

## RECONSTRUCTING PLEISTOCENE CLIMATIC CHANGE FROM THE OXYGEN ISOTOPE COMPOSITION OF SEDIMENTS: A REVIEW

by

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### ABSTRACT

The isotope composition of ocean sediments is the chief data source for Pleistocene climatic changes. It is generally believed that the  $^{18}\text{O}/^{16}\text{O}$  ratio of a sample indicates the global total of glacial ice at the time the sample was deposited. This is roughly correct, but numerous complicating factors limit the accuracy of the isotope proxies.

Oxygen isotopes in ocean sediments provide a continuous record of climate changes over the past several hundred thousand years (see, for example, Hays and others (1976) or Emiliani (1978)). The ratio of  $^{18}\text{O}$  to  $^{16}\text{O}$  is measured in the calcium carbonate shells of small marine fossils and the isotope composition is expressed at  $\delta^{18}\text{O}$ , the deviation in parts per thousand of the isotope ratio of the sample compared with the

isotope ratio of a convenient standard. For the most part, the  $\delta^{18}\text{O}$  record is similar in ocean sediment cores recovered from different parts of the world (Fig.1). It is generally believed that this record reflects the global total of glacial ice;  $^{16}\text{O}$  is preferentially deposited onto ice sheets so that higher values of  $\delta^{18}\text{O}$  in the ocean sediments indicate higher ice volume. In the following discussion we evaluate this interpretation and conclude that it is correct to "first order", but that (1) effects other than ice volume may control part of the amplitude of the  $\delta^{18}\text{O}$  signal, and (2) even considering the effect of volume alone, the time record of the  $\delta^{18}\text{O}$  signal is probably distorted compared with the true ice volume record.

Originally it was believed that  $\delta^{18}\text{O}$  measures the

### TIME SERIES

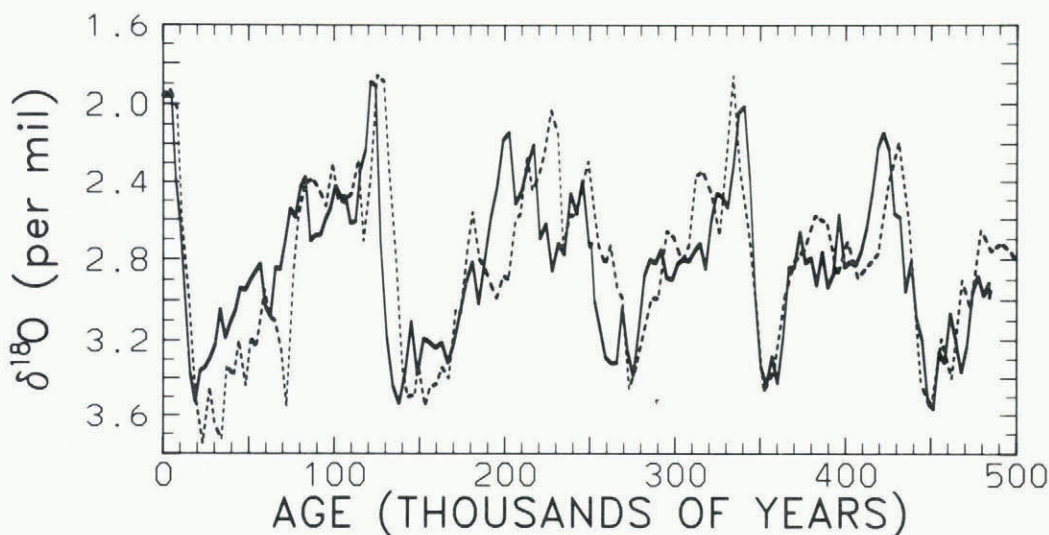


Fig.1. Oxygen-isotope composition of ocean-floor sediment samples vs age of samples. Note that time increases to the left and  $^{18}\text{O}$  content increases downward; thus lower ice volume and warmer climates are indicated by  $\delta^{18}\text{O}$  values near the top of the graph. Solid line indicates data of Hays and others (1976), dashed line, those of Emiliani (1978). Despite their independent compilations, the two data sets are remarkably similar.

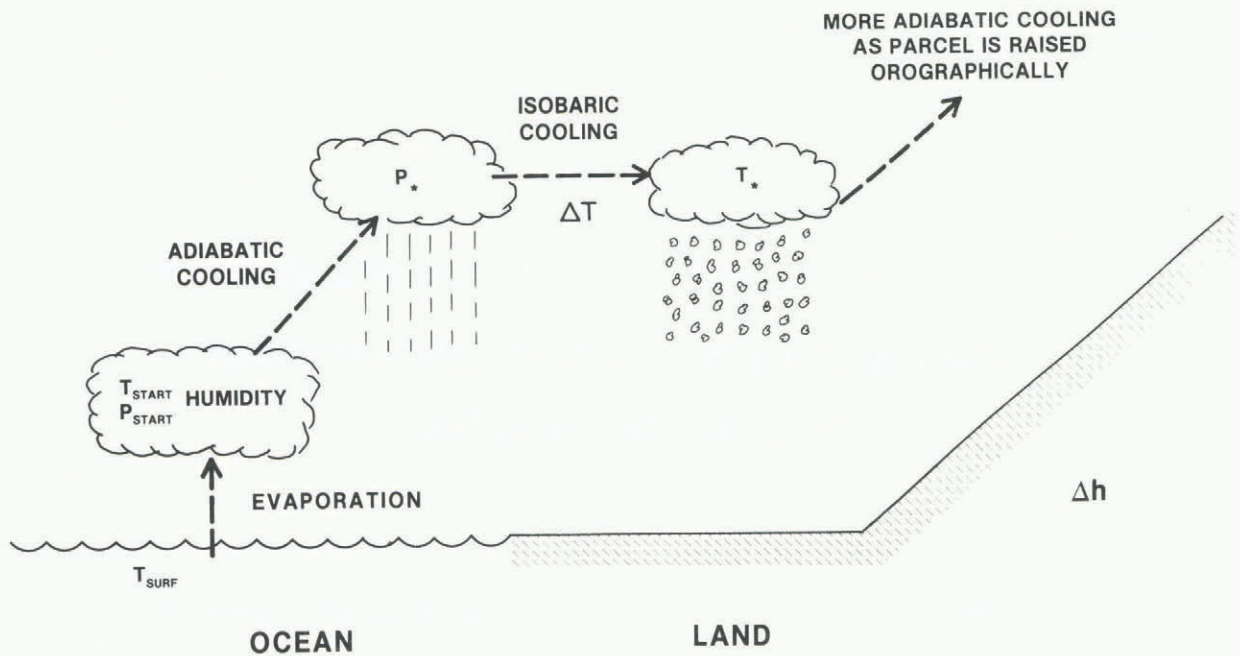


Fig.2. Main features of model for predicting isotopic composition of precipitation. The model is based on concepts discussed by Dansgaard (1964). A parcel of air and water vapor is formed by evaporation from the ocean at temperature  $T_{START}$ , raised adiabatically to pressure  $P^*$ , then cooled isobarically by an amount  $\Delta T$ . Upon encountering elevated terrain (e.g. a continental ice sheet) the parcel is raised to a distance equal to the height of the ice sheet at the point of precipitation  $\Delta h$ . The  $^{16}O/^{18}O$  ratios in the parcel and in precipitation from the parcel are followed throughout this process.

temperature of the water in which the calcium carbonate shells were formed. An increase of roughly 0.2‰/oo takes place for each 1 K decrease in temperature (Epstein and others 1953). Note that this effect operates in the same "direction" as the effect of ice volume, so that colder climates mean higher  $\delta^{18}O$ . The observed increase in  $\delta^{18}O$  between today's conditions and those of a glacial maximum is in the range from 1.5 to 2.0‰/oo for the fossils of bottom-dwelling as well as surface-dwelling organisms (Shackleton 1967). The inferred temperature change, over 7 K, is far too large for bottom waters and thus the temperature effect cannot be the prime contributor to  $\delta^{18}O$  variations, at least in bottom-dwelling organisms.

The ice volume effect can account for most of the amplitude of  $\delta^{18}O$  variations provided that there is enough isotopic fractionation between ocean water and glacial ice. From measurements of the isotopic composition of rain and snow at the present time and of ice from Greenland and Antarctica, it has been inferred that the great Pleistocene ice sheets were from about 30 to 40‰/oo lighter than ocean water (Olausson 1965, Dansgaard and Tauber 1969). Since the fraction of ocean water stored in the Pleistocene ice sheets was about 0.03, the ice volume effect would then account for a change of from roughly 0.9 to 1.2‰/oo, or about two-thirds of the observed variation.

We have recently obtained results from a model of isotopic fractionation in evaporation and precipitation (Fig.2) which support this conclusion (Covey and Haagenson\*). Results of the model depend strongly on (a) the degree to which non-equilibrium processes act in evaporation, (b) the height to which saturated air is raised after its formation, and (c) the temperature at which precipitation is formed. After obtaining representative values for these parameters

by calibrating the model against isotope data from the present, we extend the calculations to predict the isotopic composition of precipitation onto the Laurentide ice sheet during the last glacial maximum. Our results (Fig.3) imply that near the ice divide,  $\delta^{18}O$  relative to ocean water should have been  $\sim 30$ ‰/oo. Since most of the ice in an equilibrium ice sheet originates near the ice divide, the conclusion is that the isotopic fractionation/ice volume effect is large enough to account for most of the  $\delta^{18}O$  signal in oceanic sediments.

It should be noted, however, that the sensitivity of isotope composition to the meteorological details of evaporation and precipitation does introduce significant uncertainties into the interpretation of a related paleoclimate proxy, namely the isotope composition of ice cores from the Greenland and Antarctic ice sheets. The amount of  $^{18}O$  in ice is found to decrease during glacial periods; this has been interpreted as indicating colder temperatures at the ice sheet surface during these times (Dansgaard and others 1971). Our results (Fig.3) show that a lowering of  $\delta^{18}O$  may indeed be caused by a decrease in temperature, but it may also be caused by a shift in precipitation patterns, e.g. by increased uplift of precipitating air masses independent of terrain elevation. Thus the ice-core data are more ambiguous than has been hitherto appreciated.

It is important to assess the ice volume effects of the Antarctic as well as the Pleistocene ice sheets. During a glacial maximum the exposure of the Antarctic continental shelf would allow the ice sheet to expand. An approximate 10% increase in ice volume results from a shelf exposure of 0.5° latitude. The change in volume is much less than the volume of the Pleistocene ice sheets though its effect on the  $\delta^{18}O$  record would be magnified by the fact that Antarctic ice is isotopically very light.

In addition to the temperature and ice volume effects, many other factors such as variations of isotope composition within the oceans, non-equilibrium incorporation of oxygen isotopes into marine animal

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Covey C, Haagenson P L A model of oxygen isotope composition of precipitation: implications for paleoclimate data

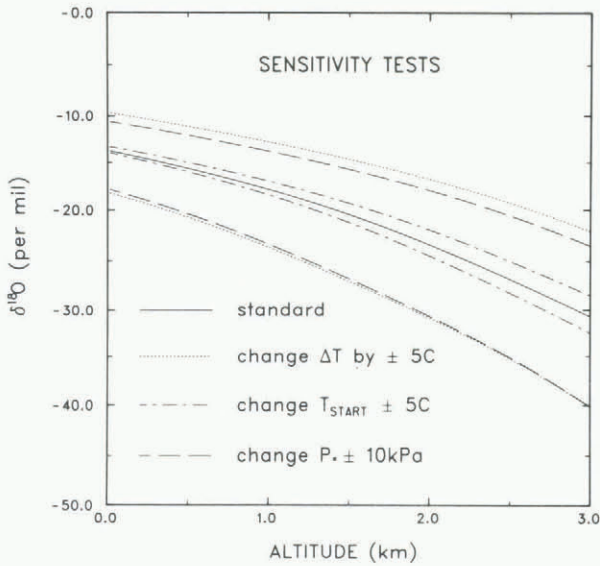


Fig.3. Isotopic composition of precipitation onto a continental ice sheet as a function of the height of the ice. Results are shown for seven model runs. "Standard" case uses values based on trajectory calculations for contemporary storms:  $T_{START} = 10\text{ C}$ ,  $P^* = 75\text{ kPa}$ ,  $\Delta T = 5\text{ C}$ , and an  $8\text{‰}$  enhancement of isotopic fractionation due to kinetic effects during the initial evaporation. Additional curves are for changes in parameter values from the "standard" set, as described in the figure: the curves above the "standard" result from decreasing  $\Delta T$ , increasing  $T_{START}$ , and increasing  $P^*$ ; those below the "standard" result from changes of the opposite sign. The lowest two curves show that raising the parcel further (lowering  $P^*$  by 10 kPa) can have the same effect as lowering the surface temperature at which precipitation occurs (increasing  $\Delta T$  by 5 C).

shells, and preferential dissolution of isotopically lighter shells can influence the  $\delta^{18}\text{O}$  signal. Berger and Gardner (1975) estimate that each of these effects could contribute up to  $0.5\text{‰}$ . Surprisingly few laboratory studies have been performed on cultured foraminifera to quantify effects such as these (Erez and Luz 1982).

Other effects may distort the time series of  $\delta^{18}\text{O}$  obtained from sediment cores away from the true ice volume record. Bioturbation preferentially diminishes the record of higher-frequency fluctuations (Goreau 1980, Dalfes and others in press). Furthermore, we must consider the changing isotopic composition of the ice sheets themselves: the ice sheets were probably isotopically heavier (less negative  $\delta^{18}\text{O}$ ) at their inception than at their maximum extent (Olausson 1981). Thus the apparent record of ice growth inferred from ocean sediment  $\delta^{18}\text{O}$ , which implicitly assumes a constant isotopic composition for the ice sheets, will lag the true ice volume record. Estimates based on simple isotope budgets for the Pleistocene ice sheets indicate that this lag time could be as long as several thousand years; similar effects may create a lag between ice decay and its appearance in the  $\delta^{18}\text{O}$  record (Olausson 1981, Mix and Ruddiman in press).

More detailed calculations with an ice-sheet/ice-flow model (Covey and Schneider in press) imply that the distortions between ice volume and  $\delta^{18}\text{O}$  record may be even greater because in a real ice sheet a large amount of the ice has a very long residence time. Covey and Schneider use a zonally symmetric ice-sheet model with power law creep and instant bedrock adjustment.  $\delta^{18}\text{O}$  of deposition is specified as a function of the height of the ice sheet, and isotopes are advected by the ice flow. Preliminary results for the

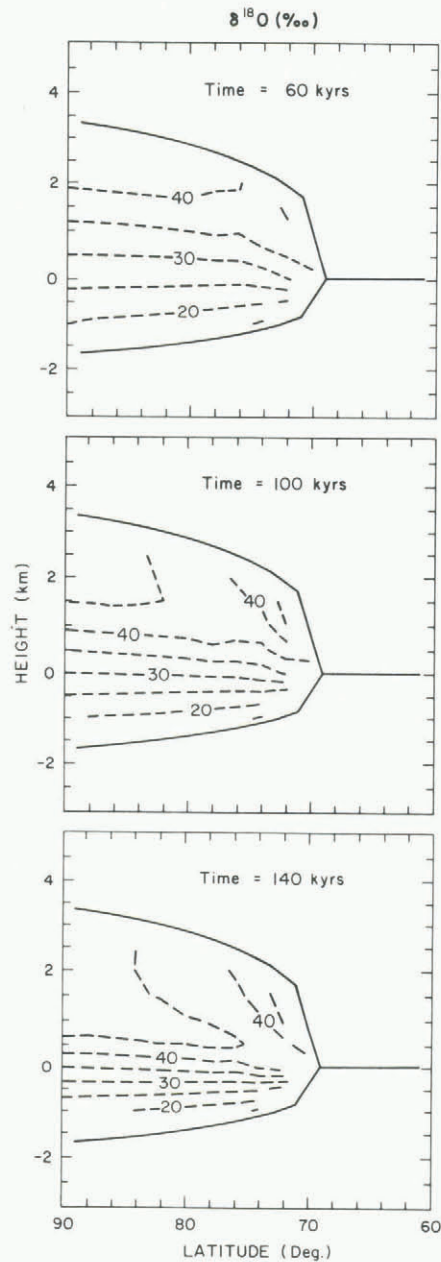


Fig.4. Flow of isotopes through a growing ice sheet initialized at zero at time = 0. Details of the calculation are given in Covey and Schneider (in press).

Antarctic ice sheet are shown in Figure 4. Long after the ice sheet reaches an equilibrium height, a large amount of (relatively) isotopically heavy ice remains trapped below the bedrock level.

In summary, it is apparent that the  $\delta^{18}\text{O}$  record from ocean sediment cores primarily reflects the changing ice volume during the Pleistocene. The isotope record is probably good enough to identify the major cycles of glacial/interglacial variations, i.e. the cycle of  $10^6\text{ a}$  and other "Milankovitch peaks". But caution should be exercised in examining the details, particularly in interpreting phase lags between ice volume and other climatic variables such as temperature, and in "tuning" the results of climate models to correlate with  $\delta^{18}\text{O}$  series.

ACKNOWLEDGEMENT

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