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Huascaran $\delta^{18}O$ as an indicator of tropical climate during the Last Glacial Maximum

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Abstract. The interpretation of the isotopic composition of Andes tropical mountain glaciers is reconsidered using a simple Rayleigh distillation model. It is shown that Andes snowline depression and Huascaran ice core $\delta^{18}O$ are consistent with a 3°C drop in sea surface temperature during the Last Glacial Maximum, accompanied by an increase from 40% at present to 60% during the LGM in the annual mean proportion of atmospheric water lost to continental runoff. Broecker's interpretation of Huascaran $\delta^{18}O$ as a paleo-hygrometer is discussed.

1. Introduction

The behavior of the tropics during the Last Glacial Maximum (LGM) has emerged as a crucial test of models used to understand the climate system [*Guilderson et al.*, 1994; *Broecker*, 1994, 1995; *Broccoli and Manabe*, 1987]. Tropical glaciers provide an important window into past behavior of this region. The Huascaran ice core recovered by [*Thompson et al.*, 1995] represented a breakthrough, providing a continuous record of $\delta^{18}O$ in tropical precipitation going from the Holocene back to the LGM. *Thompson et al.* tentatively suggested LGM temperatures 8-12°C cooler than present, based on an empirical $\delta^{18}O$-temperature relation [*Dansgaard*, 1965]. We present a re-interpretation based on *a priori* physics, which shows that the isotopic data does not demand such a large cooling. Simple models of the type employed herein provide a complement to more intricate studies based on general circulation models [*Joussaume et al.*, 1984; *Hoffman et al.*, 1998], while avoiding the pitfalls inherent in empirical $\delta^{18}O$-T relations. We consider in detail only the implications of Huascaran data for the LGM, but the treatment illuminates the issues involved in interpretation of tropical glacial isotope data from other sites, and referring to other times.

2. Model description

Our calculation is similar to that used by [*Boyle*, 1997] to interpret Greenland ice-core $\delta^{18}O$. The usual interpretation of such data is in terms of empirical relations between present $\delta^{18}O$ and temperature [*Dansgaard*, 1965]. However, fractionation is a function not of the final temperature of a parcel when it precipitates onto the glacier, but rather of the *difference* between the initial temperature at which the parcel picks up moisture and the final temperature; more precisely, it depends on the ratio of the initial and final water vapor content, which in turn is closely related to temperature. Insofar as one can expect the temperature of the source air to be different in a glacial world, isotopic paleothermometry based on empirical relations drawn from the present climate is suspect. Boyle employed an equilibrium Rayleigh distillation process, based on source air with a temperature related to the tropical sea surface temperature and precipitating air with a temperature related to the glacier's surface temperature.

The distillation model is less problematic in the tropical case than for polar ice. The premise of a global fractionation process involving a source vapor at a unique tropical temperature is meteorologically implausible for polar regions, and owing to the large extratropical temperature gradient, a small shift of the location of source water could lead to a large shift of glacial $\delta^{18}O$. Boyle used the calculation only to determine the intercept of the $\delta^{18}O$ - T relation, continuing to rely on empirical relations to determine its slope. For the tropics, one has a much better handle on the temperature and humidity of the source air, since it is virtually certain that the moisture snowed out on Huascaran comes from somewhere in the tropics (defined as the region 25N-25S), and most probably the tropical Atlantic. Further, the temperature and humidity of the tropical boundary layer are relatively horizontally uniform, so that one can tolerate quite large errors in the location of the source air without incurring serious errors in the final isotopic composition. We therefore carry out our interpretation of the Huascaran record without recourse to the empirical slope. In addition, we allow for pre-fractionation due to re-distillation on the way to the Andes, associated with water loss due to runoff. This effect is crucial to the large annual isotopic cycle observed in the modern-day high Andes [*Grootes et al.*, 1989].

The air parcel which ultimately drops precipitation onto the glacier is presumed to begin life in the marine boundary layer at sea level pressure p_s, with a temperature T_s , a specific humidity q_0, and vapor isotopic composition δ_0. A fraction r of the initial vapor is lost by runoff of precipitation over the continent on its way to the foothills, and the remaining fraction $(1-r)$ evapotranspirates back into the continental boundary layer. Thus, the inland boundary layer vapor has composition $\delta_1 = (1+\delta_0)(1-r)^{\alpha_1 - 1} - 1$ where α_1 is the equilibrium coefficient at the temperature at which the prefractionation occurs. The precipitation deposited at Huascaran is presumed to form when the prefractionated pool of moisture is lifted to an altitude z_p, which must lie above the surface of the glacier (6048m). The temperature $T(z_p)$ is assumed to be related to T_s by the moist adiabat starting from the lifted condensation level corresponding to q_o and T_s. The moist adiabat closely fits $T(z)$ for the present tropics [*Xu and Emanuel* 1989] , and in the absence of compelling evidence to the contrary one should assume the same for the LGM. The air parcel is assumed saturated when it

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Table 1. Parameter summary. Values in [ ] are model-inferred, and values in () are assumed.

	Holocene	LGM
Sea Level	0m	-120m
p_s	1012mb	1022mb
r (ann. mean)	[.4]	[.6]
h	80%	(80%)
T_s	298K	[295K]
Snow Line (0C)	4800m	3850m ± 50m
δ_L	$1^o/_{oo}$	$2.24^o/_{oo}$
δ_0	$-11.4^o/_{oo}$	$\delta_L - (\epsilon^* + (1-h)C_k)$
z_p	[6300m]	(6300m)
Glacier $\delta^{18}O$	$-17^o/_{oo}$	$-23^o/_{oo}$

precipitates, whence the final moisture content is given by the saturation specific humidity at $T(z_p)$. The final isotopic composition of the vapor and precipitation can then be obtained by integrating Eqn. (12) of [*Jouzel*, 1986] subject to initial condition $\delta = \delta_1$, using fractionation coefficients from [*Jouzel*, 1986]. Because the coefficients are only weakly dependent on temperature, our results are not very sensitive to the temperature at which the prefractionation occurs. The prefractionation step was carried out at $T = T_s - 5°C$. Standard formulae reviewed in [*Xu and Emanuel* 1989] were used to compute the saturation vapor pressures and moist adiabat. The temperatures aloft range from well above freezing to well below freezing. Whether one is condensing into liquid or ice affects the saturation vapor pressure, and also the equilibrium fractionation coefficient. We incorporated the effects of supersaturation by using a weighted average between the liquid and ice saturation vapor pressures which shades over between the all-liquid formula at 0°C to the all-ice formula a -20°C linearly with temperature. The same weighting coefficients were used to effect the transition between liquid and ice fractionation coefficients.

To determine δ_0 in terms of the isotopic composition of the sea water, δ_L, we use the Craig-Gordon model [*Craig and Gordon*, 1965]. In a steady state this yields $\delta_0 = \delta_L - (\epsilon^* + (1-h)C_k)$, where ϵ^* is the equilibrium fractionation increment, h is the boundary layer relative humidity, and C_k is the diffusive resistance coefficient (determined from present observations).

3. Application to Huascaran

As noted by [*Broecker*, 1997], tropical glaciers provide us with two bits of information about the climate of the LGM: snowline, and glacial $\delta^{18}O$. The correct interpretation of snowline data was discussed by [*Betts and Ridgway*, 1992], and suggests a 2-3C sea level cooling at the LGM. We use the snowline and $\delta^{18}O$ to constrain the surface temperature and the runoff fraction r. To close the problem, one needs observations or assumptions regarding h and z_p.

Present conditions are summarized in *Table 1.* With a surface temperature of 298K, the thermodynamic model matches the observed present day freeze line. The stated δ_0 is supported by coastal vapor data quoted in [*Moreira et al.*, 1997]. r is constrained by the observed annual cycles of precipitation in the foothills of the Andes, and of recent snow deposited in Andes glaciers. Rainfall data implies that 80% of the original water is lost to runoff in the rainy season, but that virtually all rainfall re-evaporates into the boundary layer during the dry season [*Grootes et al.*, 1989]. The $\delta^{18}O$ of recent Huascaran snowfall ranges from about $-27^o/_{oo}$to $-13^o/_{oo}$([*Thompson et al.*, 1995], Fig. 6A). Using our model, these observations imply a precipitation altitude of 6000m during the wet season, and 6600m in the dry season. To match the annual average composition of $-17^o/_{oo}$we fix the precipitation height at the mean of the seasonal extremes (6300m) and take $r = .4$. This is close to runoff estimates based on streamflow and precipitation data [*Moreira et al.*, 1997; *Salati et al.*, 1979].

During the LGM, snowline was depressed and the $\delta^{18}O$ of glacial precipitation was more negative. Known parameters for the LGM [*Betts and Ridgway*, 1992; *Thompson et al.*, 1995; *Broecker*, 1997] are summarized in Table 1. For the moment we assume that glacial age h was the same as at present. Currently, h varies little over a wide range of conditions encountered in the tropics (e.g. [*Broecker*, 1997]). It could change appreciably in response to extreme changes in the low level wind [*Betts and Ridgway*, 1992]. The implications of h variation will be taken up in the Discussion.

What could we say on the basis of isotopic data alone? Figure 1 shows $\delta^{18}O$ vs. height for LGM conditions as calculated by the Rayleigh model with r fixed at its present value, for various surface temperatures. As surface temperature decreases, the $\delta^{18}O$ aloft becomes more negative. This occurs partly because the temperature lapse rate becomes steeper as the atmosphere gets colder. Further, the saturation vapor pressure curve is not exactly exponential in temperature, with the consequence that cooling a parcel from, say, 290K to 270K wrings out a greater fraction of the original water than cooling it from 295K to 275K. If we keep the precipitation height and runoff ratio fixed at current values, then a surface temperature of 292.5K matches the LGM isotopic data. Allowing precipitation height to change, temperatures colder than 291.25K can still be ruled out, as they would require a precipitation height below Huascaran ground level. On the warm side, an LGM surface temperature of 297K, as implied by CLIMAP [*CLIMAP*, 1976], would require the precipitation height to increase dramatically to 7500m. If one allows the runoff ratio to change, however, the isotopic constraint on temperature becomes weak. If r=0 (i.e. the LGM runoff picture year round becomes like the present dry season), then temperatures as cold as 288.5K can be accomodated. On the warm side, if r increases to .72 (year round like the present wet season), then a surface temperature of 298K can be accomodated without any change in precipitation altitude from the present value.

If we invoke the snowline constraint, then the thermodynamic model fixes the LGM sea surface temperature at 295K, as in [*Betts and Ridgway*, 1992]. Then, in order to account for the isotopic data, the runoff ratio must increase to 60% if we demand that the precipitation height remain at its present value. However, if the precipitation height rises to 6653m in glacial times, the runoff ratio need increase to only 50%, and if the precipitation height were 6950m, the runoff could remain at its present value of 40%. In order to drop the runoff ratio to 30% one must increase the precipitation height to 7176m.

A substantial increase in precipitation height is hard to justify for upslope precipitation, so the isotopic data leads us to infer a greater runoff percentage during the LGM. This is

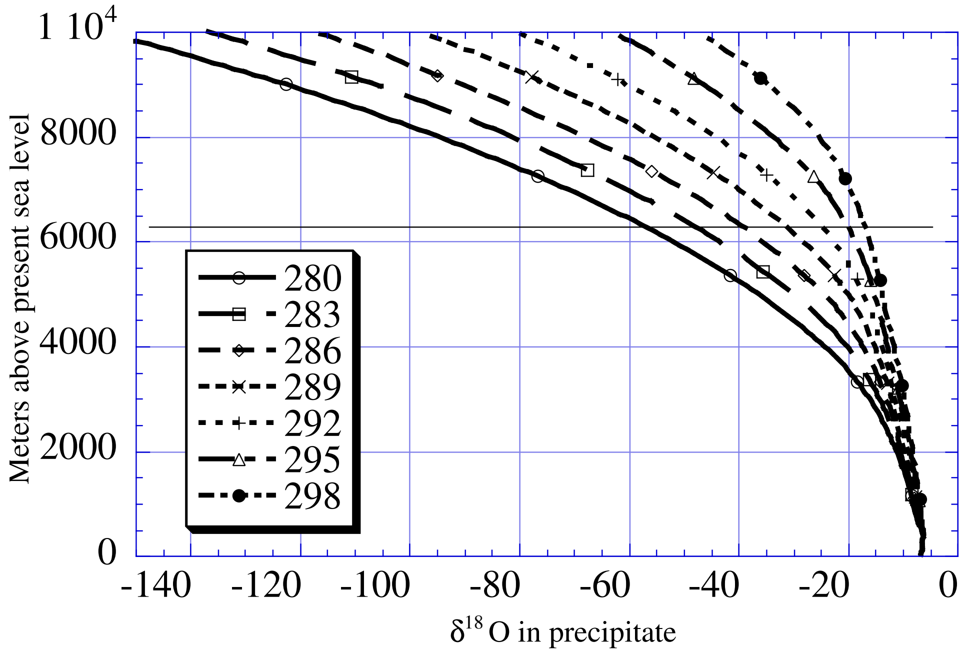

Figure 1. Computed isotopic composition of precipitation, as a function of the altitude at which the precipitation forms. The runoff ratio is held fixed at 40%.

a surprising result. In the colder glacial climate, one expects less precipitation because there is less moisture in the air. Hence, one might expect the whole year to behave more like the present dry season, during which time little or no precipitated water is lost to runoff. We speculate that changes in vegetation during the LGM led to less efficient evapotranspiration. Some of the apparent increase in runoff may actually reflect a shift from transpiration to evaporation. Transpiration doesn't fractionate [*Moreira et al.*, 1997], while evaporation preferentially releases isotopically light water. If water is 100% recycled this doesn't matter, but if recycling is partial a shift to evaporation increases pre-fractionation even if the runoff ratio is fixed. Significantly, forest is dominated by transpiration, while grasslands are dominated by evaporation [*Moreira et al.*, 1997]. The interpretation of isotopic fluctuations in snowfall would then be dependent on time scale, since short-lived climate fluctuations might not allow the turnover in vegetation needed to increase the pre-fractionation.

4. Discussion

Broecker, [1997] has argued that the Huascaran data implies that the mid-tropospheric relative humidity was lower at glacial times than it is today. This would support a role for anomalous water vapor radiative feedbacks in enhancing tropical climate fluctuations. The thermodynamics used by Broecker differ only inconsequentially from our moist adiabat calculation. Recognizing that cooling compatible with snowline data does not yield enough fractionation to account for the observed LGM isotopic lightness, Broecker posits the removal of sufficient additional water from the air parcel to yield the desired lightness; the water thus removed is translated into a lowering of the water vapor content of the air over Huascaran. The problem with this interpretation is that one can only remove water from the air parcel by forming precipitation, and precipitation forms only if the parcel is lifted and cooled enough to bring it to saturation. Hence, Broecker's argument only says that the water vapor content was lower at the altitude at which the parcel becomes saturated and at which the precipitation falling on Huascaran forms. This is, in essence, an inference of the precipitation height. It constitutes information about the water content at a single point of those special trajectories which happen to drop snowfall at Huascaran, and is an insufficient basis for drawing inferences about water vapor content elsewhere in the tropics.

The general peril in isotopic paleoclimate proxies is that the data may reflect a change in the source vapor due to a minor circulation change, rather than a widespread change in a major climate variable such as temperature or runoff. While this problem is less severe in the tropics than in the extratropics, it is by no means absent. Suppose that the temperature profile is held fixed at that which is consistent with the LGM snowline, but that the temperature and relative humidity of the source air in the isotope calculation is varied. If T_s is held fixed while h is reduced to 60% (which is well below the lowest Atlantic value currently observed from 20N to 20S), then kinetic effects reduce δ_0 to -14.6$^o/_{oo}$, but this is offset by the reduction in q_0, whence our estimate of r is reduced inconsequentially, to 59%. If h is fixed at 80% but the source water temperature is reduced from 295K to 285K (a difference comparable to the maximum present day difference between coastal Brazil at 10S and coastal Africa at 20S) then the reduction in q_0 dominates, reducing the altitude-effect fractionation, and our estimate of r increases to 77%. This possibility notwithstanding, we think it most likely that the Huascaran data reflects conditions in the Western Atlantic, as it is unlikely that vapor could retain its initial isotopic stamp after travelling thousands of kilometers in the boundary layer at speeds of around 5m/s.

The interpretation of Huascaran $\delta^{18}O$ is less constrained than one would like, and additional effects further complicate the picture. Evaporative enrichment of the firn could be canceling some of the pre-fractionation. The kinetic corrections to the Rayleigh model, discussed in [*Jouzel and Merlivat*, 1984], are only crucial at polar temperatures, and falling or retained ice crystals do not re-equilibrate with ambient vapor [*Jouzel*, 1986]. However, the tropical regime permits mixed phase water/ice clouds, in which both kinetic effects and re-equilibration effects due to retained water could be important. The retained-water effect causes precipitation to be less isotopically light than would be expected on the basis of open-system Rayleigh distillation [*Jouzel*, 1986]. Thus, the true precipitation heights may be somewhat greater than we estimated, with the $\delta^{18}O$ being brought back to observed values through the retained-water effect. Because kinetic fractionation of HDO procedes differently from $H_2^{18}O$, HDO measurements in tropical ice cores would narrow the possibilities, and also help detect changes in boundary layer relative humidity. Some of the ambiguity between runoff and precipitation height effects can be resolved through a better understanding of the synoptic setting in which Andes snowfall occurs, such as may derive from new field campaigns [*Hardy et al.*, 1998] and satellite-based rainfall observations such as *TRMM*.

However these issues play out, our results point clearly to the importance of accurate treatment of runoff and land surface processes, and to the compatibility of the isotopic data with LGM tropical temperatures on the order of 3C cooler than at present. Isotopic data taken alone does not strictly rule out a colder tropics, but it does not require it either.

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