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**983 and oxygen isotope records from ODP Site the Jaramillo Subchron: palaeomagnetic Brunhes boundary and the boundaries of Geomagnetic palaeointensities and
astrochronological ages for the Matuyama** −

J. E. T. Channell and H. F. Kleiven

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Geomagnetic palaeointensities and astrochronological ages for the Geomagnetic palaeointensities
and astrochronological ages for the
Matuyama–Brunhes boundary and the
boundaries of the Jaramillo Subchron: and astrochronological ages for the
Matuyama–Brunhes boundary and the
boundaries of the Jaramillo Subchron:
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oundaries of the Jaramillo Subchron:
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records from ODP Site 983 **records from ODP Site 983**
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We have measured relative geomagnetic palaeointensity proxies, palaeomagnetic directions, and δ^{18} O for the 700-1100 ka interval from ODP Site 983 (Gardar Drift, We have measured relative geomagnetic palaeointensity proxies, palaeomagnetic directions, and $\delta^{18}O$ for the 700–1100 ka interval from ODP Site 983 (Gardar Drift, North Atlantic), where mean sedimentation rates are *ca* rections, and δ^{18} O for the 700–1100 ka interval from ODP Site 983 (Gardar Drift,
North Atlantic), where mean sedimentation rates are *ca*. 13 cm kyr⁻¹. The age model
was generated by matching the benthic δ^{18} O was generated by matching the benthic δ^{18} O data to the Ice Volume Model and confirmed by tuning the precessional components of both signals. For the Matuyama– was generated by matching the benthic δ^{18} O data to the Ice Volume Model and confirmed by tuning the precessional components of both signals. For the Matuyama-
Brunhes boundary (MBB) and the boundaries of the Jaramill firmed by tuning the precessional components of both signals. For the Matuyama-
Brunhes boundary (MBB) and the boundaries of the Jaramillo Subchronozone, the
duration of the polarity reversal process, defined by virtual g duration of the polarity reversal process, defined by virtual geomagnetic polar latitudes of less than 45° , is *ca*. 5 kyr. Whereas the generally accepted astrochronological estimates for the boundaries of the Jarami duration of the polarity reversal process, defined by virtual geomagnetic polar latitudes of less than 45°, is ca. 5 kyr. Whereas the generally accepted astrochronological estimates for the boundaries of the Jaramillo Subc tudes of less than 45°, is *ca*. 5 kyr. Whereas the generally accepted astrochronological estimates for the boundaries of the Jaramillo Subchronozone lie within the polarity transitions as recorded at Site 983, the astroc estimates for the boundaries of the Jaramillo Subchronozone lie within the polarity
transitions as recorded at Site 983, the astrochronological age for the Matuyama-
Brunhes polarity transition (780 ka) is *ca*. 5 kyr olde transitions as recorded at Site 983, the astrochronological age for the Matuyama–
Brunhes polarity transition (780 ka) is $ca.5 \text{ kyr}$ older than the onset of this transition at Site 983 (775 ka). The polarity reversals li Brunhes polarity transition (780 ka) is ca. 5 kyr older than the onset of this transition at Site 983 (775 ka). The polarity reversals lie within palaeointensity lows, with abrupt recovery of palaeointensity post reversal. sition at Site 983 (775 ka). The polarity reversals lie within palaeointensity lows,
with abrupt recovery of palaeointensity post reversal. There is no progressive ('saw-
tooth') decrease in palaeointensity within the Jara with abrupt recovery of palaeointensity post reversal. There is no progressive ('sawtooth') decrease in palaeointensity within the Jaramillo Subchronozone or between
the top of the Jaramillo and the MBB, but rather, within tooth') decrease in palaeointensity within the Jaramillo Subchronozone or between
the top of the Jaramillo and the MBB, but rather, within polarity chrons, several
short intervals of low palaeointensity which sometimes coi I the top of the Jaramillo and the MBB, but rather, within polarity chrons, several short intervals of low palaeointensity which sometimes coincide with high-amplitude secular variation. Orbital (100 and 41 kyr) periods ar short intervals of low palaeointensity which sometimes coincide with high-amplitude
secular variation. Orbital (100 and 41 kyr) periods are present in the palaeointensity
record. As they are not obviously attributable to c secular variation. Orbital (100 and 41 kyr) periods are precord. As they are not obviously attributable to climathey may be a feature of the geomagnetic field itself.

they may be a feature of the geomagnetic field itself.
Keywords: geomagnetic secular variation; geomagnetic palaeointensity; Matuyama-Brunhes boundary; Jaramillo Subchron; ODP Site 983; Iceland Basin

1. Introduction

1. Introduction

Ocean Drilling Program (ODP) Site 983 (60.40 $^{\circ}$ N, 336.36 $^{\circ}$ E, 1983 m water depth),

located close to the crest of the Gardar Drift south of Iceland (figure 1), was drilled located close to the crest of the Gardar Drift south of Iceland (figure 1), was drilled
in late July 1995. The three holes drilled at the site provided a complete composite Ocean Drilling Program (ODP) Site 983 (60.40° N, 336.36° E, 1983 m water depth),
located close to the crest of the Gardar Drift south of Iceland (figure 1), was drilled
in late July 1995. The three holes drilled at the si located close to the crest of the Gardar Drift south of Iceland (figure 1), was drilled
in late July 1995. The three holes drilled at the site provided a complete composite
section to the base of the Olduvai Subchronozone

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Deposition in the Gardar Drift has been driven by thermohaline circulation associ-
ated with Norwegian Sea Overflow Water (NSOW) spilling over the Iceland–Scotland Deposition in the Gardar Drift has been driven by thermohaline circulation associated with Norwegian Sea Overflow Water (NSOW) spilling over the Iceland–Scotland
Ridge, Accumulation rates have been enhanced by horizontal a **IATHEMATICAL,
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CIENCES** ated with Norwegian Sea Overflow Water (NSOW) spilling over the Iceland–Scotland Ridge. Accumulation rates have been enhanced by horizontal advection of fine mate-Ridge. Accumulation rates have been enhanced by horizontal advection of fine mate-
rial suspended in the nepheloid layer (Wold 1994; McCave & Tucholke 1986). The
Gardar Drift accumulated throughout the Pleistocene at rate rial suspended in the nepheloid layer (Wold 1994; McCave & Tucholke 1986). The rial suspended in the nepheloid layer (Wold 1994; McCave & Tucholke 1986). The Gardar Drift accumulated throughout the Pleistocene at rates of 5–20 cm kyr⁻¹, and provides one of the most detailed Pleistocene climatic re and provides one of the most detailed Pleistocene climatic records retrieved from the North Atlantic. The composite section was generated shipboard using magnetic the North Atlantic. The composite section was generated shipboard using magnetic
susceptibility, GRAPE density and reflectance data from the measurement systems
track (MST) to correlate the three holes and splice together THE ROYAL
SOCIETY susceptibility, GRAPE density and reflectance data from the measurement systems
track (MST) to correlate the three holes and splice together an optimal (complete
and undisturbed) record of the sedimentary sequence (see Hag susceptibility, GRAPE density and reflectance data from the measurement systems track (MST) to correlate the three holes and splice together an optimal (complete track (MST) to correlate the three holes and splice together an optimal (complete and undisturbed) record of the sedimentary sequence (see Hagelberg *et al.* (1992) for review of methods). A preliminary magnetic polarity and undisturbed) record of the sedimentary sequence (see Hagelberg *et al.* (1992) for
review of methods). A preliminary magnetic polarity stratigraphy was generated on
the 'archive' halves of core sections from each hole review of methods). A preliminary magnetic polarity stratigraphy was generated on
the 'archive' halves of core sections from each hole, using the shipboard pass-through
magnetometer (Shipboard Scientific Party 1996) after the 'archive' halves of core sections from each hole, using the shipboard pass-through magnetometer (Shipboard Scientific Party 1996) after demagnetization at a single peak alternating field (usually 25 mT). The high peak alternating field (usually 25 mT). The high rate of core recovery and the need peak alternating field (usually 25 mT). The high rate of core recovery and the need
to avoid processing bottlenecks necessitated this abbreviated demagnetization treat-
ment. Subsequent shorebased treatment of $7 \text{ cm$ to avoid processing bottlenecks necessitated this abbreviated demagnetization treat-
ment. Subsequent shorebased treatment of 7 cm^3 discrete samples, collected ship-
board in plastic cubes, served to ground-truth the **PHILOSOPHICAL**
TRANSACTIONS board in plastic cubes, served to ground-truth the shipboard magnetic stratigraphy (Channell $\&$ Lehman 1999). \overline{c}

ard in plastic cubes, served to ground-truth the shipboard magnetic stratigraphy

thannell & Lehman 1999).

Here we report the magnetic properties and δ^{18} O of the 700–1100 ka interval at

the 983. The objective is to (Channell & Lehman 1999).
Here we report the magnetic properties and δ^{18} O of the 700–1100 ka interval at
Site 983. The objective is to place the palaeomagnetic directional and palaeointensity
records on a firm isotop Here we report the magnetic properties and δ^{18} O of the 700–1100 ka interval at Site 983. The objective is to place the palaeomagnetic directional and palaeointensity records on a firm isotopic age model in order to d Site 983. The objective is to place the palaeomagnetic directional and palaeointensity
records on a firm isotopic age model in order to document the behaviour of the
geomagnetic field during the 400 kyr interval which incl geomagnetic field during the 400 kyr interval which includes the Matuyama–Brunhes boundary (MBB) and the Jaramillo Subchron.

2. Oxygen isotope stratigraphy

The interval from 700 to 1200 ka was sampled continuously (nominally at 2 cm inter-The interval from 700 to 1200 ka was sampled continuously (nominally at 2 cm intervals) for stable isotopic analyses of benthic and planktic foraminifera. The benthic foraminiferal isotopic analyses were performed on the t The interval from 700 to 1200 ka was sampled continuously (nominally at 2 cm intervals) for stable isotopic analyses of benthic and planktic foraminifera. The benthic foraminiferal isotopic analyses were performed on the t vals) for stable isotopic analyses of benthic and planktic foraminifera. The benthic
foraminiferal isotopic analyses were performed on the taxon *Cibicidoides*, principally
made up of *C. wuellerstorfi*. The planktic foram foraminiferal isotopic analyses were performed on the taxon *Cibicidoides*, principally made up of *C. wuellerstorfi*. The planktic foraminiferal isotopic analyses were performed on *Neogloboquadrina pachyderma* (sinistral made up of *C. wuellerstorfi*. The planktic foraminiferal isotopic analyses were per-
formed on *Neogloboquadrina pachyderma* (sinistral). Both species were selected from
the >150 µm size fraction. The abundance of *N. pa* formed on *Neogloboquadrina pachyderma* (sinistral). Both species were selected from
the >150 µm size fraction. The abundance of *N. pachyderma* (sinistral) allowed the
planktic record to be acquired at a 2-5 cm (*ca.* 50 planktic record to be acquired at a 2-5 cm (ca. 500 year) resolution, whereas the spacspp. ing of the benthic δ^{18} O is 10 cm or greater due to the relative scarcity of *Cibicidoides* spp.
Stable isotope measurements of *N. pachyderma* (s.) specimens were made at the

spp.
Stable isotope measurements of *N. pachyderma* (s.) specimens were made at the
University of Bergen on a Finnigan MAT251 coupled to an automated carbonate
preparation device whereas the isotope measurements of *Cibic* Stable isotope measurements of *N. pachyderma* (s.) specimens were made at the University of Bergen on a Finnigan MAT251 coupled to an automated carbonate preparation device, whereas the isotope measurements of *Cibicidoid* University of Bergen on a Finnigan MAT251 coupled to an automated carbonate
preparation device, whereas the isotope measurements of *Cibicidoides* spp. were car-
ried out with a VG Isogas PRISM mass spectrometer at the Uni preparation device, whereas the isotope measurements of *Cibicidoides* spp. were carried out with a VG Isogas PRISM mass spectrometer at the University of Florida. A small number of benthic stable isotope measurements were ried out with a VG Isogas PRISM mass spectrometer at the University of Florida. A
small number of benthic stable isotope measurements were made at Scripps Institu-
tion of Oceanography on a Finnigan MAT252 and at the Unive small number of benthic stable isotope measurements were made at Scripps Institu-
tion of Oceanography on a Finnigan MAT252 and at the University of Cambridge on
a VG Isogas PRISM. Isotope data from all these laboratories a VG Isogas PRISM. Isotope data from all these laboratories were calibrated using the NIST (NBS) 19 standard, and values are reported relative to PDB.

Oxygen isotopic stages 18-35 can easily be identified in the Site 983 benthic record the NIST (NBS) 19 standard, and values are reported relative to PDB.
Oxygen isotopic stages 18–35 can easily be identified in the Site 983 benthic record
by matching to other deep-sea sediment $\delta^{18}O$ records (figure 2) Oxygen isotopic stages 18–35 can easily be identified in the Site 983 benthic record
y matching to other deep-sea sediment δ^{18} O records (figure 2). Orbital variability in
¹⁸O is clearly present in the Site 983 reco by matching to other deep-sea sediment $\delta^{18}O$ records (figure 2). Orbital variability in $\delta^{18}O$ is clearly present in the Site 983 record, indicating that Milankovitch forced ice-
volume changes are controlling the δ^{18} O is clearly present in the Site 983 record, indicating that Milankovitch forced ice-
volume changes are controlling the large-scale fluctuations. Mean sedimentation rates
at Site 983 are three times higher than a).

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> Figure 1. Map showing the location of ODP Site 983 at $60^{\circ}24'$ N, $23^{\circ}38'$ S and 1983 m water
denth. Site 983 is located near the head of the Gardar Drift which extends to the south-southwest Figure 1. Map showing the location of ODP Site 983 at $60^{\circ}24'$ N, $23^{\circ}38'$ S and 1983 m water
depth. Site 983 is located near the head of the Gardar Drift which extends to the south-southwest,
following the path of d depth. Site 983 is located near the head of the Gardar Drift which extends to the south-southwest, following the path of deep overflow currents (Norwegian Sea Overflow Water, NSOW, as shown
by arrows) from the Iceland–Faeroe Ridge. Dashed line shows crest of Gardar Drift. Map after
Manley & Caress (1994) and McCave *et* by arrows) from the Iceland–Faeroe Ridge. Dashed line shows crest of Gardar Drift. Map after

Manley & Caress (1994) and McCave *et al.* (1980). Bathymetry in metres.
When compared with other records (figure 2), many additional isotope events indi-
cating sub-orbital variability are well defined in the Site 983 re When compared with other records (figure 2), many additional isotop cating sub-orbital variability are well defined in the Site 983 record.
Within isotopic stage 18, we observe supplementary peaks both in the

cating sub-orbital variability are well defined in the Site 983 record.
Within isotopic stage 18, we observe supplementary peaks both in the planktic and cating sub-orbital variability are well defined in the Site 983 record.
Within isotopic stage 18, we observe supplementary peaks both in the planktic and
benthic records (figure $3a$) which have previously been observed o Within isotopic stage 18, we observe supplementary peaks both in the planktic and
benthic records (figure 3*a*) which have previously been observed only in the planktic
record of core MD900963 (figure 2), where they were benthic records (figure 3*a*) which have previously been observed only in the planktic
record of core MD900963 (figure 2), where they were interpreted to be related to
precession (Bassinot *et al.* 1994). The stage 19–18 record of core MD900963 (figure 2), where they were interpreted to be related to precession (Bassinot *et al.* 1994). The stage 19–18 transition displays a succession of light $\delta^{18}O$ peaks superimposed on the trend tow precession (Bassinot *et al.* 1994). The stage 19–18 transition displays a succession of light $\delta^{18}O$ peaks superimposed on the trend towards glacial values. These parallel, sub-orbital oscillations of the planktic and light δ^{18} O peaks superimposed on the trend towards glacial values. These parallel, sub-orbital oscillations of the planktic and benthic records imply that the isotopic
signal was rapidly transferred from surface water to depth; indicating that surface
and deep waters were strongly coupled, perhaps as a r signal was rapidly transferred from surface water to depth; indicating that surface and deep waters were strongly coupled, perhaps as a result of rapid waxing and waning of ice sheets in the near surroundings accompanied and deep waters were strongly coupled, perhaps as a result of rapid waxing and waning of ice sheets in the near surroundings accompanied by glacial meltwater
fluxes. Similar shorter-period fluctuations occur in the stage 21–20 transition, and
there are also excursions in δ^{18} O near the stage 25– Isotope stage 21 stands out as an interglacial abruptly interrupted by cool periods Isotope stage 21 stands out as an interglacial abruptly interrupted by cool periods Isotope stage 21 stands out as an interglacial abruptly interrupted by cool periods *Phil. Trans. R. Soc. Lond.* A (2000)

Figure 2. Benthic δ^{18} O records for isotopic stages 18–35 (700–1200 ka): Ocean Drilling Program
Site 983 (this paper), ODP Site 677 (Shackleton *et al.* 1990), ODP Site 607 (Ruddiman *et*
 el 1980), ODP Site 929 (Bick *Figure 2. Benthic* δ^{18} O records for isotopic stages 18–35 (700–1200 ka): Ocean Drilling Program
Site 983 (this paper), ODP Site 677 (Shackleton *et al.* 1990), ODP Site 607 (Ruddiman *et al.* 1989), ODP Site 929 (Bic Site 983 (this paper), ODP Site 677 (Shackleton *et al.* 1990), ODP Site 607 (Ruddiman *et al.* 1989), ODP Site 929 (Bickert *et al.* 1997), ODP Site 659 (Tiedemann *et al.* 1994), and the planktic δ^{18} O record from M al. 1989), ODP Site 929 (Bickert *et al.* 1997), ODP
the planktic δ^{18} O record from MD900963 (Bassinot *exaccording to Shackleton <i>et al.* (1990) and this paper.

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Figure 3. (*a*) Planktic (solid symbols) and benthic (open symbols) δ^{18} O records from ODP Site
983 compared with the Ice Volume Model (thick line) calculated after Imbrie & Imbrie (1980) Figure 3. (a) Planktic (solid symbols) and benthic (open symbols) δ^{18} O records from ODP Site
983 compared with the Ice Volume Model (thick line) calculated after Imbrie & Imbrie (1980).
(b) After initial fit of benth Figure 3. (*a*) Planktic (solid symbols) and benthic (open symbols) δ^{18} O records from ODP Site 983 compared with the Ice Volume Model (thick line) calculated after Imbrie & Imbrie (1980).
(*b*) After initial fit of b 983 compared with the Ice Volume Model (thick line) calculated after Imbrie & Imbrie (1980).

(b) After initial fit of benthic $\delta^{18}O$ record to the Ice Volume Model: output of a Gaussian filter

centred on 20 kyr (0.05 centred on 20 kyr (0.05 kyr^{-1}) with a 0.02 kyr^{-1} bandpass applied to the Ice Volume Model (thick line), benthic δ^{18} O record (thin line), and planktic δ^{18} O record (dashed line). (*c*) After centred on 20 kyr (0.05 kyr⁻¹) with a 0.02 kyr⁻¹ bandpass applied to the Ice Volume Model
(thick line), benthic δ^{18} O record (thin line), and planktic δ^{18} O record (dashed line). (c) After
tuning of the filter (thick line), benthic $\delta^{1\circ}$ O record (thin line), and planktic $\delta^{1\circ}$ O record (dashed line). (c) After tuning of the filtered (20 kyr) records in figure 2b: output of a Gaussian filter centred on 20 kyr $(0.05 \text{ kyr}$ (0.05 kyr^{-1}) with a 0.02 kyr^{-1} bandpass applied to the Ice Volume Model (thick line) and the δ^{18} O record (thin line).

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which define a precession-related tripartite stage 21. Isotope analyses from ODP Site 929 and MD900963 (figure 2) also suggest a partition of stage 21, but the Site 983 which define a precession-related tripartite stage 21. Isotope analyses from ODP Site 929 and MD900963 (figure 2) also suggest a partition of stage 21, but the Site 983 record exhibits significantly greater amplitude $(1{-$ 929 and MD900963 (figure 2) also suggest a partition of stage 21, but the Site 983 record exhibits significantly greater amplitude $(1-0.5\%)$ and differs from other deep sea records, which generally show stage 21 as a war record exhibits significantly greater amplitude $(1-0.5\%)$ and differs from other deep
sea records, which generally show stage 21 as a warm and stable period. Also note-
worthy is the interval between stages 26 and 30, wh sea records, which generally show stage 21 as a warm and stable period. Also note-
worthy is the interval between stages 26 and 30, where significant millennial scale
variability is superimposed upon five light benthic $\$ variability is superimposed upon five light benthic δ^{18} O peaks (more than 1% amplitude), attributed to precession-related forcing. Such instability, with pacing indistinguishable from that of the last glacial cycle, appears to characterize all observed tude), attributed to precession-related forcing. Such instability, with pacing indistinguishable from that of the last glacial cycle, appears to characterize all observed climate states during the Mid-Pleistocene interval, guishable from that of the last glacial cycle, appears to characterize all observed
climate states during the Mid-Pleistocene interval, suggesting that sub-orbital vari-
ability has been a fundamental part of the climate s climate states during the Mid-Pleistocene interval, suggesting that sub-orbital variability has been a fundamental part of the climate system in the North Atlantic region. This view is supported by recent results from the ability has been a fundamental part of the climate system in the North Atlantic
region. This view is supported by recent results from the Early Pleistocene (Raymo
et al. 1998) and Late Pleistocene (McManus *et al.* 1999) region. This view is supported by recent results from the Early Pleistocene (Raymo *et al.* 1998) and Late Pleistocene (McManus *et al.* 1999) in the North Atlantic region; implying that climatic instability on sub-orbita *et al.* 1998) and Late Pleistocene (McManus *et al.* 1 implying that climatic instability on sub-orbital timerity intervals throughout the Pleistocene.
Notable differences between the Site 983 records plying that climatic instability on sub-orbital time-scales existed during glacial and
erglacial intervals throughout the Pleistocene.
Notable differences between the Site 983 records and the other records spanning
e Mid-P

interglacial intervals throughout the Pleistocene.
Notable differences between the Site 983 records and the other records spanning
the Mid-Pleistocene interval (figure 2) can be attributed to the greater resolution of the Site 983 record. Local climatic processes taking place in the sub-polar North the Mid-Pleistocene interval (figure 2) can be attributed to the greater resolution
of the Site 983 record. Local climatic processes taking place in the sub-polar North
Atlantic are clearly superimposed on the global ice v of the Site 983 record. Local climatic processes taking place in the sub-polar North
Atlantic are clearly superimposed on the global ice volume signal at Site 983. As
mentioned above, some of these sub-orbital events coul mentioned above, some of these sub-orbital events could result from local processes transferring the highly negative δ^{18} O surface waters downward either through melt water injection or brine formation (Vidal *et al*. 1998; Dokken & Jansen 1999).

3. Age model

Site 983 provides the most detailed Mid-Pleistocene δ^{18} O record retrieved from the North Atlantic and it allows us to compare in detail palaeoclimatic oscillations with Site 983 provides the most detailed Mid-Pleistocene δ^{18} O record retrieved from the North Atlantic, and it allows us to compare in detail palaeoclimatic oscillations with variations in orbital forcing and thus test th Site 983 provides the most detailed Mid-Pleistocene δ^{18} O record retrieved from the North Atlantic, and it allows us to compare in detail palaeoclimatic oscillations with variations in orbital forcing and thus test th North Atlantic, and it allows us to compare in detail palaeoclimatic oscillations with variations in orbital forcing and thus test the accuracy of the orbitally derived time-
scale of Shackleton *et al.* (1990). Figure 3*a* **MATHEMATICAL,
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scale of Shackleton *et al.* (1990). Figure 3*a* shows the planktic and benthic oxygen
isotope records from Site 983 together with an isotope records from Site 983 together with an ice volume simulation using the model
of Imbrie & Imbrie (1980). We chose to use the latter as our tuning target, because
it has proved to be a powerful target for the Late P isotope records from Site 983 together with an ice volume simulation using the model of Imbrie & Imbrie (1980). We chose to use the latter as our tuning target, because it has proved to be a powerful target for the Late P 1987; Bassinot *et al.* 1994) as well as for the Early Pleistocene (Shackleton *et al.* 1990). This model assumes that the rate of climate response (growth or decay of ice sheets) is proportional to the magnitude of summe 1990). This model assumes that the rate of climate response (growth or decay of ice sheets) is proportional to the magnitude of summer insolation forcing at 65° N. We constructed this target curve (figure 3*a*) using ice sheets) is proportional
We constructed this target
Berger & Loutre (1991).
Our procedure was firs Berger & Loutre (1991).
Our procedure was firstly to match the benthic $\delta^{18}O$ record from Site 983 to

Berger & Loutre (1991).
Our procedure was firstly to match the benthic δ^{18} O record from Site 983 to
the target using linear interpolation between tie points (figure 3*a*). This proce-
dure was fairly straightforward Our procedure was firstly to match the benthic δ^{18} O record from Site 983 to the target using linear interpolation between tie points (figure 3*a*). This procedure was fairly straightforward because the precession-rel the target using linear interpolation between tie points (figure 3*a*). This proce-
dure was fairly straightforward because the precession-related oscillations are well
expressed in the $\delta^{18}O$ records from Site 983. Wh dure was fairly straightforward because the precession-related oscillations are well expressed in the $\delta^{18}O$ records from Site 983. When assigning the tie points, we gave the glacial-interglacial transitions higher pri expressed in the δ^{18} O records from Site 983. When assigning the tie points, we gave the glacial-interglacial transitions higher priority than the centres of glacial or interglacial intervals. We assumed constant sedi gave the glacial-interglacial transitions higher priority than the centres of glacial or interglacial intervals. We assumed constant sedimentation rates between tie points, resulting in a change in sedimentation rate more interglacial intervals. We assumed constant sedimentation rates between tie points,
resulting in a change in sedimentation rate more or less abruptly at these control
points (figure 4). This assumption may be realistic whe resulting in a change in sedimentation rate more or less abruptly at these control
points (figure 4). This assumption may be realistic when the tie points correspond
to glacial–interglacial transitions, when environmental points (figure 4). This assumption may be realistic when the tie points correspond
to glacial–interglacial transitions, when environmental changes might be expected
to effect the sedimentation rates. Moreover, assigning ti to glacial–interglacial transitions, when environmental changes might be expected
to effect the sedimentation rates. Moreover, assigning tie points at high-amplitude
transitions is probably more accurate than the designati *Phil. Trans. R. Soc. Lond.* A (2000) **Phil.** Trans. R. Soc. Lond. A (2000)

age (kyr)
Figure 4. Age–depth map and interval sedimentation rates for the composite section at ODP
Site 983. The thin line connecting open circles indicates interval sedimentation rates after initial Figure 4. Age-depth map and interval sedimentation rates for the composite section at ODP
Site 983. The thin line connecting open circles indicates interval sedimentation rates after initial
fit of the benthic δ^{18} O r Figure 4. Age-depth map and interval sedimentation rates tor the composite section at ODP
Site 983. The thin line connecting open circles indicates interval sedimentation rates after initial
fit of the benthic $\delta^{18}O$ r Site 983. The thin line connecting open circles indicates interval sedimentation rates after initial
fit of the benthic $\delta^{18}O$ record to the Ice Volume Model. The thick line represents the interval
sedimentation rates fit of the benthic δ^{10} or ecord to the lce Volume Model. The thick line represents sedimentation rates after final tuning using the 20 kyr filter outputs (see text). The plots (continuous thin lines) are essentially identical l
Table 1.

trough in intervals displaying considerable high-frequency, low-amplitude variability. trough in intervals displaying considerable high-frequency,
Tie points for this initial age model are given in table 1.
Down to stage 22, the benthic $\delta^{18}O$ record and the ide bugh in intervals displaying considerable high-frequency, low-amplitude variability.

e points for this initial age model are given in table 1.

Down to stage 22, the benthic $\delta^{18}O$ record and the ice volume curve are

Down to stage 22, the benthic δ^{18} O record and the ice volume curve are very similar, and a good match can be achieved interpreting stage 18 as containing two

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¹⁰³⁴ *J. E.T.ChannellandH.F. Kleiven*

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precession cycles and stage 21 as containing three precession cycles. When the extra
peak in stage 23 is tuned to a precession oscillation and the double peaks of stage precession cycles and stage 21 as containing three precession cycles. When the extra
peak in stage 23 is tuned to a precession oscillation and the double peaks of stage
29 are interpreted to be part of one precession cycle precession cycles and stage 21 as containing three precession cycles. When the extra
peak in stage 23 is tuned to a precession oscillation and the double peaks of stage
29 are interpreted to be part of one precession cycle peak in stage 23 is tuned to a precession oscillation and the double peaks of stage 29 are interpreted to be part of one precession cycle, the tuning down to stage 29 is a fairly obvious. Once the stage 29 is tuned, howeve 29 are interpreted to be part of one precession cycle, the tuning down to stage 29 is
a fairly obvious. Once the stage 29 is tuned, however, we need to assign a precession
cycle to the transition between stages 29 and 30 a fairly obvious. Once the stage 29 is tuned, however, we need to assign a precession
cycle to the transition between stages 29 and 30 in order keep pace with the Ice
Volume Model. There is some ambiguity in this interval cycle to the transition between stages 29 and 30 in order keep pace with the Ice Volume Model. There is some ambiguity in this interval in the benthic $\delta^{18}O$ record, but the planktic $\delta^{18}O$ record can be matched wit Volume Model. There is some ambiguity in this interval in the benthic δ^{18} O record,
but the planktic δ^{18} O record can be matched with confidence to the Ice Volume
Model. To respect the number of precession-related Model. To respect the number of precession-related peaks in the Ice Volume Model, we assigned two precession cycles to the major δ^{18} O peak found in stage 31, and tuned stage 33 to the low ice volume peak at 1120 ka. Model. To respect the number of precession-related peaks in the Ice Volume Model, we assigned two precession cycles to the major $\delta^{18}O$ peak found in stage 31, and tuned stage 33 to the low ice volume peak at 1120 ka. we assigned two precession cycles to the major δ^{18} O peak found in stage 31, and tuned stage 33 to the low ice volume peak at 1120 ka. As our isotope records do not extend beyond the peak of isotope stage 35, it was d tuned stage 33 to the low ice volume peak at 1120 ka. As our isotope records do
not extend beyond the peak of isotope stage 35, it was difficult to pick an end point
for our tuning procedure. But, the close correlation bet not extend beyond the peak of isotope stage 35, it was difficult to pick an end point
for our tuning procedure. But, the close correlation between Site 983 and Site 677
(figure 2) (the latter also tuned to the Ice Volume for our tuning procedure. But, the close correlation between Site 983 and Site 677 (figure 2) (the latter also tuned to the Ice Volume Model) allowed this task to be completed, and we thus conclude that our record spans t (figure 2) (tł
completed, a
to 1170 ka.
This first s mpleted, and we thus conclude that our record spans the time-interval from *ca*. 710
1170 ka.
This first step was followed by a fine-tuning of the extracted precession components
the benthic δ^{18} O record to the preces

This first step was followed by a fine-tuning of the extracted precession components of the benthic $\delta^{18}O$ record, to the precession component of the Ice Volume Model. of the benthic δ^{18} O record, to the precession component of the Ice Volume Model.
The Ice Volume Model and the benthic (and planktic) δ^{18} O record were passed
through a Gaussian filter centred on 20 kyr (0.05 kyr The Ice Volume Model and the benthic (and planktic) δ^{18} O record were passed The Ice Volume Model and the benthic (and planktic) $\delta^{18}O$ record were passed
through a Gaussian filter centred on 20 kyr (0.05 kyr⁻¹) with a 0.02 kyr⁻¹ bandwidth
(figure 3*b*). The resulting match of filter-output $\overline{\delta}$ through a Gaussian filter centred on 20 kyr (0.05 kyr^{-1}) with a 0.02 kyr^{-1} bandwidth (figure 3b). The resulting match of filter-outputs led to a slight adjustment of the original match of the unfiltered benthic (figure 3*b*). The resulting match of filter-outputs led to a slight adjustment of the original match of the unfiltered benthic $\delta^{18}O$ to the Ice Volume Model. The resulting depth-age map (figure 4) indicates that sedi original match of the unfiltered benthic $\delta^{18}O$ to the Ice Volume Model. The resulting depth-age map (figure 4) indicates that sedimentation rates vary in a range from 5 to 22 cm kyr⁻¹, tending to be lower during gla depth-age map (figure 4) indicates that sedimentation rates vary in a range from 5 to 22 cm kyr⁻¹, tending to be lower during glacial intervals. The Ice Volume Model and the final tuned benthic $\delta^{18}O$ record were the to 22 cm kyr⁻¹, tending to be lower during glacial intervals. The Ice Volume Model
and the final tuned benthic $\delta^{18}O$ record were then filtered again (figure 3*c*), and the
match between the two filtered records are and the final tuned benthic δ^{18} O record were match between the two filtered records are in accurate tuning solution has been obtained. accurate tuning solution has been obtained.
4. Magnetic properties

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Magnetic properties were largely determined from U-channel samples collected from
the archive halves of the composite section U-channels have a $2 \times 2 \text{ cm}^2$ square Magnetic properties were largely determined from U-channel samples collected
the archive halves of the composite section. U-channels have a $2 \times 2 \text{ cm}^2$ section are up to 1.5 m in length with a clin-on lid constituting Magnetic properties were largely determined from U-channel samples collected from
the archive halves of the composite section. U-channels have a $2 \times 2 \text{ cm}^2$ square
cross-section, are up to 1.5 m in length, with a clipthe archive halves of the composite section. U-channels have a $2 \times 2 \text{ cm}^2$ square cross-section, are up to 1.5 m in length, with a clip-on lid constituting one of the sides (Tauxe *et al.* 1983). Magnetic remanence of the archive halves of the composite section. U-channels have a $2 \times 2 \text{ cm}^2$ square cross-section, are up to 1.5 m in length, with a clip-on lid constituting one of the sides (Tauxe *et al.* 1983). Magnetic remanence of U-channels was measured at 1 cm intervals using the 2G Enterprises small-access passsides (Tauxe *et al.* 1983). Magnetic remanence of U-channels was measured at 1 cm
intervals using the 2G Enterprises small-access pass-through magnetometers at Gif-
sur-Yvette (France) and the University of Florida (see W sur-Yvette (France) and the University of Florida (see Weeks *et al.* 1993). The response functions of the three orthogonal magnetometer pick-up coils yield effecresponse functions of the three orthogonal magnetometer pick-up coils yield effec-
tive U-channel lengths in the sensing region of 4.2 cm for vertical and horizontal (X
and Y) and 6.2 cm for the axial direction (Z). Hence ive U-channel lengths in the sensing region of 4.2 cm for vertical and horizontal $(X$ tive U-channel lengths in the sensing region of 4.2 cm for vertical and horizontal (X and Y) and 6.2 cm for the axial direction (Z). Hence, although measurements were made at 1 cm intervals downcore, there is an inherent and Y) and 6.2 cm for the axial direction (Z) . Hence, although measurements were
made at 1 cm intervals downcore, there is an inherent smoothing in the measure-
ment procedure. Volume magnetic susceptibility (κ) was m made at 1 cm intervals downcore, there is an inherent smoothing in the measure-
ment procedure. Volume magnetic susceptibility (κ) was measured on U-channels at
1 cm intervals using a 45 mm diameter loop. Natural reman ment procedure. Volume magnetic susceptibility (κ) was measured on U-channels at 1 cm intervals using a 45 mm diameter loop. Natural remanent magnetization (NRM) was first measured in conjunction with stepwise alternat 1 cm intervals using a 45 mm diameter loop. Natural remanent magnetization (NRM) was first measured in conjunction with stepwise alternating field (AF) demagnetization. Anhysteretic remanence (ARM) and then isothermal re \sim was first measured in conjunction with stepwise alternating field (AF) demagnetization. Anhysteretic remanence (ARM) and then isothermal remanence (IRM) were imposed on the samples, and these remanences were progressivel zation. Anhysteretic remanence (ARM) and then isothermal remanence (IRM) were
imposed on the samples, and these remanences were progressively AF demagnetized.
The ARM was acquired in a 100 mT alternating field with a 0.05 imposed on the samples, and these remanences were progressively AF demagnetized.
The ARM was acquired in a 100 mT alternating field with a 0.05 mT bias DC field, and the IRM was acquired in a 500 mT DC fiel The ARM was acquired in a 100 mT alternating field with a 0.05 mT bias DC field, and the IRM was acquired in a 500 mT DC field. ARM and IRM values were used to normalize the NRM for variations in concentration of remanence and the IRM was acquired in a 500 mT DC field. ARM and IRM values were used to normalize the NRM for variations in concentration of remanence carrying grains, and hence generate the palaeointensity proxies.

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Figure 5. Orthogonal projection of alternating field demagnetization data for samples from Figure 5. Orthogonal projection of alternating field demagnetization data for samples from
Hole 983A. Open and solid symbols indicate projection on the vertical and horizontal planes,
respectively. The NRM intensity befor Figure 5. Orthogonal projection of alternating field demagnetization data for samples from
Hole 983A. Open and solid symbols indicate projection on the vertical and horizontal planes,
respectively. The NRM intensity befor Hole 983A. Open and solid symbols indicate projection on the vertical and horizontal planes,
respectively. The NRM intensity before demagnetization (J_0) is indicated, and the peak demagnetization fields are indicated in respectively. The NRM intensity before demagnetization (J_0) is indicated, and the netization fields are indicated in mT. The hole, core, section and depth in section total depth in metres below the sediment-water interf

total depth in metres below the sediment–water interface (mcd) are indicated.
Orthogonal projections of AF demagnetization data indicate that the characteristic \sim magnetization component is isolated at peak fields of 20–25 mT after removal of low Orthogonal projections of AF demagnetization data indicate that the characteristic
magnetization component is isolated at peak fields of 20–25 mT after removal of low
coercivity component attributed to the drilling proces magnetization component is isolated at peak fields of $20-25$ mT after removal of low
coercivity component attributed to the drilling process (figure 5). About 10% of the
remanence remain after demagnetization in peak remanence remain after demagnetization in peak fields of 70 mT and the median destructive field (20-30 mT) is consistent with magnetite as the principal remanence carrier. The plot of anhysteretic susceptibility agai destructive field $(20-30$ mT) is consistent with magnetite as the principal remanence destructive field (20–30 mT) is consistent with magnetite as the principal remanence
carrier. The plot of anhysteretic susceptibility against susceptibility (figure 6*a*) can
be used to assess the uniformity in grain size carrier. The plot of anhysteretic susceptibility against susceptibility (figure $6a$) can
be used to assess the uniformity in grain size of magnetite. The reasonably tight
grouping along a line emanating from the origin o *Phil. Trans. R. Soc. Lond.* A (2000)

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Figure 6. (*a*) Anhysteretic susceptibility (κ_{ARM}) plotted against volume susceptibility (*k*) for Figure 6. (*a*) Anhysteretic susceptibility (κ_{ARM}) plotted against volume susceptibility (κ) for the late Matuyama composite section at Site 983, (*b*) hysteresis ratio plot for samples from the late Matuyama c Figure 6. (a) Anhysteretic susceptibility (K_{ARM}) plotted against volume susceptibility (K) for
the late Matuyama composite section at Site 983. M_r , saturation remanence; M_s , saturation mag-
late Matuyama composite s late Matuyama composite section at Site 983. M_r , saturation remanence; M_s , saturation magnetization; H_c , coercive force; H_{cr} , remanent coercivity; SD, single domain; PSD, pseudo-single domain; MD, multidomain (plot style after Day *et al*. (1977)).

domain; MD, multidomain (plot style after Day *et al.* (1977)).
with fairly uniform magnetite grain size in the 5-10 µm range (King *et al.* 1983).
Hysteresis ratios lie in the pseudo-single domain (PSD) field (figure 6b) with fairly uniform magnetite grain size in the 5–10 μ m range (King *et al.* 1983).
Hysteresis ratios lie in the pseudo-single domain (PSD) field (figure 6*b*) according to Day *et al.* (1977) with fairly uniform
Hysteresis ratios lie
Day *et al.* (1977).
Volume magnetic vsteresis ratios lie in the pseudo-single domain (PSD) field (figure 6b) according to
y *et al.* (1977).
Volume magnetic susceptibility measured at 1 cm intervals along U-channel sam-
es varies by a factor of about 5 (fig

Day *et al.* (1977).
Volume magnetic susceptibility measured at 1 cm intervals along U-channel samples varies by a factor of about 5 (figure 7). The prominent warm isotopic stages (19, Volume magnetic susceptibility measured at 1 cm intervals along U-channel samples varies by a factor of about 5 (figure 7). The prominent warm isotopic stages (19, 21, 25 and 31) have relatively low susceptibility, due par ples varies by a factor of about 5 (figure 7). The prominent warm isotopic stages (19, 21, 25 and 31) have relatively low susceptibility, due partly to dilution by calcium carbonate. IRM and ARM also vary by a factor of a carbonate. IRM and ARM also vary by a factor of about 5, with a less pronounced carbonate. IRM and ARM also vary by a factor of about 5, with a less pronounced
correlation with the δ^{18} O record (figure 7). The magnetite grain size sensitive param-
eter ARM/ κ (figure 7) varies in a narrow rang correlation with the δ^{18} O record (figure 7). The magnetite grain size sensitive parameter ARM/ κ (figure 7) varies in a narrow range (as expected from figure 6*a*) indicating no pronounced variations in magnetite eter ARM/ κ (figure 7) varies in a narrow range (as expected from figure 6*a*) indicating no pronounced variations in magnetite grain size. The Brunhes Chronozone at Site 983 has very similar magnetic properties (see C ing no pronounced variations in magnetite grain size. The Brunhes Chronozone at

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Figure 7. Site 983 magnetic data, planktic δ^{18} O record (dashed line) and percent carbonate data plotted against depth (mcd). IRM and ARM intensities are plotted after demagnetization at peak fields of 35 m^T. Promin Figure 7. Site 983 magnetic data, planktic δ^{18} O record (dashed line) and percent carbonate data plotted against depth (mcd). IRM and ARM intensities are plotted after demagnetization at peak fields of 35 mT. Prominen plotted against depth (mcd). IRM a
peak fields of 35 mT. Prominent inte
carbonate after Ortiz *et al.* (1999). % carbonate after Ortiz *et al.* (1999).
983 are characterized by rather uniform magnetite grain size in the few micrometres

(less than $10 \mu m$) grain size range. Channell *et al.* (1998) used the thermal demagne-983 are characterized by rather uniform magnetite grain size in the few micrometres (less than 10 μ m) grain size range. Channell *et al.* (1998) used the thermal demagnetization of IRM to demonstrate the low levels of (less than 10 μ m) grain size range. Channell *et al.* (1998) used the thermal demagnetization of IRM to demonstrate the low levels of haematite in the Brunhes Chron at Site 983. The optimal conditions for palaeointensi Site 983. The optimal conditions for palaeointensity determinations (see King *et al.* 1983; Tauxe 1993) are that fine-grained magnetite be the exclusive remanence carrier and that concentrations of these grains vary by n 1983; Tauxe 1993) are that fine-grained magnetite be the exclusive remanence carrier and that concentrations of these grains vary by no more than a factor of 10. These conditions appear to be satisfied in the Brunhes and conditions appear to be satisfied in the Brunhes and late Matuyama chronozones at \sim Site 983.

5. Magnetization component directions

The magnetic overprint attributed to the drilling process is removed at peak alternating fields of $20{\text -}25$ mT (figure 5). The origin of this secondary component is not

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age (kyr)
Figure 8. Component declinations and inclinations, and corresponding maximum angular
deviation values, computed at 1 cm intervals and placed on the tuned isotope age model Figure 8. Component declinations and inclinations, and corresponding maximum angular deviation values, computed at 1 cm intervals and placed on the tuned isotope age model.

deviation values, computed at 1 cm intervals and placed on the tuned isotope age model.
well understood. It is probably a viscous remanence (VRM) produced by magnetic well understood. It is probably a viscous remanence (VRM) produced by magnetic
fields within the drill string, possibly augmented by some sort of stirred remanence
(StRM) produced by drilling disturbance. As mentioned abov well understood. It is probably a viscous remanence (VRM) produced by magnetic
fields within the drill string, possibly augmented by some sort of stirred remanence
(StRM) produced by drilling disturbance. As mentioned abov fields within the drill string, possibly augmented by some sort of stirred remanence
(StRM) produced by drilling disturbance. As mentioned above, archive halves of the
composite section were demagnetized at peak fields of (StRM) produced by drilling disturbance. As mentioned above, archive halves of the composite section were demagnetized at peak fields of 25 mT to produce the ship-
board polarity stratigraphy. In the stratigraphic in composite section were demagnetized at peak fields of 25 mT to produce the ship-
board polarity stratigraphy. In the stratigraphic interval discussed here, the NRMs of
all U-channels collected from the archive halves board polarity stratigraphy. In the stratigraphic interval discussed here, the NRMs of all U-channels collected from the archive halves of the composite section were step-
wise demagnetized in peak fields of 25, 30, 35, 40 all U-channels collected from the archive halves of the composite section were step-
wise demagnetized in peak fields of 25, 30, 35, 40, 45 and 60 mT. Further demag-
netization steps were carried out for individual U-chann netization steps were carried out for individual U-channels. In order to compute the characteristic magnetization component, the standard three-dimensional leastnetization steps were carried out for individual U-channels. In order to compute
the characteristic magnetization component, the standard three-dimensional least-
squares line-fitting routine (Kirschvink 1980) was applied the characteristic magnetization component, the standard three-dimensional least-
squares line-fitting routine (Kirschvink 1980) was applied each 1 cm downcore to the
25-60 mT demagnetization interval. The maximum angular squares line-fitting routine (Kirschvink 1980) was applied each 1 cm downcore to the
25–60 mT demagnetization interval. The maximum angular deviation values are gen-
erally less than 10° (figure 8) indicating that the comp 25–60 mT demagnetization interval. The maximum angular deviation values are generally less than 10° (figure 8) indicating that the components are well defined in this demagnetization interval. The declinations and in netization component indicate that the interval comprises the base of the Brunhes demagnetization interval. The declinations and inclinations of the characteristic magnetization component indicate that the interval comprises the base of the Brunhes
Chronozone to just below the Jaramillo Subchronozone (f netization component indicate that the interval comprises the base of the Brunhes
Chronozone to just below the Jaramillo Subchronozone (figure 8). Several intervals of
high-amplitude secular variation are apparent in the t Chronozone to just below the Jaramillo Subchronozone (figure 8). Several intervals of high-amplitude secular variation are apparent in the top of the Matuyama Chronozone and within the Jaramillo Subchronozone. The mean in high-amplitude secular variation are apparent in the top of zone and within the Jaramillo Subchronozone. The mean if
with the expected mean inclination at the sampling site. *Phil. Trans. R. Soc. Lond.* A (2000)

composite depth, metres (mcd)
Figure 9. NRM/ κ after demagnetization (of NRM) at peak fields in the 25-45 mT range,
NRM/ARM after demagnetization (of both remanences) at peak fields in the 25-45 mT range Figure 9. NRM/ κ after demagnetization (of NRM) at peak fields in the 25–45 mT range,
NRM/ARM after demagnetization (of both remanences) at peak fields in the 25–45 mT range,
and NRM/IRM after demagnetization (of both NRM/ARM after demagnetization (of both remanences) at peak fields in the $25{-}45$ mT range, and NRM/IRM after demagnetization (of both remanences) at peak fields in the $25{-}45$ mT range.

6. Palaeointensity estimates

Palaeointensity proxies are constructed by normalizing the NRM intensity by a Palaeointensity proxies are constructed by normalizing the NRM intensity by a
parameter (such as κ , ARM or IRM) to compensate for variations in the concen-
tration of remanence carrying grains. The chosen normalizer s Palaeointensity proxies are constructed by normalizing the NRM intensity by a
parameter (such as κ , ARM or IRM) to compensate for variations in the concen-
tration of remanence carrying grains. The chosen normalizer s parameter (such as κ , ARM or IRM) to compensate for variations in the concentration of remanence carrying grains. The chosen normalizer should activate the same grain population that carries the NRM. Susceptibility $(\$ tration of remanence carrying grains. The chosen normalizer should activate the same grain population that carries the NRM. Susceptibility (κ) is sensitive to large multidomain (MD) grains and small (superparamagnetic, not important remanence carriers (in the case of MD grains) or not able to carry multidomain (MD) grains and small (superparamagnetic, SP) grains which are either
not important remanence carriers (in the case of MD grains) or not able to carry
remanence (in the case of SP grains). ARM and IRM activate not important remanence carriers (in the case of MD grains) or not able to carry
remanence (in the case of SP grains). ARM and IRM activate remanence carrying
fine-grained magnetite. The mean grain size of the population o remanence (in the case of SP grains). ARM and IRM activate remanence carrying
fine-grained magnetite. The mean grain size of the population of magnetite grains
activated by IRM might be expected to be larger than for grai ie-grained magnetite. The mean grain size of the population of magnetite grains
tivated by IRM might be expected to be larger than for grains activated by ARM.
In figure 9, we plot NRM/ κ , NRM/ARM and NRM/IRM for five activated by IRM might be expected to be larger than for grains activated by ARM.
In figure 9, we plot NRM/κ , NRM/ARM and NRM/IRM for five demagnetiza-
tion steps in the 25-45 mT range. Note that the value of the ratio is In figure 9, we plot NRM/ κ , NRM/ARM and NRM/IRM for five demagnetiza-
tion steps in the 25–45 mT range. Note that the value of the ratio is determined for
a particular peak demagnetization field applied to NRM, and AR tion steps in the 25–45 mT range. Note that the value of the ratio is determined for
a particular peak demagnetization field applied to NRM, and ARM or IRM. The
variability of normalized remanence is similar for each norm a particular peak demagnetization field applied to NRM, and ARM or IRM. The variability of normalized remanence is similar for each normalizer (figure 9). Surprisingly, the NRM/ARM values are more variable than the NRM/ $\$ variability of normalized remanence is similar for each normalizer (figure 9). Surprisingly, the NRM/ARM values are more variable than the NRM/ κ or NRM/IRM values (figure 9). It appears that the NRM/ARM record is unde prisingly, the NRM/ARM values are more variable than the NRM/ κ or NRM/IRM values (figure 9). It appears that the NRM/ARM record is under normalized and that the coercivity spectra of NRM and ARM are more dissimilar th values (figure 9). It appears that the NRM/ARM record is under normalized and

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Figure 10. Site 983 virtual geomagnetic polar (VGP) latitudes computed from magnetization Figure 10. Site 983 virtual geomagnetic polar (VGP) latitudes computed from magnetization components (figure 8), mean NRM/IRM scaled to μ T and represented as virtual axial dipole moment (VADM) and normalized remanence Figure 10. Site 983 virtual geomagnetic polar (VGP) latitudes computed from magnetization
components (figure 8), mean NRM/IRM scaled to μ T and represented as virtual axial dipole
moment (VADM), and normalized remanence moment (VADM), and normalized remanence (paleointensity record) from ODP Hole 851D (Meynadier *et al.* 1994). Bathymetry in metres.

(Meynadier *et al.* 1994). Bathymetry in metres.
sity proxy. The arithmetic mean of the five values of NRM/IRM (for the 25–45 mT
demagnetization range) provides the palaeointensity proxy sity proxy. The arithmetic mean of the five values of NRM/I
demagnetization range) provides the palaeointensity proxy.
Constable & Tauxe (1996) suggested a means of scaling y proxy. The arithmetic mean of the five values of NRM/IRM (for the 25–45 mT
magnetization range) provides the palaeointensity proxy.
Constable & Tauxe (1996) suggested a means of scaling sedimentary relative
laeointensit

demagnetization range) provides the palaeointensity proxy.
Constable & Tauxe (1996) suggested a means of scaling sediments
palaeointensity records using the assumption that the axial dipole (g_1^0) at the time of reversa $\begin{smallmatrix} 0 & 0 \\ 1 & 0 \end{smallmatrix}$ goes to zero
the non-axial-Constable & Tauxe (1996) suggested a means of scaling sedimentary relative palaeointensity records using the assumption that the axial dipole (g_1^0) goes to zero at the time of reversal and that field intensity at that palaeointensity records using the assumption that the axial dipole (g_1^0) goes to zero
at the time of reversal and that field intensity at that time is due to the non-axial-
dipole field (NAD). They use an estimate of 7 at the time of reversal and that field intensity at that time is due to the non-axial-
dipole field (NAD). They use an estimate of $7.5 \mu T$ for the strength of the NAD, then
multiply the sedimentary palaeointensity record dipole field (NAD). They use an estimate of $7.5 \mu T$ for the strength of the NAD, then
multiply the sedimentary palaeointensity record by a constant factor which sets the
average transitional palaeointensity to $7.5 \mu T$. multiply the sedimentary palaeointensity record by a constant factor which sets the average transitional palaeointensity to 7.5μ T. The resulting scaled palaeointensity record for Site 983 is shown in figure 10. The ave average transitional palaeointensity to $7.5 \mu T$. The resulting scaled palaeointensity
record for Site 983 is shown in figure 10. The average palaeointensity over the scaled
record is 47.5 μ T, which is comparable with record for Site 983 is shown in figure 10. The average palaeointensity over the scaled
record is 47.5 μ T, which is comparable with the expected mean dipole field intensity at the site latitude (54.2 μ T). The average record is 47.5 μ T, which is comparable with the expected mean dipole field intensity at the site latitude (54.2 μ T). The average virtual axial dipole moment (VADM) over the record is 6.8×10^{22} A m² with value sity at the site latitude
over the record is 6.8 >
reversals (figure 10).
The MBB and the beer the record is 6.8×10^{22} A m² with values falling to ca . 1×10^{22} A m² at polarity
versals (figure 10).
The MBB and the boundaries of the Jaramillo Subchronozone occur within palaeo-
censity lows, as do se

reversals (figure 10).
The MBB and the boundaries of the Jaramillo Subchronozone occur within palaeo-
intensity lows, as do several intervals of high-amplitude secular variation (figure 10).

in

The palaeointensity low at about 793 ka that predates the directional change at the MBB (figure 10) appears to be the same as the pre-reversal palaeointensity low recorded in equatorial Pacific and North Atlantic cores (K *<u>ITHEMATICAL</u>* The palaeointensity low at about 793 ka that predates the directional change at The palaeointensity low at about 793 ka that predates the directional change at the MBB (figure 10) appears to be the same as the pre-reversal palaeointensity low recorded in equatorial Pacific and North Atlantic cores (K the MBB (figure 10) appears to be the same as the pre-reversal palaeointensity low
recorded in equatorial Pacific and North Atlantic cores (Kent & Schneider 1995; Hartl
& Tauxe 1996). Other palaeointensity records for the recorded in equatorial Pacific and North Atlantic cores (Kent & Schneider 1995; Hartl & Tauxe 1996). Other palaeointensity records for the Late Matuyama Chron are from the Pacific and Indian Oceans (Meynadier *et al.* 1994 & Tauxe 1996). Other palaeointensity records for the Late Matuyama Chron are from
the Pacific and Indian Oceans (Meynadier *et al.* 1994) and from Core K78030 from
the central equatorial Pacific (Laj *et al.* 1996; Verosu the Pacific and Indian Oceans (Meynadier *et al.* 1994) and from Core K78030 from
the central equatorial Pacific (Laj *et al.* 1996; Verosub *et al.* 1996). All these records
are from piston cores with mean sedimentation the central equatorial Pacific (Laj *et al.* 1996; Verosub *et al.* 1996). All these records
are from piston cores with mean sedimentation rates of a few (1–3) cm kyr⁻¹ in
contrast to mean sedimentation rates of *ca.* 1 are from piston cores with mean sedimentation rates of a few $(1-3)$ cm kyr⁻¹ in contrast to mean sedimentation rates of *ca*. 13 cm kyr⁻¹ in this interval at Site 983 (see figure 4). Although the records of Meynadier ROYAL contrast to mean sedimentation rates of ca. 13 cm kyr⁻¹ in this interval at Site 983 (see figure 4). Although the records of Meynadier *et al.* (1994) can be correlated from the Indian Ocean to the Pacific Oceans (see a (see figure 4). Although the records of Meynadier *et al.* (1994) can be correlated from the Indian Ocean to the Pacific Oceans (see also Valet & Meynadier 1998), these records cannot be clearly correlated in detail to th from the Indian Ocean to the Pacific Oceans (see also Valet & Meynadier 1998),
these records cannot be clearly correlated in detail to those from Core K78030, and
none of these low resolution palaeointensity records can be these records cannot be clearly correlated in detail to those from Core K78030, and
none of these low resolution palaeointensity records can be correlated in detail to Site
983. An example of these correlation problems is from central equatorial palaeointensity records can be correlated in detail to Site
983. An example of these correlation problems is shown in figure 10, where the record
from central equatorial Pacific ODP Hole 851D (Meyna THE 1 983. An example of these correlation problems is shown in figure 10, where the record
from central equatorial Pacific ODP Hole 851D (Meynadier *et al.* 1994) is shown with
the Site 983 record. The progressive decrease in from central equatorial Pacific ODP Hole 851D (Meynadier *et al.* 1994) is shown with the Site 983 record. The progressive decrease in palaeointensity within the Jaramillo Subchronozone in the Meynadier *et al.* (1994) rec PHILOSOPHICAL
TRANSACTIONS Subchronozone in the Meynadier *et al.* (1994) record was contributing evidence to Subchronozone in the Meynadier *et al.* (1994) record was contributing evidence to the 'sawtooth' hypothesis of palaeointensity in which geomagnetic palaeointensity was thought to decrease within polarity chrons until cri the 'sawtooth' hypothesis of palaeointensity in which geomagnetic palaeointensity
was thought to decrease within polarity chrons until critically low values triggered
the polarity reversal process. The hypothesis continues was thought to decrease within polarity chrons until critically low values triggered
the polarity reversal process. The hypothesis continues to be the focus of debate,
however, records showing this progressive decay in pal č the polarity reversal process. The hypothesis continues to be the focus of debate,
however, records showing this progressive decay in palaeointensity can, according
to some authors, be explained by delayed remanence acqui however, records showing this progressive decay in palaeointensity can, according
to some authors, be explained by delayed remanence acquisition (see Mazaud 1996;
Kok & Tauxe 1996; Meynadier *et al.* 1998). Recently, Valet to some authors, be explained by delayed remanence acquisition (see Mazaud 1996; Kok & Tauxe 1996; Meynadier *et al.* 1998). Recently, Valet *et al.* (1999) have made the case, based in absolute palaeointensity determinat Kok & Tauxe 1996; Meynadier *et al.* 1998). Recently, Valet *et al.* (1999) have made the case, based in absolute palaeointensity determinations from the Canary Islands, that the interval between the top of the Jaramillo the case, based in absolute palaeointensity determinations from the Canary Islands,
that the interval between the top of the Jaramillo and the MBB is characterized by
an 'overall tendency' of the field to decrease. It is i that the interval between the top of the Jaramillo and the MBB is characterized by
an 'overall tendency' of the field to decrease. It is important to note that no clear
overall decay of palaeointensity is apparent within t an 'overall tendency' of the field to decrease. It is important to note that no
overall decay of palaeointensity is apparent within the Jaramillo Subchronozone
between the Jaramillo Subchronozone and the MBB at Site 983 (f

between the Jaramillo Subchronozone and the MBB at Site 983 (figure 10).
 7. Ages of polarity reversals

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Since 1990, the ages of Late Miocene to Pleistocene polarity reversals have been
revised due to astrochronological estimates from Italian land sections (e.g. Hilgen Since 1990, the ages of Late Miocene to Pleistocene polarity reversals have been
revised due to astrochronological estimates from Italian land sections (e.g. Hilgen
1991a b) from ODP Site 677 (Shackleton *et al.* 1990) an Since 1990, the ages of Late Miocene to Pleistocene polarity reversals have been
revised due to astrochronological estimates from Italian land sections (e.g. Hilgen
1991*a*, *b*), from ODP Site 677 (Shackleton *et al.* 199 revised due to astrochronological estimates from Italian land sections (e.g. Hilgen 1991*a, b*), from ODP Site 677 (Shackleton *et al.* 1990) and from ODP Leg 138 (Shackleton *et al.* 1995). These new ages are systematica 1991*a*, *b*), from ODP Site 677 (Shackleton *et al.* 1990) and from ODP Leg 138 (Shackleton *et al.* 1995). These new ages are systematically older than the previously accepted ages (e.g. Berggren *et al.* 1985), which f (Shackleton *et al.* 1995). These new ages are systematically older than the previously accepted ages (e.g. Berggren *et al.* 1985), which followed the K-Ar chronology of Mankinen & Dalrymple (1979). The generally accepte ously accepted ages (e.g. Berggren *et al.* 1985), which followed the K-Ar chronology
of Mankinen & Dalrymple (1979). The generally accepted astrochronological ages for
the MBB and the boundaries of the Jaramillo Subchron of Mankinen & Dalrymple (1979). The generally accepted astrochronological ages for
the MBB and the boundaries of the Jaramillo Subchron (Shackleton *et al.* 1990) are
indicated in figure 11 together with the Site 983 virt indicated in figure 11 together with the Site 983 virtual geomagnetic polar (VGP) Ξ latitudes placed on the $\delta^{18}O$ age model.

The duration of the polarity reversals, as determined by the length of time for latitudes placed on the $\delta^{18}O$ age model.
The duration of the polarity reversals, as determined by the length of time for
which the VGP latitudes are less than 45° , is *ca*. 5 kyr for each reversal (figure 11).
Th The duration of the polarity reversals, as determined by the length of time for which the VGP latitudes are less than 45° , is ca. 5 kyr for each reversal (figure 11). This estimate is comparable with the magnetic dif which the VGP latitudes are less than 45° , is *ca*. 5 kyr for each reversal (figure 11).
This estimate is comparable with the magnetic diffusion time of the Earth's inner
core, estimated to be *ca*. 3 kyr, and is con This estimate is comparable with the magnetic diffusion time of the Earth's inner core, estimated to be $ca.3 \text{ kyr}$, and is consistent with the idea that diffusion through the inner core is necessary to stabilize the reve For the *base* of the *base* of the Jaramillo Subchronozone, the Site 983 estimate (figure 11) is
For the *base* of the Jaramillo Subchronozone, the Site 983 estimate (figure 11) is
positiont with the generally accepted as

the inner core is necessary to stabilize the reversing field (see Gubbins 1999).
For the *base* of the Jaramillo Subchronozone, the Site 983 estimate (figure 11) is
consistent with the generally accepted astrochronologica For the *base* of the Jaramillo Subchronozone, the Site 983 estimate (figure 11) is consistent with the generally accepted astrochronological estimate (1070 ka) (Shackleton *et al.* 1990). Singer *et al.* (1999) gave 4 consistent with the generally accepted astrochronological estimate (1070 ka) (Shack-
leton *et al.* 1990). Singer *et al.* (1999) gave ⁴⁰Ar/³⁹Ar ages from lavas recording
the Jaramillo Subchron in the Punaruu Valley, *Phil. Trans. R. Soc. Lond.* A (2000)

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Figure 11. Virtual geomagnetic polar (VGP) latitudes (open symbols) and benthic δ^{18} O record (solid symbols) across (a) the Matuyama–Brunhes boundary (MBB) and (b), (c) the boundaries of the Iaramillo Subchronozone. B Figure 11. Virtual geomagnetic polar (VGP) latitudes (open symbols) and benthic δ^{10} O record (solid symbols) across (a) the Matuyama–Brunhes boundary (MBB) and (b), (c) the boundaries of the Jaramillo Subchronozone. B (solid symbols) across (*a*) the Matuyama–Brunhes boundary (MBB) and (*b*), (*c*) the boundaries
of the Jaramillo Subchronozone. Bold lettering indicates isotopic stage numbers. Thick vertical
lines indicate the generally lines indicate the generally accepted astrochronological ages for these polarity chron boundaries (Shackleton *et al.* 1990).

the Jaramillo Subchronozone (1053 \pm 6 ka) is midway between the earlier ⁴⁰Ar/³⁹Ar
determinations of Spell & MacDougall (1992) and Izett & Obradovich (1994) For the Jaramillo Subchronozone (1053 \pm 6 ka) is midway between the earlier ⁴⁰Ar/³⁹Ar determinations of Spell & MacDougall (1992) and Izett & Obradovich (1994). For the *top* of the Jaramillo Subchronozone, the Site 98 the Jaramillo Subchronozone (1053 \pm 6 ka) is midway between the earlier ⁴⁰Ar/³⁹Ar
determinations of Spell & MacDougall (1992) and Izett & Obradovich (1994). For
the *top* of the Jaramillo Subchronozone, the Site 98 determinations of Spell & MacDougall (1992) and Lzett & Obradovich (1994). For
the *top* of the Jaramillo Subchronozone, the Site 983 estimate (figure 11) is again
close to the generally accepted astrochronological estimat

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al. 1990). The ⁴⁰Ar/³⁹Ar age given by Singer *et al.* (1999) (986 \pm 5 ka) is consistent with these estimates. The process of checking astrochronological age estimates al. 1990). The ⁴⁰Ar/³⁹Ar age given by Singer *et al.* (1999) (986 \pm 5 ka) is consistent with these estimates. The process of checking astrochronological age estimates for reversal boundaries using ⁴⁰Ar/³⁹Ar met tent with these estimates. The process of checking astrochronological age estimates
for reversal boundaries using ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ methods is somewhat circular because the
best way of calibrating the ${}^{40}\text{Ar}/{}^{39}\text{$ for reversal boundaries using ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ methods is sc
best way of calibrating the ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ standards is by
astrochronological estimates (see Renné *et al.* 1994).
The age of *onset* of the Matuvama– st way of calibrating the ⁴⁰Ar/³⁹Ar standards is by matching the ⁴⁰Ar/³⁹Ar and
trochronological estimates (see Renné *et al.* 1994).
The age of *onset* of the Matuyama–Brunhes reversal at Site 983 (775 ka) and the

astrochronological estimates (see Renné *et al.* 1994).
The age of *onset* of the Matuyama–Brunhes reversal at Site 983 (775 ka) and the midpoint of the reversal (772.5 ka) (figure 11) are younger than the mean ⁴⁰Ar/³⁹ midpoint of the reversal (772.5 ka) (figure 11) are younger than the mean ⁴⁰Ar/³⁹Ar age of 778.7 \pm 1.9 ka given by Singer & Pringle (1996). The apparent age for the MBB from Site 983 is also younger than the generall age of 778.7 \pm 1.9 ka given by Singer & Pringle (1996). The apparent age for the MBB from Site 983 is also younger than the generally accepted astrochronological estimate (780 ka) (Shackleton *et al.* 1990), and the esti age of 778.7 \pm 1.9 ka given by Singer & Pringle (1996). The apparent age for the MBB
from Site 983 is also younger than the generally accepted astrochronological estimate
(780 ka) (Shackleton *et al.* 1990), and the esti (780 ka) (Shackleton *et al.* 1990), and the estimate (778.0 \pm 1.7 ka) by Tauxe *et al.* (1996) based on 19 oxygen isotope/magnetic records combined with the ⁴⁰Ar/³⁹Ar estimate of Singer & Pringle (1996). At Site 9 (1996) based on 19 oxygen isotope/magnetic records combined with the ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ estimate of Singer & Pringle (1996). At Site 983, the MBB occurs at the young end of the isotopic stage 19 (compare figure 11 with estimate of Singer & Pringle (1996). At Site 983, the MBB occurs at the young end
of the isotopic stage 19 (compare figure 11 with fig. 3 of Tauxe *et al.* (1996)). The
sedimentary sections where the MBB ages have been de % of the isotopic stage 19 (compare figure 11 with fig. 3 of Tauxe *et al.* (1996)). The sedimentary sections where the MBB ages have been determined by astrochronology (e.g. ODP Sites 677 and DSDP Site 607) have lower se sedimentary sections where the MBB ages have been determined by astrochronology (e.g. ODP Sites 677 and DSDP Site 607) have lower sedimentation rates than Site 983. Note that ODP Site 677 does not have a polarity stratigra (e.g. ODP Sites 677 and DSDP Site 607) have lower sedimentation rates than Site 983. Note that ODP Site 677 does not have a polarity stratigraphy and the age of reversals was deduced by correlation to DSDP Hole 552A and D 983. Note that ODP Site 677 does not have a polarity stratigraphy and the age of reversals was deduced by correlation to DSDP Hole 552A and DSDP Site 607 (Shackleton *et al.* 1990). The younger apparent age of the MBB at of reversals was deduced by correlation to DSDP Hole 552A and DSDP Site 607 (Shackleton *et al.* 1990). The younger apparent age of the MBB at Site 983 could be attributed to the sedimentary record of polarity reversals b attributed to the sedimentary record of polarity reversals being shifted down-section. attributed to the sedimentary record of polarity reversals being shifted down-section
by a finite lock-in depth for magnetic remanence acquisition. The resulting time delay
of remanence acquisition might be greater for low by a finite lock-in depth for magnetic remanence acquisition. The resulting time delay
of remanence acquisition might be greater for lower sedimentation rates, as the lock-
in depth would have greater temporal significance of remanence acquisition might be greater for lower sedimentation rates, as the lock-
in depth would have greater temporal significance than in a higher sedimentation
rate sequence (such as Site 983). This possible explan in depth would have greater temporal significance than in a higher sedimentation
rate sequence (such as Site 983). This possible explanation for the relatively young
age for the MBB at Site 983 is inconsistent with the fin rate sequence (such as Site 983). This possible explanation for the relatively young
age for the MBB at Site 983 is inconsistent with the findings of Tauxe *et al.* (1996),
who concluded that the isotope-based (astrochron age for the MBB at Site 983 is inconsistent with the findings of Tauxe *et al.* (1996), who concluded that the isotope-based (astrochronological) ages for the MBB do not show any systematic variation with sedimentation ra who concluded that the isotope-based (astrochronological) ages for the MBB do not show any systematic variation with sedimentation rate, implying shallow (few cm) magnetization lock-in depths.
For the Brunhes Chron $(0-72$ show any systematic variation with sedimentation rate, implying shallow (few cm)

magnetization lock-in depths.
For the Brunhes Chron $(0-725 \text{ ka interval})$ at Site 983, Channell *et al.* (1998) documented *ca*. 100 kyr and *ca*. 41 kyr power in the NRM/IRM record and suggested that the 41 kyr power may be due *IATHEMATICAL,
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CIENCES* For the Brunhes Chron (0–725 ka interval) at Site 983, Channell *et al.* (1998) documented *ca*. 100 kyr and *ca*. 41 kyr power in the NRM/IRM record and suggested that the 41 kyr power may be due to the geomagnetic field documented *ca*. 100 kyr and *ca*. 41 kyr power in the NRM/IRM record and suggested
that the 41 kyr power may be due to the geomagnetic field itself (see also Yamazaki
1999). The rationale for this conclusion was that no s that the 41 kyr power may be due to the geomagnetic field itself (see also Yamazaki
1999). The rationale for this conclusion was that no significant 41 kyr power was seen
in the magnetic concentration parameters such as IR 1999). The rationale for this conclusion was that no significant 41 kyr power was seen
in the magnetic concentration parameters such as IRM, which would be expected to
be sensitive to lithologic/climatic variability. In t in the magnetic concentration parameters such as IRM, which would be expected to
be sensitive to lithologic/climatic variability. In these Brunhes records, 100 kyr power
was found to be ubiquitous in all magnetic concentr be sensitive to lithologic/climatic variability. In these Brunhes records, 100 kyr power
was found to be ubiquitous in all magnetic concentration parameters (including IRM)
reflecting the strong influence of the *ca*. 100 was found to be ubi
reflecting the stron
during this time.
Spectral analysis reflecting the strong influence of the *ca*. 100 kyr orbital period on climate/lithology during this time.
Spectral analysis of the mean NRM/IRM record for the 700–1100 ka interval from

during this time.

Spectral analysis of the mean NRM/IRM record for the 700–1100 ka interval from

Site 983 indicates power peaks at periods close to 41 kyr $(0.0244 \text{ kyr}^{-1})$ and just

oreater than 100 kyr $(0.01 \text{ kyr}^{-1$ Spectral analysis of the mean NRM/IRM record for the 700–1100 ka interval from
Site 983 indicates power peaks at periods close to 41 kyr $(0.0244 \text{ kyr}^{-1})$ and just
greater than 100 kyr (0.01 kyr^{-1}) (figure 12). The p Site 983 indicates power peaks at periods close to 41 kyr. $(0.0244 \text{ kyr}^{-1})$ and just
greater than 100 kyr (0.01 kyr^{-1}) (figure 12). The power spectrum for IRM features a
broad power peak close to 41 kyr. It is inter greater than 100 kyr (0.01 kyr^{-1}) (figure 12). The power spectrum for IRM features a
broad power peak close to 41 kyr. It is interesting to note the absence of 100 kyr power
in IRM, presumably reflecting the diminishe broad power peak close to 41 kyr. It is interesting to note the absence of 100 kyr power
in IRM, presumably reflecting the diminished influence of this period on climate at
this time (in the so-called 41 kyr world) relativ \bigcirc in IRM, presumably reflecting the diminished influence of this period on climate at
 \bigcirc this time (in the so-called 41 kyr world) relative to its strong influence in the Brunhes.

The squared coherency between N The squared coherency between NRM/IRM and IRM is not significant at periods
The squared coherency between NRM/IRM and IRM is not significant at periods
greater than *ca*. 12 kyr (figure 12) implying that the 41 kyr power i The squared coherency between NRM/IRM and IRM is not significant at periods
greater than $ca.12 \text{ kyr}$ (figure 12) implying that the 41 kyr power in NRM/IRM is
not be due to climatic influence on the palaeointensity record greater than $ca. 12 \text{ kyr}$ (figure 12) implying that the 41 kyr power in NRM/IRM is
not be due to climatic influence on the palaeointensity record (through IRM) but
rather to the geomagnetic field itself. The $ca. 100 \text{ kyr$ not be due to climatic influence on the palaeointensity record (through IRM) but
rather to the geomagnetic field itself. The ca 100 kyr power in NRM/IRM is also,
in the records documented here, not easily attributed to c rather to the geomagnetic field itself. The ca 100 kyr power in NRM/IRM is also in the records documented here, not easily attributed to climate/lithology as it absent in IRM, and therefore, may also be attributed to the *Phil. Trans. R. Soc. Lond.* A (2000)

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Figure 12. Power spectra using the Blackman–Tukey method with a Bartlett window for NRM/IRM (black continuous line) and $\text{IRM}_{35\text{ mT}}$ (grey continuous line) in the 700–1100 ka Figure 12. Power spectra using the Blackman–Tukey method with a Bartlett window for NRM/IRM (black continuous line) and IRM_{35 mT} (grey continuous line) in the 700–1100 ka
interval at Site 983. Squared coherence between NRM/IRM (black continuous line) and IRM_{35 mT} (grey c
interval at Site 983. Squared coherence between NRM/IRM
insignificant coherence at frequencies less than 0.08 kyr⁻¹.

insignificant coherence at frequencies less than 0.08 kyr^{-1} .
We thank N. J. Shackleton and D. A. Hodell for mass spectrometry support and advice concern-We thank N. J. Shackleton and D. A. Hodell for mass spectrometry support and advice concerning the age model during research stays (H.F.K.) at University of Cambridge and University of Florida. We are very grateful to E. J We thank N. J. Shackleton and D. A. Hodell for mass spectrometry support and advice concerning the age model during research stays (H.F.K.) at University of Cambridge and University of Florida. We are very grateful to E. J Florida. We are very grateful to E. Jansen, C. Laj, C. Kissel and A. Mazaud for logistical support and scientific advice. C. Laj, B. Clement and N. J. Shackleton provided helpful comments on
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Program for ODP Leg 162 and by the Norwe the manuscript. This work was supported by NSF grant EAR 98-04711, the US Science Support

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