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## Latitudinal control of astronomical forcing parameters on the high-resolution clay mineral distribution in the 45°-60° N range in the North Atlantic Ocean during the past 300,000 years

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**Abstract.** The clay mineralogy of four 5.5- to 13.5-m-long cores sampled between 45° and 60°N in the North Atlantic Ocean has been investigated at high latitudes within a well-constrained chronostratigraphic scale. Cross-correlation spectral analyses have been performed on both clay mineral and  $\delta^{18}\text{O}$  planktonic records. Detrital clay minerals display strong signals which are coherent with the  $\delta^{18}\text{O}$  record, within the three main Milankovitch frequency bands (eccentricity, obliquity, and precession). The climatic control on clay mineral sedimentation largely depends on the latitudinal location of the sediment cores. The 100,000-year signal occurs as a uniformly acting factor, whereas the 41,000-year signal dominates clay sedimentation at high latitudes and the 23,000-year signal dominates at midlatitudes. We suggest that the latitudinal variations of the orbital forcing on the detrital clay mineral distribution in the North Atlantic Ocean not only result from climatic control of the intensity of physical and chemical weathering, but also from latitudinal control on the detrital clay supply linked to influences of the high-latitude wind-driven and midlatitude ocean-driven transportation processes, respectively.

### Introduction

The North Atlantic Ocean is strongly influenced by long-term climatic variations between 40° and 65°N, where the largest temperatures changes in the world ocean develop due to the vicinity of large continental ice -sheets [Ruddiman and McIntyre, 1981a,b]. The circulation of large volumes of deep and intermediate water originating in the Norwegian and Greenland Seas is modified by glacial-interglacial cycles. The  $\delta^{13}\text{C}$  records indicate that the vertical and hydrological structures and circulations of the deep North Atlantic were different. The penetration depth of the ventilated North Atlantic Deep Water decreased, whereas poorly -ventilated water originating in the southern hemisphere invaded the deep oceanic basins [Duplessy *et al.*, 1988; Oppo and Fairbanks, 1990].

The North Atlantic Ocean is surrounded by large, high-latitude continental masses (Figure 1). The deep-sea clay sedimentation in this ocean is mainly controlled by detrital supply from the American and European continents [Biscaye, 1965; Griffin *et al.*, 1968; Rataev *et al.*, 1969; Lisitzin, 1972]. The variations of the terrigenous input during the Quaternary depended on the size of

continental ice -sheets and on sea -level changes, both of which modified coastline and continental erosion areas. During glacial periods, the extensive high -latitude continental areas surrounding the North Atlantic Ocean allowed the growth of large continental ice -sheets which strongly affected the detrital clay input [Chamley, 1989]. Previous works have shown the relationships existing between the variations of clay composition and the long-term climatic evolution in Atlantic Ocean sediments [Chamley, 1979; Robert and Maillot, 1983; Thiébaud *et al.*, 1989]. More recent works have demonstrated that the Quaternary terrigenous clay input was controlled by the size of the continental ice -sheets and by the intensity of the physical weathering of the surrounded continental masses in both the South Atlantic Ocean [Petschick *et al.*, 1996] and the Norwegian Sea [Froget *et al.*, 1989].

Previous studies performed in the Arabian Sea and the Indian Ocean have proved that clay sedimentation was orbitally forced during the last 1.5 m.y. [Fagel *et al.*, 1992; Fagel, 1994]. The present study focuses on high -resolution clay sedimentation in the North Atlantic Ocean and its dependence on short-time climatic evolution during the last 300,000 years. We used the spectral analysis techniques on high -resolution clay records (time spacing of 200 to 2000 years), in the frame of a well-constrained chronostratigraphic scale defined by accelerator mass spectrometry  $^{14}\text{C}$  dating for the last 35 kyr [Cortijo *et al.*, 1997; Vidal *et al.*, 1997] and by correlation with the SPECMAP stacked  $\delta^{18}\text{O}$  signal for older periods [Martinson *et al.*, 1987].

### Materials and Methods

#### Materials

Four sediment cores from the 1990 Paleocinat I cruise were studied (Table 1, Figure 1). Two cores were sampled in the

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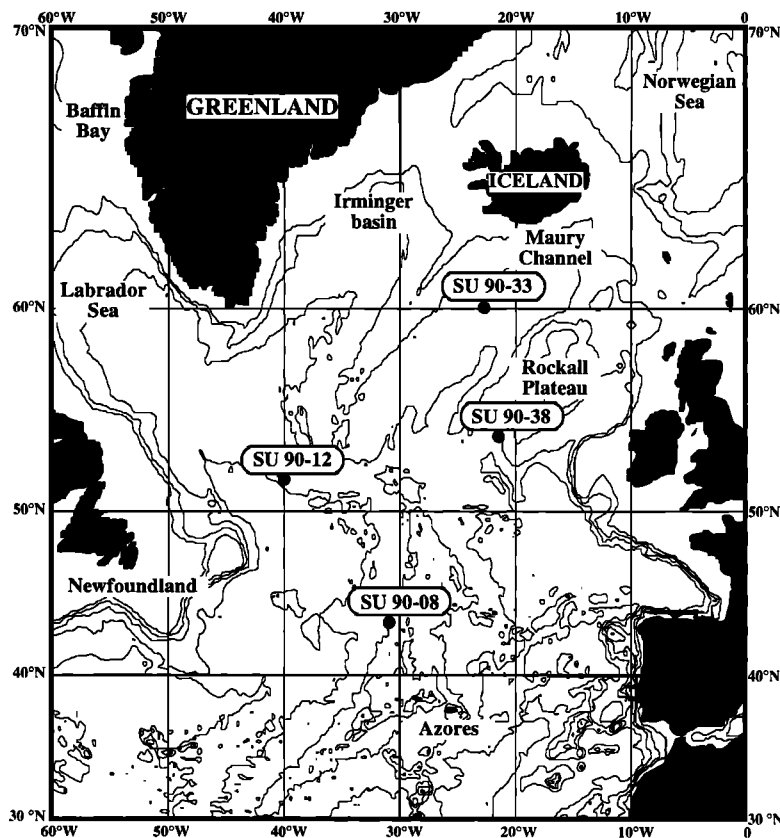


Figure 1. Study area and core locations.

eastern Atlantic basin : the southernmost core SU 90-08, situated near the Azores on the western flank of the mid-oceanic ridge, and the core SU 90-12, situated near the mouth of the Labrador Sea, on a seamount at 1000 m above the seafloor. The two other cores are located in the western basin : the core SU 90-38, situated near the Rockall Plateau, and the Northernmost core SU 90-33, located on the southern margin of Iceland (Figure 1).

### Chronostratigraphy

For each site the timescale is based on the comparison between the benthic and/or planktonic  $\delta^{18}\text{O}$  records and the spectral mapping SPECMAP stack [Martinson *et al.*, 1987] using the Analyseries software [Paillard *et al.*, 1996]. The age-depth relations are given in Table 2 and drawn in Figure 2. A linear interpolation was applied between the stratigraphic levels

identified, assuming that the sedimentation rate was constant between these levels. The  $\delta^{18}\text{O}$  isotopic analysis of foraminifera was measured on an automatic carbonate preparation line coupled to a Finnigan MAT 251 mass spectrometer (Centre des Faibles Radioactivités, Gif-sur-Yvette, France). Isotopic stages and stage boundaries were deduced from isotopic curves, magnetic susceptibility, and grey-level reflectance [Grousset *et al.*, 1993; Cortijo *et al.*, 1995]. Isotopic stages 7, 5, 3 and 1, are considered as interglacial periods, and isotopic stages 8, 6, 4, and 2 are considered as glacial periods.

Cores SU 90-08 and SU 90-12 were sampled every 2-10 cm, corresponding to a sampling interval ranging from 0.2 to 2 kyr for SU 90-08 and from 0.5 to 2.5 kyr for SU 90-12. Samples were taken every 5 cm along the core SU 90-33, corresponding to a sampling interval of about 1 kyr. The core SU 90-38 was sampled every 5-10 cm, pointing to a sampling interval of 1-2 kyr (Table 1).

Table 1. Core Characteristics

Core	Latitude	Longitude	Depth, m	Length, m	Age of Bottom Sediments, kyr	Sample Number	Sample Spacing by Age, kyr	Sediment Rate, cm/kyr <sup>-1</sup>
SU 90-08	43°31'2 N	30°24'5 W	3080	12.27	285.5	246	0.2-2.0	4.2
SU 90-12	51°52'6 N	39°47'4 W	2950	5.47	188	76	0.5-2.5	2.9
SU 90-38	54°05'4 N	21°04'9 W	2900	11.42	228	147	1-2	5.0
SU 90-33	60°34'4 N	22°05'1 W	2400	13.53	272	268	1	5.1

**Table 2.** Age-Depth Relation for the Studied Cores

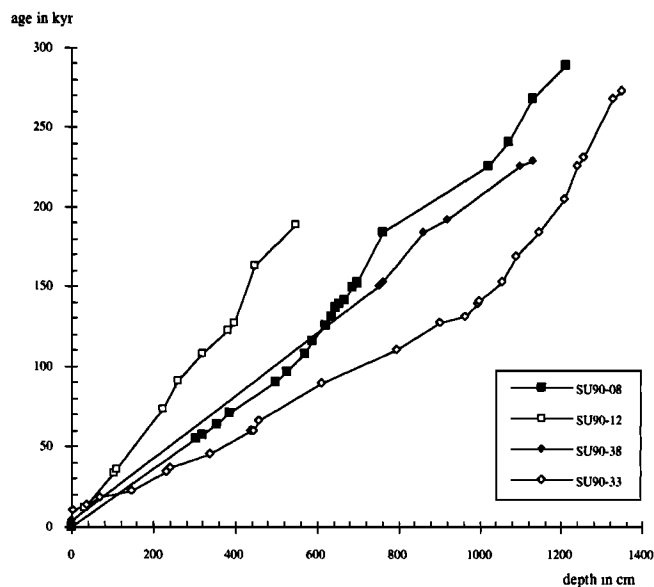
SU 90-08		SU 90-12		SU 90-33		SU 90-38	
Depth, cm	Age, ky	Depth, cm	Age, ky	Depth, cm	Age, ky	Depth, cm	Age, ky
0	0.0	29.5	11.4	3	10.4	0	4.1
302	54.8	104	33.2	36	13.5	750	150.4
320	57.6	109	35.6	68	18.6	760	152.3
352	64.1	220.5	73.2	145	22.5	860	183.4
383	71.1	258.5	90.9	231	34.1	920	191.4
497	90.1	319.5	107.5	240	36.2	1100	225.2
526	96.4	380	122.2	336	45.0	1130	228.3
569.5	107.6	398.5	126.6	439	59.5		
589.5	115.9	448	162.8	444	60.0		
620	125.0	547.5	188.3	455	66.6		
633.5	131.1			609	89.4		
644.5	136.6			795	110.3		
653	139.0			900	127.0		
667	141.3			963	131.1		
685	149.3			994	139.2		
698.5	152.1			999	140.3		
699	152.6			1054.5	152.3		
759	183.4			1090.5	168.5		
1019	225.2			1144.5	183.4		
1071	240.2			1209	204.3		
1130	267.5			1240.5	225.2		
1210	288.5			1254	230.9		
				1326	267.5		
				1348.5	272.7		

## Clay Minerals

About 840 samples were submitted to X-ray diffraction analysis (XRD) on a Philips PW1710 X-ray diffractometer from  $2.49^\circ$  to  $32.5^\circ 2\theta$  using a copper  $\alpha$  radiation. The method was conducted as follows. All samples were first decalcified with a 0.2 N chlorhydric acid. The excess acid was removed by repeated centrifuging. The clay-size fraction less than  $2 \mu\text{m}$  was isolated by settling and then oriented on glass slides. Three tests were performed on the oriented mounts: (1) natural sample, (2)

glycolated sample, and (3) sample heated at  $490^\circ\text{C}$  for 2 hours [Holtzapffel, 1985]. The clay minerals were identified by their basal reflections [Brindley and Brown, 1980].

The peak height, which is the parameter used for the spectral analysis, is obtained by measuring directly on XRD diagrams the height of one characteristic peak for each relevant clay mineral, on the glycolated sample after subtraction of the background noise ( $12\text{\AA}$  for 10-14v mixed-layer,  $10\text{\AA}$  for illite,  $3.57\text{\AA}$  for kaolinite, and  $3.53\text{\AA}$  for chlorite). The reproducibility of the technical work and measurements was tested: five oriented amounts, prepared from the same sample, were submitted three times to XRD. The relative margin of error is  $\pm 10\%$ .

**Figure 2.** Age-depth relation for the studied cores.

## General Data and Clay Mineral Results

The foraminiferal  $^{18}\text{O}/^{16}\text{O}$  isotopic ratios (expressed as  $\delta^{18}\text{O}$  versus Pee Dee belemnite (PDB)) have been measured on an automated carbonate preparation line coupled to a Finnigan MAT 251 mass spectrometer. The mean external reproducibility of powdered carbonate standards is  $\pm 0.05\%$ . Data are reported using as reference NBS 19  $\delta^{18}\text{O}$  of  $-2.20\%$  versus PDB [Coplen, 1988]. Both planktonic foraminiferal species *Neogloboquadrina pachyderma* (left coiling), shell sizes  $200\text{--}250 \mu\text{m}$ , and *Globorotalia bulloides* (shell sizes  $250\text{--}315 \mu\text{m}$ ), were analyzed, depending on the latitude of the core. The  $\delta^{18}\text{O}$  planktonic records were plotted on the same timescale as clay mineralogy in regard with illite, chlorite, and kaolinite peak heights (Figures 3-6).

### Core SU 90-08

The clay mineral fraction is mainly composed (Table 3) of illite (37%) and smectite (33%). Chlorite and kaolinite are less

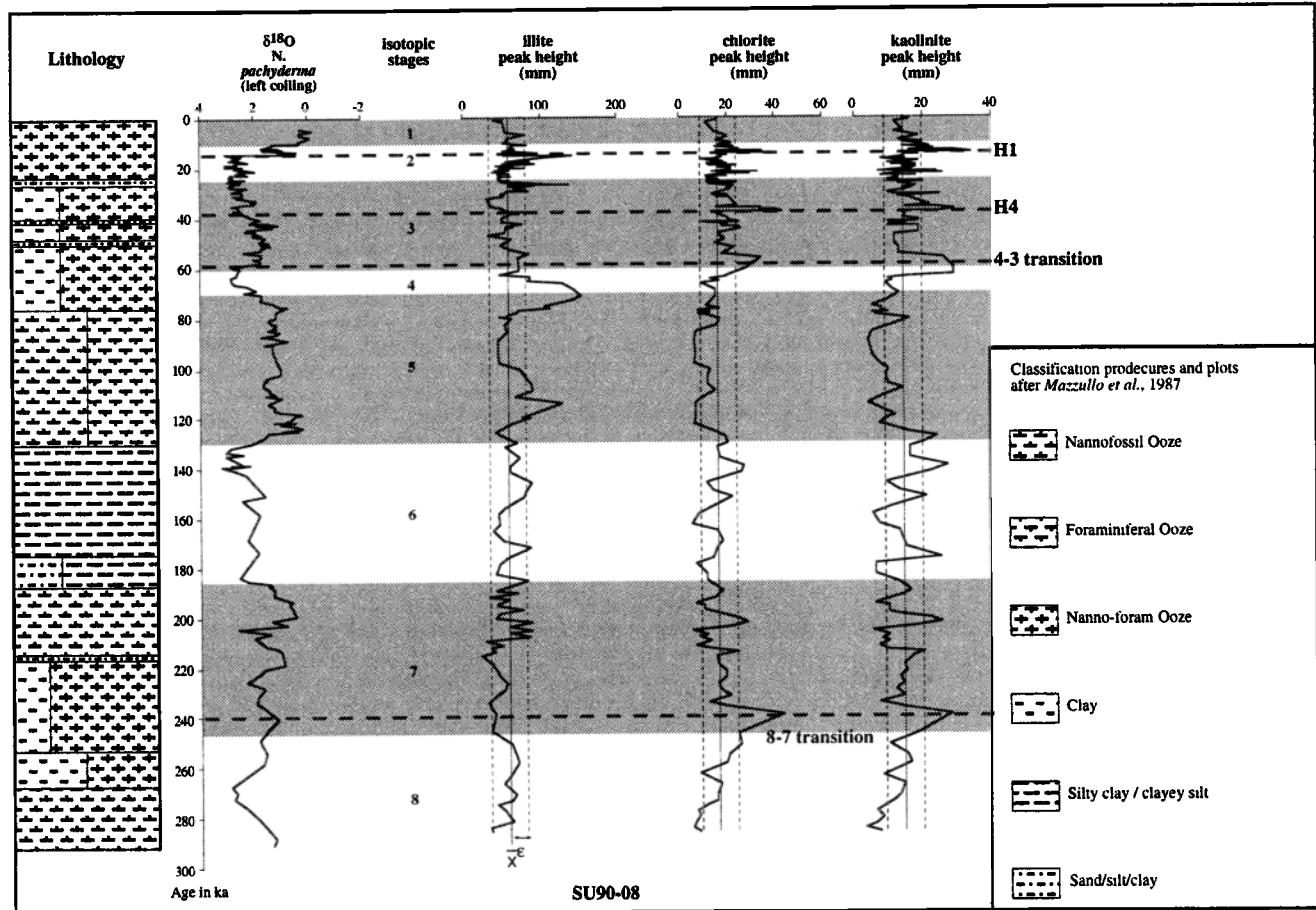


Figure 3. Core SU 90-08. The  $\delta^{18}\text{O}$  of *Neogloboquadrina pachyderma* (left coiling) and peak heights (in millimeters) of illite, chlorite and kaolinite (interpolated data) plotted versus time and isotopic stages. Letter X denotes average (solid line) and letter  $\epsilon$  denotes standard deviation (dashed line). Interglacial and glacial periods are outlined by shaded areas. H1 and H4 are Heinrich events 1 and 4 [Heinrich, 1988].

abundant, with 14% and 11% of the clay fraction, respectively. Illite-smectite random mixed-layer minerals (10-14s) constitute 5% of clay minerals [Bout-Roumazeilles, 1995].

The illite peak height values vary slightly from one sample to another. They remain inside the confidence interval except at the beginning of stage 5, at the 5-4 transition, at the 3-2 transition, and during stage 2 where the highest values are registered (Figure 3). The latter interval refers to specific sediments located between  $40^\circ$  and  $55^\circ\text{N}$  in the North Atlantic Ocean which contain a series of layers rich in detrital ice-rafted material, unusually poor in foraminifera and called "Heinrich layers" [Heinrich, 1988].

Isotopic stages 8 and 5 exhibit the lowest peak heights of chlorite and to a lesser extent of kaolinite, whereas stages 7 and 6 are characterized by higher values for these mineral species. A few discrete high values are situated at the 8-7 transition, at the 4-3 transition and within stages 3-2 (Figure 3). Two of the four more recent peaks correspond to Heinrich events H1 and H4.

#### Core SU 90-12

The clay mineral fraction is dominated by average values of 34% illite (Table 3) and 21% chlorite. The illite-vermiculite mixed-layer minerals represent 18% of the clay mineral fraction.

Smectite (15%) and kaolinite (12%) are less abundant [Bout-Roumazeilles, 1995].

The variations of illite, chlorite, and to a lesser extent of kaolinite, are very similar. The illite, chlorite, and kaolinite peak height values fluctuate inside the confidence interval from one sample to another, except at the 4-3 and 6-5 stage transitions, which are marked by illite values higher than the standard deviation (Figure 4).

The illite-vermiculite mixed-layer (10-14v) peak height values vary slightly, with a higher range in the upper part of the core during stages 3 and 2, in relation with Heinrich deposits (H2, H4, and H5). Two low values occur within stages 6 and 5 and at the 6-5 stage transition (Figure 4).

#### Core SU 90-38

The clay mineral fraction is mainly composed of illite (39%) and smectite (35%). chlorite (14%) and kaolinite (12%) are less abundant, whereas the illite-smectite random mixed -layers occur as trace amounts (Table 3).

Once again the illite, chlorite, and kaolinite peak height variations are very similar and roughly parallel the successive isotopic stages (Figure 5). The values display very short-time

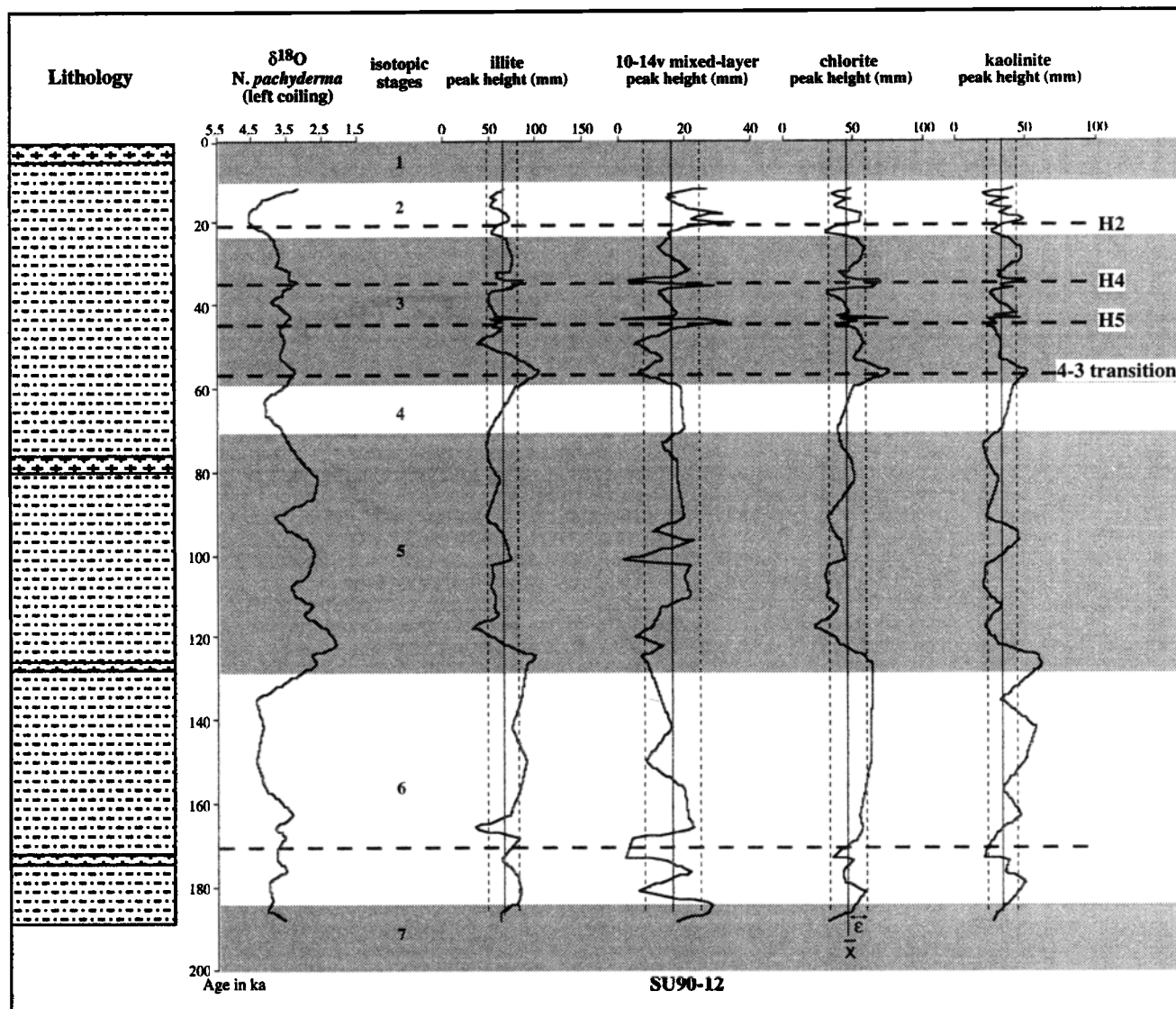


Figure 4. Core SU 90-12. The  $\delta^{18}\text{O}$  of *Neogloboquadrina pachyderma* (left coiling) and peak heights (in millimeters) of illite, illite-vermiculite (10-14v) random mixed -layers, and chlorite and kaolinite (interpolated data) plotted versus time and isotopic stages. Letter X denotes average (solid line) and letter  $\epsilon$  denotes standard deviation (dashed line). Interglacial and glacial periods are outlined by shaded areas. H2, H4, and H5 are Heinrich events 2, 4 and 5 [Heinrich, 1988].

modifications less apparent than in the two previous cores, with distinct periods of high and low values. Glacial stages 6, 4, and 2 are characterized by the highest values, whereas interglacial stages 7, 5, and 1 show the lowest values. Local variations occur, especially expressed by a strong decrease of all clay mineral peak heights during stage 5.5 (a warmer episode inserted in an interglacial period) and by a characteristic increase of all peak heights during the cold stage 6.2.

#### Core SU 90-33

Smectite constitutes as an average 59% of the clay mineral fraction. The other minerals are less abundant : Illite represents 20% of the clay mineral fraction, whereas the percentages of

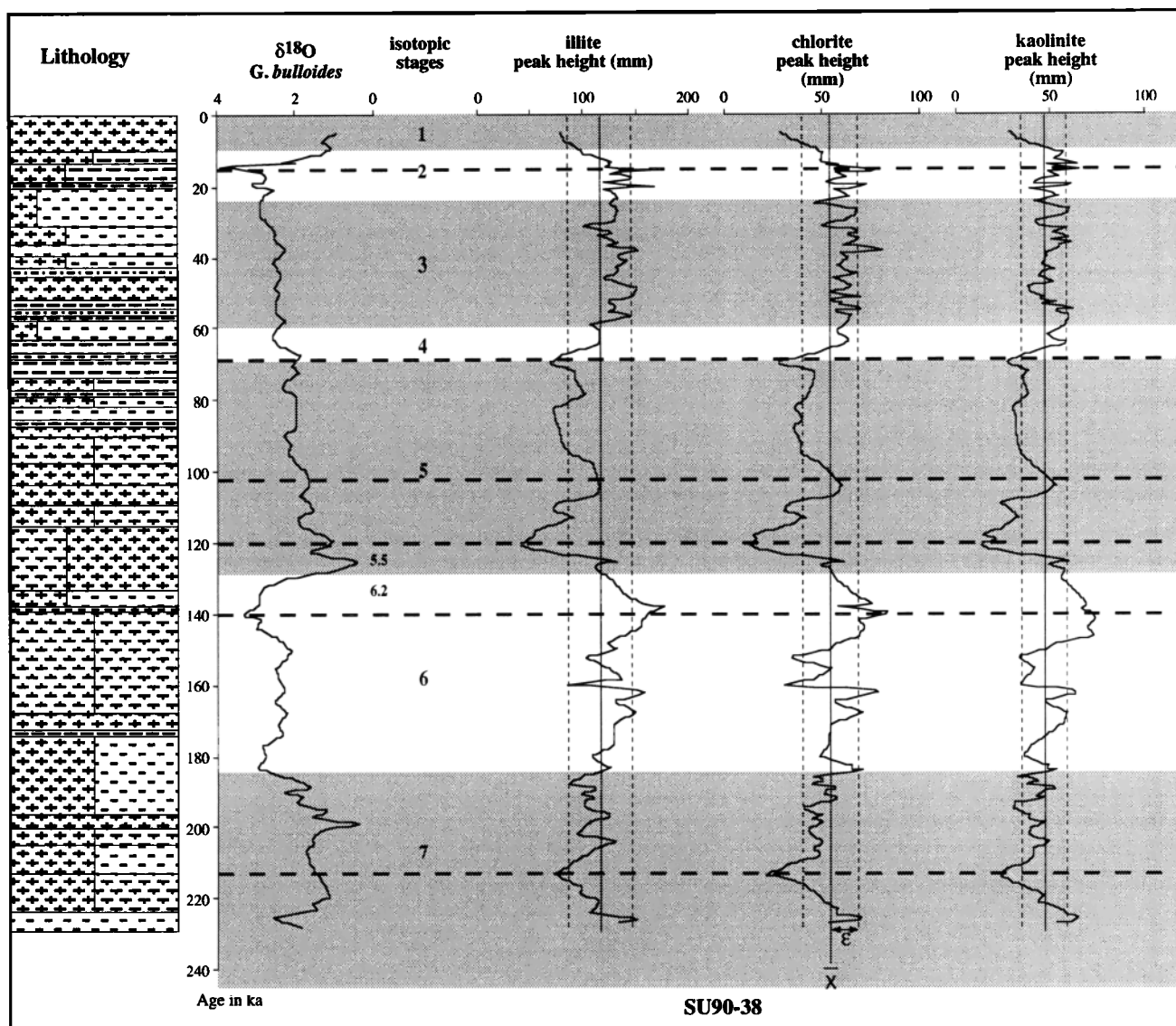
chlorite, of kaolinite, and of illite-smectite random mixed -layers represent 10% or less (Table 3).

The illite peak height variations parallel those of chlorite and kaolinite (Figure 6). The beginning of isotopic stages 7 and 3 corresponds to the highest positive deviations from the mean value. Except for a level situated within stage 6, the whole core is characterized by a low range of peak height variations, the values being situated very close to the average value.

### Spectral Analysis Methods and Results

#### Methods

The data have been processed by spectral analysis, in order to examine the response of the clay sedimentology to climate in the



**Figure 5.** Core SU 90-38. The  $\delta^{18}\text{O}$  of *Globigerina bulloides* and peak heights (in millimeters) of illite, chlorite, and kaolinite (interpolated data) plotted versus time and isotopic stages. Letter X denotes average (solid line) and letter  $\epsilon$  denotes standard deviation (dashed line). Interglacial periods are outlined by shaded areas.

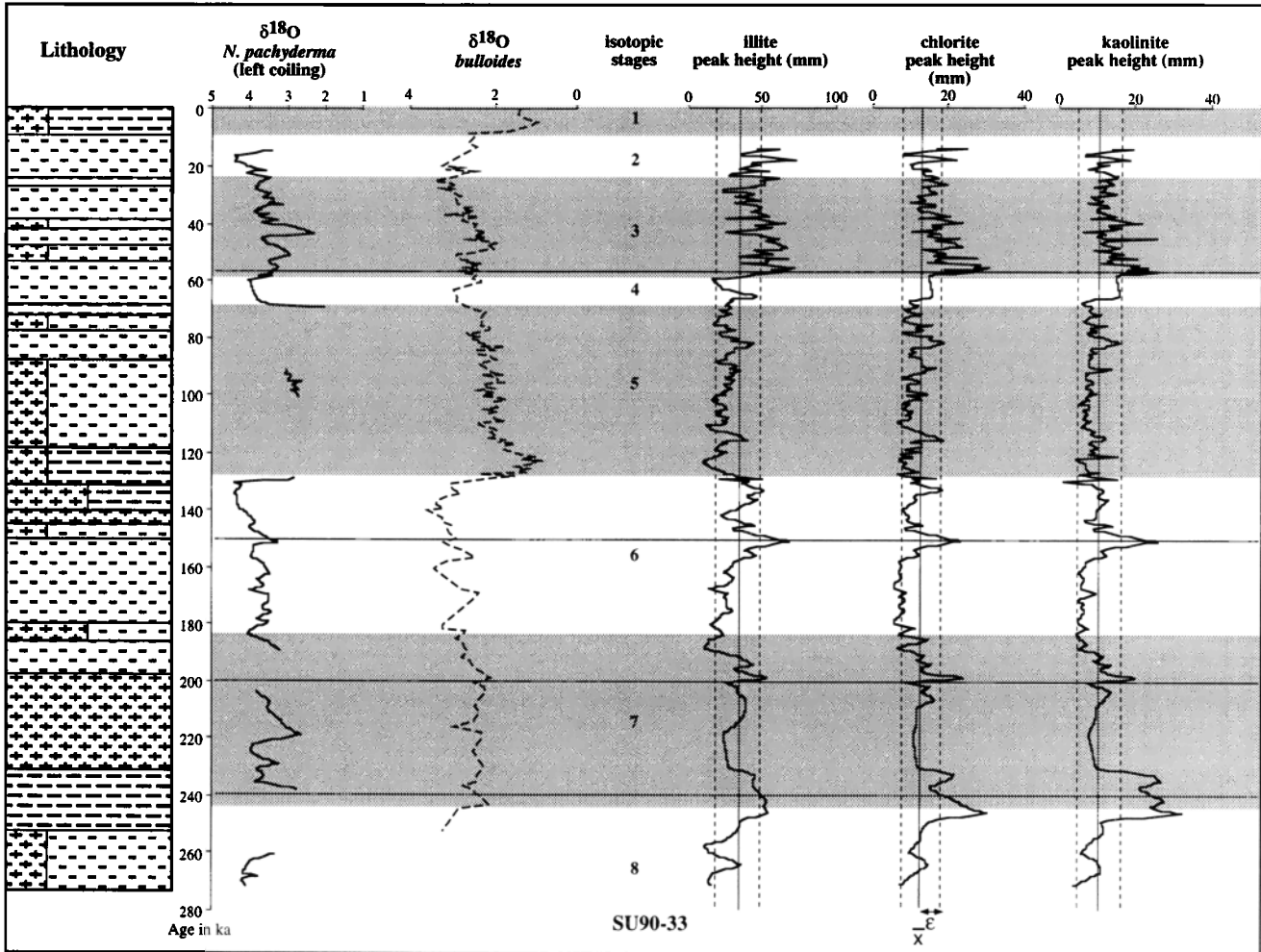
main frequency bands of the orbital forcing, about  $1/23 \text{ kyr}^{-1}$  (precession),  $1/54 \text{ kyr}^{-1}$  and  $1/41 \text{ kyr}^{-1}$  (obliquity), and of ice-volume variations (about  $1/100 \text{ kyr}^{-1}$ ). The spectral analyses were performed on the following terrigenous clay minerals: chlorite, illite, kaolinite, and illite-vermiculite mixed-layers. The data on smectite were not processed by spectral analysis because smectite could be either continent-derived or volcanic-islands-derived.

The  $1/23 \text{ kyr}^{-1}$  frequency represents the precession forcing for which the insolation is modulated by the relative position of the perihelion and of the summer solstice. The  $1/41 \text{ kyr}^{-1}$  frequency is the obliquity forcing band. It results from the distribution of solar energy between high and low latitudes. In Quaternary sediments, the  $1/100 \text{ kyr}^{-1}$  frequency is now considered as an internal oscillation of the coupled ice-atmosphere-ocean system linked to the long-time constant of the ice sheet response to the orbital

forcing [Imbrie *et al.*, 1992, 1993]. Notice that the temporal windows considered in the present work (176.5-281 kyr) are too short for fully reliable results in that frequency range and will be discussed only for general trends.

Spectral analysis was done by the *Blackman and Tuckey*, [1958] method (hereinafter referred as Blackman-Tuckey) with a Bartlett-type window providing a 80% confidence interval [Jenkins and Watts, 1968; Bringham, 1974]. The Analyseries package [Paillard *et al.*, 1996] was used. It includes the ARAN software for Blackman-Tuckey spectral analysis developed at Brown University for the SPECMAP program and communicated by W. Prell.

Owing to the fact that sedimentation rates are not constant, the time spacing between the samples is not regular. In order to obtain a regular time sampling, all data were linearly interpolated



**Figure 6.** Core SU 90-33. The  $\delta^{18}\text{O}$  of *Neogloboquadrina pachyderma* (left coiling), of *Globigerinabulloides* (dashed line) for interglacial periods, and peak heights (in millimeters) of illite, chlorite, and kaolinite (interpolated data) plotted versus time and isotopic stages. Letter X denotes average (solid line) and letter  $\epsilon$  denotes standard deviation (dashed line). Interglacial periods are outlined by shaded areas.

at a 0.5-kyr interval. The spectral analysis was performed on the largest period where both clay minerals and  $\delta^{18}\text{O}$  have been analyzed : between 4.5 and 285.5 ka for SU90-08, between 11.5 and 188 ka for SU 90-12, between 5 and 228 ka for SU 90-38, and between 14.5 and 272 ka for SU 90-33.

The studied frequencies range from 0 to 0.2 kyr<sup>-1</sup> with a step of 0.003 kyr<sup>-1</sup>. The bandwidth is 0.01 kyr<sup>-1</sup> for SU 90-08, 0.015 kyr<sup>-1</sup> for SU 90-12, 0.013 kyr<sup>-1</sup> for SU 90-38, and 0.011 kyr<sup>-1</sup> for SU 90-33. Results of the spectral analysis are presented as variance powers. The variance powers of both clay peak heights and

**Table 3.** Mean Mineralogical Composition of Studied Cores Based on 246 Samples for SU 90-08, on 76 Samples for SU 90-12, on 268 Samples for SU 90-33 and on 147 Samples for SU 90-38

Core	Chlorite		Illite		10-14v		10-14s		Smectite		Kaolinite	
	Average	s.d.	Average	s.d.	Average	s.d.	Average	s.d.	Average	s.d.	Average	s.d.
SU 90-08	14%	±3	37%	±7	0%	-	5%	±2	33%	±11	11%	±3
SU 90-12	21%	±4	34%	±4	18%	±8	0%	-	15%	±8	12%	±2
SU 90-38	14%	±3	39%	±7	0%	-	0%	-	35%	±11	12%	±2
SU 90-33	10%	±4	20%	±5	0%	-	4%	±2	59%	±11	7%	±2

Here 10-14v is illite-vermiculite mixed layer, and 10-14s is illite-smectite mixed layer; s.d. is standard deviation.



foraminiferal isotope data were normalized by the integrated variance which corresponds to the sum of all the variance powers between 0 and  $0.2 \text{ kyr}^{-1}$ . Each frequency band is reported taking as center the observed maximum variance for specific band, not the orbital frequency. Differences between these two values are within the errors and are attributed to uncertainties in the timescale and to analytical noise. The mean Blackman-Tuckey coherency and phase were calculated by cross-spectral analysis between the clay mineral peak heights and the  $\delta^{18}\text{O}$  signal in the three major frequency bands :  $1/100 \text{ kyr}^{-1}$ ,  $1/41 \text{ kyr}^{-1}$ , and  $1/23 \text{ kyr}^{-1}$  [Imbrie *et al.*, 1992; Imbrie *et al.*, 1993; Labeyrie *et al.*, 1996]. Coherency is considered significant (at the 80% level) if larger than 0.52. Phases are reported and discussed only for the more significant bands (highest variance and significant coherency), therefore mostly within the  $1/100 \text{ kyr}^{-1}$  band, considered as the "ice sheet climatic forcing" band. The phase origin is taken at the minimum ice volume (minimum foraminiferal  $\delta^{18}\text{O}$ ) and reported as leads (negative phases) of the clay minerals relative abundance when compared to the ice-sheet volume changes. We calculated also the phase relative to the insolation change for discussion of the large precessional response of the clay minerals in the northwestern core SU 90-12. Following Imbrie *et al.* [1992], calculations were done for June 15 insolation at  $65^\circ\text{N}$  (origin for June 15 at perihelion) using the Berger [1978] algorithm as available in the Analyseries package. Phases in radians may be converted to degrees or kiloyears (one period =  $360^\circ = 2\pi \text{ rad}$ ) for each of the frequency bands.

## Results

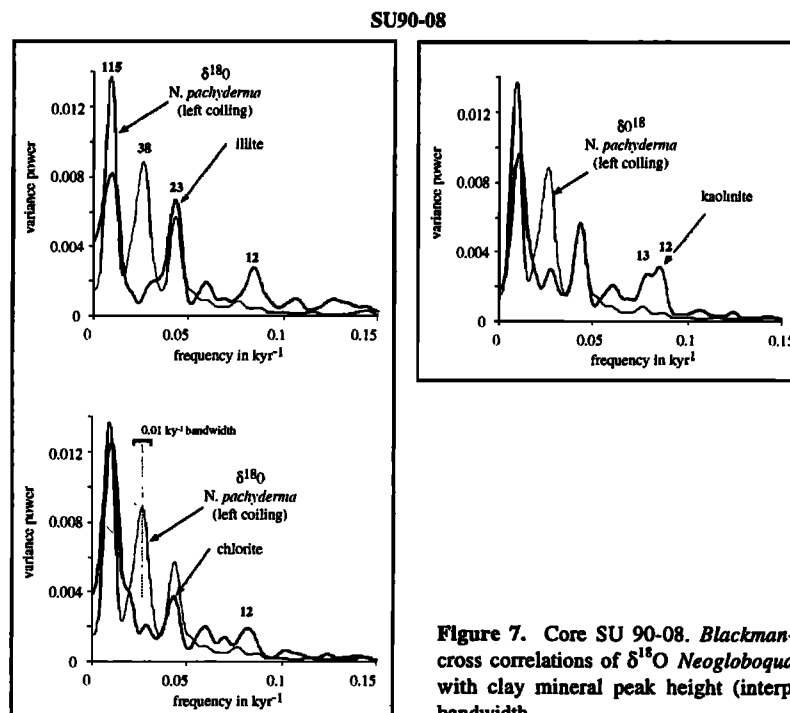
**Core SU 90-08.** The core located at  $44^\circ\text{N}$  on the western side of the Mid-Atlantic Ridge reveals a  $\delta^{18}\text{O}$  record of planktonic foraminifera marked by three major frequencies (Figure 7), with variance power maxima around  $1/115 \text{ kyr}^{-1}$  (the  $1/100 \text{ kyr}$  ice -

sheet forcing),  $1/38 \text{ kyr}^{-1}$  (the  $1/41 \text{ kyr}$  obliquity), and  $1/23 \text{ kyr}^{-1}$  (precession) (Table 4).

The illite spectrum shows two main signals around  $1/115 \text{ kyr}^{-1}$  with 25% of the integrated variance and around  $1/23 \text{ kyr}^{-1}$  with 18% of the integrated variance (Figure 7, Table 4). The coherencies between illite and  $\delta^{18}\text{O}$  records are slightly higher ( $0.57 < \text{Cov} < 0.62$ ) than the confidence level for the two main frequency bands (0.52 for a 80% confidence interval) [Jenkins and Watts, 1968]. An additional frequency is observed around  $1/12 \text{ kyr}^{-1}$  on the illite record but also on the chlorite and kaolinite records. It represents less than 7% of the total variance but has already been observed in previous studies on rock-magnetic properties [Robinson *et al.*, 1995] as well as from grain-size parameters on core SU 90-33 [Revel *et al.*, 1996]. The latter authors have attributed this frequency to either ice-raftered events [Robinson *et al.*, 1995] or changes in the intensity of the bottom circulation.

The chlorite and the kaolinite records are mainly controlled by the  $1/115 \text{ kyr}^{-1}$  and  $1/23 \text{ kyr}^{-1}$  frequencies (Table 4, Figure 7). The coherency between the chlorite and the  $\delta^{18}\text{O}$  records ranges from 0.56 to 0.65 and is therefore slightly superior to the confidence level of 0.52, whereas the consistency between the kaolinite and  $\delta^{18}\text{O}$  records is significant in the two main frequency bands ( $\text{Cov} = 0.65$ ).

Changes in all three minerals lead the ice volume changes ( $\delta^{18}\text{O}$ ) quite considerably (about 2-2.5 rad, i.e. about 30 kyr) in the  $1/100 \text{ kyr}^{-1}$  band. Errors are large. Thus the similarity in the early response of illite, chlorite, and kaolinite to the ice-sheet evolution, when compared to the marine paleoclimate proxies (which respond approximately in phase with ice volume [Imbrie *et al.*, 1992, 1993]), does indicate that the accumulation of these minerals in the core is not controlled directly by the ice volume, but by the growth or decay of the ice-sheets : The clay mineral relative abundance is larger during periods of ice-sheet decay and smaller during periods of ice-sheet growth.



**Figure 7.** Core SU 90-08. Blackman-Tuckey [1958] spectral analysis cross correlations of  $\delta^{18}\text{O}$  *Neogloboquadrina pachyderma* (left coiling) with clay mineral peak height (interpolated data) with a  $0.01 \text{ kyr}^{-1}$  bandwidth.

**Table 4.** Blackman-Tuckey Spectral Analysis Cross Correlations of  $\delta^{18}\text{O}$  of *Neogloboquadrina pachyderma* (Left Coiling) With Clay Minerals for Core SU 90-08

1/ Frequency, kyr			Variance Power			Covariance,
$\delta^{18}\text{O}$ <i>pachyderma</i>	Clay Mineral*		$\delta^{18}\text{O}$ <i>pachyderma</i>	Clay Mineral*		Confidence Level = 0.52
115		104	32		25	0.57
38	illite	-	27	illite	-	-
23		23	15		18	0.62
115		101	32		36	0.56
38	chlorite	-	27	chlorite	-	-
23		23	15		10	0.65
115		98	32		26	0.65
38	kaolinite	37	27	kaolinite	-	-
23		23	15		15	0.65

\*Peak height is in millimeters.

To summarize, the widely consistent response in the  $1/100 \text{ kyr}^{-1}$  frequency band ( $0.57 < \text{Cov} < 0.62$ ) of the chlorite, illite, and to a lesser extent kaolinite results from the variations of the northern continental ice -volume which dominated all climatic records in the North Atlantic Ocean during the last 500 kyr [Ruddiman and McIntyre, 1984; Imbrie et al., 1992].

The illite, kaolinite, and chlorite records also express in a significant way the precession forcing in the  $1/23 \text{ kyr}^{-1}$  frequency band, which is consistent with the  $\delta^{18}\text{O}$  planktonic record ( $\text{Cov} = 0.65$ ). It reveals the dominant precessional forcing at midlatitudes ( $44^\circ\text{N}$ ) on the detrital clay sedimentation during the last 300 kyr. The phase response in the precession band corresponds for the three clay minerals (illite, chlorite, and kaolinite) to a lead of about  $-1.2 \text{ rad}$  (3 kyr) when compared to ice volume, or a lag of  $0.4 \text{ rad}$  (1 kyr) when compared to insolation; the clay minerals follow therefore the "early response group" of Imbrie et al. [1992, 1993], in a way similar to some of the parameters linked with surface-deep ocean exchanges and southern hemisphere sea surface temperature.

**Core SU 90-12.** The core situated at  $52^\circ\text{N}$  near the entrance of the Labrador Sea displays a  $\delta^{18}\text{O}$  record characterized by three main frequencies, with power variance maxima around  $1/128 \text{ kyr}^{-1}$  (46% of integrated variance),  $1/42 \text{ kyr}^{-1}$  (26%), and  $1/23 \text{ kyr}^{-1}$  (12%). The temporal windows (176.5 kyr) are too short for reliable results in the  $1/100 \text{ kyr}^{-1}$  frequency range and will be discussed only for general trends.

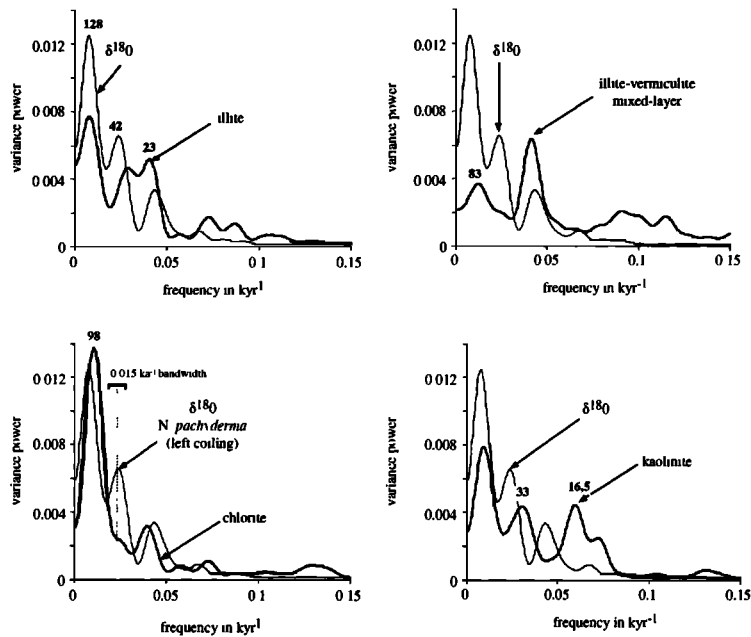
The illite record shows three main frequencies around  $1/128 \text{ kyr}^{-1}$ ,  $1/42 \text{ kyr}^{-1}$ , and  $1/23 \text{ kyr}^{-1}$  (Table 5, Figure 8). The precession forcing dominates the illite record with 18% of the integrated variance. The coherency between the planktonic  $\delta^{18}\text{O}$  and the illite records is highly significant in the  $1/128$  frequency band ( $\text{Cov} = 0.85$ ) and less significant in the  $1/23 \text{ kyr}^{-1}$  band ( $\text{Cov} = 0.55$ ) (Table 5). The additional  $1/35 \text{ kyr}^{-1}$  frequency probably represents a harmonic of studied time -interval ( $176.5 \text{ kyr} / 5$ ), but we notice that a nearby frequency ( $1/31 \text{ kyr}^{-1}$ ) has already been reported in previous works and attributed to polar ice fluctuations [Ruddiman et al., 1984; Pisias and Rea, 1988; Revel et al., 1996]. The phase response of the illite, when compared to

**Table 5.** Blackman-Tuckey Spectral Analysis Cross Correlations of  $\delta^{18}\text{O}$  of *Neogloboquadrina pachyderma* (Left Coiling) With Clay Minerals for Core SU 90-12

1/ Frequency, kyr			Variance Power			Covariance,
$\delta^{18}\text{O}$ <i>pachyderma</i>	Clay Mineral*		$\delta^{18}\text{O}$ <i>pachyderma</i>	Clay Mineral*		Confidence Level = 0.52
128		128	46		31	0.84
42	illite	35	26	illite	17	-
23		25	12		18	0.55
128		83	46		15	0.63
42	is 10-14v	-	26	is 10-14v	-	-
23		24	12		24	0.54
128		98	46		48	0.88
42	chlorite	-	26	chlorite	-	-
23		25	12		10	0.58
128		111	46		30	0.84
42	kaolinite	33	26	kaolinite	17	-
23		-	12		-	-

\*Peak height is in millimeters.

## SU90-12



**Figure 8.** Core SU 90-12. *Blackman-Tuckey* [1958] spectral analysis cross correlations of  $\delta^{18}\text{O}$  *Neogloboquadrina pachyderma* (left coiling) with clay mineral peak height (interpolated data) with a  $0.015 \text{ kyr}^{-1}$  bandwidth.

minimum ice volume, is similar to core SU 90-08 (a lead of about 2.5 rad) in the ice response band and corresponds to a lead of about 0.7 rad (2 kyr) in the precession band. This is, if confirmed, a little smaller than for core SU 90-08 and would indicate a larger influence of the ice-sheet at that latitude in the precession band.

The chlorite record is dominated by the  $1/128 \text{ kyr}^{-1}$  and  $1/23 \text{ kyr}^{-1}$  frequencies (Figure 8, Table 5). The coherencies between the chlorite and the  $\delta^{18}\text{O}$  records are significant in the main frequency bands ( $0.57 < \text{Cov} < 0.88$ ).

The kaolinite record displays three main frequencies with variance maximum around  $1/128 \text{ kyr}^{-1}$  (30% of the variance),  $1/33 \text{ kyr}^{-1}$ , and  $1/16.5 \text{ kyr}^{-1}$ . The coherency between kaolinite and the  $\delta^{18}\text{O}$  records is highly reliable in the  $1/128 \text{ kyr}^{-1}$  frequency band ( $\text{Cov} = 0.89$ ). The additional frequency at  $1/33 \text{ kyr}^{-1}$  has already been observed in the illite record and is attributed to polar ice fluctuations; it is associated with its harmonic at  $16.5 \text{ kyr}$  ( $33 \text{ kyr} / 2$ ). However its integrated variance is greater than the 33-kyr peak. It probably means that this peak is a harmonic or a nonlinear combination of orbital frequencies. The phase responses of both chlorite and kaolinite are similar to illite, as in core SU 90-08.

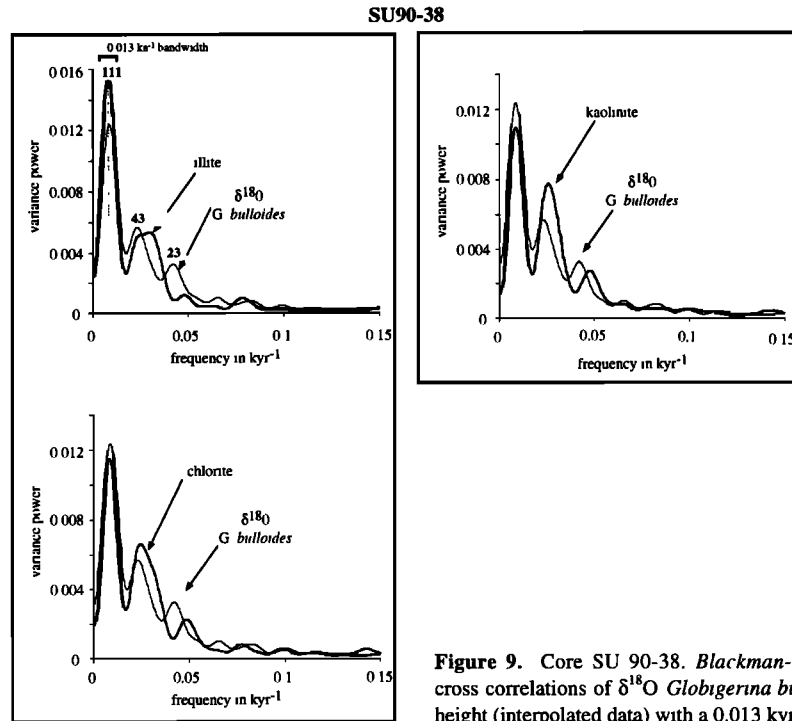
The illite-vermiculite mixed-layer record is mainly controlled by the  $1/23 \text{ kyr}^{-1}$  frequency related to the precession. It reaches 24% of the integrated variance, whereas the  $\delta^{18}\text{O}$  record represents only 12% of the variance in the same frequency band. The coherency between the  $\delta^{18}\text{O}$  and the illite-vermiculite mixed-layer records is barely significant (0.54) but is much higher when cross-correlation is done directly with insolation (0.78). The phase in the precession band corresponds to a lead of the illite-vermiculite mixed-layer by about 2.5 rad ( $80^\circ$ , or about 9 kyr) to maximum northern summer insolation. Taking in account the errors involved (at least 0.3 rad or 1 kyr), this means that the maximum illite-vermiculite mixed-layer fluxes in that core are in

phase with the minimum northern summer insolation or maximum northern winter insolation. The  $1/83 \text{ kyr}^{-1}$  frequency revealed by the spectral analysis is tentatively attributed to a harmonic of studied time-interval ( $176.5 \text{ kyr} / 2$ ), taking into account that the eccentricity band cannot be expressed for that time series.

To summarize, the illite, chlorite, and especially illite-vermiculite-mixed-layer temporal distributions are controlled by the  $1/23 \text{ kyr}^{-1}$  precession frequency and are coherent with the isotopic record in this frequency band. The illite, chlorite, and kaolinite are also controlled by the frequencies near to  $1/100 \text{ kyr}^{-1}$  linked to the dynamic of the northern continental ice-sheet [Ruddiman and McIntyre, 1984; Imbrie et al., 1992]. The isotopic and detrital clay mineral signals are significantly consistent within this frequency band (Table 5). The obliquity-driven  $1/42 \text{ kyr}^{-1}$  frequency which is recorded by the  $\delta^{18}\text{O}$  signal does not appear in the clay records at the  $52^\circ\text{N}$  latitude where core SU 90-12 has been drilled.

**Core SU 90-38.** The core located at  $54^\circ\text{N}$  in the eastern basin near Rockall plateau displays a foraminifera  $\delta^{18}\text{O}$  record which expresses again the three main climatic frequencies. The variance maxima are situated around  $1/111 \text{ kyr}^{-1}$  (40% of the variance),  $1/43 \text{ kyr}^{-1}$  (21%), and  $1/23 \text{ kyr}^{-1}$  (12%) (Figure 9, Table 6). The temporal window (176.5 kyr) is too short for reliable results in the  $1/100 \text{ kyr}^{-1}$  frequency range and will be discussed only as far as general trends are concerned.

The illite record shows a well-marked signal (Figure 9) associated with the  $1/111 \text{ kyr}^{-1}$  frequency and parallels the isotopic record in this frequency band ( $\text{cov.} = 0.83$ ). It also reveals a poorly-defined frequency with a variance maximum located around  $1/43 \text{ kyr}^{-1}$ . The coherency between the illite and  $\delta^{18}\text{O}$  records in this frequency band is significant ( $\text{Cov} = 0.80$ ). The illite spectrum shows a weak signal in the precession



**Figure 9.** Core SU 90-38. Blackman-Tuckey [1958] spectral analysis cross correlations of  $\delta^{18}\text{O}$  *Globigerina bulloides* with clay mineral peak height (interpolated data) with a 0.013 kyr<sup>-1</sup> bandwidth.

frequency band which represents less than 5% of the integrated power. The phase response for illite in the ice volume band (2.8 rad  $\pm$ 0.9) is the same as for the southwestern cores, with maximum input of illite during periods of retreat of the large ice sheets.

The chlorite and kaolinite curves are very similar. They display large dominant frequencies with variance maxima around 1/111 kyr<sup>-1</sup> and 1/43 kyr<sup>-1</sup>. The coherencies between the  $\delta^{18}\text{O}$  of planktonic foraminifera and the chlorite-kaolinite records are highly significant within the two frequency bands (0.81 < Cov < 0.88). In addition, the chlorite and kaolinite records express a third frequency with variance maximum located around 1/23 kyr<sup>-1</sup>. The corresponding integrated variances represent less than 5% of the total signal. The phase responses for chlorite and kaolinite are the same as for illite.

To summarize, the temporal distribution of the detrital clay mineral abundance is mainly controlled by the ice-forcing-driven frequency (1/100 kyr<sup>-1</sup>) and by the obliquity frequency (1/41 kyr<sup>-1</sup>) at 54°N. The 1/23 kyr<sup>-1</sup> precession frequency is almost not expressed by the clay mineral record. The coherencies between clay and isotopic records are highly significant, ranging from 0.80 to 0.88.

**Core SU 90-33.** The core located at 60°N south of Iceland displays a  $\delta^{18}\text{O}$  record marked by three major frequencies with variance maximum situated around 1/135 kyr<sup>-1</sup> (39% of the integrated variance), 1/39 kyr<sup>-1</sup> (20%), and 1/23 kyr<sup>-1</sup> (10%) (Figure 10, Table 7).

The illite curve shows two mean frequencies with variance maximum located around 1/135 kyr<sup>-1</sup> and 1/39 kyr<sup>-1</sup> (Figure 10). The coherencies between illite and  $\delta^{18}\text{O}$  records are significant

**Table 6.** Blackman-Tuckey Spectral Analysis Cross Correlations of  $\delta^{18}\text{O}$  of *Globorotalia bulloides* With Clay Minerals for Core SU 90-38

1/ Frequency, kyr		Variance Power		Covariance,	
$\delta^{18}\text{O}$ <i>bulloides</i>	Clay Mineral*	$\delta^{18}\text{O}$ <i>bulloides</i>	Clay Mineral*	Confidence Level = 0.52	
111		123	41		0.83
43	illite	34	22	illite	0.80
23		-	12	-	-
111		119	41		0.88
43	chlorite	41	22	chlorite	0.85
23		21	12	-	-
111		115	41		0.86
43	kaolinite	39	22	kaolinite	0.81
23		21	12	-	-

\*Peak height is in millimeters.

(0.62 < Cov < 0.64) in both frequency bands. The phase response of illite in that core for the ice volume frequency band (1/100 kyr<sup>-1</sup>) is the same as for other cores.

The chlorite and the kaolinite distributions display similar patterns, with two main frequencies situated around 1/135 kyr<sup>-1</sup> and 1/39 kyr<sup>-1</sup> (Figure 10, Table 7). The coherencies between the clay minerals (i.e., chlorite and kaolinite) and δ<sup>18</sup>O records are very low (0.36 < Cov < 0.47) and located below the confidence level of 0.52 in the 1/135 kyr<sup>-1</sup> frequency band (Table 7). By contrast, the well-marked signal expressed by the kaolinite record and that of chlorite in the 1/39 kyr<sup>-1</sup> frequency band are consistent with that provided by the δ<sup>18</sup>O signal (Table 7).

To summarize, the detrital clay mineral distribution is controlled by frequencies close to 1/100 kyr<sup>-1</sup>, linked to the northern continental ice-volume variations, but the signals are weaker and less coherent with the δ<sup>18</sup>O record than in the three other cores. By contrast, the illite, chlorite, and kaolinite records present a large signal coherent with the δ<sup>18</sup>O results (0.61 < Cov < 0.70) in the obliquity band. The precession frequency is almost absent from the spectrum, which indicates a preferential action of the high-latitude climate forcings (with a source north of 60°N).

## Discussion

The typically terrigenous or land-derived clay minerals illite and chlorite mainly result from physical or moderate weathering of old bedrocks at high latitudes, under present-day cold climatic conditions [Biscaye, 1965; Millot, 1970; Griffin et al., 1968; Rateev et al., 1969; Lisitzin, 1972; Chamley, 1975, 1989]. During glacial stages, when climatic conditions were colder, continental weathering intensified and the detrital input toward the ocean tended to increase [Chamley, 1989].

In the eastern basin, the main sources of chlorite and illite are the Scandinavian shield, Scotland, Ireland, and the eastern Greenland [Moyes et al., 1964; Biscaye, 1965; Griffin et al., 1968]. In the Western basin, chlorite and illite derive from the Precambrian and Paleozoic igneous and sedimentary rocks of the North American continent, the Baffin Island and the Western Greenland [Piper and Slatt, 1977; Petersen and Rasmussen, 1980; Thiébault et al., 1989].

In the northernmost Atlantic Ocean at latitudes higher than 45°N, kaolinite is essentially inherited from adjacent landmasses

where it formed during preglacial times and derives from old paleosols [Darby, 1975; Naidu et al., 1982; Sancetta et al., 1985; Chamley, 1989; Thiébault et al., 1989]. In the eastern basin, the main sources of kaolinite are the soils of Scotland [Wilson et al., 1984], the Mesozoic areas around the Barents Sea [Kuhlemann et al., 1993], and southeastern Svalbard [Elverhøi, 1979]. In the western basin, kaolinite derives from old sedimentary formations of the North American continent [Boyd and Piper, 1976; Piper and Slatt, 1977].

The illite-vermiculite random mixed-layered minerals mainly result from pedogenic processes which worked during Quaternary times under moderate chemical weathering. They seem to especially result from the alteration of mica in an alkaline environment [Millot, 1970] or from pedogenic transformation in soils under temperate climatic conditions [Berry and Johns, 1966]. Their main sources are located in the North American sedimentary basins [McCartan, 1988], on the coastal plain of north Virginia, and in the Canadian Appalachian Mountains [Yang and Hesse, 1991], Quebec [Aargast, 1991] and the Adirondack Mountains, New York [April et al., 1986]. It is also identified as reworked material in fluvial sediments on the western coast of Greenland [Petersen and Rasmussen, 1980].

The comparison between the results of the spectral analysis expressed by terrigenous clay minerals which constitute continental indicators and by the δ<sup>18</sup>O of planktonic foraminifera which represents the changes in ice volume and/or sea level demonstrates that both the orbital forcing and ice-sheet development control the changes in detrital clay mineral input. It also reveals that this climatic control on clay sedimentation varies with the latitudinal location of the investigated cores within the North Atlantic Ocean.

In a general way, the detrital clay minerals resulting essentially from physical weathering processes (illite, chlorite) are chiefly controlled by high-latitude climatic phenomena (1/41 kyr<sup>-1</sup> obliquity frequency), whereas illite-vermiculite mixed-layers resulting from chemical weathering are controlled by low-to middle-latitude processes (dominant 1/23 kyr<sup>-1</sup> frequency).

In the southernmost core SU 90-08, located at 44°N (Figure 1), the land-derived clay minerals (illite-chlorite-kaolinite) are mainly controlled by precession and by the continental ice-volume (Table 8). At this latitude, obliquity does not seem to directly influence the clay mineral sedimentation. At site SU 90-

**Table 7.** Blackman-Tuckey Spectral Analysis Cross Correlations of δ<sup>18</sup>O of *Neogloboquadrina pachyderma* (Left Coiling) With Clay Minerals for Core SU 90-33

1/ Frequency, kyr			Variance Power			Covariance,
δ <sup>18</sup> O <i>pachyderma</i>	Clay Mineral*		δ <sup>18</sup> O <i>pachyderma</i>	Clay Mineral*		Confidence Level = 0.52
135		111	40		30	0.64
39	illite	45	20	illite	20	0.62
23		-	10		-	-
135		135	40		33	0.48
39	chlorite	45	20	chlorite	18	0.61
23		-	10		-	-
111		104	40		29	0.36
43	kaolinite	46	20	kaolinite	26	0.70
23		-	10		-	-

\*Peak height is in millimeters.

SU90-33

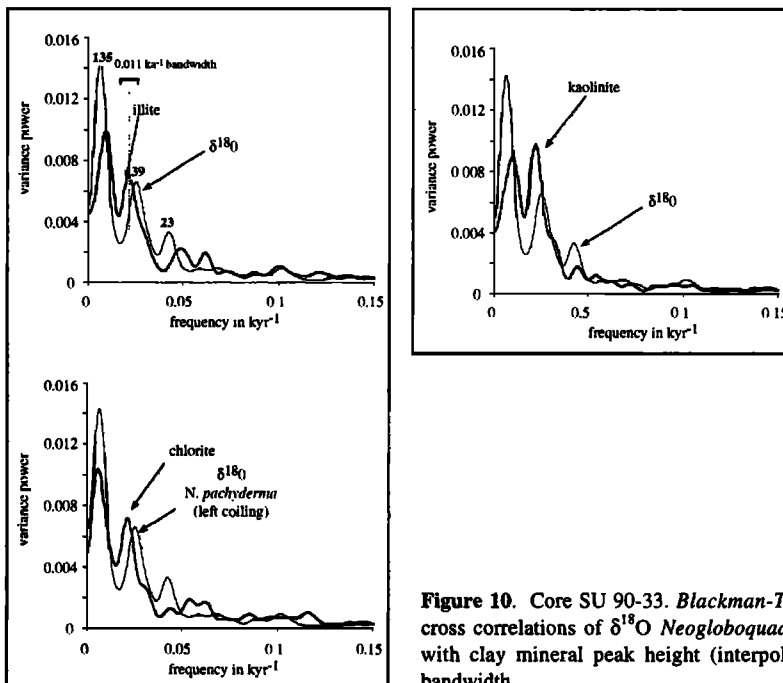


Figure 10. Core SU 90-33. Blackman-Tuckey [1958] spectral analysis cross correlations of  $\delta^{18}\text{O}$  *Neogloboquadrina pachyderma* (left coiling) with clay mineral peak height (interpolated data) with a  $0.011 \text{ kyr}^{-1}$  bandwidth.

12, located at  $51^\circ\text{N}$ , the clay mineral variations are submitted to the same forcing, precession, and ice -volume variations dominating the illite and chlorite signals, whereas the influence of obliquity is not detectable (Table 8). At  $54^\circ\text{N}$  (core SU 90-38) the terrigenous signal is mainly controlled by obliquity and continental ice -volume (Figure 9, Table 8) whereas the role of precession becomes minor. In the northernmost core SU 90-33, located at  $60^\circ\text{N}$ , the obliquity variations are very dominant in the detrital signal (Table 8), whereas the precession signal disappeared (Figure 10).

To summarize, the responses of illite, chlorite, and kaolinite in the three main climatic frequency bands are different in the southwestern and northeastern basins. The  $1/100 \text{ kyr}^{-1}$  frequency is dominant, with a very homogenous phase response of the detrital clay mineral in the two basins. The land-derived clay mineral variations are also controlled by the  $1/41 \text{ kyr}^{-1}$  obliquity

frequency in the northeastern basin and by the  $1/23 \text{ kyr}^{-1}$  precession frequency in the southwestern basin, with a nearly opposite phase response in this precession band. The dominance of the  $1/100 \text{ kyr}^{-1}$  and  $1/41 \text{ kyr}^{-1}$  frequencies in the Northeastern area (cores SU 90-38 and SU 90-33) would indicate that the relative proportion of illite, chlorite, and kaolinite is controlled by the growth and/or decay of the Fenno-Scandinavian ice -sheet. The relative abundance of illite, chlorite, and kaolinite is higher during the ice -sheet decay. It could result from the intensification of the runoff or from the physical weathering of the ice-free bedrocks by the strong easterly winds. The  $1/23 \text{ kyr}^{-1}$  frequency modulation of the illite, chlorite, and kaolinite relative proportion in the southwestern basin (cores SU 90-08 and SU 90-12), which is in phase with the maximum northern summer insolation, could be linked with a northward trend of the westerly winds or of the surface oceanic currents. As the variability in the  $1/23 \text{ kyr}^{-1}$  frequency band is relatively small in deep oceanic parameters ( $\delta^{13}\text{C}$  of benthic foraminifera in core SU 90-08 (L. Vidal, unpublished results, 1996) CFR-Gif), modifications of the deep water masses are not directly involved in order to explain the strong signal observed for the detrital clay minerals in the precession band.

The illite-vermiculite mixed -layers, which are significant in amount only in the northwest Atlantic core SU 90-12, are controlled by low- to middle-latitude processes (dominant  $1/23 \text{ kyr}^{-1}$  frequency). It is surprising not to find any obliquity forcing for this signal, considering the latitude of the core ( $52^\circ\text{N}$ ) and its proximity to the Laurentide ice -sheet. As discussed above, the illite, chlorite, and kaolinite present a nearly opposite phase in the precession band (maximum relative abundance near maximum northern summer insolation) for both cores SU 90-08 and SU 90-12. Different processes need therefore to be considered for these two different responses to the climatic forcing. In core SU 90-12 the maximum illite-vermiculite mixed -layer mineral fluxes are about in phase with the maximum winter insolation. This is the

Table 8. Summary of Detrital Clay Mineral Spectral Analysis Within the Three Main Orbital Frequency Bands Versus Latitudinal Setting of Cores

	SU 90-08 ( $44^\circ\text{N}$ )		SU 90-12 ( $51^\circ\text{N}$ )		SU 90-38 ( $54^\circ\text{N}$ )		SU 90-33 ( $60^\circ\text{N}$ )		
	E	O	P	E	O	P	E	O	P
Illite	H	V	H	V	V	H	H	V	
Chlorite	V	V	V	V	V	V	V	V	
Kaolinite	H	V	H		V	V	H	V	
Mixed-layer				V					

Here mixed-layer is illite-vermiculite. Abbreviations are E, eccentricity band ( $1/1000 \text{ kyr}^{-1}$ ); O, obliquity band ( $1/41 \text{ kyr}^{-1}$ ); P, precession band ( $1/23 \text{ kyr}^{-1}$ ); H, high variance power; V, very high variance power.

response which is observed in other cores from the northwest Atlantic Ocean for summer sea surface temperatures [Ruddiman and McIntyre, 1981a]. It is also the response of the alkenone fluxes to the sediment in core SU 90-08 [Villanueva et al., 1995].

Both these proxies show a dominant  $1/23 \text{ kyr}^{-1}$  variance peak. Ruddiman and McIntyre [1981a] attributed this behavior to the sensitivity of the western north Atlantic wind pattern to the oscillations of the southern limit of the Laurentide ice-sheet, itself modulated by the changes in snow accumulation: The higher the winter insolation at low latitudes, the higher the evaporation and snow precipitation over the southern limit of the ice-sheet. Input of the illite-vermiculite mixed-layer minerals could therefore be modulated by the strength of the westerly winds, transporting that mineral from the North American continent. According to the location of the main source of 10-14 $\mu$  mixed-layers, near the Labrador Sea [Bout-Roumazeilles, 1995], it would mean that the illite-vermiculite mixed-layer supply increased during the continental ice-sheet melting which is induced by the precession-driven poleward heat transport. Other mineral modifications are more or less driven by latitudinal effects.

As a result, the relative proportion of the main detrital clay minerals in the North Atlantic Ocean are mainly controlled by precession ( $1/23 \text{ kyr}^{-1}$ ) south of  $52^\circ\text{N}$  and by obliquity ( $1/41 \text{ kyr}^{-1}$ ) north of  $52^\circ\text{N}$ . The ice-volume variations appear as a global forcing factor independent of the latitude (Figure 11). The results of the spectral analysis expressed by the continental parameters represented by detrital clay minerals are directly comparable with those obtained by Ruddiman and McIntyre [1984] on the variations of sea-surface temperatures. The spectral analysis of the sea-surface temperatures indicates that very large changes occurred during the late Quaternary in the spectral response of North Atlantic surface waters between  $40^\circ$  and  $60^\circ\text{N}$ , and between  $24^\circ\text{W}$  and  $46^\circ\text{W}$  in both the eastern and western Basins. A dominant precessional control (23000 years) occurred south of  $45^\circ\text{N}$  and declined rapidly northward, a dominant obliquity control (41000 years) prevailed north of  $45^\circ\text{N}$  and decreased southward, whereas a global "ice" control (100000 years) was present in the whole area investigated. We found the prevailing 23-kyr band, with the same phase as the changes in sea surface temperature, until  $52^\circ\text{N}$  in the northwestern Atlantic.

Two main hypotheses could explain the latitudinal variations registered and responsible for the climatic control on the detrital clay sedimentation. The first hypothesis implies that the climatic control primarily influences the clay production. The formation of illite-vermiculite mixed-layer minerals by chemical weathering in little-evolved soils would depend, whatever the latitude, on the low-latitude insolation and moisture which should be mainly controlled by the precession. In a similar way, the illite and chlorite supply to the ocean would strongly depend on the high-latitude climatic conditions (obliquity) which directly affect the physical weathering intensity. The detrital clay spectral signal should then express a obliquity-driven control, whatever the latitude. According to this first hypothesis, the differences observed in the detrital clay spectral signal between high and middle latitudes are not easily explained.

The second hypothesis refers to the fact that the climatic control frequently affects the transportation conditions of the detrital clay particles, as shown especially in the Mid-Atlantic Reykjanes Ridge area where clay minerals (i.e., smectite) were

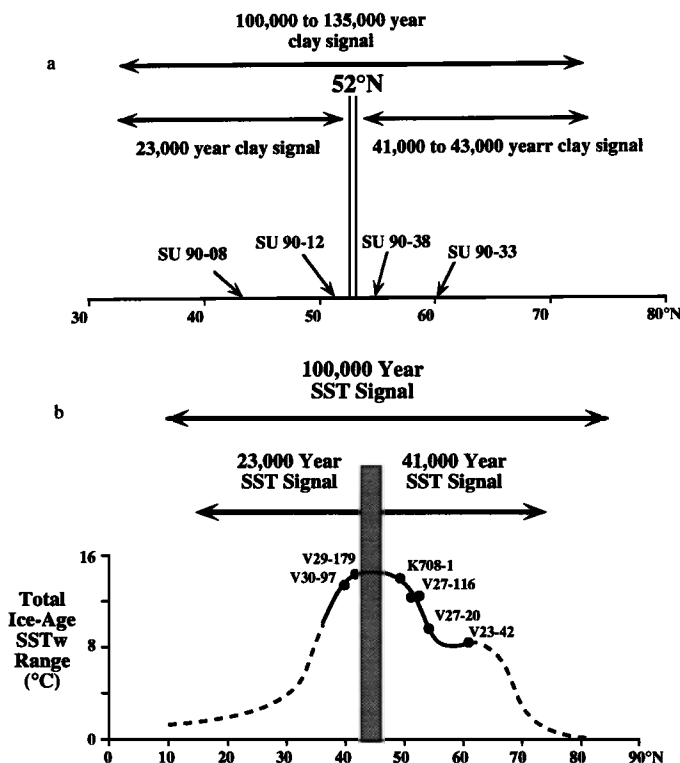


Figure 11. Summary of (a) the spectral analysis results versus latitude compared with (b) the sea surface temperature spectral analysis results from Ruddiman and McIntyre [1984].

primary transported by ice-rafting during the last glacial period, whereas they were transported by bottom currents since 13 ka [Gehrke et al., 1996] and in the Southern Ocean as shown by Robert and Kennett [1992] and Petschick et al. [1996]. According to such a possibility, some of the ideas developed by Ruddiman and McIntyre [1984] would explain the latitudinal variations of the detrital clay spectral signal. At high latitudes, illite and chlorite would be mainly wind-driven, whereas the ocean-driven transportation of detrital clay (illite, chlorite, kaolinite, mixed-layers, etc.) would predominate at midlatitudes. A significant wind-driven clay particles transport (illite and chlorite mainly) originating from the North American continent has already been reported, especially in the western North Atlantic basin [Folger, 1970; Emery et al., 1974; Prospero, 1981; Grousset et al., 1983; Grousset and Chesselet, 1986; Grousset et al., 1988; Grousset and Biscaye, 1989]. The transition between a precession-driven and an obliquity-driven clay supply occurs between  $51^\circ52'\text{N}$  and  $54^\circ05'\text{N}$ . This very sharp transition seems to result partly from the location of cores. The cores located below  $52^\circ\text{N}$  (SU 90-08 ( $43^\circ\text{N}$ ,  $30^\circ\text{W}$ ) and SU 90-12 ( $51^\circ\text{N}$ ,  $39^\circ\text{W}$ )) are situated in the western basin, whereas those located at latitudes higher than  $52^\circ\text{N}$  (SU 90-38 ( $54^\circ\text{N}$ ,  $21^\circ\text{W}$ ) and SU 90-33 ( $60^\circ\text{N}$ ,  $22^\circ\text{W}$ )) are situated in the eastern basin. Moreover, the main atmospheric and oceanic surface circulation patterns have a general SW-NE orientation which would enhance the differences observed between the two oceanic basins. We have, however, no arguments to support an ocean-driven transport. The respective influence of the reasons envisaged through both these hypotheses has to be further investigated.

## Conclusion

1. High resolution clay minerals and foraminifera  $\delta^{18}\text{O}$  records from four cores situated between  $44^\circ$  and  $60^\circ\text{N}$  in the North Atlantic Ocean have been compared over the last 300,000 years by using cross-correlation spectral analysis. The spectral responses of the detrital clay distribution display strong signals associated with the three main climatic frequencies. Within these frequency bands the detrital clay minerals are representative of continental conditions; the  $\delta^{18}\text{O}$  typical of marine sedimentary components usually show good coherencies which range between 0.55 and 0.88.

2. These results, which are based on a very high resolution sampling associated with a well-constrained chronostratigraphic scale, give additional evidence of the climatic control on the terrigenous clay mineral sedimentation revealed by previous works [e.g., Chamley, 1989; Fagel, 1994].

3. The climatic control identified strongly depends on the latitudinal and longitudinal distribution of the sediments investigated, which is enhanced by the main SW-NE trend of both dominant winds and oceanic surface circulation patterns. In the eastern basin, at high latitudes located north of  $52^\circ\text{N}$ , the obliquity is dominant, whereas south of  $52^\circ\text{N}$  the precession effects dominate. The ice-volume variations determine a general

control on the clay mineral sedimentation whatever the latitudinal location of the studied series.

4. We suggest that the latitudinal variations of the climatic control on detrital clay sedimentation result from two main processes. The climatic variations affect first, the intensity of both physical and chemical weathering on the continent and second, the transportation conditions of the terrigenous clay minerals by wind and marine currents. The comparison of our results with those performed on sea -surface temperatures [Ruddiman and McIntyre, 1984] suggests that in the North Atlantic Ocean, the supply of detrital clay deriving from rock erosion mainly depends at high latitude on a wind-driven transfer, and at midlatitudes on ocean currents. The poleward heat oceanic transport from low latitudes is partly responsible for the ice -sheet melting, which releases nepheloids originating from the Labrador Sea and supplying the (10-14v) random mixed -layers [Bout-Roumazeilles et al., 1996].

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