

Eolian Evidence for Spatial Variability of Late Quaternary Climates in Tropical Africa¹

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Study of the eolian fraction of late Quaternary sediments from the tropical Atlantic reveals that two modes of long-term climate variability have existed in tropical Africa during the last 150,000 yr. Tropical northwest Africa (i.e., the southwestern Sahara and Sahel) was driest during glaciations and stadets, but wetter than at present during interglaciations and interstadets. This may be a response to ice sheets at higher latitudes, via equatorward displacement of the westerlies and the subtropical high. In contrast, central equatorial Africa (southeast of the Sahara) was most arid during interstadets and times of ice growth, and most humid during deglaciation. Wet periods in this area correspond to insolation maxima in northern hemisphere summer. A 23,000-yr precessional rhythm is suggested, supporting a direct link between African Monsoon intensity and orbitally modulated insolation. The late Holocene is the only time observed when both areas are arid during an interglacial episode. This may reflect, in part, anthropogenic disturbance of late Holocene climates. © 1985 University of Washington.

INTRODUCTION

Many climatic proxy records from tropical Africa indicate conditions more arid than today during the last glacial maximum, and more humid than today during the early Holocene. Because most records available from land-based sections are relatively short and difficult to date, evaluation of long-term causes of climate change (such as "Milankovitch" orbital forcing) has proved difficult. Unlike continental records, deep-sea sediments can be dated and correlated directly to ice volume fluctuations throughout the Pleistocene using oxygen-isotope stratigraphy. In pelagic sediments off northwest Africa, windblown freshwater diatoms (Diester-Haass *et al.*, 1973; Parmenter and Folger, 1974) and quartz silt (Kolla *et al.*, 1979; Sarnthein *et al.*, 1981) are most abun-

dant during glacial episodes. Parmenter and Folger (1974) ascribed the increased eolian silica in pelagic sediments to glacial aridity in the tropics.

The purpose of this paper is to monitor spatial variability of tropical African climate changes using eolian biosiliceous material in deep-sea sediments. We examine the eolian record of freshwater diatoms of the genus *Melosira* and opal phytoliths from the tropical Atlantic through the last 150,000 yr. *Melosira* in eolian sediments come from deflation of diatomaceous deposits in dry lake beds. Thus, they are an indirect indicator of lake levels, and therefore of aridity. Opal phytoliths are silica bodies in the leaves of vascular plants (abundant in grasses; Twiss *et al.*, 1969) that are injected into the atmosphere during dry-season brush fires. Their presence in deep-sea sediments off Africa reflects the position of the tall-grass savannah, and wind directions.

MODERN CLIMATOLOGY

The climate of tropical Africa ranges from humid (Gulf of Guinea coast) to very

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arid (the Sahara). Rainfall is highly seasonal, occurring in Northern Hemisphere winter north of the Sahel, and in Northern Hemisphere summer to the south (Thompson, 1965). An anti-Hadley Cell cross circulation associated with the Tropical Easterly Jet enhances subsidence over North Africa and restricts summer rains of the African Monsoon to south of 14°N (Flohn, 1966; Flohn and Nicholson, 1980). Vegetation belts reflect the moisture gradient (Agwu and Beug, 1982). Evergreen rain forests occur near the coast of the Gulf of Guinea and in the Zaire basin. To the north, rain forests are succeeded by the Sudano-Guinean deciduous tropical forests and tall-grass savannah, the Sahelian short-grass savannah, and finally the Sahara, which harbors little vegetation.

The seasonality of rainfall results in two seasonal dust plumes (Fig. 1). During northern hemisphere summer, the "Dunkelmeer," or "Dark Sea" plume occurs off northwest Africa at about 10°–25°N. Dust in the summer plume comes mainly from the western and central Sahara and its borders (Kalu, 1979; Sarnthein *et al.*, 1981). In northern winter, the "Harmattan Haze" occurs from about 15°N to 5°S (Kalu, 1979; Prospero, 1981; our Fig. 1). Dust in the winter plume originates mainly along the southeast edge of the Sahara (Kalu, 1979; Prospero, 1981). Because the literature is inconsistent in the use of the terms "Harmattan" and "Dunkelmeer" (e.g., Nicholson, 1982; Tetzlaff *et al.*, 1982), we refer to the two plumes by the Northern Hemisphere seasons in which they occur. Westward transport of dust in the atmosphere occurs primarily in middle altitude (1.5- to 4-km) easterly winds (Tetzlaff and Wolter, 1980).

RESULTS

Sample preparation methods are described fully in Pokras (1984). Briefly, a measured weight (about 0.2 g) of dried sediment is treated with HCl and H₂O₂, to remove carbonates and organic matter. The

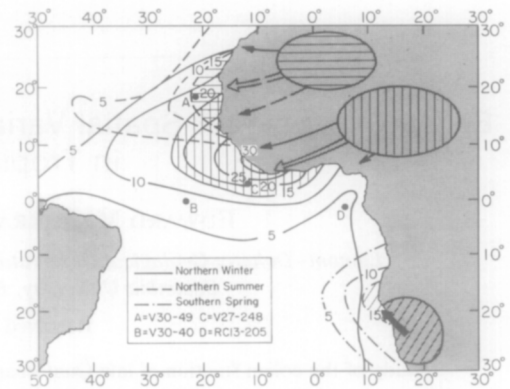


FIG. 1. Approximate source areas for dust in the summer (northern) and winter (southern) plumes, and the southwest African plume of southern spring (Kalu, 1979; Pêwé, 1981). Arrows indicate typical trajectories of dust toward the tropical Atlantic. Contours represent percentage of shipboard observations that reported atmospheric haze (after Prospero, 1981). Also shown are locations of deep-sea cores in which 150,000-yr time series were generated. Core V30-49 is under the summer plume, V30-40, and RC13-205 lie under the winter plume.

treated sediment is suspended in a known volume of water and an aliquot is vacuum filtered onto a 0.8- μ m Nuclepore filter membrane. A portion of the filter is mounted on a glass slide using Carmount 165 mounting medium. A number of fields of view are examined (between 25 and 200, depending on the abundance of material). The fraction of the original sample examined is calculated, and raw counts are converted to numbers per gram of dry sediment. As most *Melosira* valves in deep-sea sediments are fragmented, whole valves and fragments are combined. Precision ($\pm 1\sigma$) is 8% when abundances exceed 5×10^6 /g, and 20% when abundances are between 1×10^5 and 5×10^6 /g.

Distributions of *Melosira* and phytolith abundances were determined for (1) modern samples from atmospheric dust and core-top sediments, (2) the last glacial maximum, 18,000 yr B.P., and (3) time series of the last 150,000 yr. All records are dominated by *Melosira granulata*, a widely distributed species adapted to shallow, turbid lakes and the littoral zones of larger lakes

(Gasse *et al.*, 1983). Stratigraphic control and a direct link to ice volume are provided by $\delta^{18}\text{O}$ analyses of foraminifera and radiocarbon dates of bulk carbonate (Mix and Fairbanks, in press; A. C. Mix, unpublished data).

Abundance counts are not transformed to mass accumulation rates because sedimentation rates and bulk densities are not known precisely for many of the cores examined for the time slices. Presentation of per gram data may thus cause some inaccuracies in the map patterns. Except where noted later in V27-248, this is unlikely to lead to substantial errors in the time series, as abundance changes in both *Melosira* and phytoliths (typically over an order of magnitude) far exceed temporal variations in sedimentation rate in the cores we have used.

Modern Distributions

Samples of airborne dust collected during Cruise RC24-03 of R/V *Robert D. Conrad* in March 1983 show that *Melosira* constitute up to 38% of the $>4\ \mu\text{m}$ eolian fraction in the equatorial Atlantic (Table 1; Fig. 2A). Phytoliths are much less common, especially in samples farther from Africa. *Melosira* increase in relative abundance seaward. Although phytoliths and whole *Melosira* are similar in size (10–20 μm), the dense phytoliths remain airborne less readily than *Melosira* valves, which are hollow, highly perforate, and commonly fragmented. The numerous marine diatoms in dust sample RC24-03-4 are probably locally derived, due to relatively heavy seas during this deployment of the dust trap.

In core-top sediments, the highest *Melosira* abundances (tabulated in the appendix) occur under the winter plume from 0° – 10°N (Fig. 2A). A second *Melosira* maximum off the African coast near 20°S (Fig. 2A) reflects dust transport from the margins of the Namib/Kalahari Desert system, primarily in southern hemisphere spring (Prospero, 1981). Although there is no distinct maximum of *Melosira* abundance at 15° –

TABLE 1. COMPOSITION OF EOLIAN DUST IN THE EQUATORIAL ATLANTIC, CRUISE RC24-03, MARCH 1983; $>4\text{-}\mu\text{m}$ FRACTION

Station	Dates	Location		Duration (hr)	Lithogenic silt (%)	% <i>Melosira</i>	% Other freshwater diatoms	% Opal phytoliths	Marine/nonmarine diatoms
		Starting	Ending						
1	3/15–3/16	$4^\circ 08' \text{N } 6^\circ 11' \text{W}$	$3^\circ 32' \text{N } 7^\circ 44' \text{W}$	13	67.6	25.6	5.6	1.0	0.02
2	3/16–3/17	$3^\circ 23' \text{N } 7^\circ 45' \text{W}$	$2^\circ 08' \text{N } 10^\circ 38' \text{W}$	22.5	71.1	26.1	2.2	0.6	0.03
4	3/22–3/23	$1^\circ 15' \text{S } 11^\circ 49' \text{W}$	$1^\circ 53' \text{S } 11^\circ 22' \text{W}$	13	56.7	38.2	4.6	0.5	0.40

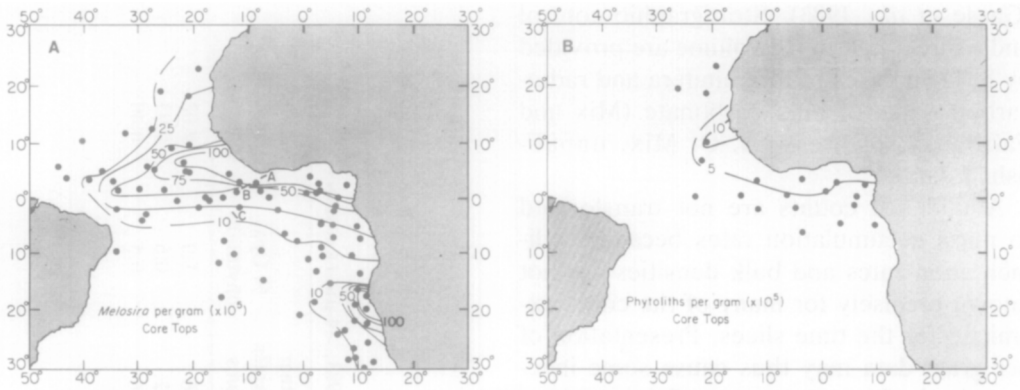


FIG. 2. (A) *Melosira*/g ($\times 10^5$) in core-top sediments. Maximum abundances in the northern Gulf of Guinea and Sierra Leone Basin correspond to maximum dust transport in the winter plume. Because *Melosira* are less common in the summer plume, there is no distinct maximum north of 12°N. A, B, and C represent airborne dust samples 1, 2, and 4 taken during R/V *Robert D. Conrad* Cruise RC24-03, March 1983. Dashed lines indicate inferred contours. (B) Opal phytoliths/g ($\times 10^5$) in core-top sediments. Highest phytolith abundances occur west of the tall-grass savannah belt of Africa. Phytoliths decrease rapidly offshore because they are less easily transported than *Melosira* valves. Dashed line indicates inferred contour.

20°N, substantial concentrations do extend this far north, confirming the presence of *Melosira* in the summer plume as reported by Folger *et al.* (1967). Melia (1984) also observed high numbers of freshwater diatoms in the northern Gulf of Guinea. In addition, he reported high abundances along the Africa coast near 15°N. This is not evident in our data due to the low density of the core tops in that region.

Phytoliths, which are less common in pelagic sediments than *Melosira*, decrease rapidly in abundance seaward (Fig. 2B; Melia, 1984). As with the airborne dust samples, core-top abundances show that phytoliths are less readily transported by the wind than are *Melosira* (Figs. 2A, B). Maximum concentrations of phytoliths (appendix; also Melia, 1984) occur off the tall-grass savannah belt (Agwu and Beug, 1982). Abundance of phytoliths in pelagic sediments thus appears to reflect proximity to grasslands, especially the tall-grass savannah. Long-term changes in rainfall should control the position of the grass belt, which should move northward (toward the Sahara) during wet times, and southward during dry times. The reflection of the grass-belt position by phytoliths in deep-

sea sediments may also be affected by wind directions.

Last Glacial Maximum

Maximum numbers of *Melosira* at 18,000 yr B.P. occur near the modern maximum at 6°–8°N (Fig. 3A). *Melosira* abundances are essentially the same as today's north of 8°N, but are consistently lower than at present in the Gulf of Guinea. Phytoliths (Fig. 3B) are more abundant at 18,000 yr B.P. in all but the three northernmost cores. The phytolith increase is pronounced throughout the equatorial region, although the maximum remains near 10°N. This is essentially the opposite of the change in *Melosira* abundance.

Time Series

Piston cores for time-series study sample both the summer and winter plumes (core locations in Fig. 1). Core V30-49 (18°26'N, 21°05'W, 3093 m depth) is from near the center of the summer plume. Cores V30-40 (0°12'S, 23°09'W, 3706 m), RC13-205 (2°17'S, 5°11'E, 3731 m), and V27-248 (3°03'N, 11°49'W, 4543 m) are under the winter plume.

RC13-205 has an existing $\delta^{18}\text{O}$ stratig-

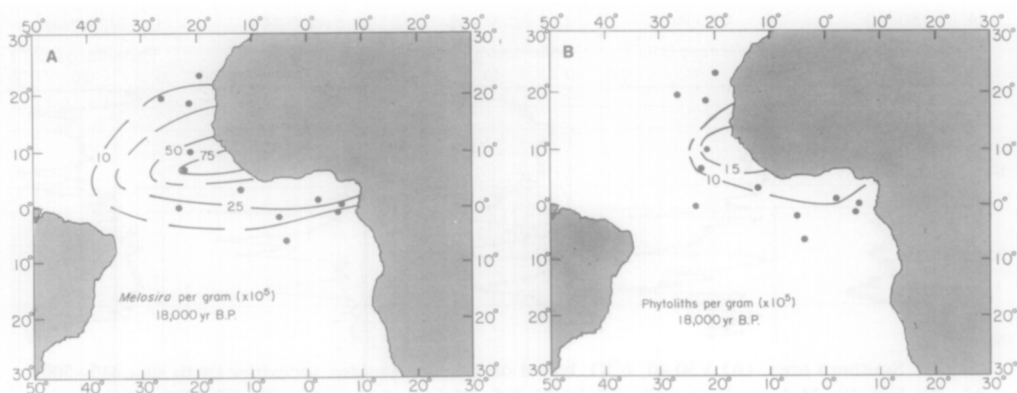


FIG. 3. (A) *Melosira/g* ($\times 10^5$) at the last glacial maximum, 18,000 yr B.P. Abundances in the Gulf of Guinea are lower than at present (Fig. 2A). Abundances north of 10°N are essentially the same as those in core tops. Transport in the summer plume was thus at approximately modern levels. Dashed lines indicate inferred contours. (B) Phytoliths/g ($\times 10^5$) at the last glacial maximum, 18,000 yr B.P. The maximum remains near the modern maximum (Fig. 2B). Abundances are higher than modern in the Gulf of Guinea, suggesting a southward expansion of the grass belt during glacial time. Dashed lines indicate inferred contours.

raphy (Shackleton's data in Morley, 1977). New oxygen-isotope data are from *Cibicides wuellerstorfi* in V30-49 (Fig. 4), and from *Globigerinoides sacculifer* (with sac, 415–500 μm) in V30-40 (Fig. 5). Although low carbonate content in V27-248 precludes isotope analyses, fluctuations of the percentage of CaCO_3 in this core allow correlation to similar records in the other cores. We illustrate the data versus age, using the SPECMAP chronology (Imbrie *et al.*, 1984).

In the northern area, *Melosira* are most abundant in glacial and stadial stages 2, 4, 5b, 5d, and 6, and the upper Holocene (Fig. 4). The upper Holocene is anomalous, as it contains the only interglacial or interstadial *Melosira* maximum at this site within the last 150,000 yr. Phytoliths are most common at times of relatively low ice volume in the middle Holocene, stage 3, and substages 5a and 5c. The correlation to ice volume is not perfect, however, as no significant phytolith maximum occurs in substage 5e.

Time series from the southern area are quite different from that in the northern area. *Melosira* in RC13-205, V27-248, and V30-40 are most abundant just after the ice-

volume minima at substages 5e, 5c, and 5a, and late in stage 1 (Fig. 5). In V30-40, major *Melosira* peaks in stage 5 occur at 118,000 yr B.P. (substage 5e/5d boundary), 97,000 yr B.P. (5c/5b boundary), and 76,000 yr B.P. (5a/4 boundary), all ± 4000 yr. They lag the isotopic (\approx ice volume) minima by 4000 ± 2000 yr. A *Melosira* maximum at the core tops is clearly present in RC13-205 and V27-248, but is weak in V30-40. Thus, the modern event is probably not as severe as

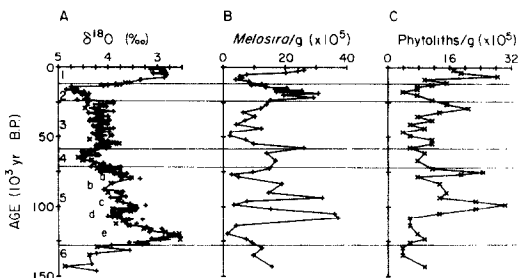


FIG. 4. Northern area (V30-49). (A) $\delta^{18}\text{O}$, based on *Cibicides wuellerstorfi*. (B) *Melosira/g* ($\times 10^5$). *Melosira* maxima indicate arid times at glacial maxima, when lakes desiccated and deflated. Abundant *Melosira* in the late Holocene are unique in the last 150,000 yr, reflecting anomalous aridity for this area in full interglacial conditions. (C) Opal phytoliths/g ($\times 10^5$). Phytolith maxima indicate northward movement of the grass belt during interglacial and interstadial maxima, possibly except for substage 5e.

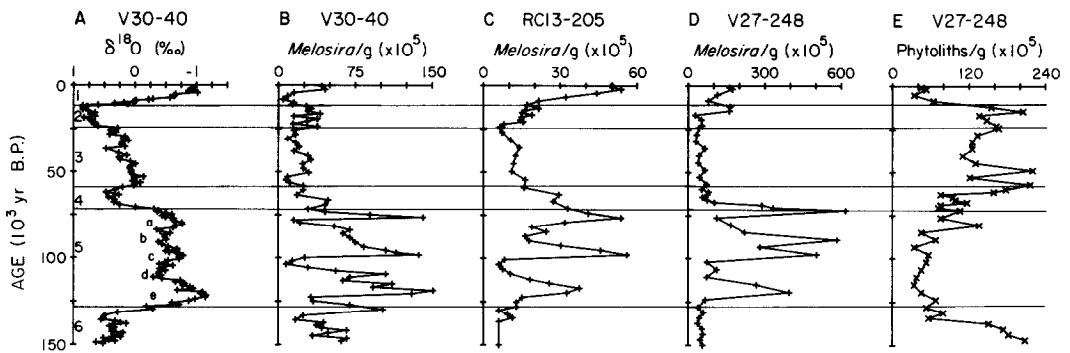


FIG. 5. Southern area. (A) V30-40, $\delta^{18}\text{O}$, based on *Globigerinoides sacculifer* (with sac, 415–500 μm). (B) V30-40, *Melosira/g* ($\times 10^5$). (C) RC13-205, *Melosira/g* ($\times 10^5$). (D) V27-248, *Melosira/g* ($\times 10^5$). *Melosira* maxima in these records reflect increased aridity in central equatorial Africa. In contrast to the northern area, these maxima occur during ice-growth phases at the stage 5e/5d, 5c/5b, 5a/4 boundaries, and in the latest Holocene. (E) V27-248, Opal phytoliths/g ($\times 10^5$). Here, phytolith maxima indicate glacial aridity and movement of the grass belt southward, replacing woodlands in tropical West Africa.

those in stage 5. Smaller *Melosira* maxima occur in all three cores in glacial stage 2, and in stage 6 in V30-40. The stage 3 interstade shows no strong or reproducible fluctuations in *Melosira* records from the southern area.

Phytolith abundances in the southern area were determined in V27-248, the core closest to the dust source (Fig. 5). Here, phytoliths are most common in stage 2, in lower stage 3 and/or upper stage 4, and in stage 6. Abundances are moderate in stage 4 and substage 5b, and lowest in the mid-Holocene and substage 5e interglaciations.

DISCUSSION

Transport and Preservation

Kolbe (1957) speculated that freshwater diatoms such as *Melosira* are transported to the deep sea by fluvial rather than eolian processes. The particulate load of rivers is deposited rapidly upon entering the ocean as hemipelagic sediments on continental rises and turbidites on abyssal plains, but not as pelagic sediments away from the continents (e.g., Damuth, 1977). River outflow is unlikely to transport freshwater diatoms to the site of V30-40, near the crest of the Mid-Atlantic Ridge 2500 km from land. The similarity of *Melosira* time series in the three equatorial cores (Fig. 5) argues for a single transport mechanism to all three

sites, because it is highly unlikely that fluvial transport or ocean-floor sediment redistribution would lead to such similar signals in such widely separated sites. Furthermore, sampling of airborne dust (Delaney *et al.*, 1967; Melia, 1984; our results) clearly shows that *Melosira* are readily transported by winds. We thus suggest that *Melosira* abundances in open-ocean pelagic sediments are primarily a function of eolian transport.

Because both whole and fragmented *Melosira* valves are counted, dissolution or mechanical breakage could lead to a false increase in abundance. If breakage were the sole cause of high abundances, whole valves should be more abundant in samples with fewer fragments, yielding a negative correlation. The correlations between whole valves and fragments are, however, positive: $r = 0.36$ ($n = 50$) for RC13-205, $r = 0.39$ ($n = 50$) for V27-248, and $r = 0.58$ ($n = 58$) for V30-49. The different responses of *Melosira* and phytolith abundances through time (Figs. 4 and 5) also argue against substantial control by dissolution, which should affect both indicators in a core in the same sense.

Climatic Significance

Comparison of the modern and 18,000 yr B.P. map patterns for phytoliths (Figs. 2B

and 3B) shows that the maximum concentration is near 10°N at both times. This does not necessarily mean that the grass belt has remained in its present location despite marked changes in continental climate. Late Holocene climate of northwest Africa is more similar to the glacial (dry) pattern than the interglacial (wet) pattern (Street and Grove, 1979; Sarnthein and Koopmann, 1980). The decrease of phytoliths in V30-49 in the latest Holocene may represent a recent southward movement of the northern edge of the grass belt. No corresponding increase in phytolith abundance is seen in V27-248, suggesting that the southern margin of the grass belt did not migrate significantly in recent times. Through the last 150,000 yr, inverse correlation of phytoliths from the northern and southern areas (Figs. 4 and 5) indicates that both margins of the grass belt probably move southward in the relatively arid glacial times and northward in the relatively humid interglacial times. An alternative (but not mutually exclusive) explanation could be that winds were more northerly in glacial times, and easterly in interglacial times.

The phytolith peak in stages 3 and 4 of core V27-248, rather than just in stage 4, may reflect dilution due to high sedimentation rates in stage 4 of V27-248. In V30-49, however, the *Melosira* maximum and the phytolith minimum also occur near the stage 4/3 boundary, suggesting that peak aridity in tropical northwest Africa may have occurred at that time, rather than during the maximum stage 4 glaciation.

Interpretation of the *Melosira* data is somewhat different than that of the phytoliths due to the long-distance transport possible for *Melosira*. In V30-49, increased *Melosira* abundance in glacial time (Fig. 4) reflects glacial aridity and desiccation of lakes bordering the western Sahara. High aridity is indicated by *Melosira* between 24,000 and 13,000 yr B.P., decreasing to a minimum near 9000 yr B.P. This is consistent with the phytolith data, with records of low glacial-age lake levels in northwest

Africa (Maley, 1977; Giresse, 1978; Street and Grove, 1979; Servant and Servant-Vildary, 1980; Nicholson and Flohn, 1980; Street-Perrott and Harrison, 1984), and with other data from marine sediments (Parmenter and Folger, 1974; Sarnthein, 1978; Kolla *et al.*, 1979; Rossignol-Strick and Duzer, 1979; Agwu and Beug, 1982; Sarnthein *et al.*, 1982).

In agreement with past studies, our data indicate that prior to the late Holocene, aridity in northwest Africa is linked to times of high ice volume (Fig. 6). Three mechanisms could explain this linkage. First, global climate changes associated with expanded high-latitude ice cover could yield tropical cooling and aridity in Africa. Second, the topographic effects of the Northern Hemisphere ice sheets (Nicholson and Flohn, 1980) and the southern position of the North Atlantic Polar Front (Ruddiman and McIntyre, 1976) would probably displace the westerlies and the subtropical high southward. This would enhance upper-level convergence and subsi-

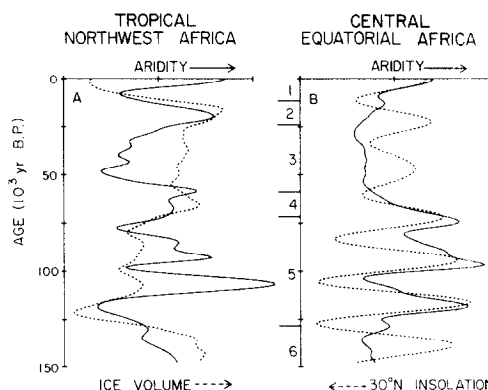


FIG. 6. Summary of climate changes in tropical Africa indicated by the eolian record. (A) Aridity in tropical northwest Africa (smoothed *Melosira* record from V30-49, solid) superimposed on a smoothed ice-volume curve (average of $\delta^{18}\text{O}$ data in V30-49 and V30-40, dashed). Except for the late Holocene, aridity maxima generally occur during glacial stades. (B) Aridity in central equatorial Africa (smoothed average of southern-area *Melosira* curves, solid) superimposed on a 30°N May–July insolation curve (dashed). Aridity generally occurs during summer insolation minima. Reduced-amplitude fluctuations during glaciations suggest that the response to insolation is nonlinear, or that ice volume contributes to the climatic forcing (see discussion in text).

dence over northwest Africa, and thus increase the areal extent and the intensity of aridity in northwest Africa (Newell and Kidson, 1984). Third, glacial aridity could be a relatively local response to colder sea-surface temperatures along West Africa, caused by enhanced coastal upwelling and equatorward advection of cold water in the Canary Current (Thiede, 1977). We favor the second hypothesis, because spatial variability of the climate response within Africa argues against a single hemispheric or global signal. The possible effects of cold sea-surface temperatures off northwest Africa, however, cannot be ruled out. Modeling studies suggest that tropical atmospheric circulation may be sensitive to changes in sea-surface temperature (e.g., Voice and Hunt, 1984).

In the southern cores (Fig. 5), the *Melosira* records indicate aridity during ice-growth phases after interglacial/interstadial maxima of substages 5e, 5c, and 5a, and during the late Holocene. Relatively humid conditions are indicated by low *Melosira* abundances during deglaciation events at the stage 6/5e, 5d/5c, 5b/5a, and 2/1 boundaries, and during much of stages 3 and 4. The minor *Melosira* peaks during stage 2 in V30-40 and V27-248, and during stage 6 in V30-40, appear synchronous with *Melosira* maxima in V30-49. This may indicate that the influence of the summer plume extended farther southward (that is, that winds were more from the north than the east), or that aridity associated with extremely high ice volume extended into the source area for the winter plume at those times.

Other data corroborate our inference of relatively wet conditions in equatorial Africa during deglacial events and dry conditions during global ice growth. High outflow is suggested from the Niger River during the stage 2/1 transition (Pastouret *et al.*, 1978) and from the Zaire River during the stage 6/5 transition (Zachariasse *et al.*, in press). Lake levels near the equator were

high from 12,500 to 7500 yr B.P., while lakes farther north reached high levels only after 10,000 yr B.P. (Talbot and Hall, 1981; Street-Perrott and Roberts, 1983). Rossignol-Strick (1983) argues that sapropels in the eastern Mediterranean resulted from Nile outflow enhanced by monsoonal rains during deglaciation 11,000, 82,000, 106,000, and 127,000 yr B.P.

Melosira minima from the southern cores occur at 11,000, 85,000, 104,000, and 129,000 yr B.P. (± 4000 yr), near summer insolation maxima (Berger, 1978) at 11,000, 82,000, 104,000, and 127,000 yr B.P. (Fig. 6). Maxima occur at 76,000, 97,000, and 118,000 yr B.P. (± 4000 yr), near times of summer insolation minima at 71,000, 93,000, and 115,000 yr B.P. (Fig. 6). This may suggest that the 23,000-yr precession cycle that dominates insolation in this area may also be present in tropical climates on land, with maximum response at times when precessional amplitude is high (Berger, 1978). A possible mechanism is an enhanced African Monsoon during episodes of maximum summer insolation in the Northern Hemisphere (Kutzbach, 1981; Rossignol-Strick, 1983).

The asymmetric "sawtooth" nature of the *Melosira* records, with rapid rises and more gradual declines, and the minimal fluctuations of the southern-area signal in glacial episodes, however, suggest that the relationship between insolation and the deep-sea data is nonlinear. It is not clear whether this nonlinearity lies in the climate system, or in the recording of climate by *Melosira*. The sawtooth pattern may reflect (1) rapid deflation of *Melosira* immediately after lake dessication when the area of exposed diatomaceous deposits is highest, or (2) bioturbation in the deep-sea sediments, which leads to asymmetrical downward translation of sharp abundance peaks (Berger and Heath, 1968). If either of these effects are important, the slight (but not statistically significant) lead of *Melosira* peaks in the southern area before Northern Hemi-

sphere summer insolation minima (Fig. 6) may be an artifact.

The lack of strong *Melosira* variations in the southern area during stages 2, 3, and 4 may reflect partial forcing by ice volume such that the source area is less sensitive to aridity when ice volume is high. If this is true, the effect of high-latitude ice volume on central equatorial Africa is to suppress aridity. This is the exact opposite of the effect observed in V30-49 for tropical northwest Africa. It is consistent with our preferred mechanism for glacial aridification of northwest Africa by southward displacement of the subtropical high. Alternatively, the availability of *Melosira* may be greatly reduced for only slightly reduced lake-level fluctuations associated with lower amplitude insolation variations.

The late Holocene is the only time in the last 150,000 yr when arid conditions occur in both tropical northwest Africa and central equatorial Africa during interglacial times. Our data are consistent with those of Sarnthein and Koopmann (1980), which show a late Holocene increase of eolian quartz silt in deep-sea sediments, and with many studies on the continent that indicate lower lake levels at present than in the middle Holocene. The late Holocene increase in *Melosira* in V30-49 could conceivably result from increased wind intensity (Sarnthein *et al.*, 1982), but the corresponding decrease in phytolith abundances can not. A change in African aridity is required.

Decreasing summer insolation over the last 6000 yr may have decreased the strength of the African Monsoon, resulting in late Holocene aridity throughout tropical Africa (Kutzbach, 1981). While this scenario may hold for central equatorial Africa, it seems unlikely to explain aridity in tropical northwest Africa, because *Melosira* in core V30-49 has no such response to larger insolation changes earlier in the record. Instead, late Holocene aridity may in part reflect anthropogenic disturbance of

climate during the last few thousand years. In historical times, nomadic herdsman have caused significant deforestation and soil degradation south of the Sahara by cutting trees to make charcoal and burning brush to clear grazing lands (National Research Council, 1983). This practice may have started as early as several thousand years ago. Fire was used by early man for game drives (Sagan *et al.*, 1979). Domestic herd animals were widespread after 6000 yr B.P. (Harlan, 1982). Although there is some disagreement as to when agriculture began in tropical west Africa, cultivation of grains and other crops may have occurred prior to 3000 yr B.P. (Phillipson, 1982; Flight, 1976).

Charney *et al.* (1975) suggested that "biogeophysical feedback" could amplify the climatic effects of recent anthropogenic disturbances in this area. We suggest that this mechanism could have amplified late Holocene aridification begun by insolation change and anthropogenic disturbance. Nicholson and Flohn (1980) also speculated that anthropogenic aridification may have begun as much as 5000–8000 yr ago. If this is true, the climatic result of man's presence is striking in its amplitude, which in V30-49 appears as large as the glacial/interglacial variations. *Melosira* abundance may, however, overestimate the climatic effects of anthropogenic disturbance. If soils formed from dry lake beds were tilled for agriculture, this could disturb crusts that inhibit soil deflation (Talbot, 1980) and increase the susceptibility of *Melosira* valves to eolian transport. The idea of anthropogenic amplification of late Holocene aridity in tropical northwest Africa is speculative. It does, however, account for the fact that interglacial aridity in this region is contrary to the overall pattern of the last 150,000 yr, as revealed by the *Melosira* data in V30-49.

CONCLUSIONS

Our data suggest that two modes of climate variability existed in tropical Africa during the last 150,000 yr. Except in the

latest Holocene, the climate of tropical northwest Africa (e.g., the southwestern Sahara and Sahel) has been most arid during glaciations and stades, and most humid during interglaciations and interstades. The co-occurrence of aridity in this area with high global ice volumes suggests that the climate of tropical northwest Africa is linked to that of the higher latitude ice sheets. Our preferred explanation for this linkage is southward displacement of the westerlies and subtropical high by the topographic effect of the ice sheets and southern position of the North Atlantic Polar Front. Enhanced convergence over northwest Africa and a stronger Tropical Easterly Jet would suppress rainfall in this area.

In central equatorial Africa, maximum aridity has occurred during ice-growth phases following interglacial maxima (substage 5e/5d, 5c/5b, and 5a/4 boundaries) and in the latest Holocene. Maximum humidity occurred during deglaciation (6/5e, 5d/5c, 5b/5a, and 2/1 boundaries). Wet periods occurred consistently at times of high summer insolation. *Melosira* abundance histories, with an apparent 23,000-yr precessional rhythm modulated by eccentricity, support theories that call for direct control of African Monsoon intensity by seasonal contrast of insolation.

The late Holocene is the only time in the last 150,000 yr when both tropical northwest Africa and central equatorial Africa are very arid during a peak interglaciation. Anomalous late Holocene aridity in tropical northwest Africa may reflect climatic feedback forced in part by anthropogenic disturbance. If this is true, it suggests that the tropical climate system is quite sensitive to modifications made by man.

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APPENDIX

Abundances of *Melosira* Valves and Phytoliths in Core-Top Sediments ($\times 10^5$)

Core	Core tops	
	<i>Melosira</i> /g	Phytoliths/g
EN66-10	180.0	10.2
RC13-205	50.4	4.2
V22-193	39.4	10.6
V23-98	22.9	1.5
V26-41	20.3	1.3
V27-238	12.1	1.9
V27-240	9.1	2.3
V27-248	161.0	5.7
V29-144	100.0	1.5
V29-153	66.7	5.3
V30-40	42.4	0.8
V30-49	26.1	1.7
RC13-203	232.0	4.9
RC13-206	12.1	3.8
V12-79	13.6	1.5
V27-230	18.2	2.6
V27-242	36.4	4.9
V30-43	57.6	5.6
V19-293	—	5.7
RC13-183	33.0	
RC13-184	8.3	
RC13-189	75.0	
RC13-190	86.4	
RC13-194	63.6	
RC13-212	1.5	
RC13-214	9.1	
RC13-215	6.1	
RC13-217	15.1	
RC13-222	6.1	
RC13-224	75.7	
RC13-226	133.0	
RC13-227	18.2	
RC13-228	51.5	
RC13-229	45.4	
RC17-37	6.8	
V16-25	3.0	
V16-35	0.1	
V16-204	7.6	
V19-241	18.2	
V19-243	5.3	
V19-250	0.1	
V19-253	12.1	
V19-265	54.5	
V19-276	6.1	
V19-277	12.1	
V19-280	0.1	
V19-283	69.7	
V19-287	30.3	
V20-203	18.2	
V20-204	12.1	
V20-236	1.1	

Core	Core tops	
	Melosira/g	Phytoliths/g
V22-34	14.4	
V22-38	0.8	
V22-168	0.2	
V22-170	0.4	
V22-174	4.5	
V22-177	6.1	
V22-188	9.8	
V25-57	12.9	
V25-60	20.1	
V26-53	45.4	
V26-55	0.8	
V27-177	40.1	
V27-178	65.9	
V27-223	3.0	
V27-224	16.7	
V27-234	239.0	
V27-239	18.2	
V27-249	24.2	
V29-149	159.0	
V29-150	156.0	
V29-156	39.4	
V30-27	19.7	
V30-29	47.7	
V30-36	91.7	
V31-2	75.7	
V31-130	14.4	

Core	Depth (cm)	18,000 yr B.P.	
		Melosira/g	Phytoliths/g
EN66-10	29.5	90.9	13.6
RC13-205	50.0	14.8	8.3
V22-193	50.0	60.6	19.3
V23-98	70.0	8.9	1.1
V26-41	43.5	16.3	1.0
V27-238	55.0	6.1	6.8
V27-240	70.0	42.4	6.8
V27-248	60.0	27.3	14.0
V29-144	75.0	24.2	7.6
V29-153	60.0	33.3	14.4
V30-40	66.0	37.9	1.5
V30-49	80.0	17.8	0.4

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