Geophys. J. Int. (1997) **129**, 75–94

Magnetostratigraphy and astronomical calibration of the last 1.1 Myr from an eastern Mediterranean piston core and dating of short events in the Brunhes

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Accepted 1996 November 14. Received 1996 November 10; in original form 1994 November 14

SUMMARY

A 37 m long piston core (KC-01B) was collected from the Calabrian Ridge in the Ionian Sea. Astronomical calibration of sapropels and sapropelic signals—based on rock-magnetic and geochemical properties—yields a very accurate time-frame for the last 1.1 Myr, with the exception of the interval (650–850 ka) straddling the Brunhes/Matuyama boundary. We used the oxygen isotope data as an additional constraint on our age model.

We arrive at an age of 812 ka for the Brunhes/Matuyama boundary, which indicates that its position is determined by delayed acquisition of NRM by some 40 kyr. The age of the top of Jaramillo subchron (1001 ± 3 ka) is slightly older than that of Shackleton, Berger & Peltier (1990); that of its bottom (1072 ± 5 ka) concurs with their estimate.

In the Brunhes Chron, four short reversal excursions (CR0–3) were found and dated (CR0 261 ± 3 ka, CR1 318 ± 3 ka, CR2 515 ± 3 ka, CR3 573 ± 3 ka). In all cases, a high-coercivity reversed component is overlapped to a varying degree with a low-coercivity normal component. Three of the reversal excursions (CR1–3) occur in zones where diagenetic processes were shown to be minimal. The fourth (CR0), in which a very large coercivity overlap occurs, is not well expressed and is found in a zone where some dissolution of magnetic minerals has occurred.

The existing literature on reversal excursions is reviewed, and some of the older records with an oxygen isotope or sapropel record have been redated. CR0 could correspond to one of the Fram Strait excursions (Nowaczyk *et al.* 1994). CR3 corresponds to the Emperor, which is shown to be equivalent to the Big Lost reversal excursion. For CR1 and CR2 we propose the names Calabrian Ridge 1 and Calabrian Ridge 2, since correlation with existing reversal excursions is uncertain. The Blake was not detected because extensive diagenesis has occurred in the corresponding interval. All reversal excursions observed in KC-01B correspond to periods of minimal (relative) palaeointensity in the independently dated record of Valet & Meynadier (1993). This seems to imply that reversal excursions are more likely to occur in periods with a relatively large non-dipole field contribution

Key words: Brunhes, geomagnetism, magnetostratigraphy.

1 INTRODUCTION

During oceanographic Cruise MD69, organised within the framework of the European Union scientific program MAST (Marflux and Palaeoflux), a long piston core labelled KC-01B with a length of 37 m was taken from the Calabrian Ridge (36°15.25′ N, 17°44.34′ E), at a water depth of 3643 m. The aim of the Marflux project is the reconstruction of biogeochemical cycles in the Mediterranean and eastern North Atlantic, within a high-resolution biostratigraphic and magnetostratigraphic framework. Core KC-01B has been chosen as the reference core for the Ionian Sea by the Marflux participants, not only because it is one of the longest piston cores ever taken in the Mediterranean, but also because it contains the oldest sapropels (dark organic-rich layers) obtained so far in a piston core. Hence, it narrows the gap between the piston-core sapropels and those found in land-based sections. The oldest sapropel in piston cores is S12 in core[1]RC9–181 (Ryan 1972; Vergnaud-Grazzini, Ryan & Cita 1977), with an age of $\approx$464 ka based on an astronomically derived monsoon index (Rossignol-Strick 1983), or 481 ka based on the correlation with precessional cycles (Hilgen 1991), which was later refined to 483 ka based on correlation with the insolation target curve (Lourens *et al.* 1996). The youngest one in a landsection is sapropel *v* in the Crotone section (Zijderveld *et al.* 1991; Hilgen 1991), with an age of 1280 ka (Lourens 1994; Lourens *et al.* 1996). Core KC-01B was sampled for magnetostratigraphic purposes to provide age constraints for the high-resolution biostratigraphy, using interpolation between astronomically calibrated sapropel ages (Hilgen 1991) and reversal ages of the Brunhes/Matuyama and Jaramillo subchrons (Shackleton *et al.* 1990). The resulting calcareous nannofossil biochronology is discussed by Castradori (1993).

During the magnetic measurements, several short events in the Brunhes Chron were found. The global recognition and identification of these events is severely hampered by the lack of an accurate and reliable age control. We decided to date the observed events in core KC-01B by astronomical calibration, rather than trying to correlate them with known events with often uncertain age, and thus avoid contributing to the 'reinforcement syndrome' (Watkins 1968). The calibration as reported in this paper involves the matching of sapropels and sapropel patterns—which form an expression of climate induced by precession and eccentricity (see Hilgen 1991)—to the astronomical solutions. The strong sapropelitic association of the anhysteretic remanent magnetization (ARM) and some geochemical parameters (S, Ba) facilitates the recognition of sapropelitic signals in cases where an actual sapropelitic lithology is not apparent. Furthermore, we add a discussion on many short events or 'reversal excursions' (Merrill & McFadden 1994) in the Brunhes Chron. We assign new ages to several of them based on a reinterpretation of the original data (including oxygen isotope stages and sapropel patterns). Our astronomical calibration also provides new and independent astronomical ages for the upper and lower Jaramillo boundaries, while the position of the Brunhes/Matuyama boundary (at 812 ka)—on the basis of the oxygen isotope record of planktonic foraminifera from core KC-01B—has clearly suffered from delayed NRM acquisition.

1.1 Lithology and sampling

Core KC-01B consists of hemipelagic sediments with intercalated sapropel and tephra layers. They form an alternation of grey, greenish, olive-coloured, yellowish, white and beige shades. Sapropel layers are black to dark green, and at several levels black spots or brownish olive colours are found which may point to a hardly developed or a re-oxidized sapropel (de Lange, Middelburg & Pruysers 1989; Pruysers *et al.* 1991, 1993; Higgs *et al.* 1994). The sediments represent different depositional and palaeoredox conditions, ranging from oxic-suboxic (beige and white) via suboxic-anoxic (grey) to anoxic (sapropels, black). These cyclically varying conditions govern early diagenetic processes in which dissolution, migration and subsequent precipitation of elements like Fe, Mn and Ba may occur. This may have important consequences for the (delayed, post-depositional) formation of magnetic minerals, as was shown for the Pliocene marls of Sicily and Calabria (Van Hoof & Langereis 1991; Van Hoof *et al.* 1993). Following this earlier work, multivariate statistical methods (fuzzy c-means clustering and non-linear mapping) were used by Dekkers *et al.* (1994) to analyse early diagenetic effects in core KC-01B, using rock-magnetic and geochemical data.

After recovery, the piston core was cut into 1 m segments and split lengthwise into two halves. Both were stored under N_2 at 4 °C. One half was used for subsampling in Utrecht (the Netherlands), where the core segments were kept under these conditions until the actual sampling. The other half was intended for archive purposes. Palaeomagnetic samples (8 cm^3 perspex cylinders) were taken at 10 cm intervals, corresponding to an average resolution of $\approx$3000 yr; in total, 337 samples were taken. For every sample, NRM_{20} (natural remanent magnetization, subscript 20 denotes the intensity after alternating field demagnetization at 20 mT), ARM (29 μT DC bias field, 100 mT peak AC field) and initial susceptibility (χ_{in}) were determined, and a number of hysteresis loops and thermomagnetic runs were performed on selected (by colour) samples. At the same level as the palaeomagnetic samples, a sample was taken for geochemical analyses, including the determination of Ca (recalculated to $CaCO_3$), Ba, S and Mn. Rock-magnetic and geochemical data are discussed in detail by Dekkers *et al.* (1994); their multivariate statistical analysis shows that the sediments of the core fall into two categories: one is mainly related to diagenesis (dissolution-precipitation category), the other to depositional conditions (lithology category).

Oxygen isotope stratigraphy has been derived from the planktonic foraminifera *Globigerinoides ruber*, picked at 5 cm intervals as an average. Measurements have been performed on a Finnigan MAT 251 mass spectrometer, using the procedure described by Duplessy (1978). The oxygen isotope stages can be clearly recognized, and extend back to at least stage 31.

2 RESULTS

2.1 Demagnetization of NRM

The NRM was demagnetized by stepwise alternating fields (AF) of 5, 10, 20, 30, 40, 50 and 60 mT, and measured on a 2G superconducting magnetometer (Model 740-R). On the basis of the results from AF demagnetization of every sample in the lower half of the core, we fully demagnetized only alternate samples in the upper half of the core, whereas for the remaining samples the direction after demagnetization at 20 mT was taken. Only if there were anomalous (e.g. reversed) directions were the remaining steps up to 60 mT also carried out, in order to better determine the characteristic remanent magnetization (ChRM) direction.

In the vast majority (85–90 per cent) of samples, the interpretation of AF demagnetization diagrams is straightforward and yields a ChRM showing a univectorial decay to the origin for both normal and reversed samples (Figs 1a,b,d). Typically, a small (10–20 per cent of the total NRM intensity) and randomly directed component is removed at 5 mT, which

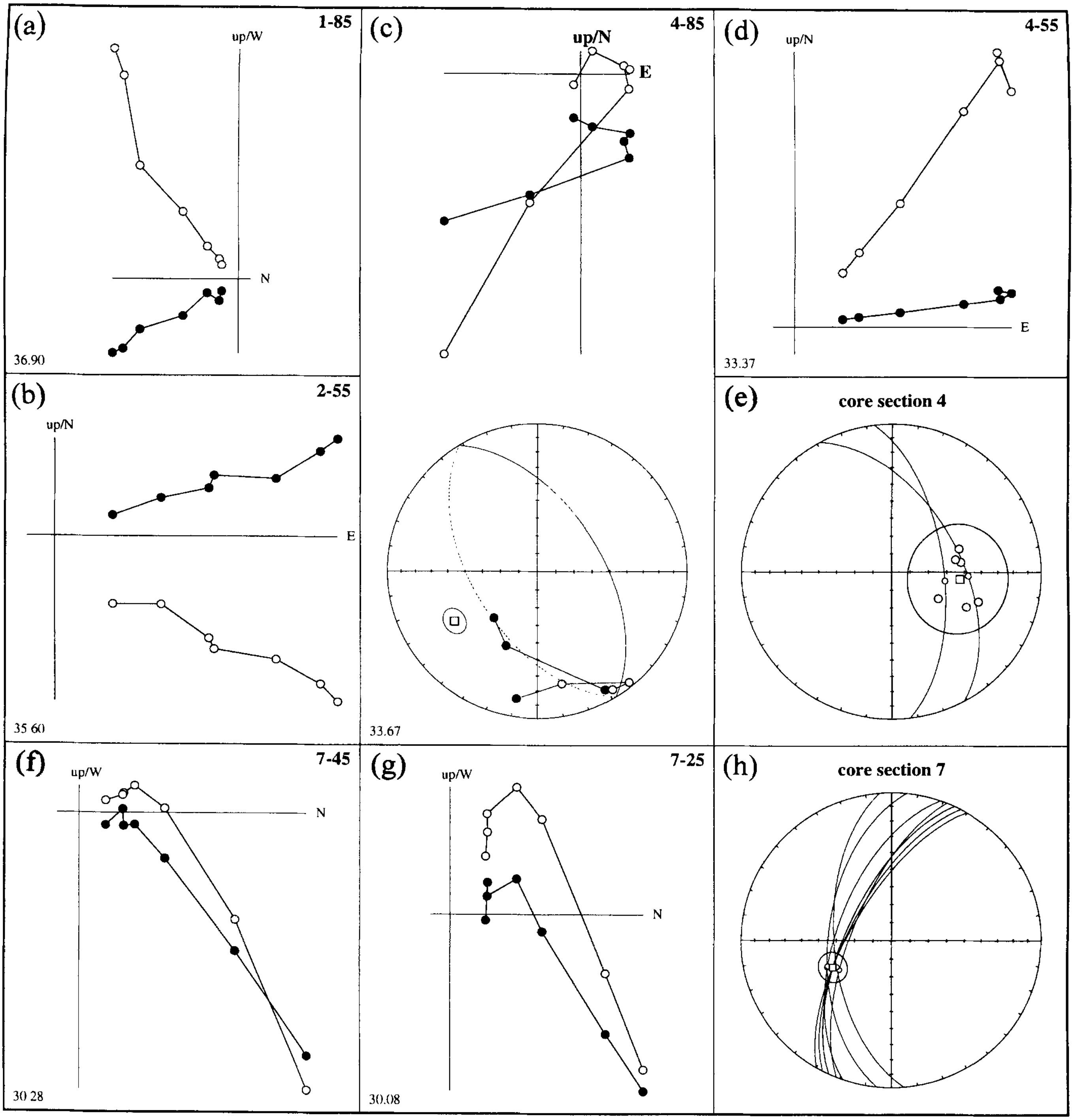


Figure 1. Examples of alternating field NRM demagnetization diagrams, using steps of 5, 10, 20, 30, 40, 50 and 60 mT. Open (closed) circles denote projection on the vertical (horizontal) plane. In several cases, a high coercive component is not well resolved, even though a reversed polarity is apparent (c,f,g). Occasionally, we have used great circles to approximate the reversed direction; in (c), the square and ellipse denote the normal to the great circle plane and its maximum angular deviation. Reliably determined directions (as in d) have been used as setpoints (large circles in e) to determine a best-fitting direction (small circles in e and h) from the great circles (after McFadden & McElhinny 1988). Lower left in each diagram: subbottom level.

we interpret as a soft laboratory-induced viscous component. Most of the NRM is removed at fields of 60 mT, which suggests magnetite as the dominant carrier of the NRM; this is confirmed by the rock-magnetic properties and the thermomagnetic treatment (Dekkers *et al.* 1994).

Starting at subbottom level 32.5 m up to 29.5 m, AF demagnetization increasingly shows the existence of two components (Figs 1c,f,g): a low-coercivity normal component, and a high-coercivity reversed component. The latter component is often not well resolved, in which case we determined the best-fitting great circle through the two components (Fig. 1c). Where possible, we used well-determined directions ('setpoints', like in Fig. 1e) in combination with the great-circle solutions from the same core segment for a great-circle analysis (McFadden & McElhinny 1988) to approximate the reversed component. In two core segments, we had to revert to great circles only (Fig. 1h), since no reliable setpoints could be determined, even though it is clear that there is a two-component NRM, with the high-coercivity component being reversed (Figs 1f,g).

The ChRM directions are shown in Fig. 2. The core segments were not oriented with respect to each other. After NRM demagnetization, each core segment was oriented using average values of declination and inclination to obtain an internally consistent record with normal (reversed) polarity having on average a declination of 360° (180°). In the topmost part of the core (2–4 m subbottom level) declinations within a core segment show a large but consistent divergence; here, the orientations of individual segments have been adjusted arbitrarily to obtain a smooth transition between adjacent core segments. This large swing in declination is interpreted to represent physical rotation of the (soft and water-saturated) sediment caused by the initial spinning of the piston core before it is halted by sediment friction.

The entire Brunhes Chron appears to be recorded; the Brunhes/Matuyama boundary occurs at a depth of 29.85 m. In addition, the Jaramillo subchron has been recorded, between 33.52 and $\approx 35.10 \pm 0.10$ m subbottom. This shows that core KC-01B contains a sedimentary record of approximately the last 1.1 Myr.

2.2 The occurrence of 'reversal excursions' during the Brunhes Chron

A (nearly) reversed direction has been recorded at four levels in the Brunhes Chron (Fig. 2). We informally label them CR0 (14.83 m subbottom level, CR = Calabrian Ridge), CR1 (16.55 m), CR2 (22.72 m) and CR3 (24.05 m). CR0 occurs adjacent to a core break and was initially not considered further. The three other levels are well away from a core break. To rule out a single deviating sample, we have taken additional samples at 2.5 cm intervals around these levels. The CR0 level was also subsampled at a later stage because the ChRM directions are generally very consistent across core breaks throughout core KC-01B (see Fig. 2). The demagnetization results of these additional samples confirm that there are indeed short intervals with a (nearly) reversed polarity (Fig. 3). In many cases, the small reversed component is largely overprinted by a low-coercivity normal component, and its direction can often not be reliably determined. The detailed records of the subsampled core segments (Fig. 4) show that the intervals with anomalous directions are small, some 10–20 cm.

The intervals containing CR1, CR2 and CR3 belong to the lithology category (Dekkers *et al.* 1994), implying that no or minor diagenesis has occurred. CR0 is allocated to a cluster where some dissolution has occurred. Apparently, reversed directions can still be preserved in such zones. We feel that these results are evidence of several non-normal intervals within the Brunhes Chron. In the reversed interval between Jaramillo and Brunhes, there are two levels that show a stable normal polarity (Fig. 2). These levels, however, are closely associated with either a sapropel or with a sapropel-like signal in the rock-magnetic and geochemical parameters. Therefore, we refrain from interpreting these levels as true normal polarity. More likely, they represent a Brunhes overprint through delayed NRM acquisition, related to early diagenetic precipitation of magnetite through Fe-migration (Van Hoof *et al.* 1993) close to or just above an anoxic (sapropel) layer (Dekkers *et al.* 1994; Higgs *et al.* 1994).

2.3 Rock-magnetic and geochemical signature of sapropels

The rock-magnetic parameters ARM and χ_{in} and the geochemical parameters Ba and S show a distinct signature with respect to sapropels (Fig. 5; see also Dekkers *et al.* 1994). Very low values of ARM (and also NRM_{20}) can be associated with dissolution of magnetic minerals just below a sapropel, while typically just above a sapropel precipitation of magnetic minerals may occur, explaining the often higher than average values of ARM (and NRM_{20}). Similarly, χ_{in} values are lower just below a sapropel, but minima are less distinct because changes in (ferrimagnetic) susceptibility are buffered by the paramagnetic contribution of the clay minerals. Ba and S show very conspicuous maxima in association with the sapropels. Sulphur maxima (predominantly pyrite) indicate anoxic conditions. Barium is a productivity proxy—it occurs as barite ($BaSO_4$). Under severely anoxic conditions it may become mobile again because of undersaturation with respect to barite as a consequence of sulphate reducing conditions; it precipitates again where conditions are more oxic (i.e. with higher sulphate concentrations; see Van Os, Middelburg & de Lange 1991). Therefore, it is an attractive element to trace so-called reoxidized ('burnt-down') sapropels: the organic material and the high sulphide content have disappeared because of oxidation, but the high Ba content has remained. This is clearly illustrated by van Santvoort *et al.* (1997), who studied several intervals with a sapropelic signature (high Ba, low ARM) in geochemical detail. They have shown that indeed many of these intervals represent sapropels or sapropelic intervals that have been entirely oxidized.

Because of its apparent sensitivity to anoxic conditions, ARM is also a good indicator of 'ghost-sapropels', even if no actual sapropelitic lithology is visible. For instance, ARM shows a strong minimum just below level S″ and a distinct maximum just above it (Fig. 5). Similarly, the level where we would expect S10 is characterized by distinct minimum values in ARM, NRM and χ_{in}. Sulphur maxima are also found at other levels, and are indicative of anoxic conditions and hence could indicate 'ghost-sapropels'. In the S and Ba profiles, for instance, there are indications for S3 and S10 (Fig. 5).

3 SAPROPEL CHRONOLOGY

Core KC-01B contains the oldest sapropels recorded in a piston core. The youngest sapropel (S1) was not recovered. S3 was not found either, although there are geochemical indications (high S, Ba) for a sapropel signal at ≈ 6.8 m subbottom, which could correspond to S3, while palynological and micropalaeontological data suggest a sapropelic interval at ≈ 6.85 m (Rossignol-Strick, personal communication 1996). Sapropels actually identified are S4–S9, S11 and S12. In addition to

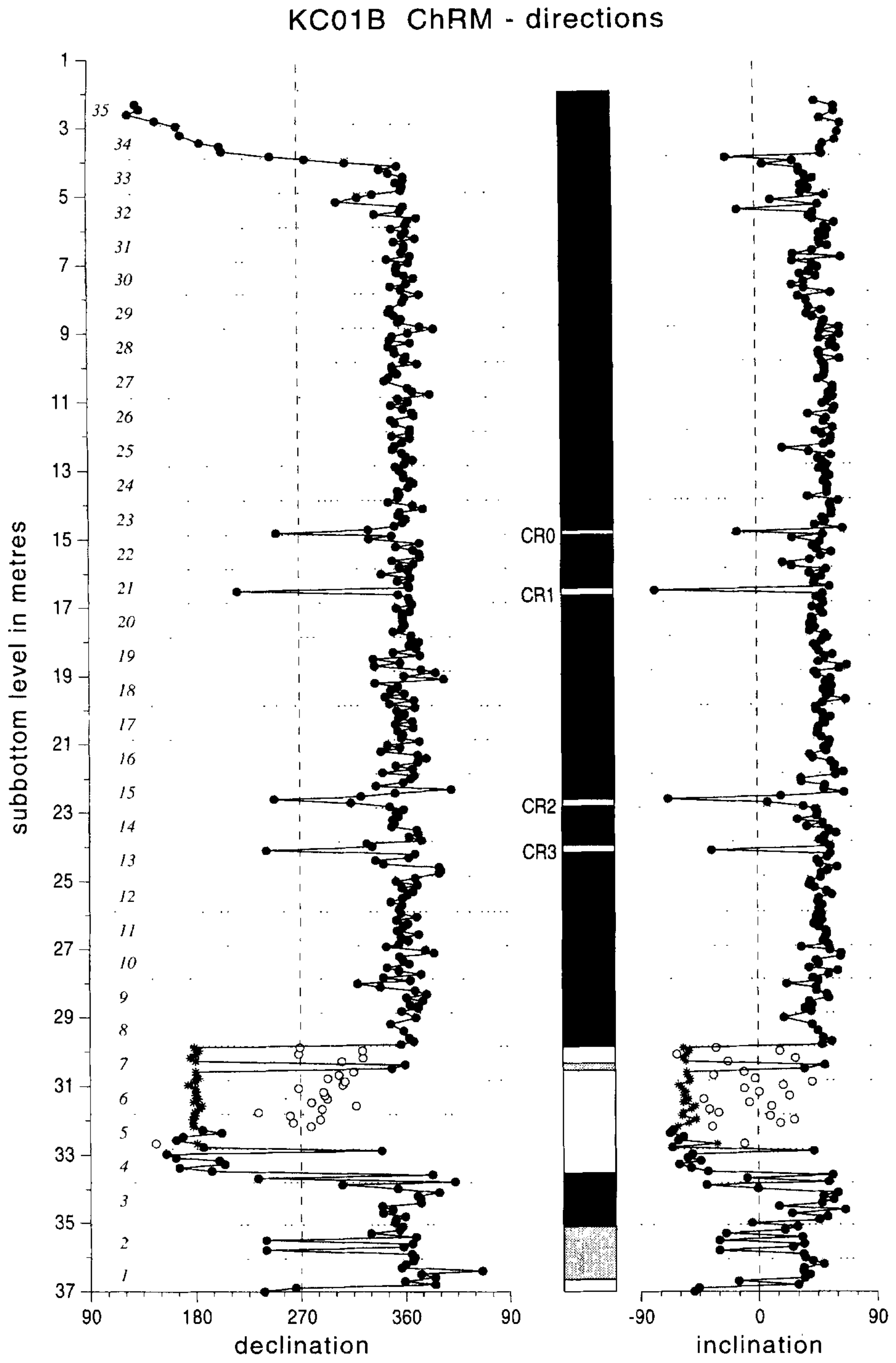


Figure 2. Declination and inclination of the characteristic remanent magnetization (ChRM) for core KC-01B versus subbottom level in metres, after rotation per core segment to a consistent (normal or reversed) direction. Solid (open) circles denote reliable (largely unresolved) directions; asterisks denote the great-circle solutions. Black (white) is normal (reversed) polarity; grey is uncertain polarity. At four levels a (nearly) reversed direction was found; they have been labelled CR0–3 (CR = Calabrian Ridge). Core segments (dotted lines) are numbered 1–35.

these, there are four older sapropels, here informally named Sa, Sb, Sc and Sd (Fig. 5). Furthermore, there are two levels with a clear lithological sapropelitic signature, S′ (dark brownish olive layer) and S″ (a layer with sapropel traces).

The oldest sapropel recorded before the sampling of KC-01B (S12) was found in core RC9-181 (Ryan 1972). This core was first dated by Wollin *et al.* (1971) by correlating the inclination record with that of core V12-122 for which a dated climate curve was already available (by extrapolation from a radiometrically dated horizon of 126 ka). Ryan (1972) used the same climatic method to date RC9-181, and concluded that indeed the inclination changes correlated well with those in V12-122, and that the inferred events (Blake, Jamaica and Levantine) were the same in both cores. The oxygen isotope

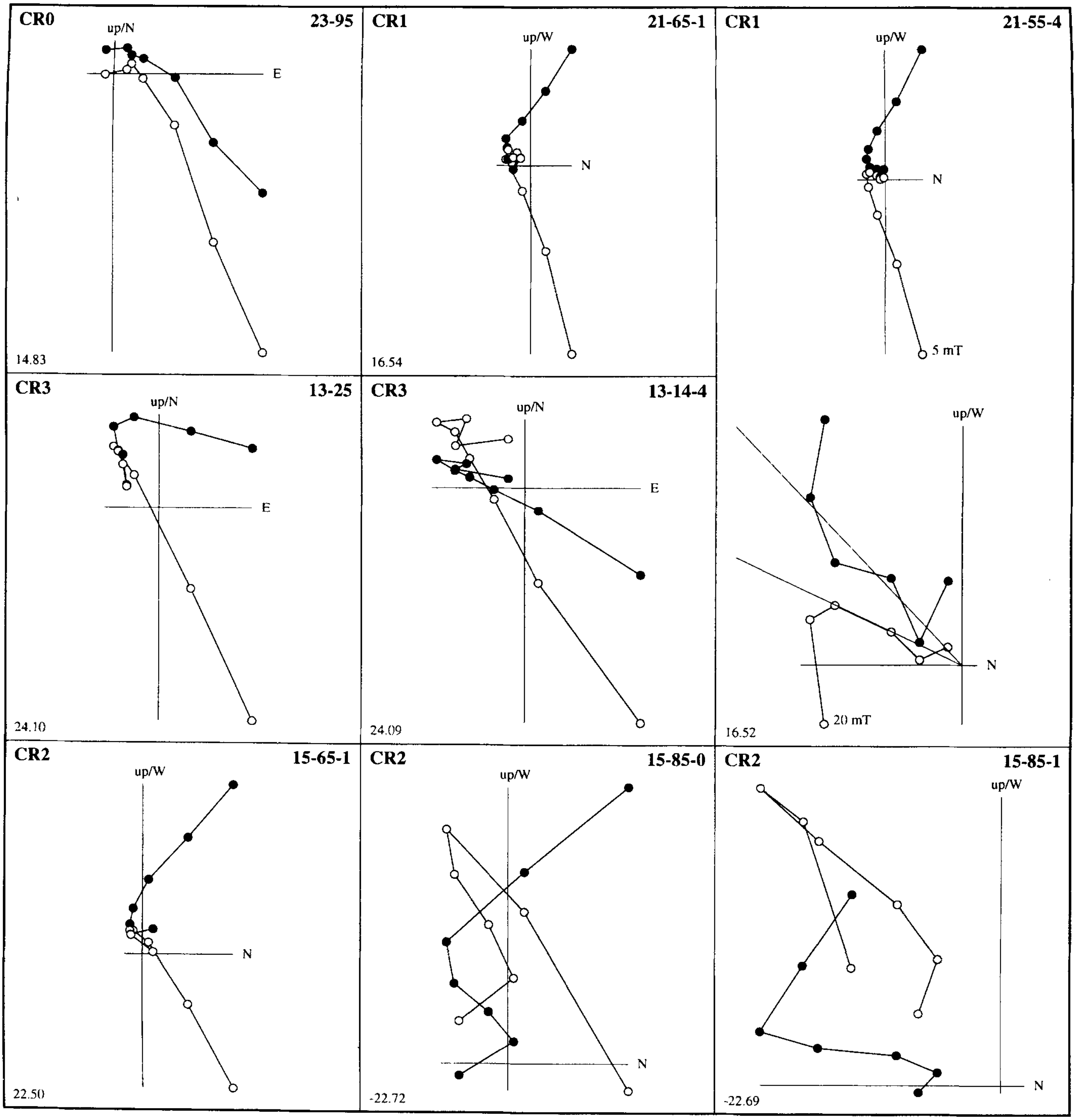


Figure 3. Examples of alternating field NRM demagnetization diagrams, using steps of 5, 10, 20, 30, 40, 50, 60 mT, of additional samples taken in the intervals containing CR0 to CR3. See also caption to Fig. 1.

record of core RC9-181 was later established by Vergnaud-Grazzini *et al.* (1977), and a sapropel chronology could be established according to a procedure outlined by Rossignol-Strick (1983, 1985). She attributed sapropel formation to African monsoons as an immediate response to orbital insolation. Her monsoon index was derived from a low-latitude insolation gradient while she used orbital chronology to determine the ages of the sapropels. Following her pioneering work, Hilgen (1991) correlated sapropel patterns to the astronomical solutions using slightly different inferred relationships between the sapropel and orbital cycles. He suggested that sapropel clusters correlate with maxima in the eccentricity cycle and individual sapropels with minimum peak values of the precession index. Using a 4 kyr time lag of the climate response, sapropels S1 to S12 were dated using the correlative peak of the precession index as calculated by Berger (1978).

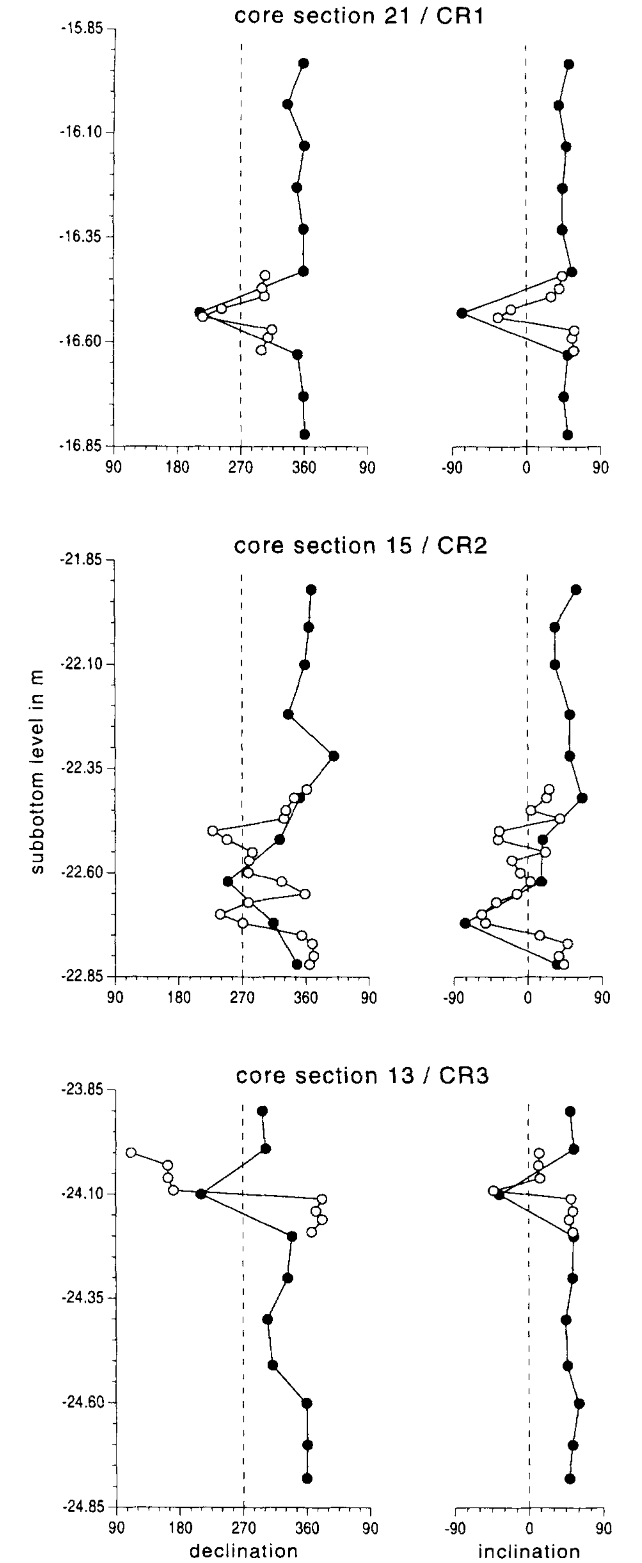


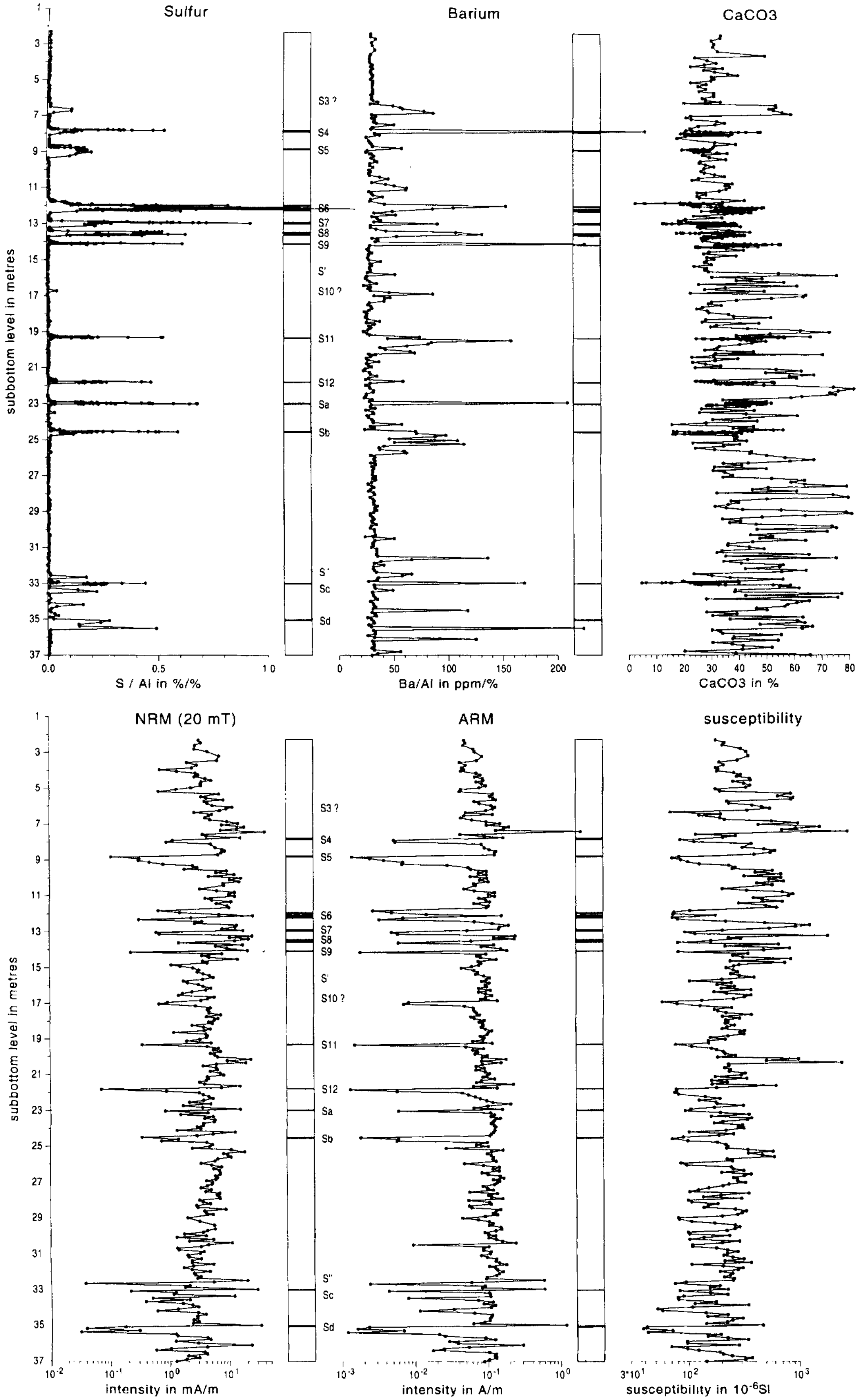
Figure 4. Declination and inclination of the ChRM of the original (closed circle) and additional (open circle) sampling of CR1–CR3. The directions are determined directly, not using great-circle analysis.

Lourens *et al.* (1996) recently followed Hilgen (1991) in his correlation, but used 65°N summer insolation rather than precession, but they arrived at essentially the same ages. Discrepancies between the insolation ages of Rossignol-Strick (1983) and the precession ages of Hilgen (1991) are minor—a few kyr—except for sapropels S10 (respectively 291 and 331 ka) and S12 (respectively 464 and 481 ka).

Here, we follow the method of Lourens *et al.* (1996) who use the 65°N summer insolation target curve calculated from the astronomical solution La90 (Laskar 1990; Laskar *et al.* 1993) with present-day values for the dynamical ellipticity of the Earth and tidal dissipation by the Sun and Moon. Lourens *et al.* (1996) argue that this solution results in the best fit with the geological record; for the construction of climatic proxy time series they introduced a 3 kyr time lag between maxima in summer insolation and sapropel mid-points.

We have used the actual sapropels as well as a number of 'ghost-sapropels' in our correlation to the insolation curve. In detail, it is observed that ARM minima are typically found just below a sapropel, and ARM and Ba maxima just above it. Consequently, we have taken as a ghost-sapropel a level in agreement with those observations. For most of the core, recognition of sapropels and sapropel patterns and their correlation to insolation is quite straightforward. Only in the interval 26–30 m subbottom is the correlation less certain, because the S, Ba and ARM signals do not display characteristic peaks which can be tied to insolation maxima, while the carbonate, NRM and susceptibility signals are ambiguous. In this interval, the correlation relies strongly on the oxygen isotope record; in particular, the position of the Brunhes/Matuyama boundary in stage 20, at ≈812 ka, imposes a strong constraint (Fig. 6).

The oxygen isotope record was tuned to the ice sheet model of Imbrie & Imbrie (1980) using the 65°N July insolation curve of Berger & Loutre (1991). This target curve introduces some non-linearity in the response to insolation forcing because of different time constants for the growth and decay of ice sheets. This may explain in part the generally small differences—typically ranging from 0 to 5 kyr and averaging 2 kyr—between the oxygen isotope curve tuned to the ice model and the sapropel pattern tuned to the 65°N summer insolation curve of La90. Small differences are probably also caused by slight inaccuracies in the position of the sapropel signal (i.e. insolation maximum) based on Ba and ARM: the sampling resolution of 10 cm and a sedimentation rate of 2–3 cm kyr^{-1} may cause the position to be in error by some 2–3 kyr. A significant discrepancy of 22 kyr between the two age models concerns the correlation of sapropel S12. Initially, we correlated S12 to the insolation maximum at 483 ka—following Lourens *et al.* (1996) and thus Hilgen (1991)—because this maximum is more prominent than the maximum at 461 ka. Spectral analysis of the oxygen isotope record, however, revealed a better fit if we used the maximum at 461 ka. This is in agreement with the earlier results of Rossignol-Strick (1983), but also agrees with the correlation of the sapropel palynology of KC-01B and the monsoon index (Rossignol-Strick, personal communication 1996). Also in the 600–800 ka interval there are some small discrepancies between the two age models: some 10 kyr at 618 and at 785 ka. Here, we retain the sapropel/insolation model. We emphasize that both methods have been independently derived: even though the sapropel/insolation model was aided by constraints imposed by the isotope record in the 26–30 m subbottom interval, calibration points are still only determined by correlation of the sapropel signal to the insolation maximum. A detailed comparison between the two age models and their



Sulfur
Barium
CaCO3
subbottom level in metres
S3 ?
S4
S5
S6
S7
S8
S9
S'
S10 ?
S11
S12
Sa
Sb
S"
Sc
Sd
S / Al in %/%
Ba/Al in ppm/%
CaCO3 in %
NRM (20 mT)
ARM
susceptibility
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susceptibility in 10^-6 SI

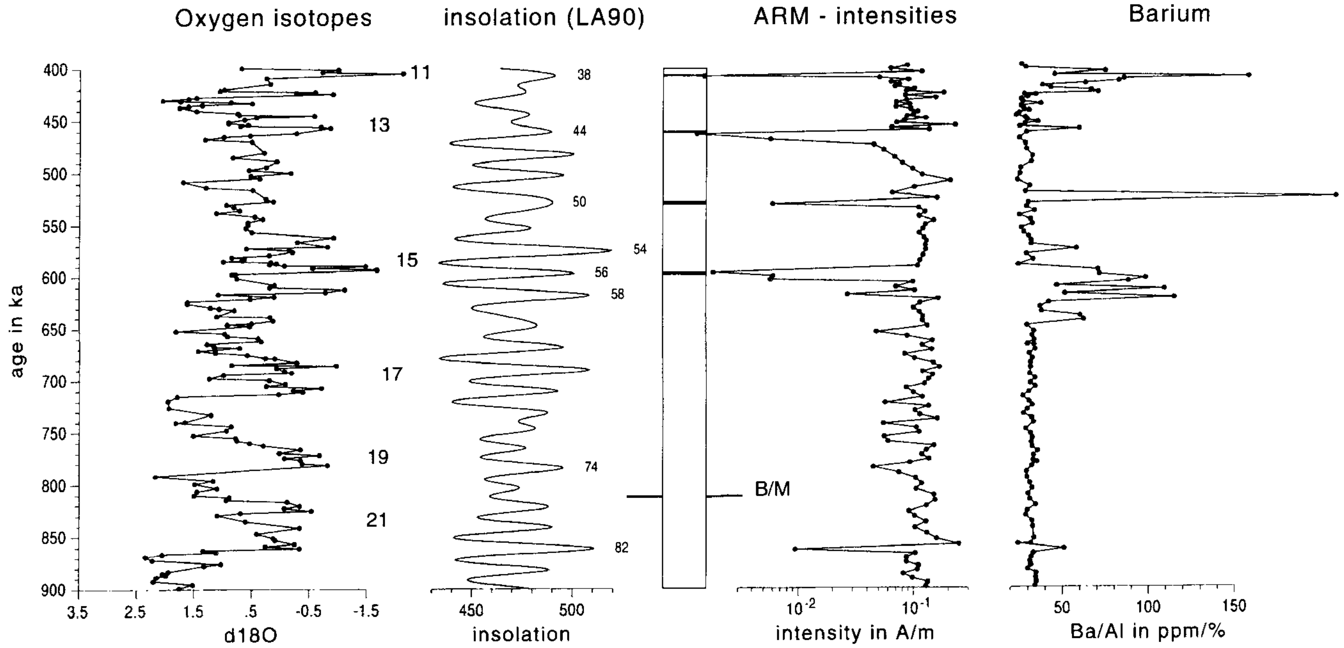

Figure 6. Oxygen isotope, insolation curve, ARM intensities and Barium record of KC-01B. The records are compared to the 65°N summer insolation of solution La90 (Laskar 1990) with present-day values for dynamical ellipticity and tidal dissipation, lagged by 3 kyr for better comparison with the records. The numbers in the isotope curve denote oxygen isotope stages, chronologically tuned to the ice volume model of Imbrie & Imbrie (1980). The (even) numbers in the insolation curve denote insolation maxima (Table 1) according to Lourens *et al.* (1996). In the column, black lines denote sapropels; the position of the Brunhes–Matuyama (B/M) with respect to the isotope record (Stage 20) is indicated.

tuning strategies will be discussed elsewhere. Reconciling the two age models in detail will not result in any significant change in the ages of reversal excursions and biostratigraphic datum levels.

The calibration points used are summarized in Table 1, while Fig. 7 shows the ARM and Ba record converted to the time domain and compared with the insolation curve. We propose a new sapropel labelling that has the advantage that if new sapropels are found, they need not be renamed or renumbered: we tie them to the corresponding maximum in the insolation curve. Maxima in insolation are even-numbered starting from the Recent, while minima are odd-numbered. We denote actually observed sapropels with an upper-case S, ghost-sapropels with a lower-case s. For example, sapropel S11 becomes Si38, whereas S' becomes si26. Upon its actual recording as a sapropel, it becomes Si26.

The resulting astronomically calibrated ages for the calcareous nannofossil events (Table 2) differ slightly from the ages obtained by Castradori (1993), who used interpolation between the published ages of sapropel S9 (238 ka; Hilgen 1991) and the Brunhes/Matuyama and upper Jaramillo reversal boundaries (780 and 990 ka; Shackleton *et al.* 1990). Also, the ages given in Dekkers *et al.* (1994) differ by several thousand years from those presented here because we have refined our correlation and, in addition, used more calibration points (Table 1). We can also derive independent astronomical ages for the recorded positions of the Brunhes/Matuyama (BM) and Jaramillo boundaries, as well as for the short geomagnetic events in the Brunhes Chron. These ages, and their correlation to existing events are discussed in the following sections.

4 ASTRONOMICAL AGES FOR THE BRUNHES/MATUYAMA BOUNDARY AND THE JARAMILLO SUBCHRON

4.1 The position of the Brunhes/Matuyama boundary, 812 ka

In core KC-01B, the BM boundary is dated at 812 ka. The 812 ka age is based on its position in stage 20 (Fig. 6). The BM boundary interval in core KC-01B is located in a zone where only minor diagenesis was thought to have occurred (Dekkers *et al.* 1994) so that delayed NRM acquisition is less important. Nevertheless, delayed NRM acquisition has probably caused the reversal to appear older, because the obtained age of 812 ka for the BM boundary is older than the generally accepted age of 780 ka. A recent age estimate of the BM boundary is derived from astronomical fine-tuning of the $\delta^{18}O$ record of core MD900963 by Bassinot *et al.* (1994) and is 775 ± 10 ka. The presently most accurate age estimate of the BM boundary is 777.9 ± 1.8 ka, as derived by Tauxe *et al.*

Figure 5. Geochemical (S, Ba and $CaCO_3$) and rock-magnetic records (NRM_{20mT}, ARM and susceptibility) of core KC-01B. In particular, S, Ba and ARM show a strong correlation with sapropels (black layers in the columns). Sapropels have been recognized as the earlier known sapropels S4 to S12 (Ryan 1972); four older sapropels have been labelled Sa, Sb, Sc, Sd. A sapropelitic lithology (sapropel traces, dark brown olive colour) has been found at levels S″ and S′. So-called 'ghost-sapropels' are recognized on the basis of their geochemical and/or rock-magnetic signature (Table 1) and were used for the correlation to the insolation curve (Figs 6 and 7).

Table 1. Sapropels (S1–12, and Sa–d) and ghost-sapropels (S′, S″) and ages (age1) of S1–12 according to Hilgen (1991). All sapropels have been renamed after their correlative maximum of the insolation curve. Si plus an even number denotes insolation maxima from Recent downwards (minima are odd-numbered); upper-case (Si) denotes a sapropel, lower-case (si) denotes a ghost-sapropel or sapropelic signature based on rock-magnetic and geochemical properties. Asterisks denote the levels that have been used for our age model. Prominent features in the various records are indicated. The youngest sapropel in a land section is sapropel *v* in the Vrica section (Zijderveld *et al.* 1991; Lourens et al. 1996).

level (m)	sapropel (S)	sapropel (i-cycle)	age (ka)	lithology	rock magnetic and geochemical properties
	S1	**Si2**			
6.70	**S3**	* **Si8**	81		Ba, S, chi
7.82	**S4**	* **Si10**	102	sapropel	Ba, S, ARM, NRM, chi
8.80	**S5**	* **Si12**	124	sapropel	Ba, S, ARM, NRM, chi
12.08	**S6**	* **Si16**	172	sapropel (double)	Ba, S, ARM, NRM, chi
12.91	**S7**	* **Si18**	195	sapropel	Ba, S, ARM, NRM, chi
13.50	**S8**	* **Si20**	217	sapropel	Ba, S, ARM, NRM, chi
14.08	**S9**	* **Si22**	240	sapropel	Ba, S, ARM, NRM, chi
15.78	**S'**	* si26	288	thin black layer	Ba, NRM, chi
16.89	**S10**	* **Si30**	331	brownish olive	Ba, S, ARM, NRM, chi
19.30	**S11**	* **Si38**	407	sapropel	Ba, S, ARM, NRM, chi
21.78	**S12**	* **Si44**	461	sapropel	Ba, S, ARM, NRM, chi
22.97	**Sa**	* **Si50**	529	sapropel	Ba, S, ARM, NRM, chi
24.10		* si54	575		Ba, NRM, chi
24.52	**Sb**	* **Si56**	597	sapropel	Ba, S, ARM, NRM, chi
25.10		* si58	618		Ba, ARM, chi
29.00		* si74	785		ARM, NRM, chi
30.45		* si82	862		Ba, ARM
31.60		* si86	908		Ba, chi
32.57	**S''**	* si90	955	sapropel traces	Ba, S, ARM, NRM, chi
33.01	**Sc**	* **Si92**	976	sapropel	Ba, S, ARM, NRM, chi
33.42		* si94	997		Ba, S, ARM, NRM, chi
34.10		* si96	1027		S, ARM, NRM, chi
34.55		* si98	1048		Ba
35.05	**Sd**	* **Si100**	1070	sapropel	S, ARM, NRM, chi
35.50		* si102	1091		Ba, S, ARM, NRM, chi
36.25		* si104	1111		Ba, ARM, NRM
36.85		* si108	1144		Ba
GAP BETWEEN PISTON CORES - LAND SECTIONS					
	v	**Si122**	1280	sapropel	see Lourens et al. (1996)

(1996) based on a compilation of 18 estimates of the age of the boundary which were tied to an astronomical timescale derived from the ice-sheet model of Imbrie & Imbrie (1980). The use of the ice-volume model of Berger *et al.* (1995) would shift the astronomical age estimate by some 2 kyr to older ages. The youngest—and thus probably the most reliable—age estimates ($\approx$775 ka) place the boundary in the middle of stage 19 (or stage 19.2 according to the revised numbering of Bassinot *et al.* (1994)), or in insolation cycle si73. A compilation of 10 $^{40}Ar/^{39}Ar$ ages by Tauxe *et al.* (1996) results in an average of 778.2$\pm$3.5 ka, which compares favourably with their astronomical age of 777.9$\pm$1.8 ka; the combination of the two independent data sets yields an overall mean of 778.0$\pm$1.7 ka ($N=28$).

4.2 The Jaramillo subchron, 1000–1070 ka

In core KC-01B, our best age estimate of the upper Jaramillo is 1001$\pm$3 ka (Table 2), interpolated between si94 and si96 (Table 1). We have taken the highest level where the change from normal to reversed polarity occurs (33.52 m subbottom), and assume that the level of reversed polarity just below is caused by delayed NRM acquisition. This is supported by the results of Dekkers *et al.* (1994).

Determining the position of the lower Jaramillo reversal in core KC-01B, however, is not straightforward. It occurs in an interval with strong sapropelitic signals (Fig. 5), and most of the interval belongs to the diagenetic category. Our best estimate is therefore the last level above the interval where anomalous inclinations (and some declinations) occur, at $\approx$35.10 m subbottom, having an age of 1072$\pm$5 ka, interpolated between Si100 and si102. This is at a higher level than the level we provided earlier to Castradori (1993; see also Dekkers *et al.* 1994) and which was based on a preliminary assessment of ChRM directions. Below sapropel Si100 is a 40 cm thick zone where dissolution has occurred. Deeper, two distinct Ba-peaks occur and hence the NRM signal in this part of the core is very probably not primary. Re-examination of

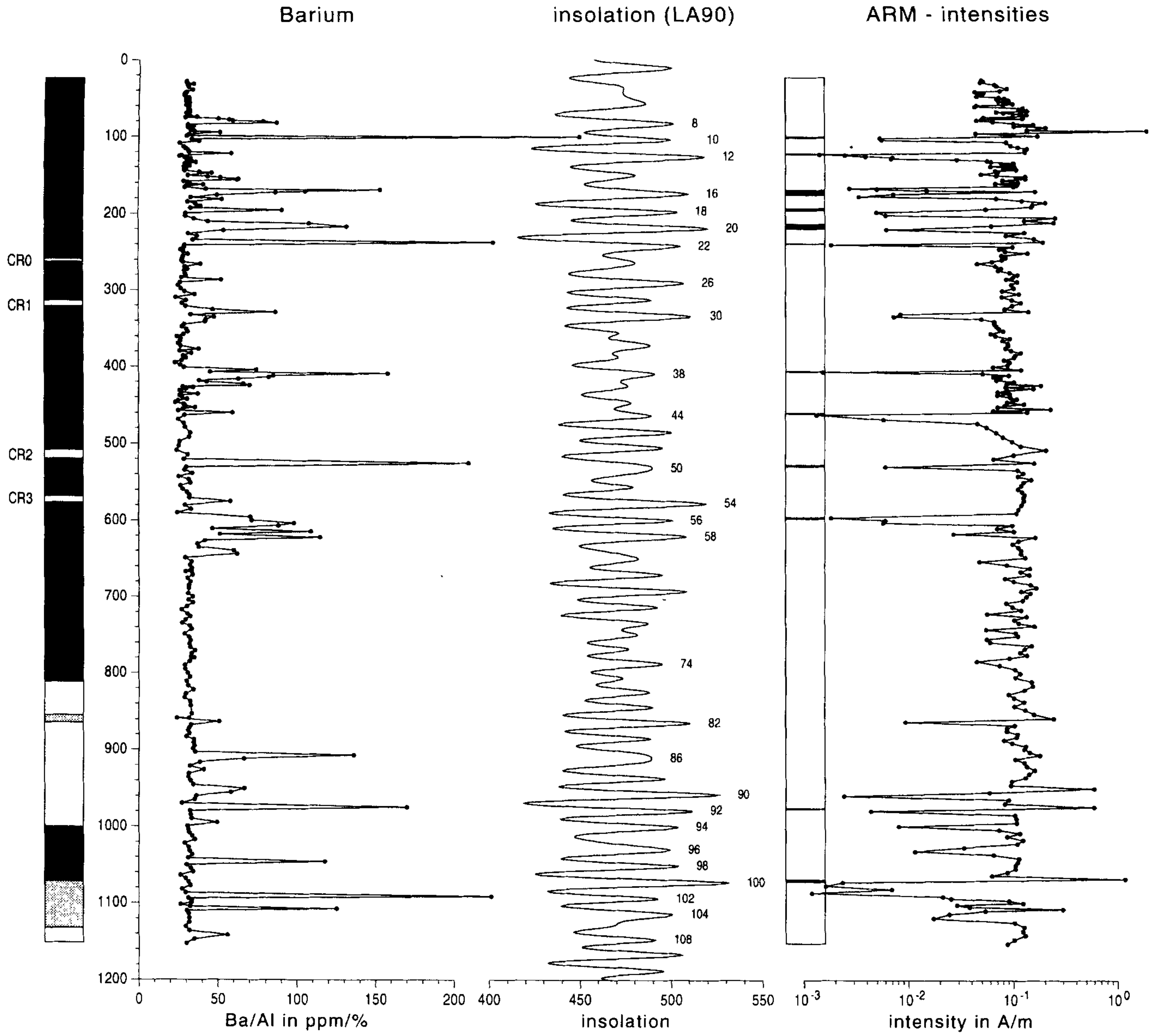

Figure 7. Conversion of the subbottom levels of KC-01B to ages in ka, using the tie-point (sapropels or ghost-sapropels) from Table 1 by correlating them to distinct minima in the calculated insolation curve. The correlation points (insolation maxima) that have been used for our age model are indicated by a number, reflecting the maxima from Recent downwards. These numbers have been used for a new (Si) sapropel coding (see Table 1, and text).

the demagnetization diagrams showed that there are indeed some small and largely unresolved reversed components remaining after treatment with the highest fields (up to 60 mT).

Shackleton *et al.* (1990) derived ages for the Jaramillo boundaries of 990 and 1070 ka. From their paper, we estimate the errors in the age of the two boundaries at ≈ 10 kyr. The upper Jaramillo occurs in ODP Site 607 (and in Site 552) in oxygen isotope stage 27, and the lower Jaramillo in oxygen isotope stage 31. The age for the upper Jaramillo in KC-01B (1001 ± 3 ka) is slightly older than that of Shackleton *et al.* (1990). The most likely age for the lower Jaramillo reversal boundary in KC-01B is within the error of Shackleton *et al.* (1990). Therefore, we retain the Shackleton *et al.* (1990) ages as the currently best ages for the Jaramillo, keeping in mind the difficulties encountered in KC-01B.

5 ASTRONOMICAL AGES OF GEOMAGNETIC EVENTS IN THE BRUNHES CHRON

A relevant discussion on the reality of geomagnetic events was recently given by Merrill & McFadden (1994). They argue that short geomagnetic events in the Brunhes are not the same as full reversals between (sub)chrons and they prefer the term 'reversal excursions' rather than 'reversal events'. They concluded that there is no convincing evidence for even a single

Table 2. Ages of calcareous nannofossil datums and geomagnetic events. Level is subbottom level in metres, age1 is from Castradori (1993) for the nannofossils and from Shackleton *et al.* (1990) for the reversal boundaries. Age2 is the age according to the sapropel chronology of Table 1.

		level (m)	age 1 (ka)	age 2 (ka)
base 'acme' small	*Gephyrocapsa*	14.95	265	265
P. lacunosa	LAD	20.80	468	440
Gephyrocapsa	sp. 3 LAD	24.14	584	577
Gephyrocapsa	sp. 3 FAD	32.71	944	962
CR0		14.83		261
CR1		16.55		318
CR2		22.72		515
CR3		24.05		573
Brunhes / Matuyama		29.85	780	812
upper Jaramillo		33.52	990	1001
lower Jaramillo		35.10	1070	1072

reversal event during the Brunhes and that there are only two well-documented reversal exursions: the Laschamp ($\approx$40 ka) and the Blake (110–120 ka). They suggest that reversal excursions may occur when the dipole field intensity is low, reflecting a larger non-dipole to dipole field ratio, an implication which is supported by the observations of Valet & Meynadier (1993). These authors suggested that in their composite palaeointensity record from ODP Sites 848 and 851—which is well-dated by astronomical tuning—many of the reported excursions correlate with low field intensity intervals. The redating of reversal excursions from recent compilations (Champion, Dalrymple & Kuntz 1988; Nowaczyk *et al.* 1994), however, requires a critical (re)examination of the existence and age of the reported reversal excursions.

The global recognition of short geomagnetic events/excursions in the Brunhes (or Matuyama) Chron is severely hampered by the lack of an accurate and reliable age control. This may lead, and often has led, to the well-known 'reinforcement syndrome' (Watkins 1968). Fortunately, the quality of age control of sediment cores is rapidly increasing, in particular because of the increasing quality and resolution of oxygen isotope records, which are in turn increasingly better constrained in age (Imbrie *et al.* 1984; Shackleton *et al.* 1990; Bassinot *et al.* 1994). In core KC-01B, four reversal excursions are found, with ages of 261$\pm$3 ka (CR0), 318$\pm$3 ka (CR1), 515$\pm$3 ka (CR2) and 573$\pm$3 ka (CR3), based on the current astronomical calibration. The present section focuses on the correlation with reversal excursions reported elsewhere.

5.1 Chronology of reversal excursions

In their compilation of reversal excursions in the Brunhes Chron, Champion *et al.* (1988) recognized eight excursions, four of which were—at that time—also found in lavas and radiometrically dated. In a more recent compilation, Nowaczyk *et al.* (1994) presented a table of 13 possible reversal excursions during the last 500 kyr, including the Mono Lake (Denham & Cox 1971) and Lake Mungo (Barbetti & McElhinny 1972) excursions. In addition, they included some new excursions from the Norwegian–Greenland Sea and Fram Strait that seem well-documented.

When correlating reversal excursions, it must be realized that many of the older ages in the literature have been based on linear interpolation between Recent and the BM boundary. The accepted age of the BM boundary has changed considerably over the years (from 690 to 780 ka), now having a best estimate of 778 ka (Tauxe *et al.* 1996). A correction to the presently preferred age of the BM boundary should at least be applied. Since sedimentation rates within the Brunhes may vary widely, however, we believe that a better age assignment is needed than simple linear interpolation. Although many geomagnetic events seem to have been recorded in numerous cores from all oceans, only in relatively few of the older cores is a good age control available in the form of a δ^{18}O record (e.g. Wollin *et al.* 1971) or of a sapropel record (e.g. Ryan 1972; Vergnaud-Grazzini *et al.* 1977; Rossignol-Strick 1983). The older δ^{18}O records, however, were not calibrated and tuned to the astronomical solutions as, for instance, was done by Shackleton *et al.* (1990) for the ODP 677 record. Hence, when correlating reversal excursions of which the ages are based on older δ^{18}O records, these records must be (re)calibrated and their reliability assessed. Often, an oxygen isotope record cannot be established because of the absence or (later) dissolution of planktonic and benthonic foraminifera. Therefore, in many recent studies (nannofossil) biostratigraphic dating is often used, as calibrated to the oxygen isotope record in other, similar areas and hence to absolute time. Although these biostratigraphies are not always synchronous between the high/middle/low latitudes (Raffi *et al.* 1993), they provide, in general, a reliable chronostratigraphic framework (see e.g. Bleil & Gard 1989; Nowaczyk & Baumann 1992). Additional dating may occasionally come from ^{14}C, ^{10}Be, ^{230}Th or fission-track dating. In several cases, a reversal excursion is recorded in lava flows which can then be dated by K/Ar or ^{40}Ar/^{39}Ar methods, for instance the Big Lost (Champion *et al.* 1988) or the West Eifel excursion (Schnepp & Hradetzky 1994).

5.2 Revised chronology of cores V12–122 and RC9–181

We have taken two of the older cores—V12–122 and RC9–181—for two reasons. (1) An oxygen isotope record is available for both V12–122 (Wollin *et al.* 1971) and RC9–181 (Vergnaud-Grazzini *et al.* 1977) and the latter core has in addition a sapropel chronology (Ryan 1972). This allows us to re-examine the original chronology for these two cores. (2) These cores are the 'type sections' of several of the earliest reported reversal excursions: the Jamaica, Levantine and Emperor (Ryan 1972).

We have recalibrated core V12–122 by correlating and matching the δ^{18}O record to the astronomically tuned isotope record of ODP Site 677 (Shackleton *et al.* 1990). For this, we have taken the most prominent peaks in both records, typically the minima and maxima of isotope stages 5, 7, 9, 11 and 13. The lowermost part is less well resolved, and an error of some 10 kyr is easily introduced (Fig. 8). For the calibration of core RC9–181, we have used the sapropel chronology reported in the present paper, assigning ages to the midpoints of S1–S12 according to Table 1 (Fig. 8). If we compare the transferred ages of both cores, it appears that Ryan (1972) was correct in correlating the Blake and Jamaica reversal excursions in cores V12–122 and RC9–181.The Levantine reversal excursion, however, does not correlate at all with the reversal excursion (now) at *c.* 310–320 ka in core V12–122, as was originally proposed. The present correlation (Fig. 8) is in agreement with the oxygen isotope record of Vergnaud-Grazzini *et al.* (1977). For instance, the isotope stage 9 peak is found at S10 (Si30), and the most prominent peak of isotope stage 11 at S11 (Si38). Hence, a revision of the ages of these earlier reported (and other) reversal excursions seems warranted.

5.3 Ages of reversal excursions in the Brunhes Chron

We refrain from discussing in detail the reversal excursions younger than the Blake (110–120 ka). Several of them seem well documented, like the Mono Lake and Laschamps reversal excursions. The Mono Lake excursion is only found in western North America and in the Arctic seas, and has an age of $\approx$25 ka (see Nowaczyk *et al.* 1994; and references therein), while the Laschamps excursion seems well established and (radiometrically) dated in France and Iceland (Levi *et al.* 1990) and Hawaii (Holt, Kirschvink & Garnier 1996), and it is also found in the Arctic seas (Bleil & Gard 1989; Nowaczyk & Baumann 1992; Nowaczyk *et al.* 1994). The age of the Laschamps excursion ranges from 40 to 45 ka. Finally, a reversal excursion at 70–80 ka seems to have been recorded in the Norwegian–Greenland Sea (Bleil & Gard 1989) which was also found elsewhere in the Arctic seas (see Nowaczyk *et al.* 1994).

The (absence of the) Blake reversal excursion in core KC01B will be discussed in Section 6.2. Between the Blake reversal excursion and the BM boundary, many short geomagnetic events seem to have occurred. Here, we discuss some of the better dated evidence for such short reversal excursions, which often seem to be related to intervals of low relative palaeointensity (Valet & Meynadier 1993; hereafter referred to as VM93). Starting with the youngest reversal excursion, we assign age ranges according to our current best estimate. We also label the reversal excursions found in core KC-01B.

5.3.1 Blake, 110–120 ka

The Blake reversal excursion was first reported by Smith & Foster (1969, in the Carribean Sea, NW Atlantic and Indian Ocean), who assigned an age of 108–114 ka to it. The Blake has subsequently been found at many places (see the compilations of Champion *et al.* 1988; Nowaczyk *et al.* 1994) and appears to occur globally. In the Mediterranean, the Blake event is amongst others recorded in core RC9–181 (Ryan 1972), where negative and shallow inclinations were found in the entire interval between sapropels S4 and S5 (Si10 and Si12, 102–123 ka; Table 1). Tucholka *et al.* (1987) found the Blake event in another five piston cores from the Mediterranean, three of which contained sapropels; again, the Blake was found between sapropels S4 and S5. They used oxygen isotopes (stages 5e–5d) to assign an age of 117 ka and a duration of $\approx$7 kyr. Records of the Blake have also been reported in Arctic cores from the Norwegian–Greenland Sea (Bleil & Gard 1989) and from the Fram Strait (Nowaczyk & Baumann 1992). The most recent age estimate of the Blake comes from a record in Chinese Loess (Zhu *et al.* 1994), based on correlation to the oxygen isotope record and on thermoluminescence dates, from 117.1 ± 1.2 to 111.8 ± 1.0 ka. There is little direct or accurate evidence from radiometric dating of the Blake event. The closest K/Ar and Ar/Ar ages are perhaps 128 ± 33.0 ka for the Laguna basalt flow of Champion *et al.* (1988), and the occurrence of reversed flows on Hawaii younger than 132 ± 32 ka (Holt *et al.* 1996). The minimum intensity interval in the VM93 relative palaeointensity curve would be in agreement with an interval of 110–120 ka.

5.3.2 The Biwa events

Some evidence for the Blake and other events was originally thought to come from Lake Biwa in Japan. In the 200 m

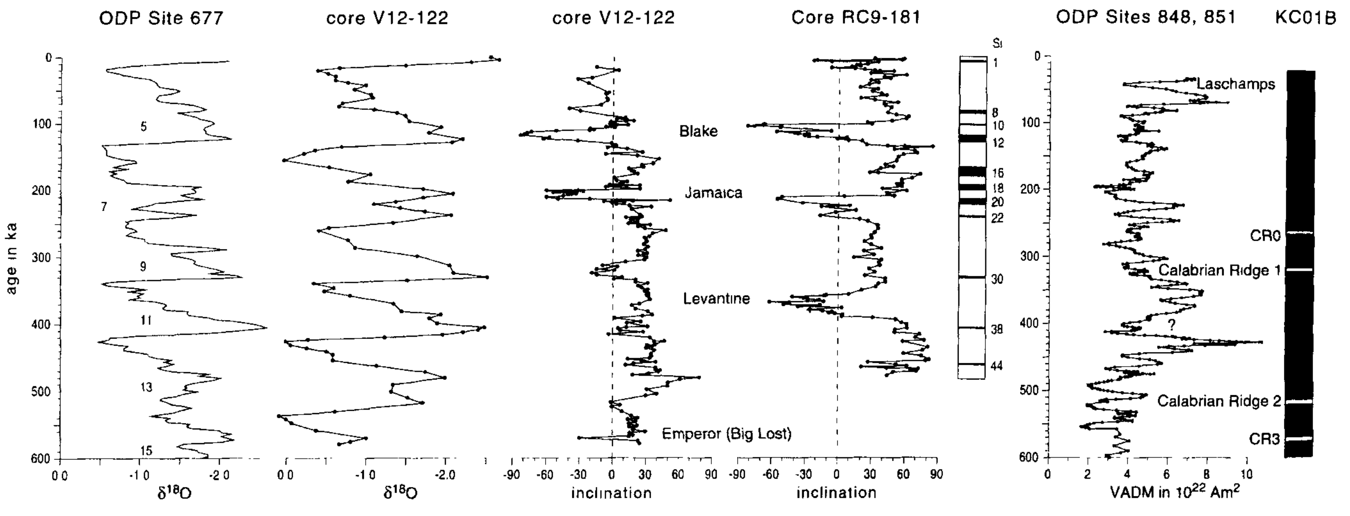

Figure 8. Age recalibration of cores V12–122 and RC9–181 by using the oxygen isotope record of V12–122 and correlation to the astronomically tuned isotope record of ODP Site 677 (Shackleton *et al.* 1990), and by using the sapropel record (black levels in the column) of RC9–181 and the tie-points of Table 1. The rightmost figure is the relative palaeointensity record of Valet & Meynadier (1993), showing that almost all reversal excursions (labelled by their names from the literature) and the Calabrian Ridge reversal excursions (CR0–3) occur during intensity minima.

core—which was 'broken up into pieces more than 200 in number'—Kawai *et al.* (1972) found five events (A, B, C, D and E), the latter ones being called Biwa I (event C), Biwa II (D) and Biwa III (E). Kawai *et al.* (1972) assigned an age of ≈ 110 ka to Biwa event B, on the basis of the extrapolation of ^{14}C ages. They concluded that event B represented the Blake reversal excursion. They also assigned ages to Biwa I (≈ 180 ka) and Biwa II (≈ 295 ka).

In a subsequent and more detailed study of a 1422 m core from Lake Biwa, close to the first one, no Biwa events were found. Their absence was attributed to ineffective AF demagnetization and distortion of the soft sediment (Torii *et al.* 1983). Later, Kawai (1984) interpolated the Biwa events in the original 200 m core between tephra layers fission-track dated by Nishimura & Yokoyama (1973), arriving at the following ages: event B, 100 ka; Biwa I, 160 ka; Biwa II, 310 ka; and Biwa III, 380 ka. Hayashida (1984) correlated volcanic ashes—in association with reversed directions in Pleistocene deposits—to ash layers (BB195 and BB207, near the Biwa I event) from the Lake Biwa core, and inferred an age of 170 ka for the reversed interval.

In their recent re-examination of the Lake Biwa tephras, however, Machida, Arai & Yokoyama (1991) showed that volcanic ash layers around Biwa I are correlative to widespread tephra layers occurring in the last interglacial stage. They derived new ages for events B (≈ 60 ka), Biwa I (≈ 110 ka), Biwa II (≈ 200 ka) and Biwa III (≈ 250 ka). Thus, the Biwa I event would correlate to the Blake reversal excursion. Meyers, Takemura & Horie (1993) obtained the same conclusion by examining stratigraphic records of palaeoclimatic indices and their correlation with marine glacial-interglacial cycles. Recent work by Hayashida (1995) has shown that nearby lake cores could be correlated to the top 100 m of the Lake Biwa core (containing event B and Biwa I), but he found no evidence for any significant excursions or reversals. These results, together with the data of Torii *et al.* (1983) on the 1422 m Lake Biwa core, provide negative evidence for the occurrence of reversal excursions in central Japan. We hesitate to include the Biwa events as reliably determined and well-dated reversal excursions.

5.3.3 Albuquerque at 155–165 ka?

Several reversal excursions have been reported around 160–170 ka, coinciding with a minimum in VM93. There is some evidence for an excursion found in Alaska sediments below the Old Crow Tephra which has a reported FT age of 149 ± 13 ka (Westgate 1988). A K/Ar radiometric age with a weighted mean of 155 ± 47 ka was found by Geissman *et al.* (1989) for a reversal excursion in volcanoes from Albuquerque, New Mexico, but its error and individual measurements (ranging from 30 ± 600 ka to 170 ± 50 ka) do not distinguish it from the Jamaica or the Blake. Recently, however, this excursion has been redated using ^{238}U-^{230}Th at 156 ± 29 ka ($2\ \sigma$) by Peate *et al.* (1996). Further evidence has been found by Nowaczyk & Baumann (1992) in the Fram Strait (152–160 ka).

At Long Valley in California, Liddicoat (1990) found a reversal excursion at ≈ 280 ka, but this was corrected to between 160 and 180 ka (Note added in proof). It now appears, however, that this is the same excursion as the Pringle Falls excursion (Herrero-Bervera *et al.* 1994), which correlates to the Jamaica (see below). Similarly, the age of the Summer Lake excursion found by Negrini *et al.* (1994) between 180 and 190 ka was later corrected (Negrini *et al.* 1996) to the same Pringle Falls excursion. There is a distinct relative palaeointensity minimum around 160 ka in VM93, but we feel that confirmation of a reversal excursion during this time interval must await further evidence and more accurate (isotope) dating.

5.3.4 Jamaica, 205–215 ka

The Jamaica reversal excursion, named by Ryan (1972) after its discovery in core V12–122 from the Jamaica Ridge, is also clearly seen in RC9–181 (see Fig. 8). Recalibration of the original $\delta^{18}O$ record in core V12–122 yields an age slightly older than 200 ka, while in RC9–181 the most negative inclinations are seen just above sapropel Si20 (217 ka), and thus in a typical precipitation zone and implying that the actual reversal excursion may be somewhat younger. The redated Biwa II (≈ 200 ka)—if it exists—has a similar age. Our best age estimate for this reversal excursion is therefore 205–215 ka, slightly older than the Champion *et al.* (1988) estimate. In this interval, Valet & Meynadier (1993) report low values for the relative palaeointensity (see Fig. 8). Recent $^{40}Ar/^{39}Ar$ dating of an ash layer directly below the Pringle Falls reversal excursion (Herrero-Bervera *et al.* 1989) gives an age younger than 218 ± 10 ka (Herrero-Bervera *et al.* 1994). As noted above, this excursion is found at several places in western North America (Summer Lake and Long Valley). Pringle Falls thus correlates to the Jamaica. Other firm evidence for this reversal excursion is found on Hawaii, where event C of Holt *et al.* (1996) has an age between 232 ± 4 ka and 200 ± 9 ka. The Mamaku ignimbrite in New Zealand shows a reversal excursion having a reported fission-track age of 230 ± 12 ka (Shane, Black & Westgate 1994). Recent Ar/Ar dating of this ignimbrite, however, shows a slightly younger age of 220 ± 10 ka, thus probably correlating it to the Jamaica/Pringle Falls excursion (Tanaka *et al.* 1996). The Jamaica reversal excursion thus seems globally and firmly established.

5.3.5 Fram Strait/CR0, 255–265 ka?

There is very little evidence in the literature for the CR0 reversal excursion at 261 ± 3 ka. Only Nowaczyk & Baumann (1992) present evidence for a reversal excursion in cores PS 1535–10 KOL and PS 1235–2 KOL from the Fram Strait, and they assign an age of 252–262 ka to this interval. We also note that the redated Biwa III excursion has a similar age (≈ 250 ka). Additional confirmation of this excursion is certainly required, also because there is no clear evidence in the VM93 curve for a palaeointensity minimum at this age: only a small local minimum occurs at 260 ka.

5.3.6 Calabrian Ridge 1, 318 ± 3 ka

The reversal excursion Calabrian Ridge 1 (CR1) has an age of 318 ± 3 ka. Also in core V12–122 there is a reversal excursion, with negative inclinations around 320 ka (Fig. 8). Some additional evidence may come from the Norwegian–Greenland Sea, where Bleil & Gard (1989) found a reversal excursion with an age of between 330 and 340 ka. The most conspicuous interval with minimum palaeointensity values in VM93 is between 310 and 320 ka. We consider CR1 to be a reliably determined reversal excursion with an accurate age, and propose to name it the Calabrian Ridge 1 excursion.

5.3.7 Levantine, 360–370 ka?

The Levantine event was named by Ryan (1972) as the reversal excursion occurring in RC9–181 (Fig. 8). He correlated it to the interval of shallow inclinations in core V12–122, which according to our recalibration is clearly older (400–420 ka; Fig. 8). For the time being, our best estimate for the Levantine reversal excursion is an age interval of 360–370 ka. There is little confirmation for this excursion elsewhere in the world, except perhaps an excursion ranging between 330 and 340 ka in the Norwegian–Greenland Sea (Bleil & Gard 1989). For this age range, the VM93 record gives values that are higher than average, but at 360–370 ka there is a short interval with (only relative) minimum intensities.

5.3.8 A reversal excursion at 400–420 ka?

In core V12–122, the excursion with shallow inclinations, formerly correlated to the Levantine by Ryan (1972), is now found in the interval 400–420 ka (Fig. 8). The corresponding interval in VM93 is a distinct, very low relative palaeointensity interval. Perhaps, in future studies with a reliable age control, a reversal excursion will be found in this interval, but at present the evidence is poor.

5.3.9 Calabrian Ridge 2, 515 ± 3 ka

The reversal excursion CR2 is the first reliable and accurate evidence for this short and nearly reversed interval, having an age of 515 ± 3 ka. Some other evidence is perhaps found in core V12–122, where shallow inclinations occur between 515 and 525 ka. Wilson & Hey (1981) recognized a tiny wiggle within the central anomaly across the Galapagos Ridge, with an interpolated age of 493–504 ka (Cande & Kent 1992) which could correlate with CR2 (or alternatively with the distinct intensity minimum at ≈490 ka in VM93). Recently, Schnepp & Hradetzky (1994) calculated a mean K/Ar age of 510 ± 30 ka for an excursion in the West Eifel volcanics. Also, there is a distinct mimimum-intensity interval in VM93 ranging between 515 and 525 ka. Awaiting further confirmation, we propose the name Calabrian Ridge 2 reversal excursion, after its discovery in core KC-01B.

5.3.10 Emperor (Big Lost/CR3), 560–570 ka

With respect to the Emperor reversal excursion of Ryan (1972), we seem to have come full circle. Ryan (1972) provisionally named the short (indeed, one negative inclination value only) reversal excursion in core V12–122 the Emperor event. Champion *et al.* (1981) reported anomalous directions from Idaho lava flows with a K/Ar age of 460 ka, and argued that they represented the Emperor. Later, Champion *et al.* (1988) revised this age to 565 ± 14 ka, and named it the Big Lost event. They argued that the Emperor reversal excursion of Ryan (1972) probably had an age of 460 ka, and could be no older than 480 ka. According to the correlation of the V12–122 $\delta^{18}O$ record to that of Shackleton *et al.* (1990), this age is appreciably older, and arrives at *c.* 570 ka (Fig. 8). Thus, the Big Lost corresponds to the Emperor again. In KC-01B, we find the well-defined reversal excursion CR3, with an age of 573 ± 3 ka (Table 2), which we believe represents the Emperor reversal excursion. This event does not correlate very well to a distinct low palaeointensity interval (550–560 ka) in VM93, possibly because of delayed NRM acquisition and thus appearing older. Our best estimate of the Emperor is within the age range of 560 to 570 ka.

6 DISCUSSION

6.1 True geomagnetic behaviour or artefacts?

It is often difficult to determine whether reversal excursions are real or artefacts. These difficulties arise from disturbances during the coring and sampling and/or from a complicated NRM acquisition process.

Physical disturbance during the coring or sampling procedure may play an important role, but in modern studies considerable effort is directed to avoiding those sources of error, or to assessing and documenting their influence on the observed results. In long piston cores the topmost few metres are often disturbed (see, for instance, the directional swing in declination in KC-01B) and should not be used for palaeomagnetic purposes. In older studies, however, any possible disturbance is not easy to determine. Ideally, a box core, short piston core and a long piston core should be taken at the same site to minimize the problems which may arise in the top of the core. Even in a box core the topmost few centimetres often yield disturbed NRM directions, although a stable palaeomagnetic direction according to the site latitude is already fixed at a subbottom depth of only 5 cm in Arctic box cores (Nowaczyk, personal communication 1995). In addition, a portion of these few centimetres is still above the lock-in depth zone and the (detrital or authigenic) magnetic particles have not yet recorded the ambient geomagnetic field.

Complications in the NRM acquisition process can arise from the depth of the lock-in zone which may vary, for instance, with the sedimentation rate or lithology. Estimates range from 7 cm (Glass *et al.* 1991) to 16 cm (deMenocal, Ruddiman & Kent 1990; Schneider, Kent & Mello 1992), but recently Tauxe *et al.* (1996) have argued that a lock-in depth of a few centimetres is more consistent with available data. At very low sedimentation rates this may represent a considerable amount of time, but for core KC-01B possible differences in lock-in depth are hardly significant. More important is the possibility of delayed NRM acquisition (Van Hoof & Langereis 1991) caused by early diagenetic processes (Van Hoof *et al.* 1993). These processes certainly play an important role in core KC-01B, and it can be shown that particular intervals have undergone severe dissolution and/or precipitation of Fe-bearing minerals (Dekkers *et al.* 1994), resulting in the complete obliteration of a reversal excursion like the Blake (see next section) or, alternatively, in an apparent longer duration of a reversal excursion. The apparent longer duration of a reversal excursion depends on its occurrence and 'true' duration versus the extent and timing of early diagenesis—which in turn depends on factors such as climatic variation (an)oxic depositional environment, organic matter supply, sedimentation rate, and so forth.

The possibility of a self-reversal (see Merrill & McFadden 1994) is very remote in KC-01B. Neither a PDRM mechanism, nor a growth-CRM mechanism can yield a self-reversed direction. Alteration of existing magnetic minerals can yield spurious directions (e.g. Heider & Dunlop 1987), but in KC-01B the reversal excursions are found in levels where diagenetic alteration is minimal. Furthermore, laboratory alteration experiments (Heider & Dunlop 1987) were conducted at high temperature, and natural examples where alteration of the orginal NRM is suspected (alteration of seafloor basalts) are always tied to high-temperature processes.

Many reversal excursions have been reported with an apparently long duration, even to the extent that, were these durations real, one would expect to find them quite easily even with a moderately high sampling resolution. Nevertheless, many of the reported excursions have not been found in numerous cores (see Harrison 1974), nor in lavas. The rarity of observations of reversal excursions in volcanic rocks is not surprising, however, if one takes into account the extent and number of studies in sediments (e.g. DSDP/ODP and many other cores, together with long-core and U-channel processing techniques) and those in lavas. The relatively few observations of reversal excursions in lavas would plead for a short duration and is probably related to the small chance that a lava flow extrudes during a reversal excursion. The short duration of the reversal excursions is furthermore confirmed by the observations in KC-01B.

The currently existing data on reliably dated reversal excursions do not allow us to accept or reject their global occurrence, although there seems to be a bias towards high-latitude cores recording numerous and relatively long-duration (10–20 kyr) reversal excursions. For instance, from the 159 cores taken in the Southern Ocean by Goodell & Watkins (1968), $\approx$ 70 cores showed one or more reversed polarity intervals during the Brunhes Chron. Similarly, recent studies of Arctic cores (Bleil & Gard 1989; Nowaczyk & Baumann 1992; Nowaczyk *et al.* 1994) show several reversal excursion intervals that are usually not found elsewhere on the globe. These Arctic studies show high sedimentation rates and high intensities because of their lithogenic (detrital) origin, and they conform to modern palaeomagnetic standards. Possibly, Arctic sediments may be more reliable recorders of non-dipole field behaviour than many of their equatorial or mid-latitude counterparts with low sedimentation rates and increased chance of diagenesis and subsequent obliteration of short events (Nowaczyk, personal communication 1995).

6.2 The missing Blake reversal excursion, 110–120 ka

In core KC-01B, no negative or even shallow inclinations occur in the eight samples between Si10 (S4) and Si12 (S5) where the Blake reversal excursion is expected, despite the average resolution of 2.5 kyr. We have taken an almost continuous set of additional samples in the lower two-thirds of this interval to be sure that we had not missed the Blake. None of the additional samples shows any significantly deviating directions (Fig. 9). This is not surprising because almost the entire interval between SPT5 and 6 falls into the dissolution/precipitation category of Dekkers *et al.* (1994), except for two samples at 8.10–8.30 m subbottom (108–112 ka).

The various records of the Blake reversal excursion in the Mediterranean, however, are not identical and indicate that diagenesis has occurred. Contrary to Ryan (1972), Tucholka *et al.* (1987) mainly found negative or shallow inclinations in a small zone just above Si12 (S5) in three of their cores (the MD cores; see also Tric *et al.* 1991), while in their KET cores the interval of negative inclinations extends to more than halfway between Si10 and Si12 (we used their oxygen isotope records to estimate the equivalent positions of Si10 and Si12). The Blake was probably recorded more reliably in the KET cores than in the MD cores because of the absence of sapropels in the former, pointing to less anoxic conditions and probably less diagenetic alteration. The very short interval above Si12 in the MD cores is probably a diagenetic precipitation zone, similar to the situation in KC-01B. This would imply that the true Blake reversal excursion may have happened at any younger age, but that early diagenetic processes (dissolution) have subsequently obliterated (most of) the 'true event', while some of it was recorded by delayed NRM acquisition in the precipitation zone just above Si12. Indeed, in two of the MD cores (MD84461 and MD84640) only a few negative/shallow inclinations are found in the first few centimetres above this sapropel. We recall that delayed acquisition can result in apparently older ages for reversals or reversal excursions but not in apparently younger ages.

The occurrence of the Blake event in the 100–110 ka interval, however, is 'on the young side' of most reported ages, which favour the interval 110–120 ka; not all of these ages are easy to explain by sedimentary artefacts (see, e.g. Nowaczyk *et al.* 1994; and references therein). Another mechanism that can distort and even obliterate reversal excursions is PDRM resetting through a subsequently higher geomagnetic field. If reversal excursions indeed occur when the geomagnetic dipole field intensity is low as suggested by available data (Valet & Meynadier 1993; Merrill & McFadden 1994; Nowaczyk *et al.* 1994; Roberts, Verosub & Negrini 1994), then their record can

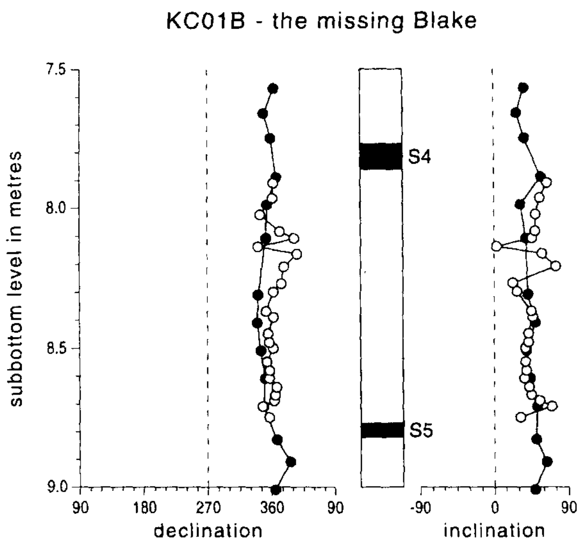

Figure 9. Declination and inclination of the ChRM from the detailed resampling of the Blake interval, showing that the Blake has not been recorded, because of early diagenetic dissolution. Solid symbols denote the original samples, open symbols are the additional samples with a higher resolution.

be (partially) reset by the subsequent geomagnetic field with a higher intensity. This mechanism was suggested by Coe & Liddicoat (1994) to explain features of the Mono Lake reversal excursion. An experimental analogue describing the behaviour of magnetite particles in curing epoxy resin in various fields reported by Løvlie (1994) shows that this mechanism can indeed occur.

7 CONCLUDING REMARKS

We summarize the reversal excursions during the Brunhes Chron with reliable age estimates and global occurrence in Fig. 10. When assessing the reliability of reversal excursion records, a thorough understanding of potentially biasing/disturbing effects should be made. These include disturbances during the coring and sampling procedure, but also disturbances of the actual recording of the NRM. Progressive demagnetization (and not blanket demagnetization) should unambiguously indicate deviating (reversed) directions, while anomalous behaviour of single samples should be confirmed by detailed resampling. Furthermore, before accepting a reversal excursion as real, diagenetic artefacts must be ruled out. Particularly in cyclic or non-homogeneous lithologies, delayed NRM acquisition can be important. An apparently long duration of a reversal excursion or its obliteration is probably related to early diagenetic precipitation or dissolution. Even if PDRM rather than a growth-CRM can be made plausible as the origin of NRM, PDRM resetting through a later geomagnetic field with a higher intensity probably plays a role in any sediment (Coe & Liddicoat 1994).

Once a reversal excursion is reliably established, the next important aspect is an accurate chronology. A truly accurate (astronomical) chronology depends on many elements, for example on synchronous recording of climate response in a variety of sedimentary environments, synchronous deposition of planktonic foraminifera (for $\delta^{18}O$ records) after bioturbation under different sedimentation rates (deMenocal *et al.* 1990), and delay of NRM acquisition depending on diagenesis. Furthermore, in globally correlating reversal excursions, an implicit assumption is that they occur at the same time. This, however, need not be true if excursions are expressions of a large non-dipole to dipole field ratio. Perhaps the same excursion (or period of low dipole field) occurs at a certain time at location A but, for example, several thousand years later (or not at all) at location B. It is interesting to note that current data on the Brunhes–Matuyama boundary suggest a difference of several thousand years of the midpoint of the reversal between two geographic regions (Tauxe *et al.* 1996). The currently available reversal excursion database, with its inaccurate chronology, however, is still too fragmentary for any definitive answer in this direction. Fortunately, the development of astronomically tuned time frames (e.g. Bassinot *et al.* 1994; Lourens *et al.* 1996) allows an increasingly accurate age determination of reversal excursions.

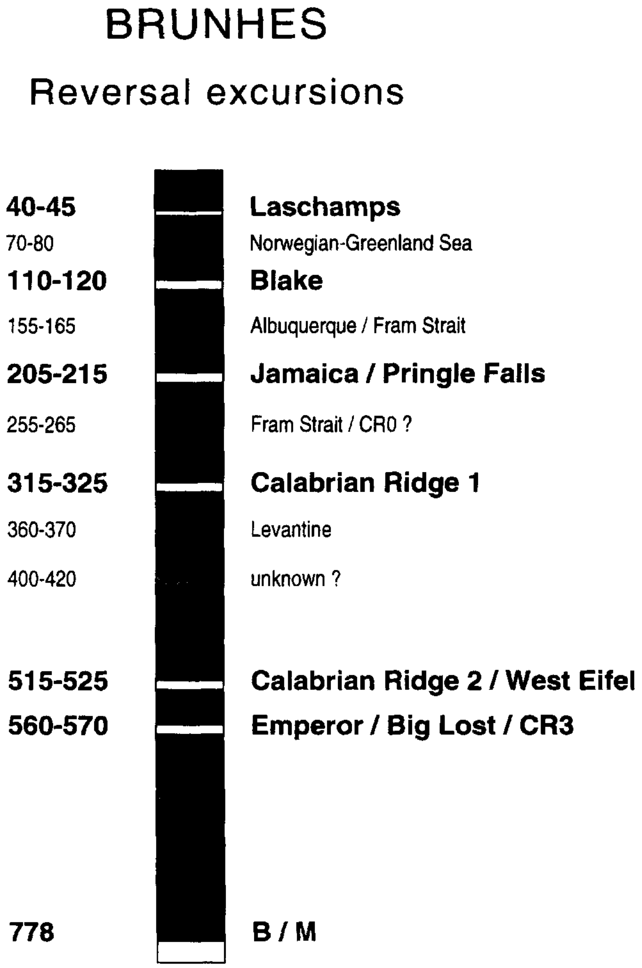

Figure 10. Reversal excursions during the Brunhes. The excursions that we believe are well-dated and occur globally are indicated in bold. We have indicated in small print the excursions that are not as certain, or which occur so far only in restricted regions. The ages are discussed in the text; the age of the Brunhes–Matuyama (B/M) is according to Tauxe *et al.* (1996).

ACKNOWLEDGMENTS

Piet-Jan Verplak and Henk Meijer assisted in the palaeomagnetic measurements. Hans Zijderveld critically read an earlier version of the manuscript. Nick Shackleton, Lucas Lourens and Jean-Pierre Valet kindly provided their data on ODP 677, insolation curve and relative palaeointensity. The review of Norbert Nowaczyk was very valuable, while the information provided by Akira Hayashida on the Biwa events proved to be useful. We are grateful for the comments of Martine Rossignol-Strick, particularly for those on monsoonal forcing. The isotopic data were obtained through the additional support of the Commissariat à l'Energie Atomique and Centre National de la Recherche Scientifique. We thank L. Labeyrie for providing facilities for the isotopic analyses, and B. LeCoat and J. Tessier for technical assistance. MJD gratefully acknowledges support of the Royal Netherlands Academy of Sciences and Arts in the form of a fellowship. This research was supported by the EU Marine Science and Technology (MAST) programme, contracts 900022C (Marflux) and MAS2-CT93-0051 (Palaeoflux).

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