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

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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., this issue), 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 Eurasia.

Keywords: Lake Baikal - paleomagnetism - relative paleointensity - geomagnetic excursions – age model - Late Quaternary

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 Eurasia.

2. Materials 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. 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. 1), far from turbiditic influences (e.g. Charlet et al., this issue). 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 Posolskoe Bank, and the Vydrino Shoulder (Table 1, Figure 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., this issue) 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., this issue). 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 2).

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 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., this issue).

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:

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

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., this issue).

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

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

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., this issue). 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. (this issue), HIRM is a climatic proxy, revealing the detrital (perhaps eolian) input in Lake Baikal sediments.

The ratio of SIRM intensity to low field susceptibility ($\mathrm{SIRM}/\kappa_{LF}$) was calculated in order to monitor the possible presence of greigite (an authigenic iron sulphide) already evidenced in Lake Baikal sediments (Demory et al., this issue). A high $\mathrm{SIRM}/\kappa_{LF}$ 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).

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. 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. 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., this issue). 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. 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.

4. Results

4.1. Rock magnetism

A detailed discussion on rock magnetic analyses is given in Demory et al. (this issue). 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. 5).

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. 4A, 4B and 4C), the demagnetisation step with 30 mT was sufficient for removing the viscous overprint from the NRM.

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. 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. 5 outer right), is unstable in sediments of Lake Baikal

(Fig. 2 and Demory et al., this issue). 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. 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.

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 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. 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. 4B). Samples from diatomaceous sediments are often characterized by a weak and unstable magnetisation (Fig 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. 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. 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. 6B and C). 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 Fig. 7, together with the correlation scheme introduced in Fig 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.

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.

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. 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.

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. 9).

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., this issue) 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., this issue). 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 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.

5. Discussion

5.1 Lake Baikal responses 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. 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. 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 ca. 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 Eurasia 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.

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 Eurasia 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.

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 Eurasia. 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. 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.

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.

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Charlet et al. this issue

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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 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 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.

Figure captions

Figure 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.

Figure 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 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 $\delta^{18}O$ 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 $\delta^{18}O$ record represent marine isotope stages (MIS). (Data available at: doi:10.1594/GFZ/ICDP/CON/2004….)

Figure 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}.

Figure 5. Three down-core variations of normalised relative paleointensity after diagenetic correction and κ_{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). (Data available at: doi:10.1594/GFZ/ICDP/CON/2004….)

Figure 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 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.

Figure 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 $\delta^{18}O$ record from ODP 677 (Shackleton et al., 1990), see also Figure 3. Diagenetic features such as dissolution of magnetite and

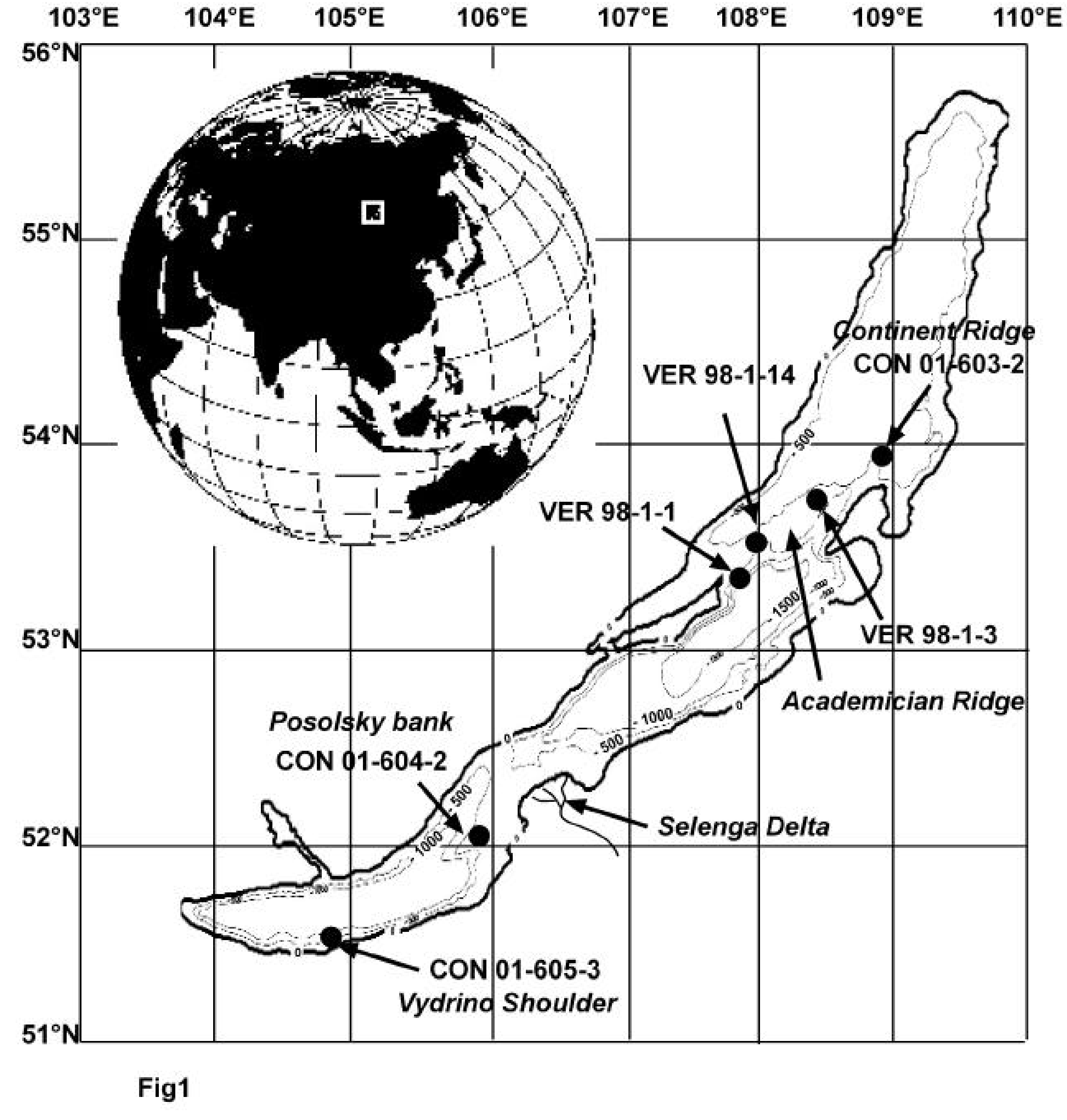
mineralization of greigite are marked according to Figure 4. Laschamp and Iceland Basin excursions are indicated as La. and Ic., respectively. (Data available at: doi:10.1594/GFZ/ICDP/CON/2004….)

Figure 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). (Data available at: doi:10.1594/GFZ/ICDP/CON/2004….)

Figure 9. Diagram of depth versus age based on relative magnetic paleointensity correlations (Fig. 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. (Data available at: doi:10.1594/GFZ/ICDP/CON/2004….)

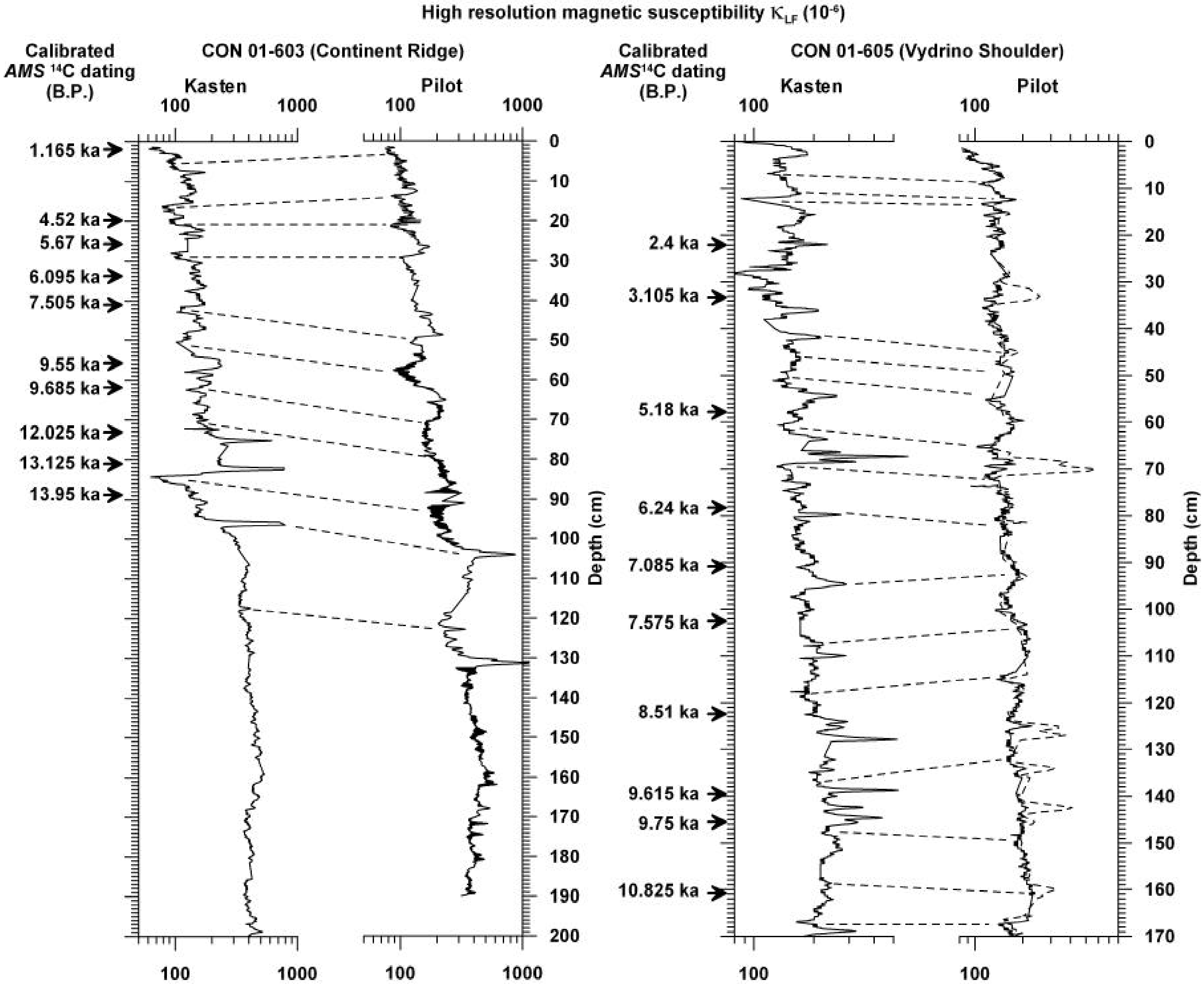
Figure 10. This figure provides a focus on the 160 ka-200 ka time window. Here the relative paleointensity record of CON 01-603-2, of ODP Site 984, (Channell, 1999), i.e. its reference curve for dating, as well as the relative paleointensity record from ODP Site 983 (Channell et al., 1997) are plotted 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.

Figure 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.

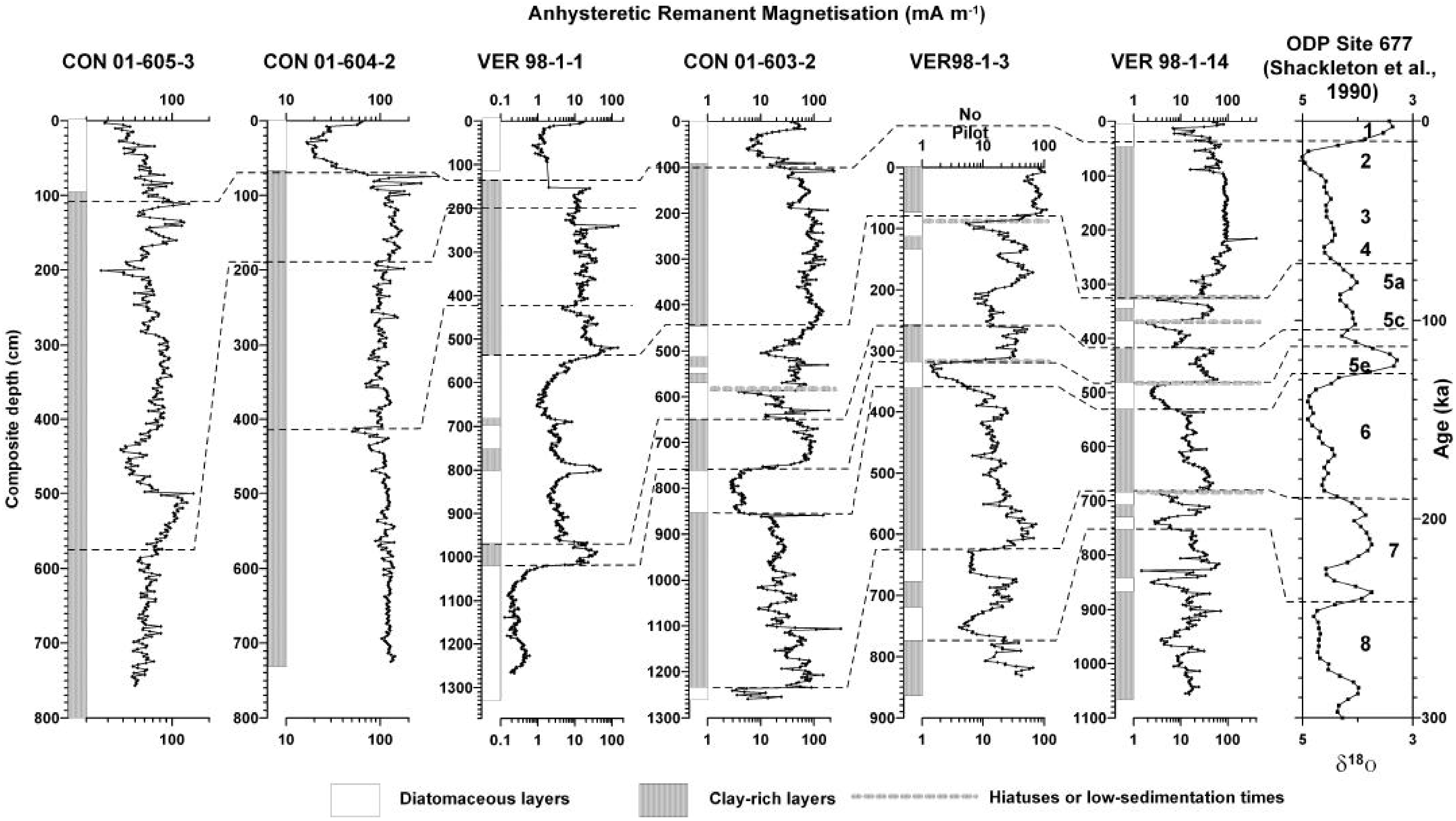


103°E
104°E
105°E
106°E
107°E
108°E
109°E
110°E
56°N
55°N
54°N
53°N
52°N
51°N
Continent Ridge
CON 01-603-2
VER 98-1-14
VER 98-1-1
VER 98-1-3
Academician Ridge
Posolsky bank
CON 01-604-2
Selenga Delta
CON 01-605-3
Vydrino Shoulder
500
1000
1500
0

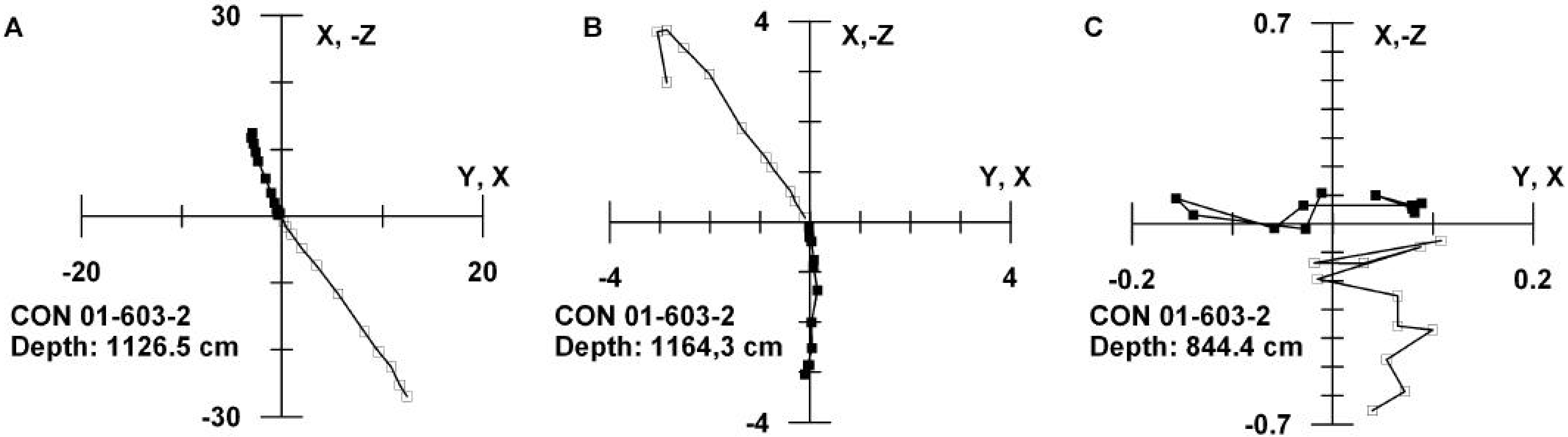
Fig1

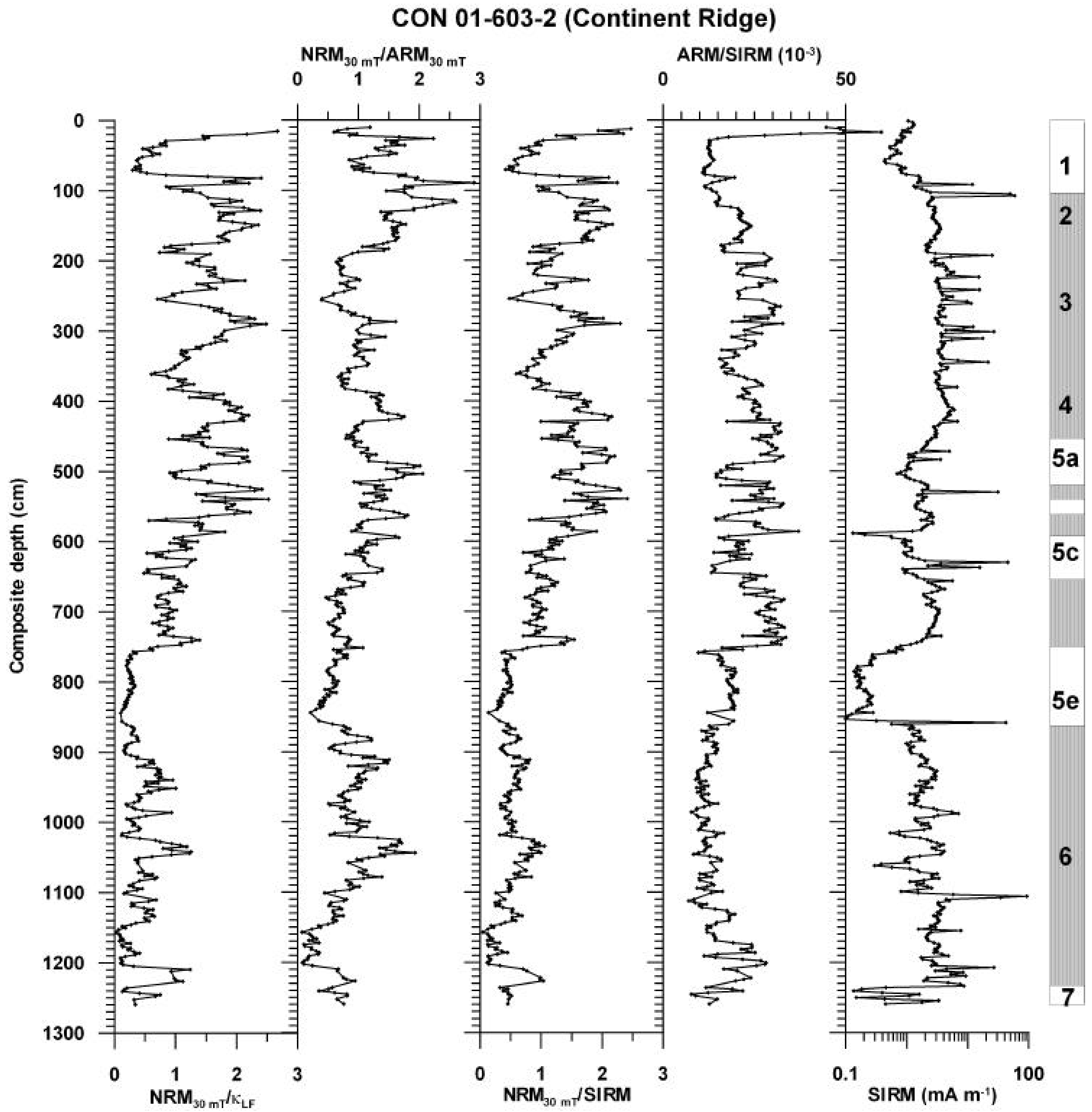


High resolution magnetic susceptibility κ_{LF} (10^{-6})
Calibrated AMS ^{14}C dating (B.P.)
CON 01-603 (Continent Ridge)
Kasten
Pilot
1.165 ka
4.52 ka
5.67 ka
6.095 ka
7.505 ka
9.55 ka
9.685 ka
12.025 ka
13.125 ka
13.95 ka
Depth (cm)
CON 01-605 (Vydrino Shoulder)
2.4 ka
3.105 ka
5.18 ka
6.24 ka
7.085 ka
7.575 ka
8.51 ka
9.615 ka
9.75 ka
10.825 ka

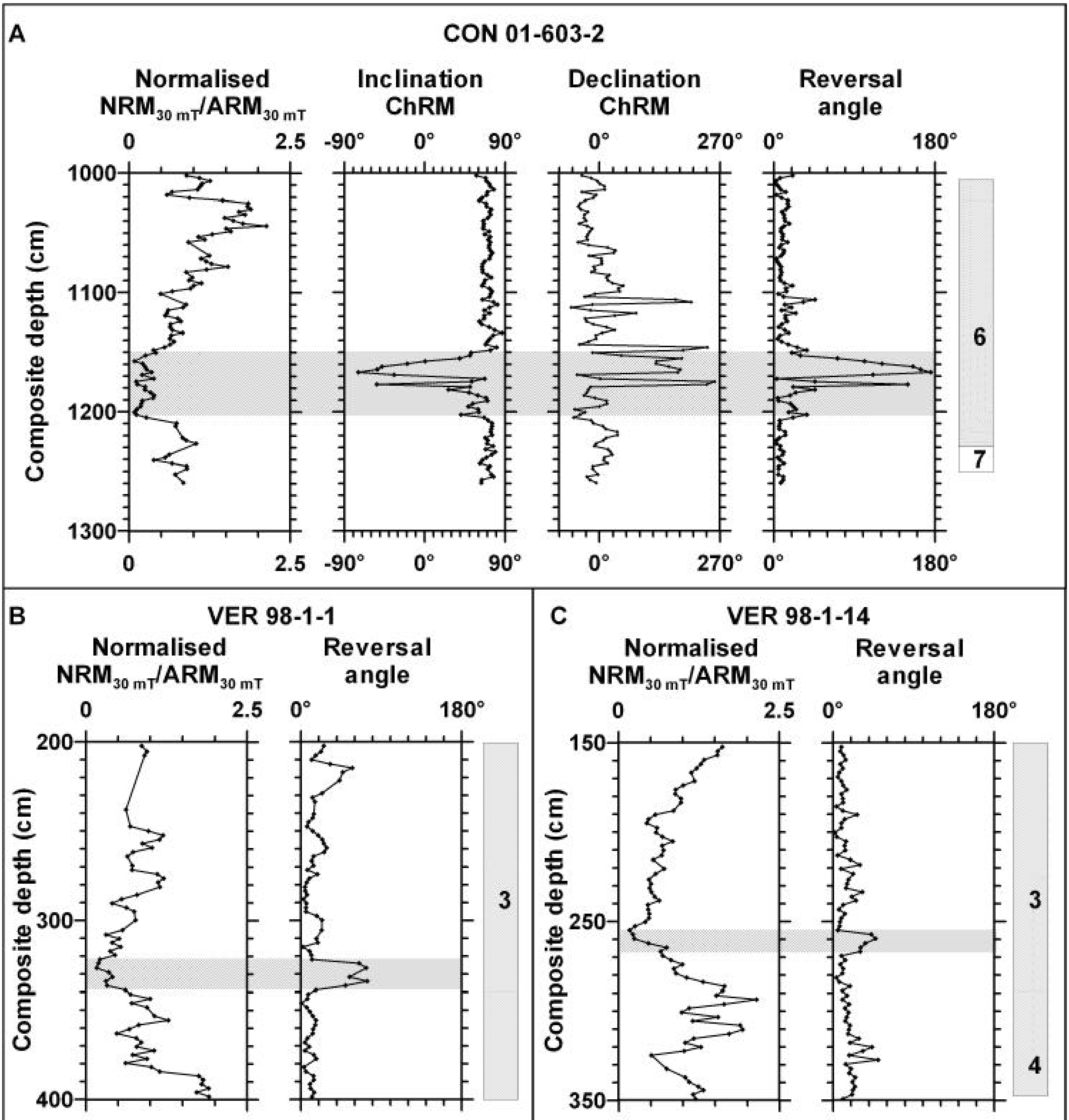


Anhysteretic Remanent Magnetisation (mA m-1)
CON 01-605-3
CON 01-604-2
VER 98-1-1
CON 01-603-2
VER98-1-3
VER 98-1-14
ODP Site 677 (Shackleton et al., 1990)
No Pilot
Composite depth (cm)
Age (ka)
δ18O
1
2
3
4
5a
5c
5e
6
7
8
Diatomaceous layers
Clay-rich layers
Hiatuses or low-sedimentation times

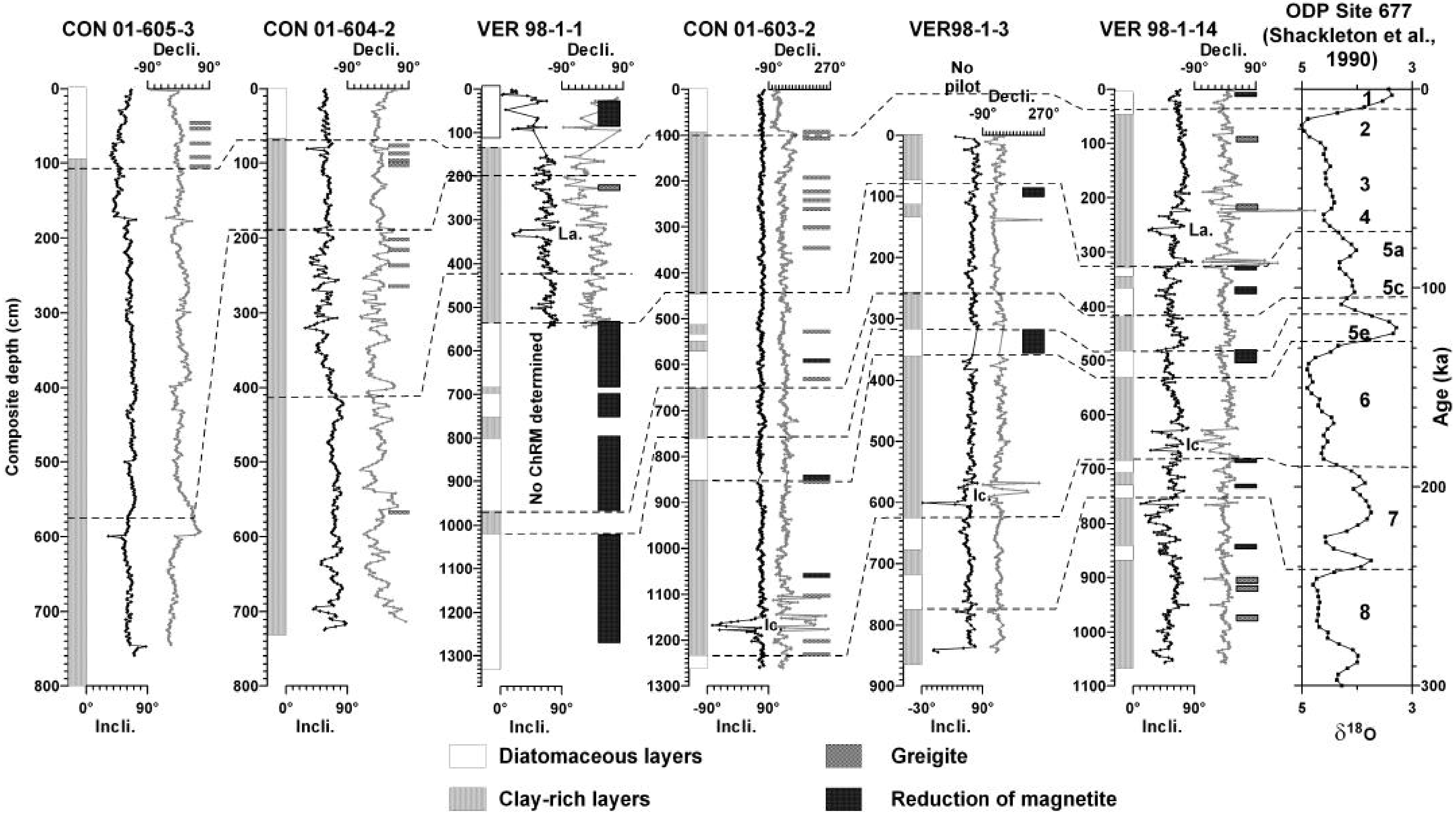

A
30
X, -Z
Y, X
-20
20
-30
CON 01-603-2
Depth: 1126.5 cm
B
4
X,-Z
Y, X
-4
4
-4
CON 01-603-2
Depth: 1164,3 cm
C
0.7
X,-Z
Y, X
-0.2
0.2
-0.7
CON 01-603-2
Depth: 844.4 cm



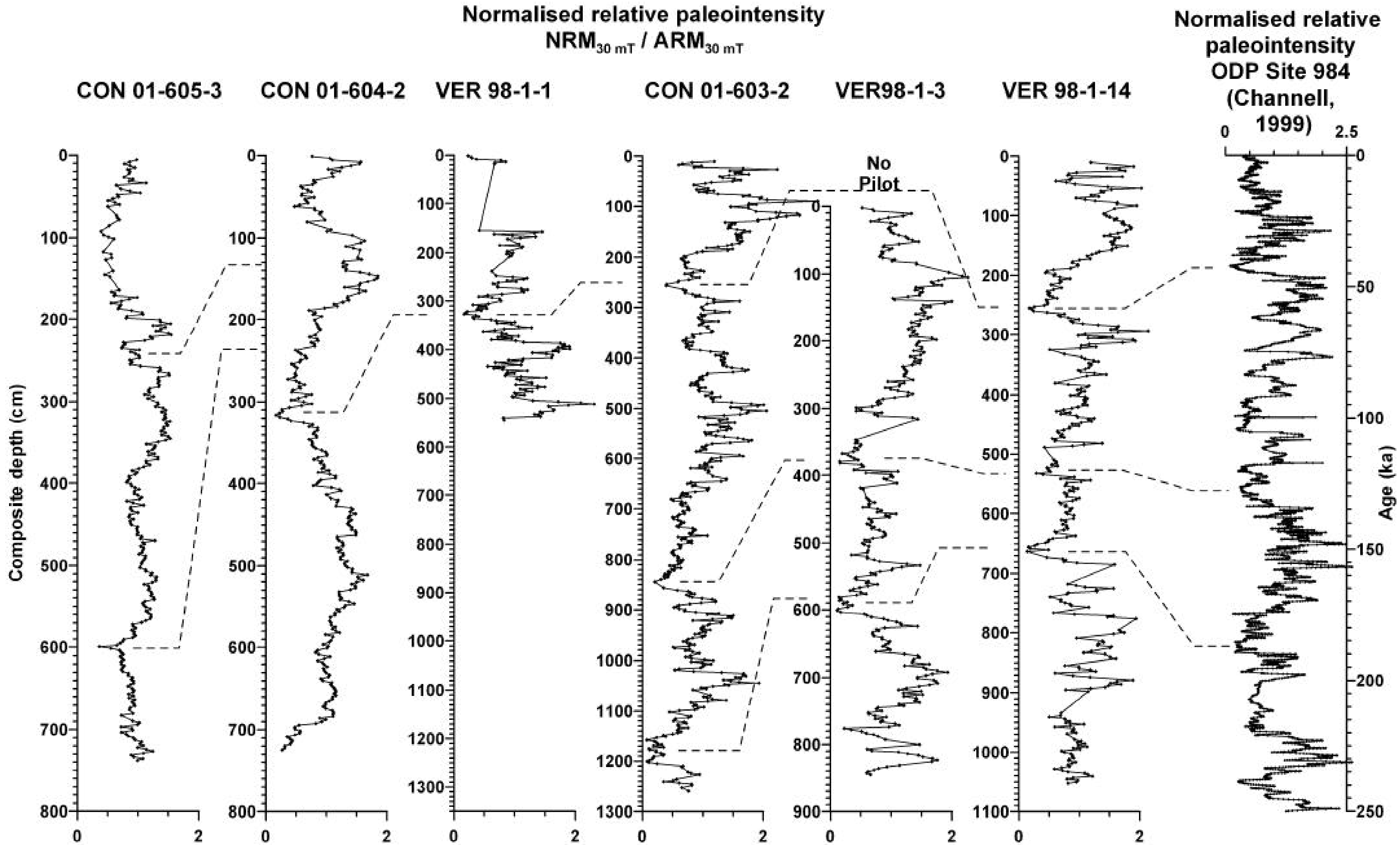
CON 01-603-2 (Continent Ridge)
NRM30 mT/ARM30 mT
ARM/SIRM (10-3)
Composite depth (cm)
NRM30 mT/κLF
NRM30 mT/SIRM
SIRM (mA m-1)
1
2
3
4
5a
5c
5e
6
7



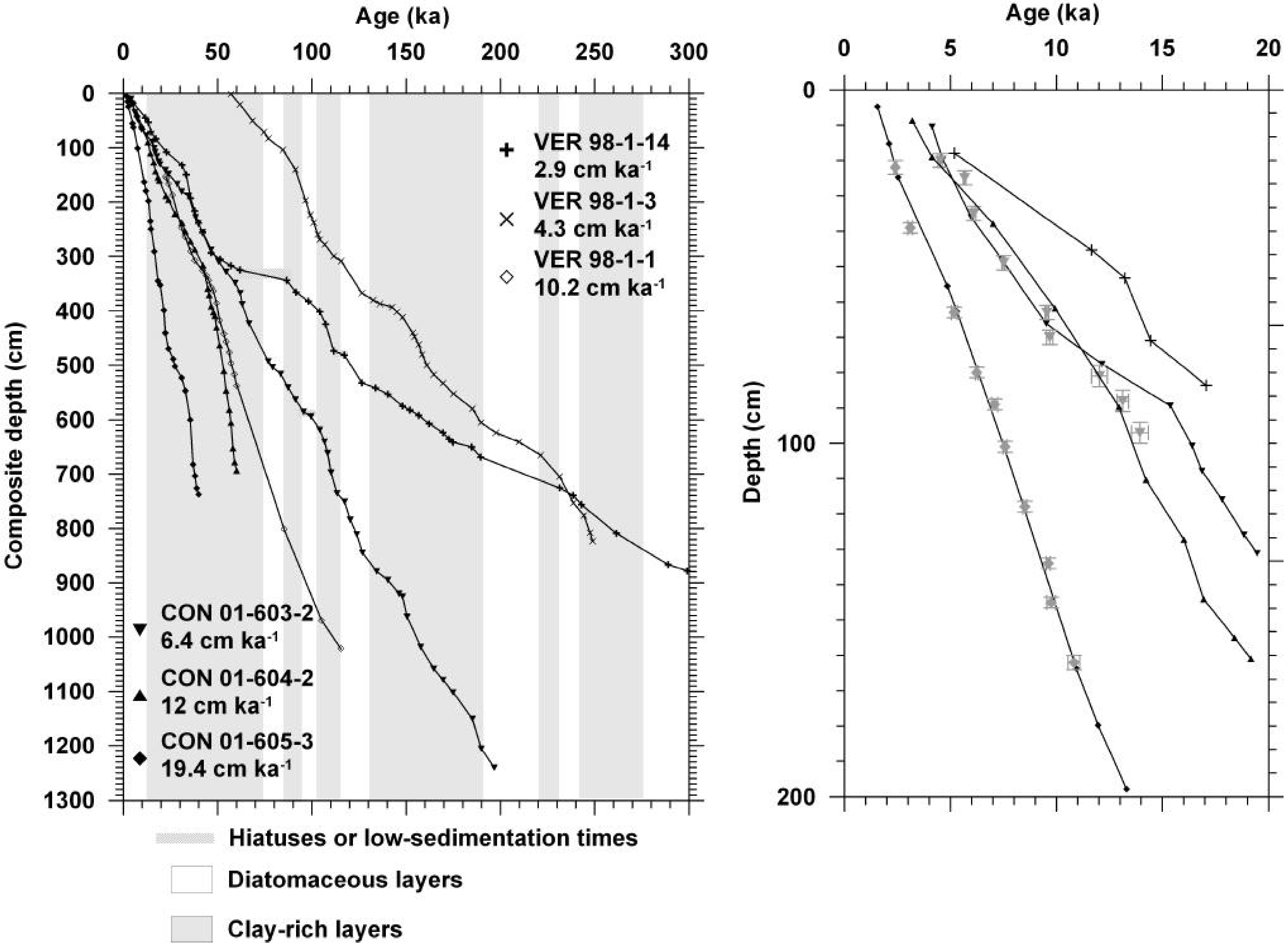
A
CON 01-603-2
Normalised
NRM30 mT/ARM30 mT
Inclination
ChRM
Declination
ChRM
Reversal
angle
0
2.5
-90°
0°
90°
0°
270°
0°
180°
Composite depth (cm)
1000
1100
1200
1300
6
7
B
VER 98-1-1
Normalised
NRM30 mT/ARM30 mT
Reversal
angle
200
300
400
3
C
VER 98-1-14
Normalised
NRM30 mT/ARM30 mT
Reversal
angle
150
250
350
3
4



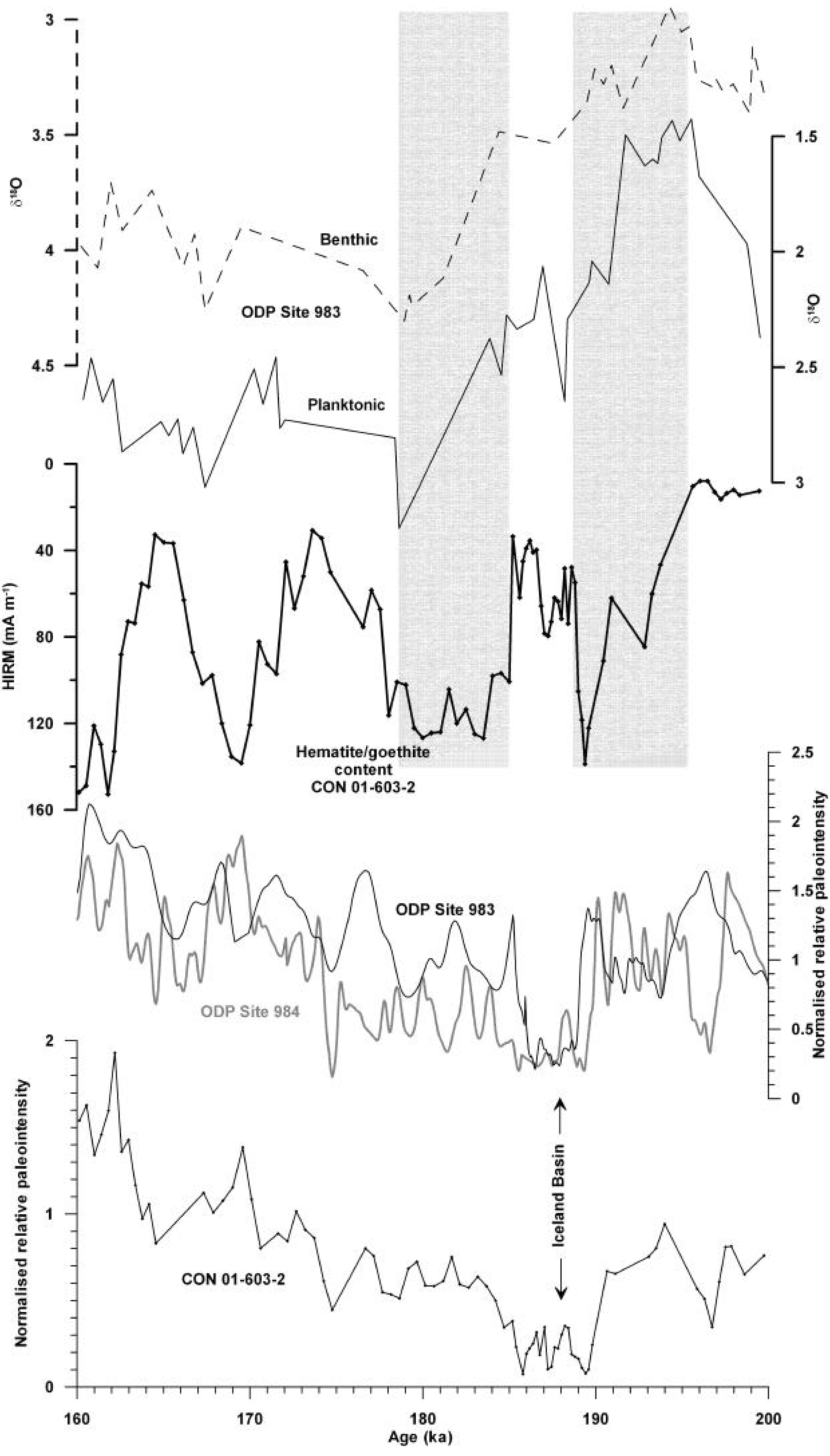
CON 01-605-3
CON 01-604-2
VER 98-1-1
CON 01-603-2
VER98-1-3
VER 98-1-14
ODP Site 677 (Shackleton et al., 1990)
Decli.
Incli.
Composite depth (cm)
Age (ka)
δ18O
No pilot
No ChRM determined
La.
Ic.
1
2
3
4
5a
5c
5e
6
7
8
Diatomaceous layers
Clay-rich layers
Greigite
Reduction of magnetite



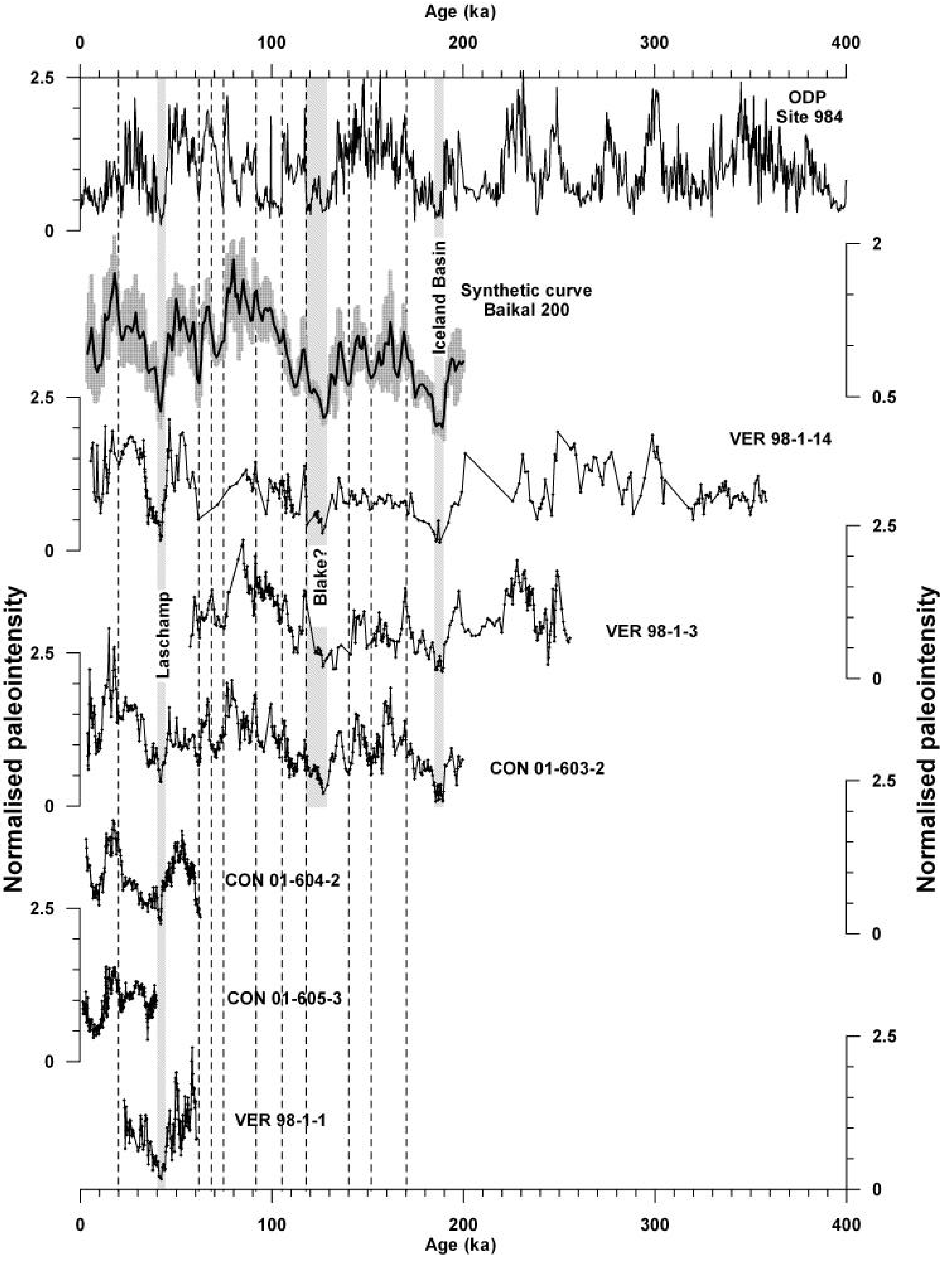
Normalised relative paleointensity
$NRM_{30\ mT}$ / $ARM_{30\ mT}$
CON 01-605-3
CON 01-604-2
VER 98-1-1
CON 01-603-2
VER98-1-3
VER 98-1-14
Normalised relative paleointensity ODP Site 984 (Channell, 1999)
No Pilot
Composite depth (cm)
Age (ka)



Age (ka)
Composite depth (cm)
VER 98-1-14
2.9 cm ka-1
VER 98-1-3
4.3 cm ka-1
VER 98-1-1
10.2 cm ka-1
CON 01-603-2
6.4 cm ka-1
CON 01-604-2
12 cm ka-1
CON 01-605-3
19.4 cm ka-1
Hiatuses or low-sedimentation times
Diatomaceous layers
Clay-rich layers
Age (ka)
Depth (cm)



δ¹⁸O
Benthic
ODP Site 983
Planktonic
δ¹⁸O
HIRM (mA m⁻¹)
Hematite/goethite content CON 01-603-2
ODP Site 983
ODP Site 984
Normalised relative paleointensity
Iceland Basin
CON 01-603-2
Normalised relative paleointensity
Age (ka)



Age (ka)
0
100
200
300
400
2.5
0
ODP
Site 984
2
0.5
Synthetic curve
Baikal 200
Iceland Basin
2.5
0
VER 98-1-14
2.5
0
Blake?
Laschamp
VER 98-1-3
2.5
0
CON 01-603-2
2.5
0
CON 01-604-2
2.5
0
CON 01-605-3
2.5
0
VER 98-1-1
Normalised paleointensity
Normalised paleointensity
0
100
200
300
400
Age (ka)