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Atmospheric oxygen 18 and sea-level changes

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Abstract

Past isotopic composition of atmospheric oxygen ($\delta^{18}O_{atm}$) can be inferred from the analysis of air bubbles trapped in ice caps. The longest record covers the last 420 ka (thousand of years) at the Vostok site in East Antarctica. It shows a strong modulation by the precession and striking similarities, but also noticeable differences, with the deep-sea core oxygen 18 record from which changes in the oxygen content of sea-water ($\delta^{18}O_{sw}$) and in sea-level can be derived. Indeed, $\delta^{18}O_{atm}$ is driven by complex fractionation processes occuring during respiration and photosynthesis. Both $\delta^{18}O_{atm}$ and its difference with respect to $\delta^{18}O_{sw}$ (the Dole effect) are influenced by factors such as the ratio of oceanic and terrestrial productivities which may have significantly changed between different climates. Also, the response time of $\delta^{18}O_{atm}$ to oceanic changes should be taken in consideration but this parameter itself depends on biospheric activity. We review the various aspects of the link between the $\delta^{18}O_{atm}$ and the $\delta^{18}O_{sw}$ signals. We also examine the approach followed by Shackleton (Science (2000)) for deriving sea-level change from the $\delta^{18}O_{atm}$ Vostok record, assuming that the phase between this record and insolation changes is constant and that the Dole effect is a fraction of the precessional component of the $\delta^{18}O_{atm}$ signal. Glaciological constraints on the Vostok chronology and the complexity of the Dole effect show that those two assumptions are quite probably too simplistic. © 2001 Published by Elsevier Science Ltd.

1. Introduction

The $^{18}O/^{16}O$ ratio of the oxygen in the atmosphere (denoted $\delta^{18}O_{atm}$ and expressed in the δ scale relative to the V-SMOW standard which has an $^{18}O/^{16}O$ isotope ratio of 2005.2×10^{-6}) is naturally enriched with respect to the oxygen isotopic ratio in sea water ($\delta^{18}O_{sw}$). Most of this enrichment is due to the activity of marine and terrestrial biospheres that transmit the $\delta^{18}O$ signature of oceanic and continental waters to the atmosphere via respiration and photosynthesis. The difference between $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$, the Dole effect ($\Delta = \delta^{18}O_{atm} - \delta^{18}O_{sw}$), has a present-day value of ~ 23.5‰ (Kroopnick and Craig, 1972).

Our current knowledge and interpretation of $\delta^{18}O_{atm}$ and of the Dole effect in the past rely on measurements of the isotopic composition ($^{18}O/^{16}O$) of air bubbles trapped in ice cores. The most exciting result of the pioneering work performed on the Dome C Antarctic ice core (Bender et al., 1985) was to reveal similar variations of $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$ during the last deglaciation. This finding suggested that despite the complexity of the fractionation processes involved (see below), $\delta^{18}O_{atm}$ might be mainly governed by the changes in $\delta^{18}O_{sw}$ resulting from the waxing and waning of the Northern Hemisphere ice sheet. It opened the possibility of using the $\delta^{18}O_{atm}$ record to correlate ice cores and deep-sea core records, once accounted for the turnover time of atmospheric oxygen. This possibility was fully exploited over the last glacial—interglacial cycle, first to examine the timing of CO_2 and ice volume changes during the last deglaciation (Sowers et al., 1991) and second (Sowers et al., 1993) to place Vostok and deep-sea core records on a common temporal framework back to 135 kyr BP (thousands of years Before Present).

From the small glacial–interglacial variability in the Dole effect, Bender et al. (1994) inferred that relative rates of primary production in the land and marine realms have been relatively constant. They also pointed out that most periodic variability of the Dole effect is in the precession band. The presence of a strong precessional signal, directly observed in the $\delta^{18}O_{atm}$ record, was remarkably confirmed once this record was extended over the last two glacial–interglacial cycles

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(Jouzel et al., 1993, 1996) and more recently back to 420 kyr BP (Petit et al., 1999). However, the $\delta^{18}O_{atm}$ record then showed less similarities with $\delta^{18}O_{sw}$ than observed for the first cycle (Jouzel et al., 1996). In particular, the Vostok data revealed an unexpected high variation of the Dole effect during the insolation maximum around 175 kyr BP (Mélières et al., 1997; Malaizé et al., 1999).

The strong modulation of $\delta^{18}O_{atm}$ by the precession has implications for dating purposes that have been explored in two different ways. Petit et al. (1999) noted the obvious similarity between $\delta^{18}O_{atm}$ and insolation. These authors extended the Vostok timescale based on the glaciological model developed by Ritz (1992), verifying that the dated $\delta^{18}O_{atm}$ record presents the correct number of precessional cycles (see Table 1 of Petit et al., 1999). In contrast, Shackleton (2000) applied the orbital tuning approach devised by Imbrie et al. (1984) to the Vostok $\delta^{18}O_{atm}$ record assuming that its phase relative to precession is constant all along the record (see also Pépin et al., in press). Shackleton then derived a $\delta^{18}O_{sw}$ ocean water record as the difference between the Vostok $\delta^{18}O_{atm}$ record and the Dole effect. For this purpose, he assumed that Dole effect changes are linearly related to orbital parameters and can be considered to be a part of the precession component of $\delta^{18}O_{atm}$ (the other part being related to the ice volume signal) with a small additional obliquity component.

In the present article we examine some of the assumptions made by Shackleton (2000) concerning the link between $\delta^{18}O_{atm}$ and sea-level change. This analysis is largely based on the results of two ongoing studies. The first one concerns an inverse modeling method developed to date the Vostok ice core, which suggests that the phase between $\delta^{18}O_{atm}$ and precession is not constant through time (Parrenin et al., in press). The second one deals with the modeling of the Earth's Dole effect using a hierachy of models (Hoffmann et al., submitted) and points to the complexity of the mechanisms involved.

2. Glaciological constraints on the phasing between $\delta^{18}O_{atm}$ and insolation

In this section we examine the assumption of constant phase between $\delta^{18}O_{atm}$ and insolation. Several approaches have been used to date the Vostok ice core: the glaciological timescale was developed using an ice flow model (Ritz, 1992) and the assumption that accumulation changes depend on temperature changes. The latter assumption is based on the fact that the amount of precipitating snow in Central Antarctica first depends on the amount of water vapor in the atmosphere and can be related to the derivative of the saturation vapor pressure (Lorius et al., 1985; Ritz, 1992; Jouzel et al., 1993, 1996; Petit et al., 1999). The age of the gas, younger than the age of the ice, due to the fact that air bubbles are trapped when the firn closes off at depth, is estimated using the firnification model of Barnola et al. (1991). Vostok chronologies were also derived by correlation with marine records dated by orbital tuning using either dust content (Petit et al., 1990), temperature changes (Shackleton et al., 1992), $\delta^{18}O_{atm}$ (Sowers et al., 1993) or CO_2 concentration (Raymo and Horowitz, 1996), and by direct orbital tuning (Waelbroeck et al., 1995; Salamatin et al., 1998; Shackleton, 2000).

Each of these dating methods has advantages and drawbacks. Orbitally tuned chronologies have associated uncertainties relatively constant all along the record with an estimated range of $\sim$4–5 ka (Martinson et al., 1987; Waelbroeck et al., 1995). However, the use of this method limits the climatic information that can be inferred from the phasing between orbitally tuned series and insolation curves or other climate records. Such a limitation also holds true for chronologies obtained by correlation to a reference time series. Moreover, both orbitally tuned and correlated chronologies may imply locally unrealistic accumulation changes that prevent to correctly estimate the duration of successive climatic events. Indeed, once accounted for the accumulation change due to the temperature change, the accumulation values derived from Shacketon's chronology may jump by more than a factor of two between successive time periods (for example, there is such a large jump around 135 kyr BP). Although changes in the oceanic origin of the Vostok precipitation, in the atmospheric circulation and in other processes may significantly affect snow accumulation in central Antarctica such large jumps are difficult to explain.

Instead, glaciological timescales are consistent with respect to ice flow laws and to the assumption that accumulation changes can be related to temperature changes. However, the dating error increases with depth and, in order to keep this error acceptable, control points (i.e. depths with assigned ages) are needed along the profile. Those control points are themselves derived from correlation with other records or with insolation curves. Thus a glaciological timescale is not truly independent of some assumptions on the phasing between the ice core records and other time series. Also, a glaciological timescale is based on assumptions that are subject to discussion such as the use of the present-day temporal deuterium—temperature relationship to interpret the Vostok deuterium record and infer accumulation changes. Moreover, it relies on poorly defined values of the melting and the sliding rates at the bottom of the ice sheet. As a result, the uncertainty of the current Vostok GT4 glaciological timescale may be as high as $\pm$15 ka (Petit et al., 1999).

Parrenin et al. (in press) recently proposed to combine the advantages of the glaciological approach and of the orbital tuning in order to minimize the limitations mentioned above. Following Petit et al. (1999), these authors do not make any assumption about the phase relationship between insolation and Vostok records but simply assume that the number of precessional cycles in those records can be correctly counted. This appears straightforward considering how clearly this cycle is imprinted in the $\delta^{18}O_{atm}$ series (Jouzel et al., 1996; Petit et al., 1999). To express this assumption the ice and gas chronologies are assigned to pass through a succession of large doors with a width of ± 6 ka ($\sim 1/4$ of precessional cycle).

To define these doors, Parrenin et al. (in press) first derived an orbitally tuned Vostok timescale using the same information (orbital signature present in the $\delta^{18}O_{atm}$ record) as Shackleton (2000) and the same assumption of constant phase between this record and $\delta^{18}O_{atm}$ insolation changes. The two timescales also use the same scaling method between orbitally derived control points (at a given depth, the annual layer thickness is that of GT4 multiplied by the ratio of the durations between the two neighboring control points). Surprisingly, they, however, significantly differ over some intervals with differences higher than 4 ka at the start of termination II (135 kyr BP). We attribute this large difference to the fact that Shackleton's timescale is not optimally adjusted to the insolation signal over this period.

One major advantage of this inverse method is the determination of confidence intervals, not only on the ages themselves, but also on the duration of given events. The phase lags between $\delta^{18}O_{atm}$ and insolation can then be estimated and do indeed significantly vary with time. Over the last two climatic cycles, this phase lag varies from ~ 4 to 10 ka. This is further illustrated by the large range (from 19 to 31 ka) derived from the inverse method for the duration of the successive $\delta^{18}O_{atm}$ cycles (Fig. 1) whereas this duration is assumed to be equal to 23 ka when the dating is performed by orbital tuning. Even if the inverse method has its own uncertainties linked, in particular, to the parameterization of accumulation change (Parrenin et al., in press), those results seriously contradict the assumption of constant phase between $\delta^{18}O_{atm}$ and insolation which is a key assumption of the orbital tuning approach.

One would like to find arguments either supporting or allowing to discard this assumption of constant phase. For example, the presence of a strong precessional signal in the methane record suggests that methane concentrations are quite probably influenced by moonsonal activity (Chappellaz et al., 1990). An examination of the phasing between $\delta^{18}O_{atm}$ and methane indicates that the phase between the two records is far from being constant. Chappellaz et al. (1997) noted that the lag between the methane peak and the following $\delta^{18}O_{atm}$ minima varied between 4 and 8 ka over the last climatic cycle but the range of variation is even larger (from practically no lag up to 10 ka) for the full record (note that those estimates are essentially independent of the accuracy of the timescale). Obviously, this variability may simply reflect that the processes involved are more complex for methane, influenced by high latitude production and possibly by rapid climate changes (Chappellaz et al., 1993), than for $\delta^{18}O_{atm}$. Alternatively, it could be viewed as reflecting complex processes for both atmospheric quantities and thus be taken as an additional argument indicating that the assumption made by Shackleton (2000) is too simplistic.

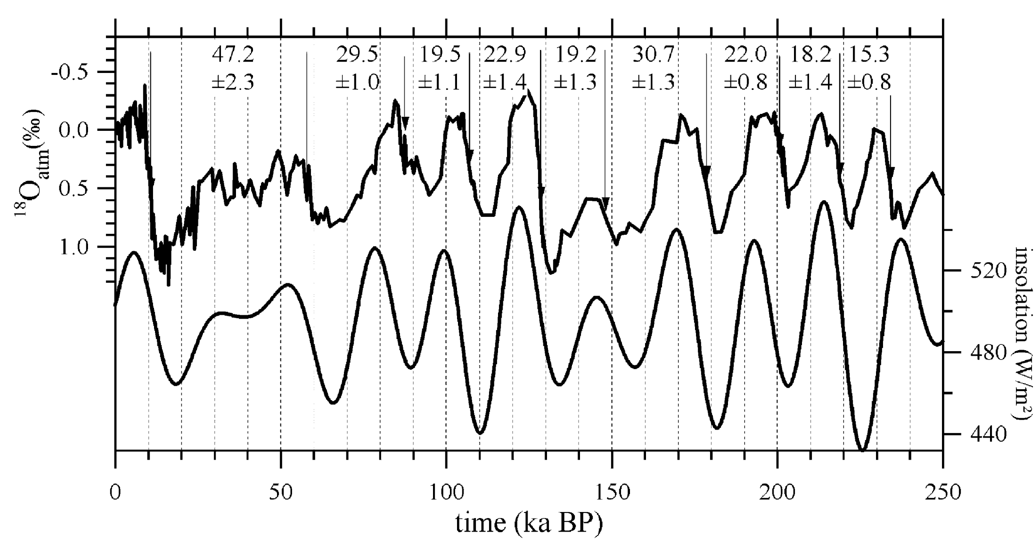

Fig. 1. Duration of the $\delta^{18}O_{atm}$ cycles as estimated from a glaciological timescale derived by an inverse method; the lower curve corresponds to the June 21 insolation change at 65°N (adapted from Parrenin et al., in press).

3. The complexity of the Earth's Dole effect

In the following, we test the hypothesis that the Dole effect could be modeled as a fraction of the precessional component of the $\delta^{18}O_{atm}$ signal. The isotopic enrichment of atmospheric oxygen with respect to seawater mainly results from the fact that during respiration most species (both autotrophic respiration, that is by plants, and heterotrophic respiration, that is by bacteria and other micro-organisms) preferentially use the light ^{16}O oxygen isotope, leading to an enrichment in the atmospheric ^{18}O. The discrimination during marine respiration differs when this respiration takes place within or below the euphotic zone whereas the terrestrial respiratory fractionation depends on the type of respiration (dark respiration, Mehler reaction or photorespiration). Accounting for these various factors, Bender et al. (1994) estimated that global isotope effects amount to about 18.9‰ for the respiration of the marine biosphere and to about 18.0‰ for the respiration of the terrestrial biosphere. Instead, the $\delta^{18}O$ of the oxygen rejected during photosynthesis has the same value as the oxygen of the source water (i.e. a modern value close to 0‰). The marine Dole effect is thus estimated to be 18.9‰. The source water for photosynthesis on land is taken from the leaves. The leaf $\delta^{18}O$ ($\delta^{18}O_{leaf}$) depends on many local parameters, such as the relative humidity and the $\delta^{18}O$ of water in the soil and in the atmosphere. In addition, fractionation processes occuring at the surface of the leaf include equilibrium and kinetic effects that contribute to enhance the terrestrial Dole effect and depend on temperature and relative humidity. Using a value of 4.4‰ for the productivity-weighted yearly average $\delta^{18}O$ of leaf water (Farquhar et al., 1993), Bender et al. (1994) ended up with a terrestrial Dole effect of 22.4‰. These authors also mentioned that the isotopic exchange between CO_2 and O_2 in the stratosphere slightly lowers $\delta^{18}O_{atm}$ (by $\varepsilon = {\sim}0.4$‰). The total Dole effect $\mathit{\Delta}$ is then estimated as $\mathit{\Delta} = (\Delta_o P_o + \Delta_t P_t)/(P_o + P_t) - \varepsilon$, where $\mathit{\Delta}_o$ and $\mathit{\Delta}_t$ are, respectively, the oceanic and terrestrial Dole effects and P_o and P_t the oxygen fluxes of marine and terrestrial productivities. Accounting for the currently estimated annual average values of P_o and P_t (10.6 and 20.4×10^{15} moles/year, respectively), they obtained a best present-day estimate of 20.8‰ for $\mathit{\Delta}$. They noted the difficulty to correctly predict the observed Dole effect of 23.5‰ (Kroopnick and Craig, 1972) and specifically pointed out to the difficulty of estimating the $\delta^{18}O$ of leaf water as the current biggest uncertainty.

The Vostok $\delta^{18}O_{atm}$ record has been used to estimate the Dole effect over the last (Bender et al., 1994) and the previous (Malaizé et al., 1999) climatic cycles and to infer information on the ratio of past marine and terrestrial productivities. This approach requires to account for the turnover time of atmospheric oxygen, τ, calculated as the ratio of the oxygen atmospheric inventory, 3.7×10^{19} moles, divided by the total oxygen fluxes, $P_o + P_t$, i.e $\tau = 1.2$ ka for the values of present-day productivities adopted by Bender et al. (1994). The method is illustrated in Fig. 2a which shows our current best estimate of the Dole effect over the last two climatic cycles calculated using the $\delta^{18}O_{sw}$ record of Waelbroeck et al. (this volume) and the $\delta^{18}O_{atm}$ Vostok record dated by an inverse method (Parrenin et al., in press). Fig. 2b illustrates the approach of Shackleton (2000) who, alternatively, proposed to derive sea-level change simply assuming that the changes in the Dole effect can be linearly related to orbital parameters.

The success of such a simplification might not be warranted given the complexity of the processes involved. The oxygen fluxes associated to marine and terrestrial productivities, as well as the isotopic content of leaf water, are highly variable both spatially and temporally. Those variabilities are well illustrated by the modeling approach recently followed by Hoffmann et al. (submitted) in which the $\delta^{18}O$ in precipitation and in water vapor is simulated by atmospheric general circulation models (AGCMs) implemented with water isotopes. $\delta^{18}O_{leaf}$ is then calculated assuming steady state conditions during plant respiration (Craig and Gordon, 1965). Productivity on land is estimated using Terrestrial Biochemical Models that predict the biome distribution. The marine Dole effect is calculated using a global ocean carbon cycle model that includes a plankton sub-model to estimate the net marine primary productivity and oxygen consumption. The distribution of oceanic $\delta^{18}O$ is also accounted for. The results (between 20.9‰ and 22.4‰) are closer to the observations (23.5%) than the best guess of Bender et al. (1994) of 20.8‰, mainly due to a larger global mean $\delta^{18}O_{leaf}$ computed by both isotopic AGCMs (Hoffmann et al., in press). The remaining difference between observations and model results may be attributed to the fact that the daily cycle of the relative humidity, a parameter that strongly influences $\delta^{18}O_{leaf}$, is missing. Accounting for this daily cycle in a reasonable way, eliminates this inconsistency (Hoffmann et al., submitted).

The implicit assumption of Shackleton (2000) is that the terrestrial and/or the marine productivity are either constant through time or entirely driven by precessional changes through the influence of low latitude circulation changes (principally monsoonal precipitation and wind forcing) on this productivity. Otherwise, it is unlikely that a combination of the terrestrial and the oceanic contribution to the total Dole effect could result in a precessional signal. However, model results show that due to combined changes in high latitude precipitation $\delta^{18}O$ and productivity, the variability in the terrestrial Dole effect forced by high latitudes (and thus not driven by precession) is far from being negligible. Fig. 3 shows the zonal contribution to the terrestrial Dole effect both

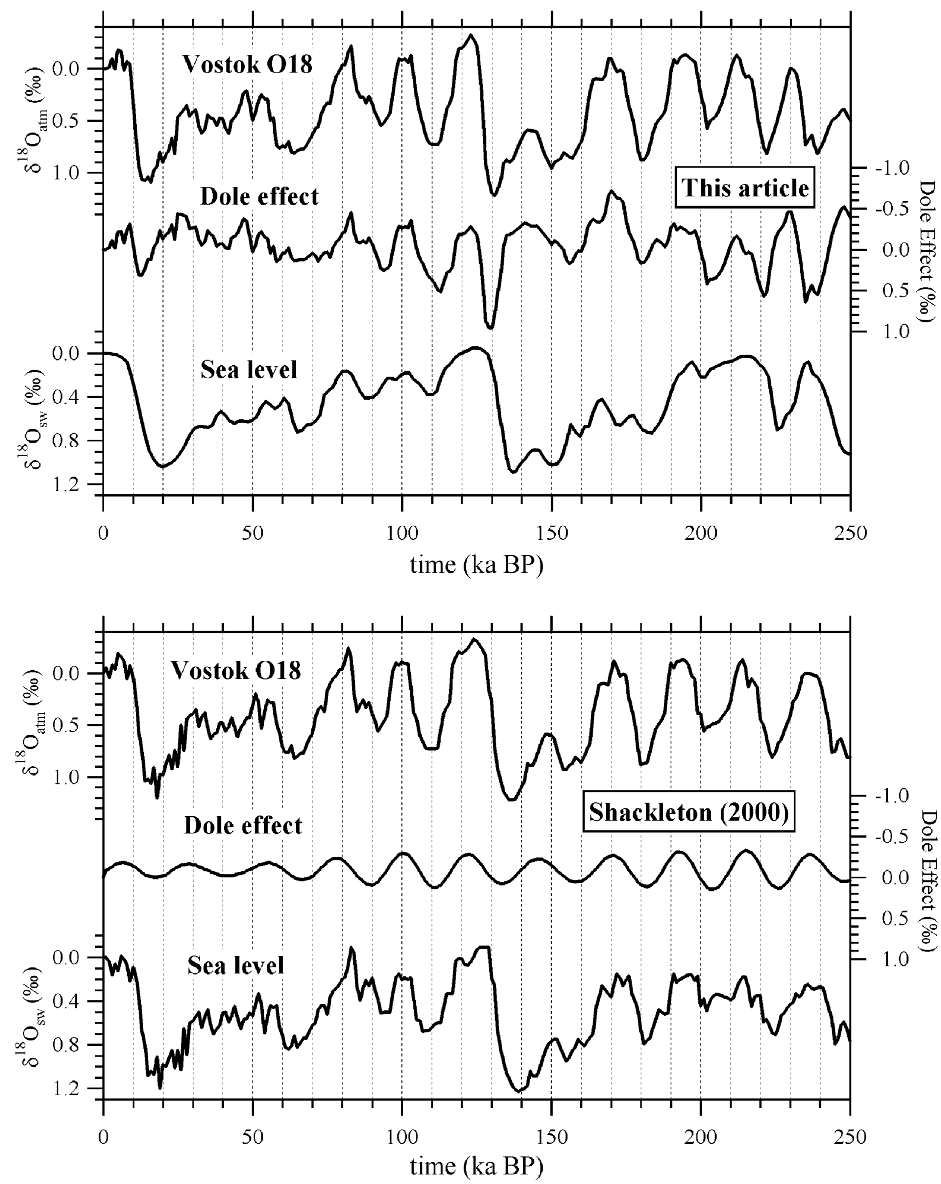

Fig. 2. This figure showing $\delta^{18}O_{atm}$, the Dole effect and $\delta^{18}O_{sw}$ over the last 250 ka illustrates the approach adopted in this article and in Shackleton (2000). In this article (upper panel), we follow Bender et al. (1994) in calculating the Dole effect (see text) whereas Shackleton (2000) assumes that the Dole effect can be expressed as a function of orbital parameters.

for modern and full glacial conditions. Though low latitudinal insolation is quite similar for both time periods, large changes in the latitudinal contribution to the Dole effect have been simulated: considerably lower productivity in high latitudes during glacial times is giving more weight to low latitudes characterized by quite enriched leaf water. The terrestrial Dole effect under glacial conditions is thus larger (by about 3‰ in the upper simulation) compensating to a large extent the globally lower terrestrial productivity. The Dole effect remains approximately stable.

Obviously, given the various uncertainties and the complexity of the entire system, this modeling of the Dole effect is not capable to prove that the view of Shackleton (2000) is wrong. However, it helps to identify at least one mechanism influencing the Dole effect which is not driven by precession. One can also think of transient effects such as those linked with plant colonization of the boreal landscape saturated with $\delta^{18}O$ depleted water following glacial terminations which should temporarily reduce the Dole effect.

The same is true for the oceanic part of the computation. Shackleton's assumption would require either no or purely precessionally driven changes in the marine productivity contribution. Instead, Malaizé et al. (1999) invoked changes in marine productivity to explain the unexpected minimum of the Dole effect that occurred around 175 kyr BP (marine stage 6.5 event).

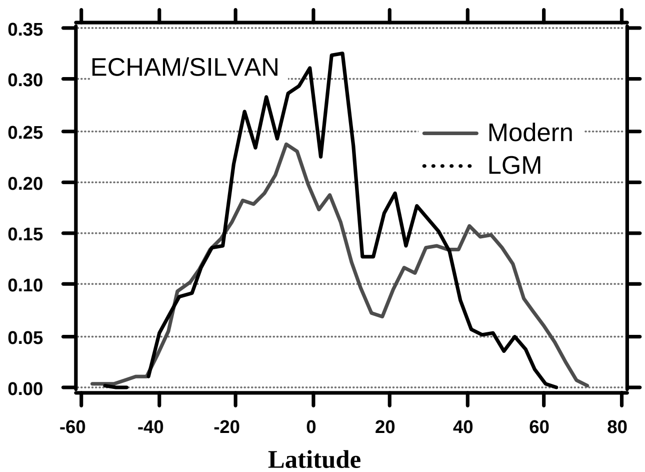

Fig. 3. The terrestrial Dole effect as a function of latitude for present-day conditions (continuous line) and for the last glacial maximum (relative contribution per 10° of latitude).

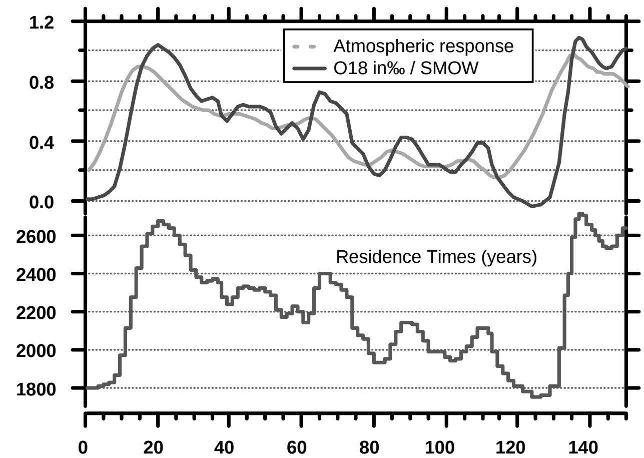

Fig. 4. The lower curve shows the atmospheric oxygen turnover time (in years) derived from model estimates of total productivity and scaled with respect to sea-level change (see text). The upper curves show the oceanic $\delta^{18}O$ record (solid line adapted from Waelbroeck et al., this volume) and the atmospheric response (dotted line).

Alternatively, a recent GCM simulation at 175 kyr BP recently showed that despite glacial conditions monsoonal activity was strong (Masson et al., 2000) which might explain the unusual minimum in the Dole effect. As pointed out at the EPILOG meeting by Bard, the period around stage 6.5 should give a good opportunity to test the validity of sea-level reconstructions. Shackleton's approach then predicts a relatively high sea-level stand (Fig. 2b) whereas the study of cave deposits indicates lower sea level (Bard et al. (2000)).

Modeling the Dole effect for past climates (Hoffmann et al., in preparation; see also Leuenberger, 1997) shows that the total oxygen fluxes, $P_o + P_t$ (i.e., the sum of the marine and terrestrial productivities) and in turn, the turnover time of atmospheric oxygen, τ, are climate dependent. Model estimates of marine and terrestrial productivities show that, largely due to a decrease of terrestrial productivity, $P_o + P_t$ can decrease by up 30% from present-day to glacial conditions (Hoffmann et al., pers comm). To illustrate the consequences of a variable τ, we simply scaled $P_o + P_t$ by a linear function of sea level between the present-day and glacial values and used a box model to compute the atmospheric response to an oceanic $\delta^{18}O$ change. This atmospheric response (and thus the oceanic fraction of the Dole effect) shows a variable lag with respect to the oceanic signal and is smoother than this signal (Fig. 4). Although this approach is very preliminary, it suggests that τ is larger than the value of 1.2 ka adopted by Bender et al. (1994) and variable. Still those variations are not large enough to explain the variable phase of $\delta^{18}O_{atm}$ with respect to insolation found by Parrenin et al. (in press) as the variability of this phase is much higher than the variability of τ.

That τ varies with time has implications for the use of $\delta^{18}O_{atm}$ as a tool to correlate the ice core and deep-sea core record (Sowers et al., 1991, 1993; Broecker and Henderson, 1998; Petit et al., 1999). Similarities between the two records first result from the fact that $\delta^{18}O_{sw}$ changes are fully transmitted to the atmosphere as far as the oceanic part of the Dole effect is concerned. Also, the isotopic content of tropical and equatorial precipitation is directly influenced by the change in the isotopic composition of the oceanic source (Jouzel et al., 1994). In turn this explains why the overall variability of the Dole effect is relatively weak. It also explains why the amplitudes of $\delta^{18}O_{atm}$ changes parallel to those of $\delta^{18}O_{sw}$ changes when those changes are large as for deglaciations (Petit et al., 1999). Such parallel behavior is thus not surprising despite the anticipated large changes in the ratio of terrestrial and oceanic productivities, and thus in the Dole effect, during terminations. Still, the interpretation would gain in accuracy if the variation of τ (which can vary from about 2.7–1.7 ka during such a transition according to the model results) was correctly taken into account.

4. Conclusion

In this article we have reviewed the current understanding of the processes that drive the variations in atmospheric oxygen 18 through time and how they relate to sea-level changes. This review has been done in the light of the study of Shackleton (2000) who elegantly derived a sea-level curve from this $\delta^{18}O_{atm}$ record. We critically assessed the two key assumptions that underline this approach, namely that the phase of $\delta^{18}O_{atm}$ relative to precession is constant along the entire record and that changes in the Dole effect can be assumed to be a fraction of the $\delta^{18}O_{atm}$ precessional component. The results obtained from an inverse method applied to date the Vostok ice core, fully accounting for glaciological constraints (Parrenin et al., in press) are difficult to reconcile with the assumption of a constant phase between $\delta^{18}O_{atm}$ and precession. Our current knowledge of the various processes governing changes in $\delta^{18}O_{atm}$ and in the Dole effect either from the pioneering work of Bender et al. (1994) or from recent model developments (Hoffmann et al., in press) points out to their complexity and makes difficult to believe that changes in the Dole effect can be simply parameterized as a linear response to precession. Also, the fact the oxygen atmospheric turnover time was significantly larger during glacials than initially suggested by Bender et al. (1994) and variable through time (Leuenberger, 1997; Hoffmann et al., pers comm) should be taken into account. This also holds true when the $\delta^{18}O_{atm}$ record is used to correlate ice cores and deep-sea cores as done, for example, for glacial terminations (Petit et al., 1999).

To go further in the understanding of the relationship between $\delta^{18}O_{atm}$ and $\delta^{18}O_{sw}$, various possibilities can be explored. First, one strong support in favor of Shackleton's approach is that sea-level estimates extracted from the $\delta^{18}O_{atm}$ Vostok record agrees well with sea-level observations over the last climatic cycle. Such a comparison should be extended further back in time in focusing in particular on the period around 175 kyr BP for which there is a possible disagreement with other methods. Second, the problem of ice core dating is crucial. However, despite the application of an inverse method, the Vostok core is not ideally suited to account for glaciological constraints. This is mainly because this drilling site is not located on a dome so that ice found at depth comes from snow precipitated upslope, seriously complicating the estimation of accumulation changes through time (Parrenin et al., in press). Instead, snow is of local origin, both for the japenese Dome F ice core which extends back to ~330 kyr BP (Dome F group, 1998) and for the Dome C EPICA core drilling now underway in the frame of a European collaboration. Records from those two cores should allow a more accurate assessment of the phasing between $\delta^{18}O_{atm}$ and precession than provided by Vostok data. Third, the complex modeling developed by Hoffmann et al. (submitted) opens the way to estimate the Dole effect for various time slices in the past. A series of experiments are currently being conducted (Hoffmann et al., in preparation) which should shed light on the relationship between the precession and the Dole effect. Also, it is planned to equip the CLIMBER model, a model of intermediate complexity which allows performing long transient simulations (Petoukov et al., 2000), with isotopic modules describing both water and oxygen isotopic cycles. This should allow to examine the link between the Dole effect and the precession and provide a continuous estimate of the variation in the oxygen turnover time.

References

Bard, E., Antonioli, F., Rostek, F., Silenzi, F., Schrag, D., 2000. The penultimate glaciation as viewed from geochemical data measured in a submerged Speleothem from Italy, EOS Supplement, AGU Spring meeting.

Barnola, J.M., Pimienta, P., Raynaud, D., Korotkevich, Y.S., 1991. CO_2 climate relationship as deduced from the Vostok ice core: a re-examination based on new measurements and on a re-evaluation of the air dating. Tellus 43B, 83–91.

Bender, M., Labeyrie, L.D., Raynaud, D., Lorius, C., 1985. Isotopic composition of atmospheric O_2 in ice linked with deglaciation and global primary productivity. Nature 318, 349–352.

Bender, M., Sowers, T., Labeyrie, L.D., 1994. The Dole effect and its variation during the last 130,000 years as measured in the Vostok core. Glob. Biog. Cycles 8, 363–376.

Broecker, W.S., Henderson, G.M., 1998. The sequence of events surrounding termination II and their implications for the causes of glacial interglacial CO_2 changes. Paleoceanography 13, 352–364.

Chappellaz, J., Barnola, J.-M., Raynaud, D., Korotkevich, Y.S., Lorius, C., 1990. Ice-core record of atmospheric methane over the past 160,000 years. Nature 345, 127–131.

Chappellaz, J., Blunier, T., Raynaud, D., Barnola, J.M., Schwander, J., Stauffer, B., 1993. Synchronous changes in atmospheric CH_4 and greenland climate between 40 and 8 kyr BP. Nature 366, 443–445.

Chappellaz, J., Brook, E., Blunier, T., Malaizé, B., 1997. CH_4 and $\delta^{18}O$ of records from Greenland ice: a clue for stratigraphic disturbance in the bottom part of the Greeland Ice Core Project and the Greeland Ice Sheet Project 2 ice-cores. Journal of Geophysical Research 102, 26 547–26 557.

Craig, H., Gordon, A., 1965. Deuterium and oxygen 18 variations in the ocean and the marine atmosphere. In: Consiglie Nazionale delle Ricerec (Eds.), Stable isotopes in Oceanic Studies and Paleotemperatures, Laboratorio di Geologia Nucleare, Pisa, Italy, pp. 9–130.

Farquhar, G.D., Lloyd, J., Taylor, J.A., Flanagan, L., Syvertsen, J.P., Hubick, K.T., Wong, S., Ehleringer, J.R., 1993. Vegetation effects on the isotopic composition of oxygen in atmospheric CO_2. Nature 363, 439–443.

Hoffmann, G., et al., A model for the Earth's Dole effect, global geochemical cycle, submitted for publication.

Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L., Shackleton, N.J., 1984. The orbital theory of Pleistocene climate: support from a revised chronology of the marine $\delta^{18}O$ record, In: A.L. Berger et al. (Eds.), Milankovitch and Climate, Part 1. D. Riedel, Hingham, MA.

Jouzel, J., et al., 1993. Extending the Vostok ice-core record of paleoclimate to the penultimate glacial period. Nature 364, 407–412.

Jouzel, J., Koster, R.D., Suozzo, R.J., Russell, G.L., 1994. Stable water isotope behaviour during the LGM: a GCM analysis. Journal of Geophysical Research 99, 25 791–25 801.

Jouzel, J. and 13 others, 1996. Climatic interpretation of the recently extended Vostok ice records. Climate Dynamics 12, 513–521.

Kroopnick, P., Craig, H., 1972. Atmospheric oxygen: isotopic composition and solubility fractionation. Science 175, 54–55.

Leuenberger, M., 1997. Modeling the signal transfer of seawater $\delta^{18}O$ to the $\delta^{18}O$ of atmospheric oxygen using a diagnostic box model for terrestrial and marine biosphere. Journal of Geophysical Research 102, 26 841–26 850.

Lorius, C., Jouzel, J., Ritz, C., Merlivat, L., Barkov, N.I., Korotkevitch, Y.S., Kotlyakov, V.M., 1985. A 150,000-year climatic record from Antarctic ice. Nature 316, 591–596.

Malaizé, B., Paillard, D., Jouzel, J., Raynaud, D., 1999. The Dole effect over the last two glacial–interglacial cycles. Journal of Geophysical Research 104, 14 199–14 208.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore, T.C., Shackleton, N.J., 1987. Age dating and the orbital theory of the Ice Ages: development of a high-resolution 0–300,000 years chronostratigraphy. Quaternary Research 27, 1–30.

Masson, V., Braconnot, P., Cheddadi, R., Jouzel, J., Marchal, O., de Noblet, N., 2000. Simulation of intense monsoons under glacial conditions. Geophysical Research Letters 27, 1747–1750.

Mélières, M.A., Rossignol-Strick, M., Malaizé, B., 1997. Relation between low latitude insolation and $\delta^{18}O$ change of atmospheric oxygen for the last 200 kyr as revealed by Mediterranean sapropels. Geophysical Research Letters 24, 1235–1238.

Parrenin, F., Jouzel, J., Waelbroeck, C., Ritz, C., Barnola, J.M., Dating the Vostok ice core by an inverse method. Journal of Geophysical Research, in press.

Pépin, L., Raynaud, D., Barnola, J.M., Loutre, M.F. Hemispheric roles of climate forcings during glacial–interglacial transitions as deduced from the Vostok record and LLN-2D model experiments. Journal Geophysical Research, in press.

Petit, J.R., Mounier, L., Jouzel, J., Korotkevitch, Y., Kotlyakov, V., Lorius, C., 1990. Paleoclimatological implications of the Vostok core dust record. Nature 343, 56–58.

Petit, J.R., et al., 1999. Climate and atmospheric history of the past 420,000 years from the Vostok Ice Core, Antarctica, Nature 399, 429–436.

Petoukov, V., Ganapolski, A., Brokvin, V., Claussen, M., Eliseev, A., Kubatzki, C., Rahmstorf, S., 2000. CLIMBER-2: a climate system model of intermediate complexity. Part I: model description and performance for present climate. Climate Dynamics 16, 1–17.

Raymo, M.E., Horowitz, M., 1996. Organic carbon paleo-pCO_2 and marine-ice core correlations and chronology. Geophysical Research Letters 23 (4), 367–370.

Ritz, C., 1992. Un modele thermo-mecanique d'evolution pour le bassin glaciaire Antarctique Vostok-Glacier Byrd: sensibilite aux valeurs des parametres mal connus.

Salamatin, A.N., Lipenkov, V.Y., Barkov, N.I., Jouzel, J., Petit, J.R., Raynaud, D., 1998. Ice core age dating and paleothermometer calibration on the basis of isotopes and temperature profiles from deep boreholes at Vostok station (East Antarctica). Journal of Geophysical Research 103, 8963–8977.

Shackleton, N.J., 2000. The 100,000-year Ice-age cycle identified and found to lag temperature, CO_2 and orbital eccentricity, Science.

Shackleton, N.J., Le, J., Mix, A., Hall, M., 1992. Carbon isotope records from Pacific surface waters and atmospheric carbon dioxide. Quaternary Science Reviews 11, 387–400.

Sowers, T., Bender, M., Raynaud, D., Korotkevich, Y.S., Orchado, J., 1991. The $\delta^{18}O$ of atmospheric O_2 from air inclusions in the Vostok oce core: timing of CO_2 and ice volume change during the penultimate deglaciation. Paleoceanography 6, 669–696.

Sowers, T., Bender, M., Labeyrie, L.D., Jouzel, J., Raynaud, D., Martinson, D., Korotkevich, Y.S., 1993. 135,000 year Vostok—SPECMAP common temporal framework. Paleoceanography 8, 737–766.

Waelbroeck, C., Jouzel, J., Labeyrie, L., Lorius, C., Labracherie, M., Stievenard, M., Barkov, N.I., 1995. A comparison of the Vostok ice deuterium record and series from Southern Ocean core MD 88-770 over the last two glacial–interglacial cycles. Climate Dynamics 12, 113–123.

Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J.C., McManus, J.F., Balbon, E., Labracherie, M. Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. Quaternary Science Reviews, this volume.