opal, TOC, and TS wt %, semi-quantitative Fe and Ti contents. These proxies trace the biogenic productivity and the detrital input, but are also indicative of early diagenetic conditions, which control e.g. the preservation of organic matter and the redox conditions, and would help to constrain the processes responsible for modifications in magnetic mineral assemblages.

A



Clay-rich layer	% (weight fraction)
quartz	13
titanite (Al poor)	9
magnetite/chromite	70
titanomagnetite	0.5
hematite	7.5

B



Diatomaceous layer	% (weight fraction)
quartz	38
titanite (Al rich)	29
magnetite/chromite	17
titanomagnetite	6
hematite	10

Figure IV.8: X-ray diffraction patterns and semi-quantitative approach of the different phases for magnetic extracts; A. extract from the clay-rich layer (interval 527.5 to 529.9 cm, core VER 98-1-1), characterised by 70 % of magnetite and by a titanite poor in aluminium; B. extract from diatomaceous layer (interval 583 to 590 cm, core VER 98-1-1), characterised by only 17 % of magnetite and by a titanite rich in aluminium.

IV.7.1. Tracers of the detrital input

In selected intervals, we measured titanium and iron contents in parallel to rock magnetic parameters (Fig. IV.9). Titanium content are a good reflection of detrital input since minerals containing titanium are not very sensitive to dissolution. Iron however is rather mobile and involved in the redox history of highly porous sediments: the spike of iron observed on top of the sedimentary column (Fig. IV.9A) marks the redox front. We observed a strong similarity between the titanium and HIRM curves: the detrital input decreases from the late glacial to the Holocene. In ancient sediments, HIRM and titanium display similar variations with high values in glacials and low values in interglacials (Fig. IV.9B). Notably, no significant HIRM change is observed at the transition between oxidising and reducing conditions in the sediment (Fig. IV.9A). This implies that HIRM is not affected by redox conditions and further confirms that the "hard" magnetic mineral content is the best tracer of detrital input (Peck et al., 1994). On the other hand, the S-ratio seems to be related to the redox conditions in the sediment (see section IV.7.2 below). The ARM has also to be considered with caution as it is mainly influenced by the ferrimagnetic contribution, which is itself influenced by post depositional processes. This is seen in Figure IV.9, where ARM variations are partly influenced by S-ratio variations.

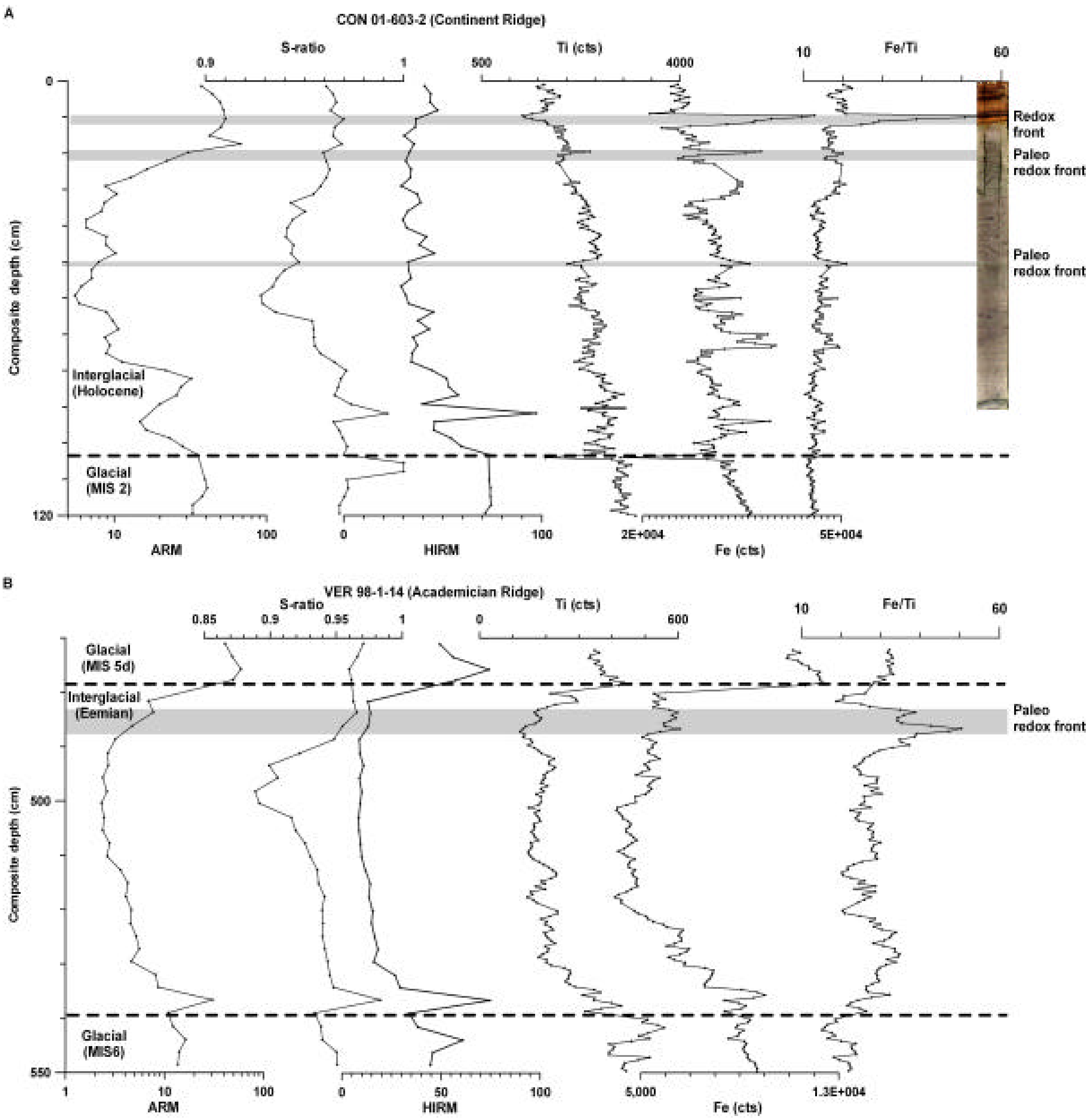


Figure IV.9: Downcore variations of rock magnetic parameters (ARM, S-ratio and HIRM), and XRF titanium, iron counting and ratio Fe/Ti for two different sedimentary sequences, redox and paleo-redox fronts are highlighted in grey; A. section at site CON 01-603-2 covering the Holocene and MIS 2, on the photo of the core the limit oxic/anoxic sediments is well documented; B. section at site VER 98-1-14 covering MIS 5d, Eemian and MIS 6.

IV.7.2. Dissolution of magnetic minerals

The difference by an order of magnitude between ARM values in glacial and interglacial sediments (Fig. IV.3) suggests the dissolution of magnetic minerals in the sediment. The observed difference cannot result only from low detrital input and dilution of the lithogenic component by opal (30 wt %) and pore water (40 wt %) (Fig. IV.3).

The topmost sediments are oxidised since the entire water column of Lake Baikal is oxygen saturated throughout the year (Weiss et al., 1991). High ARMs as well as high S-ratios in this oxidised zone (Fig. IV.9A) are partly due to fine-grained biogenic magnetite (Peck and King, 1996). The redox boundary between the oxic zone at the top and the anoxic zone below is marked by a downward decrease of the ARM and S-ratio values while the HIRM (i.e. the quantity of "hard" magnetic minerals such as hematite) remains constant. The decrease in the S-ratio is, therefore, due to a decrease of "soft" magnetic minerals (magnetite), which is linked to processes in the redox front and results from an in-situ process.

Processes driving selective dissolution of "soft" magnetic particles under suboxic and anoxic conditions in the pore water are still debated. In lakes characterised by sediments rich in organic carbon, hematite has been shown to dissolve first (Williamson et al., 1998). In subarctic lakes (Snowball, 1993) and in Lake Baikal (the present study), the fine grained magnetite dissolves easier than hematite under similar conditions with relatively low organic carbon contents (up to 2 wt %) and low sedimentation rates. Due to high surface/volume ratio, the small grains of bio-magnetite (Peck and King, 1996) and magnetite, which underwent maghemitization under oxic/suboxic conditions, are very sensitive to dissolution processes (Smirnov and Tarduno, 2000). The dissolution of soft magnetic minerals (magnetite) after burial below the redox front is probably mediated by bacteria in the presence of organic carbon. We suggest two types of transformation in interglacial sediments of Lake Baikal. First, the dissolution of magnetic minerals is the main supplier of soluble Fe^{2+} feeding the redox front (Granina et al., 2004). Secondly, a high concentration of dissolved opal in the pore water facilitates the transformation of magnetite into smectite (Florindo et al., 2003). This would corroborate the occurrence of a secondary smectite in Lake Baikal sediments (Fagel et al., 2003).

The dissolution of soft magnetic minerals is observed in recent sediments but also in other interglacial sediments such as the Kazantsevo (equivalent to the Eemian in Europe) (Figure IV.9B). We observed low S-ratio below a peak of iron over titanium interpreted as a paleo-redox front. Therefore dissolution of soft magnetic minerals is a feature also observed in ancient interglacial sediments and may result from similar processes.

Nevertheless this dissolution is only observed at the end of some interglacial intervals. Dissolution events do not trace a basin-wide signal in Lake Baikal but result from local phenomena since they cannot be correlated from one core to another core. Furthermore, they are observed just below preserved paleo-redox fronts, as highlighted by spikes of iron over titanium (Fig. IV.9). The paleo-redox fronts are themselves observed just below zones of condensed sedimentation rates. Indeed, several studies have demonstrated a strong relationship between a decrease of the sedimentation rates and preservation of paleo-redox fronts (Deike et al., 1997; Granina et al., 2000; Granina et al., 2004). In the present study, there are further pieces of evidence for changes in the sedimentation rates. According to *AMS* ^{14}C measurements (Piotrowska et al. in press) performed on a Kasten core taken at site CON 01-603-2 and transferred to the present study by susceptibility correlation (Demory et al., submitted-b), a general decrease of the sedimentation rates was observed in the topmost section of the sedimentary column: ~ 10 cm ky^{-1} between 98 and 80 cm depth and ~ 4.5 cm ky^{-1} between 25 and 2 cm depth. This decrease is likely to be underestimated since we have not considered the sediment compaction processes. Moreover, abrupt changes between interglacial and glacial sediments may be hiatuses, as in core VER 98-1-14, where we found many events of soft magnetic mineral dissolution and sediment discontinuities (Fig. IV.3).

Changes in sedimentation rates are also observed in core VER 98-1-1 located on the slope between Academician Ridge and the central basin (Fig. IV.1). Interglacial intervals here are very thick and most probably related to redeposition events (Ceramicola et al., 2002), while glacial intervals have similar extension as determined from other cores at the Academician Ridge (Fig. IV.4). In this core also, the S-ratio decreases strongly on top of interglacial stages (Fig. IV.10).

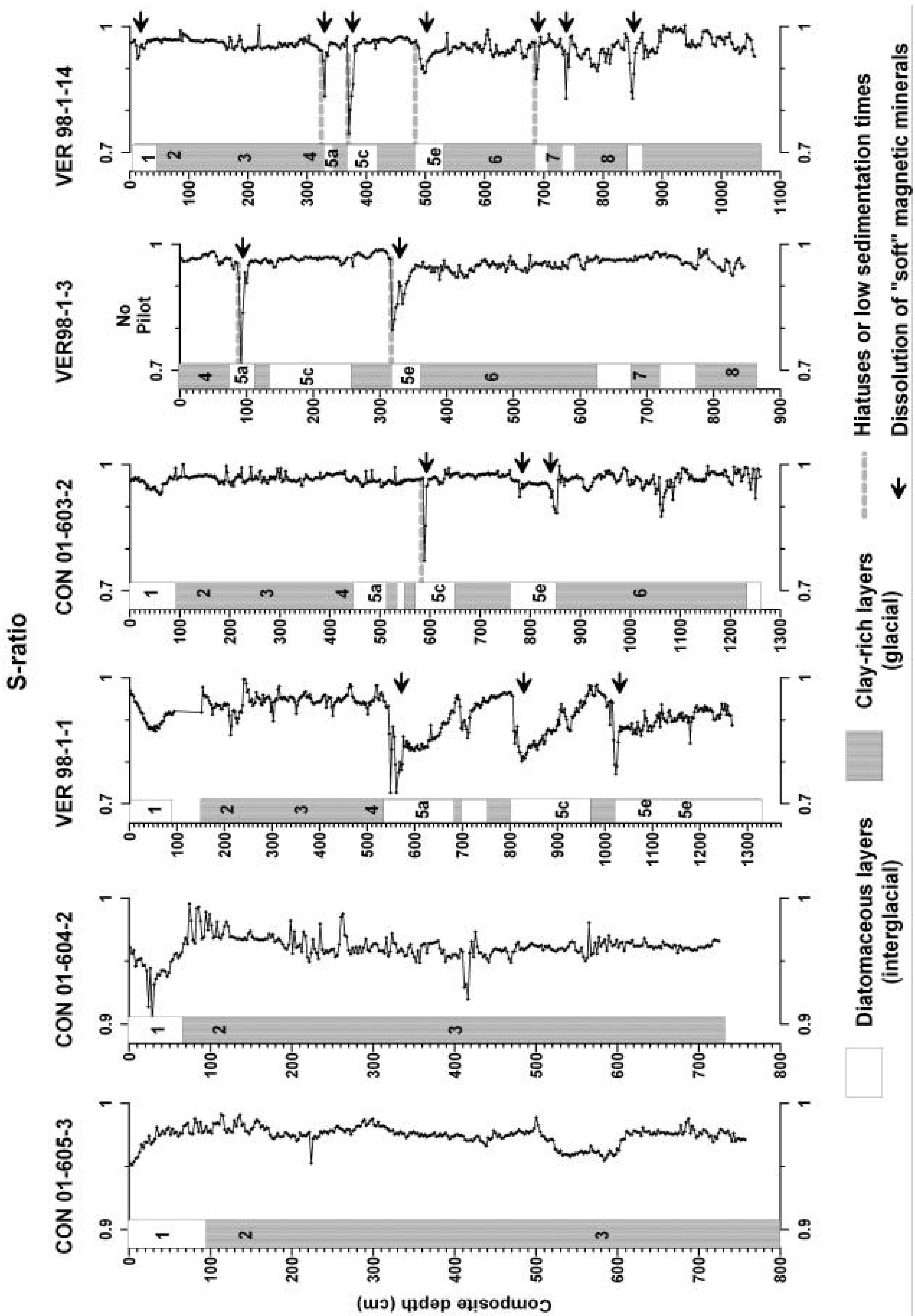

S-ratio
VER 98-1-14
VER98-1-3
No Pilot
CON 01-603-2
VER 98-1-1
CON 01-604-2
CON 01-605-3
Composite depth (cm)
Diatomaceous layers (interglacial)
Clay-rich layers (glacial)
Hiatuses or low sedimentation times
Dissolution of "soft" magnetic minerals

Figure IV.10: Downcore variations of the S-ratio and simplified lithological description. Here MIS are denoted by numbers in the lithological columns. The arrows show spikes of low S-ratios.

Processes responsible for low sedimentation or hiatuses were suggested after high resolution seismic investigations. Ceramicola et al. (2002) observed many events of winnowing which could be responsible for sedimentation rate decay, while local slumping cannot be neglected as a cause for hiatuses. The low cohesivity of the highly porous interglacial sediments would easily lead to a destabilisation of the sediments. A decay of the biogenic productivity (less opal input), which would not be immediately compensated by the glacial type detrital input, could be another reason for the sedimentation decrease at the end of interglacials.

IV.7.3. Occurrence of greigite

Authigenesis of greigite is another characteristic of the sedimentary sequences of Lake Baikal. As summarized by Dekkers et al. (2000), the greigite can form diagenetically after deposition of sediments during bacterial reduction or when hydrothermal fluids are present. Greigite is more likely to grow beyond the redox front during sulphate reduction, although the influence of hydrothermal fluids activated in the tectonic setting of Lake Baikal cannot be ruled out. Increased presence of greigite (high SIRM/κ_{LF}) coincides with maximum sulphur contents observed at the beginning of interglacial stages (Fig. IV.11A). At similar levels in another sediment core of Lake Baikal, Watanabe et al. (2004) observed pyrite mineralization. They attributed these pyrite-rich levels to mineralization at sediment/water interface under anoxic bottom water conditions. However, we prefer to interpret the greigite as a result of magnetite transformation when sulphate reduction occurs in the interglacial sediments. Peak sulphur contents would therefore be due to sulphur mineralisation within the sediment and would not result from an enrichment of the sediment in sulphur at the sediment/water interface.

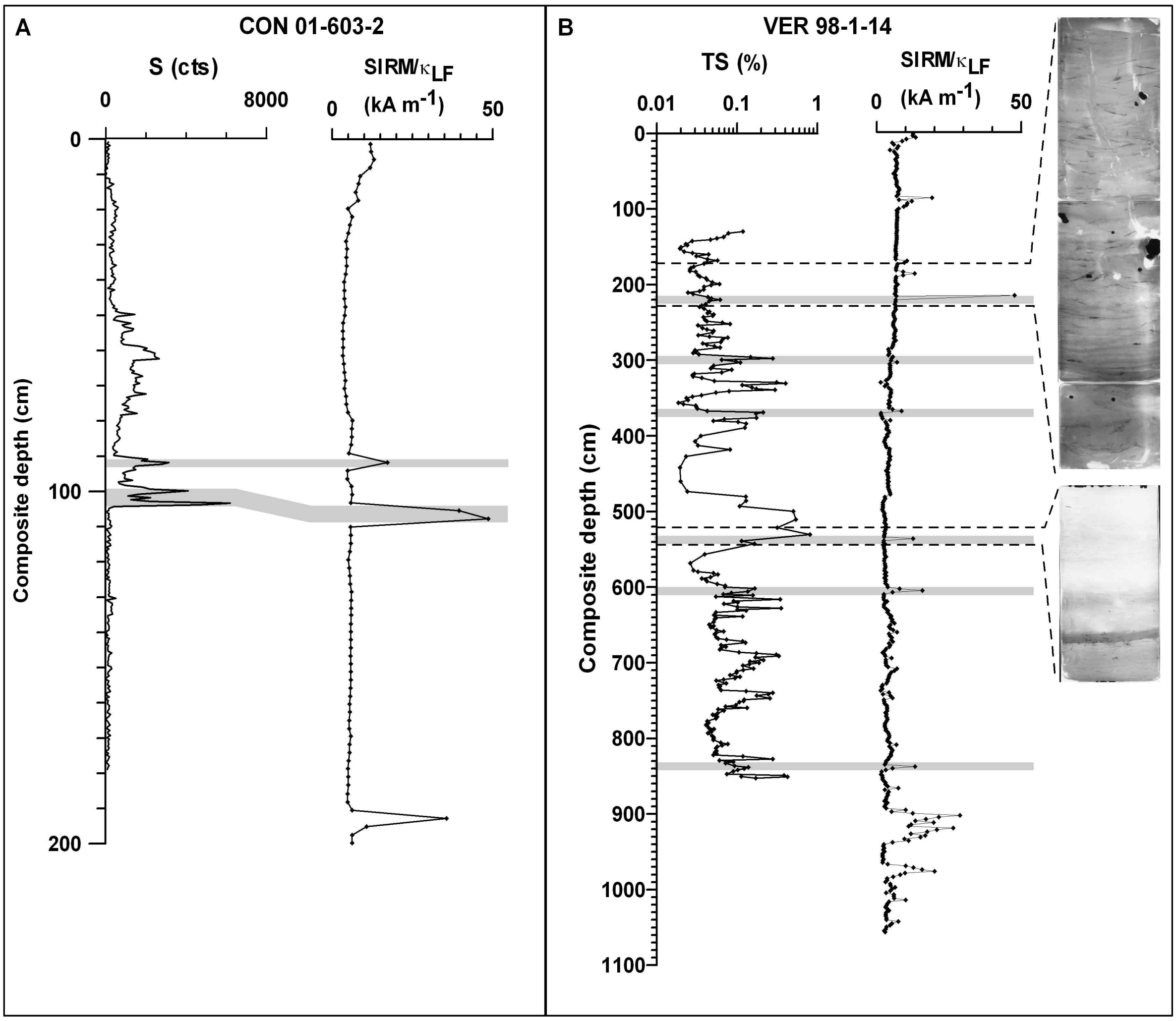
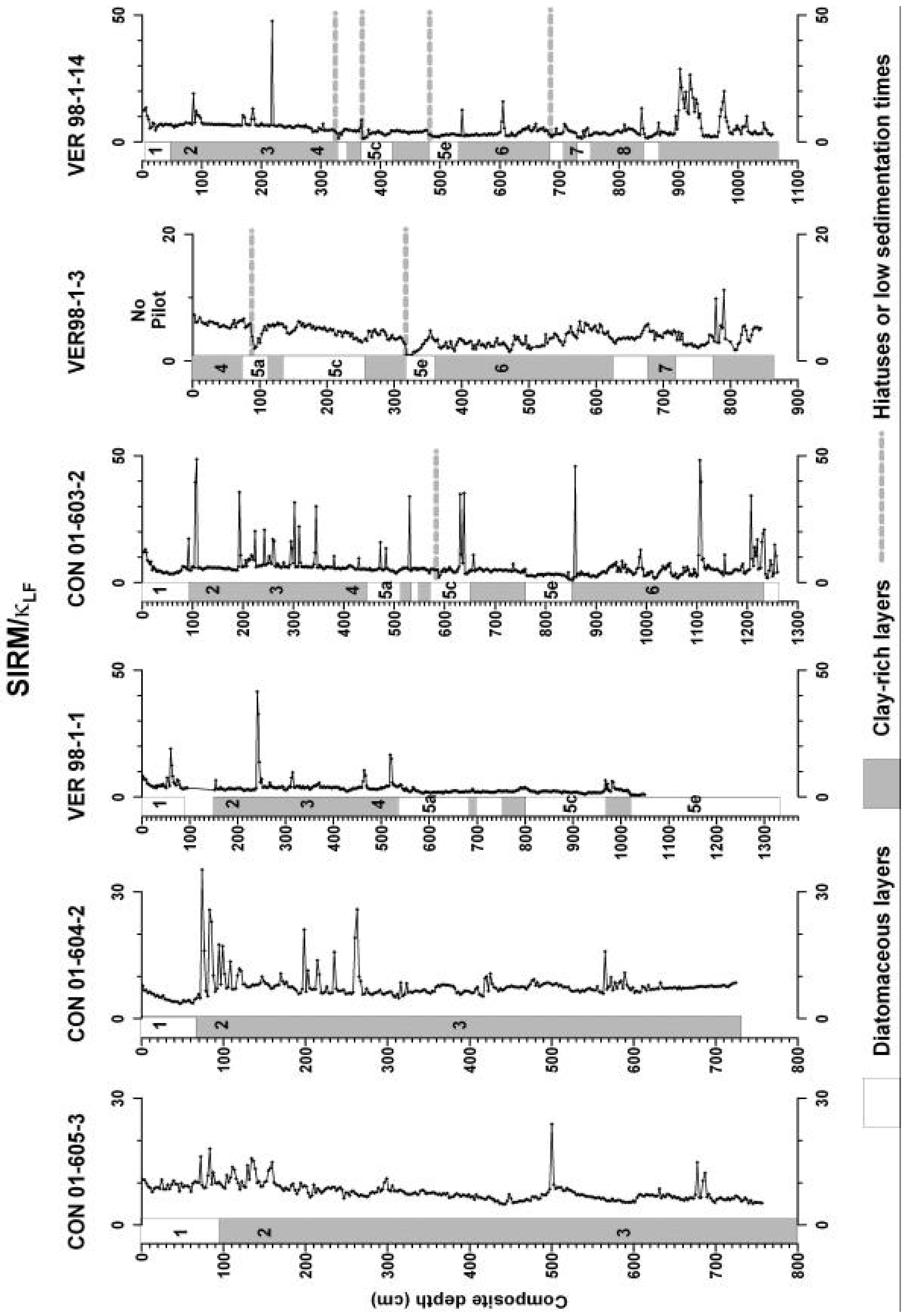


Figure IV.11: Downcore variations of sulphur and SIRM/k_{LF} (high values are considered to indicate greigite); A. topmost part of the sedimentary sequence from site CON 01-603-2, relative sulphur content is estimated by XRF counting; B. Full record for sedimentary sequence from site VER98-1-14, total sulphur content. For A. and B. the grey lines show the correlation between increase of sulphur and presence of greigite. Two intervals are documented by radiographs showing the scattered distribution of the greigite in glacial sediments and the layer rich in sulphur at the transition glacial/interglacial.

Higher abundance of greigite during glacial intervals also coincides with small increases of the S content (Fig. IV.11B). Greigite levels in glacial sediments cannot be correlated between cores (Fig. IV.12), which suggests that greigite concentrations are driven by local processes. We suggest that faecal pellets could be a suitable microenvironment for sulphate reduction. While greigite could potentially act as proxy for faecal pellets in glacial sediments, unfortunately we can not rely on this possible indicator since the greigite is very sensitive to onshore alterations after sampling (Snowball and Thompson, 1990).

Figure IV.12: Downcore variations of the SIRM/k_{LF} (high values are considered to indicate greigite) and simplified lithological description. Here MIS are denoted by numbers in the lithological columns.



IV.8. Conclusion

Continuous rock magnetic measurements, complemented by X-ray diffraction and TEM on magnetic extracts, were performed on cores from Lake Baikal covering several glacial and interglacial stages. Compared to geochemical data, the rock magnetic study showed that:

(i) HIRM, i.e. hematite+goethite content, remains the best estimate for the detrital input in Lake Baikal, whereas parameters estimating the magnetic mineral concentration, like magnetic susceptibility or ARM, are partly influenced by post-depositional processes such as magnetite dissolution.

(ii) Magnetite dominates the magnetic signal in glacial sediments except for a few enhancements of greigite, probably confined to reductive microenvironments.

(iii) Magnetic assemblages vary in interglacial sediments with predominance of magnetite ± greigite at the beginning of interglacial intervals and a predominance of hematite at the end of some of the interglacial intervals. For interglacial sediments, we propose a model of the magnetic signal involving mainly in situ processes. Due to less wind, the detrital input in Lake Baikal is low compared to the glacial, and therefore there are less detrital magnetic minerals as observed in the HIRM curves (Fig. IV.9). During interglacial stages the biogenic productivity is high and the bottom water of Lake Baikal is known to be well oxygenated. In the oxidised bottom sediments, a magnetite is generated by magnetotactic bacteria (Peck and King, 1996). After burial, the biomagnetite is passed by the redox front (rich in iron) and then stays in a suboxic to anoxic environment where organic carbon is preserved. In the case of a constant sedimentation rate, the burial is quite fast. Consequently, the magnetite could be preserved or transformed into greigite once sulphate reduction occurs. This scenario is proposed for the beginning of the interglacials. A second scenario occurs when the sedimentation rate decreases. In this case, magnetite stays longer just below the redox front. This allows a slow dissolution of magnetite and a significant precipitation of iron at the redox front. This is expressed by the preservation of a paleo-redox front as observed in the present study as well as in previous studies (e.g. Deike et al., 1997; Granina et al., 2004) and by low S-ratios, showing the preferential dissolution of soft magnetic minerals.

Acknowledgments

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V. High-resolution magnetostratigraphy of late Quaternary sediments from Lake Baikal, Siberia: timing of intracontinental paleoclimatic responses.

F. Demory, N. R. Nowaczyk, A. Witt and H. Oberhänsli

Submitted to Global and Planetary Change

Abstract

Sediment cores retrieved from 6 locations in Lake Baikal were subjected to a paleomagnetic study in order to establish detailed age models based on correlations of relative paleointensity records. Additional data were provided by calibrated accelerator mass spectrometry (*AMS*) ^{14}C dating, as well as by documentation of geomagnetic excursions like Laschamp at ~ 42 ka and Iceland Basin at ~ 185 ka. Few intervals were affected by diagenetic features like selective reductive dissolution of magnetite and greigite mineralization (Demory et al., submitted-a), and those that were, were left out of paleointensity records. These records were tuned to the well-dated paleomagnetic record from ODP Site 984 (Channell, 1999). The complex shape of the resulting depth/age curves highlights the need for a high resolution age model. We focused on the climatic boundary between marine isotopic stage (MIS) 7 and 6 where the Iceland Basin paleomagnetic excursion is clearly documented in the North Atlantic (Channell et al., 1997) and in Lake Baikal (Oda et al., 2002; present study). During this period, we provide evidence for a return to cold conditions in the Lake Baikal region simultaneous to the sea surface cooling, but earlier than the global ice volume change observed in North Atlantic planktonic and benthic $\delta^{18}O$ records, respectively. The classical strategy of age model reconstruction, based on direct correlation of the climatic record from Lake Baikal sediments with the marine $\delta^{18}O$ reference curves is shown here to be unreliable. Moreover, this strategy does not consider (i) the non linearity of the age model in Lake Baikal sediments and (ii) the time lags between the global ice volume change and sea surface cooling observed in $\delta^{18}O$ marine records. Finally, the "Baikal 200" compilation of the paleointensity records established in this study provides a 200 ka-long synthetic paleomagnetic record for Central Asia.

Keywords: Late Quaternary sediments - Lake Baikal - paleomagnetism - relative paleointensity - geomagnetic excursions – age model – intracontinental climatic changes

V.1. Introduction

Dating intracontinental climatic records is still a challenge (e.g. Kukla et al., 1997; Forsstrom, 2001) because the correlation with the well documented marine records are still uncertain. In the scope of dating Late Quaternary sediment studies, magnetostratigraphy retains its position as a powerful stratigraphic tool. This is mainly due to the new technique of correlating high-resolution records of relative geomagnetic paleointensity variations. Reference curves now available from nearly all regions of the earth indicate that most of these variations can be considered to constitute global signals. Thus, despite some dating discrepancies, good quality records of relative paleointensity show a reproducible pattern, which allows the establishment of global stacks (Guyodo and Valet, 1996) and furnishes a correlation tool at the global scale (Channell et al., 2000). In addition, the recent cosmogenic nuclide production records recently produced for the last 300 ka further corroborate the paleointensity variations (Thouveny et al., 2004; Carcaillet et al., 2004). The quality of a relative paleointensity record depends on the type, amount, quality, and preservation of the carriers of the remanent magnetisation. They ideally should be of a low coercive and fine-grained mineral (such as magnetite) of sufficient concentration with only moderate variation (King et al., 1983; Tauxe, 1993). In this respect, most of Lake Baikal sediments are suitable for the reconstruction of high-quality relative paleointensity records (Peck et al., 1996).

In Late Quaternary times, several full reversals of the paleomagnetic field were suggested, mostly from Arctic marine records (Nowaczyk et al., 1994; Nowaczyk and Antonow, 1997; Nowaczyk and Frederichs, 1999). However, they are rarely found in sediments from other regions. This could be either due to a significant non-dipole contribution of the field during excursions (Langereis, 1999) or due to peculiar redox conditions affecting sediments (notably lake sediments) erasing the primary paleomagnetic information. Indeed, mono-domain ferromagnetic minerals, carrying the paleomagnetic directional variation, are sometimes badly preserved in lake sediments (Snowball, 1994). Although geomagnetic excursions are not frequently documented in most studied sedimentary sequences, recent compilations (Nowaczyk et al., 1994; Langereis et al., 1997; Thouveny et al., 2004) and detailed studies focussing on given excursions (e.g. Nowaczyk et al., 2003) yield a clearer image about the timing and nature of their occurrence. Therefore, if documented, paleomagnetic excursions can provide important tie points in order to anchor records of relative paleointensity. In Lake Baikal, a full reversal of the geomagnetic dipole is documented (Hayashida and Yokoyama, 1995; Oda et al., 2002) and is attributed to the Iceland Basin event dated at ~ 185 ka in marine sediments (Channell et al., 1997). Therefore, Lake Baikal sediments can be utilised as an important intra-continental paleomagnetic archive.

In this study we present new data on this excursion and other directional features, while their relationship to relative paleointensity variations is discussed. In addition, we outline the role of diagenetic processes in terms of preservation of geomagnetic excursions and of the fidelity of relative paleointensity variations. Emphasis is placed on the Lake Baikal sedimentary reaction to global climatic changes using high-resolution age models ensuing from paleomagnetic correlation and rock magnetic parameters. Finally, the compilation of the different relative paleointensity records allows us here to establish a new, synthetic reference curve for Central Asia.

V.2. Material and methods

The intra-continental Lake Baikal (eastern Siberia) is located in a rift valley, from 51°N to 56°N and from 104°E to 110°E (Fig. V.1). It consists of three deep basins separated by two underwater highs. The lake has accumulated up to 7500 m of sediments since at least the Miocene (Hutchinson et al., 1992). In this study, six sites were cored on geomorphologic highs, on margins of the basins (Fig. V.1), far from turbiditic influences (e.g. see Charlet et al. submitted). The coring was performed in 1998 and 2001, yielding a total of six piston and five pilot cores from the Academician Ridge, the Continent Ridge, the Posolsky Bank, and the Vydrino Shoulder (Table V.1, Fig. V.1). The sedimentary sequences are composed of thick clay-rich (up to 80 wt %) greenish layers alternating with thick diatomaceous (30 wt % of opal, (Fagel et al., 2003; Demory et al., submitted-a) olive colour layers. The pore-water content is high in the diatomaceous layers (up to 60 wt %) and low in the clay rich layers (30 wt % (Fagel et al., 2003; Demory et al., submitted-a). Due to the coring techniques employed, sediments from the top of the piston cores were lost, and thus sedimentary sequences were completed by taking a combination of pilot and piston cores (Table V.2).

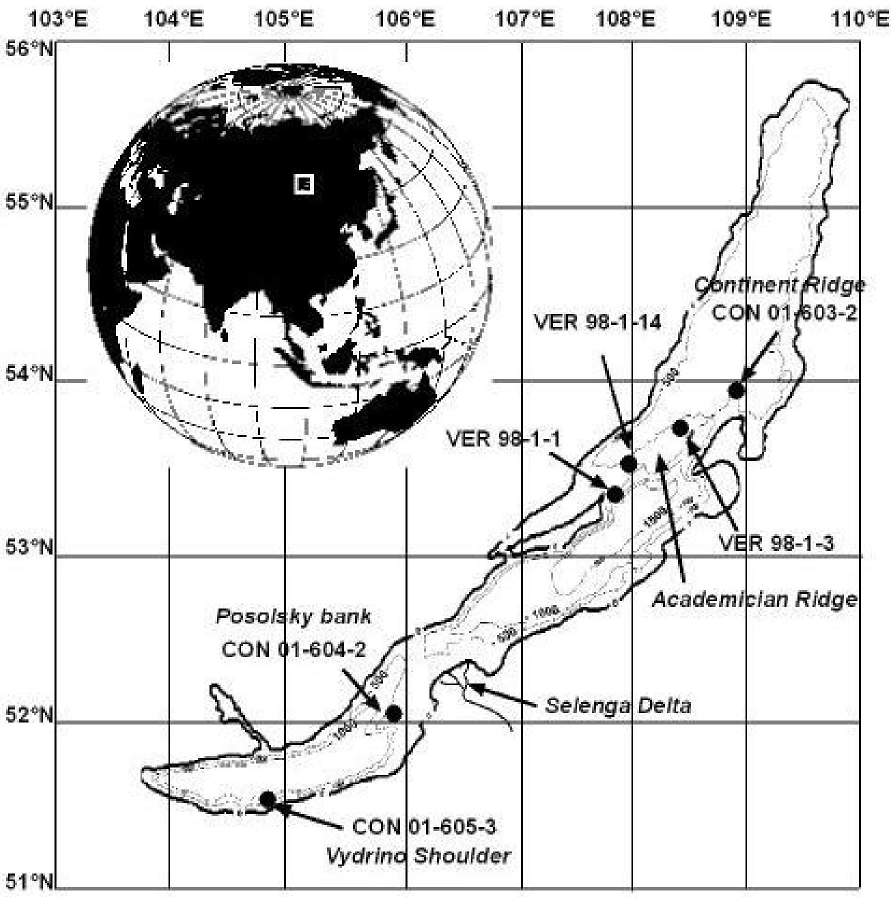


Figure V.1. Simplified bathymetric map of Lake Baikal showing the coring locations. From each site a piston core (~ 10 m long) and a pilot core (~ 2 meters long, when available) were investigated.

Core name and type	Location	Water depth (m)	Latitude	Longitude	Length (cm)	Number of samples
VER 98-1-14 Piston	Academician Ridge	412	53°31'23''N	107°58'10''E	983	383
VER 98-1-14a Pilot	Academician Ridge	412	53°31'23''N	107°58'10''E	192	80
VER 98-1-3 Piston	Academician Ridge	373	53°44'56''N	108°19'02''E	831	331
VER 98-1-1 Piston	Academician Ridge	245	53°23'36''N	107°55'22''E	1120	470
VER 98-1-1a Pilot	Academician Ridge	245	53°23'36''N	107°55'22''E	100	41
CON 01-603-2 Piston	Continent Ridge	386	53°57'48''N	108°54'47''E	1127	479
CON 01-603-2a Pilot	Continent Ridge	386	53°57'48''N	108°54'47''E	190	82
CON 01-604-2 Piston	Posolsky Bank	133	52°04'46''N	105°51'27''E	622	268
CON 01-604-2a Pilot	Posolsky Bank	133	52°04'46''N	105°51'27''E	188	82
CON 01-605-3 Piston	Vydrino Shoulder	675	51°35'06''N	104°51'17''E	1052	453
CON 01-605-3a Pilot	Vydrino Shoulder	675	51°35'06''N	104°51'17''E	173	74
					Total:	2743

Table V.1. Core number and type, location, water depth, latitude and longitude, length, and number of samples of the sediment cores used in this study. In the following Figures, when the type of core is not mentioned, the name corresponds to the pilot-piston core composite.

Site	Overlap (cm)	Depth correction for the piston (cm)
VER 98-1-14	108.1	84.7
VER 98-1-3	No	No
VER 98-1-1	No	150
CON 01-603-2	56.6	133.9
CON 01-604-2	84.0	104.5
CON 01-605-3	39.9	133.3

Table V.2. Overlap between pilot and piston cores and depth correction for the piston cores. It should be noted that, in order to minimise core compression, data from the lowermost part of the pilot core has been preferred for the overlapping section. No pilot was retrieved from site VER 98-1-3 and no overlap was observed at site VER 98-1-1.

High-resolution records of magnetic susceptibility were obtained from the topmost of the cores using a Bartington MS2E sensor mounted to an automatic logging device (Nowaczyk, 2001). Measurements were performed every 1 mm on 1 cm thick slabs taken for x-ray radiographies.

The cores were quasi-continuously sampled with 6.2 cm^3 plastic boxes pushed into the split surface, side by side every ~ 2.5 cm, yielding a total of 2743 samples (Table V.1). All samples were subjected to stepwise alternating field demagnetisation and detailed rock magnetic analyses. Measurements of magnetic low field bulk susceptibility (κ_{LF}) and anisotropy of magnetic susceptibility (AMS) were performed using a Kappabridge KLY-3S (AGICO Brno, Czech Rep.). AMS was used to determine the orientation of the magnetic fabric. Shape and orientation of the anisotropy ellipsoid was used to detect whether disturbances were from the coring process or from natural origin (e.g. Rees, 1965; Demory et al., submitted-a).

Measurements and stepwise demagnetisation of natural remanent magnetisation (NRM, up to 100 mT) and anhysteretic remanent magnetisation (ARM up 65 mT) were performed using a fully automated 2G Enterprises DC-SQUID 755 SRM long-core system with an in-line tri-axial degausser. NRM and ARM demagnetisations were processed at the same alternative field levels.

The directions of the characteristic remanent magnetisation (ChRM) were determined using principal components analysis on the demagnetisation results of the NRM (Kirschvink, 1980). The ARM was produced along the z-axis of the samples with 50 µT static field and 100 mT alternating field amplitude, using a 2G Enterprises 600 single-axis demagnetiser with an in-line ARM coil. Isothermal remanent magnetisations (IRM) were imprinted with a 2G Enterprise 660 pulse magnetizer and measured with a Molyneux MiniSpin fluxgate magnetometer. The samples were exposed to a field of 2 T, which is sufficient for reaching saturation. The IRM acquired at 2 T is therefore defined as the saturation isothermal remanent magnetisation (SIRM). After applying this maximum field, all the samples were magnetised in a field of 0.3 T in the opposite direction. The fraction of high-coercivity magnetic minerals was estimated by calculating the ratio:

$$\text{S-ratio} = \frac{1}{2}\left(1 - \frac{\text{IRM}_{-0.3\text{T}}}{\text{SIRM}_{2\text{T}}}\right)$$

A S-ratio = 1 is obtained when only low-coercivity particles occur. S-ratio decreases to (nearly) zero with increasing proportion of high coercivity particles like hematite or goethite (Bloemendal et al., 1992). In the present study, the S-ratio is used to detect intervals affected by selective reductive dissolution of magnetite (Demory et al., submitted-a).

In addition, we calculated the hard IRM (HIRM) intensity:

$$\text{HIRM} = (\text{SIRM}_{2\text{T}} + \text{IRM}_{-0.3\text{T}})/2$$

This parameter estimates the quantity of high coercivity magnetic minerals (hematite/goethite), minerals that are little influenced by early diagenesis in Lake Baikal

(Demory et al., submitted-a). HIRM was successfully used as an estimator of the dust input in North Atlantic (Thouveny et al., 2000) and in Antarctica (Maher and Dennis, 2001). According to Peck et al. (1994) and Demory et al. (submitted-a), HIRM is a climatic proxy, revealing the detrital (perhaps eolian) input in Lake Baikal sediments.

The ratio of SIRM intensity to low field susceptibility (SIRM/κ_{LF}) was calculated in order to monitor the possible presence of greigite (an authigenic iron sulphide) already evidenced in Lake Baikal sediments (Demory et al., submitted-a). A high SIRM/κ_{LF} ratio can be indicative for greigite (Roberts, 1995). Since greigite has a coercivity force between 44.8 mT and 94.8 mT (Roberts, 1995), we used the ratio (ARM_{50mT}-ARM_{65mT})/ARM_{0mT} for estimating strong losses of remanence at intermediate demagnetisation levels. In addition, a remanence component acquired perpendicular to the applied ARM field direction can be observed. This phenomenon is due to the magnetocrystalline anisotropy of greigite, which induces a deviation of the ARM vector from the applied field direction (parallel to the z-axis). The presence of greigite can be simply monitored by calculation of the ARM inclination (within sample coordinates).

V.3. Initial chronology

The topmost part (~ 1 m) of the investigated sedimentary column was dated by accelerator mass spectrometry (*AMS*) ^{14}C measurements performed on pollen (Piotrovska et al., in press). The pollen were extracted from kasten cores retrieved from sites CON 01-603 and CON 01-605 (Fig. V.1). Conventional ages were calibrated (Calibration range for 95.4 % probability) using the OxCal programme (Ramsey, 1995) and the atmospheric ^{14}C data from (Stuiver et al., 1998). As the present paleomagnetic study was performed on piston cores and their pilot cores only, correlation of high-resolution magnetic susceptibility logs was used in order to transfer *AMS* ^{14}C dating from kasten to pilot cores (Fig. V.2). However, unexpected positive ages were systematically observed at the sediment surface. Positive ages are attributed to a loss of sediments on top of the cores due mainly to the coring technique.

In a first approximation, the recognition of the global climatic variations reflected in physical properties of Lake Baikal sediments was used to estimate the time span covered by the cores investigated in the present study, as classically done in Lake Baikal (Colman et al., 1995; Williams et al., 1997). Clay-rich sediments represent glacial stages whereas diatomaceous sediments are characteristic of interglacial stages. In parallel, the concentration in ferromagnetic particles is high in the clay-rich (lithogenic) layers and low in the diatomaceous (biogenically dominated) layers. The low values are due to dilution of the lithogenic material (more biogenic materiel and less detrital input) and to dissolution of magnetic particles (Demory et al., submitted-a). Up to date, low field magnetic susceptibility (κ_{LF}) has been the most used proxy for core correlation in Lake Baikal (e.g. Sakai et al., 2000). In the present study, we preferred to use the ARM intensity. Indeed the ARM is only influenced by the contribution of the ferromagnetic particles and provides much more detail than magnetic susceptibility, which is smoothed or has even negative values when biogenic silica (diamagnetic substance) dominates, as in interglacial sediments. Therefore, ARM intensities were correlated to climatic records such as $\delta^{18}O$ variations measured in marine sediment records (Fig. V.3) in order to establish a preliminary age model of medium-resolution. However, this approximate and conceptual age model considers neither hiatuses of sedimentation nor a possible time lag between global climatic transitions recorded in marine and lacustrine sediments.

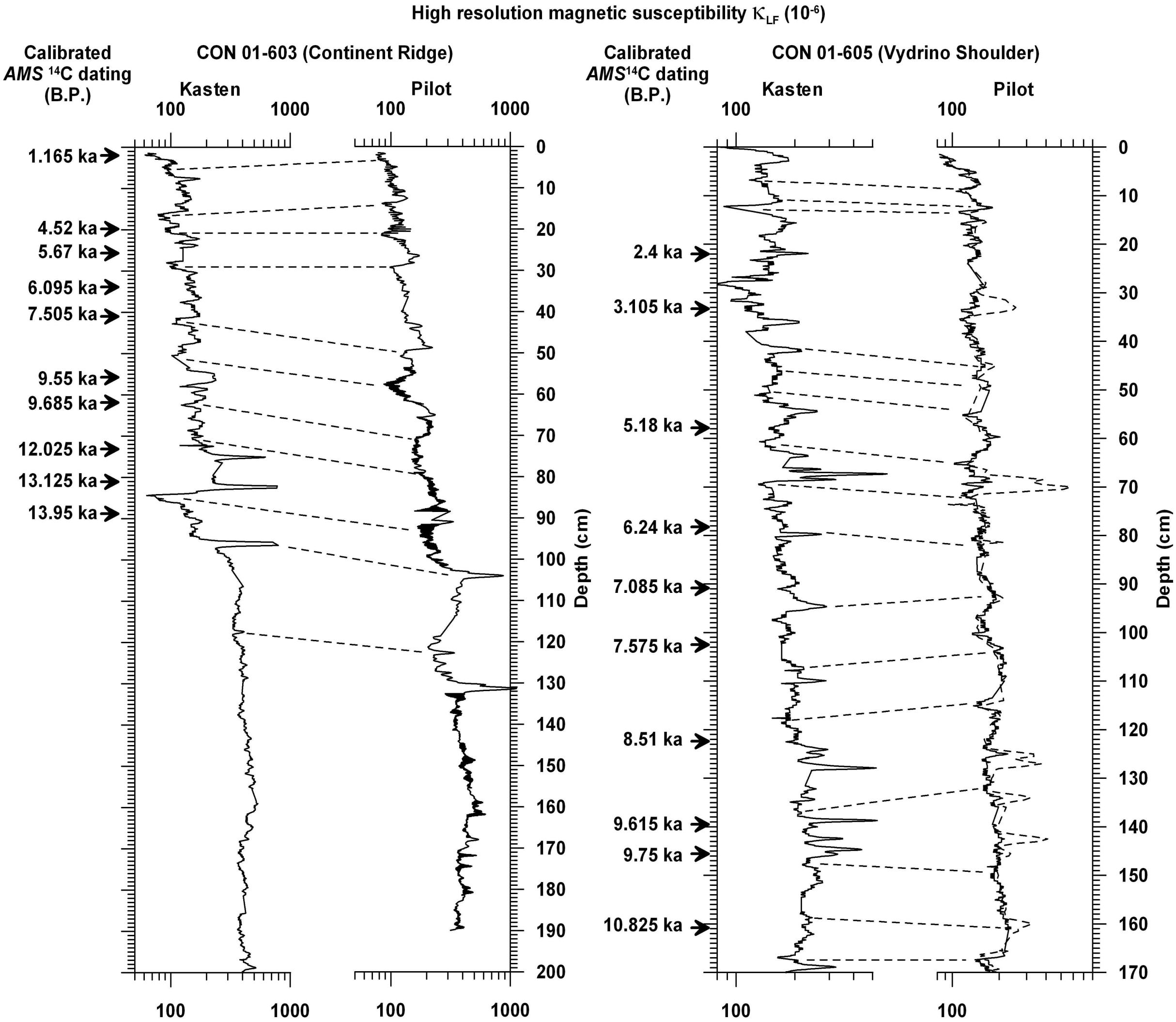
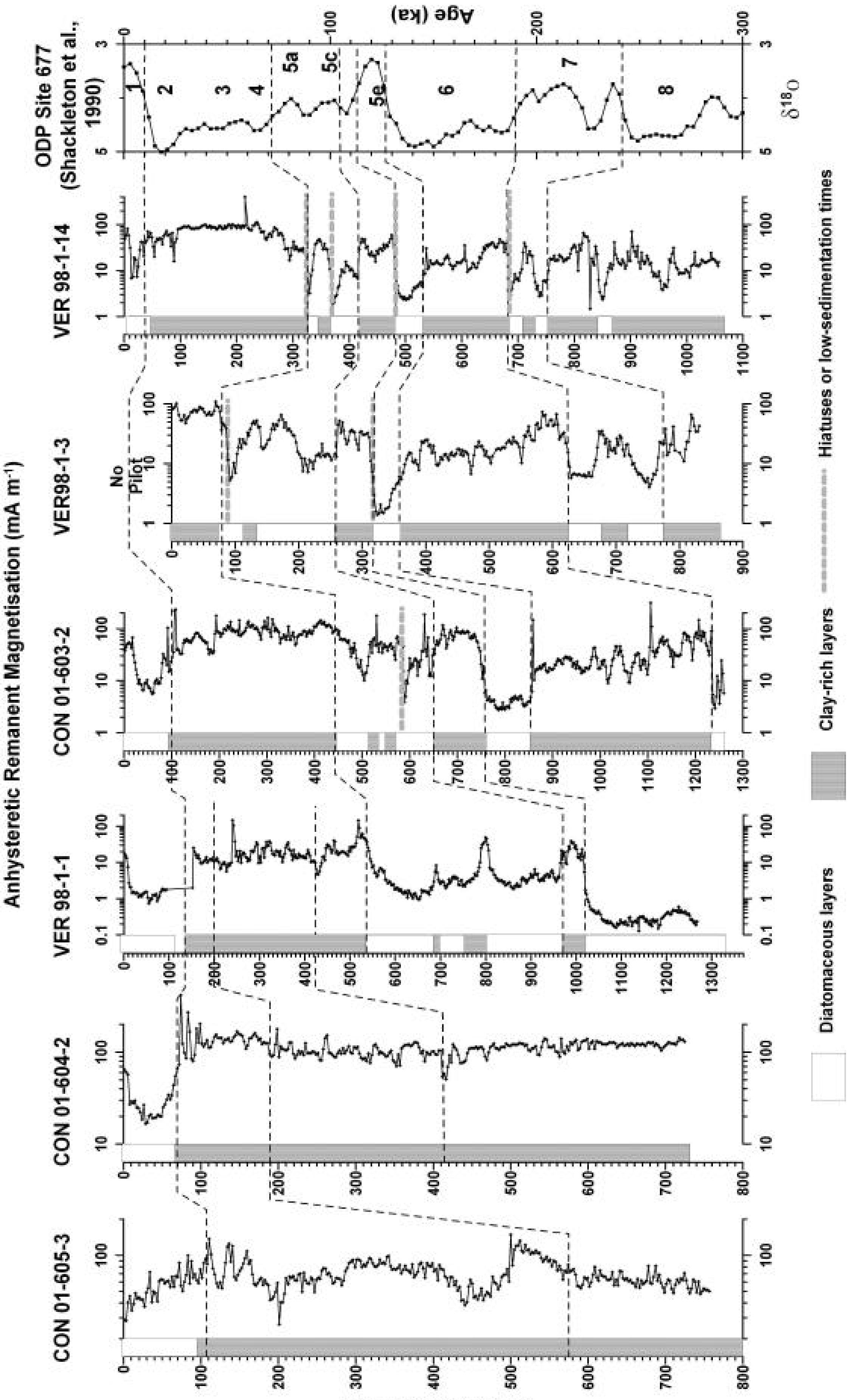


Figure V.2. Inter-correlation of down-core variations of the high-resolution magnetic susceptibility (0.1 cm steps) obtained for kasten and pilot cores from sites CON 01-603 and CON 01-605 (left). This allowed the transfer of AMS ^{14}C dating performed on kasten cores to the pilot cores subjected to paleomagnetic investigations. Note the discrepancies in the magnetic susceptibility curves from the pilot core from site CON01-605 (Vydrino Shoulder) measured in 2001 and 2003, respectively (right). Several large peaks visible in the first measurement from 2001 (dashed lines) disappeared after a two-year-long storage. This is a first hint for the presence the ferromagnetic, chemically unstable greigite.

Figure V.3. Down-core variations of anhysteretic remanent magnetisation and simplified lithological description of the investigated cores. Conceptual inter-correlation between ARM and simplified lithology, and correlation to the $d^{18}O$ record from ODP Site 677 (Shackleton et al., 1990), assuming that clay-rich layers with high magnetic concentration represent glacial periods, and that diatomaceous layers with low magnetic concentration represent interglacial periods. Numbers in the $d^{18}O$ record represent marine isotope stages (MIS).


ODP Site 677 (Shackleton et al., 1990)
Age (ka)
$\delta^{18}O$
1
2
3
4
5a
5c
5e
6
7
8
VER 98-1-14
VER98-1-3
No Pilot
CON 01-603-2
VER 98-1-1
CON 01-604-2
CON 01-605-3
Anhysteretic Remanent Magnetisation (mA m-1)
Composite depth (cm)
Diatomaceous layers
Clay-rich layers
Hiatuses or low-sedimentation times

V.4. Results

V.4.1. Rock magnetism

A detailed discussion on rock magnetic analyses is given in Demory et al. (submitted-a). Here, they are presented only briefly. The deformations in the unlaminated sediments of Lake Baikal are hardly be detected visually. Therefore, the quality of the cores was checked with AMS characterised by the lengths and orientation angles of the three principal axes of the anisotropy ellipsoid. In most of the cores, a flat lying oblate ellipsoid was observed, which is typical for compacted and undisturbed sediments. According to rock magnetic analyses, 3 main minerals carry the signal: (1) A low coercive magnetite, observed in almost the entire cores; (2) Locally in interglacial sediments, magnetite is dissolved a high coercivity hematite dominates the signal; and (3) Mineralization of greigite was observed at the bottom of interglacial sediments and sporadically distributed in glacial sediments (see SIRM spikes in Fig. V.5).

V.4.2. Establishment of relative paleointensity records

Relative paleointensity records were established after detailed inspection of the NRM demagnetisation pattern, the variations in ferromagnetic concentration and the variation in the rock magnetic assemblages. According to Zijderveld diagrams (Fig. V.4A, V.4B and V.4C), the demagnetisation step with 30 mT was sufficient for removing the viscous overprint from the NRM.

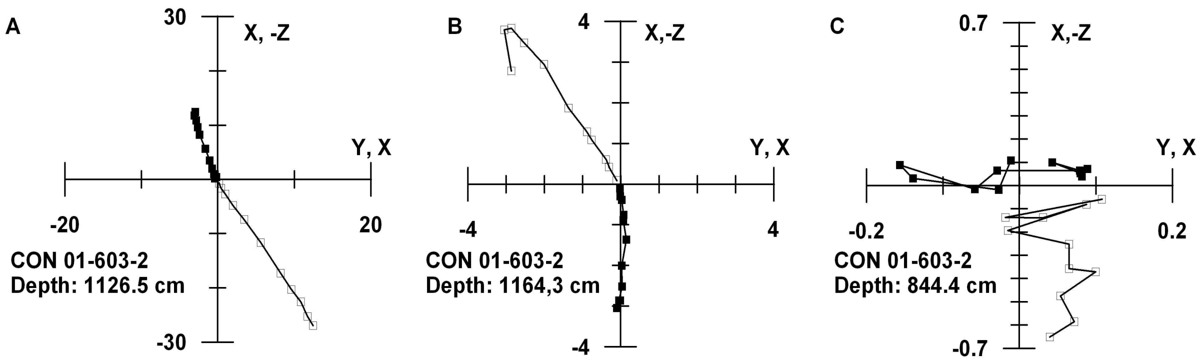


Figure V.4. Three representative vector endpoint diagrams; A. sample from a clay-rich layer, exhibiting normal polarity; B. sample from a clay-rich layer, exhibiting reverse polarity, with the separation of the stable remanence after removal of the normal viscous overprint; C. Instable remanence of a sample from a diatomaceous layer, characterised by a low S-ratio, indicative of reductive magnetite dissolution: see text for details. Axis labelling is in mA m^{-1}.

As shown in e.g. Frank et al., (2002), the dissolution of magnetite alters the quality of paleointensity records. In the present study, the remanent magnetisation is very unstable when the S-ratio is low (Fig. V.4C). Consequently, relative paleointensities determined in intervals where magnetite dissolution is apparent, and in intervals without dissolution are not comparable. Furthermore, the greigite carries a secondary, chemical remanent magnetisation, which competes or obscures the primary paleomagnetic signal (Roberts, 1995). The greigite, visible as spikes in SIRM records (Fig. V.5 outer right), is unstable in sediments of Lake Baikal (Fig. V.2 and Demory et al., submitted-a). Thus, we excluded all samples showing evidence for magnetite dissolution (S-ratios < 0.93) and for greigite mineralization (SIRM/κ_{LF} > 10 kA m^{-1}) from the data sets before establishing relative paleointensity records.

The concentration in ferromagnetic particles is much lower in the interglacial sediments than in glacial sediments. Therefore, the NRM needs to be normalised by a concentration-related parameter such as κ_{LF}, ARM or SIRM, in order to generate relative paleointensity records free of the effects of concentration (Fig. V.5). The three intensity records look quite similar in the topmost part of the sedimentary column, but intensity values are lower in the bottom of the column when using κ_{LF} or SIRM instead of ARM as a concentration parameter. Low values of ARM/SIRM are also observed in the bottom of the sedimentary column while SIRM values remain constant. The discrepancies between the relative paleointensity records result from a relative high amount of coarse magnetic grains (high values of SIRM), which lower the intensity carried by small magnetic particles preferentially contributing to the ARM. Therefore, ARM was preferred as the concentration parameter, and the ratio NRM/ARM was calculated to estimate the variations of relative paleointensity.

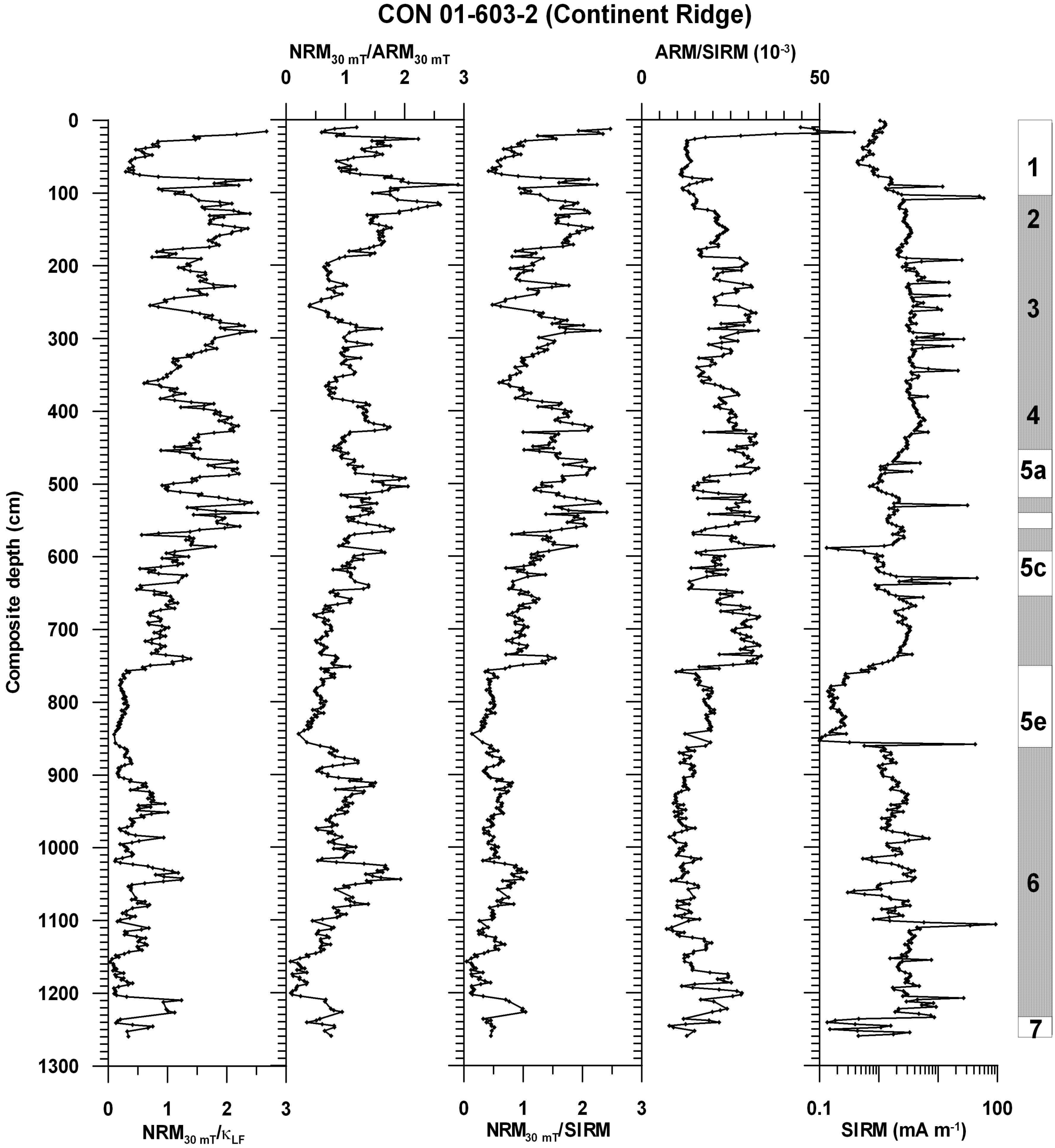


Figure V.5. Three down-core variations of normalised relative paleointensity after diagenetic correction and k_{LF}, ARM and SIRM using as parameters of concentrations. Down-core variations of the ARM/SIRM (also after diagenetic correction) and SIRM for CON 01-603-2. Diatomaceous layers are marked in white and clay-rich layers are marked in light grey. The numbers correspond to the marine isotope stages (MIS).

V.4.3. Directional variations of the geomagnetic field

The paleomagnetic directions, defined by principal components analysis, were rotated in order to adjust the mean declination to zero. Typical demagnetisation behaviours are shown in Figure V.4. Most samples from lithogenic sediments are characterized by a single vector component with a northerly declination and a downward inclination as expected at a northern hemispheric site. In some parts, a weak and soft overprint is visible (Fig. V.4A). A geomagnetic excursion is evident in the composite core CON 01-603-2 at 1165 cm. Here, a reversed direction with a southerly declination and an upward inclination, superimposed by a weak overprint parallel to the present day field direction of normal polarity, is observed (Fig. V.4B). Samples from diatomaceous sediments are often characterized by a weak and unstable magnetisation (Fig. V.4C) so that no ChRM direction could be determined.

In order to condense the paleomagnetic information, the reversal angle was used for monitoring directional anomalies of the geomagnetic dipole, in addition to ChRM inclinations and declinations. The reversal angle is defined as the solid angle between the determined ChRM direction of a sample and the field direction of a geocentric axial dipole of normal polarity at the coring site, i.e. 0° (180°) for a pure normal (reversed) dipole direction, ± 20° due to the paleosecular variation. Reversal angles reaching ~ 180° can be observed in several samples between 1150 cm and 1190 cm of core CON 01-603-2 (Fig. V.6A). All samples carrying an intermediate to reversed ChRM direction were partly overprinted by a soft component parallel to the current field direction of normal polarity (Fig. V.4B). Thus, it excludes the possibility of occasionally, misoriented samples. According to the medium-resolution age model, this excursion occurs just after the onset of marine isotope stage (MIS) 6, during a period of low relative paleointensity. A similar paleomagnetic excursion was already witnessed at the same lithostratigraphic level in Lake Baikal by Oda et al. (2002). These authors attributed the excursion to the Iceland Basin Event recorded between 186 and 189 ka in marine sediments (Channell et al., 1997). A second deviation of the reversal angle reaching about 90°, which is at least a clear intermediate direction, is noticed in the middle of MIS 3, in cores VER98-1-1 and VER98-1-14 (Fig. V.6B and V.6C). This deviation also occurs during a period of low relative paleointensities, and could be related to the Laschamp excursion (Bonhommet and Babkine, 1967) reported from other sedimentary sequences (Vlag et al., 1996).

ChRM inclination and declination records of all investigated cores are plotted in Figure V.7, together with the correlation scheme introduced in Figure V.3 and indication of diagenetically affected intervals. Inclination and declination records could provide information on paleosecular variations (periodicity $=10^5$ years and directional variability < 20°, according to Butler (1992)). In the present study, we did not interpret inclination and declination records in terms of paleosecular variations since slight sediment disturbances could produce slight deviations of ChRM declinations and inclinations.

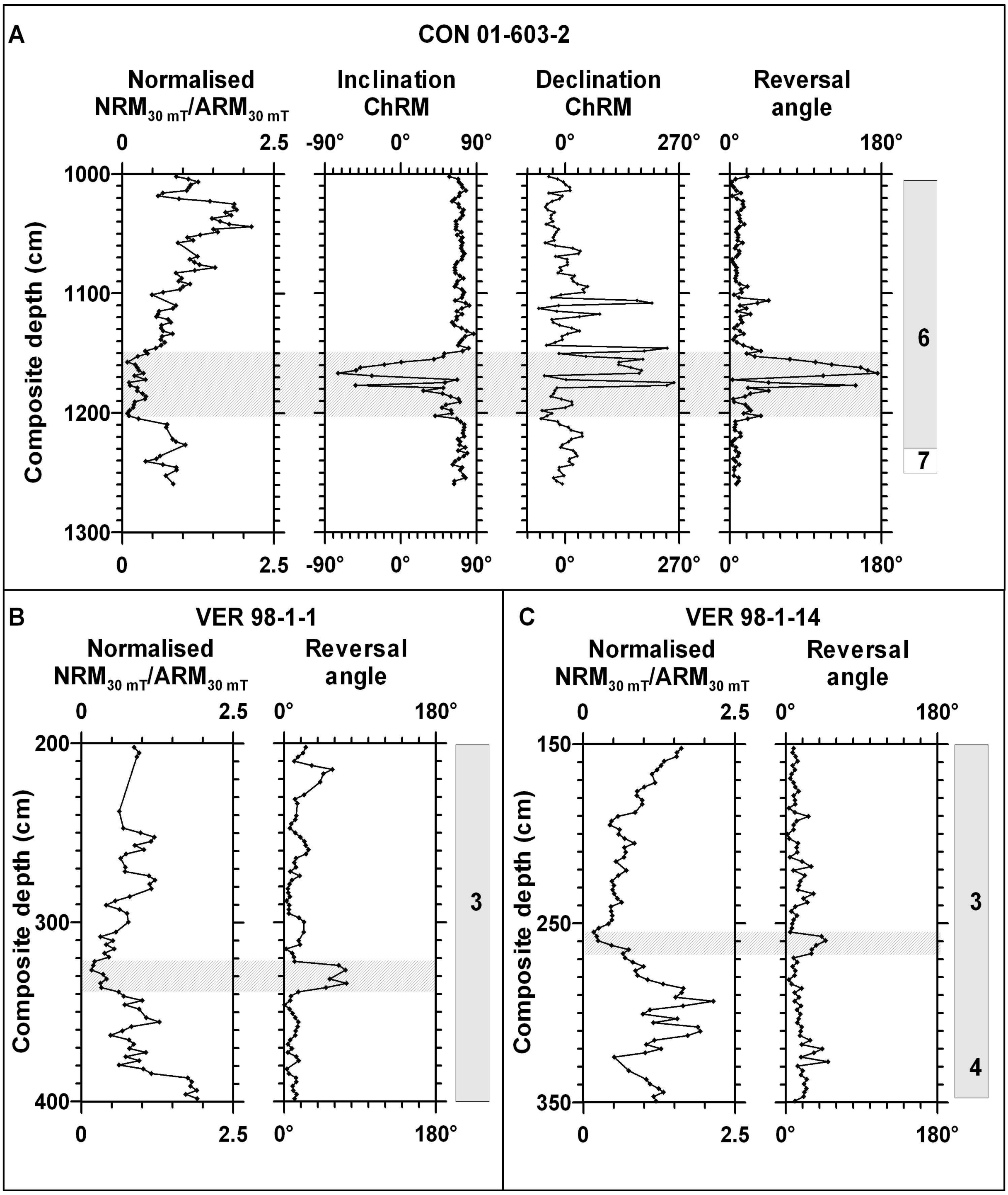


Figure V.6. Selected intervals of down-core variations of normalised relative paleointensity, ChRM inclination and declination, and the reversal angle. A = core CON 01-603-2. Numbers in the simplified lithological column indicate marine isotope stages (MIS), after Figure V.3. The paleomagnetic data show a geomagnetic excursion with a short, but full, reversal of the local field vector at the beginning of MIS 6; B = core VER 98-1-1. In this case the excursion represented by a strong deviation of the reversal angle during a period of low intensity occurs in MIS 3 and corresponds to the Laschamp event; C = core VER 98-1-1. In this case, the excursion is also represented by a strong deviation of the reversal angle during a period of low intensity. Again this occurs in MIS 3 and corresponds to the Laschamp event.

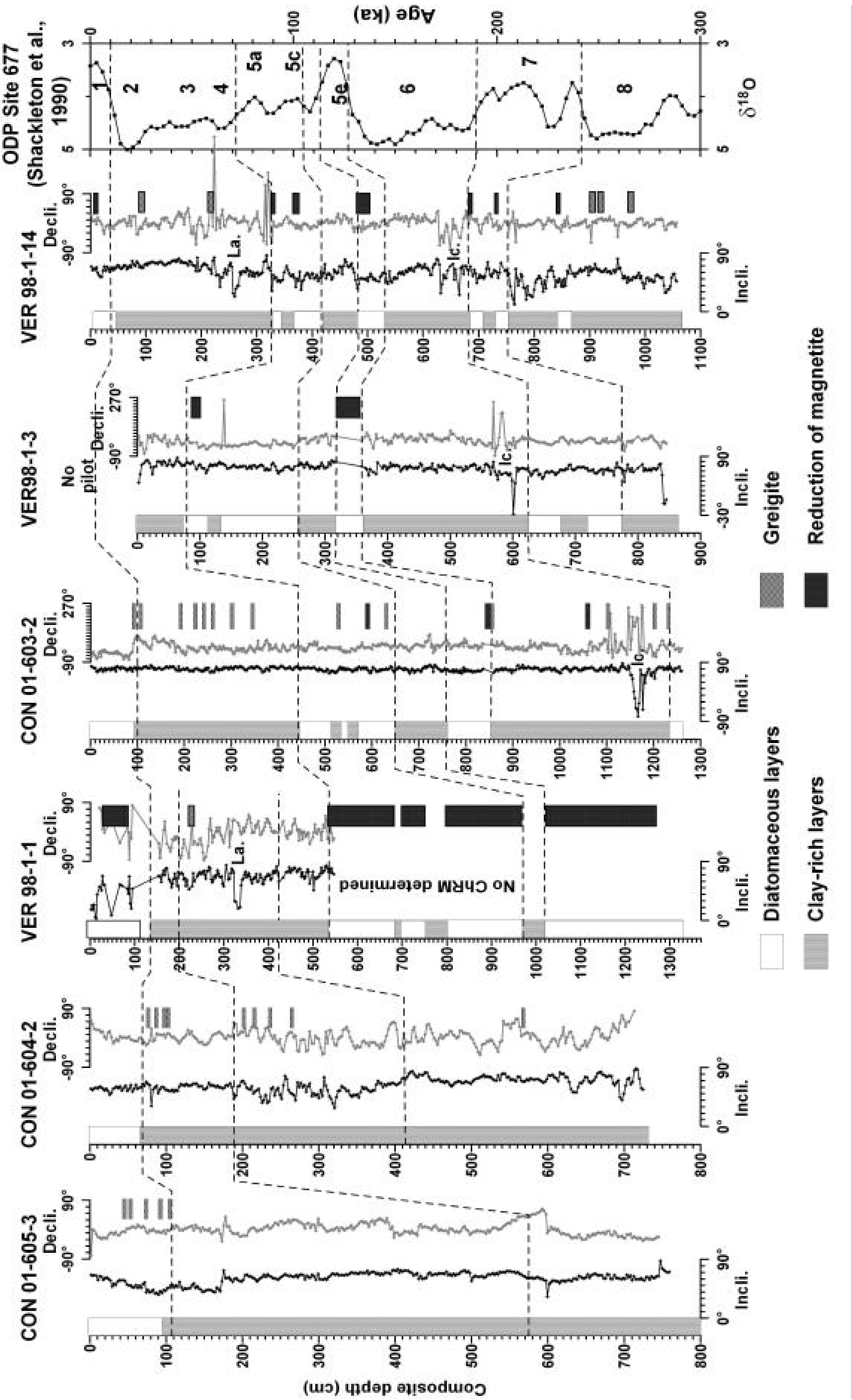

ODP Site 677 (Shackleton et al., 1990)
Age (ka)
$\delta^{18}O$
VER 98-1-14
VER98-1-3
No pilot
CON 01-603-2
VER 98-1-1
No ChRM determined
CON 01-604-2
CON 01-605-3
Decli.
Incli.
La.
Ic.
Composite depth (cm)
Greigite
Reduction of magnetite
Diatomaceous layers
Clay-rich layers

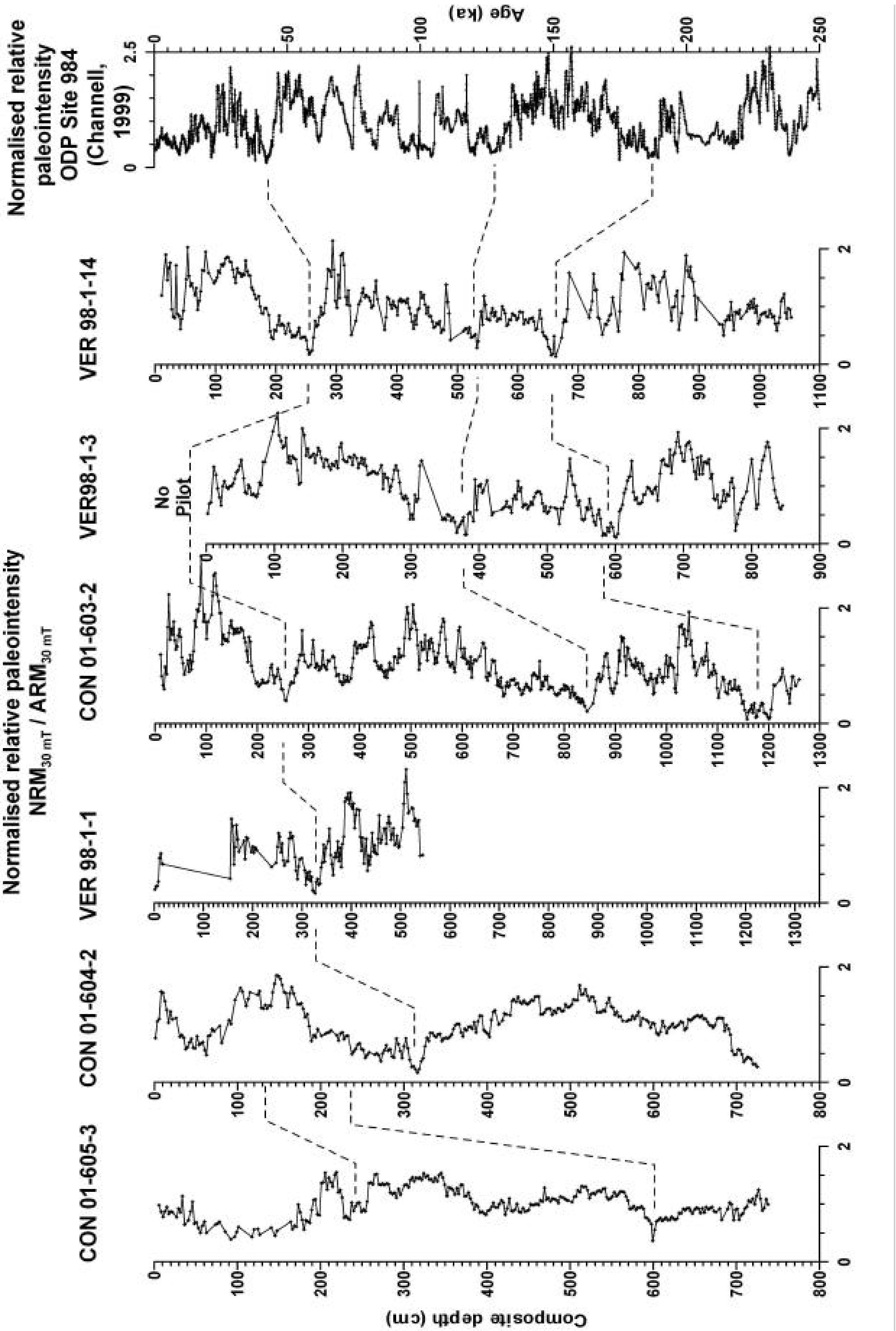
Figure V.7. Down-core variations of the inclination and declination of ChRM and simplified lithological columns of all investigated cores. Dashed lines mark some correlation levels of the cores from Lake Baikal with the dated d^{18}O record from ODP 677 (Shackleton et al., 1990), see also Figure V.3. Diagenetic features such as dissolution of magnetite and mineralization of greigite are marked according to Figure V.4. Laschamp and Iceland Basin excursions are indicated as La. and Ic., respectively.

The Iceland Basin excursion, documented as a full reversal in core CON 01-603-2, corresponds to large deviations of the ChRM inclination and declination in cores VER 98-1-3 and VER 98-1-14. Neither greigite mineralization nor magnetite dissolution occurs in the intervals where the Laschamp and Iceland Basin excursions are observed. Finally, in the time window where the Blake excursion is expected, around 120 ka, according to e.g. Nowaczyk et al. (1994) or Langereis et al. (1997), there is no directional relic of this event. This is probably due to the diatomaceous type of sediments deposited in this interval, which are strongly affected by magnetite dissolution. Such dissolution processes affect mainly the smallest particles of magnetite, which are the main carrier of the directional variability of the geomagnetic field.

V.4.4. Age model based on paleomagnetic correlations

In addition to *AMS* ^{14}C dating and the geomagnetic excursions, the age model was completed and refined by tuning the relative paleointensity records to the equivalent record from ODP Site 984 (Channell, 1999) (Fig. V.8). The relative paleointensity variations in Lake Baikal and ODP Site 984 are well correlated. This confirms the global geomagnetic field origin of the relative paleointensity variations documented in the present study. In addition, it shows that local sedimentary variations have no effect on the paleomagnetic records.

Figure V.8. Down-core variations of normalised relative paleointensity after removal of intervals affected by diagenesis (magnetite dissolution and/or greigite formation), and correlation to the relative paleointensity record from ODP Site 984 (Channell, 1999).


Normalised relative paleointensity ODP Site 984 (Channell, 1999)
Age (ka)
VER 98-1-14
VER98-1-3
No Pilot
CON 01-603-2
VER 98-1-1
CON 01-604-2
CON 01-605-3
Normalised relative paleointensity NRM30 mT / ARM30 mT
Composite depth (cm)

The long, continuous paleointensity record (CON 01-603-2) has a resolution of ~ 350 years and the resolution of its correlation with the reference curve is of ~ 3.5 ka, or, ~ every 10 samples (55 points of correlation span the last 200 ka). Such high resolution for telecorrelation might be more critical because of possible delays caused by differing remanence acquisition processes, local sedimentary or unresolved geochemical influences and regional non-dipolar components of the geomagnetic field. Despite this uncertainty, the fact the Iceland Basin event (with an estimated duration of about 3 ka) is very well documented both in North Atlantic sediments (Channell et al., 1997; Channell, 1999) and in Lake Baikal (Oda et al., 2002 and this study), suggest that telecorrelation between North Atlantic and Lake Baikal paleomagnetic records is possible. The paleomagnetic correlation allows us therefore, to refine the age model with a high precision for each individual sedimentary sequence (Fig. V.9).

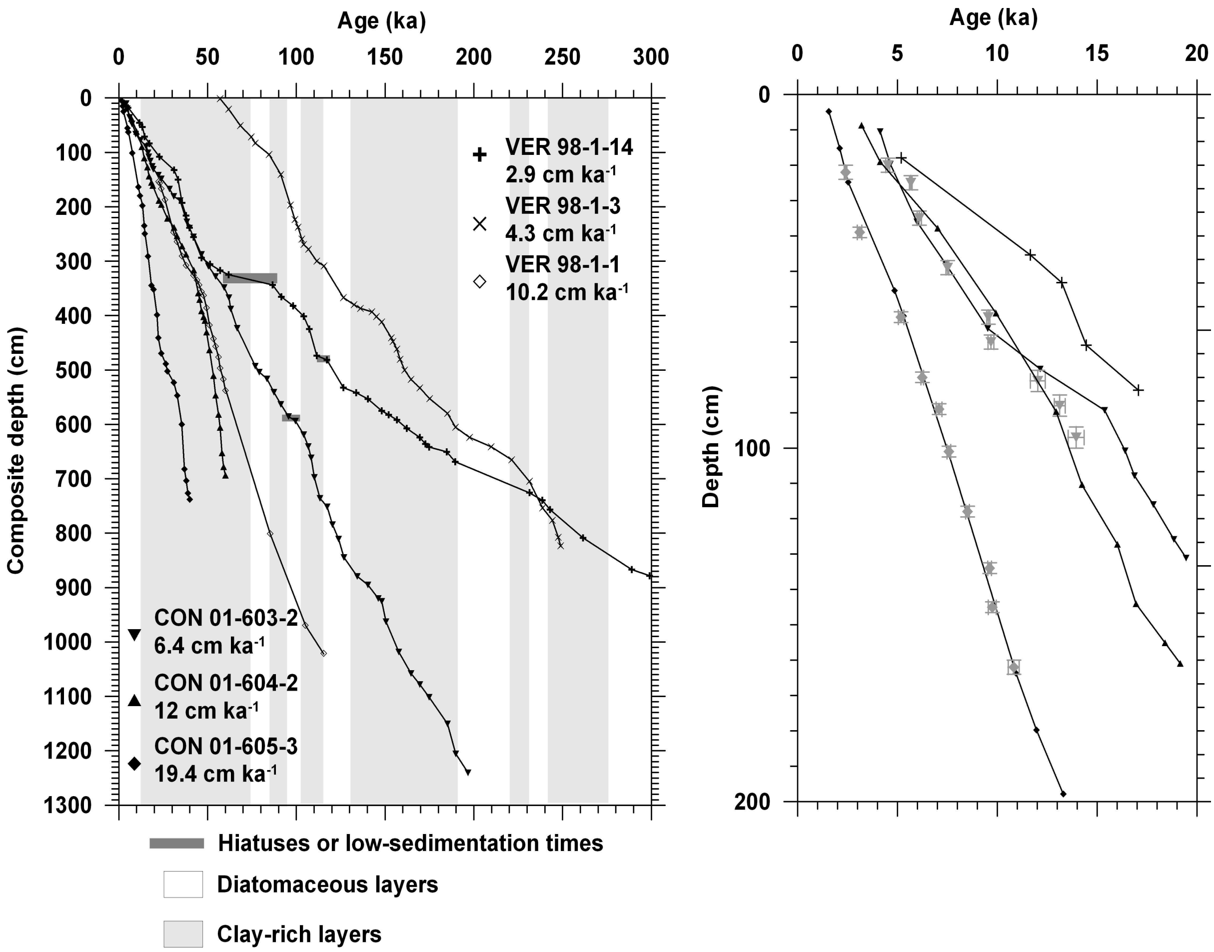


Figure V.9. Diagram of depth versus age based on relative magnetic paleointensity correlations (Fig. V.8) for all six investigated sites. In the figure, the average sedimentation rates are also indicated. The grey bands correspond to the time span of stadial+glacial periods. To the right, the blow-up of the area of the diagram covering 200 cm, shows the AMS ^{14}C dating for the last 20 ka, including error bars.

The average sedimentation rates show lower values for sites far from river influences (VER 98-1-3: 4.3 cm ka^{-1}, VER 98-1-14: 2.9 cm ka^{-1} and CON 01-603-2: 6.4 cm ka^{-1}). Rates are higher on slopes (VER 98-1-1: 10.2 cm.ka^{-1}), close to the Selenga Delta (CON 01-604-2: 12 cm ka^{-1}), or close to a canyon system (CON 01-605-3: 19.4 cm.ka^{-1}, see Charlet et al. submitted) due to the higher detrital input.

Age/depth curves of individual sites show non-linear shape, suggesting that sedimentation rates are variable with time. As reported in previous studies (Colman et al., 1996), changes of sedimentation rates are not related to climatic changes. In general, Lake Baikal seems to have

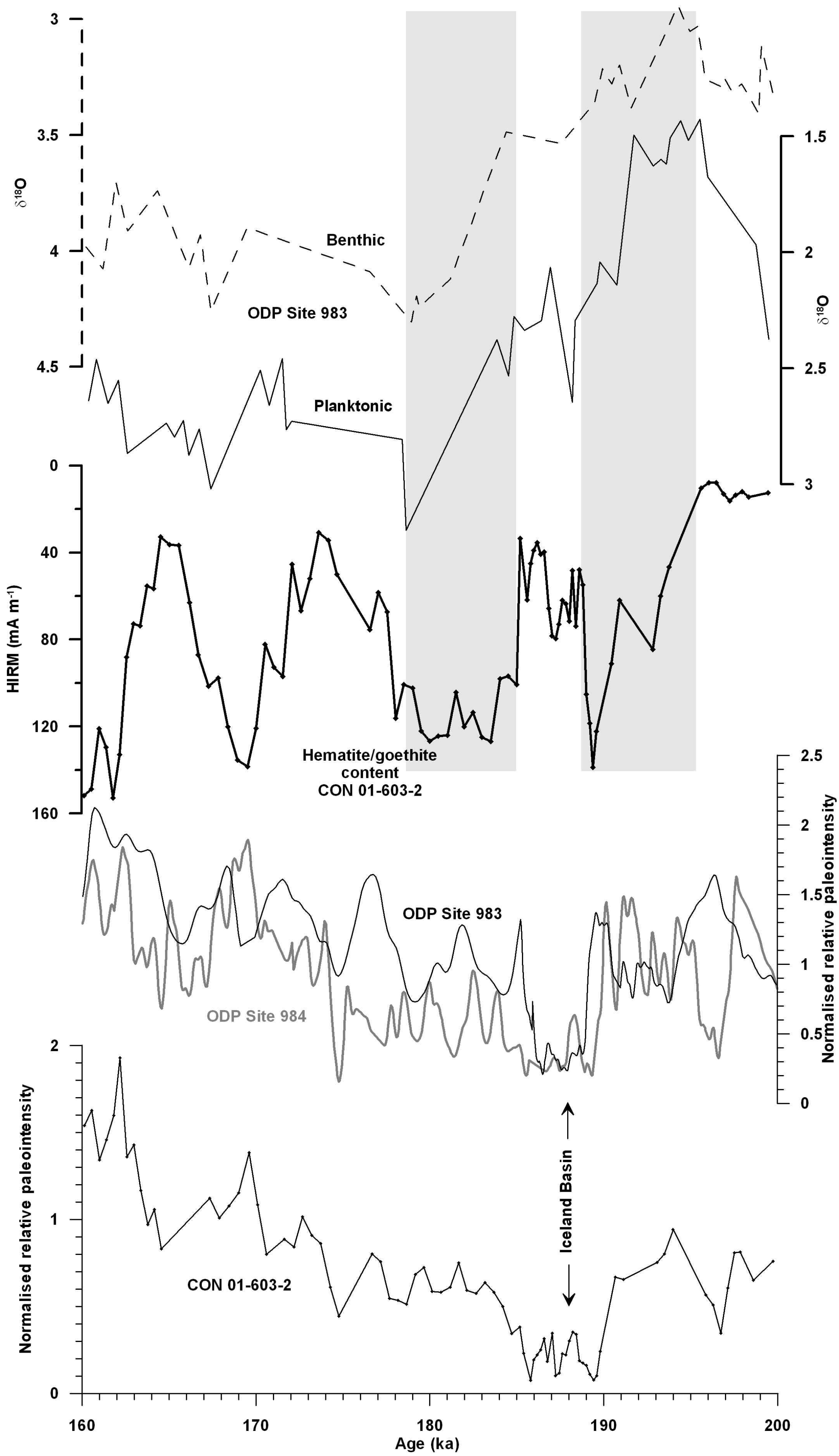
a rather constant sedimentary flux whatever the climatic regime, with high biogenic productivity during warm periods, compensating the detrital input, which is more important during the cold periods. Strong decreases of the sedimentation rate and even gaps are localised on top of some interglacial intervals. However, these gaps cannot be correlated from one to another core. Hence, gaps are not lake-wide features related to external processes. More likely, the variations of sedimentation rates are either due to periods of low or no sedimentation (Deike et al., 1997) resulting from winnowing (Ceramicola et al., 2002) or due to local slumping linked to tectonic activity in the Lake Baikal region (Charlet et al., submitted). Indeed, the interglacial sediments, rich in water and diatoms, are easily destabilised.
Consequently, considering the many variations evidenced by the high-resolution paleomagnetic correlations, correlations like those presented in Figure V.3 lead to a highly simplified age model. Although sedimentation rate changes are not obviously climatically induced, their omission would lead to strong misdating of climatic events.

V.5. Discussion

V.5.1. Lake Baikal response to global climatic change

In order to characterise Lake Baikal sedimentary responses to global climatic changes that may be recorded in marine sediments, we compared our paleomagnetically dated climate-proxy record from Lake Baikal with benthic and plankontic $\delta^{18}O$ curves of ODP Site 983, a site close to ODP Site 984. The neighbouring site was chosen for comparison because although the quality of the ODP site 984 paleomagnetic record is high, its $\delta^{18}O$ records are of lower quality than those of ODP site 983. Synchronous paleomagnetic variations observed in ODP sites 983 and 984 sediments (Fig. V.10) show that the premise of our age model based on paleomagnetic correlation is identical, if the reference curve used for correlation is from ODP site 983. We can, therefore, compare climatic records from ODP site 983 and Lake Baikal. The climatic proxy used for Lake Baikal sediment is the HIRM record since it displays the detrital input variations (Peck et al., 1994).
Here we have focused on the transition of MIS 6/7 in the sedimentary sequence of CON 01-603-2, as close to this climatic transition, the pronounced Iceland Basin event enables a detailed comparison of Lake Baikal and ODP Site 983 climatic records (Fig. V.10). At MIS 6/7 transition in ODP Site 983, $\delta^{18}O$ values increase in two steps: (1) a strong increase of the planktonic $\delta^{18}O$ occurs at ~ 192 ka documenting a rapid sea surface cooling; (2) a second increase, affecting both planktonic and benthic $\delta^{18}O$ records, is observed c. 7 ka later, at ~ 185 ka, which is related to the global ice volume change. The strong increase (1) correlates with a strong increase of HIRM in Lake Baikal record, while the increase (2) observed in the planktonic and benthic $\delta^{18}O$ record from ODP Site 983 correlates with a moderate increase of HIRM. In Lake Baikal, the strongest cooling (marked by a significant increase of HIRM) occurs synchronously with the sea surface cooling observed in the North Atlantic at ~ 192 ka. Therefore, the main cooling event in Central Asia occurs ~ 7 ka before the global ice volume change recorded in the benthic $\delta^{18}O$ from North Atlantic sediments and dated at ~ 185 ka.

Figure V.10. This figure provides a focus on the 160 ka-200 ka time window. Here the relative paleointensity record of CON 01-603-2, of ODP Site 984, (Channell, 1999), i.e. its reference curve for dating, as well as the relative paleointensity record from ODP Site 983 (Channell et al., 1997) are plotted together with the HIRM record of CON 01-603-2 and planktonic and benthic $\delta^{18}O$ record from ODP Site 983. The records are plotted on their respective time scales. The location of the Iceland Basin event is also denoted by the double arrow, and the two step increase in $\delta^{18}O$ by grey rectangles.



δ18O
3
3.5
4
4.5
Benthic
ODP Site 983
Planktonic
1.5
2
2.5
3
δ18O
HIRM (mA m-1)
0
40
80
120
160
Hematite/goethite
content
CON 01-603-2
ODP Site 983
ODP Site 984
Normalised relative paleointensity
2.5
2
1.5
1
0.5
0
Normalised relative paleointensity
2
1
0
Iceland Basin
CON 01-603-2
160
170
180
190
200
Age (ka)

V.5.2 Mismatch between dating of Iceland Basin events

The dating of the Iceland Basin excursion (177 to 183 ka) performed by Oda et al. (2002), who used X-ray CT tuned to $\delta^{18}O$, is different from the age range given by Channell et al. (1997) of 186 to 189 ka. The time-lag between Central Asia climatic change and global ice volume change at transition MIS6/7 presented in section 5.1 is probably responsible for the dating discrepancy. Indeed, Oda et al. (2002) undertook correlation with the benthic $\delta^{18}O$ record at ODP site 677 (Shackleton et al., 1990) for establishing their age model, and dated the Iceland Basin at the beginning of the MIS 6. However, Channell et al. (1997) performed a correlation of planktonic and benthic $\delta^{18}O$ records with SPECMAP (Martinson et al., 1987). Neglecting both cooling phases when tuning the $\delta^{18}O$ records to SPECMAP could have biased the dating of ODP site 983 and 984 paleomagnetic records. However coupled paleomagnetic and cosmonuclide geochemistry studies have recently corroborated the location of the Iceland Basin event at the end of the MIS 7 (Thouveny et al. 2004; Carcaillet et al., 2004)

Different lock-in of the geomagnetic field variations in the different sediments could also contribute to the misdating. Indeed lock-in depths exist but they are now well quantified by comparing cosmogenic nuclides and paleomagnetic variations (Carcaillet et al., 2004). The sedimentation rate is about 2.5 times lower in core CON 01-603-2 than in ODP Site 984. If we consider empirically the same lock-in depth for both North Atlantic and Lake Baikal sediments, the paleomagnetic information should be recorded in older sediments in CON 01-603-2 than in ODP Site 984. However, with regard to the age model used by Oda et al. (2002), the Iceland Basin event is recorded in younger sediments in Lake Baikal than in ODP Site 984. The time lag between the excursion dated by Oda et al. (2002) in Lake Baikal and the same excursion dated by Channell et al. 1997 at ODP Site 983 is, therefore, likely due to a misidentification of climatic boundaries.

V.5.3. Stacked records versus individual records.

We established a mastercurve “Baikal 200” of relative paleointensity, which represents a new synthetic paleomagnetic archive for Central Asia. The synthetic record is composed of mean values of the 6 records with respect to a sliding time window of 2 ka. This compilation has been restricted to the last 200 ka in order to maintain a representative population of points (between 2 and 68). However, we present this synthetic record together with individual records and the reference paleointensity curve (Fig. V.11) for the following reasons:

- Each relative paleointensity record has a different resolution (e.g. sedimentation rates in CON 01-605-3 are five times higher than in VER 98-1-14). During stack procedure, smoothing of the data had the effect of lowering the resolution of the paleomagnetic information.
- This stack does not provide more information on timing of the geodynamo changes since the records are tuned to ODP Site 984.

The relative paleointensity record CON 01-603-2 (Continent Ridge) is the best individual paleomagnetic record for the last 200 ka developed for Lake Baikal sediments. It contains only few, and short, hiatuses. Furthermore, it is accurately dated with AMS ^{14}C dating performed on sediments from a parallel core from the same site. The record CON 01-605-3 (Vydrino Shoulder) shows the highest resolution and is also anchored by AMS ^{14}C dating.

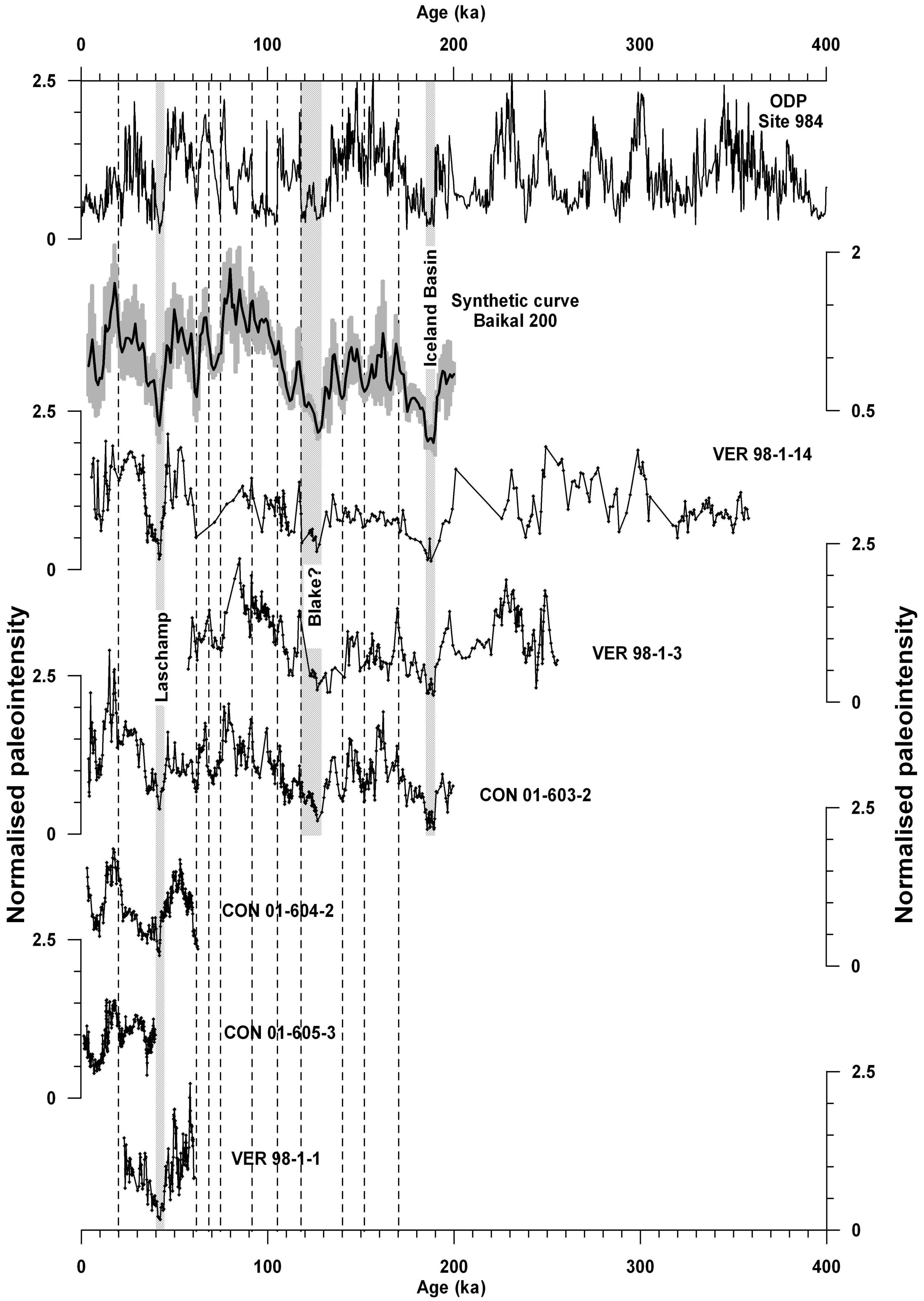


Figure V.11. Here we show diagrams of paleointensity versus age of all the sedimentary sequences of the present study, of the synthetic curve resulting from its compilation from other curves, and of the reference curve from ODP Site 984 (Channell, 1999). For the compilation, data have been averaged using a sliding window of 2 ka (the variance is marked by the grey shadow). Dashed lines show some of the correlations. The grey lines show the location of the low paleointensities related to geomagnetic excursions. Note that the lowest paleointensities in the time span of Blake are at c. 129 ka.

V.6. Conclusions

Measurements of the anisotropy of magnetic susceptibility revealed the excellent quality of the recovered cores selected for the present study. Rock magnetic analyses allowed detection of sediment intervals affected by reductive dissolution of magnetite and/or the diagenetic formation of secondary greigite. Although geomagnetic excursions are remarkably preserved (Laschamp at about 42 ka and Iceland Basin at about 185 ka), others are missing mostly due to diagenetic processes. After exclusion of these diagenetically affected zones from the data sets, high-quality relative paleointensity records could be obtained. Correlating these records to other well-dated records, especially ODP Site 984, and supported by *AMS* ^{14}C dating, high-resolution age models for the investigation of Lake Baikal sediment cores could be derived. The average sedimentation rates range from 2.9 to 19.4 cm per ka. Larger variations in the sedimentation rates point to a high diversity and complexity of the sedimentary environments within Lake Baikal. The complex shape of the age model curves due to low sedimentation makes the age model based on tuning climatic records of Baikal with marine $\delta^{18}O$ records unsuitable at high resolution. Another observation deduced from our age model, is the timing between Lake Baikal and marine climatic responses. At the MIS7/MIS6 transition where the Iceland Basin excursion is a strong tie point, the climatic conditions in the Lake Baikal region turn colder, at the same time as sea surface cooling occurs in the North Atlantic. Since sea surface temperature change recorded by planktonic $\delta^{18}O$ occurs earlier than the global ice volume change recorded by benthic $\delta^{18}O$, it implies a bias in any age model based on tuning climatic records from Lake Baikal sediments with benthic $\delta^{18}O$ records from marine sediments.

Acknowledgments:

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VI. Detrital input traced by grain size distribution and rock magnetism in Late Quaternary sediments from Lake Baikal

VI.1. Introductory remarks

Transport of detrital particles into Lake Baikal is controlled by eolian and fluvatile processes though their contribution has not yet been quantified. Reworking of older detrital material is a third process, which must be addressed though its quantification is difficult. As earthquakes are quite frequent and slopes along the rift flanks are steep, turbidites are common in the centre of basins. For minimising the reworking a coring site with hemipelagic sedimentation was selected. However, redeposition of reworked detrital components due to strong currents within the water body and near the lake bottom can not be ruled out for hemipelagic settings.
The balance between the eolian and fluviatile detrital input is entirely controlled by the regional hydrology. Today the seasonally varying moisture distribution, which regulates the precipitation in the catchment area of Lake Baikal, is mostly driven by the Westerlies. Mean daily precipitation during summer exceeds winter precipitation by a factor of 10. Occasionally strong summer rainfalls lead to flood events in the hinterland (Heim et al., submitted), which increase the suspension discharge into the lake significantly. The lake is feed by many rivers covering a huge catchment (560 000 km^2), which spreads mostly eastward and south eastward of the lake (Fig. VI.1A). The three main tributaries are Selenga, Barguzin and Upper Angara Rivers.
As to the eolian input of detritus, we have to expect some increased activities from time to time during summer and winter. During the summer season, eolian input may increase due to exceptional strong atmospheric turbulence responsible for dust events due to extreme heating of the surface. During dry winter coarser silt and fine sand particles are carried by saltation on the ice cover from the nearby shores.
To date, pollen (Demske et al., 2002; Demske et al., submitted) and diatoms assemblages (e.g. Mackay et al., 1997; Khursevich et al., 2001) are the most commonly used climatic proxies for reconstructing climate variability at Lake Baikal. Only few studies deal with detrital proxies, like clay mineralogy (Yuretich et al., 1999) and grain size analyses. Francus (1998) carried out the first grain size study on turbiditic sediments from Lake Baikal using the image analysis technique and refining it (Francus and Karabanov, 2000). Recently results of a laser assisted grain size analysis on hemipelagic sediments from the BDP section covering several million years and 140 ka respectively have been reported (Kashiwaya et al., 2001; Ochiai and Kashiwaya, 2003). In all the studies the data were used to constrain parameters such as temperature and moisture distribution in Lake Baikal. Fagel et al. (2003) notably confirmed that clays cristallinity and weathering is related to the temperature. This has lead to conclusion that interglacial periods were more humid than glacial periods. Ochiai and Kashiwaya (2003) showed further that abundance of fine particles increased during cold periods.
We discuss data from a laser-assisted grain size analysis of the detrital fraction > 2 µm from sedimentary sequence CON01-603-2, retrieved from the Continent Rise in the Northern Basin of Lake Baikal (Fig. VI.1B). Moreover, the following approach will be used for interpretation. We will use temperature and moisture indexes as reconstructed from pollen assemblages (Karabanov et al., 2000; Demske et al., submitted; Granoszewski et al., submitted) for comparison with grain size data, characterising environmental changes during interglacial periods such as Holocene and Eemian. These indexes may help to understand the distribution of grain size in time and shed lights on processes driving the detrital input into Lake Baikal.

A

B

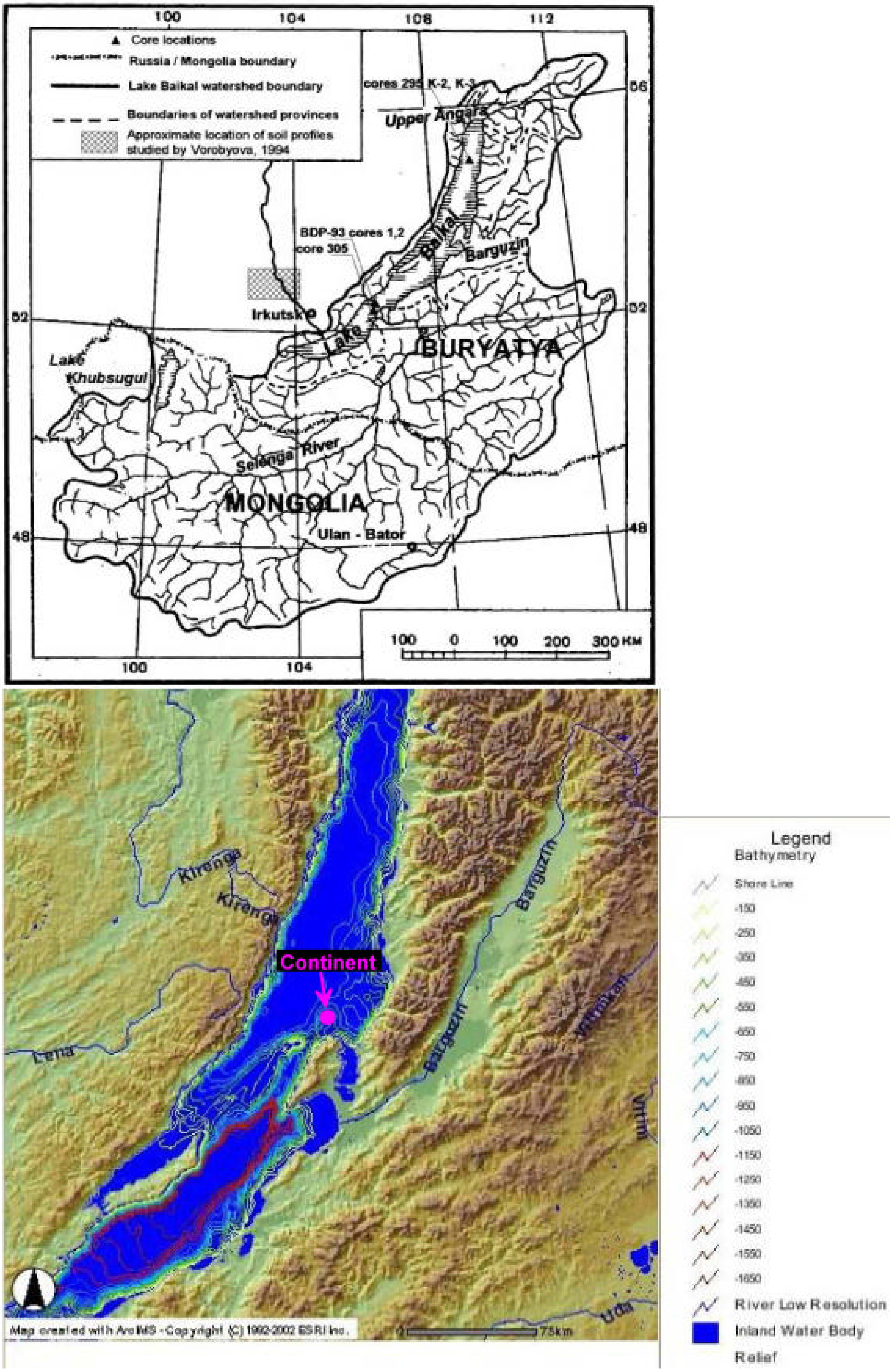


Figure VI.1: A. Catchment Area of Lake Baikal (after; Karabanov et al., 2000b); B. simplified bathymetric map around site Continent, relief of its surrounding and main rivers. The site of coring, Continent, is marked by the dot.

The study is complemented by selected rock magnetic parameters. We will use ARM shown to be a tool for characterising the lithology. HIRM, which estimates the contribution of ferromagnetic minerals and the quantity of hematite, will be used to trace the detrital input into Lake Baikal. Before using the rock magnetic data we ruled out those biased by early diagenesis (Demory et al., submitted-a; see pages 31 to 50).
After evaluation of the different parameters ensued from this multidisciplinary study, selected grain size and rock magnetism parameter could reflect, if any, teleconnections with north Atlantic.

VI.2. Material and methods

VI.2.1. Material

For the grain size study, we choose a sedimentary sequence composed of a pilot core and a piston core retrieved in 2001 from the Continent Ridge (Fig. VI.1B). This sequence consists of hemipelagic sediments deposited on the Continent Ridge, a geomorphological high located in the prolongation of the Academician Ridge where most of the cores studied earlier have been retrieved (BDP cores, see e.g. Khursevich et al., 2001; Müller et al., 2001; Prokopenko et al., 2001; and Vereshagin cores, see e.g. Oda et al., 2002; Fagel et al., 2003; Watanabe et al., 2004). The sedimentary record of the studied cores is nearly continuous and covers a time span from the Holocene to the interstadial equivalent to the marine isotope stage MIS 7. The cold periods (glacials+stadials) are characterised by clay-rich sedimentation whereas the warm periods (interglacials+interstadials) are characterised by diatomaceous sedimentation. The uppermost section, continuously sampled at 2 cm intervals, was precisely dated using *AMS*^{14}C for first 15 ka (Piotrowska et al., In press). Beyond 15 ka we used paleomagnetic data (Demory et al., submitted-b; see pages 51 to 70). The reference curve established by Channell (1999) at ODP site 984 was used to date the Kazantsevo (Eemian time equivalent). For dating the time window 15 to 110 ka, we used the paleomagnetic intensity record from the Labrador Sea, Core MD 95-2024 (Stoner et al., 2000). The record of Stoner et al. (2000) only covers 110 ka but seems to be better dated than the Channell (1999) record since the former is tuned to the high resolution chronology of GISP 2 (Grootes et al., 1993) and the latter is tuned to the low resolution chronology of the orbitally tuned SPECMAP curve (Martinson et al., 1987). The results of both age models are shown in Figure VI.2 which shows only minor difference between the two calibration curves.

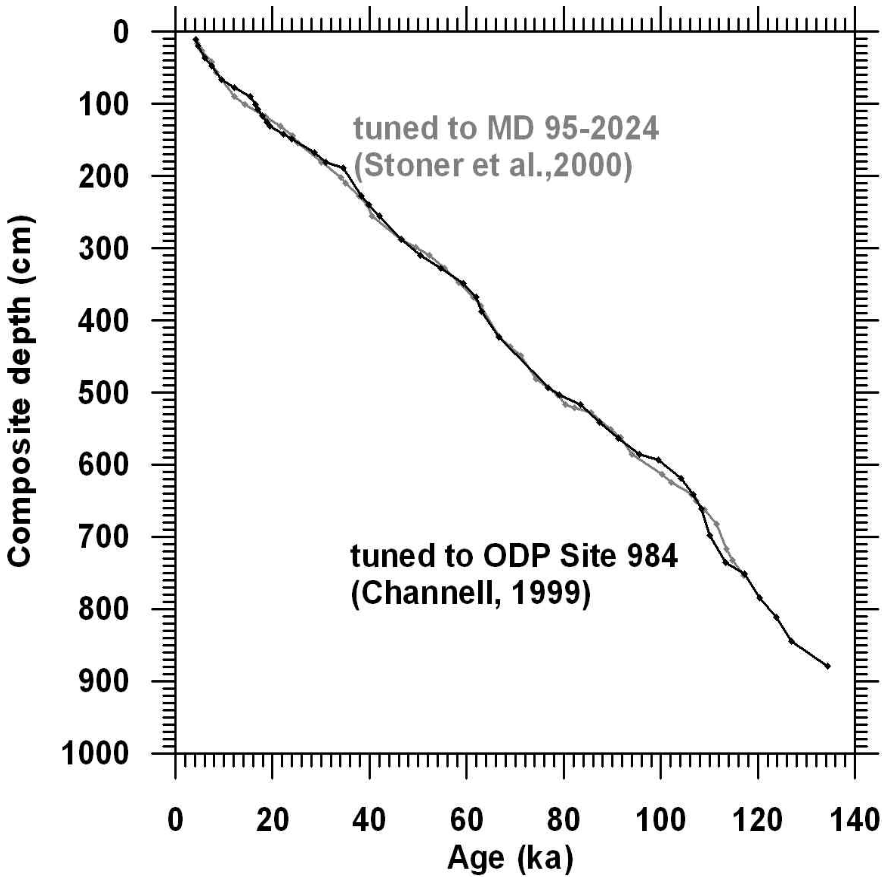


Figure VI.2: Age model (depth versus age) for the sedimentary column CON 01-603-2 based on calibrated AMS ^{14}C (Piotrowska et al., In press) and on paleomagnetic correlation (Demory et al., submitted-b) using the reference paleomagnetic curves from Iceland Basin (ODP Site 984, (Channell, 1999), chronology based on SPECMAP (Martinson et al., 1987)) and from Labrador Sea (MD95-2024 (Stoner et al., 2000) chronology based on GISP2 (Grootes et al., 1993)).

VI.2.2. Preparation of samples for laser-assisted grain size analysis

We quantified the different grain size fractions (particles > 2 µm, clays, opal) on 448 samples. For grain size analyses the samples were prepared as follows. 1 g of freeze dried sediment was soaked in a H_2O_2 solution (5 %) in order to remove the organic carbon. Then, the clay fraction was separated by centrifugation. In the next step, the biogenic silica was dissolved at 90°C, while shaking the sample for 5 hours in a 2M solution of Na_2CO_3. To remove the carbonates and some authigenic hydroxides, the remaining sediment, soaked in a HCl solution (1.1N), was treated with ultrasonic for 1 hour. For 56 samples, which still contained diatoms and spicules of sponge after treatment with Na_2CO_3, the remaining opal was separated with a density separation (at 2.32 g/cm^3) using sodium poly-tungstate, a heavy salt, diluted in water. A density of 2.32 is necessary for a complete separation. The clay fraction as well as the detrital fraction > 2 µm were dried and thereafter weighted. The quantity of opal digested by Na_2CO_3 was measured with ICP-OES at the University of Potsdam. The error bar is ± 5%.
Grain size distribution of the detrital fraction > 2µm was measured using Malvern Mastersizer 2000 equipment at the University of Lille. The principle of measurement is presented in Chapter III, "Background and methods". The grain size distribution is displayed as the mode, which is the most represented size in the distribution and also as the mean of its cumulative curve.
Separation of the clay and silt fractions is difficult to complete. After multiple separation steps an irreducible fraction of clays persists in the size fraction >2 µm. However, given the small

and constant amount (~ 5 %) of clay remaining in the silt fraction (see Fig. VI.3) we consider that it does not affect our interpretations.
In terms of downcore variations, spikes involving one data point are excluded from interpretations since such spikes could result from an artefact due, for instance, to sample preparation. Using tests of reproducibility after laboratory treatment, an estimated error bar of ± 1 µm for the mode seems realistic, while the error bar on the quantification of each fraction is considered to be ± 5 % (see Chapter III, "Background and methods").

VI.2.3. Rock magnetism

The rock magnetic parameters used in this study involve measurements of Anhysteretic Remnant Magnetisation (ARM). This artificial magnetisation was imprinted with 50 µT static field and 100 mT alternating field, using a 2G Enterprises 600 single-axis demagnetiser with an in-line ARM coil. ARM was measured using a fully automated 2G Enterprises DC-SQUID 755 SRM long-core. For the rock magnetism, discrete samples were taken using 6.2 cm^3 plastic boxes. We present the record of ARM, which represents the ferromagnetic mineral content of the sediment. In Lake Baikal, this parameter highlights the glacial/interglacial alternation. Indeed ARM is high during cold periods due to high detrital input and low during warm periods. The low values are due to less detrital input, dilution by biogenic opal and post-depositional dissolution of ferromagnetic particles (Demory et al., submitted-a; see pages 31 to 50).
Isothermal Remnant magnetisation (IRM) was imprinted with a 2G Enterprise 660 pulse magnetizer at 0.3 T and at 2 T. The field at 2 T saturates the sample and results in the saturation isothermal remnat magnetisation (SIRM). These magnetisations were measured with a Molyneux MiniSpin fluxgate magnetometer. The quantity of high coercivity minerals (hematite or goethite) is estimated by the High Isothermal Remnant Magnetisation HIRM, which results from the following relationship:
$HIRM = (IRM_{2T} - IRM_{0.3T})/2$
This parameter estimates the detrital input such as eolian particles in ice records (Maher and Dennis, 2001) or in Lake Baikal sediments (Peck et al., 1994) though its accuracy is still disputed (e.g. Liu et al., 2002).

VI.3. Results

In Figure VI.3 we present averaged grain-size distributions of the detrital fraction > 2 µm for selected time periods and different sediments. Fine silts generally dominate the grain size distribution of the detrital fraction > 2µm. The silt fraction is fine (with a mode around 9 µm) during glacial and interglacial optimums while the silt fraction coarsens (with a mode around 10.5 µm) during climatic transitions. Coarse grains (> 63 µm) are restricted to cold and temperate climate periods. This grain size shifts the ratio mean/mode towards higher values. In the same figure, we also present the grain-size distribution of detrital material trapped in fresh snow above the Lake Baikal ice cover. This dust, which is of eolian origin, is dominated by fine silt particles like the detrital particles > 2 µm extracted from Lake Baikal sediments. The clay contribution is weak although no treatment (i.e. no grain size separation) was applied to this eolian sediment.

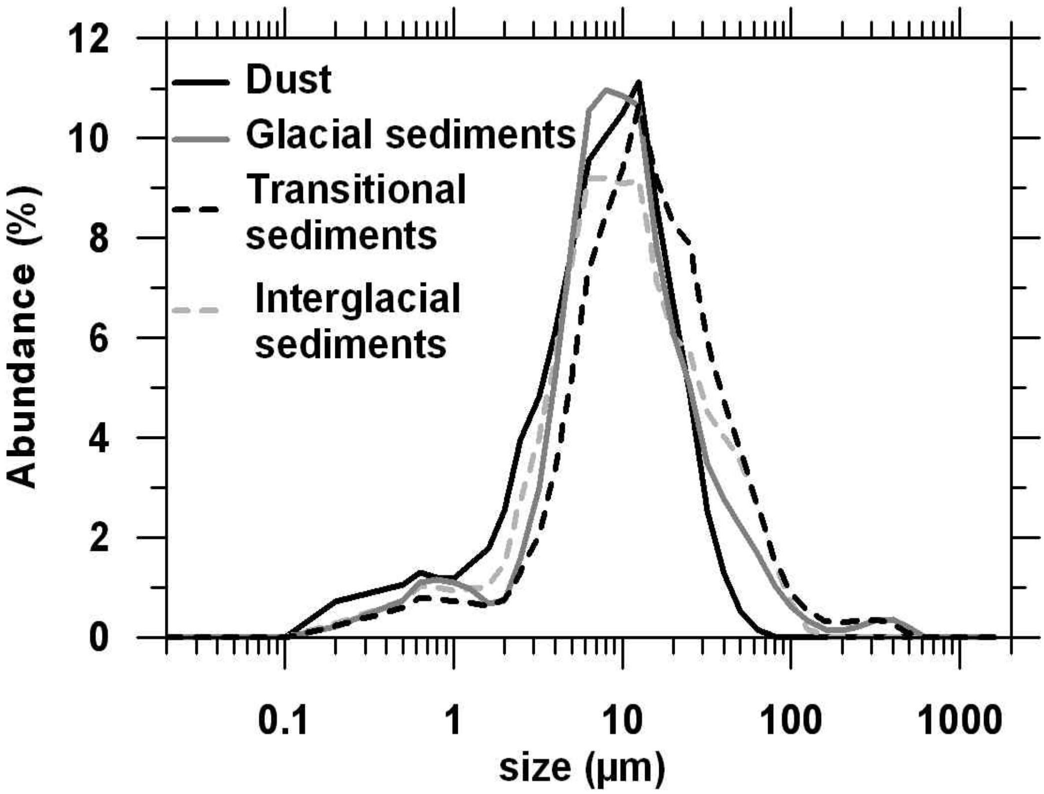


Figure VI.3: Grain size distribution of the detrital fraction > 2 µm for 4 samples: one from glacial period, one from an interglacial period, one from a transitional period and one from sediments trapped in fresh snow covering Lake Baikal. The error bar on the mode is ± 1 µm. The average error bar given by Malvern manufacturer on the measurement is 1 %, but considering individual classes, the error bar reaches 2 % between 900 and 5 µm and 6 % for the finer fractions.

The clay and opal contents generally mimic the classical clay-rich/diatom-rich alternation, which reflects the cold/warm alternation (Fig. VI.4). Cold periods are characterised by high amounts of clay (up to 80 %) and by an absence of opal whereas warm periods are characterised by low amounts of clay (down to 40 %) and high amounts of opal (up to 25 %). An increase in the silt size and quantity is observed at the beginning of the Kazantsevo (from 133 ka to 127 ka) and during late Sartan (late Glacial) and early Holocene (from 13.5 ka to 3 ka). The relative abundance of coarse sand particles is evidenced in the varying ratio of mean to mode. This ratio is higher during cold periods than during warm periods as the coarse size fraction increases during cold periods.

To summarise, the detrital characteristics can be grouped in three assemblages:

The *first group*, typical for glacial and stadial intervals, shows high clay (up to 80 %) and sand (>63 µm) contents (mean/mode reaching 1.3) and a low silt content (down to 15 %). Silts get finer (with a mode around 9 µm) by the end of the glacial/stadial periods.

A *second type*, common during interstadials, at the transitions glacial/interglacial (Terminations 2 and 1) and the early Holocene, shows a high variability. The size and abundance of detrital particles > 2 µm increases (mode from 9 to 10 µm and quantity from 15 to 30 %) while the quantity of clay and sand decreases (respectively from 80 to 40 % and a mean/mode from 1.3 to 0.9).

In a *third* assemblage, the size and amount of silt is minimal while the quantity of clay and sand is low. This type characterises the middle Kazantsevo and late Holocene.

As for the paleomagnetic signals, the ARM record reflects quite well changes between cold/warm conditions typically showing higher values during cold periods. HIRM also traces quite clearly the alternation between cold and warm intervals indicating high values during cold periods. The HIRM displays peak values during the interstadial, which is considered to be time equivalent to the MIS 3.

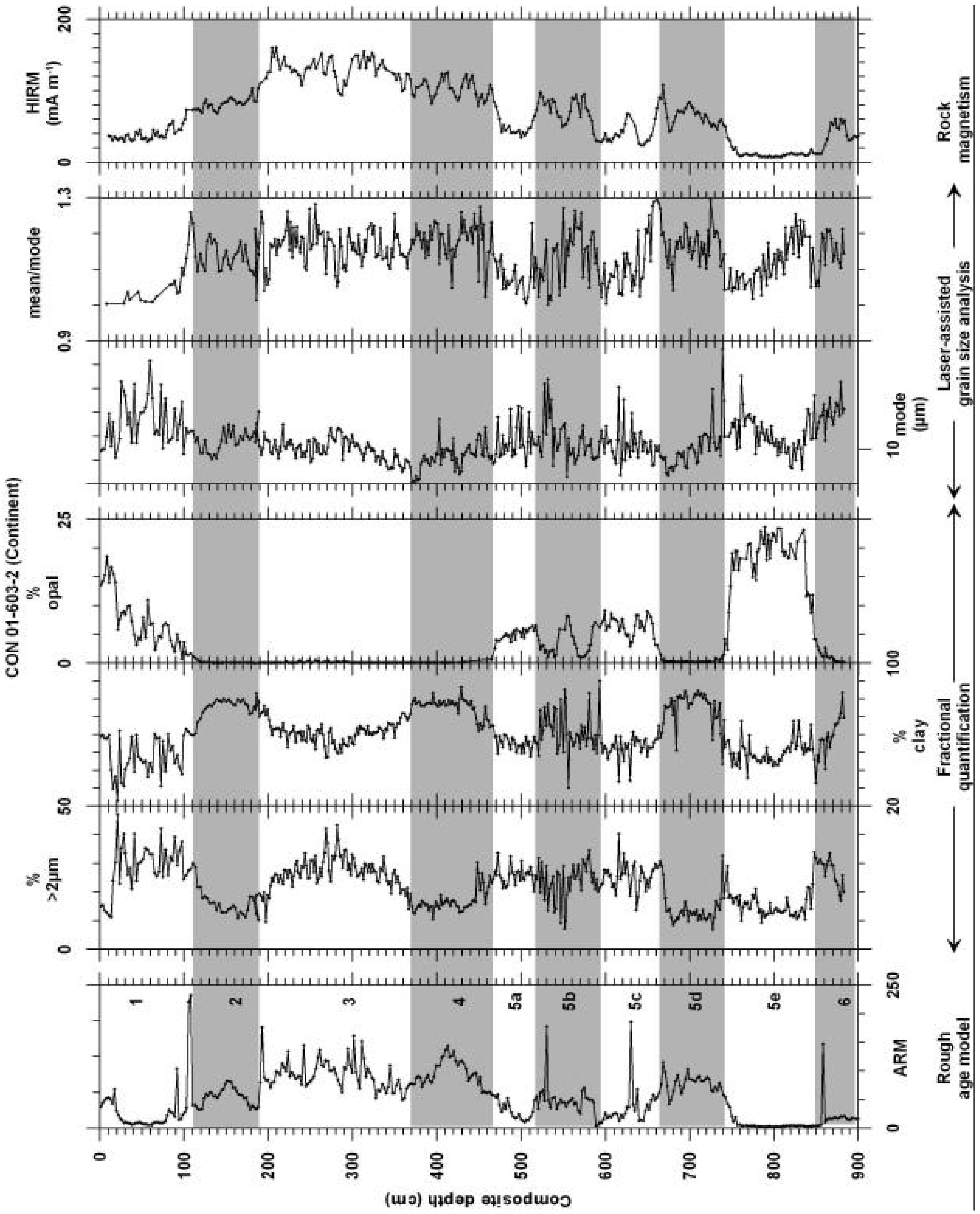
VI.4. Evaluation of the results

We focused on Kazantsevo and Holocene records and the related Terminations I and II, to understand processes driving the detrital input during interglacial and transitions from glacials into interglacials. The studied time slices are well dated with a high resolution age model ensued from paleomagnetic correlations (Demory et al., submitted-b; pages 51 to 70) and calibrated *AMS*^{14}C dating (Pietrowska et al. in press). First we evaluate the changes observed during the early through to late Weichselian, and following that we discuss the grain size results from the two interglacials Holocene and Kazantsevo. Grain size distribution patterns of the interglacials are compared with temperature and moisture indices as inferred from Holocene and Kazantsevo palynological data (Demske et al., submitted; Granoszewski et al., submitted). For the glacials and stadials no pollen data are available because the vegetation coverage is very sparse and of minimal diversity. In the next step we compare selected rock magnetic properties with the grain size distribution pattern in order to assess the potential of rock magnetic proxies for reconstructing the allochthonous input. Finally, we estimate which parameters contain a signal which could be correlated with events occurring at global scale, and we compare these records to North Atlantic climatic records.

VI.4.1. Variations in detrital components between the Holocene and the Kazantsevo

The cold Weichselian period (light grey in Fig. VI.4**,** from time equivalent MIS 2 to time equivalent MIS 5d) is characterised by high amounts of clay and sand (high mean/mode ratio). Karabanov et al. (1998) suggested that erosion of surrounding mountains by glaciers is the main supplier of clay minerals into Lake Baikal. The presence of glaciers was important on the eastern side of the Lake Baikal northern basin and glacigenic particles like tills entered into Lake Baikal (Back and Strecker, 1998). Nevertheless, significant production of clays by glaciers only holds true during periods with wet glaciers. It is rather unrealistic to assume that the glaciers where wet during a glacial period at a latitude where permafrost was widespread (Vogt and Larqué, 2002). Thus glacial abrasion cannot be the dominant mechanism contributing to the clay size fraction in lake sediments during cold maxima. We propose that wind is a much more efficient media to transport the fine fraction to the lake. Coarse particles (sand), which are abundant in glacial sediments, originate from physical glacial erosion and permafrost degradation. Indeed, the permafrost that developed around Lake Baikal under cold climatic conditions is responsible for cryoturbation, (also well described around Lake Baikal, see Chlachula, 2001). This cryoturbation facilitates the production of detrital particles, notably sandy particles. Such process for sand generation is well known in present-day cold and arid regions (Mountney and Russell, 2004). Furthermore, the transport of sand and clay particles by wind is facilitated by the availability of such particles. As moisture was rather low, Siberia has never been covered by ice sheets during cold periods (e.g. the recent synthesis by Svendsen et al., 2004). Furthermore, as already mentioned, the glaciers were restricted to the Barguzin mountain range (Back and Strecker, 1998), located on the eastern coast of the northern basin of Lake Baikal.

Figure VI.4: Downcore variations of the bulk results for the sedimentary column CON 01-603-2. From left to right: ARM, quantification of the different fractions (Opal, Clays, detrital particles > 2 µm), mode and mean/mode resulting from the grain size analysis performed on the detrital fraction > 2 µm and quantity of hematite+goethite (HIRM). The numbers to the left correspond to the equivalence with MIS. The grey rectangles correspond to the stadial and glacial periods.


CON 01-603-2 (Continent)
HIRM (mA m-1)
mean/mode
mode (µm)
% opal
% clay
% >2µm
ARM
Composite depth (cm)
1
2
3
4
5a
5b
5c
5d
5e
6
Rock magnetism
Laser-assisted grain size analysis
Fractional quantification
Rough age model

A major source for the coarse grains was most probably the nearby beach sands exposed along the shoreline. These coarse grains are possibly brought by wind driving the particles by saltation over the ice, which today is covering the lake in the Northern Basin from late December through late May. During glacial time the ice cover must have lasted much longer, possibly as long as 10 months to a year. Travelling distance to the core site CON01-603-2 is 12 km. As the grains mostly reach the open lake by saltation on ice, this mechanism is enhanced when snow cover is minimal or absent. The mobilisation of coarse particles by wind would be facilitated when the snow cover at the shore and the adjacent hinterland is minimal or absent. Thus the variable content in coarse particles is also considered as tracer of snow at this latitude. Furthermore the presence of coarse particles (> 63 µm) could indicate that wind dynamics increased when the lake was covered with ice during most of the year during glacials and stadials. To provide a minimum wind speed for sand motion is rather difficult since it mostly depends on friction forces, which are hard to estimate. According to Karabanov et al. (1998), such “ice rafted” debris are released to the lake bottom when the ice melts in late spring/early summer. Francus and Karabanov (2000) observed evidences, that aggregates of coarser grains melt through the ice during winter when the ice is bare of snow. These aggregates probably reach the bottom sediments as frozen grain lumps (Francus and Karabanov, 2000).

As the clay content is high, the content in silts (63- 2µm) is low during cold periods. Moreover silts are fine during these periods (mode around 9 µm). The lowest mode is systematically observed at the end of cold periods (Figure VI.4) in Lake Baikal sediments. To date, we may suggest that local winds were mostly at a minimum. Occasionally increased wind dynamics occurred during winter time as revealed by the presence of coarse particles in glacial sediments.

Besides temperature, humidity is another important aspect of the climate evolution. Karabanov (2000b) uses diatoms and chrysophyts to estimate moisture changes in the catchment area of Lake Baikal. However, monitoring studies of primary production in the lake have shown that besides light availability, biological factors, like predator conditions (Jewson and Granin, 2000) are very important. Recent studies show that primary producers in the lake react to biological factors rather than moisture changes in the hinterland directly (e.g., Mackay et al., 2003; Ryves et al., 2003). Hence, the terrestrial vegetation is a direct and far more sensitive indicator for the reconstruction of moisture conditions than limnological proxies. Therefore moisture indices developed from pollen analyses are used for interpretations in the following two sections.

VI.4.2. Variations in the grain size during Termination 1 and the Holocene

The ending of the last glacial is well pronounced in the clay content and the coarse detrital fraction as both fractions decrease significantly at about 13.5 ka BP (Fig. VI.5). The decrease of the coarse fraction is probably related to an increase of the snow cover in the hinterland. Indeed increased accumulation results from local transport of coarse particles by wind as proposed in the previous chapter. Yet fewer particles become airborne at the source because of the snow cover.

At 13.5 cal. ka BP, the quantity of silts and their size increases and displays a high variability. This trend is interrupted at about 10 ka BP when the moisture availability diminishes as shown in the moisture index based on vegetation changes (Demske et al., submitted). It shows, therefore, that there is a link between moisture and silt input at the Continent Ridge located in the North Basin. However, the moist period, as established from vegetation data, indicates a later onset than reported from other studies (Karabanov et al., 2000b). Indeed, following the humidity index of Demske et al. (submitted), humidity increase only at ~ 13.5

cal ka BP. Karabanov et al. (2000b) who used diatoms, chrysophyts and spicules of sponge assemblages as climatic proxies, date the beginning of moist conditions at ~ 18 ^{14}C ka BP. Part of this discrepancy could originate from dating problems. Karabanov et al. (2000b) based his age model on ^{14}C dating performed on bulk organic carbon (Colman et al., 1996) while the ^{14}C data used in this study and in Demske et al. (submitted) are produced from pollen extracts (Piotrowska et al., in press). The latter proved to be an excellent media for dating as time offsets can be ruled out (Piotrovska, person. comm., 2004). Hence, we attribute this discrepancy to the age model of Karabanov (2000). As changes in the lakes bioproduction (Karabanov et al., 2000b) are more or less coeval, assuming that the error of the *AMS*^{14}C dating based on TOC is minimal (Colman et al., 1996), this indicates that shortly before Termination I the system recorded a fundamental change in dynamics. So far it is unclear what caused the change and which season and proxy were most affected by climatic changes. Evidently, diatom and chrysophyt assemblages are more affected by changes during the late spring and summer seasons and are, therefore, rather controlled by water temperature. As insolation increases strongly around 17 ka BP (e.g. the recent synthesis on Milankovitch cycles by Kukla and Gavin, 2004) this could have trigged changes in the lake's bioproduction. As to the detrital input, it is controlled by both eolian and fluviatile dynamics, which highly depend on the regional climatic conditions and dynamics. Evidently both are coupled to insolation changes too. As hydrology and wind patterns change significantly during different seasons, these factors obviously control aerosol and river discharge during autumn and summer storms or during snow melting in late spring and flooding during summer.

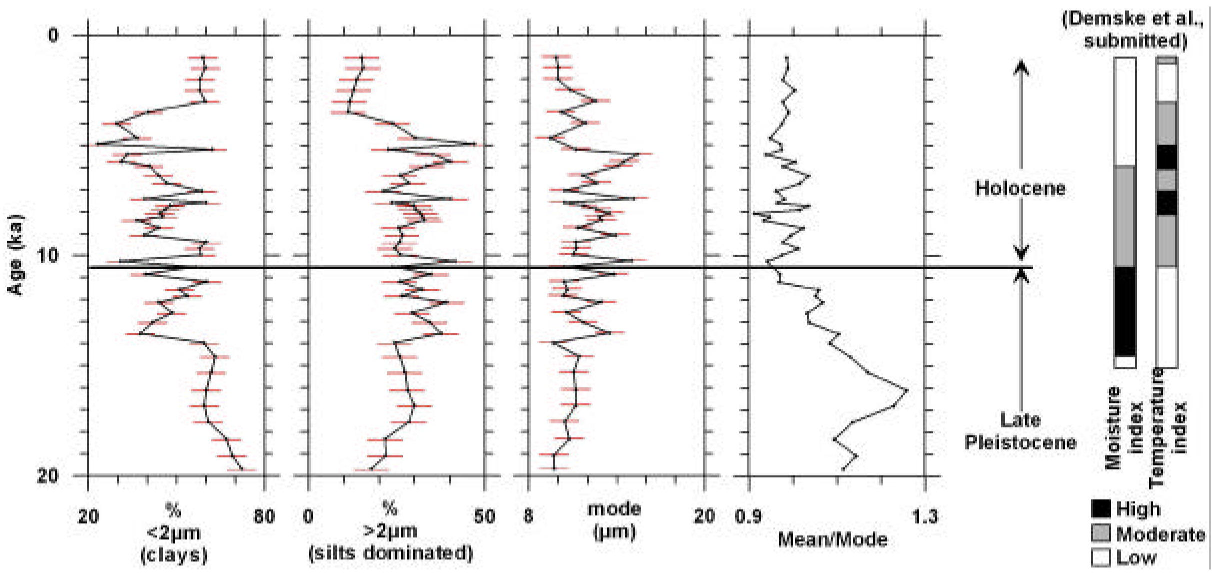


Figure VI.5: Focus on the last 20 ka for the sedimentary column CON 01-603-2. The dating of the first 15 ka is based on calibrated AMS ^{14}C. Percentage of clay (particles < 2µm) with error bar, of the detrital fraction > 2 µm (and its mode) with error bars and modal/mode. These are compared with moisture and temperature indices estimated by Demske et al. (submitted).

From 10 cal. ka BP to 3 cal. ka BP, the quantity and variability of silts increase again until a strong decrease at 3 cal. ka BP, which is marked by colder conditions in the temperature index as based on the vegetation (Demske et al., submitted). At the same time we also observe an increase in clay content yet another characteristic for cooler conditions as discussed earlier. The moisture index is also very low during this period. It seems, therefore, that during the Holocene the mechanism responsible for delivering high input and leading to a coarsening of silts is weakened when dryer and colder conditions occur.

This cooling phase, here recorded by higher clay content and a minimum of the silt size after 3 cal ka BP is known from earlier studies of Lake Baikal sediments (Karabanov et al., 2000b). This is in good accordance with high resolution pollen analyses performed by Demske et al.

(submitted) who showed that after the warm optimum (3.7-2.9 cal. ka BP) of the Holocene, cooler conditions are recognized in pollen assemblages.

VI.4.3. Termination II and the Kazantsevo

Here the grain size data are also compared with moisture and temperature indices established by palynologists in the same core (Granoszewski et al. 2004) (Fig. VI.6). A decrease in clay and coarse particle contents is observed at the transition between the glacial and the Kazantsevo interglacial from 135 to 127 ka which roughly corresponds to Termination II, which is dated in marine records as covering the interval 135 to 125 ka (e.g. Gouzy et al., 2004). This interval is also characterised by a high amount of silt and coarsening of the silt size fraction. This change is matched by an increase of moisture as deduced from the vegetation pattern. Once more, the period of high moisture corresponds to a strong decrease in the quantity of coarse particles depicted in the mean/mode ratio. Termination II and I show similar evolutions in grain size pattern, probably linked to comparable changes in the hydrology at the transition from glacial into interglacial.

Above Termination II, the size and quantity of silts decrease. The low quantity of silts, which is characteristic for the Kazantsevo, is interpreted as resulting from the optimum of vegetal development, in accordance with palynological data (Granoszewski et al. submitted). This vegetal development in the hinterland would blanket potential detrital sources and would hamper a reworking of silt particles. The resolution of the present study does not allow for detection of the mid-Kazantsevo cooling (at ~ 120 ka) which is documented in many other studies from Siberia (e.g. Karabanov et al., 2000a, Granoszweski et al., submitted; Rioual et al., submitted) and Europe (e.g. Rioual et al., 2001);

The shift towards a cold period (time equivalent to MIS 5d) starts at about 120 ka, with an increase in clay and sand and decrease in the quantity and size of silt. From the grain size characteristics, the paroxysm (cold maximum) of the stadial following the Kazantsevo is reached at about 113 ka, and shows a high clay content.

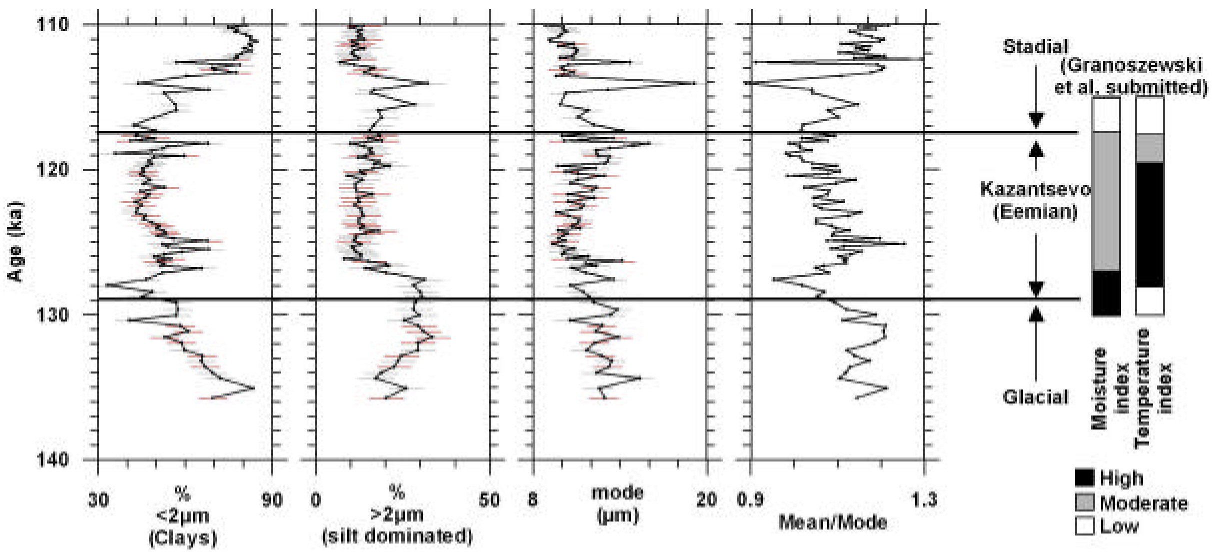


Figure VI.6: Focus on the window 110-140 ka (termination II, Eemian and MIS 5d) for the sedimentary column CON 01-603-2. From left to right, percentage of clay (particles < 2µm) with error bar, of the detrital fraction > 2 µm (and its mode) with error bars and modal/mode. These are compared with moisture and temperature indices estimated by Granoszeski et al. (submitted).

VI.4.4.Comparison with rock magnetic proxies

Figure VI.4 displays that in Termination I the ARM, i.e. the contribution of ferromagnetic minerals to the signal, gently decreases until the end of the high silt input (at about 3 ka). In Termination II, the ARM remains stable until a decrease in the quantity of silt. This observation highlights the problem of location of climatic boundaries in the sedimentary sequence using grain size proxies. If we consider the steep decrease of clay as the true climatic boundary, it does not correspond to the ARM decrease, which is progressive due to the presence of silts. This suggests that rock magnetic proxies can reflect the detrital input but without grain size discernment would lead to misidentification of climatic boundaries.

The high content of hematite-goethite (high HIRM) also highlights to a certain extent the cold periods (equivalent to MIS 5d, 5b, 4 and 2) while a low HIRM is characteristic of warm periods (equivalent to MIS 1, 5a, 5c and 5 e) (Fig. VI.4). This observation is not valid for the sediments equivalent to MIS 3, which shows a high HIRM. This interval is considered to be temperate rather than cold. as biogenic productivity e.g. (Swann et al., submitted) reaches a medium level. However, hematite-goethite is abundant while silt is the most common size fraction during this stage. Therefore, the hematite-goethite and silts abundances are not systematical indicators for cold periods. The increased hematite-goethite and silts input in sediments equivalent to MIS 3 may result from a local change in the sedimentation dynamics, which is not well understood yet.

The high amount of hematite-goethite in time equivalent MIS 3, somehow supports the previously proposed interpretation of this proxy, as until recently, hematite was used in Lake Baikal as the best proxy for eolian input during cold periods (Peck et al., 1994). However, based on Siberian loess records (e.g. Chlachula, 2003), far-distant eolian activity in time equivalent MIS 3, is known to be rather low. Thus the hematite, normally used as proxy reflecting far-distant eolian transport can also partly reflect local eolian redistribution. Moreover, time equivalent MIS 3 is known to have had a relatively warm and moist climate, similar to the present-day at least during an interval extending from 40 to 24 ka according to the archeological investigations carried out in the Baikal area (Chlachula, 2001).

VI.4.5. Teleconnections to the North Atlantic

At Terminations I and II, we have seen that detrital input changes: increased size and quantity of silts are linked to increased moisture around Lake Baikal. This increase of moisture seems not to be related to moisture produced in the nearest vicinity (Morley et al., submitted) and has to be placed in the context of North Hemispheric atmospheric changes. As the limit of Monsoon influence is located far to the south of Lake Baikal even in summer (e.g. Xiao and An, 1999), we considered that Lake Baikal is mostly influenced by Westerlies. The moisture transport towards Baikal region is therefore probably due to Westerly winds, which seem to strengthen at the transition from glacial to interglacial. This moisture coming from North Atlantic is known to originate from Gulf Stream activity bringing moisture from tropical to high latitudes. During glacial/interglacial transitions, the Gulf stream weakens because of fresh water input at high latitude resulting from ice melting. Moreover, the increased temperature of surface water implies more moisture formation (see, for the mechanisms, Driscoll and Haug, 1998). In Western Europe, pollen records display an increase of moisture occurring in two steps at about 13.5 and 10 ^{14}C ka BP (Peñalba et al., 1997). Furthermore its transport toward Lake Baikal may have been more efficient. One driving process could be that at the end of glacial periods, the Fennoscandian ice sheet starts melting. This implies a decrease of the albedo and therefore a decay or more probably a shift of the high pressure system over this region. A redistribution of pressure cells, i.e. their decay or shift linked to global warming, allows to establish new pathways for Westerlies (Khotinsky, 1984), which possibly carries moisture more efficiently into Central Siberia.

In the Weichselian, our HIRM record, i.e. the quantity of Hematite record, displays many pronounced spikes. These spikes match with spikes observed in our mean/mode record, i.e. the abundance of sand record. Looking for a justification of these so interpreted aridity events, we observed that hematite and sand spikes correlated well with Ca^{2+} input events recorded in Greenland ice record GISP 2 (Fig. VI.7). The Ca^{2+} recorded in the ice cores, is known to be of eolian origin (Mayewski et al., 1997). The source for Greenland dust is known to be Central Asiatic deserts such as the Taklamakan (Kang et al., 2003), as demonstrated by isotope tracers (Biscaye et al., 1997). Lake Baikal is not on the trajectory of the dust transfer from Central Asia to Greenland, which is known to run over China (e.g. Xiao and An, 1999). The connection between both events is, therefore, indirect. From this observation, a question arises: what causes synchronicity between sand and hematite input events in Lake Baikal and dust input events in Greenland ice? An answer could be that the events could be related to simultaneous particle production resulting from dry and cold events over Central Asia and Siberia. These paleoclimatic conditions are probably due to an extremely developed Siberian High pressure system. This interpretation matches quite well with the observations for the last century. From these observations, it has been recently shown that intensification of Siberian High may strengthen the transport of dust aerosols to Greenland (Kang et al. 2003). Placed in a seasonal context, we consider that these periods are characterised by long and dry winters. The reduced humidity would favour the production of eolian particles, which would be mobilised during spring storm events, the finest articles would reach high atmosphere and would be brought eastward by the Jet Stream and could even reach Greenland.

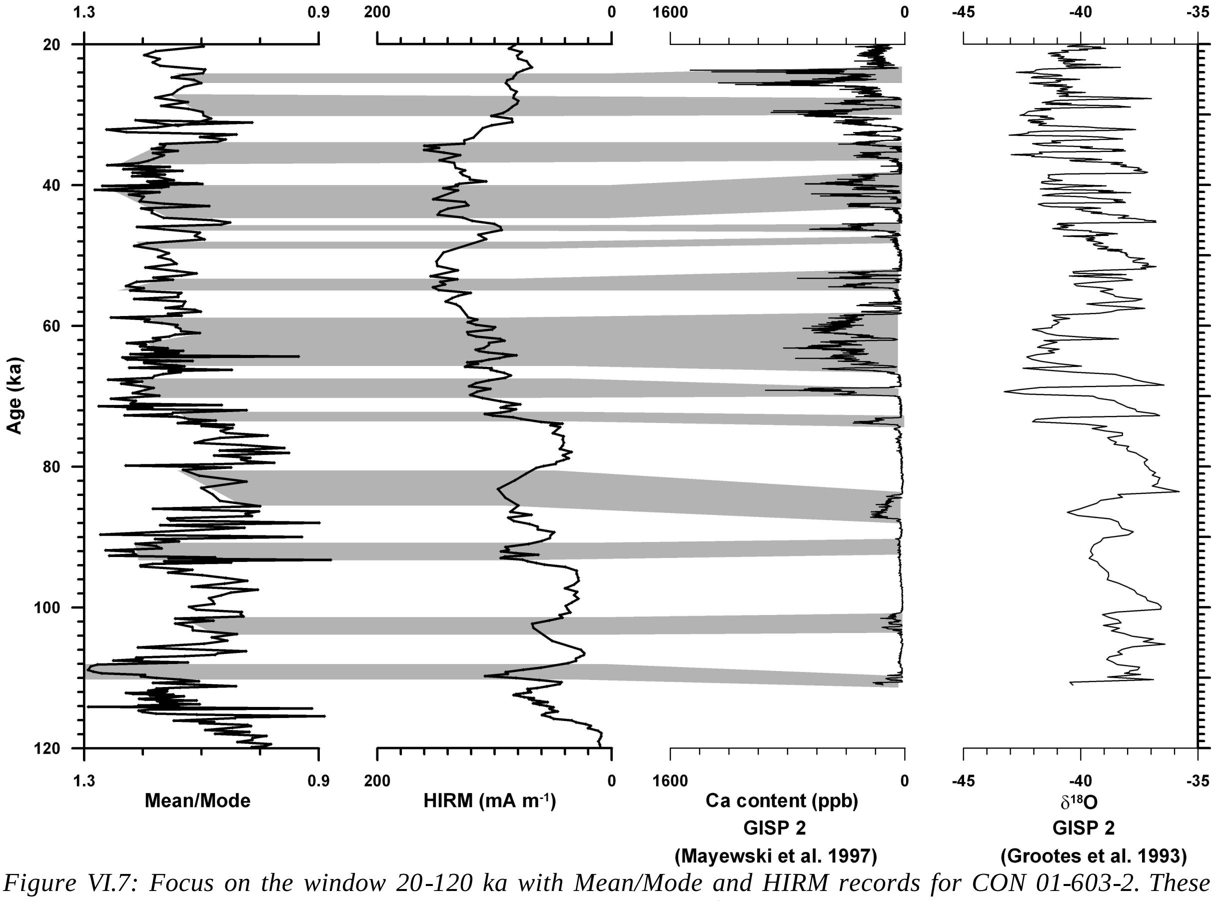


Figure VI.7: Focus on the window 20-120 ka with Mean/Mode and HIRM records for CON 01-603-2. These records are in face of Ca content (Mayewski et al., 1997) and $d^{18}0$ (Grootes et al., 1993) from GISP2. Shaded areas show the match between Ca^{2+} input in Greenland ice and sand and hematite input in Lake Baikal.

VI.5. Discussion

The climatic implications of the detrital input variation in Lake Baikal are well established in the present study with notably atmospheric teleconnections to North Atlantic. Some processes responsible for the input of fractions such as clay and sand are better constrained: Clays, abundant in cold-period sediments, and previously considered to originate from the mountain erosion by ice cover (Karabanov et al., 1998), are rather considered, in the present study, as resulting partly from (far-distant?) eolian input during winter, and partly from river input during the summer, when there is ice melting and permafrost thawing. The coarse particles, also abundant in cold-period sediments would result from local wind transport, bringing these particles onto the ice cover of Lake Baikal. Both fractions would reflect aridity of the hinterland. Nevertheless the mean of transport for silts particles, which are abundant during Terminations I and II, has to be discussed in more details. The high quantity of silts could be linked to melting events of mountain glaciers as soon as there is some warming (as seen in the diatoms by Karabanov et al., 2000b). Resultant rivers would bring the fine detritus to our site. The following question needs to be answered: Can silts be transported by flow currents shallower than 300 m depth to the Continent Ridge? The nearest shore is located at 12 kilometres eastward of Continent Ridge. Throughout different climatic periods, water level change did not exceed 10 metres (Back and Strecker, 1998). This change little affects the detrital assemblage at Continent Ridge since the horizontal shore line displacement has always been quite restricted as the coastal slope is steep. This coast is cut by many small rivers, which probably bring a lot of silt particles into Lake Baikal. If the settling of the silts follows the Stokes law under ideal conditions, i.e. without the influence of a water current, for ideal particles (concerning shape and grain size) and for an un-layered water column, the silts would settle in the basin located between the Continent Ridge coring site and the shore. Nevertheless, the physical characteristics of the Lake Baikal water body are complex and many parameters have to be considered. The Coriolis force dominates underwater currents in Lake Baikal basins (Botte and Kay, 2002). These currents form counter-clockwise cells, which would drive the particles from the shore to the NE, away from the Continent Ridge. In addition, vertical movements of water are also important when ice breaks up in spring. At this time, the slightly warmer water (approx.4° C) close to the shore plunges deeply, allowing for ventilation of the Lake Baikal bottom water (Weiss et al., 1991). This sinking water would facilitate the downward transfer of particles from the surface directly to the basin. One additional parameter controlling the migration of particles is the biogenic productivity. Indeed recent studies have shown a direct relationship between the speed of downward fluxes for particles and biogenic productivity (Bory et al., 2001). Organic carbon facilitates aggregation of particles in faecal pellets and biofilms, which results in faster settling. This implies a rapid settling of the particles especially during interglacials and climatic transitions, when higher biogenic productivity is present in Lake Baikal. All these parameters could reduce the influence of river input at Continent Ridge.

Besides river input, another way of bringing detrital particles to the deposition site could be to trap them in the fresh snow present on the ice cover of Lake Baikal whenever there is moisture and related snowfall. Then the particles are released when the ice cover melts. Present day observations emphasize the occurrence of such a process: grain-size distribution of detrital particles trapped in the fresh snow has been measured in the present study. The mode of the distribution is at about 11.3 µm that is in the range of fine silt. In the context of higher moisture, the availability of particles for wind transport would be decreased and restricted to certain area. Nevertheless detrital particles trapped in fresh snow are necessary for an eolian origin. This mechanism for silt input into Lake Baikal would explain the good match between increased silt input and increased moisture recognised in the sediment: it would implicitly record the presence of snow on the ice cover of Lake Baikal.

VI.6. Conclusion

This integrated study, coupling laser-assisted grain size and rock magnetism helps to better constrain the different processes responsible for the detrital input in Lake Baikal.

Drought is the main factor responsible for eolian particle production and availability of sediments for wind transport during cold periods. During these periods, the detrital input is mainly composed of clay and sand. From our discussion, it became apparent that clays may have been transported principally by wind throughout the year, and only partly brought by mountain glaciers melting in summer. Sandy particles observed in glacial sediments trace the aridity near the shore, where they get reworked more easily when the hinterland has a minimal snow cover. This sand is possibly whirled by local winds and transported by saltation on the ice cover of Lake Baikal. A local process is responsible too for the detrital assemblage during warm optima of the last interglacial (Kazantsevo), which shows low silt content due to strong vegetal development blanketing the potential sedimentary sources.

The mode of transport for the silt fraction, which increases along with increased moisture in Termination I and II, may be diverse and is not yet well constrained. Nevertheless, we highlighted that part of the silt input could be attributed to eolian transport and fresh snow trapping, this interpretation being based on present day observations. Many arguments are in favour of a limited influence for fluviatile activity on detrital input at the study site but this has to be supported by sediment trap observations.

The present study shows that detrital proxies have strong climatic implications with notably obvious teleconnections between Siberia and the North Atlantic. A detrital signal mainly reflecting atmospheric circulation patterns seems to be observed in our hemipelagic sediments during cold periods and at climatic transitions Terminations I and II. Firstly, in the Weichselian, sand and hematite input events occur simultaneously with dust events bringing eolian particles from Central Asia to Greenland ice. These events would mark cold and dry events in central Asia and Siberia probably related to extension of the Siberian High Pressure Cell over especially long winters. Secondly, in transition such as Termination I and II, the increase in silts and the decrease in sand and clay highlight higher sedimentary dynamics related to an increase of moisture over Lake Baikal. This increased moisture would result from new westerly pathways related to the Fennoscandian and Siberian high pressure system decay.

In the present study rock magnetic parameters as indicators for detrital input is further used: ARM variation in Termination I and II show that ARM is influenced by both clay and silt fractions. This observation shows that using ARM as a climatic marker could lead to misidentification of climatic boundaries. Only HIRM remains a good marker for detrital input as spikes can be related to dust events. Nevertheless HIRM shows highest values in sediments equivalent MIS 3 as also evident in the silt content. These results are not understood yet but contradict the previous use of HIRM, which was proposed to be the best marker for cold periods (Peck et al., 1994). Indeed, many studies show that time equivalent MIS 3 was relatively warm and moist in the Lake Baikal region (Swann et al. submitted).

VII. Conclusions and Perspectives

In this thesis we acquired and interpreted data of different topics such as paleomagnetism and rock magnetism and sedimentology (grain size analysis). Combining these different methods provides tools for (1) decrypting the climatic variations as recorded in Lake Baikal sediments and (2) constraining the timing of the changes.

The multidisciplinary approach of magnetic mineralogy i.e. using rock magnetic data combined with other investigations, such as TEM, XRD, XRF, allows us to better define and understand the processes responsible for rock magnetic assemblages. With this combined approach the relative influences of the detrital input and post-depositional processes have been further characterised showing the following results:
Fine magnetite generally dominates the magnetic signal. This signal is high in glacial sediments, with a predominance of magnetite of detrital origin, while it is diluted in interglacial sediments by a decrease in detrital input and an increase of opal. The contrast of the magnetic signal intensity mimics the lithological variations and therefore can be used for correlating cores. However, the present rock magnetic study shows that early diagenesis affects the magnetic mineralogy in specific intervals. On the one hand, low susceptibility in interglacial sediments can be partially due to magnetite dissolution. On the other hand fine spikes of high susceptibility are observed at glacial/interglacial transitions and randomly distributed in glacial sediments. These are due to diagenetic neoformation of greigite. Post-depositional dissolution and mineralization preserve information on the paleo-redox state of the sediment which, at least partly, is modulated by changes in sedimentation rates. Besides being of interest for understanding diagenesis, characterising post-depositional processes is important since this allows us to identify a miscorrelation of magnetic susceptibility. In addition post depositional processes directly affect the quality of the paleomagnetic records. The rock magnetic study shows that filtration of the data to establish the paleomagnetic record is necessary since the homogeneity of rock magnetic properties is interrupted by effects of the early diagenesis. The principle of filtration, which consists in excluding intervals affected by diagenesis from the paleomagnetic records, should be systematically applied to future paleomagnetic investigations on Lake Baikal sediments.

Our paleomagnetic approach was innovative in Lake Baikal: instead of using a correlation of climatic events to date paleomagnetic records, we used paleomagnetic correlation to date climatic events. The establishment and correlation of the paleomagnetic records with reference curves from ODP Site 984 (Channell, 1999) and MD 95-2024 (Stoner et al., 2000), provide high-resolution age models for 6 sedimentary columns from Lake Baikal. These age models were anchored by *AMS* ^{14}C dating (Piotrowska et al., in press) but also by geomagnetic excursions such as Laschamp and Iceland Basin events, documented in this thesis as a strong directional deviation and as a full reversal, respectively. Using these age models, we have shown that, at the MIS 6/7 transition, climatic records from Lake Baikal sediments assess a Central Asia cooling synchronous with the surface water temperature change recorded in planktonic $\delta^{18}O$ measured in North Atlantic sediments. Consequently, the Baikal region cooling occurred earlier than the global ice volume change, which is recorded in benthic $\delta^{18}O$ profiles several thousand years after the sea surface cooling revealed by planktonic $\delta^{18}O$ profiles. This implies misdating when the age model, as it is classically done in Lake Baikal, is based on correlation between paleoclimatic profiles from Lake Baikal sediments and benthic $\delta^{18}O$ profiles from marine sediments. Therefore, a paleomagnetically based age model provides a much more robust chronology, which is now used by the Continent project members and which will be hopefully used in future studies.

We also established a stack of different relative paleointensity records. This stack, which covers the last 200 ka is a brand new paleomagnetic reference curve for Central Eurasia, circa 50 ° N. It can be used for geodynamo models and provides information on intensity and directional variability. Two geomagnetic excursions are preserved in Lake Baikal sediments and there documentation will contribute to constrain the geodynamo behaviour during such events. The paleomagnetic reference curve we established in this study is the longest high-resolution paleomagnetic record ever established for Lake Baikal. Therefore, it will be used as a reference for correlating not only at a regional scale but also at a global scale. The paleomagnetic database will be complemented by records established for other Baikal cores by other CONTINENT partners, Petr Pruner and Jaroslav Kadlec of the Institute of Geology of the Academy of Sciences of Czech Republic, Prague. The goal is to compile paleomagnetic records of equivalent resolution in order to create an extended paleomagnetic reference curve of several hundred thousand years for Central Eurasia. In addition, radionuclide measurements (^{10}Be, ^{36}Cl), reporting on the cosmic flux rates which seem to be related to the paleointensity field (Wagner et al., 2000; Beer et al., 2002), could confirm the reliability of the relative paleointensity records. This has been successfully done, for instance, on Portuguese Margin sediments (Thouveny et al., 2004; Carcaillet et al., 2004) and could be used as a climatic proxy. Indeed previous low-resolution radionuclide measurements (^{10}Be) revealed the great potential of Lake Baikal sediments for such investigations (Horiuchi et al., 2001).

The age model that ensued from paleomagnetic correlation was used to date detrital input profiles measured directly (grain size analysis) or derived fom rock magnetic parameters. Our grain size analysis approach is complex and tedious since it consists of separating the sediments into different fractions like clay, opal and particles >2µm via multiple steps. This separation of the different detrital fractions as well as the laser-assisted measurement of the grain size distribution on the detrital fraction >2µm improved our conceptual understanding of the influence of different transport processes and yielded insights into climatic implications not necessary revealed by biogenic proxies. With the discussion concerning grain size results we concluded that we had to reconsider some earlier statements about detrital input to Lake Baikal. We consider, for instance, that the high clay sedimentation during cold periods mainly results from eolian input, rather than, as proposed earlier, erosion by glaciers. Some lines of thought are proposed which indicate that, during cold periods, the glaciers are rather stable and the melting period is quite restricted in time. Though glacial erosion occurs, it is most probably less important than previously assumed. We further propose that the sand faction is an excellent tracer for aridity: We observed that this fraction is more abundant during glacial and cooler periods than during warmer intervals. Sand grains and small rock pebbles are known to be "ice rafted" debris and transported on the ice cover of Lake Baikal by local winds. We propose that they mainly enter the sedimentation realm when there is little snow on the ice and at the shore. Comparison with vegetal cover reveals that minimum silt size and silt quantity in the Eemian marks the vegetal optimum when the potential sources for detrital material are blanketed. We observed that the silt distribution is a very sensitive proxy during climatic transitions. Silt size and silt quantity increased during Terminations I and II indicating an increase of the sedimentary dynamics linked to an increase of moisture as reconstructed by the palynologists (Demske et al, submitted and Granoszewsky et al., submitted). Nevertheless the identification of transport processes for this silt fraction is still under discussion as it can result from glacial and fluvial as well as eolian input. From preliminary results, it was apparent that eolian sediments could be the important source since we observed that the silt which is trapped today in fresh snow on the ice cover on Lake Baikal has a similar grain size as the silt measured during the Terminations. However, this needs further work.

Links between Lake Baikal and North Atlantic regions are observed for cold periods, when dry and cold events are characterised by sand and hematite input in Lake Baikal which are synchronous with dust input in Greenland Ice. This may point to teleconnections between Central Eurasia and North Atlantic events. Moreover the increase of moisture observed during Terminations I and II are probably due to new pathways for the Westerlies resulting from a decay of the Fennoscandian high pressure system linked to ice sheet melting but also due to a reorganisation of flow patterns in Central Siberia. Seasonally, this modification is materialised by shorter winter due to increased insolation. Nevertheless some open questions remain: a minimum in silt size, for instance is systematically observed at the top of cold periods. This observation cannot be explained as yet.

The characterisation of the grain size distribution of selected detrital fractions shows the potential of this proxy for paleoclimatic studies. Nevertheless, the preparation protocol has to be improved since some artefacts linked to the formation of pellets sporadically alter the quality of the detrital signal. Using fresh sediments instead of freeze-dried sediments should prevent the formation of pellets. Further grain size investigations, and improvements in lab protocols where necessary, are being undertaken by Matthias Zopperitsch (University of Potsdam).

Besides the improvement of the preparation protocol, the detrital input study has to be complemented by other proxies such as Nd-Sm isotopes. Their measurements will inform on the relative influence of different detrital sources (Matthias Zopperitsch, in prep.). The synthesis of this combined work could qualify the first interpretations for grain size results and answer open questions on the dynamics of detrital input in Lake Baikal.

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