CHAPTER 6
Evaluation of African aridity and North Atlantic Dansgaard-Oeschger cycles during Pliocene MIS101-95

Abstract
High-resolution planktonic and benthic oxygen isotope data of the Mediterranean marine land-based section of San Nicola (Sicily/Italy) and eastern Mediterranean ODP Leg 160 Site 967 are presented together with colour reflectance (CR) and magnetic susceptibility (MS) data of the same sites and of DSDP Leg 94 Site 607 and ODP Leg 162 Site 981 (North Atlantic) throughout marine oxygen isotope stages (MIS) 101-95. Visual inspection of the climate records indicates a relationship between Mediterranean temperature and aridity with North Atlantic temperature and ice rafting history both on a glacial-interglacial (41 kyr) as well as a stadial-interstadial (6-7 kyr) scale and possibly millennial (3.5 kyr) scale. Spectral analysis of the Mediterranean δ¹⁸O records indicates high variance at periodicities similar to those observed in the ice volume record (i.e. 80 kyr, 41 kyr, and 28 kyr; Chapter 5), whereas spectra of the Mediterranean sediment property (CR and MS) records indicate additional variance within the precession band and at periodicities equal to harmonics and combination tones of primary precession, suggesting a low latitude forcing component. On the contrary, frequency-modulation may explaining the 8-12 kyr periodicities in δ¹⁸O and the non-stationary character of this signal in the sub-Milankovitch range, the modulator being probably the ~80 kyr (ice volume) component.
6.1. Introduction

Mediterranean climate during late Pliocene marine oxygen isotope stage 100 (MIS100) underwent rapid fluctuations similar to those associated with the Dansgaard-Oeschger (D-O) and Heinrich events (HE) of the late Pleistocene (Chapter 2). Distinct changes in Mediterranean sea surface temperature (SST) and deep convection occurred in conjunction with changes in SST, thermohaline circulation (THC) and ice rafting in the North Atlantic. This concurrence points to the coupling of Mediterranean and North Atlantic climate on millennial to Milankovitch time scales, either directly through the inflow of Atlantic surface waters into the Mediterranean or, more importantly, indirectly through an atmospheric connection with the North Atlantic climate system. Both these couplings are controlled by rapid oscillations in the Arctic-Greenland ice cap, where in particular severe (winter) winds from the Alps, similar to the present-day bora and mistral winds, during cold European climate conditions (cold D-O phases) resulted in downwelling of Mediterranean surface waters due to intensified cooling.

In addition to the strong winter cooling and the vigorous deep convection in the central Mediterranean during cold D-O and stadial intervals, grain size data and geochemical proxies point to an increase in aeolian dust supply of Saharan origin to the Mediterranean during these intervals. Maximum dust supply from the Sahara occurs during intervals of maximum ice rafting in the North Atlantic (Chapters 3 and 4) indicating a linkage between low-latitude African aridity/wind strength and North Atlantic cooling during MIS100 (Chapters 3 and 4). Such a relationship has been observed in the Mediterranean record of the past 50 kyr [Allen et al., 1999; Moreno et al., 2001; Combourieu-Nebout et al., 2002; Sanchez-Goni et al., 2002; Moreno et al., 2004] and a strong atmospheric cross-latitudinal linkage between mid and high latitude has been suggested.

In order to further evaluate the relationship between low-latitude and high-latitude climate changes on sub-Milankovitch time scales during the major phase of Northern Hemisphere glaciation in the late Pliocene, we extended the benthic and planktonic foraminiferal δ¹⁸O records of San Nicola and ODP Site 967 (Chapters 2 and 3) up to MIS95. The new isotope data were subsequently compared with high-resolution magnetic susceptibility and colour reflectance records of the same sites as well as those of ODP Site 969 (Mediterranean) and ODP Site 981 and DSDP Site 607 in the North Atlantic. The Mediterranean oxygen isotope data suggest SST and deep water changes during MIS98 and MIS96 that are similar to the D-O and H-like climate variations observed during MIS100 (Chapters 2 and 3). The colour reflectance data of the North Atlantic sites suggest ice rafting episodes during MIS100, 98 and 96 on a D-O to Heinrich timescale (Chapter 3), whereas the magnetic susceptibility and colour reflectance records in the Mediterranean are probably controlled by changes in dust supply from the African continent (Chapter 4). To test for cyclicity in the observed...
climate variability records we applied spectral analysis and band-pass filtering in the time as well as in the frequency domain on the long time series.

6.2. Materials and Methods

6.2.1. Sections and sampling
Samples have been derived from the Mediterranean land-based marine section of Monte San Nicola (Sicily/Italy) and ODP Leg 160 Site 967 Hole A (Figure 6.1). Monte San Nicola is situated 10 km north of the coastal town of Gela in southern Sicily (Italy) (for detailed location map and description see [Rio et al., 1994]). San Nicola (SN) consists of a succession of ~160 m of rhythmically bedded marly limestones and marls from the Trubi and Monte Narbone Formation. Sapropels of the O, A, B and C clusters [Verhallen 1987; Zijderveld et al., 1991] can easily be recognised in the section. Depth of deposition was estimated to be 800-1000 m [Bonaduce and Sprovieri 1984; Rio et al., 1994] and the average sedimentation rate is ~8 cm/kyr [Sprovieri et al., 1986]. The Gelasian GSSP (Global boundary Stratotype Section and Point) is formally defined in the San Nicola section at the top of the A5 sapropel, also termed the SN marker bed [Rio et al., 1994].

Marine oxygen isotope stages (MIS) 96-100 are well defined in the central Mediterranean by prominent incursions of the cold water planktonic foraminifer Neogloboquadrina atlantica [Zachariasse et al., 1990]. At San Nicola, these stages are visible in the field as meter-scale grey-white alternations in between sapropel A5 and the B
sapropel group with grey layers reflecting glacial stages 100, 98 and 96. Light-coloured decimetre thick bands are visible within the dark glacial stages. The base of MIS101 is characterised by a dark layer which represents the ghost sapropel that corresponds to insolation cycle i-246 [Lourens et al., 1996].

The interval of MIS101 to MIS95 was sampled in two trajectories with the top part of MIS100 being sampled twice in stratigraphic overlap. Sampling of the older part (MIS101-99) started ~2 m above the San Nicola bed (Chapter 2), while the younger part (MIS100-95) was sampled in the other trajectory starting ~5 m above the SN marker bed. The sampling trajectory was carefully chosen as to ensure a continuous and undisturbed succession. The weathered surface was cleaned and only ‘fresh’ sediment was sampled using an electric water-cooled drill. Generally, two cores of 2.5 cm diameter were drilled per sample level (every 3 cm) resulting in 123 and 341 sample levels for the lower and upper trajectory, respectively. The magnetic susceptibility of the samples was measured on a Kly-2 magnetic facility at the paleomagnetic laboratory Fort Hoofdijk, Utrecht University, to check continuity of the section and build a composite depth profile. Three specimens of each drilled sample were measured in duplicate and results were averaged.

ODP Leg 160 Site 967 (34°04’N, 32°43’E) was drilled in the eastern Mediterranean near the Eratosthenes seamount at a water depth of 2554 m. Three holes were drilled to recover a continuous succession. A composite depth profile was constructed by correlating colour reflectance records between different holes [Sakamoto et al., 1998]. The top 125 m consists of lower Pliocene to Holocene hemipelagic sediments containing 80 sapropels. The sapropels occur in clusters and correlate with the large-scale O, A, B and C sapropel groups as found in the land-based marine successions of the Vrica, Singa, Punta Piccola and San Nicola sections [Kroon et al., 1998; Lourens et al., 1998]. The interval between sapropel A5 and the B sapropel group is marked by a reddish colour. Glacial stages 100, 98 and 96 can be recognised as dark layers in the core. Hole A on Site 967 was sampled every centimetre in the interval 8H4-6 (hereafter referred to as Site 967). The continuity of the section was checked using core photographs and colour reflectance data of all holes (A-C). Additionally, colour reflectance was measured on the half core every centimetre and on the individual samples using a hand-held Minolta CM 503i spectrophotometer.

For stable isotope analysis of the San Nicola samples about 20 specimens of the benthic foraminifer *Uvigerina peregrina* and 50 specimens of the planktonic foraminifer *Globigerinoides ruber* were hand picked from a split of the >212 µm size fraction. The analysis was carried out at Utrecht University stable isotope facility where an ISOCARB common bath carbonate preparation device linked on-line to a VG SIRA24 mass spectrometer is operated. Isotope values were calibrated to the PeeDeeBelemnite (PDB) scale. Analytical precision was determined by replicate analyses and by comparison to the international (IAEA-CO1) and in house carbonate
standard (NAXOS). Replicate analyses showed standard deviations of < 0.06‰ and < 0.1‰ for \(\delta^{13}C\) and \(\delta^{18}O\), respectively.

6.2.2. Age model

The studied interval at ODP Site 967 has been astronomically dated by correlating the Ti/Al record interpreted in terms of the relative contribution of aeolian versus fluvial material to the 65°N summer insolation curve of solution La93 [Lourens et al., 2001]. Lourens et al. [2001] assumed an in-phase relationship between Ti/Al and 65°N summer insolation, resulting in 3-k.yr younger ages for the i-cycles than the lagged ages of Lourens et al. [1996]. Here, the La04(1,0) astronomical solution with present-day values for the Earth’s tidal dissipation and dynamical ellipticity [Laskar et al., 2004] is used, in stead of the older La90-93 solution [Laskar et al., 1993], although both solutions reveal essentially the same ages for the studied interval. Sedimentation rate at Site 967 is on average 2.5 cm/k.yr, resulting in a sample resolution of ~400 yr.

The age model for SN is based on graphical correlation between the benthic oxygen isotope records of SN and Site 967 (Figure 6.2). Tie-points are based on the recognition of glacial-interglacial transitions and stadial-interstadial variability in both benthic \(\delta^{18}O\) records. Graphical correlation yields a correlation of r=0.88 and r=0.87, respectively, between the benthic and planktonic \(\delta^{18}O\) of both locations. Linear interpolation between tie-points results in a duration of ~20 k.yr for MIS100, 18 k.yr for MIS98 and 26 k.yr for MIS96, respectively. At SN, average sedimentation rates are ~6 cm/k.yr in the interglacials and ~8 cm/k.yr in glacials resulting in an average temporal resolution of ~500 yr and ~350 yr, respectively.

The age model of Site 969 is based on correlating the Ti/Al to the 65°N summer insolation curve of the La93 solution [Wehausen 1999]. Here, we refine the initial age assessment of Wehausen by tying the colour reflectance data [Sakamoto et al., 1998] to the new high resolution colour reflectance data of Site 967 (Figure 6.3). This results in slightly modified ages.

6.2.3. Spectral analysis

Spectral analysis is carried out using the MC_CLEAN (version 2.0) frequency analysis program of Heslop and Dekkers, [2002]. MC_CLEAN is based on the CLEAN algorithm of Roberts et al. [1987], which allows to extract frequency information directly from an unevenly-spaced time series. In addition to CLEAN, Monte Carlo methods for different types of noise (here red noise) allow to generate a set of slightly different spectra from the (single) input signal. The differences between these spectra give the confidence interval around the mean spectrum.

In order to test spectral results and to trace non-stationary signals, filtering in the frequency-domain was carried out by using wavelet analysis (Matlab version 6.5,
wavelet toolbox) following the protocol of Torrence and Compo, [1998]. This routine applies a simple Morlet waveform on the normalised, equally-spaced (linearly interpolated) data. Maximum power at the 95% significance level is scaled to 1. For filtering in the time domain, we applied band-pass filters (AnalySeries, Paillard et al., [1996]), using a Gaussian filter with central frequency.

Figure 6.2: Depth-age correlation of the $\delta^{18}$O_benthos of SN (versus depth (cm)) with the $\delta^{18}$O_benthos of Site 967 (versus age (kyr)). Crosses indicate age-calibration points and horizontal dashed lines illustrate graphical correlation. $\delta^{18}$O_benthos of SN, $\delta^{18}$O_G.ruber of Site 967 and SN and sedimentation rates (overlain by average) of SN plotted versus age (kyr). Notice that all $\delta^{18}$O records are plotted on the same horizontal scale (inversed) and that benthic and planktonic $\delta^{18}$O records of SN have been shifted for clarity by respectively 0.4‰ and 1‰ towards lighter values. Labelling of stadials (grey shaded) and interstadials according to Martinson et al. [1987].
6.3. Results

6.3.1. Stable isotopes of SN and ODP Site 967

The San Nicola (SN) δ¹⁸O values for *U. peregrina* (δ¹⁸O<sub>benthos</sub>) vary from ~1.8‰ in the interglacials to ~2.8‰ in the glacials, resulting in a glacial-interglacial amplitude of ~1‰. The SN δ¹⁸O values of *G. ruber* (δ¹⁸O<sub>G.ruber</sub>) are ~0.75‰ in the interglacials with minimum values of ~1.2‰ during i-246. Maximum δ¹⁸O<sub>G.ruber</sub> values are 0.8‰ in MIS100, 0.5‰ in MIS98 and 0.25‰ in MIS96 resulting in glacial-interglacial

![Figure 6.3: Magnetic susceptibility of San Nicola, colour reflectance of Site 967, magnetic susceptibility of Site 967 and colour reflectance of Site 969 versus age (kyr). Greyscale of Site 607 and colour reflectance of Site 981 are plotted versus age (kyr) (lower record; MIS100) and versus depth (m) (upper record; MIS99-95). Notice that all colour reflectance results are plotted on an inverse y-axis. Shaded intervals and labels indicate stadials and cold D-O episodes as defined in Chapter 5. Horizontal dashed lines indicate visual correlation of MS and CR peaks between sites and labels indicate numbering of peaks.](image-url)
amplitudes of 1.55‰, 1.25‰ and 1‰, respectively. Glacial-interglacial amplitudes in δ¹⁸O_{G.ruber} of Site 967 are constant ~1.25‰ with values being ~0.5‰ (slightly lower than SN) in the interglacials and ~0.75‰ in the glacials respectively. Similar to San Nicola, minimum δ¹⁸O_{G.ruber} values (~1‰) are associated with i-246.

Benthic oxygen isotope records both reveal an overall distinct saw-tooth pattern characterised by a gradual increase during the early glacial toward maximum values during full glacial conditions and a sudden decrease at the termination. This saw-tooth pattern is less obvious in the δ¹⁸O_{G.ruber} during MIS98 and MIS96 and seems even reversed during MIS96 at Site 967. Sub-stages as defined in the δ¹⁸O_{benthos} record of SN during MIS100 (Chapter 2) can be recognised in MIS96 and are labelled according to the nomenclature established for late Pleistocene glacial where even numbers refer to stadial phases and odd numbers refer to interstadial phases [Imbrie et al., 1984; Martinson et al., 1987]. Each sub-cycle shows a saw-tooth pattern, starting with a gradual increase in isotope values that takes about 4-5 kyr followed by a sudden decrease to lighter values (with amplitudes of ~0.4-0.5‰). Values remain low for another 2-3 kyr so that the average duration of each sub-cycle is 6-8 kyr. It has been shown that the sub-cycles depicted in the δ¹⁸O_{benthos} of SN during MIS100 are associated with major ice rafting episodes in the North Atlantic and, therefore, present climate variability equivalent to the Pleistocene Bond-cycles (Chapter 3). These Bond-type cycles are also observed in the δ¹⁸O_{G.ruber} of SN but are less pronounced in the δ¹⁸O of Site 967. No such stadial-interstadial variations are observed in the δ¹⁸O during MIS98, although δ¹⁸O decrease for short episodes (~2 kyr), similar to the short decreases observed during the gradual increase in δ¹⁸O of each Bond-cycle. Again, these short episodes in MIS100 have been shown to be equivalent to Pleistocene Dansgaard-Oeschger warm phases (Chapter 3).

6.3.2. Magnetic susceptibility and colour reflectance

The San Nicola magnetic susceptibility (MS) record generally matches the weathered and fresh sediment colour changes as observed in the field (Chapter 1). High MS values (~160 SI*10⁻⁴) are associated with the darker-coloured glacial sediments and the ghost sapropel (i-246) and low MS values (~80 SI*10⁻⁴) with the lighter sediments. During glacials, MS values vary with high amplitude (~20-40 SI*10⁻⁴) showing distinct maxima with a spacing of 3-6 kyr: four extremes (1-4) can be labelled in MIS100, three in MIS98 and four in MIS96 (Figure 6.3). Given the duration of the glacials this results in a spacing of the extremes of ~5 kyr in MIS100, ~6 kyr in MIS98 and ~6.5 kyr in MIS96, respectively. Furthermore, the 1st and 2nd peak in MIS96 and the 3rd and 4th peak in MIS100 represent double peaks with a shorter spacing of ~3 kyr. Major peaks in MS correspond to maximum δ¹⁸O_{benthos} values of SN (Figure 6.4) and thus stadial and D-O cold intervals (shaded). Lower amplitude variations with irregular spacing occur during interglacial stages.
The colour reflectance (CR) and MS records of ODP Site 967 reveal a similar variability as the MS record of SN with low CR values (dark coloured sediment) being associated with glacials and the ghost sapropel and high values with interglacials (Figure 6.3). High-amplitude variations in CR are found throughout the entire interval. Distinct peaks during glacial intervals can be correlated to the MS of SN and have been labelled accordingly. Small differences are found between the two sites: in MIS100, the 3rd peak is more pronounced in Site 967 than at SN and better separated from the 2nd

![Figure 6.4: Magnetic susceptibility of San Nicola, δ¹⁸O_benthos of SN, colour reflectance of Site 967 and δ¹⁸O_benthos of Site 967 versus age. Notice that δ¹⁸O_benthos and colour reflectance results are plotted on an inverse y-axis. Shaded intervals and labels indicate stadials and cold D-O episodes as defined in Chapter 5. Horizontal stipple lines and labels indicate correlative peaks.](image)
peak. In MIS96.3, part of the additional peak (\( \ast \)) is missing at Site 967 because an interstitial water sample has been taken at this position in core 967A-8H-4 (Site 967 core photographs). In the parallel hole B of Site 967, both peaks (\( \ast \) and 2) are resolved (Shipboard data), indicating that two peaks are associated with this interval. In addition, the 3rd peak in MIS96 has lower amplitude than at SN and is less distinct. Similar to SN high CR values are related to maximum \( \delta^{18}O_{benthos} \) with the exception of those peaks in interstadial MIS100.3 and MIS96.3 (Figure 6.5). Comparison with the colour reflectance record of Site 969 (scientific shipboard party) shows that all peaks can be correlated basin-wide (Figure 6.3).

At Sites 607 and 981, high grey-scale values and low CR values (Chapter 3) are

![Figure 6.5: Greyscale record of Site 607, \( \delta^{18}O_{benthos} \) of Site 607, colour reflectance of Site 981 and \( \delta^{18}O_{benthos} \) of Site 981. Lower part of record (MIS100) is plotted versus age (kyr) (Chapter 3) and upper part (MIS99-95) of the record is plotted versus depth (m) (Site 981: data of Draut et al. [2003]; Site 607: data of Raymo et al., [1989]). Notice that \( \delta^{18}O_{benthos} \) and colour reflectance results are plotted on an inverse y-axis. Shaded intervals and labels indicate stadials and cold D-O phases as defined in the Mediterranean record. Dashed lines and labels indicate correlative peaks.](image-url)
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associated with dark glacial sediments in MIS100, 98 and 96 (Figure 6.3). Four dark intervals (1-4) are recorded within the glacial stages MIS100 and MIS96 and three with MIS98, which correspond to or are close to maximum $\delta^{18}O_{\text{benthos}}$ values ($\delta^{18}O_{\text{benthos}}$ of Site 607: Raymo et al., [1989]; $\delta^{18}O_{\text{benthos}}$ of Site 981: Draut et al. [2003]) (Figure 6.5). In contrast, the uppermost dark interval of each glacial cycle (0) does not correspond with a $\delta^{18}O_{\text{benthos}}$ maximum but with decreasing $\delta^{18}O_{\text{benthos}}$ values marking the glacial-interglacial transition.

6.3.3. Spectral results

Oxygen isotopes. The spectra of the $\delta^{18}O_{\text{benthos}}$ records reveal maximum concentration of power at or near periodicities of 80 kyr, 41 kyr and 28 kyr (Figure 6.6). This is in agreement with results as discussed in Chapter 5. No significant concentration of

![Figure 6.6: Power spectrum (versus period in kyr) of $\delta^{18}O_{\text{benthos}}$ of Site 967 and SN in relation to the 80-95% significance levels (dark grey stipple lines) and corresponding wavelet maps. Important periodicities are highlighted by horizontal grey bars. Grey intervals on the right axis of the wavelet map indicate positions of MIS100, 98 and 96.](image)
variance is resolved in the precession-band. Additionally, peaks are resolved at or near sub-Milankovitch periodicities of 12 kyr, 9 kyr, 7–6 kyr and 4–3 kyr. These are considered to be significant above the 85-90% confidence level. Results of wavelet analysis indicate that concentration of variance is spread over a relatively broad range of periodicities and not uniformly distributed throughout the time series, which explains the rather low significance levels. To pinpoint these differences in the time domain, filters with a relatively broad band were used. To evaluate the 7–9 kyr periodicity range different filters were selected to trace the signal around the upper (8.3 kyr <10 kyr <12 kyr) and lower (5.8 kyr <6.7 kyr <7.7 kyr) margins of the frequency band (Figure 6.7). The long 10 kyr filter picks up the subcycles in the benthic $\delta^{18}O$ during MIS96, while the 6 kyr filter picks up the $\delta^{18}O_{benthos}$ subcycles during MIS100 (SN) and MIS98. This is in agreement with a priori observations in the time domain (see results section $\delta^{18}O$). Higher frequency (low amplitude) variations within subcycles are picked up by fixing the filter at 3.3 kyr <3.6 kyr <3.8 kyr.

The power spectra of the $\delta^{18}O_{Globigerina}$ reveal maximum concentration of power at or near periodicities of 80 kyr and 41 kyr in agreement with those of the $\delta^{18}O_{benthos}$ (Figure 6.8). Variance at 28 kyr is observed in the spectrum of SN but absent in Site 967.
Additionally, concentration of variance around the precession-band is observed in both records, although concentrated occurs in a rather broad 28-22 kyr band, with maximum concentration during MIS98. Concentration of variance in the sub-Milankovitch band is observed at or near periodicities of 11-8 kyr, 7-5 kyr, 3-4 kyr, 2.5 kyr and 1.7 kyr. Again, variance is concentrated in rather broad bands and is not uniformly distributed throughout the time series causing rather low significance levels at Site 967. Results of wavelet analysis (Figure 6.8) and Gaussian filters (Figure 6.9) of SN indicate that variations around 10-8 kyr are important in MIS98-96, and

![Figure 6.8: Power spectrum (versus period) of δ^18O_ruber of Site 967 and SN in relation to the 80-95% significance levels (dark grey stipple lines) and corresponding wavelet maps (lower panel versus age (kyr)). Important periodicities are highlighted by horizontal grey bars. Grey intervals on the right axis of the wavelet map indicate positions of MIS100, 98 and 96.](image_url)
correspond to the sub-cycles in MIS96. On the contrary, variations ~7 kyr are more important during MIS100 (highlighted in Figure 6.6). This is similar to observations in the $\delta^{18}O_{benthos}$ where shorter cycles were observed during MIS100 compared to MIS96. In contrast to the SN $\delta^{18}O_{benthos}$, cycles in both these ranges are also detected in interglacials. This is even more evident in the filtered records of Site 967 $\delta^{18}O_{ruber}$, where the long period ~10 kyr is strongly related to interglacials and MIS98 (Figure 6.9). The shorter (6-7 kyr) is strong in the entire record except during the late stage of MIS100 and MIS96. Periodicities ~3-4 kyr in the $\delta^{18}O_{ruber}$ are related to the shorter low amplitude fluctuations in the glacials (except MIS96 at SN) and in the interglacials of Site 967.

**Magnetic susceptibility and Colour reflectance:** Spectra of the CR (Site 967 and Site 969) and MS (SN) are very different in the Milankovitch-range (Figure 6.10). At SN and Site 967 maximum concentration of variance is centred at 41 kyr with additional variance occurring around 66 kyr (Site 969) and 70 kyr (SN). Variance at 28 kyr is reduced at Site 969 and hidden in the flank of the large 41 kyr peak at SN. Variance at 23 kyr is present at SN. At Site 967 concentration of variance at 41 kyr is reduced compared to the large peaks observed at 80 kyr, 28 kyr and 21 kyr, possible due to the larger...
precession signal at this site. Results of wavelet analysis indicate, that variance at 29-19 kyr is clearly related to MIS101-98 in all records.

Peaks in the sub-Milankovitch range are resolved around periodicities of 10-9 kyr, 6.7-5 kyr, ~3.5 kyr and 2.3 kyr and are significant above the 85-90% level. Distribution of variance in the time domain looks rather difficult given the variety of peaks in the spectrum (Figure 6.11). Variance around 10 kyr seems to be related to the ghost sapropel and the upper part of MIS100-MIS99 in all records and during MIS96 in Site 967, picking up the three darkest intervals in that interval. Concentration of variance at ~6-7 kyr is related to the glacial stages, although not very clear in the lower part of SN, and to the interglacials MIS99 (Site 969 and Site 967) and MIS97 (Site 967). This observation is in agreement with observations in the depth domain and the labelling of the dark intervals accordingly (Chapter 1, i.e. during MIS96, four peaks were labelled at all sites and this results in an average spacing of 6 kyr given the duration of that glacial
Power at around 3.5 kyr is strongly related to the glacial episodes at SN and Site 967 (MIS98 only).

6.4. Discussion

6.4.1. Dust episodes in the Mediterranean

The good agreement between magnetic susceptibility, colour reflectance and bulk carbonate content ($r^2=0.96$; Chapter 4) at SN mirrors the typical cyclic character of the limestone-marl alternations in the Narbone Formation with light-coloured layers (interglacials, homogenous intervals) being carbonate-rich and dark-coloured layers (glacials, sapropels, sapropelitic layers) being carbonate-poor. As such, these changes in bulk sediment composition can be driven by (i) variations in the carbonate productivity, (ii) dilution of carbonates by aluminosilicates, (iii) dissolution of carbonates through oxidation of organic material or (iv) a combination of i-iii. In Chapter 4 it was discussed that a combination of both changes in carbonate productivity and aluminosilicate input are most likely the most important mechanisms controlling bulk sediment composition. The sediments at SN may thus represent a two-component mixing system with biogenic carbonate and terrigenous clastic material as end-members (Chapter 4). Sediments at ODP Sites 967 and 969 [Wehausen 1999] are similar in composition and changes in CR and MS at these Sites can thus be similarly
interpreted. In addition, the basin-wide registration and correlation of these signals suggests a climatic origin.

Foraminiferal fluxes at SN indicate that carbonate production was generally lower (about twice as low) during MIS100 than during the encompassing interglacials (Chapter 2). Grain size analyses of the carbonate free component at SN and geochemical proxies (Ti/Al and Zr/Al) at SN, Site 967 and Site 969 further indicate that at least part of the observed fluctuations in the CaCO$_3$ and Al content (and thus MS and CR) is related to wind-transported material with high Al values (high MS values) being related to periods of increased dust input from the African continent, either by an increase in wind strength or by aridification (Chapter 4).

![Figure 6.12: $\delta^{18}O_{benthos}$ and magnetic susceptibility of SN and Ti/Al (data Wehausen 1999) and colour reflectance of Site 969 versus age in comparison with greyscale, IRD and $\delta^{18}O_{benthos}$ of Site 607 versus age (kyr) (lower part (MIS100) of Site 607 and depth (m) (upper part (MIS99-95) data of Raymo et al., [1989]). Horizontal stipple lines and labels indicate correlative peaks.

Comparison between the Ti/Al and CR records of Site 969, and MS and CR records of SN and Site 967 reveals that a similar relationship holds for MIS98 and MIS96 (Figure 6.2 and Figure 6.12). Evidently, MS and CR peaks at SN and Site 967 have an
equivalent peak in the CR and Ti/Al records of Site 969 (Figure 6.2 and Figure 6.12). In analogue to the long-term changes in Ti/Al, Si/Al and Zr/Al series of Sites 969 and 967 which vary in concert with summer insolation [Wehausen 1999], these rapid fluctuations have been interpreted to present dry-wet alterations of circum-Mediterranean climate conditions. The episodes of enhanced dust input correspond primarily to δ¹⁸O maxima indicating a strong link with cold D-O phases and/or stadial intervals (see also Chapter 2). During the two exceptions, MIS100.3 and MIS96.3, when high dust influxes occurred at times of low δ¹⁸O values at SN, the precession cycle reaches minimum values (i.e., low background Ti/Al values in Site 967 and 969). This orbital configuration may have resulted in the exceptional low δ¹⁸O values during MIS100 and MIS96 related to a surface water temperature/salinity signal similarly as at times of sapropel formation, although less intense, whereas the increased dust flux may be related to a cold D-O phase, which operates on a much shorter time scale (Chapter 2).

6.4.2. Relation with the open ocean record

The occurrence of dark coloured layers at Sites 607 and 981 were interpreted as reflecting periods of enhanced terrigenous clastic input (ice rafted debris; IRD), which interrupt the biogenic carbonates and oozes [Carter and Raymo 1999]. Grey-scale data of different North Atlantic Sites within the IRD belt [Ruddiman et al., 1989] indicates that the occurrence of these dark sediments is restricted to glacial intervals and that variations in sediment colour in MIS100-96 can be correlated basin-wide and hence sustain the notion that CR changes are associated with distinct basin-wide phases of ice rafting during MIS100, 98 and 96.

The comparison between the CR and δ¹⁸O_benthos records of Sites 607 and 981 further indicated that the dark sediment layers correlate with or are close to maximum δ¹⁸O_benthos values in the northern Atlantic (Figure 6.5), indicating a direct relation between cold episodes (maximum ice volume) and ice rafting. Taken the average spacing of ~6 kyr between prominent dark layers suggests that these oscillations are similar in nature and spacing as the Pleistocene Heinrich events. In addition, these prominent dark layers consist of multiple advances suggesting that glacial surges appeared on even shorter time scales comparable to that of a D-O type of variability. Despite the reduced variability in and lower resolution of the Atlantic δ¹⁸O_benthos records used for stratigraphic correlation, the striking similarity between the short-term Mediterranean dust episodes and dark sediment layers at Sites 607 and 981 during MIS100, MIS98 and MIS96 (Figure 6.5, Chapter 3), indicates a strong link between circum Mediterranean and North Atlantic climates. The discrepancy of five IRD phases versus four MS/CR peaks is probably related to the additional thick IRD peak at the glacial/interglacial transition of MIS100 and MIS96, which seems therefore not related with extreme cold glacial conditions, but to melting of ice caps at the glacial-
interglacial transition. The observed relation between dust, IRD and δ¹⁸Obenthos is valid for most of the cold phases (stadials).

Exceptions seem to occur at times of the precession minima when glacial CR/greyscale values at Site 607 are relatively low, suggesting a reduced IRD input. This was already observed for interstadial MIS100.3, where low IRD abundance were associated with unusual warm surface water conditions in the mid-Atlantic and an almost interglacial deep water production in the North Atlantic as a consequence of an increase in the meridional heat transport during minimum precession (i-cycle 244; Chapter 4). The similarity of the IRD signal during insolation cycles i-240 and i-236 with the signal during i-cycle 244 suggests that conditions are comparable during these intervals and that precession-related climate signals may suppress D-O and H-type of variability during the dominantly obliquity controlled glacials (Figure 6.5; Chapter 3), although this did not lead to significant ice volume changes on a precession scale (Chapter 3).

In conclusion, despite the different age models and temporal resolution, the visual peak to peak correlation between the proxy records suggests that African dust transport to the Mediterranean varied in harmony with the waxing and waning of the ice cap at high northern latitudes throughout MIS100-96 on sub-Milankovitch time scales. Mediterranean climate during glacial stages MIS100-96 is dominated by an alternation of relatively cold phases with surface water cooling, deep convection and enhanced aeolian dust input and more temperate, stable climate phases. Dry cold phases in the Mediterranean correspond to cold stadial phases in the mid- to high-latitude Atlantic marked by changes in meridional heat transport from low latitudes and the position of the polar front and by significant input of IRD (Chapter 4). The relation between Mediterranean SST, North African dust and North Atlantic IRD points to an atmospheric connection, most likely through the trade wind system.

6.4.3. Forcing mechanisms of sub-Milankovitch climate variability

Contrasting ideas exists about the linkage between high-latitude climate and ice rafting, and low latitude winds. Hughen et al. [1996] suggested that changes in North Atlantic sea surface temperature driven by modifications in the North Atlantic thermohaline circulation influence the tropical trade wind and summer monsoon strength. By contrast, McIntyre and Molfino [1996] proposed that changes in zonal wind-driven divergence in the eastern equatorial Atlantic control the discharge of tropical surface waters from the Caribbean and Gulf of Mexico warm pool into the western boundary current of the North Atlantic subtropical gyre and further into the sub-polar Atlantic thereby producing the rapid melting of ice and hence Heinrich events. According to the authors, these changes that occurred with a period of 8.4 kyr over the last 45,000 years are caused by a nonlinear climate response to low-latitude insolation/precession forcing during minimum eccentricity.
In his original paper, Heinrich [1988] proposed that North Atlantic ice rafting events are primarily a response to precession-induced variations in seasonal insolation. He related ice rafting events to times of respectively maximum summer and maximum winter insolation, thereby stating that these events occur every 11 kyr. Our data do not provide compelling evidence that late Pliocene climate fluctuations with frequencies similar to those of the much younger Heinrich events were triggered by low-latitude forcing even though we proposed that precession-induced changes in meridional heat flux may drive variations in North Atlantic deep water formation and thus thermohaline circulation during MIS100 (Chapter 4). On the other hand our results are also not in disagreement with the model proposed by McIntyre and Molfino [1996] although their model requires interactions at a rather exceptional precession period [Berger and Loutre 1997]. Furthermore, our data are in excellent agreement with data from the late Pleistocene mid-Atlantic of Chapman and Shackleton [1998]. These data show a broad peak around 5-7 kyr in the $\delta^{18}O_{benthos}$ spectrum, which might actually be composed of two separate peaks centred around 7 and 5.7 kyr. The data also show significant concentration of variance at the semi-precession cycle and Chapman and Shackleton [1998] state that cross-equatorial heat transfer exerts a major control on the mid-latitude Atlantic, again suggesting a low-latitude control. But what could actually cause such rapid climate fluctuations on millennial timescales?

No consensus exists about the forcing mechanism underlying sub-Milankovitch climate variability, notwithstanding the huge amount of high-resolution proxy records derived from various archives, latitudes and time intervals, which substantiate sub-Milankovitch variability at similar periodicities [e.g. Yiou et al., 1994, for overview]. Different forcing mechanisms have been proposed in the literature. For example, changes in solar output [Van Geel et al., 1999; Perry and Hsu 2000; Bond et al., 2001] or tidal motion [Keeling and Whorf 2000] were used to explain climate variations in the decadal to millennial frequency range and, in particular, to account for the world-wide expression of D-O related cyclicity. However, these external forcing mechanisms are relatively weak and it remains therefore difficult to explain how they influenced the climate system. Modelling studies predict that small fluctuations in solar energy related to the 11-year solar (sunspot) cycle may affect atmospheric circulation patterns [Haigh 1994; 1996]. Also sub-harmonics of the 11-year sunspot cycle [Gauthier 1999] and amplifying mechanisms involving the role of UV variations and solar wind [Van Geel et al., 1999] have been proposed to link millennial scale climate variability to solar activity.

Alternatively, self-sustained free oscillations [Le Treut and Ghil 1983; Le Treut et al., 1988] and internal oscillations within the ocean-atmosphere system [Sakai and Peltier 1997] or the ice sheets [MacAyeal 1993; Van Kreveld and al. 2000] have been put forward to explain the D-O and H-events. Clearly northern Hemisphere ice sheets are important because high-amplitude millennial-scale climate variations occur during full glacial conditions in contrast to the relative low-amplitude variability during interglacial
times (e.g. Holocene). However, millennial-scale climate variations with similar periodicities have also been found at times that large northern Hemisphere ice sheets were absent and prior to northern Hemisphere glaciations [Ortiz et al., 1999; Steenbrink et al., 2003]. It was therefore suggested that northern Hemisphere ice sheets (or possibly even southern Hemisphere ice sheets) might act as amplifiers [Raymo et al., 1992; McIntyre et al., 2001] or resonating systems [Wara et al., 2000] causing amplification or frequency-modulation rather than triggering high-frequency climate variability. Evidently, the world-wide recognition of climate variability at D-O and H frequencies over the last 100,000 years suggests that ocean and ice sheets are amplifiers operating mostly via feedback mechanisms and internal oscillations.

A widely used explanation for sub-Milankovitch climate variability with periodicities ranging between 12 and 4 kyr is a non-linear response to primary Milankovitch forcing, which will introduce harmonics and produce combination tones [Pestiaux et al., 1988; Hagelberg et al., 1994; Ortiz et al., 1999]. Generally, the potential for such a non-linear climate response is supported by the notion that climate records contain integrated signals that hold information from more than just one season [Crowley 1992]. The presence of variations near the 2nd, 3rd and 4th harmonics of primary precession (periodicities of 12–5 kyr) was regarded evidence for non-linear climate system response to primary orbital forcing [Pokras and Mix 1987]. The delivery of the freshwater diatom Melosira to the equatorial Atlantic at half- and quarter precession frequencies suggested that African aridity and changes in the trade-wind system react in a non-linear way to precession forcing, although lake desiccation itself is a classic example of a nonlinear process [Crowley et al., 1992]. Pestiaux et al., [1988] concluded that monsoon-related processes are responsible for a nonlinear climate system response to precessional forcing in the Indian ocean. The importance of low-latitude climate forcing is strengthened by the notion that the twice-yearly overhead passage of the sun across the equator [Short and Mengel 1986; Short et al., 1991] might create a climate cycle with a semi-precession frequency. Evidently, such a low-latitude climate forcing scenario is in conflict with suggestions that internal ice sheet-ocean oscillations trigger sub-Milankovitch climate variability.

6.4.4. Harmonics and combination tones of primary Milankovitch cycles

In principle, a variety of frequencies in the sub-Milankovitch range can be generated as harmonics and combination tones of the primary Milankovitch frequencies because climatic precession is not composed of a single component but is dominated by double components around 23 and 19 kyr with periods of 22.2 and 23.8 kyr, and of 18.8 kyr, respectively (see von Dobeneck and Schmieder, [1999] for an overview). As a consequence the pattern will increase in complexity with increasing frequency although the amplitude will decrease. In palaeoclimatic timeseries, the presence of additional free oscillations [Saltzman and Sutera 1984; Le Treut et al., 1988] and stochastic noise
increases the difficulty in interpreting spectral results and separating internal climate system variability from external forcing. Gaining information about the underlying climate mechanisms from spectral analysis is, therefore, challenging.

Harmonics occur in narrow frequency bands and decrease in amplitude with increasing frequency. The only proof for the presence of harmonics is their detection in the time as well as in the frequency domain and by testing them against periodicities found in different environmental settings, climate proxies and time intervals using different spectral methods. Spectral results of SN, Site 967 and Site 969 are consistent with results from other palaeoclimatic studies [e.g. Yiou et al., 1994 for overview] and, therefore, a similar origin might be expected. In the time domain, the $\delta^{18}O$ and CR-MS time series indicate climate variations with average spacing of roughly $\sim 10$ kyr, $\sim 7$ kyr, 3-4 kyr and 1-2 kyr (see also Chapter 2 and Chapter 3). In the frequency domain, variance is concentrated around these frequencies, although in relatively broad bands.

If we consider in a first attempt that these periodicities present harmonics of the primary obliquity cycle, the 2nd to 5th harmonic (treating the fundamental as the first harmonic) would be 20.5 kyr, 13.6 kyr, 10.3 kyr and 8.2 kyr. Evidently, the 2nd and 4th harmonic of 41 kyr periodicity fall into the precession- and half-precession range and, therefore, are difficult to identify. Concentration of variance $\sim 14$ kyr is observed in the MS and CR spectra and vaguely in the $\delta^{18}O$. Variance at $\sim 8$ kyr is observed in the $\delta^{18}O$, however, the detection of a 5th harmonic would be very unlikely to be traced in palaeoclimate records in the presence of (red) noise.

Harmonics produced by the different precession components 23.8 kyr, 22.2 kyr and 18.8 kyr cover a rather broad range [e.g. Berger 1988 for overview]. Variance at the 2nd (9.4-11.9 kyr) 3rd (6.3-7.9 kyr) and 4th (4.7-5.9 kyr) harmonic are observed in the spectra but their interpretation is not straightforward. For example, the observed 6.7 kyr periodicity in the MS record of SN and the 6 kyr in the CR record of Sites 969 and 967 may have a similar origin as indicated by bandpass filtering. A periodicity of 6 kyr can either be related to the 3rd or 4th harmonic, if we start from a fundamental period of either 18.8 kyr or 23.8 kyr. The 18.8 kyr precession component reaches a deep minimum at 2.5 Ma and has as a consequence a weak expression in the insolation record. At such an astronomical configuration, the 18.8 kyr cycle may even be a doublet itself [Hinnov 2004], thereby increasing the uncertainty in the interpretation of the spectral peaks. On the contrary, the 23.8 kyr and 22.2 kyr precession components have stable amplitudes within the time interval studied, however, the presence of a 4th harmonic, necessary to explain the observed variability at 6 kyr, is unlikely. It is more likely that the lower periodicities are related to combination tones of the primary Milankovitch (precession) periodicities.

Periodicities of 7-5 kyr and 3.4 kyr may represent sum combination tones of the primary precession couplets and their harmonics. A periodicity of 7.1 kyr can be produced by the combination tone of the 18.8 kyr cycle with the 2nd harmonic of the
Similarly, the combination of the 23.8 kyr cycle with the 2nd harmonic of the 18.8 kyr cycle will produce a 6.7 kyr cycle and the combination of the 2nd harmonics of the 18.8 and 23.8 kyr cycle a 5.2 kyr cycle. A periodicity of 3.4 kyr results from combining the primary 18.8 kyr cycle with the 3rd harmonic of the 23.8 kyr cycle.

In summary, spectral results of SN, Site 967 and Site 969 proxy data suggest a strong relation between sub-Milankovitch climate variability and primary Milankovitch forcing. In the next section we will deal with the question how these signals are translated into (regional) climate variability.

6.4.5. Possible climate forcing mechanism at sub-Milankovitch timescales

Clipped precession

Climate oscillations at periodicities equal to precession harmonics have been predicted by a two-dimensional seasonal energy balance model, indicating a complex response of surface temperature to insolation forcing at equatorial latitudes where the sun passes twice a year overhead producing two insolation maxima in the annual cycle [Short et al., 1991]. Since climate is responding to the largest possible maximum insolation, independently of the month of the year (i.e. in the season that occurs nearest to perihelion), maximum summer temperature ($T_{\text{max}}$) moves from spring to fall and vice versa as perihelion interacts with the overhead passage of the sun at either equinox [Short and Mengel 1986; Short et al., 1991].

At the equator, $T_{\text{max}}$ shows a 'clipped' curve with two maxima spaced exactly half a precession cycle apart, thus potentially capable of generating a semi-precession cycle. A major drawback of this scenario is that it is only valid at the equator itself. Further away from the equator, spacing of the two $T_{\text{max}}$ becomes asymmetric until a single $T_{\text{max}}$ is reached per precession cycle at the tropics of Cancer and Capricorn. However, also the difference between $T_{\text{max}}$ and $T_{\text{min}}$ varies in a complicated way at equatorial latitudes. The difference between the strongest temperature maximum ($T_{\text{max}}$) and the strongest minimum ($T_{\text{min}}$) in the annual cycle results in a curve at the equator with maxima spaced a quarter of a precession cycle apart [Berger pers. comm.].

The difference between $T_{\text{max}}$ and $T_{\text{min}}$ (minmax curve) is shown in (Figure 6.13) for the interval between 2 and 3 Ma. This curve is dominated by eccentricity. The eccentricity detrended curve shows the typical modulation by obliquity and is very similar to the La04(1,1) 65°N summer insolation curve (lower panel) although 20 times lower in amplitude. Spectral results indicate a dominant periodicity of 5-5.8 kyr and concentration of variance at the 2nd, 3rd and 4th harmonics of the 5-6 kyr with periodicities of 2.9-2.5 kyr, 1.9-1.7 kyr and 1.5-1.4 kyr, respectively. These periodicities are also hidden in the original La04(1,1) 65°N summer insolation curve, but become statistically significant as precession is averaged out in the way described above. Although the amplitude and thus the forcing at these frequencies is small, the effect on
climate processes at low latitudes can not be ruled out and can be important in understanding sub-Milankovitch climate variability at low-latitudes and the transfer of sensible and latent heat to high-latitudes.

The ‘minmax’ curve suffers essentially from the same short-coming as $T_{\text{max}}$ in the way that the curve starts to reveal a different pattern and spectral characteristics when moved away from the equator. Construction of a ‘minmax’ curve at latitudes between 0-23.5°N results in a target curve that is highly nonlinear (Figure 6.14). For instance the eccentricity detrended ‘minmax’ curve for 5°N varies in concert with $La_{04_{(1,1)}}$ 65°N summer insolation (black line). The most important periodicities are indicated in the graph.

Figure 6.13: a) Equatorial ‘minmax’ curve (Berger, pers. comm.), eccentricity detrended ‘minmax’ (grey line) and $La_{04_{(1,1)}}$ 65°N summer insolation (black line). b) CLEAN spectrum of the eccentricity detrended ‘minmax’ curve at the equator. The most important periodicities are indicated in the graph.
summer insolation with the same amplitude but rectified and with variability at half-precession clearly visible. The resulting power spectrum is dominated by precession and semi-precession frequencies and clearly not significant additional peaks which

Figure 6.14: 'Minmax' curve (upper curve) at 5°N, eccentricity detrended 'minmax' (grey line) and La04s(1,1) 65°N summer insolation (black line). b) CLEAN spectrum of the eccentricity detrended 'minmax' curve at 5°N (lower graph on a linear, upper graph on a logarithmic scale). The most important periodicities are indicated in the graph.
correspond to periods of 7-6.7 kyr and 3.6-3.3 kyr. The latter are very similar to those found in our data, and suggest that the 7-6.7 kyr and ~3.5 kyr might be related to harmonics and combination tones of primary precession introduced in a nonlinear way.

Although the ‘N5 minmax’ curve could provide an illustration of how nonlinearities are introduced in the (low-latitude) climate system, it remains unclear whether such a mechanism could affect climate physically. Preliminary climate modelling results using a coupled model of intermediate complexity (CLIMBER-2) reveal changes in thermohaline overturning at a quarter-precession cycle of 5-6 kyr if an interactive vegetation cover and feedbacks are included [Tuenter 2004]. The underlying mechanism is not yet clear but seems to include vegetation controlled changes in low-latitude runoff into the low latitude Atlantic and resultant changes in sea surface water salinity, which in turn controls deep water formation in the North Atlantic convection areas.

The importance of vegetation feedbacks could explain why continental records, such as the pollen record from Florida over the last 50 kyr [Grimm et al., 1993] and colour reflectance records of Pliocene lacustrine successions in Greece [Steenbrink et al., 2003], often portray a more regular pacing of sub-Milankovitch climate variability around periodicities of 5-6 kyr than marine records. The example of the N5 minmax could present one way to explain why Northern Hemisphere Heinrich events do not occur regularly at a quarter precession cycle but have a the longer quasi periodic pacing of ~7 kyr.

6.4.6. Sub-Milankovitch climate variability in relation to nonlinearities in the ice sheet-ocean system

Another way to explain differences between continental and marine records is by taking nonlinearities within the ice-sheet ocean system and (nonlinear) interactions with this system into account. Different temporal responses of the various climate components to orbital forcing, diverse feedback mechanisms as well as internal oscillations indicate that the ice-sheet ocean system is highly complex.

The response times for the different climate components during MIS100-96 were evaluated in Chapter 5. The response times of the $\delta^{18}O_{benthos}$ of Site 967 and its derivative ice volume show a large lag with respect to orbital obliquity while Northern Hemisphere annual air temperature ($T_{air}$) is leading ice volume by only a few kyr. More importantly, it was shown that response times were not constant through time for all the components and may have been altered in the presence of sum and difference frequencies to the main orbital forcing. By contrast, the Ti/Al record of Site 967 does show a constant in phase relationship with orbital obliquity, which was the basic assumption in creating the age model of Site 967 [Lourens et al., 2001]. Similarly, the 41-kyr components of the MS and CR indicate constant phase leads/lags with respect to obliquity throughout MIS101-95 indicating that these climate systems were probably independent of (nonlinear) ice-sheet (ocean) dynamics.
Additionally, spectral results indicate a more regular pacing of MS and CR variations on a sub-Milankovitch scale as compared to the δ¹⁸O. Especially, the δ¹⁸O_benthos of SN reveals a broad range of spectral peaks, which might, however, be related to the same climate signal. Wavelet analysis and band-pass filtering of the SN δ¹⁸O_benthos record indicate that the spacing between sub-cycles (stadial-interstadial variations) varies through time pointing to a non-stationary behaviour. This suggests that either the time series on which the climate system acts has varied through time or that the primary signal is constant but has been frequency-modulated.

Evaluation of the δ¹⁸O_benthos of SN in the time domain indicates that the spacing of the subcycles might be controlled by the duration of the glacial, with longer spacing being associated with a longer duration of the glacial.

Frequency-modulation (FM) of the obliquity (and eccentricity) cycle caused by the ~413 kyr eccentricity has been discussed in Chapter 5. It was argued that the strong peaks at the 28 and 80 kyr observed in the δ¹⁸O_benthos of Site 967 and its derivatives, ice volume and T_air, may represent sidebands of a 41 kyr carrier (Chapter 5), although these periodicities are slightly offset from those predicted in theory [Rial and Anacleto

Figure 6.15: CLEAN spectrum and wavelet map of T_air Site 967 versus period (kyr) and T_air of San Nicola versus age. Shaded intervals in the CLEAN spectrum indicate significant periodicities. The thick stipple line indicates the modulation.
2000]. But how does this observation relate to the observed variations in cycle length of the $\delta^{18}O_{\text{benthos}}$ sub-cycles of SN? Could indeed frequency-modulation alter the cycle duration?

The Northern Hemisphere surface air temperature ($T_{\text{air}}$), which has been derived from the $\delta^{18}O_{\text{benthos}}$ of Site 967, reveals high-frequency variability superimposed on the obliquity dominated glacial interglacial variation like the $\delta^{18}O_{\text{benthos}}$ of Site 967 itself (Chapter 5). In the spectrum of $T_{\text{air}}$, peaks are observed at or near periodicities of 7, 9, 13 and 16 kyr, but the power distribution is non-stationary through time (Figure 6.15). Maximum concentration of power at a periodicity of ~9 kyr occurs during late MIS100 - early MIS99 and MIS96 while concentration of power at or near 13 and 16 kyr and around 7 kyr (or 4-5 in case of MIS101) is observed during MIS101, 98 and 95. Bifurcations of the signal from a strong signal around 9 kyr into two signals centred around a longer (13-16 kyr) and a shorter (7 kyr) periodicity occurs every 70 kyr (dotted line in Figure 6.15). This example shows that nonlinear behaviour within the climate system, cannot be excluded. However, higher order statistics (i.e. bispectral analysis) are required to substantiate such a mechanism by determining phase and frequency locking of the components. Unfortunately our record is too short to adequately resolve variance in the Milankovitch range. Preliminary results of bicoherence analysis (not shown) of the global benthic stacked record of Lisiecki and Raymo [2005] shows evidence for phase coupling as would be expected if frequency-modulation of the obliquity signal plays a role. However, these preliminary results also show that precession might have an influence although it should be realised that the stacked record has been tuned to precession and obliquity.

Getting back to our data, a ~8-9 kyr periodicity seems to play an important role in the $\delta^{18}O_{\text{benthos}}$ records of SN and Site 967, and in the derivative $T_{\text{air}}$ record. This periodicity brings us back to the model proposed by McIntyre and Molfino [1996], who stated that the 8.4 kyr periodicity observed in the zonal wind-driven divergence in eastern tropical Atlantic over the past 40 kyr represents the 2$^{nd}$ harmonic of 16.4 kyr, which is the difference tone of a short eccentricity cycle (of 72 kyr) and primary precession (22 kyr). The anomalous short eccentricity cycle should then be related to a minimum in the 400 kyr eccentricity. Berger and Loutre [1997] showed that a set of anomalous short (80-76 kyr) eccentricity cycles may produce a range of periodicities between 9.3 and 8.1 kyr. Alternatively, the 8.4 kyr cycle can be related to the difference tone between the 2$^{nd}$ harmonic of 28 kyr and the 80 kyr cycle ($2/28$ kyr-1/80 kyr = $1/16.9$ kyr) and thus probably to (internal) ice sheet oscillations (Chapter 5).

### 6.5. Conclusions

Oxygen isotope and sediment property data of San Nicola and Site 967 record climate relationships during MIS98 and MIS96 that are comparable to those observed during MIS100 (Chapters 2, 3 and 4), i.e. low latitude dust episodes as recorded by the colour...
reflectance (CR) and magnetic susceptibility (MS) can be linked with cooling cycles in the Mediterranean and the North Atlantic, and with North Atlantic ice rafting history. Spectral analysis indicates however, that these climate systems might have operated on slightly different time scales: On a Milankovitch-scale, spectra of the oxygen isotope records are dominated by 80, 41 and 28 kyr periodicities, which is in agreement with global ice volume and Northern Hemisphere annual air temperature in that interval (Chapter 5). Only the spectra of δ¹⁸O_Grubbe indicate variance at the precession period, reflecting a local sea surface salinity/temperature component. The CR and MS spectra indicate a much stronger precession component but they differ significantly from one site to the other. These differences may be explained by the shortness of the records and, probably more importantly, differences in the glacial-interglacial signature of the individual sites and the strength of the precessional signal. Spectra of these records are more similar in the sub-Milankovitch range, showing concentration of variance around 6-7 kyr close to periodicities resulting from harmonics and combinations tones of the primary precession components. Comparison with the power spectrum of a curve constructed by taking the difference between the strongest temperature maximum (T_{max}) and the strongest minimum (T_{min}) in the annual cycle at the equator indicates that differential heating related to the orbital configuration may explain the observed climate variability recorded by the MS and CR at SN and Site 967. Such a model would be in agreement with a (stationary) periodic climate signal as observed in the sediment property records of SN and Site 967. However, a link with primary precession may be considered less likely in view of the reduced precessional amplitude as a consequence of minimum eccentricity at that time.

On the contrary, spacing of δ¹⁸O subcycles, which represent the most prominent feature on a sub-Milankovitch scale in the δ¹⁸O record of the investigated time interval (see also Chapter 2), vary from one glacial to the other between 7 and 9 kyr, with the length of the cycles being related to the length of the glacial. Wavelet analysis indicates that both intensities and frequencies of the climate signal change through time, explaining the relatively broad peak in the power spectrum around 7-9 kyr. Such a non-stationary behaviour points towards nonlinear components either in the recording mechanism, the climate response or the forcing itself. Such nonlinear components could arise from frequency modulation of the primary eccentricity, obliquity and precession cycle or from internal ice sheet dynamics. An example of frequency modulation is presented for the Northern Hemisphere annual air temperature (T_{air}), a derivative of the δ¹⁸O_benthos of Site 967, which shows the presence of a 8 kyr or ~12 kyr component being modulated by a ~70 kyr periodicity. This example is more in agreement with the model proposed by McIntyre and Molfino [1996], who related zonal wind-driven divergence in the equatorial Atlantic with a periodicity of ~8.4 kyr to nonlinear interaction between precession and an anomalous short eccentricity periodicity. However, our data also clearly indicate that more work has to be done to unravel the enigmas of sub-Milankovitch variability in the climate system.