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Relation between low latitude insolation and $\delta^{18}\text{O}$ change of atmospheric oxygen for the last 200 kyrs, as revealed by Mediterranean sapropels

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Abstract. The isotopic ratio of atmospheric O_2 , $\delta^{18}\text{O}_{\text{atm}}$, deduced from ice cores, displays large fluctuations during climatic cycles. These are caused principally by changes in the oxygen isotopic ratio of sea-water, $\delta^{18}\text{O}_{\text{sw}}$, and changes in the biosphere and in the hydrological cycle. As both $\delta^{18}\text{O}_{\text{atm}}$ and $\delta^{18}\text{O}_{\text{sw}}$ coincide closely over the last 135 kyr BP, it is generally believed that $\delta^{18}\text{O}_{\text{atm}}$ is driven mainly by $\delta^{18}\text{O}_{\text{sw}}$. Here we focus on the major discrepancy, which arises between those two signals during the prior isotopic glacial stage 6, around 175 kyr BP, discrepancy which calls into question the role of changes in $\delta^{18}\text{O}_{\text{sw}}$ as the driving mechanism for $\delta^{18}\text{O}_{\text{atm}}$. We present arguments, based on the occurrence and pollen content of Mediterranean sapropels, for another source of the $\delta^{18}\text{O}_{\text{atm}}$ change: insolation at low latitudes via hydrosphere/biosphere activity.

Introduction

The changes in isotopic ratio of atmospheric oxygen, $\delta^{18}\text{O}_{\text{atm}}$, have been recorded by analysing air trapped in ice-cores [Bender *et al.*, 1985; Horibe *et al.*, 1985; Sowers *et al.*, 1991; Sowers *et al.*, 1993; Jouzel *et al.*, 1993; Bender *et al.*, 1994a]. These are caused principally by changes in the isotopic composition of sea-water, $\delta^{18}\text{O}_{\text{sw}}$, and changes in the biosphere and in the hydrological cycle [Bender *et al.*, 1994b]. Over the last 135 kyr BP, $\delta^{18}\text{O}_{\text{atm}}$ displays variations similar to those of $\delta^{18}\text{O}_{\text{sw}}$ [Sowers *et al.*, 1993; Bender *et al.*, 1994a]. During the Quaternary $\delta^{18}\text{O}_{\text{sw}}$ oscillates in response to the glacial-interglacial climatic oscillations reflecting primarily the volume in continental ice [Shackleton, 1987]. The isotopic ratio of atmospheric oxygen, $\delta^{18}\text{O}_{\text{atm}}$ is related, but not equal, to that of sea water, the difference being the Dole effect. The mechanism of this effect has recently been the subject of detailed investigations by Bender *et al.*, [1994b]. On the basis of the similarity between the two signals over the last 135 kyrs, Sowers *et al.* [1993], and Bender *et al.* [1994b] concluded that the atmospheric change is driven mainly by the sea water change and noted the existence of secondary factors associated with biosphere activity and hydrological cycle. In this scenario, variations in $\delta^{18}\text{O}$ of sea-water are transferred to atmospheric oxygen through the atmosphere-biosphere coupling. This coupling involves the O_2 cycle through biospheric processes (photosynthesis, respiration, decay), with a turnover time for O_2 through the biosphere of the order of 1.2 kyr [Bender *et al.*, 1994b].

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Recently published analyses of the deeper part of the Vostok ice-core [Jouzel *et al.*, 1993; Bender *et al.*, 1994a; Jouzel *et al.*, 1996] provide data that are inconsistent with the above interpretation. It covers the isotopic stages 6 and 7. During stage 6, the high and practically constant recorded value of $\delta^{18}\text{O}_{\text{sw}}$ (Figure 1a) indicates that the climate was fully glacial, with the ice caps at maximum extension. In the Vostok record only minor fluctuations are recorded (Figure 1e). In the interpretation of Bender *et al.* [1994b], a constant value of $\delta^{18}\text{O}_{\text{sw}}$ implies that no large amplitude variation in the $\delta^{18}\text{O}_{\text{atm}}$ signal is expected during that stage. During early glacial stage 6, however, $\delta^{18}\text{O}_{\text{atm}}$ displays a strong variation with a conspicuous minimum similar to that of interglacial stages (Figure 1f). The most direct approach to compare these two signals would be to analyse their difference. Since, however, the chronology of the two records is independent, either time scale can lead or lag by a few kyrs, and their absolute difference is hence uncertain. This question was investigated by Bender *et al.* [1994b]. It is not, in fact, necessary to know in detail the absolute chronology in order to grasp the basic mechanism: the first requirement is to know to which climatic stage a given variation belongs. In this part of the ice-core the deuterium record [Vostok project members, 1995, Jouzel *et al.*, 1996], which reflects temperature changes, together with the dust record [Jouzel *et al.*, 1996] which reflects full glacial stage maxima, allow stage 7.1 to be unambiguously attributed at the depth of 2,500 m (Figure 1e). This allows one to relate the occurrence of the conspicuous $\delta^{18}\text{O}_{\text{atm}}$ minimum at 2,300 m to the substage 6.5 occurring at 176 kyr BP [Imbrie *et al.*, 1984]. If this minimum really occurs in a glacial period, it contrasts strongly with later minima, all of which arise during warm interglacials /interstadials. In that case, the presence of this strong oscillation would show that, during stage 6, the change in $\delta^{18}\text{O}_{\text{atm}}$ is not caused by marine isotopic variations, since the latter were weak or absent during that period: the cause must be looked for elsewhere.

Here we confirm that early stage 6 was indeed fully glacial, as confirmed by the pollen content of sapropel in marine sediment of the eastern Mediterranean. We then go on to investigate another possible cause of the $\delta^{18}\text{O}_{\text{atm}}$ variation: the change in low latitude insolation.

Discussion

Glacial stages are concomitant with low eustatic sea level, which correspond to high values of $\delta^{18}\text{O}_{\text{sw}}$. However, estimates of $\delta^{18}\text{O}_{\text{sw}}$ are indirectly obtained from $\delta^{18}\text{O}$ in

foraminifera tests, which can be subject to other influences such as temperature and salinity. These uncertainties are illustrated by the difficulty of establishing the mean $\delta^{18}\text{O}_{\text{sw}}$, and therefore of reconstructing the absolute variation of the

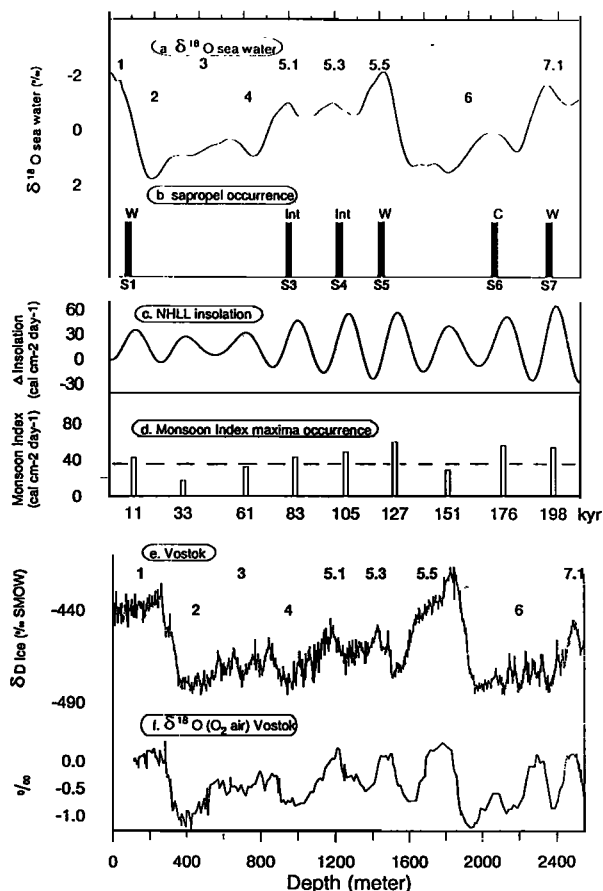


Figure 1. 1a: The $\delta^{18}\text{O}$ deep-sea record from SPECMAP Stack curve, SPECMAP time scale developed by *Imbrie et al.* [1984], based on normalized planktonic records. The corresponding isotopic stages are indicated.

1b: Sequence of sapropel in the Eastern Mediterranean sea sediment. The stratigraphic position is defined by the isotopic curve in the same cores where they are observed [*Vergnaud-Grazzini, 1977; Cita, 1977; Fontugne, 1992*]. Climate of the Mediterranean northern borderland during sapropel deposition (see Figure 2) was warm and humid (w), intermediate (int) or cold and arid (c) [*Rosignol-Strick et al., 1971; Rosignol-Strick, 1973; Rosignol-Strick, 1985; Cheddadi et al., 1995*]. 1c: Northern Hemisphere insolation in June at low latitude (23.5°N) expressed as a deviation, ΔI , from the 1950 value [*Berger, 1978*].

1d: Position and magnitude of Monsoon Index maxima as defined by *M. Rosignol Strick* [1983]. They express the strength of the African Monsoon. The threshold for sapropels is indicated.

1e: Vostok (Antarctica) isotope temperature (δD_{ice}) as a function of the depth [*Vostok project members; 1995*]. The corresponding isotopic stages are indicated.

1f: $\delta^{18}\text{O}$ of atmospheric oxygen [*Jouzel, 1993*] measured as function of depth in air bubbles trapped in the Vostok ice core (*M. Bender, private communication*). As the bubbles close at about 100 m depth, a time delay exists between the two records (1e and 1f), ranging from 2.5 to 6 kyrs depending of the climatic conditions [*Barnola, 1991*]. During stage 1 the $\delta^{18}\text{O}$ minimum arise around 7-8 kyrs BP [*Bender, 1994a; Bender, 1994b*].

sea level, or the intensity of the glacial stages. It is therefore of importance for our argument to ascertain that the oscillation of $\delta^{18}\text{O}_{\text{atm}}$ around 175 kyr BP really corresponds to glacial conditions. Mediterranean sapropel deposits provide the required confirmation.

Sapropels are discrete, black, organic-rich thin layers that were deposited basin-wide in Eastern Mediterranean sea sediments at intermittent intervals throughout the Quaternary. In these anoxic sediments, pollens are far better preserved than in the other parts of the core, and thus yield unequalled insight into the climate around the Mediterranean basin. Their sequence (labelled S1, S3, S4, S5, S6 and S7, see Figure 1b) has been correlated stratigraphically with the $\delta^{18}\text{O}_{\text{sw}}$ curve established for foraminifera in the same core [*Vergnaud-Grazzini, 1977; Cita, 1977; Fontugne, 1992*]. According to this stratigraphy in the last 200 kyrs all deposits of sapropels except one (S6) occurred during isotopically interglacial/interstadial stages (i. e. warm stages). The outstanding sapropel S6 was formed during early isotopic glacial stage 6 around 175 kyr BP in the marine chronology. Pollen records of the sapropels reveal the climate of the Mediterranean northern borderland at the time of their deposition [*Rosignol-Strick et al., 1971; Rosignol-Strick, 1973; Rosignol-Strick, 1985; Cheddadi et al., 1995*]. In core RC9-191 (Figure 2), this record confirms that the climate was warm and humid during S1, S5 and S7 (corresponding to interglacial 1, 5.5, and 7.1) moderately warm during S3 and S4 (corresponding to interstadial 5.1, 5.3) and cold and dry during S6 (corresponding to early glacial stage 6) (Figure 1b). Thus the pollen record establishes conclusively that early isotopic stage 6 displayed cold, arid, glacial conditions in the Mediterranean region during the deposition of sapropel S6. Glacial conditions for S6 are also supported from the foraminiferal assemblages in other eastern Mediterranean sediment cores [*Cita et al. 1977*]. The pollen content and foraminifera of sapropel S6 thus show that glacial conditions prevailed during early stage 6, which confirms that $\delta^{18}\text{O}_{\text{sw}}$ stays high at the same time as $\delta^{18}\text{O}_{\text{atm}}$ displays a strong variation. We can thus conclude that the variation of $\delta^{18}\text{O}_{\text{atm}}$ is not caused by marine isotopic variations during that stage.

A possible cause for the atmospheric variation is a change in insolation *via* hydro/biospheric activity, which is a mechanism of isotopic separation [*Bender, 1994b*]. Insofar as the hydro/biosphere at mean and high latitudes remains glacial during stage 6, significant changes in activity can occur only at low latitudes. The variation in hydro/biospheric activity is therefore connected to insolation changes at low latitudes only. Throughout the whole period under consideration, within the chronological uncertainty, the maxima of summer insolation at Northern Low Latitude (NLL) [*Berger, 1978*] coincide with the minima of $\delta^{18}\text{O}_{\text{atm}}$ (Figure 1). In particular, a strong maximum of summer insolation arises at 176 kyr BP during early isotopic stage 6 where the unexplained minimum in $\delta^{18}\text{O}_{\text{atm}}$ occurs. This corroborates the hypothesis that the variations in $\delta^{18}\text{O}_{\text{atm}}$ stem from low latitude hydro/biospheric changes *via* low latitude insolation.

Further support for this hypothesis comes from the sequence of sapropels, which, as discussed below, are indicators of enhanced hydrological activity at low latitudes. The link between sapropel formation and enhanced hydrological activity was first suggested by the co-occurrence of the most recent sapropel, S1, and the most humid recent phase in intertropical Africa [*Rosignol-Strick et al., 1982*]. Radiocarbon dates for the initiation of S1, established on 50 cores [*Fontugne, 1994*] are around 9 Kyr bp (10.3 Kyr BP calendar years [*Bard, 1993*]). The close coincidence of this initiation with the maximum of summer NLL insolation at 11 Kyr BP, calendar years [*Cheddadi et al., 1995*] corroborates the

scenario proposed by *M. Rossignol-Strick et al.* [1982]. The African monsoon feeds the Ethiopian highland rainfall, which the Nile river then channels into the Mediterranean sea. Modelling has shown that high insolation at low latitudes in

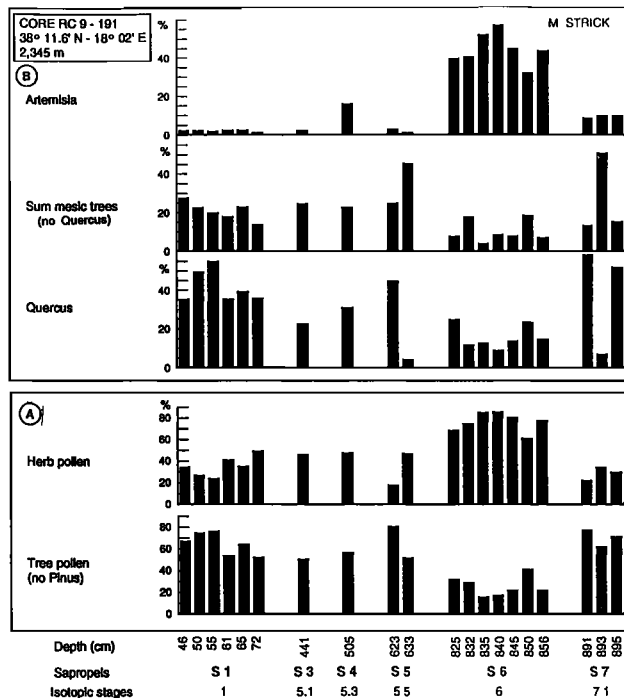


Figure 2. Pollen diagram of Late Quaternary sapropels, numbered according to ref. 11, in a deep-sea core from the Ionian sea, Eastern Mediterranean (sapropel 2 is absent). Sapropels are identified and dated (see Figure 1) by their stratigraphic position in agreement with cores with isotopic stratigraphy [Imbrie, 1984; Vergnaud-Grazzini, 1977; Cita, 1977]. The pollen sum in each sample is 300 to 500 (Pinus pollen excluded) [Rossignol-Strick et al., 1989].

A - Percentage of tree (Pinus excluded) and herb pollen. Dominance of tree pollen (S1, S3, S4, S5 and S7) indicates that the pollen source area, here the southern slopes of Italy and the Balkan peninsula, was forested at all altitudes, in response to warm and humid regional conditions (interglacial). Dominance of herb pollen (S6) indicates an open landscape, in response to dry and cold conditions (glacial period). These conditions are confirmed by the isotopic stratigraphy and comparison with pollen diagrams from Greece [Bottema, 1974; Wijmstra, 1969] and Italy [Follieri, 1988].

B - Percentage of pollen for two dominant taxa, the tree *Quercus* (oak), the grass *Artemisia* (sage-brush), and of the sum of other mesic (thriving in temperate, humid/warm conditions) trees excluding *Quercus* (*Ulmus*, *Tilia*, *Corylus*, *Alnus*, *Carpinus*, *Ostrya*, *Zelkova*, *Fagus*, *Acer*, *Betula*, *Abies*), that further document tree/herb categories. Sapropel 1 forms during the early part of the present interglacial: frost-free winters and drought-free summers in the Mediterranean domain, at least at mid-altitudes. Sapropel 3, during isotopic stage 5.1, and sapropel 4, in stage 5.3, form under interstadial climate. Sapropel 5, in stage 5.5, is clearly interglacial. In glacial stage 6, sapropel 6 reveals a regionally cold and dry climate. The vegetal cover is a sparse, sage-brush semi-desert at most elevations. Mesic trees are eliminated from mid-high elevation by low temperatures, and from low elevation by drought. Sapropel 7, in stage 7.1, displays high percentage of tree pollen (*Quercus* + other mesic trees), as do sapropels 1 and 5. During the most favourable interglacial conditions of high available moisture and warmth, the source area is forested.

the Northern summer induces heavier Asian and African monsoon precipitation [Kutzbach, 1984]. In the East Mediterranean heavy Nile discharge appears as a sudden dramatic input of a large volume of fresh water. Such water forms a low salinity surface layer that inhibits deep convection and ventilation of bottom water, which then becomes anoxic. Unoxidised refractory biodegradable detritus thus accumulates in the sediment, which becomes a sapropel. Sapropels are thus the result of increased monsoon activity through strong summer insolation. The link between sapropel formation and variation of insolation has been extended to the sapropels of the last 400 kyrs [Rossignol-Strick, 1983]. Absence of sapropels at 33, 61 and 151 kyr BP reflects lower than threshold values of the insolation maxima [Rossignol-Strick, 1983; Rossignol-Strick, 1985] (Figure 1c and 1d). Observation of Mediterranean sapropels over the last 200 kyrs is of interest for two reasons. First the sequence of sapropels is coeval with the strongest $\delta^{18}\text{O}_{\text{atm}}$ minima (Figure 1). This is consistent with our hypothesis that the minima in $\delta^{18}\text{O}_{\text{atm}}$ are driven by low latitude hydro/biospheric activity, which is a mechanism for isotopic separation, and which is illustrated by the sapropel deposits. Second, sapropels are the only physical record proving that summer NLL insolation maxima can be decoupled from the glacial/interglacial stages. Although most sapropels do coincide with a warm stage, the existence of the glacial sapropel S6 is a demonstration that a strong maximum in summer NLL insolation can coexist with a glacial stage. The same behaviour is found for the $\delta^{18}\text{O}_{\text{atm}}$ oscillations, which are decoupled from the glacial condition during early stage 6. The reason that the insolation maximum did not trigger deglaciation, an event unique in the last 200 kyrs, is a complex phenomenon that lies outside the scope of this article.

Conclusion

Since $\delta^{18}\text{O}_{\text{sw}}$ in the oceanic reservoir remains glacial we concluded that this reservoir did not make significant contribution to the observed strong oscillation in $\delta^{18}\text{O}_{\text{atm}}$. Thus for this period, the previous interpretation that atmospheric change is mainly driven by sea water change, is not tenable. Consequently the minimum observed in $\delta^{18}\text{O}_{\text{atm}}$ around 176 kyr BP has another origin. As middle and northern latitudes do not experience during that stage any drastic change (climate stays glacial as revealed by sea-level and sapropel pollen content) we suggest that this minimum can plausibly be attributed to a change in biospheric and/or hydrologic activities at low latitudes - a change which is driven by insolation. Furthermore, insolation could be the driving mechanism of the other minima, both earlier and later, since both the minima in $\delta^{18}\text{O}_{\text{atm}}$ and the sapropels all coincide with high summer northern low latitude insolation. However, the interpretation of these other minima may be more complicated, because they may also include the change of $\delta^{18}\text{O}_{\text{sw}}$ in the oceanic reservoir which can no longer be neglected as during stage 6. This leads to the final question: to what extent $\delta^{18}\text{O}_{\text{sw}}$ which is not the cause of the large change in $\delta^{18}\text{O}_{\text{atm}}$ during stage 6 is, as previously believed, the dominant factor in the variations of $\delta^{18}\text{O}_{\text{atm}}$ during the other periods? The low latitude hydro/biospheric activity, which has a potential impact as already mentioned by Bender et al., [1994b], could have an important contribution.

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