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Magnetic record of Lake Baikal sediments: chronological and paleoclimatic implication for the last 6.7 Myr

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Abstract

Magnetic remanence vectors for 1472 samples taken from a 601 m core through Lake Baikal sediments are reported along with a complete magnetic susceptibility profile obtained from a pass-through system. Matching the stable remanence directions to the standard geomagnetic polarity time scale (GPTS) provides a robust chronology from the present back to ~6.7 Ma and yields a remarkably constant sediment accumulation rate of 3.9 cm/kyr. For earlier times – represented by depths >270 m – correlation to the GPTS is more problematic. Susceptibility fluctuations reflect climatic changes that can be matched to the marine oxygen isotope pattern for the last 6.7 Myr. Spectral analysis of the resulting susceptibility time series then indicates that, for the most part, the Milankovitch obliquity signal dominates. However, when the temporal evolution of the frequency content is investigated by analyzing sequences of time windows, a complex picture emerges in which eccentricity and precession power appear during some intervals. Furthermore, there is persistent evidence for significant power in a 'non-Milankovitch' band between 28 and 35 kyr.

Keywords: Lake Baikal; magnetic susceptibility; Milankovitch cycles; paleomagnetism; Siberia

1. Introduction

Lake Baikal is the largest lake in the world, containing some 23 000 km^3 of water compared to a combined total of 22 000 km^3 for all five of

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the Great Lakes of North America. More importantly, it is the world's oldest lake, and it was never completely frozen during past glacial maxima. This makes it a prime target for the investigation of past global change, for which reason it has been intensively studied over the last decade (King et al., 1993; Peck et al., 1994, 1996; Williams et al., 1997; Antipin et al., 1998; Kuzmin et al., 1998; Kravchinsky et al., 1998; Dobretsov, 1999; Sakai et al., 2000; Minoura, 2000; Sakai et al., 2003). In 1998, the Russian–Japanese–American Baikal Drilling Project drilled a new borehole, BDP-98, on the Academician Ridge at 108°24′34″E, 53°44′48″N in a water depth of 333 m (Fig. 1). From the lake floor to a depth of 270 m, sediments were obtained by piston coring. From 270 m down to 601 m, rotary drilling was used. The hole was actually drilled down to 670 m, but no coring was done below 601 m. Average recovery was 95%. Previous drilling on the Academician Ridge in the winter of 1995–1996 resulted in the recovery of two cores, BDP-96-1 (maximum depth, 200 m) and BDP-96-2 (maximum depth, 100 m). Between them, these cores provided an excellent regional paleoclimate archive based on biogenic silica and diatom records reaching back 5 million years (Antipin et al., 1997; Antipin et al., 1998; Williams et al.,

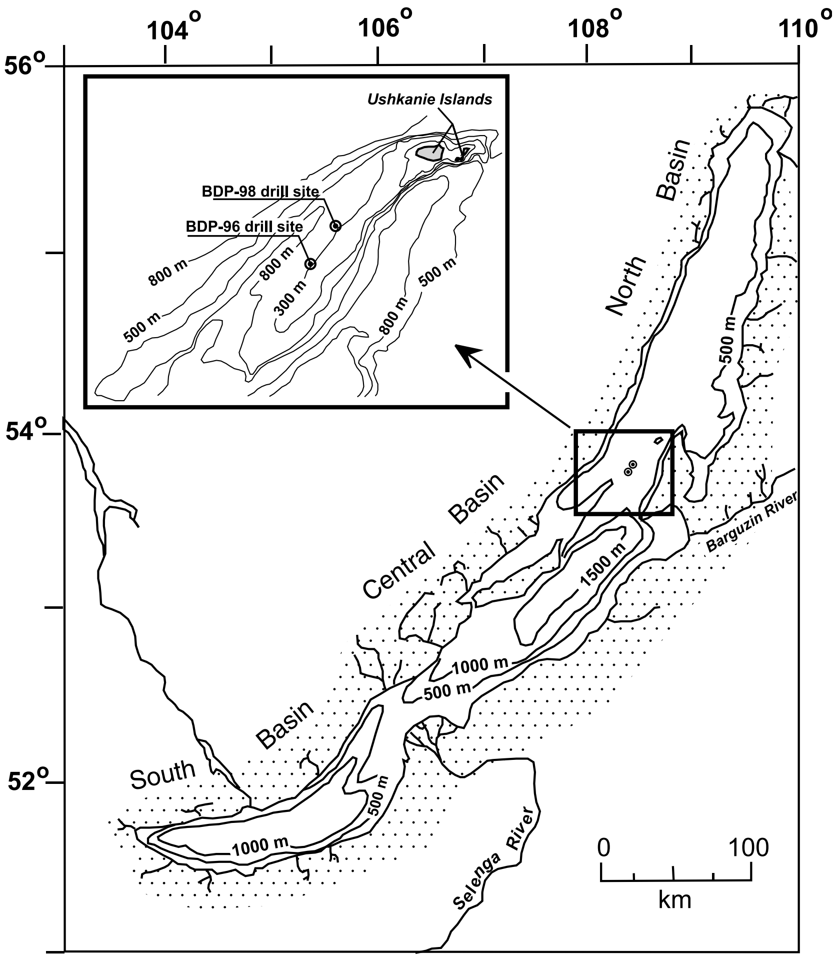

Fig. 1. Simplified map of Lake Baikal showing the 1996 and 1998 drill sites.

1997; Kuzmin et al., 2000). The aim of the 1998 drilling was to extend the Baikal sampling deeper into the Miocene sediments. The site's distant location from coarse deltaic deposits, which were identified in seismic profiles in the southeastern part of the Academician Ridge, was an important factor in selecting the new drill site.

A total of 5400 samples were taken in small plastic boxes (some 5 cm^3, others 8 cm^3) at approximately 10 cm intervals throughout the core. These were shared equally among the three participating laboratories in Irkutsk, Toyama, and Rhode Island, and thereafter measured independently. Preliminary results of this cooperative work were published by Antipin et al. (2001). Here, we report in detail the results – and interpretation – of the Russian measurements involving 1472 samples.

2. Methods

The natural remanent magnetization (NRM) of all the samples was measured with a spinner magnetometer (JR-4) or an astatic magnetometer (LAM-24). Step-wise alternating field (AF) demagnetization of pilot samples, up to 100 mT, was done with a custom-made demagnetizer. One hundred and fourteen pilot samples representing all lithological types were taken at various depths. Zijderveld diagrams and equal-area projections for these, based on at least 15 demagnetization steps, were analyzed to find appropriate demagnetization treatment for the remaining samples. These were all subjected to fields of 5, 10, and 20 mT, and in the majority of cases to 40 and 100 mT also. As a check, 20 additional pilot samples were measured and demagnetized with a three-axis 2-G cryogenic magnetometer in the paleomagnetic laboratory at the Institut de Physique du Globe de Paris. In all cases, the Paris and Irkutsk measurements were mutually compatible.

Low-field, whole-core magnetic susceptibility was measured at 3 cm intervals with a Bartington Instruments susceptibility meter generating an AF of 8 μT. A pass-through loop sensor operating at a frequency of 0.565 kHz allowed entire sections of core to be measured rapidly and non-destructively (King et al., 1993; Antipin et al., 1998).

3. Results and discussion

3.1. Geomagnetic polarity stratigraphy

The results of step-wise AF demagnetization indicate that the sediments carry a stable primary remanence with a relatively high intensity of magnetization (median = 2.2 mA/m), and a very small secondary viscous component. Typical demagnetization diagrams are shown in Fig. 2. Since the core was not oriented azimuthally, we restrict attention to the inclination values. For magnetostratigraphic purposes, this is not a severe limitation because primary inclinations are expected to be steep (~70°), making it easy to distinguish between normal and reversed polarities. However, we experienced much more difficulty in correlating the results from the rotary-drilled samples to the geomagnetic polarity time scale (GPTS) than for the piston-cored samples. It is convenient, therefore, to discuss the two types separately.

3.1.1. Piston cores (0–270 m)

After 10 mT demagnetization, the 911 samples from this upper section show a clear bimodal distribution of inclinations, with peaks near +70° and −70° (Fig. 3A). There is a significant number of intermediate values that represent either unresolved primary components or real geomagnetic features (transitional field vectors, excursions, secular variation). At 20 mT, the histogram looks essentially the same; indeed, the inclination differences ($I_{10mT}-I_{20mT}$) are less than 20° for 96% of the samples.

In Fig. 4, the 0–270 m inclination profile is correlated to the GPTS of Cande and Kent (1995). We begin by identifying the Brunhes, Matuyama, Gauss and Gilbert chrons. The lower boundaries of these are interpreted to lie at depths of 31.7, 95, 140 and 230 m, respectively. A regression line through these four points yields $d = -1.19 + 39.1a$, where d is the depth in meters and a is the GPTS age in millions of years. The fit is extremely good ($r = 0.999$). The slight zero-age

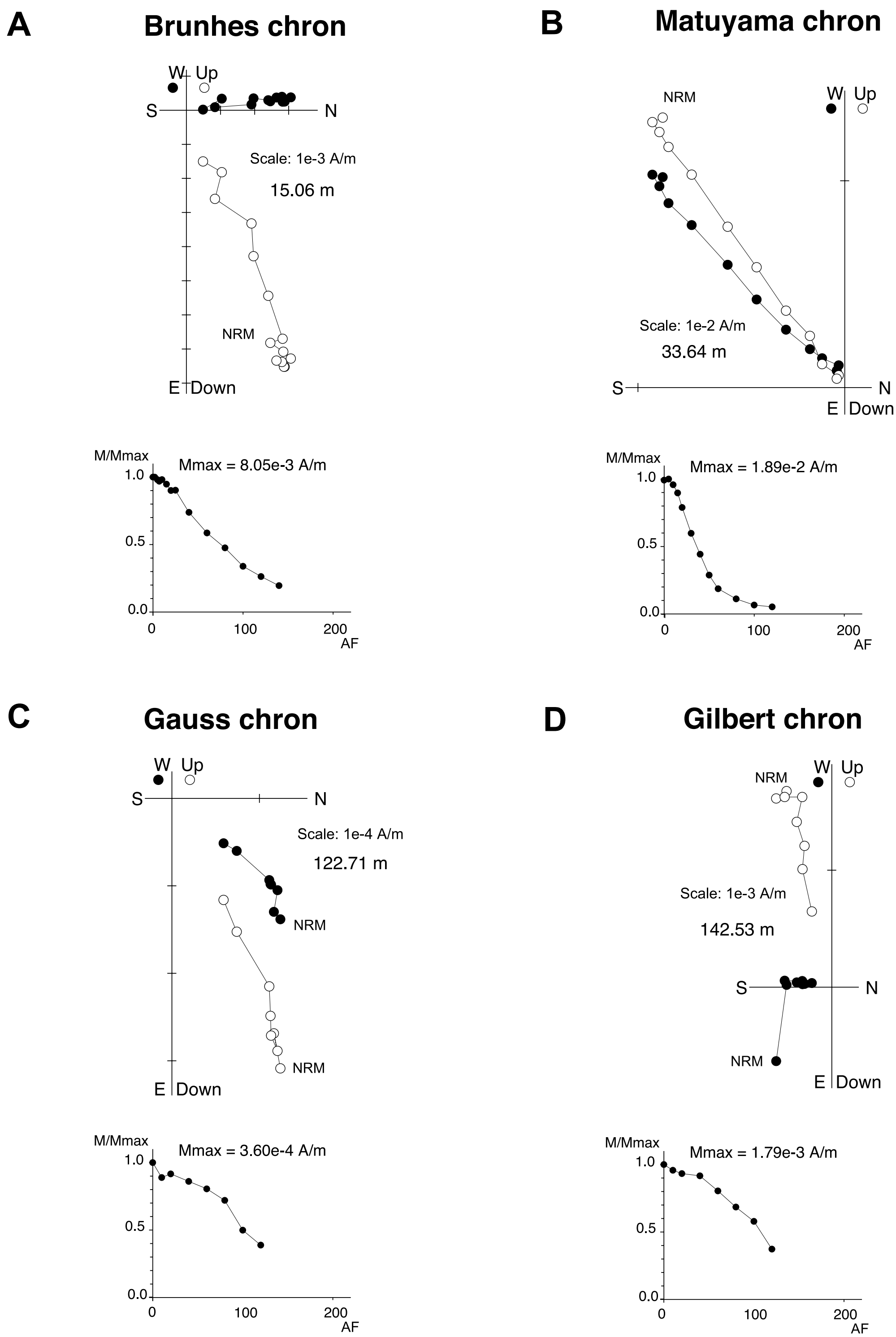

Fig. 2. Typical AF (in mT) demagnetization results. In the orthogonal vector component plots, closed/open symbols represent projections onto the horizontal/vertical plane.

misfit can be eliminated by forcing the line through the origin, which yields $d = 38.8a$ (r still equals 0.999). A zero age for the top of the core is consistent with radiocarbon data obtained from several short (4–10 m) gravity cores previously collected around the BDP-98 drilling site (Colman et al., 1996). Having established this first-order chronology, we now add in our interpreted depths for the Jaramillo, Olduvai, Kaena, Mammoth, Cochiti, Nunivak, Sidufjall, Thvera, C3A.n1 and C3A.n2 subchrons. The new regression line ($d = 39.4a$) is slightly worse ($r = 0.997$), but the correlation remains very high and again indicates that the long-term average deposition rate is 3.9 cm/kyr. By extrapolation, the deepest piston-core sample (270 m) has an age of 6.7 Ma (Fig. 5A). The relevant information is summarized in Table 1, which includes the corresponding data for cores BDP-96-1 and 2. On the right-hand side of Fig. 4, the correlation with the GPTS is extended from BDP-98 to the 1996 cores. The agreement is excellent.

Several short polarity zones may also be present (e.g. post-Jaramillo, Reunion, etc.; see Fig. 4). These are potentially important for understanding geomagnetic field behavior, but add little to the chronological control we seek here. We exclude them from consideration until results are available from other cores currently under investigation.

3.1.2. Rotary cores (270–601 m)

After 10 mT demagnetization, the samples from this part of the core yield a more confused picture. As can be seen in Fig. 3B, the two expected polarity peaks in the inclination data are much less clear than in the upper part of the core (Fig. 3A). Indeed, the entire distribution of inclinations obtained from the rotary coring is unsatisfactory. The expected normal-polarity peak is rather broad and weak, there is a strong peak at low negative values and even the expected reversed peak is shifted towards shallower values. The marked excess of negative inclinations indicates that wholesale normal (Brunhes) overprinting is not the problem. Nor can one appeal to insufficient cleaning; more than 99% of the inclination values after 20 mT treatment differ by less

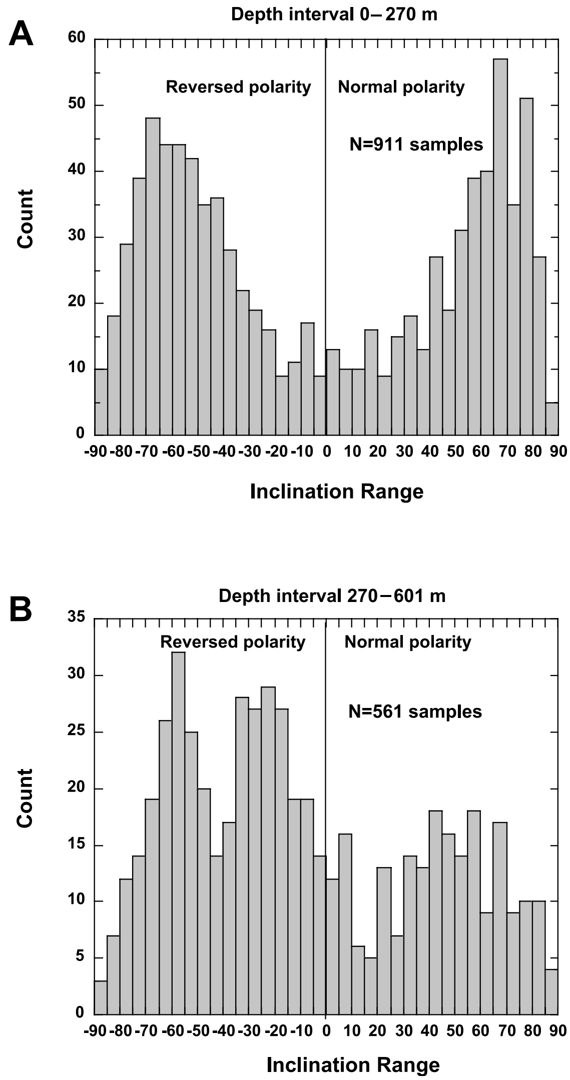

Fig. 3. Histograms of inclination values after 10 mT AF treatment (panel A for depth interval 0–270 m, panel B for 270–601 m).

than 20° from the 10 mT value. Two remaining possibilities are that the drilling process itself seriously degrades the record, or that geological factors (lithology, sedimentology, tectonics) play a role. Drilling overprints in ODP (Ocean Drilling Program) sediment cores have been reported by Sigurdsson et al. (1997), King et al. (2000) and Fuller et al. (2001). Even after AF treatment in 80 mT, they still find some evidence for the persistence of a radial overprint. Such a component

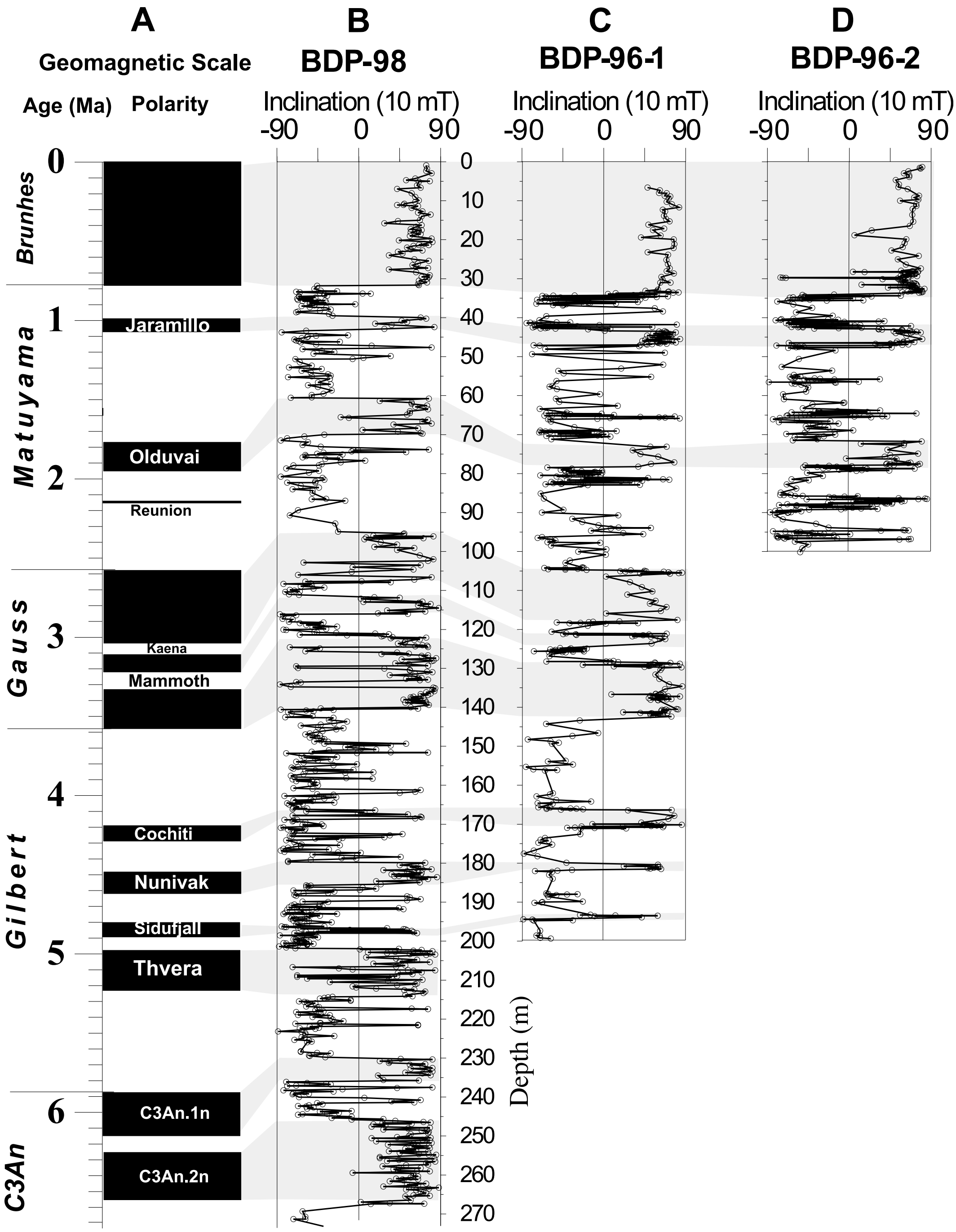

Fig. 4. Matching the inclination profiles for BDP-98 (0–270 m, panel B) and BDP-96 (holes 1 (panel C) and 2 (panel D)) to the reference geomagnetic polarity scale (panel A, after Cande and Kent (1995)). All inclination values are after 10 mT AF demagnetization.

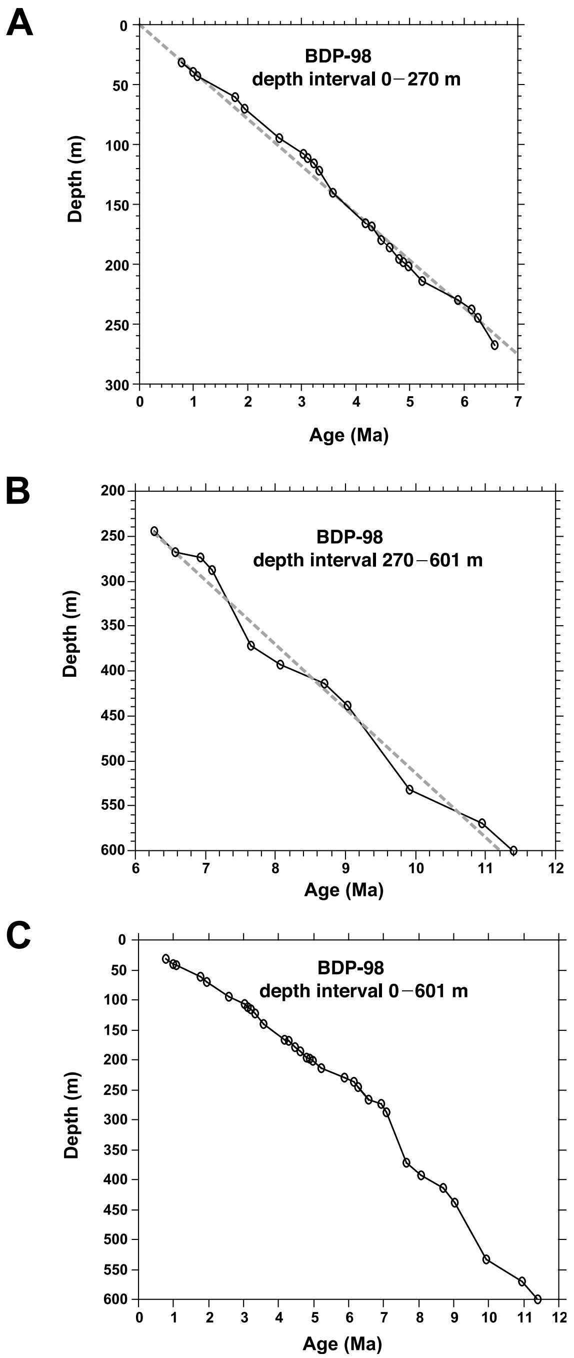

Fig. 5. Age–depth relationships for the intervals 0–270 m (A), 270–601 m (B), and for the entire hole (C).

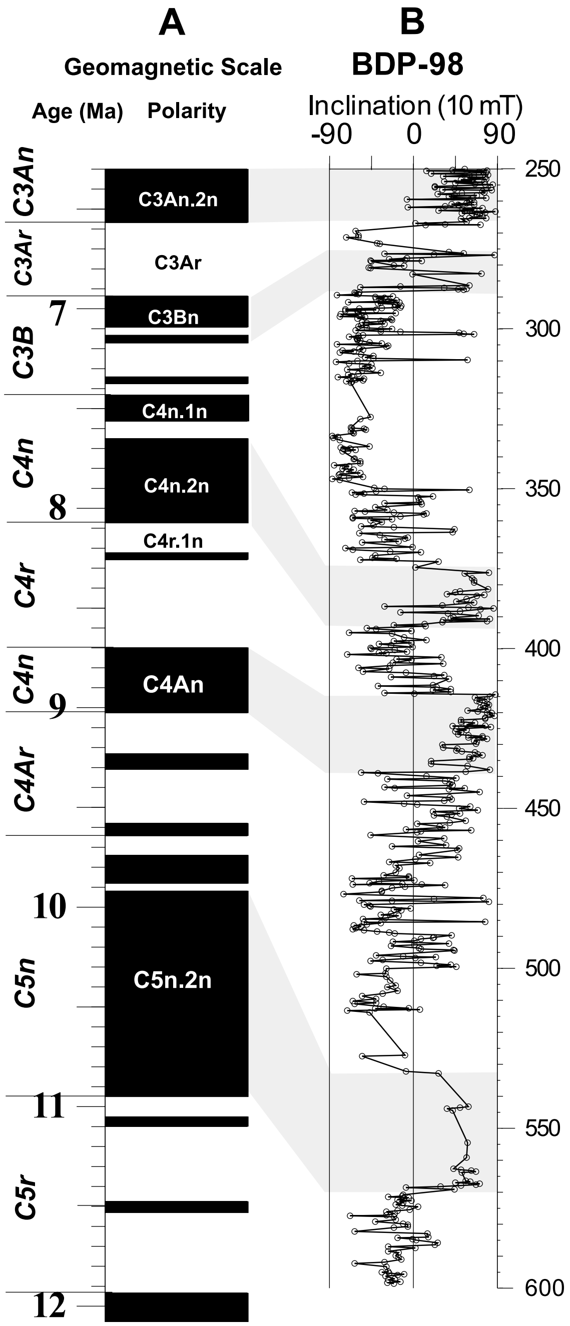

Fig. 6. Matching the inclination profile for BDP-98 (270–601 m, panel B) to the reference geomagnetic polarity scale (panel A, after Cande and Kent (1995)). All inclination values are after 10 mT AF demagnetization.

Table 1
Dating of sediments penetrated by BDP-98 and BDP-96-1 and 2 on Academician Ridge

Geomagnetic chron/event	Age (Ma)	Hole BDP-98 Depth (m)	Hole BDP-96-1 Depth (m)	Hole BDP-96-2 Depth (m)
Piston cores (0–270 m)				
Brunhes/Matuyama	0.780	31.7	33.5–33.7	33.8–34.15
Jaramillo	0.990/1.070	39.7/42.8	43.4/47	42.7/47.1
Olduvai	1.770/1.950	60.6/70	71.5/78.3	71.6/78–79.4
Matuyama/Gauss	2.581	95	104.5–104.8	
Kaena	3.040–3.110	107.5/111.5	117.7–120.7	
Mammoth	3.220–3.330	116/122	124.0–127	
Gauss/Gilbert	3.580	140	143	
Cochiti	4.180–4.290	166/168.7	166.3–171	
Nunivak	4.480–4.620	179.8/186.2	180.5–181.7	
Sidufjall	4.800–4.890	196/198	192–196	
Bottom of BDP-96 core	4.970		200	
Thvera	4.980/5.230	202/214		
C3A.n1	5.894/6.137	230/237.4		
C3A.n2	6.269/6.567	245/267.4		
Two alternative models for rotary cores (270–601 m) in hole BDP-98				
1.				
Bottom of BDP-98 core (by extrapolation)	~15.270	601		
2.				
C3Bn	6.935/7.091	274/288		
C4n.2n	7.650/8.072	372.5/393		
C4An	8.699/9.025	414.2/438.5		
C5n.2n	9.920/10.949	532.5/570		
Bottom of BDP-98 core	~11.240	601		

GPTS after (Cande and Kent, 1995).

would certainly have the potential to bias the Lake Baikal inclination pattern towards shallower values. As far as local geological factors are concerned, there is a strong possibility that the rate of sediment accumulation varied markedly, particularly in the lower parts of hole BDP-98. The average concentration of the sand–silt fraction gradually increases from 200 to 601 m depth. Especially the lowermost interval (480–601 m) differs considerably from the overlying sediments in terms of high contents of silt, sand and gravel. This lower interval also has lower water content and higher sediment densities, and is characterized by abundant fossilized plant fragments, by sand–silt interbeds and by turbidite beds (Antipin et al., 2001). These lithologic features indicate that the sediments of this interval were deposited in a near-shore environment with high inputs of terrigenous clastic material. At this time, the drill site was apparently located on a distal slope of the delta of the Paleo-Barguzin River (Zonenshain et al., 1993; Moore et al., 1997). Above 380 m, the steady increase in diatom content is paralleled by the continued decrease in the coarse fraction. This facies change represents the transition to hemipelagic sedimentation from the water column in response to the extension and deepening of the basin separating Academician Ridge from the southeastern shore of Lake Baikal. These geological factors are unlikely to be the source of the unsatisfactory inclination distribution, but the varying sedimentation rates they gave rise to could help explain the difficulty in matching the observed polarities to the GPTS.

Despite these difficulties, we attempt in Fig. 6 to correlate the inclination pattern in the lower part of the core to the GPTS of Cande and Kent (1995). First, however, we note that the (albeit unsatisfactory) inclination pattern is not only stable to 20 mT (as pointed out above), but remains essentially unchanged up to 40 mT. The correlation indicated has several difficulties, but none of the other possibilities we have tried offers any worthwhile improvements. The two most serious problems are the long (apparently) reversed intervals between ~290 and ~370 m and between ~440 and ~530 m. The resulting best-fit chronology is illustrated in Fig. 5B and implies a more rapid average sediment accumulation rate of 7.1 cm/kyr, which is consistent with the generally coarser lithology. The regression line predicts an age of 11.24 Ma at a depth of 601 m, considerably less than the 15.27 Ma obtained by simply extrapolating the accumulation rate in the upper 270 m of the core (Fig. 5C).

3.2. Magnetic susceptibility stratigraphy

King et al. (1993), Peck et al. (1994, 1996) and Kravchinsky et al. (1998) have all shown that a strong inverse correlation exists between the magnetic susceptibility and the biogenic silica content of Lake Baikal sediments. This results from the fact that non-magnetic diatomaceous organisms – which produce the silica – are much more abundant during warm periods than during cold periods. During warm intervals therefore there is a significant increase in the amount of biogenic silica deposited, which has a diluting effect that re-

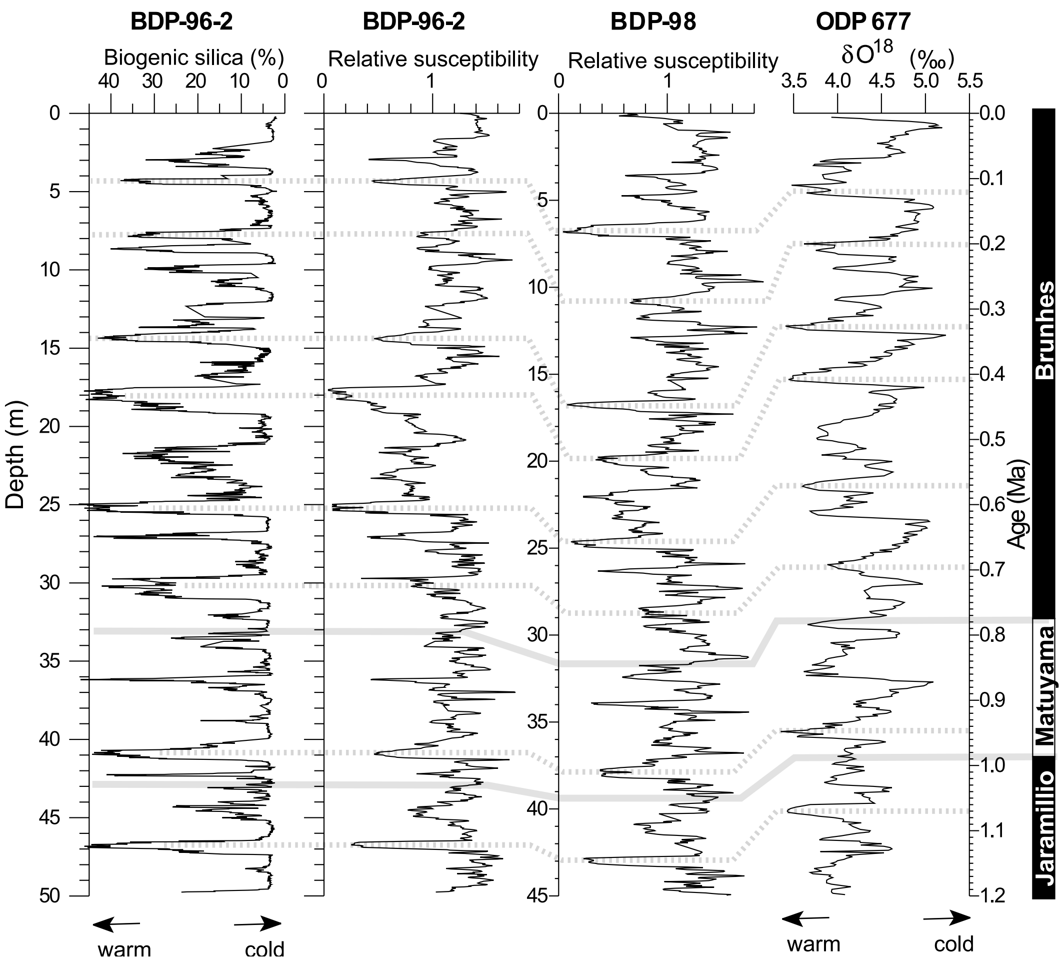

Fig. 7. Correlation between magnetic susceptibility profile from hole BDP-98, biogenic silica (Williams et al., 1997) and magnetic susceptibility (Krainov et al., 2001) for hole BDP-96-2, oxygen isotope data from ODP-677 (Shackleton et al., 1990) and the geomagnetic polarity scale (Cande and Kent, 1995). Biogenic silica values are given in wt%. Susceptibilities are plotted on a logarithmic scale. The computer program of Paillard et al. (1996) was used for establishing the correlations shown.

duces the magnetic susceptibility. A correlation is therefore expected between Lake Baikal magnetic susceptibility and oceanic oxygen isotope records. The BDP-96 magnetic data and the ODP-677 isotope results indicate that this is indeed the case (Williams et al., 1997; Antipin et al., 1998; Kuzmin et al., 1998; Peck et al., 1996; Kravchinsky et al., 1998). Despite these strong correlations, it is important to realize that dilution by biogenic silica is not the only factor able to affect the magnetic susceptibility signal. As Dearing et al. (1998) point out in their study of six short (< 40 cm) cores of surface sediments in Lake Baikal, one should consider a variety of internal and external processes, including the formation of new minerals, the dissolution of existing minerals, and changes in aeolian input. Nevertheless, the correlation between magnetic susceptibility and biosilica is sufficiently convincing that it is worth trying to use it to improve our chronological control, as follows.

The inverse relationship between silica and susceptibility is well illustrated by the BDP-96-2 data for the last 1.2 Myr, as can be seen in Fig. 7, which also includes the corresponding susceptibility results from BDP-98. The correlation between the two susceptibility profiles is extremely good, and the inverse relationship between the susceptibility and silica is clearly seen (note the reversed scale for the silica values). Synchronous increases of biogenic silica and decreases of susceptibility correspond to warm periods and are well correlated with the ODP-677 oxygen isotope curve (Shackleton et al., 1990). To match up the curves and improve the chronological control for the Baikal cores, we first assumed that the age of the sediments varies linearly between the tie points provided by the reversal stratigraphy, and only then correlated maxima and minima within each polarity zone. The final tie points used are given in Table 2, and some of them are illustrated in Fig. 7 as a visual guide. The magnetic and isotope chronologies are highly compatible, implying that delayed magnetic lock-in is not a significant problem in the Academician Ridge sediments.

For BDP-96-2, the final correlation coefficient between susceptibility and biosilica is −0.711, between biosilica and $\delta^{18}O$ it is −0.463, and between susceptibility and $\delta^{18}O$ it is +0.502. As can be seen in Fig. 7, there are significant differences between the precise shapes and amplitudes of the continental and oceanic records. For example, in the oceanic record, the $\delta^{18}O$ signal changes rather gradually from warm to cold intervals (giving rise to the well-known 'saw-tooth' pattern), whereas in the Baikal record the corresponding changes are generally more symmetrical with abrupt onset and termination of warm events. We then continued the matching process by correlating the magnetic susceptibility data of the BDP-98 core with the ODP-677 oxygen isotope curve. In this case, however, we extended the procedure down to the base of the Matuyama chron, which is the oldest polarity boundary captured in ODP-677. The resulting correlation coefficient is +0.505, essentially the same as that obtained for BDP-96-2.

Table 2
BDP-98 depth–age tie points

Depth (m)	Age (ka)	Depth (m)	Age (ka)
0	0	47.37	1199.3
3.57	48.466	48.60	1242.0
4.77	81.612	51.12	1314.6
6.86	121.16	53.33	1396.1
8.62	153.58	54.65	1439.3
10.67	199.29	55.60	1486.8
11.71	210.67	56.05	1520.0
12.91	238.00	56.65	1556.0
15.10	286.00	57.76	1613.2
16.78	326.00	59.26	1684.0
18.07	355.23	60.6[a]	1770[a]
19.81	400.60	63.19	1835.1
22.08	490.11	64.57	1869.0
23.31	529.63	67.53	1925.1
24.64	570.00	70[a]	1950[a]
26.32	611.88	70.28	1989.5
27.70	668.53	71.36	2036.6
28.50	692.61	72.98	2117.6
30.83	741.09	74.00	2145.0
31.7[a]	780.00[a]	76.01	2186.0
33.91	854.96	77.87	2227.1
37.93	950.72	82.63	2346.0
39.7[a]	990[a]	87.07	2432.0
41.31	1030.20	90.54	2518.0
42.8[a]	1070[a]	92.07	2539.9
46.54	1167.7	95[a]	2581[a]

[a] Polarity reversal boundaries.

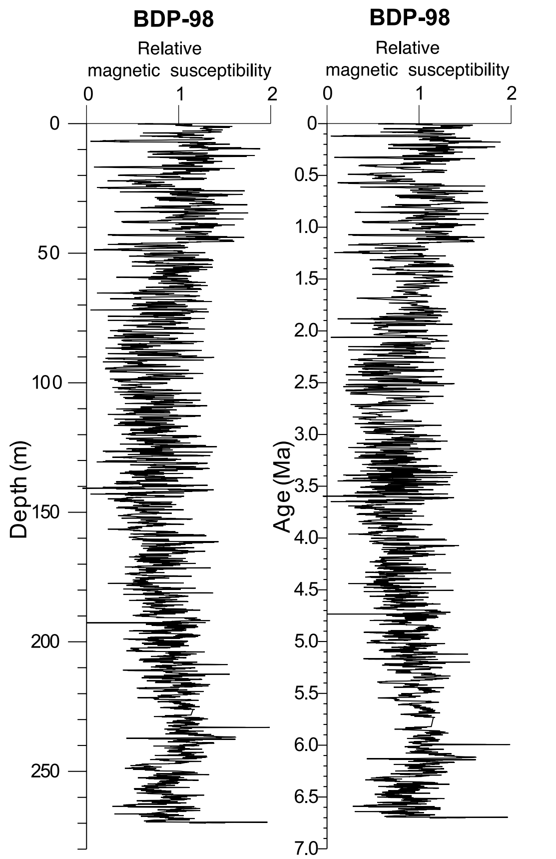

Fig. 8. The observed profile of susceptibility vs. depth for the interval 0–270 m in hole BDP-98. On the right, the corresponding profile of susceptibility vs. time is illustrated using the chronology derived in the text.

For pre-Matuyama times (below 95 m, see Table 1), we attempted to correlate our susceptibility data with longer ODP records, namely ODP-846 (Shackleton et al., 1995) and ODP-849 (Mix et al., 1995). However, the procedure was found to be unworkable because of the high frequency of fluctuations in both susceptibility and oxygen isotope records. There are simply so many options that unique tie lines cannot be drawn. Therefore our chronology for depths between 95 and 270 m is based entirely on the geomagnetic polarity boundaries. The final chronology converting susceptibility vs. depth into susceptibility vs. time is presented in Fig. 8.

3.3. *Spectral analysis*

To investigate further the climatic cyclicity of central Asia, we have performed a variety of spectral analyses of the susceptibility data from core BDP-98 and also of the biosilica and susceptibility records of core BDP-96-2 using the biosilica data of Williams et al. (1997) and the susceptibility data of Krainov et al. (2001). For the 1996 core, we investigate the entire 2.5 Myr represented, but in the 1998 core we restrict attention to the upper, piston-cored, part. In this case, we arbitrarily curtail the time series at 6.7 Ma, slightly below the base of subchron C3An.2n (see Fig. 4).

Before calculating the power spectra, we applied a least-squares smoothing from which we extracted equally spaced data points (1 kyr intervals) by an integration method using a linear function (Paillard et al., 1996). We then filtered the record using a band-pass filter (11–5000 kyr) centered on a period of 22 kyr for suppression of high- and low-frequency noise. The power spectra were then calculated by the Blackman–Tukey method with a Bartlett window, using the computer program of Paillard et al. (1996).

First, we analyzed the entire 6.7 Myr of BDP-98 (Fig. 9A) compared to the 2.5 Myr of core BDP-96-2 (Fig. 9B, susceptibility; Fig. 9C, biogenic silica). The ∼40 kyr obliquity signal is clearly seen in all three records, but there is no strong evidence for either a ∼20 kyr precession signal or a ∼100 kyr eccentricity signal. There is some power near the expected eccentricity band in the biosilica record, but nothing significant appears in either of the magnetic records. However, three peaks at periods between 28 and 35 kyr appear consistently on all the records. At the low-frequency end of the spectra, there is a clear peak in Fig. 9A near the expected 400 kyr eccentricity signal, but this peak is not found in the shorter span of 2.5 Myr represented in core BDP-96-2 (Fig. 9B,C). Finally, another long-term signal at ∼715 kyr is seen in the magnetic susceptibility record of BDP-98 (Fig. 9A) and in the biosilica data of BDP-96-2 (Fig. 9C), but does not emerge clearly from the BDP-96-2 susceptibility record (Fig. 9B).

Since the average deposition rate is ∼4 cm/kyr

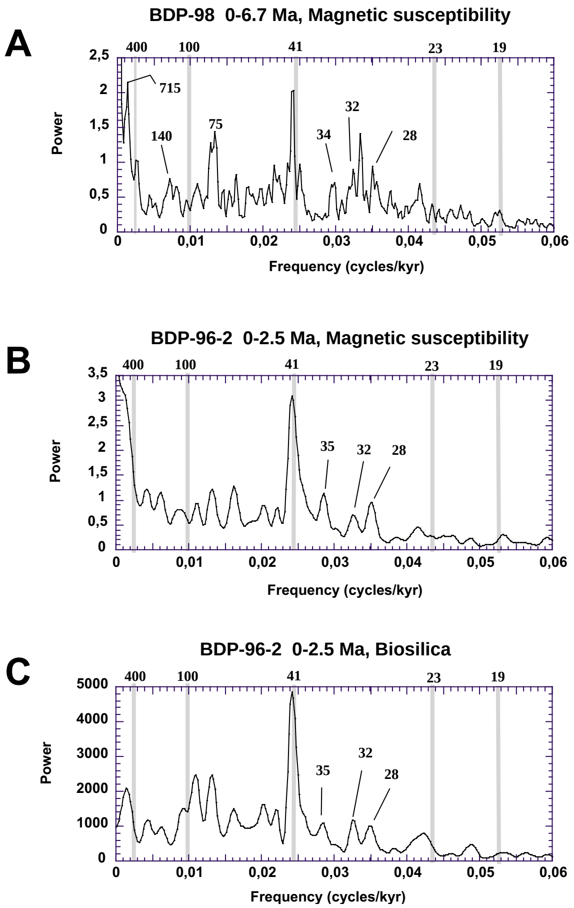

Fig. 9. Comparison of the power spectra of the BDP-98 magnetic susceptibility record for 6.7 Myr (A), the BDP-96-2 magnetic susceptibility record for 2.5 Myr (B), and the BDP-96-2 biogenic silica content for 2.5 Myr (C). Vertical gray lines correspond to the Earth's known orbital (Milankovitch) periodicities (in kyr).

and the measuring interval in only 3 cm, the lack of observed power in the precession band (~20 kyr) is not due to inadequate resolving power. Rather, it implies that the signal is simply not there, or has somehow been suppressed. Indeed, the 'missing' precession signal is a commonly encountered problem, which currently presents a serious difficulty for Milankovitch theory (Rail and Anaclecio, 2000).

The situation for the eccentricity band is more complex. Considerable variability in the duration of major glacial cycles is well documented, with published estimates of the dominant periodicities actually ranging from 80 to 130 kyr (Mix et al., 1995; Petit et al., 1999; Willis et al., 1999; Hinnov, 2000). There are several different proposals that attempt to explain such modifications in terms of either internal mechanisms (e.g. Mix et al., 1995) or external influences other than astronomical forcing (e.g. Rial, 1999). In particular, Rial (1999) points out that the period of the eccentricity signal in the ODP-806 record switches between ~80 and ~120 kyr approximately every 400 kyr. He suggests that the 100 kyr signal is frequency-modulated by the longer period 400 kyr component. The possibly relevant peaks we observe range from 75 to 143 kyr and may result from such confirmed variability in the nominally 100 kyr Milankovitch prediction.

As far as the longer periods are concerned, it should be noted that Kashiwaya et al. (2001) report a signal at ~400 kyr in their BDP-98 grain size measurements. They also find some evidence for peaks at ~600 kyr and ~1 Myr.

In the case of the 28–35 kyr power that appears consistently in all three records of Fig. 9, it should be noted that the recent analysis by Hinnov (2000) of Laskar's (1990) insolation calculations for the last 10 Myr yields obliquity peaks at 29 and 54 kyr in addition to a 'traditional' peak at 41 kyr. Furthermore, Rail and Anaclecio (2000) show that the time series of greenhouse gases and salt aerosols in the Vostok ice core yield distinct spectral peaks at ~29 and ~69 kyr as well as other, smaller, peaks surrounding the ~40 kyr obliquity signal, which they interpret as sidebands generated by frequency modulation of the obliquity signal itself.

Next, we examined the spectral content of the BDP-98 susceptibility record as a function of time by analyzing successive 2 Myr windows stepped every 1 Myr (Fig. 10) and successive 600 kyr windows stepped every 200 kyr (Fig. 11). In Fig. 10, the obliquity band is the dominant Milankovitch signal from the present back to 4 Ma, prior to which it is more muted and tends to be centered on ~45 kyr. As with the analysis of the entire 6.7 Myr record, we again see very little indication of a precession signal. Also, the possible eccentricity power presents a complicated picture, with peaks shifting between 75 and 143 kyr, somewhat

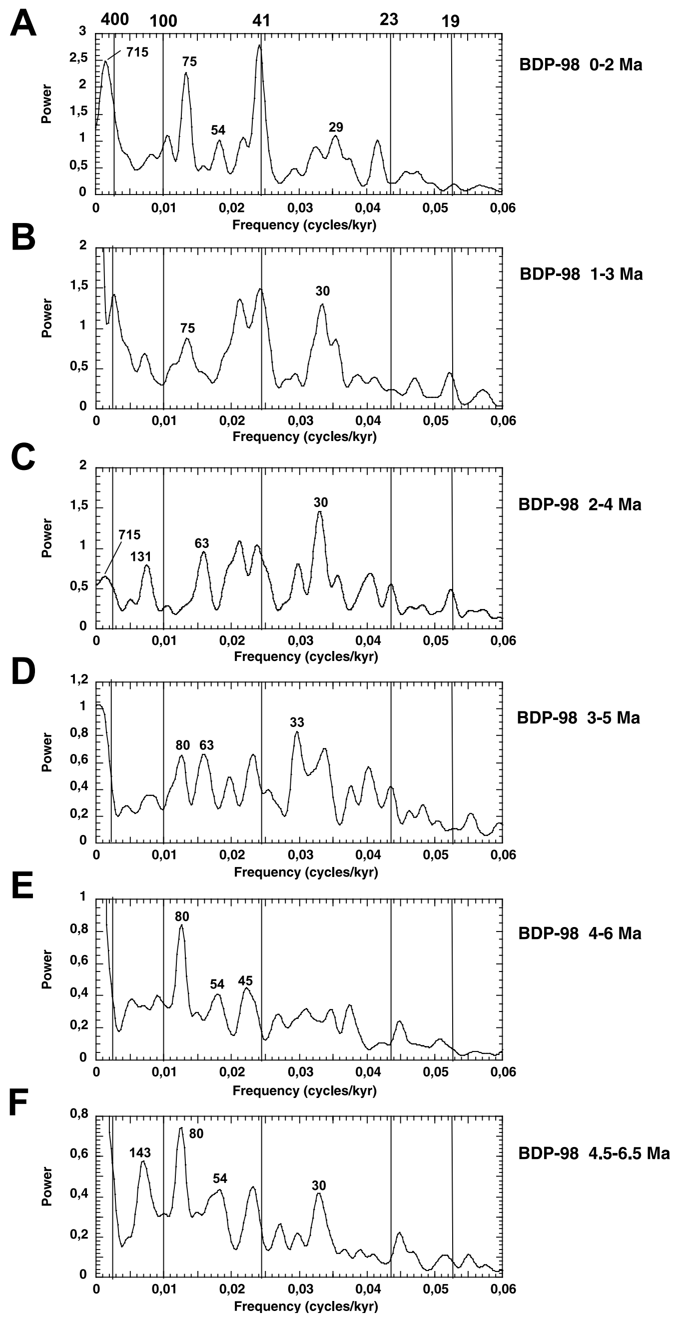

Fig. 10. Power spectra of the BDP-98 magnetic susceptibility record calculated for overlapping 2 Myr windows. Vertical gray lines correspond to the Earth's known orbital (Milankovitch) periodicities (in kyr).

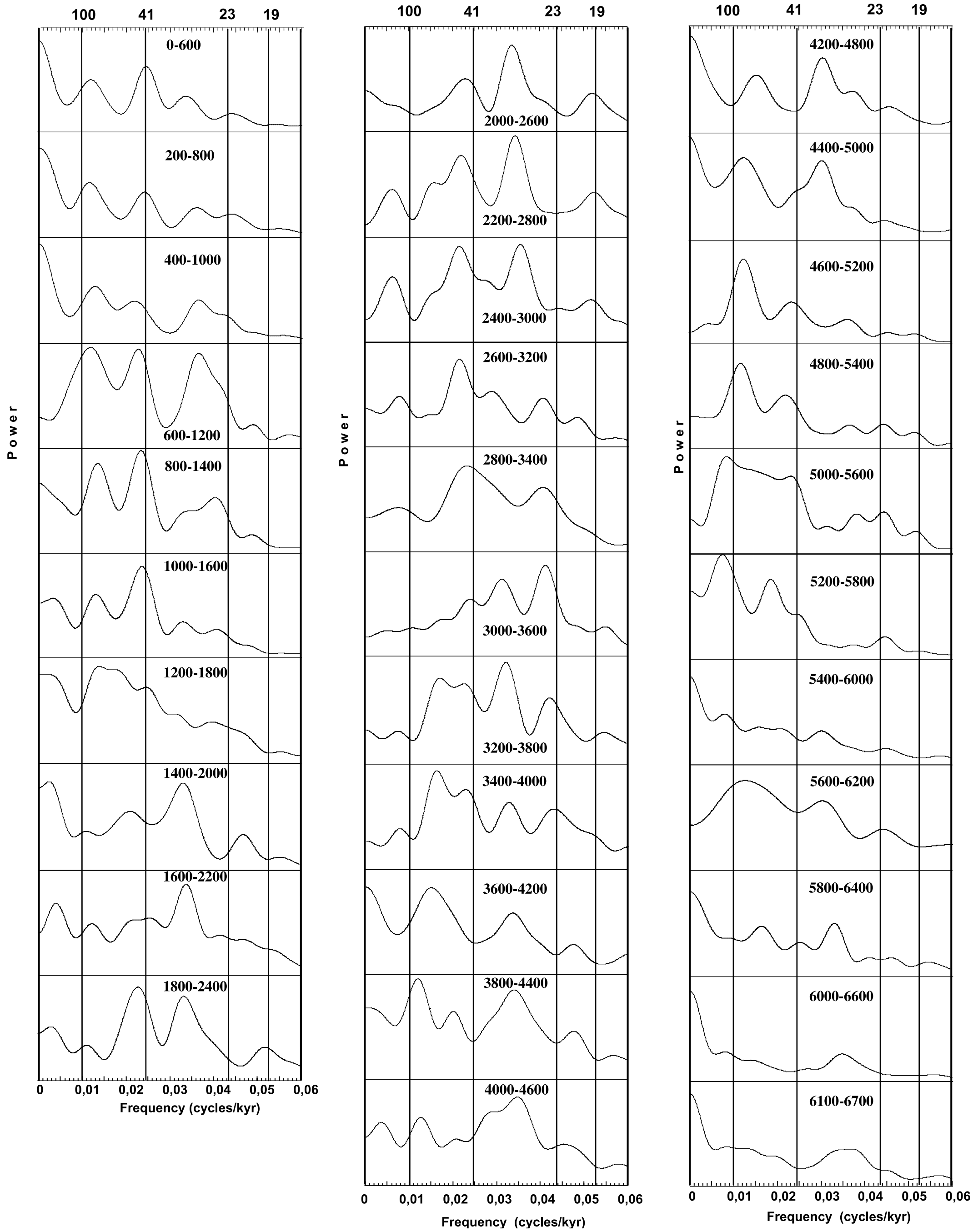

Fig. 11. Power spectra of the BDP-98 magnetic susceptibility record calculated for overlapping 600 kyr windows. Vertical lines correspond to the Earth's known orbital (Milankovitch) periodicities (in kyr). Numbers in each window correspond to time interval in ka.

broader than the range of Rial's (1999) frequency modulation hypothesis.

Hilgen et al. (1995) and Krijgsman et al. (1999) have suggested modifications to the Cande and Kent (1995) time scale. Back to 5.2 Myr, these are never larger than 18 kyr (approximately one precession cycle), but at the base of our profile (C3An.2n) they amount to some 110–140 kyr (approximately one eccentricity cycle). We therefore re-calculated the power spectra using the suggested modifications, with the particular goal of investigating the observed shift in our obliquity signal from 41 to ∼45 kyr in the lower part of the profile, as shown by Fig. 10. The spectral peaks remain essentially in the same positions, but the amplitude of the 45 kyr peak is actually somewhat increased.

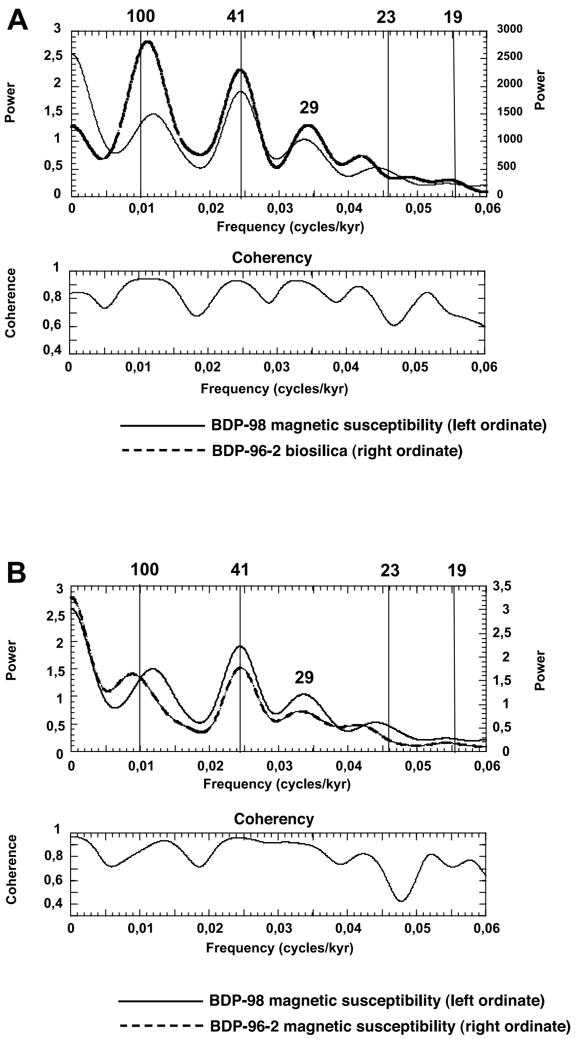

Fig. 12. Cross-spectral comparison of the BDP-98 magnetic susceptibility record with BDP-96-2 biogenic silica content (A), and with the BDP-96-2 magnetic susceptibility record (B) for the last 600 kyr. Vertical gray lines correspond to the Earth's known orbital (Milankovitch) periodicities (in kyr).

In Fig. 11, there is again a general lack of ∼20 kyr precession power, although a small peak does appear between about 2 and 3 Ma. The 100 kyr signal is relatively strong during the last million years, and between 4.6 and 5.6 Ma. Noticeable obliquity (∼40 kyr) power is present during much of the entire time, except for a short interval centered roughly on 4.5 Ma and for the earliest part of the record prior to about 5.5 Ma. For much of the entire 6.7 Myr, there is evidence of power at ∼30 kyr.

Finally, as an illustrative example, we carried out cross-spectral analyses between BDP-98 and BDP-96-2 for the interval from 0 to 600 ka. The magnetic susceptibility from BDP-98 is compared with the biogenic silica content of BDP-96-2 in Fig. 12A and with the magnetic susceptibility of BDP-96-2 in Fig. 12B. In both cases there is a strong coincidence of the obliquity signal at 41 kyr, but the eccentricity signal shifts between 83 and 110 kyr. We also note that the amplitude of the eccentricity signal for susceptibility – in both cores – is not dominant, whereas it does dominate in all the Lake Baikal biosilica sediment records for the last ∼900 kyr (Williams et al., 1997) and in all ODP oxygen isotope records (Shackleton et al., 1990; Rial, 1999; Hinnov, 2000). Once again, there is very little evidence of any precession power, but the persistent peak at 29 kyr presumably reflects the power between 28 and 35 kyr evident in all our other analyses.

4. Conclusions

Magnetic remanence results for 911 samples from the upper 270 m of Lake Baikal core BDP-98 yield a reliable geomagnetic polarity reversal chronology spanning the last 6.7 Myr and implying an average sediment accumulation rate of 3.9 cm/kyr. Spectral analysis of the corresponding magnetic susceptibility profile indicates that the most prominent Milankovitch signal is in the obliquity band centered on 41 kyr, with little or no power in the eccentricity and precession bands. A potentially important observation is the persistent appearance of significant power in a 'non-Milankovitch' band in the period range 28–35 kyr.

Remanence results from a further 561 samples spanning the depth interval 270–601 m are much more difficult to match to the standard polarity time scale, possibly because the sediments themselves are generally much coarser, having accumulated at a time when the drilling site was situated in a near-shore deltaic environment prior to tectonically controlled subsidence leading to deeper water and steady hemipelagic sedimentation. Our preferred (albeit tentative) interpretation of this lower part of the core implies a considerably higher – and more variable – sediment accumulation rate averaging about 7 cm/kyr.

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