

Cool tropical temperatures shift the global $\delta^{18}\text{O}$ -T relationship: An explanation for the ice core $\delta^{18}\text{O}$ - borehole thermometry conflict?

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Abstract. The discrepancy between central Greenland borehole temperatures and the isotopic composition of Last Glacial Maximum ice can be explained by a shift in the $\delta^{18}\text{O}$ -T relationship for the hydrological cycle linked to cooler tropical temperatures. This concept is illustrated using a simple Rayleigh distillation model. An estimate for $\alpha = \Delta\delta^{18}\text{O}/\Delta T$ (LGM-Holocene) of $-0.37 \text{ ‰}/^\circ\text{C}$ is determined with a simple graphical technique.

Introduction

The internal temperature of the central Greenland ice sheet implies that the Last Glacial Maximum (LGM) was $15\text{-}20^\circ\text{C}$ colder than the Holocene [Cuffey *et al.*, 1995; Johnsen *et al.*, 1995]. This estimate doubles that inferred from the oxygen isotopic composition of the ice calibrated from the modern spatial $\delta^{18}\text{O}$ -surface temperature relationship. The borehole thermometry result has been supported by an estimate of firn thickness (thicker firn implies colder surface temperatures) derived from a comparison of $\delta^{15}\text{N}_2$ thermal diffusion fractionation relative to $\delta^{18}\text{O}$ of the ice, as seen at abrupt climate transitions such as the end of the Younger Dryas [Severinghaus *et al.*, 1996]. These results imply that the oxygen isotope thermometer gives the wrong answer in Greenland during the past several tens of thousands of years if the modern spatial $\delta^{18}\text{O}$ -T calibration is used.

The inconsistency of thermal and isotopic estimates is puzzling because the $\delta^{18}\text{O}$ -surface temperature relationship has survived many tests in the modern world and recent past. Dansgaard [1974] established a consistent empirical relationship between annual mean surface temperature and the isotopic composition of precipitation. More detailed global data and an atmospheric general circulation model with a hydrological $\delta^{18}\text{O}$ component support a simple global $\delta^{18}\text{O}$ -T relationship [Jouzel *et al.*, 1994]. In central Greenland, temperature and $\delta^{18}\text{O}$ time histories confirm the temporal applicability of a simple Dansgaard-like spatial relationship: the $\delta^{18}\text{O}$ -T slopes determined for seasonal cycles of the past few years [Shuman *et al.*, 1995] and for centennial/millennial temporal variations associated with the "little ice age" and "medieval warm period" [Cuffey *et al.*, 1994] were about $0.5\text{-}0.6 \text{ ‰}/^\circ\text{C}$, only slightly lower than seen in the modern spatial calibration. There is some indication that this consistency may not extend to the early Holocene, however [Cuffey and Clow, in press; Cuffey *et al.*, 1995].

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Several explanations have been offered for this discrepancy [Charles *et al.*, 1994; Cuffey *et al.*, 1995; Severinghaus and Brook, 1996]. These arguments have been summarized by [Jouzel *et al.*, in press] as (1) changes in the origin of precipitation (e.g., moisture originates from a seawater source with a different $\delta^{18}\text{O}$ or temperature), (2) microphysical and atmospheric processes (with the atmospheric moisture path depending on small scale details of atmospheric transport), (3) seasonality of the precipitation (e.g., if less snow accumulates in winter than in summer), and (4) cloud/surface temperature differences (e.g., changes in the cloud inversion height).

Plausible as all of these suggestions are, all raise a fundamental difficulty for the ice core isotope thermometer: we cannot as yet specify exactly how these processes differed in the past, and their existence must be inferred *ad-hoc* from borehole-thermometry calibrations of the isotope thermometer. For example, if moisture sources change, we would have to infer these changes from a highly resolved atmospheric general circulation model (present low-resolution models being insufficient to document the sources of Greenland precipitation [Jouzel *et al.*, in press] with accurate boundary condition reconstructions (sea surface temperature and ice sheet topography). If these explanations for the discrepancy are correct, then the isotope thermometer is unreliable wherever in time or space it cannot be calibrated by borehole thermometry.

The purpose of this paper is to argue for a simpler global-scale origin for the isotope-borehole thermometry discrepancy. I propose that the slope of the $\delta^{18}\text{O}$ -T relationship for polar snow remains constant at all times, but that the intercept varies in a simple fashion determined by changes in tropical sea surface temperatures. If this explanation is correct, the ice core thermometer is reliable whenever tropical sea surface temperatures can be specified.

Discussion and Simple Rayleigh Model

Following the work of Dansgaard [1974], our understanding of the isotopic composition of atmospheric precipitation begins from the concept of progressive Rayleigh distillation from a tropical source vapor. Subsequent work has established that the details of isotope fractionation in water vapor and precipitation can be complex [Jouzel, 1986]. In particular, it is understood that (1) the mechanistic temperature recorded by ice $\delta^{18}\text{O}$ is not surface temperature but rather the temperature at cloud height (systematically offset to cooler temperatures than on the ground), and (2) the curvilinear $\delta^{18}\text{O}$ -T behavior predicted by the Rayleigh model should be modified by including a kinetic isotope fractionation, which results in

a linear $\delta^{18}\text{O-T}$ relationship (snow crystal growth is not generally thermodynamically reversible, and light ^{16}O diffuses slightly faster to the growing ice crystal than heavy ^{18}O [Jouzel and Merlivat, 1989]). With these modifications, the spatial variations of snow $\delta^{18}\text{O}$ with surface temperature are expected to follow a linear relationship. However, even with these modifications, the predicted slope is greater than observed in the modern spatial calibration [Jouzel *et al.*, in press].

For purposes of conceptual illustration, the behavior of the simple unmodified Rayleigh model is shown figure 1. It is assumed that water vapor in the tropics is fractionated from standard mean ocean water isotopic composition (SMOW, now VSMOW) with isotopic equilibrium prevailing between liquid water and water vapor. After initial equilibration, the water vapor is assumed to cease interacting with the ocean. Further changes in the isotopic composition of water vapor result from a progressive decrease in atmospheric water vapor as air cools, with the condensate immediately removed, as described by the equation:

$$\frac{R}{R_i} = f \alpha(T) - 1 \quad (1)$$

where

- T = temperature of air mass
 α = $R_{\text{liquid}}/R_{\text{vapor}}$ isotope ratio fractionation (somewhat temperature dependent)

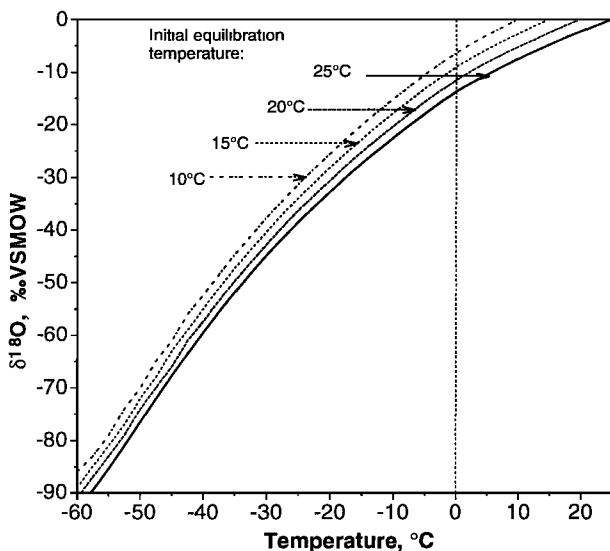


Figure 1. Simple Rayleigh Distillation Model. Equilibrium vapor pressures for water ($\geq 0^\circ\text{C}$) and ice ($< 0^\circ\text{C}$) were used to compute f . Water vapor pressure estimated for range $0 \leq T(^{\circ}\text{C}) \leq 30$ as:
 $(3.6923153078749\text{E-}07 + 1.53605633443583\text{E-}09*(T-15) - 5.41333194861509\text{E-}11*(T-15)^2 + 5.3576557376315\text{E-}16*(T-15)^3 - 1.28751659504212\text{E-}13*(T-15)^4) * \exp(0.0602300128260019*(T+273.16))$.
 Ice vapor pressure estimated for $-60 \leq T(^{\circ}\text{C}) \leq 0$ as:
 $(2.06727283917237\text{E-}07 + 6.40752162250554\text{E-}09*(T-30) - 4.20672917551061\text{E-}11*(T-30)^2 - 5.50575367980902\text{E-}13*(T-30)^3 - 4.1231149597534\text{E-}13*(T-30)^4) * \exp(0.0602644337097994*(T+273.16))$.
 Temperature dependence of α after Dansgaard [1964].

- f = the fraction of the initial water vapor remaining (set by the equilibrium vapor pressure of water, with an exponential dependence upon temperature)
 R/R_i = $^{18}\text{O}/^{16}\text{O}$ ratio of the remaining water vapor compared to its initial value

This behavior of this simple model for changing initial equilibration temperatures (figure 1) highlights the key premise of the argument presented here: the $\delta^{18}\text{O-T}$ path depends on the initial equilibration temperature. All paths start out with the first precipitation having the same isotopic composition as the ocean source water ($\delta^{18}\text{O} \approx 0\text{‰}$ relative to VSMOW). Proceeding along the cooling trajectory from the initial equilibration temperature, the paths are similar but offset relative to one another (figure 1). As noted before, water vapor transport and isotope separation is more complicated than this model implies. In particular, the empirical and kinetically-modified relationship is linear with T at temperatures where snow falls (rather than curvilinear as implied by a simple Rayleigh model).

Spatially distributed global data on the mean annual $\delta^{18}\text{O}$ of snow follow a relatively simple and uniform trajectory. In Greenland, this trajectory has a slope of $\sim 0.7\text{‰ per }^\circ\text{C}$ of surface temperature change. The system thus operates almost as if a single water vapor source was being progressively fractionated. This simplicity derives from the dominance of evaporation in tropical oceans: the tropics encompass by far the largest area of the ocean (the latitude band 20°S to 20°N encompasses 37% of the ocean surface) and the exponential dependence of vapor pressure on temperature ($\sim 5\%$ per $^\circ\text{C}$) ensures that tropical evaporation dominates the flux of water vapor into the atmosphere. This statement should not be interpreted as literally indicating that water vapor in Greenland originates from the equator; studies indicate that Greenland precipitation has a subtropical origin [e.g. Koster *et al.*, 1986].

Although the $\delta^{18}\text{O-T}$ relationship should follow a similar trajectory no matter what the initial temperature is, the absolute pathway depends on the initial temperature. When the Dansgaard relationship was established, however, it generally was believed that tropical ocean temperatures had remained relatively stable during the extreme climate cycles of the Pleistocene glaciations [e.g. CLIMAP, 1981]. Hence it was reasonable to have assumed that the same $\delta^{18}\text{O-T}$ pathway applied to both modern and LGM oceans. In recent years, several lines of evidence have suggested that LGM tropical ocean temperatures may have been as much as $\sim 4\text{--}6^\circ\text{C}$ cooler than they are at present. This cool LGM tropical temperature interpretation is based on (1) the noble gas content of tropical aquifers [Stute *et al.*, 1995; Stute *et al.*, 1992], (2) $\delta^{18}\text{O}$ and Sr/Ca in corals [Beck *et al.*, 1992; Guilderson *et al.*, 1994], and (3) snow lines and the isotopic composition of low-latitude mountain glaciers [Porter, 1979; Broecker and Denton, 1990; Thompson *et al.*, 1995]. The validity of this new view is still being debated within the paleoclimatological community, but it is sufficiently established to consider the consequences of cool tropics for the oxygen isotope composition pathway for global water vapor.

The role of moisture source temperature is well established (Aristarain *et al.*, 1986; Grootes, 1993), and discussions of the borehole thermometry/isotope conflict always mention the temperature of source moisture regions as a possible factor [Cuffey *et al.*, 1995; Jouzel *et al.*, in press]. What is new here is (1) explicit linkage between cool tropics and the global $\delta^{18}\text{O}$ -T relationship, and (2) separating the problem into two separate components (slope=site temperature change; intercept=source temperature change). Here, the change in source moisture temperature is proposed as global or large-scale regional rather than as local in origin.

LGM-Modern Comparison

As implied by figure 1, cooling of the initial source of equilibrated water vapor shifts the trajectory to parallel paths determined by the initial water vapor composition. Hence it is reasonable to assume that, if tropical sea surface temperatures did cool by 5°C , the pathway for $\delta^{18}\text{O}$ in global precipitation would have shifted to heavier values. The shift is difficult to specify precisely by simple arguments because of the complexity of water vapor transport, however there is little doubt of its sign and order of magnitude. Here, I simply assume that the $\delta^{18}\text{O}$ -T slope remains the same at all times, and that the intercept varies according to changes in tropical temperatures and global isotopic composition (figure 2). The $\sim 1\%$ positive shift in mean ocean $\delta^{18}\text{O}$ during the LGM (due to

increased ice volume) increases the shift due to cooler source waters. The resulting shift in the $\delta^{18}\text{O}$ -T trajectory has significant consequences for interpretation of the oxygen isotopic composition of LGM ice. The low-temperature region of the $\delta^{18}\text{O}$ -T relationship is expanded in figure 2 to show that if (1) central Greenland surface temperatures cooled from about -31°C to -45°C as implied by borehole thermometry and nitrogen isotope data, and (2) the trajectory followed by global water vapor shifted by -5°C as required by cooler tropics (dashed line on Fig. 2), then the change in $\delta^{18}\text{O}$ of central Greenland ice should be $\sim -40\%$, not -45% as implied by the modern $\delta^{18}\text{O}$ -T trajectory. Such a shift would result in the LGM ice having a $\delta^{18}\text{O}$ about -5% relative to the Holocene. Using this isotopic shift with the modern Dansgaard relationship (with the same intercept as today), this isotopic shift would imply a cooling of only $\sim 8^\circ\text{C}$ in surface temperature. Stated another way, this result implies that a comparison of modern and LGM isotopic values and temperatures should show an apparent $\Delta\delta^{18}\text{O}/\Delta T$ relationship of $\sim 0.37\%$ compared to the modern value of $\sim 0.70\%$. Borehole thermometry data imply an optimal model fit $\Delta\delta^{18}\text{O}/\Delta T$ of $0.35\%/^\circ\text{C}$ [Cuffey *et al.*, 1995], in close agreement with that predicted here (perhaps misleadingly so given the assumptions used for this simple estimate).

It should be emphasized again that this argument does not assume that the moisture in Greenland literally comes from the tropics; rather, the term "cool tropics" is taken as implying a global cooling of the ocean and atmosphere

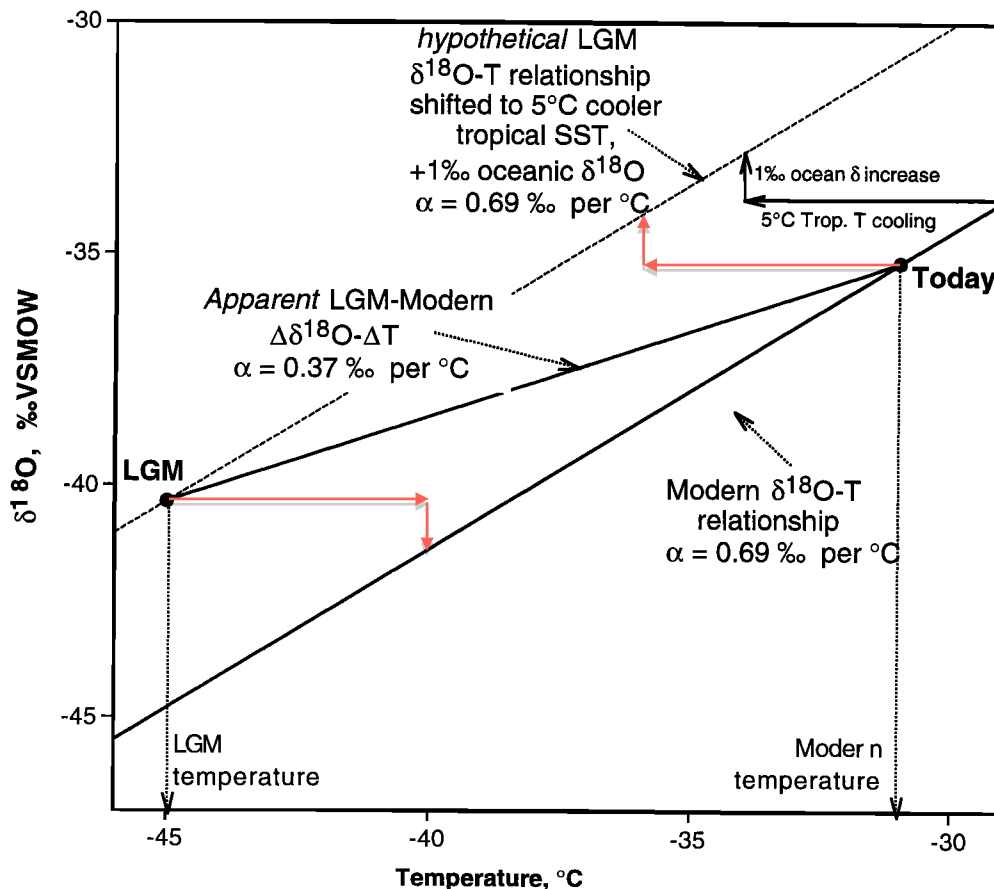


Figure 2. Proposed LGM-modern $\delta^{18}\text{O}$ -temperature offset.

system including the subtropical regions where Greenland precipitation may originate.

Summary and Implications

A shifted $\delta^{18}\text{O}$ -T trajectory is a simple consequence of cooler tropical temperatures. This shift can account for most of apparent disagreement between the borehole temperature history (and the apparent $\Delta\delta^{18}\text{O}/\Delta T$ parameter) and the LGM-Holocene temperature change inferred from $\delta^{18}\text{O}$ using the modern spatial $\delta^{18}\text{O}$ -T relationship. This interpretation does not require shifts in water vapor source regions, and it maintains the same $\Delta\delta^{18}\text{O}/\Delta T$ slope at all times. It therefore restores predictability to the ice core isotope thermometer: if we can specify tropical surface temperatures as a function of time, then we can determine a surface temperature from a contemporaneous measurement of $\delta^{18}\text{O}$ in ice. Of course other factors, including regionally differing water vapor sources, may also affect the isotopic composition of LGM ice from central Greenland; however, the significance of these changes need not be as large as suggested previously. This reinterpretation of glacial oxygen isotope values makes some testable predictions. If this model is correct, all LGM sites should follow a $\Delta\delta^{18}\text{O}/\Delta T$ relationship with the modern slope, only shifted $\sim 1\%$ vertically and $\sim 5^\circ\text{C}$ horizontally from the modern intercept. This predicted trajectory can be tested by nitrogen isotope and borehole thermometry data at additional sites throughout the world. This new interpretation also has consequences for the interpretation of other isotope thermometers influenced by precipitation, such as continental carbonate $\delta^{18}\text{O}$, tree-ring cellulose D/H, and paleo-aquifer $\delta^{18}\text{O}$. The negative shift in $\delta^{18}\text{O}$ resulting from tropical cooling and seawater $\delta^{18}\text{O}$ is estimated here as being larger than the measured positive $\delta^{18}\text{O}$ changes at sites such as Devils Hole [Winograd et al., 1985]. Hence continental isotope data should have the same large artifact as does ice in central Greenland; continental isotopic paleotemperatures will underestimate continental cooling. Compensating for this effect would also eliminate most of the apparent timing differences on termination II between Devil's Hole record and the marine record and buttress faith in Milankovitch and the SPECMAP chronology.

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