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The influence of midlatitude and tropical overturning circulation on the isotopic composition of atmospheric water vapor and Antarctic precipitation

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[1] While isotope records from ice cores are known to reflect temperature, this must be associated with zonally symmetric circulation. A new conceptual depiction of the isotopic cycling is established by considering the overturning circulation in isentropic coordinates. In this depiction, poleward transport of air and water vapor is non-diffusive, in a way that is similar to that depicted by Rayleigh models. However, the equatorward return flow is also important since it is this which is supplied with water by surface evaporation. The isotopic state emerges as a balance between evaporative supply and poleward advection, and removes the need to assume some initial source condition for an open distillation. Model experiments that simulate a wide range of circulation strengths show the isotopic composition of Antarctic snow is strongly linked to the strength of midlatitude (eddy driven) circulation, which in turn is driven by meridional temperature differences. Antarctic isotopes are largely independent of the tropical (Hadley) circulation because the rate of advective transport from the tropics to the polar region exceeds the rate at which surface sources replenish the poleward moving air stream. Across all simulations and between seasons, the relationship between δ^{18}O in Antarctica above 1500 m and local surface air temperature is found to be remarkably robust at around 0.69‰/K in winter, 0.85‰/K in summer and with a seasonal slope of 0.60‰/K. Because these slopes result from changes in circulation, isotope records from the continent interior can be considered indicative of the history of overturning circulation.

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1. Introduction

[2] The isotopic composition of ice from glaciers and ice sheets has been used successfully to deduce changes in climate on a variety of timescales [e.g., *Epstein et al.*, 1963; *Peel et al.*, 1988; *Andersen et al.*, 2004; *Stenni et al.*, 2004]. While it is conventional to interpret isotopic variations at high latitudes in terms of temperature, it is generally recognized that this is overly simplistic and other effects—such as seasonality, source region distribution and systematic change in large-scale circulation patterns—can also affect the isotopic state in ways that are inconsistent with simple single-parameter reconstructions [e.g., *Jouzel et al.*, 1997; *Krinner et al.*, 1997; *Werner and Heimann*, 2002; *Jouzel et al.*, 2003; *Brown and Simmonds*, 2004]. Linking the isotopic composition to the large-scale atmospheric circulation, *Noone and Simmonds* [2002b] showed that a large fraction of the interannual isotope signal from inland Antarctica is associated with variations in the water vapor transport in the region of the Southern Hemisphere storm track, and these changes coincide with variations in the Southern Annular Mode (SAM). Indeed, *Schneider and Noone* [2007] show this same association exists on decadal timescales if not longer. Changes in the storm track are representative of not only mean baroclinicity and storm activity but also of position and strength of the midlatitude westerly jet. As such variability that has an annular-like pattern is linked to changes in the hemispheric overturning circulation because of the requirements that conservation of westerly momentum and heat in this region impose. How changes in the overturning circulation are reflected in the isotopic composition of Antarctic snow is of concern here.

[3] There is some ambiguity in the role played by the tropical circulation in water transport to Antarctica. Evidence for a transport pathway from the tropics to Antarctica was found by *Alexander et al.* [2003] in the form of gas phase sulfur isotope anomalies, but it is unclear that a species with a short residence time, such as water, should significantly reflect this same transport pathway. Indeed a tropical water source largely disagrees with glaciological and model estimates that give a midlatitude source of Antarctic water [*Bromwich and Weaver*, 1983; *Johnsen et al.*, 1989;

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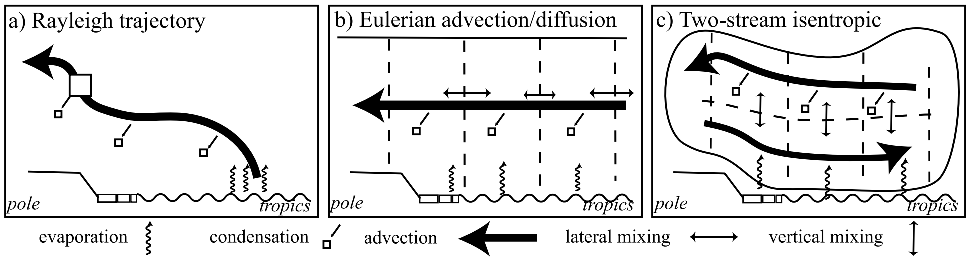

Figure 1. Schematic depiction of three alternate views of hemispheric isotope cycling. (a) A Lagrangian ("Rayleigh") distillation has continual rainout of heavy isotopes en route from some initial vapor; (b) a one-dimensional steady state is formed as a balance between rainout, evaporation and lateral dispersion due to eddies; and (c) a two-stream quasi-steady model has rainout in the mostly non-diffusive poleward branch, evaporation into a lower equatorward branch, and turbulent exchange providing vertical transport of water vapor between branches.

Koster et al., 1992; *Delaygue et al.*, 2000; *Werner et al.*, 2001; *Noone and Simmonds*, 2002b]. Nonetheless, it is plausible that a change in the circulation of the tropics can modify the circulation at higher latitudes, and thus lead to a tropical signal in the isotopic composition of polar snow even though the origin of the water may still be at midlatitudes. As a case in point, it is well established that there are some parts of the Antarctic ice sheet where accumulation correlates with variations in the mostly tropical El Nino-Southern Oscillation [*Bromwich et al.*, 2000; *Fogt and Bromwich*, 2006], and that these are reflected also in ice core isotope records [Gregory and Noone, Variability in the teleconnection between the El Nino-Southern Oscillation and West Antarctic climate, submitted to *Journal of Geophysical Research*, 2007; *Schneider and Noone*, 2007], because of large-scale reorganization of the hemispheric circulation during warm and cold phases rather than a simple equatorward shift in the water source. Further, studies that suggest isotopic evidence of a cooler tropics during glacial times are not inconsistent with weakened tropical circulation [*Boyle*, 1997; *Charles et al.*, 2001]. Detailed analysis of this possible dynamic linkage requires effective separation of the large-scale flow into that due to the mostly tropical zonal mean circulation and that of the midlatitude eddies.

[4] *Kavanaugh and Cuffey* [2003] used a zonally and vertically integrated form of the water vapor and isotope mass budget equations to partition the isotope signature into an "advective" versus "diffusive" component—the difference being that advective transport represents a spatial translation while diffusive transport tends to homogenize tracer fields through mixing. While this approach yields the conceptually satisfying result that lateral (diffusive) mixing (nominally provided by eddies) is separate from the (advective) mean flow, it is limited because of neglect of two important features of the circulation: 1) midlatitude eddies contribute not only to lateral mixing but also the (non-diffusive) mean overturning of the midlatitude atmosphere; and 2) water evaporated from the surface contributes to the poleward moving air mass only once it has been liberated from near the surface layers by turbulent mixing in the vertical. In support of the first claim, *Noone and Simmonds* [2002b] showed in model simulations that midlatitude eddies provide an advective mass flux toward the pole that ultimately contribute precipitation to the ice sheet, while the time-mean flow contributes a net sublimation from the ice sheet. This result supports observationally based estimates of polar moisture budgets [*Budd et al.*, 1995; *Turner et al.*, 1999] and is related in part to the surface outflow of cold dry air as required for mass convergence aloft [*Parish and Bromwich*, 1987; *Simmonds and Law*, 1995; *van den Broeke and van Lipzig*, 2003; *van den Broeke*, 2004]. Because the isotopic composition of vapor associated with the eddy circulation (which drives continental moisture convergence and positive surface mass balance) and mean circulations (which is associated with ablation) differ, it would be most satisfying to account for them separately. As such, one-dimensional isotope models that combine both components into a single term [*Fisher*, 1990, 1992; *Ciais and Jouzel*, 1994; *Vimeux et al.*, 2002; *Kavanaugh and Cuffey*, 2003] may be limited, especially if one were to speculate that the fractional contribution of the two to the net moisture transport would change during some paleoepoch.

[5] Illustrating this point, Figure 1 shows different conceptual views of the hemispheric water budgets. A Lagrangian trajectory model (i.e., a Rayleigh-type distillation model that follows an individual air parcel) requires some initial condition, and typically does not account for the mixing of air from a spatially disperse evaporative source along the transport route. The success of estimating polar depletion lies in accurately assuming the temperature and condensation history, and setting some reasonable initial condition [*Jouzel and Koster*, 1996; *Ciais et al.*, 1997; *Vimeux et al.*, 2002]. Such models are very constructive in providing first order understanding of isotopic signals. As an extension of this paradigm, mixing can be included by recasting the budget equation into a one-dimensional Eulerian form (i.e., rather than following a parcel, considers the passage of vapor and isotopes past locations fixed in space). Moisture can be added via evaporation as a function of latitude, and lateral mixing (by eddies) can be included as a diffusion process, as has been shown by, e.g., *Kavanaugh and Cuffey* [2003]. Here a further extension is considered that better captures underlying overturning circulation. Figure 1c shows a "two-stream" moisture budget that quantifies the poleward moving flow separately from the equatorward moving flow. Vapor in the lower (equatorward) branch is replenished by evaporation from the ocean source, while condensation in the upper (poleward) branch provides the familiar distillation. Mixing between branches allows the evaporative source to pervade the free troposphere, and thus

the lower layer acts, in some senses, to temporarily store the isotope information of the source. The capacity of the lower layer to store isotope information is dictated by the efficiency of exchange between the two layers. Once lofted, the poleward transport is mostly non-diffusive, although some lateral mixing will occur. This is in contrast to a one layer model, in which water is moved poleward immediately (without storage) and that the transport in question has a diffusive character when eddies are involved. Notice this two-stream model does not need to separate the flow into that associated with eddies or the mean flow, and is justified below. Notably the isotopic composition associated with the two one-way advective fluxes depicted in Figure 1c is not equivalent to the one two-way (i.e., diffusive-like) flux that accounts for eddies in the one-dimensional model depicted in Figure 1b. It is argued here that the two-stream model is an appropriate conceptual basis for interpretation of isotopic records because of the way it can incorporate the evaporative source in the cycling of water and its isotopes.

[6] The importance of the preceding discussion and adoption of the two-stream model stems from the realization that the strength of the poleward air mass circulation is an intimate part of the zonal mean overturning (rather than mixing) of the atmosphere, and that it can be considered a key metric of the general circulation [*Schneider*, 2005]. Central to this argument is that midlatitude eddies provide a net poleward mass flux associated with diabatic meridional circulation which directed in the same sense as a thermally direct cell [*Andrews and McIntyre*, 1976; *Johnson*, 1989; *Karoly et al.*, 1998; *Held and Schneider*, 1999]. This differs from the more familiar view in which the Ferrell cell circulates in the opposite sense, and is largely associated with adiabatic motions that have limited affect on the net poleward mass flux. The strength of this eddy driven cell conveniently quantifies the role of eddies in poleward transport of energy, momentum and gases–including water vapor and its isotopes. As such, if explicit account of this mass flux is retained in the explanation of isotopic composition, so too is the ability to consider key aspects of the circulation regime rather than resort to simple temperature reconstruction. At the same time, changes in features of the hemispheric circulation (such as the strength and location of the westerly jet and the storm tracks) can be linked to changes in the equator to pole temperature difference, which, under the assumption that tropical temperature are quite stable, is key to simple interpretations of polar isotope variations as local temperature. One might argue that a stronger temperature gradient would dictated more isotopic distillation during poleward transport so long as the source water isotope depletion remains constant. As such, explicitly taking account of the evaporative source in the two-stream model is a required advance over a simpler Rayleigh depiction in which the source water must be prescribed.

[7] This paper aims to link the isotopic composition of Antarctic snow directly to the meridional overturning circulation of the Southern Hemisphere. Of particular interest is to establish how much of the isotope signal is associated with zonal mean circulation (which dominates in the tropics and associated with the Hadley cell) and eddy driven circulation (which dominates in the midlatitudes in the region of the traditional Ferrell cell). With an atmospheric general circulation model with water isotope tracers, sets of perturbation experiments are performed to elucidate the role of the two components and to show how the ratio of the zonal advective mass flux to the rate of vertical mixing of evaporated water is a key parameter in reconstruction. By examining circulation regimes that differ greatly from the present climate, we aim to explain both how well and why classic associations between Antarctic isotope records of temperature and accumulation rates persist. To provide a context for experimental results, a brief overview of the hydrologic cycling of the Hemisphere is provided to demonstrate the relevance of the proposed two-stream conceptual model.

2. Model Simulated Circulation and Hydrology

[8] The isotopic version of the Melbourne University general circulation model (MUGCM), detailed by *Noone and Simmonds* [2002a], is a spectral model and configured to have a horizontal resolution corresponding to a transform grid of 5.25° longitude by 3.3° latitude, and to have nine levels in the "sigma" vertical coordinate. Seven of the nine levels are in the troposphere, and the upper two layers employ strong lateral damping to prohibit reflection of vertically propagating planetary waves from the model top. This limits the quality of the simulation near and above the tropopause. The validity of the isotopic simulation has been demonstrated by *Noone and Simmonds* [2002a], with applications to large-scale transport, Antarctic precipitation and the role of sea ice in interpreting ice core records further demonstrating the models credibility [*Noone and Simmonds*, 2002b, 2004]. These studies establish that the model is adequate for assessment of large-scale processes, and thus one may have confidence in the continental scale results of interest here. While the spatial resolution used here is quite modest by modern standards, the speed at which the model simulations can be completed facilitates a large number of sensitivity tests while capturing the fundamental behavior of the tropospheric circulation. For the control simulation, and all experiments described below, 600-day simulations for both perpetual 15 January and 16 July are performed, and the last 300 days are used to compile statistics.

[9] While a detailed description of the simulated climatology here is not warranted, a brief description of those modeled quantities that have bearing on discussion of perturbation experiment results is necessary and provides a link to the two-stream model. Figure 2 shows the zonal mean isentropic meridional mass transport stream function for July and January from the control simulation. The meridional mass flux stream function, ψ, is defined following *Peixoto and Oort* [1992] by the meridional velocity field (v) such that it may be evaluated using a finite difference approximation in the vertical. Specifically,

$$\frac{2\pi a \cos(\phi)}{g} v_k = \left(\frac{\psi_{k+1/2} - \psi_{k-1/2}}{\Delta p} \right) \tag{1}$$

where a is the radius of the earth, g is the acceleration due to gravity, ϕ is the latitude, Δp is the spacing between the model's sigma levels and k indicates the model layer midpoints with layer interfaces identified as "half" a layer up and down. The stream function is found by integrating

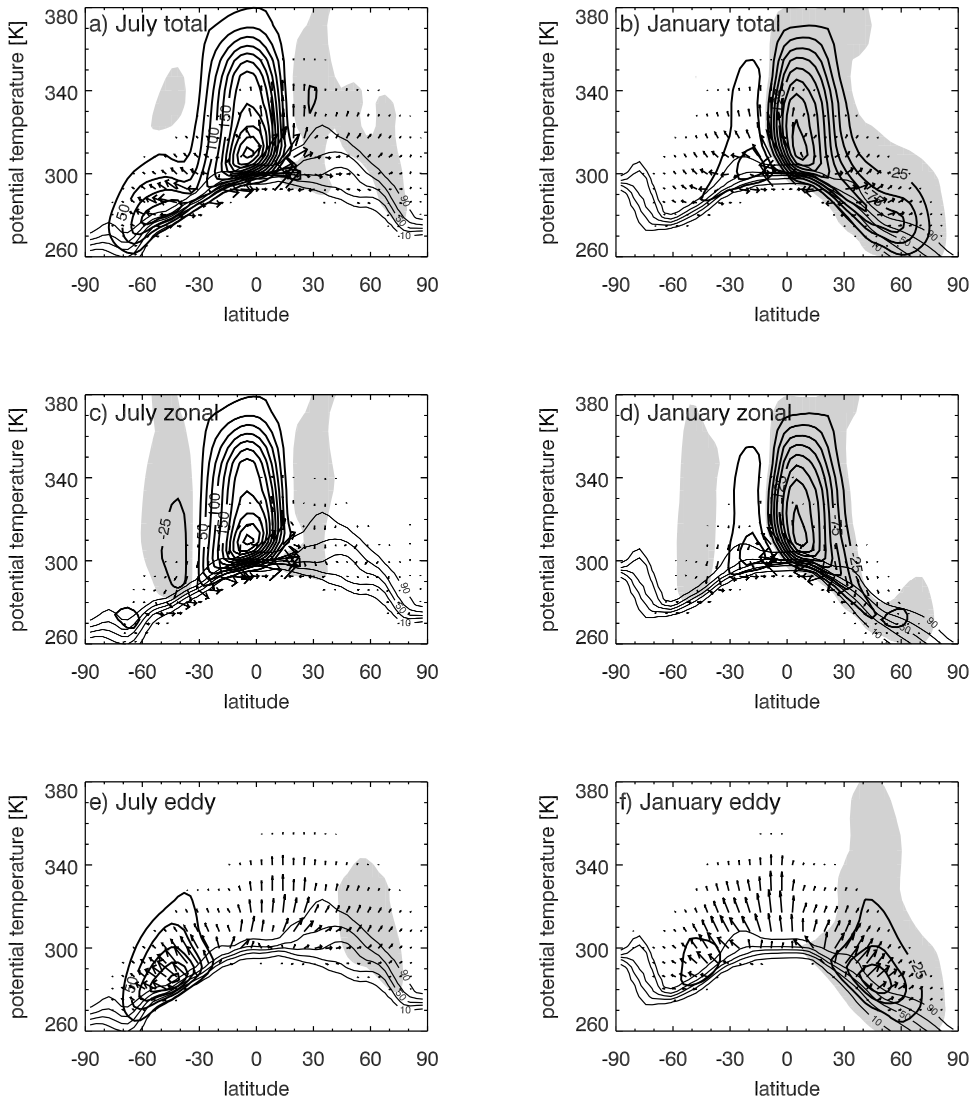

Figure 2. Isentropic mass transport stream function (contours of 10×10^9 kg/s) and water vapor flux (vectors) for (a and b) total, (c and d) zonal mean component and (e and f) eddy component. Negative values shaded and the zero contour is omitted. Panels on the left are for July and on the right for January. Contours toward the bottom of each panel are the frequency of the surface potential temperature, with the bold line showing the median potential temperature of the surface. Surface frequency contours are 10, 30, 50, 70 and 90%. The scale for moisture flux vectors is the same in all panels.

(1) from the top of the model downward with the upper boundary condition $\psi = 0$. Isentropic fields in three dimensions are computed within the model at each timestep during the simulation via Hermite cubic interpolation from the model's sigma coordinate to a high resolution potential temperature (θ) coordinate. The potential temperature coordinate is set to have 50 levels staggered to be linear in $\theta^{-\kappa}$ (where $\kappa = R_d/c_p$, R_d is the gas constant c_p is the heat capacity at constant pressure for dry air) from 220 K to 400 K, and linear thereafter to 600 K. This isentropic coordinate serves only in computing diagnostics and is advantageous because it remains constant for adiabatic processes, and thus zonal mean statistics averaged along the potential temperature surfaces show the diabatic component of circulation as vertical motion. This is important since adiabatic motion gives rise to linked pressure and density (mass) variations that when convolved with the velocity need not reflect the net mass flux. To this end it is known that large scale eddies are mostly adiabatic in character, and it is the small residual (diabatic) circulation which is important for meridional mass transport [*Tung*, 1982]. As such in potential temperature coordinates the underlying mass transport is revealed immediately by removing the adiabatic component, and in this way presents a more concise view of the so-called residual circulation [see e.g., *Holton*, 2004; *Vallis*, 2007]. Figure 2 shows the computed isentropic stream function in units of kg s^{-1}, and indicates the overturning circulation of the hemisphere as flow parallel to the contours. Notice the accent in the tropics where there is strong latent heating and decent at high latitudes associated with radiative cooling. Along with the total stream function, Figure 2 shows contributions to the

zonal mean flow and asymmetric eddies (i.e., $\psi = \overline{\psi} + \psi'$ where $\overline{\psi}$ is the zonal mean component and ψ' is that associated with deviations from the zonal mean along the isentropic coordinate). Notice specifically that the eddy component is in the same sense as the zonal mean component, which differs from the more traditional (pressure coordinate) view in which the Ferrell cell appears. The gross (one way) mass flux at any latitude is the maximum value in the vertical, while the total mass flux circulating around the cell is simply the maximum value of the entire stream function field. The sign convention is that positive values indicate anti-clockwise circulation.

[10] In a potential temperature coordinate, the Earth's surface is not constant. To indicate variations in the potential temperature of the Earth's surface, a frequency distribution is shown in the figures. Contours indicate the percentage of time that the surface potential temperature is lower than the coordinate potential temperature. For instance, the largest contour of 90% indicates that the surface potential temperature is lower 90% of the time. As such the 50% contour is the median surface potential temperature, and is shown in bold. The surface frequency is computed before taking the zonal and time means by finding what fraction of each potential temperature layer is above the surface potential temperature at each instant.

[11] In the winter hemisphere both the tropical circulation cell and the midlatitude eddy-driven circulation are modeled to be strong. Southern Hemisphere wintertime simulation of the tropical cell is stronger than observations by around 12%, while the eddy driven cell is weaker by around 25% [cf., *Schneider et al.*, 2006]. Beyond model deficiencies, this mismatch is in part due to the contrast of perpetual summer and winter simulations to the observations. The model correctly captures the much weaker summertime circulation with much diminished eddy and zonal mean contributions, as can be seen for the Northern Hemisphere in Figure 2a. While about the same as observations in the tropics, the weak Southern Hemisphere summer cell is underestimated by around 30% in the midlatitudes (Figure 2b). It is useful to note that the Antarctic tropopause is poorly modeled due to the model's vertical resolution, yet reasonably well captures the more isothermal wintertime atmosphere at the pole.

[12] Also shown in Figure 2 is the zonal mean moisture transport represented by the moisture flux in the latitude-potential temperature plane (e.g., vq, $\dot{\theta}q$ for the water vapor mixing ratio q and vertical velocity $\dot{\theta}$, and again decomposed into zonal mean and eddy components). In both hemispheres there is a net poleward transport. In the tropics this is associated with the upper branch of the tropical cell, while transport across the midlatitudes to the polar regions is associated with eddy activity and the midlatitude cell. At all latitudes eddies drive vertical transport. Again the magnitudes match reasonably well with the observations [*Schneider et al.*, 2006]. Because water vapor flux is a divergent quantity, the deviation of the moisture transport vectors from parallel with the mass flux contours is associated with local moisture divergence (i.e., condensation) and, consequently, diabatic heating. Notice that this isentropic depiction clearly separates an upper-level poleward transport path from the low-level transport that returns to the tropics, and immediately supports the conceptual view of isotope transport as being associated with two distinct one-way fluxes. Moistening in the lower branch is associated with the zonal mean flow, while condensation aloft is associated with eddies.

[13] Figure 3 shows the simulated terms contributing to the water budget, and in particular the supply by evaporation, vertical transport by turbulence in the boundary layer and loss by condensation. To illustrate the convenience gained by this isentropic depiction, we write the mass balance equation for water vapor mixing ratio (q) for some region of the atmosphere above the surface layer as the sum of evaporative supply and transport by boundary layer turbulence, condensation loss and advection:

$$\frac{\partial q}{\partial t} = \left[D_e(q_s - q) + D_t \frac{\partial^2 q}{\partial z^2} \right] - (1 - f)C + A \tag{2}$$

[14] The first term on the right represents surface evaporation where D_e is a conductance between the low level atmospheric vapor and saturated water at the surface (q_s) (Figures 3a and 3b). The diffusion term indicates turbulent transport from the surface layers to the free atmosphere with an effective diffusivity D_t (Figures 3c and 3d). We shall find it convenient to combine these two terms below. A is the rate of moisture convergence due to advection, C is the condensation rate and f is the fraction of condensate retained by evaporation of falling rain. Net water loss by precipitation loss is $P = (1 - f)C$ (Figures 3e and 3f). While f is known to be important in the tropics [*Worden et al.*, 2007], it is less important at higher latitudes and can be regarded as trivial for simplicity (i.e., $f = 0$). The advection term is the convergence of the moisture flux vectors shown in Figure 2.

[15] Latent heat exchange at the surface deposits water in the surface layers of the atmosphere (Figures 3a and 3b). The model shows this is typically below the mean potential temperature of the surface (i.e., appealing to the definition of potential temperature, when it is cold or surface pressure high). The fact that the equatorward return flow is below the mean surface potential temperature has been investigated by *Held and Schneider* [1999] in the context of momentum balance. From Figure 2 the division between the upper and lower branches of a two-stream model is seen to be near the 90% surface frequency contour. Here we find that as the return flow warms toward the equator via sensible heating (seen as in increase in potential temperature to from high to low latitude), latent heating also occurs. This result implies that water evaporated near the Antarctic coast is unlikely to have substantial contribution to the inland precipitation as the mass flux is in fact northward near the surface. Only with strong vertical mixing, say over the sea ice pack or nearby Southern Ocean, will inland precipitation be dependent on this high latitude water source. This dynamic constraint has been explored by *Noone and Simmonds* [2002b], where an argument based on buoyancy and the Froude number was used to show the difference in the isotopic composition of snow at coastal locations versus inland locations on the basis of the ability of a poleward moving air mass with a characteristic velocity to overcome thermal stratification (a potential temperature gradient) in the presence of topographic forcing. Similarly, *Noone and*

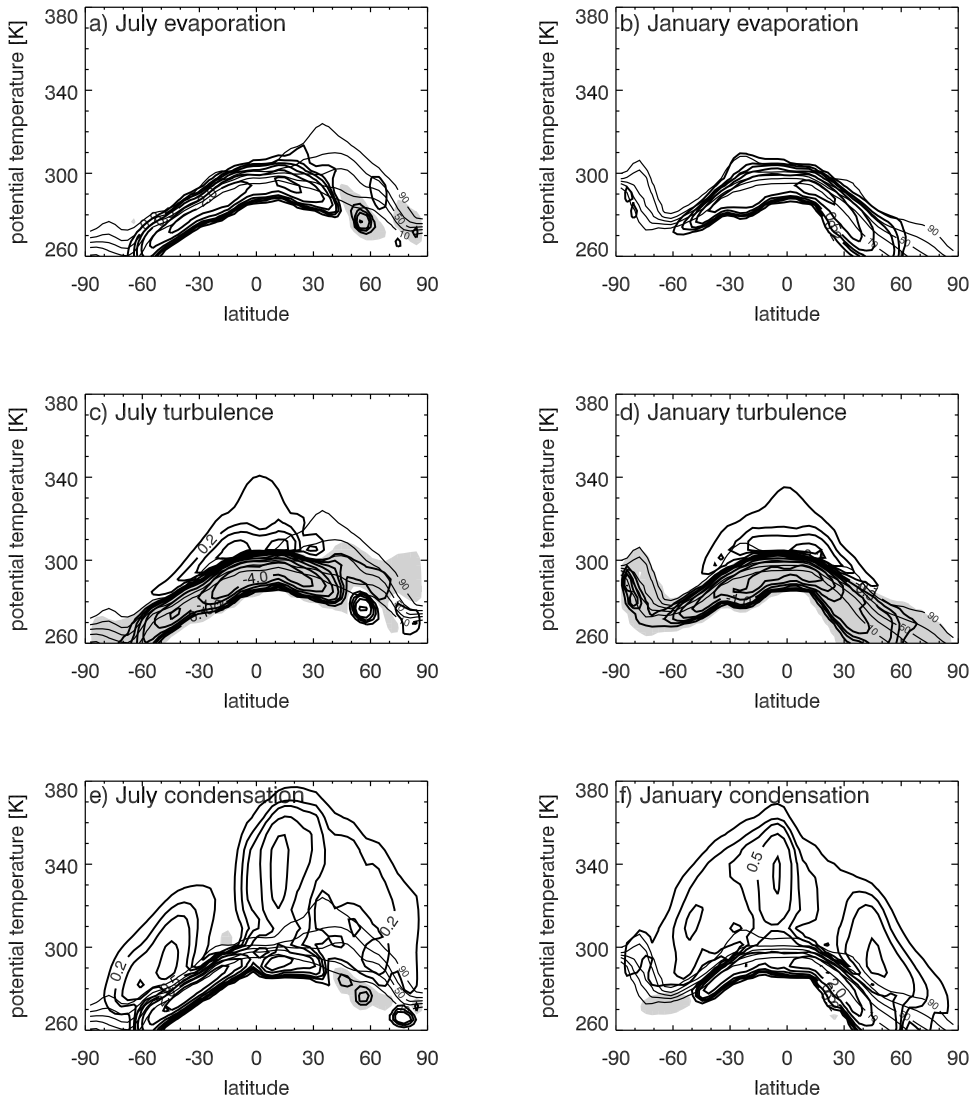

Figure 3. Components of atmospheric moisture budget as change in water vapor mixing ratio (g/kg/day). Negative values are shaded and the zero contour is omitted. Components are (a and b) surface latent heating, (c and d) transport by turbulent mixing and (e and f) divergence due to condensation. Notice sign convention that positive divergence is a loss of moisture in e and f. Panels on the left are for July and on the right for January. Surface frequency is shown as in Figure 2.

Simmonds [2004] showed that this indeed leads to sea ice having limited affect on the isotopic signature of the continent interior, although they argued also that sea ice changes the surface heating and thus modifies the turbulent mixing over the ice pack, which leads to some isotopic influence in the interior. Here the buoyancy limitations are seen clearly in Figure 2 where the surface potential temperature near the Antarctic coast is much lower than the poleward moving air stream which moves directly to the continent interior. This succinctly explains the coastally trapped minimum in deuterium excess (DXS, defined d = δD − 8δ^{18}O [*Dansgaard*, 1964]), as noted in model simulations and glaciological measurements.

[16] From Figures 3c and 3d, the turbulent mixing (i.e., boundary layer activity) is seen to transport near-surface moisture vertically to the free troposphere and into the poleward moving stream. Here, aloft, the moisture is transported poleward (and exactly horizontal if by adiabatic eddies) where it cools as it reaches the midlatitudes and condensation occurs in association with frontal convection and stratiform cloud. Moisture not entrained into the poleward moving stream by boundary layer turbulence continues to move equatorward and ultimately converges near the equator where ascent is in the region of convective clouds. Some fraction of water ascending in this region is expelled to subtropics and subsides in region of the descending branch of the tropical cell. Compared to a similar analysis of atmospheric Reanalysis data [*Schneider et al.*, 2006], these model results are qualitatively consistent, and quantitatively similar, which provides basic validation and confidence in the performance of the model's global hydrologic cycle. Compared to a traditional, isobaric, view, this isentropic view more clearly shows the key water transport components, as *Schneider et al.* [2006] have shown. In the

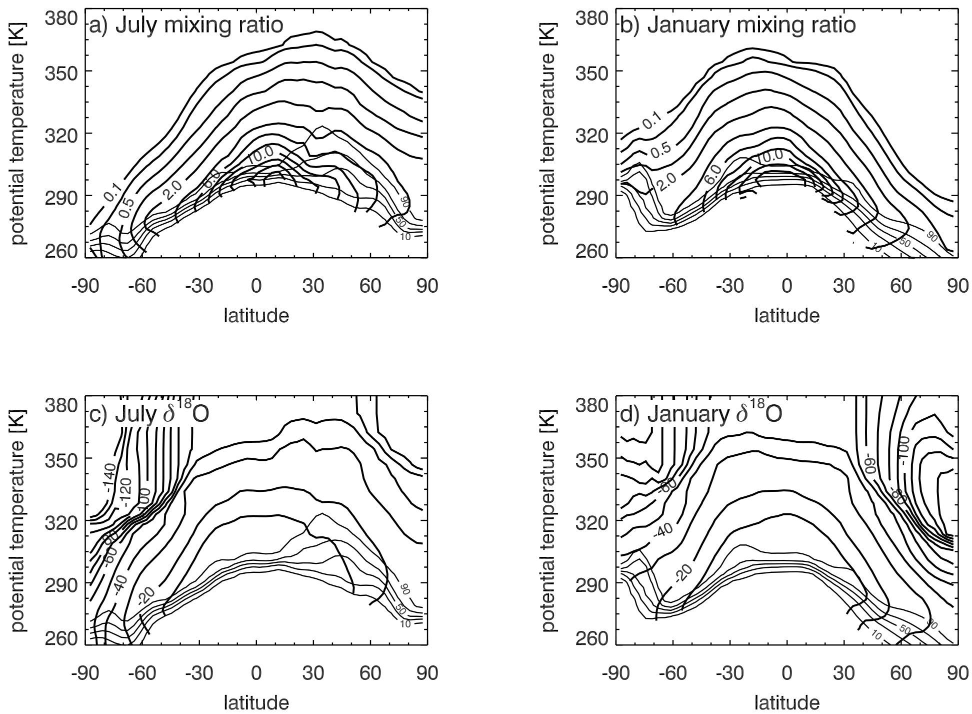

Figure 4. Modeled (a and b) atmospheric water vapor (contours of 0.1, 0.2, 0.5, 1, 2 and thereafter 2 g/kg), (c and d) isotopic composition (contour intervals of 10‰). Panels on the left are for July and on the right for January. Surface is shown as in Figure 2.

present study these results show the most important terms included in the GCM hydrological cycle are captured in the simple two-stream isotope model.

3. Depiction of Isotope Cycling

[17] The water vapor distribution and the isotopic composition of atmospheric vapor (Figure 4) is clearly understood in the two-stream context: There is a balance between the isotopic signal advected from low latitudes with condensation depleting the vapor (a Rayleigh dominated mode) and resupply of less depleted vapor from the lower layer (via non-fractionation mixing in the boundary layer) that in turn results from surface exchange (an evaporation dominated mode). *Worden et al.* [2007] examined satellite observations of HDO which supports this view, by noting that their cloudy observations are associated with poleward flow and condensation, and their clear-sky observations are associated with evaporation. Note that in both seasons in both hemispheres the near-surface water (that within the range of potential temperatures spanned by the surface frequency distribution) is dry and depleted relative to the vapor higher up. As this air moves equatorward it becomes moist and less depleted. Near the 90% surface frequency contour, both water concentration and isotopic composition are at their maximum values. This is indicative of the transition between continual gain of (less depleted) water at the surface which occurs when the surface potential temperature is low and loss of water aloft. Specifically, once further aloft and well in the poleward moving air stream, the water and isotopes become depleted again as condensation occurs with transport to colder environments. The role of the mixing between the two branches may be viewed as the mechanism that supplies surface water to the poleward moving air mass. Should this mixing, say, increase, it might be stated that the source of Antarctic precipitation has moved poleward. Similarly, should the strength of the circulation increase [i.e., poleward mass transport, or simply, advection through association with meridional wind speed in (1)], there is less opportunity for the turbulent mixing to enrich the poleward moving air stream and a lower latitude water source would result. Such an argument is in agreement with the common view of DXS variations on long (orbital) timescale, yet the present view describes the mechanism succinctly. Indeed supporting this view, *Galewsky et al.* [2005] showed that at least half the vapor in the subtropics originated from substantially higher latitudes. The behavior of the hydrological cycle described here is not seen as clearly with statistics in isobaric coordinates because the two distinct air streams are not well separated.

[18] The isotopic budget that parallels the budget for water vapor, is written in terms of $\delta = \ln R \approx R - 1$, where R is the normalized isotopes ratio $R = R'/R_{std}$, with $R' = \gamma q_i/q$, and R_{std} is a standard isotope ratio (for instance, Vienna Standard Mean Ocean Water). Subscript i denotes the heavy isotope and γ is the ratio of molecular weights of the two isotopologues in question. Rewriting (2) for isotopes in the form of a budget equation for the quantity $q_i = Rq$, the identity $\partial(Rq) = R\partial q + q\partial R$ can be applied to express the evolution of the isotope ratio R, and thus the δ. The isotopic form for net evaporative supply is $E_i = \eta D(R_s q_s - Rq)$, where $D = D_e + D_t$ and R_s is the isotope ratio of source vapor (e.g., that in equilibrium with ocean water). The discrimination for precipitation can include the effects of

rainfall evaporation by writing the isotopic composition of the evaporative flux as that in equilibrium with the liquid rain drops, and will vanish as the fraction f approaches zero. A simple form for isotopic discrimination terms follows from the approximate definition of δ. That is, $\delta_1 - \delta_2 = \ln(R_1/R_2) \approx (R_1/R_2) - 1$. As such, we find,

$$\frac{\partial\delta}{\partial t} = \frac{D}{h}[(\delta_s - \delta) - \varepsilon_s] - \frac{P}{q}\left[(\delta_p - \delta) + \varepsilon_c\right] + \frac{A}{q}\left(\overline{\delta} - \delta\right)$$

$$\left(\delta_p - \delta\right) + \varepsilon_c = (\alpha_c - 1) + f\left(1 - \frac{\alpha_c}{\alpha_e}\right) \quad (3)$$

$$\varepsilon_s = (\eta - 1)[\delta_s - \delta + (1 - h)]$$

[19] Without loss of generality, the evaporation and turbulent mixing terms have been combined as a single source term. Like the water vapor budget, the isotope budget is seen as the sum of three terms: supply by evaporation and turbulent mixing, loss by condensation and an advective component. Kinetic effects during evaporation from a liquid source are conveniently captured by ε_s, and follow *Merlivat and Jouzel* [1979], while the isotopic composition of source vapor (e.g., that in equilibrium with ocean water) is δ_s. It is useful to introduce the humidity relative to the source, $h = q/q_s$, such that it reflects the disequilibrium at which the evaporative source replenishes the vapor. Condensation occurs at equilibrium ($C_i/C = \alpha_c\, q_i/q$) with the condensed phase removed instantly from the air mass as precipitation (δ_p) (i.e., a Rayleigh process). Fractionation during condensation is given by α_c and accounts for both kinetic and equilibrium effects. The isotopic composition of vapor upstream is conveniently taken as $\overline{\delta}$. Derivation of (3) follows as an extension of the single component models used by *Worden et al.* [2007, supplementary material]. This system is closed by assuming values for D, P and A that characterize the physical aspects of the circulation and the fractionation efficiencies that characterize isotopic exchanges. Both (2) and (3) are easily solved in the time domain analytically or with numerical methods, however, for discussion here the steady state solutions can be found immediately as

$$\hat{q} = q_s + \left[\frac{A - P}{D}\right] \quad (4)$$

$$\hat{\delta} = \frac{r_T(\delta_s - \varepsilon_s) - r_p\left(\delta_p + \varepsilon_c\right) + r_A\overline{\delta}}{r_T - r_p + r_A} \quad (5)$$

[20] The term in parenthesis in (5) represents the depression of the ambient mixing ratio from saturation and immediately expresses the balance between supply and loss. This is shown as the competition between the rates of precipitation and advection and the efficiency at which turbulent transport returns the atmosphere to saturation. While the denominator in (5) can be further simplified with the aid of (4), the key result of this derivation is that the isotopic composition is best characterized as a mixture of isotopic composition of three components, which too can be considered a balance between the rate of mixing with recently evaporated low level vapor ($r_T = D/h$), the rate of replenishment by advective flow ($r_A = A/q$), and the rate of loss by precipitation ($r_P = P/q$). The term r_p has been noted by *Held and Soden* [2006] to be linked to the strength of the zonal mean Hadley circulation, and named the convective mass flux. Since the condensation rate is linked to the advection rate given some background temperature gradient and saturation, this model demonstrates the isotopic composition results simply from competition between the mixing of surface evaporated water through the boundary layer and the poleward advection of moisture. Adequately accounting for this important aspect is a central motivation of developing the two-stream conceptual model. This balance argument contrasts with more familiar Rayleigh-type isotope models in which it matters only what is chosen as the starting and ending water vapor mass (related to temperature upon making assumptions about humidity), and not the rate at which that air mass moves poleward.

[21] This alternate description of the global isotope hydrology further differs from a more conventional "trajectory" depiction because it explicitly includes the balance of source, sink and advective transport in a complete (and closed) cycle. Through understanding of the evaporative source in addition to the condensation in the context of a closed circulation, the need for selection of an initial isotopic condition that can confound simple Rayleigh distillation models [*Jouzel and Koster*, 1996] is removed. Instead the present result that emphasizes the balance between turbulent exchange and advective flux, is cast as a boundary condition problem rather than the initial value problem posed by the Rayleigh view. In the analysis below we make use of this fact by noting the advective rate is approximately proportional to the mass flux and associated with the magnitude of the mass flux stream function.

4. Forced Changes to Overturning Circulation

[22] MUGCM simulations are perturbed with sea surface temperature anomalies chosen such that strengths of the tropical cell and midlatitude cell are each modified mostly independently of the other. Zonal mean anomalies are imposed with a spatial distribution shown in Figure 5. In each case, model simulations are performed with anomalies of amplitude ±0.5, 1, 2, 3, 5 and 10 K. It should be noted the surface areas in which the SST anomalies are applied differ. As such the heating due to anomalies of the same amplitude differs, and that the anomalies are imposed over the ocean only so not perfectly zonal. The set of three tropical anomalies (each having a Gaussian shape and a width given by a standard deviation of 10° latitude) are chosen to enhance the circulation associated with the hemispherically symmetric component (TRP), as well as the asymmetric ("monsoonal") component [*Dima and Wallace*, 2003] through placement of the anomalies off the equator. Motivated by *Lindzen and Hou* [1988], these anomalies are placed near the upwelling branch of the tropical cell—north of the equator in July (10°N, TRN) and south of the equator in January (10°S, TRS). Experiments in which the eddy driven circulation is changed make use of SST anomalies that modify the midlatitude meridional temperature gradient (and thus baroclinicity) in three different ways: First, the hemispheric temperature gradient (TGR) is adjusted with the second associated Legendre polynomial (i.e., quadratic in the sine of the latitude); second, a half-Gaussian centered

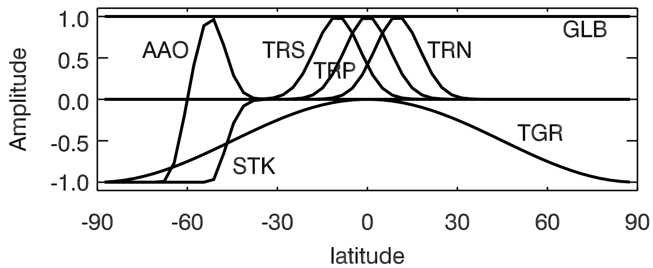

Figure 5. Structure of zonal mean sea surface temperature anomalies. Each anomaly is imposed in the model with amplitudes of ±0.5, 1, 2, 5 and 10 K. Tropical anomalies (TRS, TRP, TRN) are Gaussian with a standard deviation (half-width) of 10 degrees latitude centered on 10S, 0N and 10N. The AAO anomaly is the sum of two Gaussians of opposite sign centered on 55S and 65S, and set to be -1 south of 65S, while the STK anomaly is a half-Gaussian step change. The equator-to-pole gradient anomaly (TGR) is proportional to the square of the sine of latitude.

at 50°S with polar temperature offset to enhance the temperature gradient only in the region of the storm track (STK); and third, an anomaly constructed as the sum of Gaussians of different sign and offset by 10° latitude to yield a pattern somewhat reminiscent of the annular mode (AAO). Notice that these specifications provide trivial forcing south of 60°S since it is over the sea ice and the Antarctic continent where the surface temperature is predicted rather than prescribed. In a final experiment set, the global mean SST is changed (GLB). With 12 amplitudes, seven anomalies and two perpetual seasons (January and July), a total of 168 model experiments were performed in addition to the two seasonal control simulations. This is a computational task achievable only with moderate resolution simulations.

[23] Examination of the total mass flux confirms the anomalies produce the response in the circulation is as expected. Experiments with tropical forcing influence the tropical circulation while those with forcing in the midlatitudes exhibit modified eddy driven circulation. Common to all experiments, the responses are stronger in the wintertime when the circulation is well established, while in summer the responses are small, suggesting that the heat is more readily dissipated locally by non-dynamical influences (e.g., convection, radiation) in that season. Figure 6 shows the temperature response in July from the +5 K AAO and the +0.5 K TRN experiments, and shows changes that characterize the experimental suite. In the tropics, the increase in the strength of the zonally symmetric circulation is accompanied by stronger convection along ITCZ, which leads to warming in the upper troposphere due to latent heating. In the midlatitudes, the increase in the strength of the eddy circulation leads to warming over the Antarctic coastal region and over the continent, and is a direct response to increased heat transport from lower latitudes.

[24] For the sake of completeness, the main features of other experiment sets are described here in summary form. As expected from *Lindzen and Hou* [1988], heating anomalies located off the equator have a greater effect on the circulation strength than a heating anomaly located at the equator (that forces only the weaker hemispherically symmetric part of the circulation) and largely explains the resulting strength circulation of the asymmetric TRN and TRS experiments relative to the symmetric TRP experiment. In the TGR experiment, again changes in the strength of the eddy circulation are found as desired, but changes in the tropical circulation also result. Detailed examination reveals this is associated with wave activity in the region of the descending branch of the tropical cell. Similarly, where the global mean surface temperature is increased uniformly, changes result in both the zonal mean circulation in the tropics and the midlatitude eddy circulation. On inspection of the heat and momentum transport, this is found to be related to an increase in the stationary wave activity since the land-sea contrast has been modified by prescribing ocean temperature anomalies. Indeed this effect is seen to be larger in the Northern Hemisphere where the stationary wave contribution to the mean flow is larger than in the Southern Hemisphere.

5. Interpretation of Antarctic Isotope Records

[25] The Antarctic interior is defined to be the area mean for all model points above 1500 m elevation on the Antarctic continent. Figure 7 shows the Antarctic δ^{18}O and DXS as a function of the zonal mean and eddy circulation strength from all model experiments. Results from all experiments are plotted, and experiment sets that differ only in amplitude are joined by lines. For the discussion that follows it is not

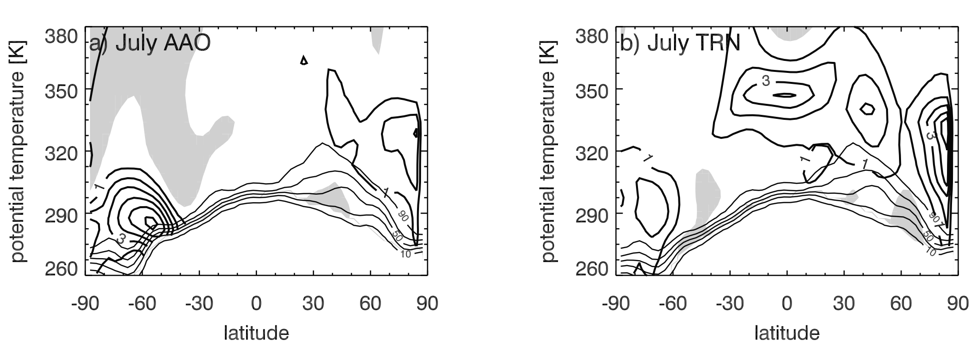

Figure 6. Experiment minus control difference in temperature in July for (a) +5 K AAO experiment and (b) +0.5 K TRN experiments. Contour interval is 1 K. Negative values are shaded and the zero contour is omitted. Surface frequency is shown as in Figure 2.

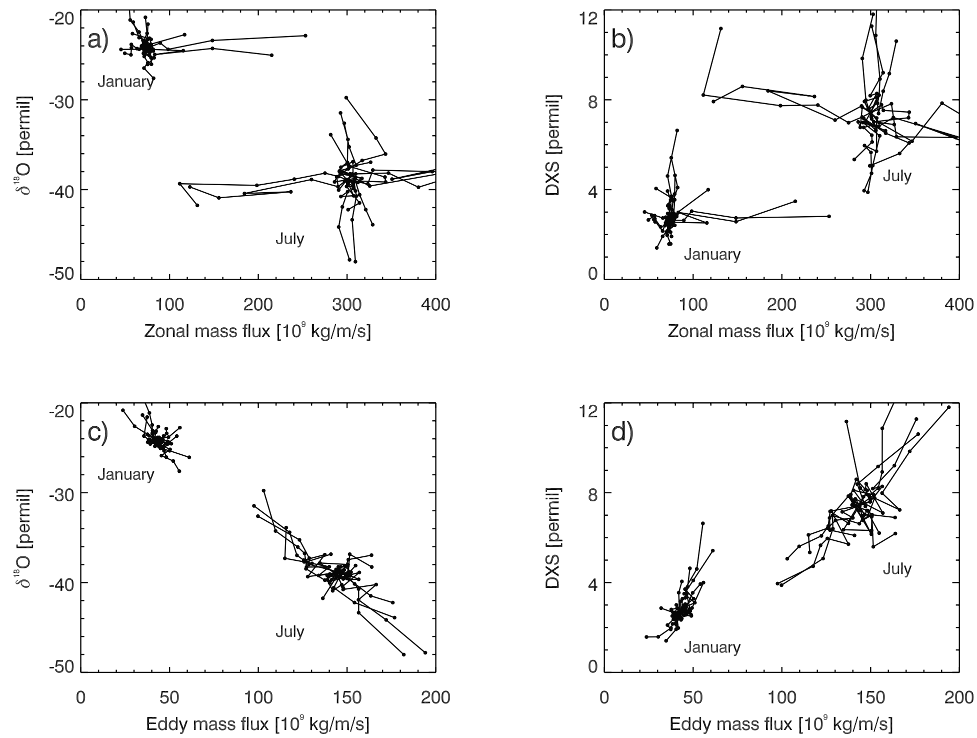

Figure 7. Antarctic interior (above 1500 m) δ^{18}O (a and c) and DXS (b and d) in precipitation as a function of zonal mean circulation (a and b) and eddy driven circulation (c and d). Individual experiments are marked as dots with experiment sets connected by lines. Clusters of points are shown for both perpetual January and July simulations.

necessary to identify specific experiment sets, only the distribution of points that characterize a wide range of climate regimes. As noted above in reference to Figure 2, the strength of each component of meridional circulation is the maximum value of the mass flux stream function (at any southern latitude). For the δ^{18}O, a clear association between the strength of the eddy circulation exists over the range of circulations examined. A similar association exists on seasonal timescales, with the weakened summertime eddy circulation associated with less depleted precipitation. The figure also reveals that the tropical circulation plays little role in the isotopic composition of Antarctic snow. In experiments where the strength of the tropical circulation is changed, the isotopic depletion is almost constant; conversely, in experiments where the isotopic depletion has changed (in association with eddy strength) the tropical circulation has changed little. Where the isotopic composition appears to change with the tropical circulation, analysis shows this to be associated with the effect of eddies on both.

[26] A similar argument follows for the DXS. Since DXS arises due to kinetic isotope effects, changes are expected where the rate of evaporation of falling rain, the amount of condensation to ice or the disequlibrium at the evaporation source has changed. As suggested by Figure 6, for the Antarctic interior this is mostly in association with temperature changes over Antarctica since warmer temperature will reduce the supersaturation within ice clouds and thus reduce the strength of kinetic isotope effects. As such, this again leads to a strong signature of the eddy circulation via its role in the temperature budget of the polar region. While there was a similar DXS change in the upper-tropospheric vapor with changes in tropical circulation and the convective warming seen in Figure 6, the influence on Antarctic precipitation is small. Similar to the δ^{18}O, there is a correlation between seasons that broadly fit this depiction since the cold temperature during winter allows stronger kinetic fractionation over the Antarctic interior. In all cases the scatter in DXS is larger than that in the δ^{18}O, reaffirming that the DXS signal is a more complex combination of condensation and evaporation histories. Specifically, in addition to changes in distillation history, the scatter suggests a stronger influence of the source conditions—including the regime of boundary layer turbulence.

[27] Motivated by this result, Figure 8 shows the DXS as a function of δ^{18}O for all experiments. The strong relationship between the two species is striking given the very wide range of circulations computed. The curvature of the scatter is reminiscent of the increase in the DXS forcing as temperature decreases and the equilibrium fractionation of the two species diverges from the nominal 8-to-1 ratio typical of meteoric water. As such, the model results suggest there is little additional information from DXS, and indeed either quantity is adequate to deduce changes in the eddy circulation strength. This result is at odds with glaciological measurement that show substantial information gain with DXS [*Vimeux et al.*, 2002], and suggests that the parameterization of kinetic fractionation in the present generation of isotope models is at least inadequate.

[28] Figure 9 shows the relationship between precipitation and temperature and the two isotope quantities. Tem-

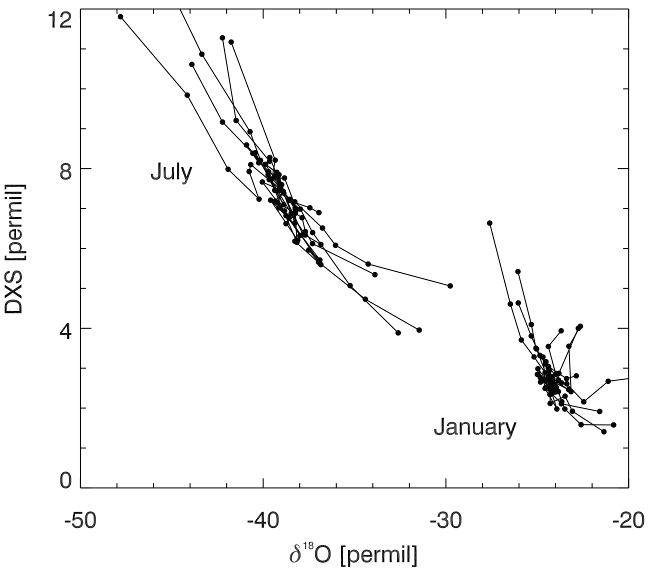

Figure 8. Covariation in δ^{18}O and DXS of Antarctic precipitation from all model experiments. Individual experiments are marked as dots with experiment sets connected by lines. Clusters of points are shown for both perpetual January and July simulations.

perature is the mean over the Antarctic continent from the lowest level of the model (about 75 m above the surface). As is well known, there is a strong association between temperature over the Antarctic continent and a single isotope both within seasons and between seasons. This figure shows that the seasonal "slope" is a reasonable estimate of the change in δ^{18}O with respect to circulation strength (and specifically the eddy strength), and therefore a useful candidate for use in empirical reconstructions on continental scales. In July the relationship is 0.85‰/K (r^2 = 0.93) and in January it is 0.69‰/K (r^2 = 0.85). Based on simulations from all experiments in both seasons, a slope of 0.60‰/K (r^2 = 0.98) is found, and appears valid across the exceedingly wide range of circulation regimes. However, beyond confidence in empiricism, the model analysis gives insight as to why this association is well founded. Specifically, both the isotopic composition and the temperature are fundamentally linked to the (eddy) circulation. Correlations with precipitation are also quite strong, but the slope within a season (especially July) differs from that between seasons, and as such there is less confidence in universality of regression models for precipitation (or, similarly, accumulation). Instead this result demonstrates that rather than simply the balance of transport processes, other differences in the hydrological cycle between seasons (particularly cloud properties and source distribution) influence the precipitation strongly.

[29] The DXS is again more difficult to interpret, although seasonal differences are consistent with seasonal temperature changes, and follow from the modeled dependence on supersaturation. However, across the range of experiments there are different slopes within the clusters of experiments, which is further evidence of the role of seasonal changes in source conditions within the different simulated circulation regimes.

6. Discussion and Conclusions

[30] Description of the isotopic composition in terms of zonal mean circulation strength is satisfying because it reflects the underlying transport mechanisms of the extratropical atmosphere. These mechanisms are captured well by a conceptual view that accounts for the two one-way transport streams. In particular, condensation occurs aloft during poleward transport and evaporation occurs during equatorward return flow along the surface, with turbulent exchange between layers. Weaker correlations between the isotopic composition of Antarctic interior snow and the strength of the overturning circulation in summer suggest the isotopic budget is less strongly controlled by large-scale transport during that time, and is more likely influenced by more regional and local sources and condensation conditions – that is, in reference to equation (5) the ratio of the advection rate to the turbulent exchange rate (i.e., r_A/r_T) is larger. In both January and July, the model results reveal the isotopic composition of Antarctic snow has little dependence on the tropical circulation. This is because the rate of cross-isentropic mixing in the midlatitudes exceeds the rate of advective transport from lower latitudes (i.e., the time required to replenish vapor by mixing is less than the timescale of advection). Thus when the poleward moving air stream reaches Antarctica, the original tropical air is almost completely diluted by water from higher latitudes. This interpretation is consistent with model and observationally based estimates of source regions, while concisely captures the essence of the underlying transport mechanisms.

[31] The methodology implicated here for interpreting the ice core record has still not addressed the importance of the details of cloud characteristics in providing a refined understanding of polar isotopes. Given the lack of simulation performance for DXS, this is one area in which DXS observations can provide specific guidance. Further, this study has not addressed the degree to which continental scale signatures are masked by local effects including pre and post depositional processes, nor the influences of organized asymmetric circulation patterns which are known to be important in considering ice records [e.g., *Schneider and Noone*, 2007]. Instead, the theoretical association between eddy strength and isotopic depletion established here strongly supports the view that the isotopic conditions are an excellent and robust proxy for polar mean temperature over a very wide range of climates. The association follows because the temperature gradient is both a consequence of eddy activity, and central to the development and maintenance of baroclinic disturbances that drive the eddy circulation. Specifically, the same circulation that transports isotopes also transports momentum and heat, which in turn controls the strength of the jet and the circulation itself. As such, should adequate balance conditions for the dynamic and thermodynamic properties of the atmosphere be assumed (for instance hydrostatic balance, approximate geostrophy and vorticity conservation), a consistent explanation of atmospheric circulation and isotopic composition can be obtained. As a case in point, *Thompson and Solomon* [2005] showed a recent trend toward the positive phase of

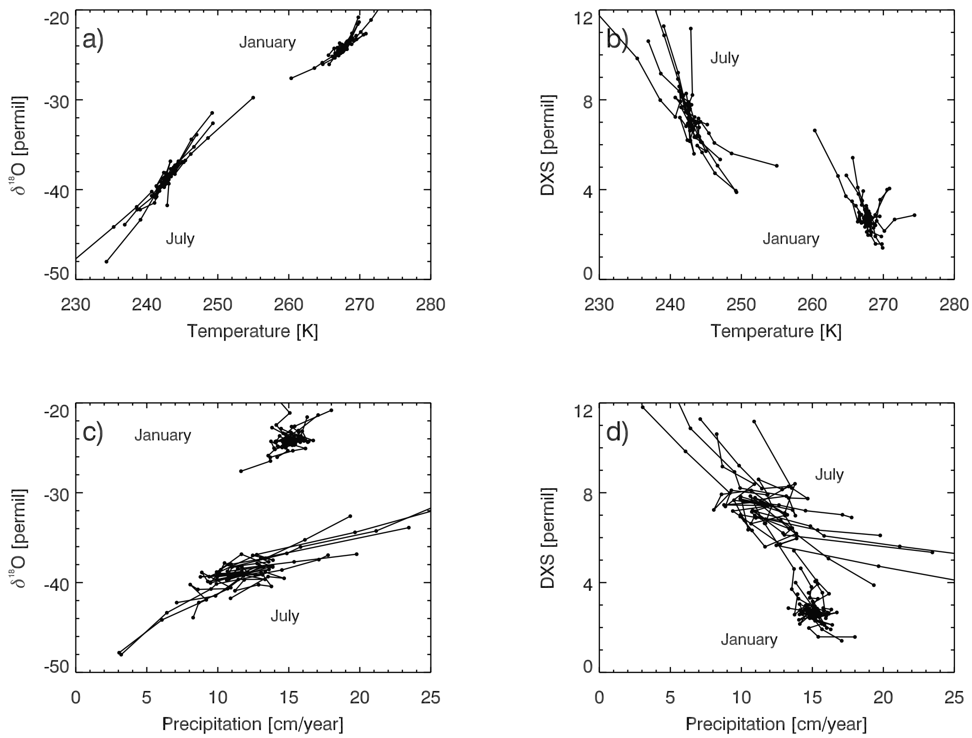

Figure 9. Variation in δ^{18}O (a and c) and DXS (b and d) as a function of temperature (a and b) and precipitation (c and d). Individual experiments are marked as dots with experiment sets connected by lines. Clusters of points are shown for both perpetual January and July simulations.

the SAM, while [*Schneider et al.*, 2005] showed an increase in the isotopic composition of an ice core in West Antarctica that was argued to be linked to the SAM. Recently, *Schneider and Noone* [2007] have confirmed this association over the Antarctic continent by considering variations in a network of high resolution cores. A stronger SAM pattern (more positive SAM index) is associated with a poleward displacement of the storm track and the intensification of the westerly jet, and is of greatest influence during the wintertime. While the mechanistic linkage between the annular mode and isotopic composition is known [*Noone and Simmonds*, 2002b], the findings here suggest that these two results can be adequately reconciled (at least qualitatively) as a decrease in the strength of the eddy circulation.

[32] In addition to polar temperature, it is also appropriate in the discussion of the DXS to consider the role of airborne particles on supersaturation and consequently the kinetic fractionation. Specifically, the supersaturation is typically parameterized in models (including the model used here) as dependent on temperature alone. Phenomenologically, supersaturation is also dependent on the abundance of suitable condensation nuclei. No account of change in such aerosols is included in the model, so the result that modeled DXS is not independent of δ^{18}O is hardly surprising and likely a model deficiency. This may explain why DXS in the present generation of isotope models is so closely linked with temperature and does not provide independent information. Since again the transport of condensation nuclei from lower latitudes will be associated with the eddy circulation, the present study suggests there is potential for polar DXS records to give insight into global scale aerosol loading and consequently an extended history of the role of aerosol effects on climate. Nonetheless, the strong associations found between a single isotope and the circulation strength suggest that isotope records from cores in the Antarctic interior can be viewed as reflecting the mean overturning circulation of the mid latitudes rather than simply the local temperature. This gives guidance to examining historical changes in the bulk features of the hemispheric circulation, such as the strength and location of the westerly jets and storm tracks through understanding the role of the overturning circulation in the momentum and heat balance of the atmosphere.

[33] **Acknowledgments.** This work was motivated in part by discussions with Kurt Cuffey of University of California at Berkeley. Conversations with Christopher Walker and Tapio Schneider at the California Institute of Technology motivated diagnosis on isentropic coordinates.

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