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B. Malaizé, D. Paillard, J. Jouzel, D. Raynaud

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B. Malaizé[1]

Graduate School of Oceanography, University of Rhode Island, Kingston
Laboratoire des Sciences du Climat et de l'Environnement, Unité Mixte de Recherche
Commissariat à l'Energie Atomique-CNRS, Gif-sur-Yvette, France

D. Paillard and J. Jouzel

Laboratoire des Sciences du Climat et de l'Environnement, Unité Mixte de Recherche
Commissariat à l'Energie Atomique-CNRS, Gif-sur-Yvette, France

D. Raynaud

Laboratoire de Glaciologie et Géophysique de l'Environnement, CNRS, St Martin d'Hères, France

Abstract. Detailed measurements of $\delta^{18}O$ of atmospheric oxygen performed on air trapped in the Vostok ice cores (Antarctica) are used to extend the record of the Dole effect over two climatic cycles (back to 240 kyr B.P.). Except for glacial terminations I and II and for an unexpected minimum occurring around 175 kyr, the Dole effect shows small variations (ΔDole within ± 0.5‰). These small variations, however, show a well-marked 23 kyr precessional periodicity, thus confirming the results obtained by *Bender et al.* [1994a] for the first climatic cycle. To explain the minimum value reached around 175 kyr, we invoke the possibility of a peak in the oceanic productivity linked to climatic events induced at low latitudes under glacial conditions.

1. Introduction

The Dole effect denotes the natural enrichment of the $^{18}O/^{16}O$ isotopic ratio of oxygen in the atmosphere with respect to its value in the ocean. Its present-day value is ~ 23.5‰ (expressed in the δ scale relative to the SMOW standard, which has an $^{18}O/^{16}O$ isotope ratio of 2005.2 x 10^{-6}). Most of this enrichment is due to the activity of marine and terrestrial biospheres which transmit the $\delta^{18}O$ signature of oceanic and continental waters to the atmosphere via mechanisms such as respiration and photosynthesis [*Dole*, 1935; *Lane and Dole*, 1956; *Dongmann*, 1974; *Förstel*, 1978; *Bender et al.*, 1985; *Sowers et al.*, 1991].

To estimate the Dole effect in the past, we need to look at paleoclimatic records of the $\delta^{18}O$ of the atmospheric oxygen (hereinafter $\delta^{18}O_{atm}$) and of the $\delta^{18}O$ of seawater (hereinafter $\delta^{18}O_{sw}$). The first information is obtained from the analysis of bubbles of ancient air trapped in ice cores. The second one comes from the study of deep sea sediment cores through the $\delta^{18}O$ of foraminifera shells, which gives access to the $\delta^{18}O_{sw}$ at the time of their formation. The Dole effect in the past is then obtained by subtracting the $\delta^{18}O_{sw}$ and $\delta^{18}O_{atm}$ signals at the same age. Uncertainties in the ΔDole (the difference between the Dole effect in the past and the Dole effect today) thus combine uncertainties associated with each of these two records but also errors in the relative chronologies of the ice and of the deep sea core records. As noted by *Bender et al.* [1994a], these latter uncertainties can indeed induce very large changes in apparent values of the ΔDole record.

In 1994, *Bender et al.* [1994a] performed a complete study of the Dole effect and of its variations over the first glacial-interglacial cycle, as inferred from the past variations of the $\delta^{18}O_{atm}$ values recorded in the air bubbles trapped in the Vostok ice cores. Their main result was that the Dole effect varied in the past but with a small amplitude. They concluded that the different processes affecting $\delta^{18}O_{atm}$ must have somehow canceled each other. However, they identified the presence of a precessional periodicity (~ 23 kyr) in the past variations of ΔDole with a remarkable relation with the summer insolation at low northern latitudes [*Bender et al.*, 1994a]. Such a precessional signal is also clearly seen in paleorecords of the marine productivity and terrestrial hydrological cycle at low latitudes [*Clemens and Prell*, 1990; *Chappellaz et al.*, 1990].

In this article we extend the pioneering work of *Bender et al.* [1994a] to the second climatic cycle using new data obtained from Vostok ice cores down to 2750 m. To assess the influence of chronological uncertainties on our estimate of the ΔDole effect, we use three different, independently derived Vostok timescales and perform additional sensitivity studies. After discussing different parameters involved in the Dole effect, we present our results and compare them with the results of Bender et al.. We confirm the conclusion of these authors concerning the presence of a strong precessional signal in the ΔDole record, and we examine its relation with the summer insolation at 20°N for the last two climatic cycles. We further discuss one striking feature of the second climatic cycle, namely, an unexpected high variation of the Dole effect during the insolation maximum around 175 kyr B.P., already mentioned by *Mélières et al.* [1997]. We examine how climatically induced changes in marine and terrestrial biospheres might have influenced ΔDole in the past.

[1]Permanently at Département de Géologie et Océanographie, Unité Mixte de Recherche 5805, Environnements et Paléoenvironnements Océaniques, Talence, France.

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2. The Dole Effect

To understand what might have influenced the Dole effect in the past, we first summarize the mechanisms involved [*Bender et al.*, 1994a]. During respiration most species use preferentially the light ^{16}O oxygen isotope, which results in an enrichment in ^{18}O for the atmospheric reservoir. Global isotope effects of ~ 18.0‰ for the respiration of the terrestrial biosphere and of ~ 18.9‰ for the respiration of the marine biosphere have been estimated [*Bender et al.*, 1994a]. During photosynthesis the $\delta^{18}O$ rejected in the atmosphere has the same value as the oxygen of the source water used by plants. Marine organisms use sea-water for photosynthesis, which has a modern value close to 0‰ whereas the source water for photosynthesis on land is taken from the leaves. The $\delta^{18}O$ of leaf water depends on many local parameters such as the $\delta^{18}O$ of water in the soil and in the atmosphere which are difficult to estimate with accuracy. For example, soil water comes from precipitation and has already fractionated during atmospheric processes [*Jouzel et al.*, 1987]. In addition, fractionation processes occurring at the surface of the leaf, including equilibrium and kinetic effects, depend on temperature and relative humidity [*Craig and Gordon*, 1965]. These processes lead to values higher than 0‰ for the photosynthetic $\delta^{18}O$ coming from most of the terrestrial biosphere, and therefore they contribute to enhance the global Dole effect. *Bender et al.* [1994a] also mentioned that the isotopic exchange between CO_2 and O_2 in the stratosphere slightly lowers $\delta^{18}O_{atm}$ (by ~ 0.4‰). Accounting for the terrestrial and marine Dole effects (and for the slight stratospheric diminution of this effect), these authors obtain a best present-day estimate of 20.8‰ and note the difficulty of correctly predicting the observed value of 23.5‰ [*Kroopnick and Craig*, 1972]. This difficulty of accounting for present-day observations should be kept in mind when interpreting past $\delta^{18}O_{atm}$ changes.

These past changes depend first on changes in $\delta^{18}O_{sw}$. This is because the isotopic composition of the ocean directly impacts on the marine biosphere (photosynthesis) but also because the isotopic content of continental precipitation is influenced by changes in the isotopic content of surface sea water, particularly in the tropics [*Jouzel et al.*, 1994] where much of the terrestrial photosynthetic production of O_2 occurs. Paleoclimatic records of $\delta^{18}O_{sw}$ show variations over time as a result of the growth and decay of the continental ice sheets. During glacial periods the atmosphere is enriched in ^{18}O relative to modern values, just like the ocean. It is not surprising that $\delta^{18}O_{atm}$ tracks $\delta^{18}O_{sw}$ at least when no change in the biospheric activity is assumed [*Sowers et al.*, 1991; *Bender et al.*, 1985]. We should, however, note that the change in the isotopic content of surface waters which mainly drives the change in the marine Dole effect (except for a small fraction occurring within more isolated parts of the ocean) [*Bender et al.*, 1994a] may differ from the change in the mean isotopic content of the ocean, in particular, at the time of meltwater events [*Leuenberger*, 1997].

In addition to changes in $\delta^{18}O_{sw}$, other processes could influence the isotopic composition of the atmospheric oxygen [*Sowers et al.*, 1991]. First, changes in the $\delta^{18}O$ of leaf water would lead to a modification of $\delta^{18}O_{atm}$ from terrestrial photosynthesis. Such changes may come from changes in the hydrological cycle leading to changes in the $\delta^{18}O$ of precipitation or from changes of environmental parameters, such as humidity [*Craig and Gordon*, 1965]. Second, a change in the ratio between terrestrial and marine productivity would lead to a variation of the Dole effect. For example, some authors have suggested that the terrestrial primary production, affected by changes in climate parameters and possibly in the CO_2 levels, could have been reduced by 25% with respect to the present time during the last glacial period [*Meyer*, 1988]. Since the terrestrial contribution to the Dole effect is more positive than the marine one, a decrease of the ratio between terrestrial and marine productivities would decrease the Dole effect. The problem is further complicated by the strong latitudinal dependency of the land Dole effect, which results primarily from the decrease of the $\delta^{18}O$ of precipitations with latitude. For example, an increase in the high-latitude/low-latitude ratio of the terrestrial productivity would lead to a decrease of the Dole effect (keeping unchanged this terrestrial productivity). Third, a parameter concerning possible changes in the land respiration factor should be considered, for this factor, as pointed out by *Leuenberger* [1997], may be itself climate dependent.

3. The $\delta^{18}O_{atm}$ Vostok Record

Five boreholes have been drilled at the Russian station Vostok (East Antarctica). In their study, *Bender et al.* [1994a] used samples from the 3G core only. We use here a compilation of measurements performed on samples from 3G, 4G, and 5G cores [*Sowers et al.*, 1991, 1993; *Bender et al.*, 1994a; *Jouzel et al.*, 1993, 1996] to obtain a detailed record of the $\delta^{18}O_{atm}$ over two climatic cycles. All these isotopic data have been obtained at the Graduate School of Oceanography, Rhode Island, following the extraction method and the analytical procedure established by *Sowers et al.* [1989]. All the $\delta^{18}O$ measurements have been corrected by using $\delta^{15}N$ measurements, for the gravitational fractionation occurring in the firn [*Sowers et al.*, 1989]. This also accounts for the thermal fractionation, if we assume that both processes lead to effects having an amplitude about twice larger for $\delta^{18}O$ than for $\delta^{15}N$ (according to *Severinghaus et al.* [1996], the thermal diffusion factor of pairs is of 0.0107 and 0.0069 for oxygen and nitrogen isotopes, respectively). Indeed, the estimate of the thermal fractionation factor for nitrogen isotopes may be twice lower (J.P. Severinghaus, personal communication, 1998), which means that only part of the correction due to thermal diffusion is taken into account for oxygen isotopes. On the other hand, effects due to thermal fractionation are very small for air bubbles trapped in the Vostok ice as there is no rapid temperature change at this site and as the associated error is negligible. For our compilation of the $\delta^{18}O_{atm}$ records, we account for the largest uncertainty of the three data sets available, i.e., 0.075‰.

A comparison of ash layers found in these three different cores reveals slight differences (up to 3 m) between depths of similar events recorded in 3G, 4G, and 5G. To combine the three records in one unique stack, we account for these slight differences in depths. The final record consists of 242 evenly spaced data points which provide a high-resolution record covering the last 240 kyr, with roughly one point every thousand years. The mixing time of the oxygen in the atmosphere has been estimated around 1.2 kyr [*Bender et al.*, 1994a; *Sowers et al.*, 1993]. This record can therefore be considered as representative of the global isotopic composition of the oxygen. This is further supported by the fact that the same $\delta^{18}O_{atm}$ signal is observed, on the first climatic cycle, in the Greenland Ice Sheet Project 2 (GISP2) ice core in central Greenland [*Bender et al.*, 1994b].

4. Estimating the Dole Effect

Following the approach of *Sowers et al.* [1993], we use the carbonate $\delta^{18}O$ record of benthic foraminifera from the V19-30 deep sea core, in the Pacific Ocean [*Shackleton and Pisias*, 1985] to estimate past variations of the $\delta^{18}O_{sw}$. This record, dated by orbital tuning, has been chosen because it is a deep record situated far from any area of cold deep water formation (such as the North Atlantic) and thus should give a realistic picture of global $\delta^{18}O_{sw}$ variations. Nevertheless, even deep sea waters could have suffered small temperature variations, hence inducing some carbonate isotopic composition variations. We account for the influence of bottom water temperature changes in extending to the second climatic cycle the correction proposed by Sowers et al. for the first climatic cycle. The V19-30 $\delta^{18}O$ record is linearly scaled in such a way to have values of 0‰ and 1.15‰ for the Holocene and the Last Glacial Maximum, respectively. By using this benthic record for estimating the marine Dole effect, we assume that these corrected $\delta^{18}O_{sw}$ variations only reflect past sea level variations and therefore can be considered as a global signal. This should be a correct assumption except at the time of large meltwater events [*Leuenberger*, 1997].

We have calculated the ΔDole effect for three different Vostok timescales. The orbitally tuned timescale (OTT) is derived from an orbital tuning approach consistent with the SPECMAP marine timescale [*Waelbroeck et al.*, 1995]. The second timescale, called the extended glaciological timescale modified (EGTmod), is based on a linear modification of the bottom part of the previous EGT Vostok glaciological timescale to get an age corresponding to marine stage 7.5 (~ 240 kyr B.P.) for a depth of 2755 m [*Jouzel et al.*, 1996]. The difference between these two timescales, giving the age of the ice, never exceeds 5 kyr. As air bubbles found in the Vostok ice are isolated from the surface several thousand years after the snow deposition, the gas trapped in the ice is younger than the surrounding ice. We calculate the difference between ice age and gas age using the model developed by *Barnola et al.* [1991], which takes into account climatic parameters specific to the deposition site (surface temperature or accumulation rate).

The third timescale, based on isotopic correlations (ICT), gives directly an estimate of the age of the gas ; it simply extends the approach adopted for the first climatic cycle by *Sowers et al.* [1993], who proposed to get a Vostok gas age chronology considering that after accounting for the atmospheric reservoir effect, changes in $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$ are in phase (thus minimizing the difference between both records, i.e., the ΔDole signal). The largest differences (~ 7 kyr) are observed between ICT and EGTmod during marine isotopic stages 5 and 7. We should keep in mind that the ICT timescale results in minimizing the Dole effect. The valididy of these different timescales will be discussed further in this paper by comparing our Dole effect results with other paleodata.

According to these three different chronologies, the $\delta^{18}O_{atm}$ record now covers approximately the last 240 kyr, enabling us to extend the study of the Dole effect to the penultimate glacial-interglacial cycle. Figure 1 shows the ΔDole records over the last two climatic cycles for two different Vostok timescales, i.e., EGTmod [*Jouzel et al.*, 1996] and ICT timescales (OTT and EGTmod giving almost similar results). Both records used to calculate these Dole effect variations, i.e., $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$, are also represented in figure 1 as well as the June 21 insolation at 20°N. We follow *Bender et al.* [1994a] in using this low-latitude signal in which the precessional parameter can easily be seen for comparison with insolation. As already pointed out by *Jouzel et al.* [1996] using EGTmod, there is a striking similarity between $\delta^{18}O_{atm}$ and this insolation curve, which holds true whatever the timescale used.

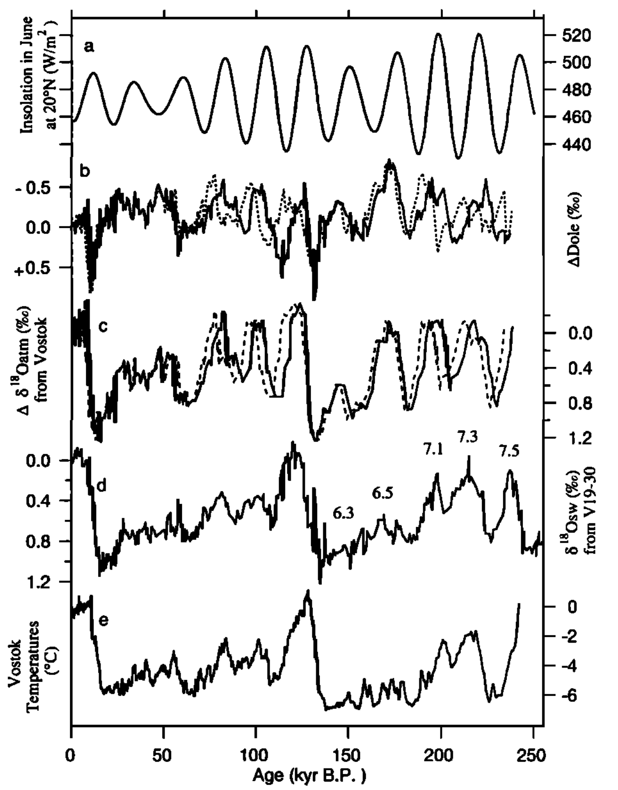

Figure 1. Variations with time of (a) the summer solstice insolation (June 21) at 20°N [*Berger*, 1978], (b) the Dole effect as calculated from $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$ records, (c) the $\delta^{18}O_{atm}$ record from gas analysis of the Vostok ice core dated with gas age timescales, (d) the $\delta^{18}O_{sw}$ record from the V19-30 core modified as described in the text, and (e) the mean atmospheric temperature at the Vostok site (Antarctica) (EGTmod timescale) as deduced from the ice deuterium content [*Jouzel et al.*, 1996]. The solid and dashed lines in figures 1b and 1C are for the Dole record calculated using the $\delta^{18}O_{atm}$ record dated with the extended glaciological timescale modified (EGTmod) and the isotopic correlation timescale (ICT) respectively. The annotations in figure 1d give the different stages as defined by *Martinson et al.* [1987].

For the last glacial-interglacial cycle we calculate a mean variation of the Dole effect (taken with respect to the present-day value) of -0.08‰ with a standard variation of ±0.27‰. *Bender et al.* [1994a] calculated a mean variation of -0.05‰, with a standard deviation of ±0.24‰. These minor differences between both studies can be related to differences between timescales and different ice samples used for the $\delta^{18}O_{atm}$ record. Our general conclusion is quite similar to the previous study: The Dole effect remains relatively stable over the last 130 kyr. The amplitude of the variations stays roughly in a ±0.5‰ interval, except for the last two glacial terminations. For the penultimate climatic cycle, four events are clearly defined. All of them can be related to an insolation maximum: the first one around 145 kyr B.P. (stage 6.3), the second one at 175 kyr B.P. (stage 6.5), the third one around 195 kyr B.P. (stage 7.1), and the fourth one around 220 kyr B.P. (stage 7.3).

A spectral analysis of the complete record confirms that a precessional signal can be extracted from the ΔDole record (Figure 2) and also gives a better statistical confidence (Table 1) than in the previous study of *Bender et al.* [1994a]. For the precessional periodicity these authors calculated a f statistic of 0.93. We found a f statistic of 0.995 for EGTmod, 0.994 for OTT, and 0.999 for ICT. A cross-spectral analysis (Blackman-Tuckey method) between the ΔDole record and the summer solstice insolation at 20°N shows also a good coherency between both signals (Table 2 and Figure 2). The maximum coherency is obtained around 23 kyr, with the ΔDole signal lagging the insolation. This lag is ~ 3.5 - 4 kyr for either OTT or EGTmod while it is of the order of 8.2 kyr for ICT (Table 2). The two extreme values (i.e. 3.5 and 8.2 kyr) define the phase lag uncertainty linked to the timescale used. In section 5.2, we will use these extreme mean values to compare our results with previous studies.

When looking more precisely at individual features of the ΔDole record, we have to be very careful at the representativeness of each event as some of them could be artificially created by uncertainties in deep sea core versus ice core chronologies. Nevertheless, the "strong event" at stage 6.5 keeps its very low value independently of the timescale used. *Bender et al.* [1994a] showed that subtracting several thousand years to a timescale could change not only the timing of events in the ΔDole effect record but also their amplitude. To better evaluate these uncertainties, we follow the same approach, by performing sensitivity studies in modifying the Vostok timescales used. We added and subtracted 5 kyr to these timescales and recalculated the ΔDole record keeping the marine timescale unchanged. The results are presented in Figure 3.

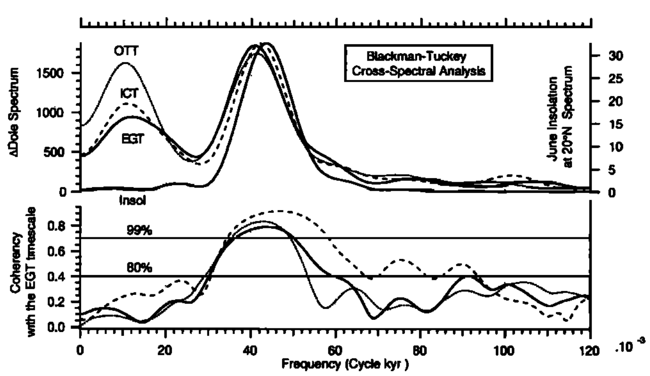

Figure 2. Blackman-Tuckey cross spectral analysis. (top) The spectra of ΔDole records (calculated with three different timescales) and of the summer solstice insolation signal at 20°N. Dashed lines are for results obtained using the orbitally tuned timescale (OTT) and the ICT timescales. (bottom) The coherency between the insolation record and these ΔDole records is shown below. Horizontal lines represent the non zero coherency (80% and 99% confidence levels).

Table 1. Spectral Analysis of the ΔDole and $\delta^{18}O_{atm}$ Records for Each Timescale.

	ΔDole		$\delta^{18}O_{atm}$	
	Period, kiloyears	*f* Statistic	Period, kiloyears	*f* Statistic
EGTmod	23.8	0.9949	23.8	0.9992
OTT	23.8	0.9939	23.3	0.9995
ICT	23.8	0.9997	23.3	0.9998

The multitaper method has been used to obtain the results [*Paillard et al.*, 1996]. EGTmod, extended glaciological timescale; OTT, orbital tuned timescale; ICT, isotopic correlation timescale.

Most events have their amplitude strongly affected by these timescale modifications. Nevertheless, each oscillation of the ΔDole record remains, with a time shift consistent with the modifications. ΔDole values during stage 6.5 always exhibit a remarkable minimum, with a value below -0.6 to -0.85‰. A closer look at the marine record can easily explain this persistent feature. During stage 6.5 the marine record does not show any large variation, while a strong minimum is clearly present in the atmospheric record (Figure 1). The difference between the two records is thus entirely insensitive to changes in the ice core chronology. These exceptionally low values of the Dole effect during stage 6.5 correspond to a particular event which has no counterpart during the first glacial-interglacial cycle. The amplitude of the other ΔDole excursions may not be as reliable.

Meanwhile, we note that amplitudes of precessional variations in the $\delta^{18}O_{atm}$ record are always larger than the ones in the marine record. The difference must therefore be significantly different from zero in this frequency band. Although this has to be considered with caution since the $\delta^{18}O_{sw}$ record deduced from the V19-30 core may not give a realistic picture of real $\delta^{18}O_{sw}$ variations (see section 4), the significant precessional variation in the ΔDole signal is further supported by the cross-spectral analysis between $\delta^{18}O_{atm}$ and ΔDole. Both records are in phase within 1.4 kyr (Table 2), showing that the ΔDole signal is dominated by the atmospheric isotopic variations.

We further investigate the phase relationship obtained

Table 2. Cross-Spectral Analysis Between the ΔDole Record Reversed and the Record of the Average June Insolation at 20° N.

	Period, kiloyears	Coherence	ΔDole-Insol Phase kiloyears	$\delta^{18}O_{atm}$–ΔDole Phase kiloyears
EGTmod	22.7	0.79	3.5	- 0.3
OTT	23.3	0.83	4.0	- 0.3
ICT	23.8	0.89	8.2	1.1

A Blackman-Tuckey method has been used for this cross spectral analysis [*Paillard et al.*, 1996]. We used the reversed ΔDole record in order to associate minima of ΔDole record with maxima of June insolation (Insol). In the last column we calculated the phase between the $\delta^{18}O_{atm}$ record and the ΔDole record for each timescale.

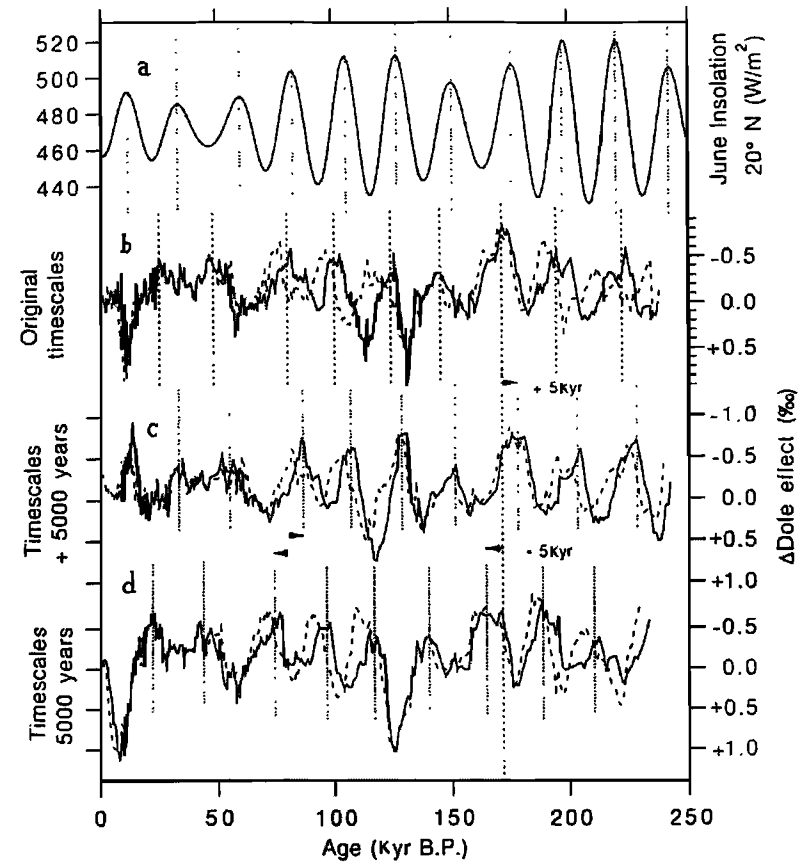

Figure 3. (a) The summer solstice insolation at 20°N [*Berger*, 1978]. The sensitivity of the ΔDole record to changes in timescales, this record being calculated (b) with the EGTmod timescale [*Jouzel et al.*, 1996] and the ICT timescale (dashed line), (c) in adding 5 kyr to the Vostok timescales (see text), and (d) in subtracting 5 kyr.

between ΔDole and June insolation records, by filtering each ΔDole record around the 23 kyr frequency and calculating the time lag between each minimum in the ΔDole and each maximum in the insolation signal (Figure 4)(Table 3). As the uncertainties, both in the absolute and relative Vostok and V19-30 chronologies, influence the estimation of these lags (see Figure 3), individual time lags are associated with a large uncertainty, probably of the order of ± 5 kyr. Bearing this in mind, it is nevertheless interesting to look at the distribution pattern of these lags versus time. Changes in the June insolation intensity generally lead to changes in the Dole effect, except for the event occurring around 220 kyr B.P., which probably results from dating problems in the deep part of the Vostok core. These time lags are consistent with spectral analysis results with approximately constant values along the whole record when using ICT. This comes as no surprise, since this chronology was built assuming a constant phase between

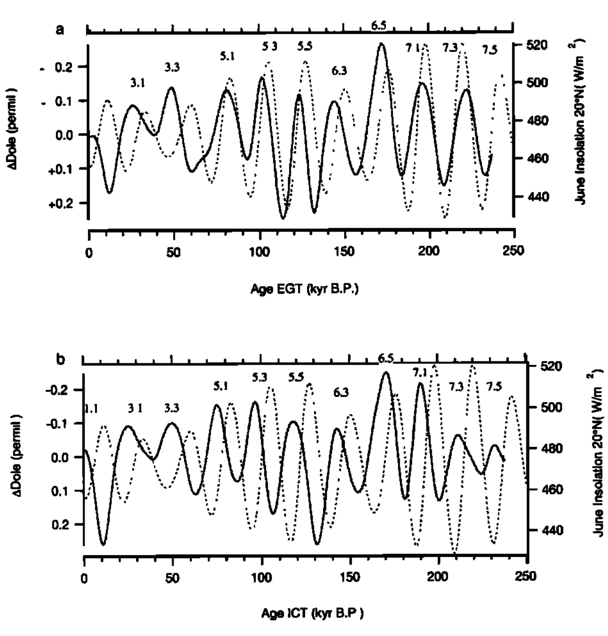

Figure 4.
Time lags between the ΔDole record and the summer solstice insolation signal at 20°N. The ΔDole records have been filtered around 23 kyr with a bandwidth of 0.04 cycles per kiloyear. Time lags between ΔDole minima and each insolation maxima are shown in Table 3. The dashed line is for the insolation signal. The solid lines are for (a) the filtered record of the ΔDole using the EGTmod timescale [*Jouzel et al*, 1993] and (b) the filtered record of the ΔDole using the ICT timescale. The annotations give the different stages as defined by *Martinson et al.* [1987].

marine and atmospheric isotopic signals. On the contrary, EGTmod and OTT tend to indicate a systematic phase difference between glacial and interglacial periods with larger lags during glacial times (3.5 - 11 kyr) than during stages 5 and 7 (-2 - 3 kyr).

5. Discussion

The main features of the complete ΔDole record in the past are the following. Except for stage 6.5, where the Dole effect reaches a minimum value down to -0.85‰, variations are quite small with respect to present-day Dole effect. Mean variations are no larger than 0.5‰ in comparison to the present 23.5‰ enrichment. These variations have a clear precessional periodicity of ~ 23 kyr, with maxima in the northern latitude summer insolation (20°N) followed by minima in the Dole effect with mean time lag values between 3.5 and 8.2 kyr (Table 2). To explain the remarkable constancy of the Dole effect, we discuss the main changes in the global biospheric activity between interglacial and glacial periods. Then we focus on small variations of the Dole effect and examine climatic changes driven at low northern latitudes by summer insolation changes. Finally, we suggest a hypothesis to explain the exceptional "Dole event" around 175 kyr B.P.

5.1. Glacial-interglacial variations

When the Earth enters a glaciation, climatic changes have an impact on both continental and marine biospheres. Many environmental parameters are changing, and the resulting effect on the global Dole effect is nontrivial. The dramatic drop in global temperature allows huge ice sheets to cover part of the Northern Hemisphere continents. The high-latitude terrestrial biospheric extension is then reduced, while the low-latitude one may eventually get enlarged by the newly emerged continental margins, thus moving the overall terrestrial biomass closer to equatorial areas. This general migration may induce an increase in the terrestrial Dole effect since precipitations are isotopically heavier in low latitudes than in high latitudes. In this line, we note that simulations of the distribution of water isotopes at the Last Glacial Maximum suggest that the isotopic content of the precipitation was then higher in tropics and subtropics than it is for present-day climate [*Jouzel et al.*, 1994; *Hoffmann et al.*, 1999], probably as a result of some "compensation effect" between high and low latitudes [*Jouzel et al.*, 1998]. This would tend to further increase the Dole effect. But much colder temperatures will also reduce notably the $\delta^{18}O$ of precipitations, at least in mid latitudes and high latitudes [*Jouzel et al.*, 1994], thus counteracting the previous change. Glacial climate is also drier and has a lower pCO_2 [*Barnola et al.*, 1987]. Less humidity in the glacial atmosphere has a direct impact on the $\delta^{18}O$ of water in the leaves, rising its isotopic composition [*Förstel*, 1978] and consequently the Dole effect. A lower CO_2 would reduce the overall terrestrial primary production and thus lower the Dole effect [*Jolly and Haxeltine*, 1997]. On the other hand, a lower CO_2 concentration during glacial period would have increased the photorespiration/carboxylation ratio and therefore raised the Dole effect (see for more details, *Bender et al.* [1994a]). To quantify all these counterbalancing effects, simulations of the terrestrial Dole effect, developed in different laboratories [*Hoffmann et al.*, Past changes in ocean and land productivity as inferred from variations in the Earth's Dole effect, submitted to *Science*, 1999](hereinafter referred to as Hoffmann et al., submitted manuscript, 1999) are necessary to better understand the contribution of the continental biosphere to the global Dole effect.

Modifications of the oceanic Dole effect are probably not coming from changes in fractionation factors involved but more likely from variations in the primary production. In their previous work, *Bender et al.* [1994a] made a complete overview of most studies of marine paleoproductivity. Analysis of tracers in low-latitude sea sediment cores seems to show a higher ocean productivity during the Last Glacial Maximum [*Sarnthein et al.*, 1988; *Marino et al.*, 1992], together with stonger upwellings [*Molfino and McIntyre*, 1990; *Mix*, 1989]. A direct consequence would be a decrease of the Dole effect. On the other hand, a decrease of the highly productive continental shelves area as a result of the sea level lowering would lead to a decrease in the global marine production and then to a higher Dole effect.

To sum up, most of the changes which occur in the biospheric activity, as well as the modifications of fractionation factors, have competing influences on the Dole effect. The rough constancy of the Dole effect suggests that the effects of all these mechanisms on the ^{18}O of the atmosphere stet almost perfectly counterbalanced.

5.2. Precessional Variations

While the ΔDole record does not show significant fluctuations associated with the 100 kyr main glacial cycle, the precessional frequency of 23 kyr is clearly present, which suggests a low-latitude mechanism. We therefore investigate the response of marine and terrestrial biospheres to changes in insolation in the low northern latitude band. Higher summer insolation has a different impact on the atmosphere over continents than it does on the atmosphere over oceans. It creates different pressure systems and leads to stronger winds in regions such as West Africa, Southwest Asia, the Arabian Sea, and the west coast of South America. It also creates higher evaporation over the oceans, enhances humidity in the atmosphere, and produces overall stronger precipitations. Such variations can be seen in paleorecords of tracers such as methane [*Chappellaz et al.*, 1990] as well as in simulated paleoclimates [*Prell and Kutzbach*, 1987]. These phenomena are associated with high monsoons occurring in the low-latitude band. They can even be detected in sediments in the Mediterranean Sea: stronger precipitations, issued from strong East African monsoons, produced large discharges of freshwater in the Mediterranean basin. These freshwater layers on the Mediterranean surface stop the vertical mixing of waters and create anoxic environments, which leads to a deposit of organically enriched sediment layers, known as sapropels [*Kallel et al.*, 1997]. The monsoon index has to exceed a certain limit to create sapropel in the Mediterranean Sea. Therefore each sapropel layer is associated with a strong monsoon event [*Rossignol-Strick*, 1983].

For the terrestrial biosphere a direct consequence of stronger precipitation is to increase the productivity. Since the low-latitude terrestrial biosphere drives $\delta^{18}O_{atm}$ toward higher values, we expect then the Dole effect to rise. However, higher humidity also minimizes the fractionation factor in the leaf water. Once again, both mechanisms have an opposite effect on the global Dole effect, and the influence of each phenomenon is difficult to quantify.

For the marine biosphere, recent studies tend to prove that monsoons have an impact via the stronger winds that they create. *Ruddiman and Mix* [1993] have worked on sea surface temperature (SST) records in the Atlantic Ocean. They made two temperature reconstructions, at 6 and 9 kyr B.P., and found colder temperatures around the west coast of Africa (Guinean Upwelling). These authors concluded that the intensity of this upwelling might have been higher at these time periods. Results obtained at 9 kyr were the most convincing ones. *Prell et al.* [1990] also showed an increase of the upwelling field in the northwestern Indian Ocean, at around 9 kyr B.P. According to climate model simulations [*Prell and Kutzbach*, 1987], this time period corresponds also to a maximum monsoon intensity. More likely, high monsoon intensity could create stronger trade winds, which could increase upwellings already existing in the low-latitude band. As the marine productivity is directly linked to the supply in nutrients, higher upwellings enhance the marine productivity. A large part of the global marine productivity is concentrated in the upwelling areas. We could then assume that such phenomena would have a global impact and would decrease the Dole effect.

Each monsoon maximum is related to a maximum of the summer insolation signal [*Prell and Kutzbach*, 1987; *Clemens and Prell*, 1990] by enhancing the pressure gradient over continents and marine areas. If we consider the summer solstice insolation signal taken at 20°N as a reference, its last maximum intensity occurred around 11.5 kyr B.P. The strongest monsoon recorded around 9 kyr B.P. can be related to the solstice insolation signal with a lag of ~ 2.5 kyr (which means that it is in phase with the insolation at the end of July). This observation is compatible with our investigation of the phase relationship between our ΔDole record and the insolation signal. Indeed, in our analysis we observe a minimum in the ΔDole Vostok record lagging the maximun 20°N insolation signal (taken in June) with a mean time lag of 3.5 or 4 kyr using the EGT or OTT timescales (but 8.2 kyr using the ICT timescale; compare mean values in Tables 2 and 3). Since the reservoir effect of the atmospheric $\delta^{18}O$ is of the order of 1.2 kyr, our results are therefore in good agreement with this monsoon hypothesis, at least when using the EGT and OTT scales.

But *Clemens and Prell* [1990] have worked on several components of sediment cores from the Arabian Sea, which are also influenced by summer monsoon. They calculated a general lag of 9 kyr between the summer solstice insolation and the maximum lithogenic grain size, which they believe is representative of the summer monsoon wind strength. Their monsoon indicator is therefore in phase with mid-November insolation. In contrast to the Holocene data, this larger time lag is more comparable to the lag we found using the ICT timescale, or to the "glacial-time" time lag. In a new study in the same area, *Clemens et al.* [1996] have found a time lag between 6 and 12 kyr, depending on the proxy used. This example highlights the difficulty of comparing different monsoon proxies, even in the same area of the ocean, and therefore the difficulty of defining a reliable monsoon maximum. A further difficulty arising in the study of local marine records is that they may be influenced by geographic shift of the monsoon as well as by its intensity, thus further complicating the interpretation of lags.

In this last study, *Clemens et al.* [1996] have also shown that this time lag has not remained constant throughout the Plio-Pleistocene but could have been influenced by the progressive ice sheet buildup over the last 3.5 million years. This may be consistent with our finding that the lag of the Dole effect with respect to precession appears to be larger during glacial periods. We should bear in mind that this lag difference could be symptomatic of a systematic chronology offset between ice core and deep sea cores or between these records and insolation changes. However, it can also represents a real feature of the variations of the Dole effect. A possible explanation would involve a shift in the timing of the summer monsoon. It has indeed already been suggested that the monsoon could be delayed by several weeks or even months in glacial periods, in particular, because of a more important seasonal extent of snow and ice cover over Tibet and Eurasia [*Barnett et al.*, 1989]. The timing of the monsoon could also be influenced by many other environmental parameters, like sea surface temperatures, CO_2 levels, or vegetation changes. A late monsoon in glacial time would respond to a seasonal insolation maximum in late summer, instead of early summer. On a precessional phase wheel a 1-month difference represents a lag of 1.9 kyr (one twelfth of 23 kyr). Such a seasonal shift for the monsoon could therefore explain our observed larger time lag during glacial periods. Furthermore, if we assume that a sizeable monsoon can occur only above a specific threshold of the summer insolation, such a threshold would be reached a few thousand years later in glacial periods since the summer insolation values are then significantly lower. It is also worth noticing that a larger time lag is expected during glacial periods if the size of the terrestrial biosphere was reduced at this time, as it has been suggested [*Meyer*, 1988]. Indeed, the 1.2 kyr turnover time of the atmospheric oxygen [*Sowers et al.*, 1993] is linked to the magnitude of the oxygen flux between the biomass and the

Table 3. Time lags between ΔDole and insolation records

Marine stages	1.1	3.1	3.3	5.1	5.3	5.5	6.3	6.5	7.1	7.3	Mean
Insol - EGT kiloyears	9.5	7.5	11	2*	3*	3.5	6	3.5	1.5*	-2	3.5
Insol - OTT kiloyears	9.5	9	8.5	3*	2.5*	4.5	3.5	4	6.5	-2.5	4
Insol - ICT kiloyears	11	8.5	10	8	8.5	8.5	7	5	8	8	8.2

*Time lags below 3 kyr occur during marine stages 5 and 7

For each minimum in the ΔDole signal corresponding to a maximum in the June insolation signal, we calculate a time lag between both records. Bold characters are for lags below 3 kyr and are systematicaly during marine stages 5 and 7. The mean time lag values are the results of the cross spectral analysis presented in Table 2.

atmosphere and therefore to the overall size of the biomass. However, a threefold or fourfold reduction in overall biomass should be invoked to explain all the observed lag difference, which is certainly unreasonable.

Although an increasing number of works suggest a direct link between the insolation signal and marine productivity [*Beaufort et al.*, 1997], the problem is far from being solved. Besides, past marine productivity reconstructions are still an extremely difficult problem. Many other parameters might also play a role in this respect. For future studies of the Dole effect it is necessary to better understand the relative importance of each mechanism.

5.3. The Stage 6.5 Event

A remarkable feature of the record, still to be explained, is the large ΔDole event occurring at 175 kyr B.P. During this stage 6.5 the $\delta^{18}O_{atm}$ record reaches a minimum, whereas no drastic variation is clear in the $\delta^{18}O_{sw}$ record. This leads to a minimum value of the Dole effect of - 0.85‰, which is larger than all other variations observed over the whole record. To explain this event, we note that two specific conditions, both having a decreasing impact on the global Dole effect, did occur simultaneously.

First, a strong monsoon event probably occurred around 175 kyr B.P., stronger than that during the first climatic cycle at an equivalent stage of the glaciation (i.e., 33 kyr B.P. and 61 kyr B.P.)[*Rossignol-Strick*, 1983]. Indeed, the summer insolation signal at 20°N was much more intense at around 175 kyr B.P. than it was during marine isotopic stages 2, 3, and 4. This high insolation might have been responsible for an intense monsoon. The hypothesis was proposed recently by *Mélières et al.* [1997], based on the existence of a concurrent sapropel layer (called S6) that these authors have correlated with the minimum of the $\delta^{18}O_{atm}$ record occurring at about the same time. Instead, no sapropel layer is found during the last glacial period between 70 and 15 kyr B.P. We suggest that this intense monsoon increased the marine productivity, therefore lowering the Dole effect.

Second, one should consider the glacial conditions existing around 175 kyr B.P. Foraminifera and pollen content, analyzed in this sapropel S6, are characteristic of glacial conditions [*Mélières et al.*, 1997]. Both low CO_2 and low deuterium content [*Jouzel et al.*, 1993] are indicators of early glacial conditions for stage 6.5. According to our hypothesis, this monsoon would be the only strong one over the last two climatic cycles, occurring under severe glacial conditions.

It is extremely difficult to quantify each factor and its role in the global Dole effect. In particular, we do not know the overall influence on the Dole effect of continental biospheric changes (or marine biospheric changes) due to monsoons. However, a rough estimate can be done to estimate the changes necessary to get the observed 0.85‰ decrease of the Dole effect with respect to its present-day value. We start from the values given by *Bender et al.* [1994a], for the present-day terrestial and marine productions (20.4 and 10.6, respectively, in 10^{15} mol/yr). To get the observed value of 23.5‰ for the present-day Dole effect [*Bender et al.*, 1994a), a value of 18.9‰ for the marine fractionation effect implies a fractionation factor over land of 26.5‰ (taking into account the stratospheric effects) (see *Bender et al.* [1994a] for a more detailed discussion). Assuming a 25% decrease of the terrestrial production during glacial periods (and no change for the marine and land fractionation factors) requires a marine production of 12.7 x 10^{15} mol/yr to get a ΔDole of -0.85‰. The same approach has been followed in using model estimates of terrestrial and marine productions as simulated by Hoffmann et al. (submitted manuscript, 1999). For present-day climate these estimates are much lower than those used by Bender et al. for both terrestrial and marine values (12.1 and 7.6 x 10^{15} mol/yr, respectively). The same assumption of a 25% decrease during glacial period then requires a marine production of 8.8 x 10^{15} mol/yr. In both cases, if the terrestrial production is decreased by 25% during glacial periods, the marine production has to increase by 15-20% to account for the observed Dole effect.

It seems difficult, but not unlikely, to consider a strong monsoon with an impact of this size (20% increase) on the marine productivity. We admit that this calculation is somewhat speculative. For example, if we were assuming a land fractionation factor of 22.4‰ as suggested by *Bender et al.* [1994a], a doubling of the marine productivity would be required to explain the 6.5 event (thus illustrating the crucial need of explaining the present-day Dole effect if we want to infer correct information on past land and marine productivities). Additional work in order to corroborate the existence of a strong monsoon event during stage 6.5, under glacial conditions, is even required. One possibility would be to look at paleoclimatic tracers of monsoon at that particular time, especially tracers of stronger winds, such as lithogenic grain size in sediment cores [*Clemens and Prell,* 1990]. Other clues could probably be obtained from atmospheric and oceanic general circulation models, which could be used to examine the response of upwellings to enhanced low-latitude insolation.

6. Conclusions

In this article we have reconstructed the Dole effect over two climatic cycles, i.e., for the last 240 kyr. Generally, the signal stays relatively stable for the whole record. Small variations hardly exceed an amplitude of 0.5‰, except for the two glacial maxima and for one exceptional event, around 175 kyr B.P. These general variations are weak with respect to the present-day value of the Dole effect, around 23.5‰. They nevertheless show a precessional periodicity, close to 23 kyr, with a strong link with the insolation signal at low northern latitudes. The June insolation signal leads the ΔDole record by a mean value ranging from 3.5 to 8.2 years, depending on the timescale used for the reconstruction. To explain the exceptional decrease of the Dole effect at 175 kyr B.P., down to -0.85‰, we invoke an unusually strong monsoon, which occurred under glacial conditions. The mechanism linking monsoonal activity and the Dole effect might involve marine productivity, via modulation of upwelling strength.

However, our understanding of the Dole effect and its fluctuations is still strongly limited by the numerous uncertainties involved. They result not only from uncertainties in deep sea cores and ice core chronologies but also from the lack of sufficiently detailed models for many mechanisms involved within the terrestrial and marine biospheres. Current modeling of the Dole effect [*Leuenberger*, 1997] also needs to be improved as exemplified by the difficulty of correctly accounting for its present-day value [*Bender et al.*, 1994*a*]. The main problem seems to arise from the estimation of the $\delta^{18}O$ of the water in leaves, which presents a high variability in space and time [*Bender et al.*, 1994a; *Keeling,* 1995]. Despite these difficulties and limitations, the present extension of the ΔDole record to the second climatic cycle further illustrates the interest of this parameter which, providing a better understanding of mechanisms involved, should give a useful insight on how the land and marine biospheres respond to climate changes.

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B. Malaizé, Département de Géologie et Océanographie, UMR 5805, EPOC, Avenue des Facultés, F33405 Talence, France. (e-mail: b.malaize@geocean.u-bordeaux.fr)

J. Jouzel and D. Paillard, Laboratoire des Sciences du Climat et de l'Environnement, UMR CEA-CNRS, CE Saclay, 91191 Gif-sur-Yvette, France. (e-mail· jouzel@lsce saclay.cea.fr; paillar@lsce.saclay.cea.fr)

D. Raynaud, Laboratoire de Glaciologie et Géophysique de l'Environnement, CNRS, BP96, 38402 St. Martin d'Hères, France. (e-mail: domraynaud@glaciog.ujf-grenoble.fr)

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