CLIMATOLOGICAL CONTROLS ON THE CHEMICAL STRATIGRAPHY OF THE MARTIAN POLAR LAYER DEPOSITS.

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Introduction

The growth of the North Polar Layered Deposits (NPLD) at the expense of tropical ice is thought to have started approximately 4.5 Myr ago [1] when the mean obliquity dropped from $\sim$35° to $\sim$25°[2]. During this epoch, the obliquity has varied between 15° and 35°, and the eccentricity between 0 and 0.13 [2]. Alternating layers seen in the NPLD in images [3] and radar data [4] provide an archive of past climate oscillations, that are widely held to be linked to orbital variations[5, 1, 6, and many others]. Ice also exists in the subsurface; both theoretical and observational work show that shallow ground-ice is abundant in the mid to high latitudes [7, 8, 9] The volume fraction and extent of the subsurface ice can alter the atmospheric energy budget by storing and releasing heat at different times, and this in turn can influence the physical and chemical evolution of the polar caps [9, 10]. On Earth, The isotopic signal in ice cores from Antarctica and Greenland have revealed past climate oscillations [11]. Mars lacks an ocean to buffer the atmospheric isotopic ratio as on Earth, thus the isotopic composition of the present atmosphere is controlled by the evaporative fluxes of the caps. Observations from orbiters and ground-based telescopes measured the atmospheric D/H ratio to be in the range of 1-8 $\times$ Vienna Standard Mean Ocean Water (VSMOW) [12, 13, 14], with an average of 4.6$\pm$0.7 [14]. The atmospheric values provide evidence for both net water loss in the long-term evolution of Mars [15, 13], as well as seasonal, cloud physics [16], and geographic source effects. Previous work [5, 17] shows a D/H record in the polar deposits is expected, although potentially complex to interpret than Earth's due to the aforementioned effects. In this work, we seek to quantify the processes that affect the D/H ratio of the accumulating ice on the PLDs, in order to provide a framework for interpreting the chemical signal in a vertical profile of polar ice.

Model

The results reported in this paper are based on simulations of the Mars LMD-GCM with the full water cycle that includes treatment of surface ice, atmospheric vapor and ice clouds and has been described in detail previously [18, 19, 16, 20, 17]. In addition to dust and water tracers [16], we use the HDO tracer [21] in both the vapor and ice phases to track the D/H ratio. Temperature-dependent isotopic fractionation of water occurs at condensation either in clouds or directly onto the surface. The equilibrium solid-vapor fractionation coefficient (α) assumed is based on lab measurements [22]. Ice clouds are taken to be radiatively passive. The simulations reported here have a horizontal resolution of 64 by 64, corresponding to 2.8125°in latitude, and 5.625°in longitude, with 29 vertical layers extending to 80 km. The simulations were performed with an assumed past ice distribution in which surface ice is present in the tropics at locations on the eastern flanks of the Tharsis rise where accumulation is predicted at high obliquity [23]. The presence of such a past tropical surface ice reservoir is supported by geologic evidence [24]. We tested thick and thin initial tropical ice layers. In the thick case, the tropical ice layer has a thickness of $\sim$60 m Polar Equivalent Layer (PEL), defined as a layer of the same volume, spread evenly from 81° to 90° latitude. This case is designed to reach steady state, to investigate the fluxes and D/H ratio of ice reaching the polar cap with a persistent tropical source. In the thin test case, we initialize the model with a $\sim$0.45 m PEL tropical ice layer, in order to investigate the different stages of the evolution of ice migration as the finite initial source is exhausted.

Results

Condensation of water vapor leads to fractionation. Figure 1 shows the zonal mean seasonal evolution of atmospheric water vapor and the D/H ratio relative to the initial value of the source. We plot three cases, that are different in orbital state and ice distribution. Atmospheric humidity is significantly higher when a tropical ice reservoir is available, and peaks near mid-summer in each pole. In general, the atmospheric vapor is isotopically depleted because the HDO is either preferentially in ice clouds or in surface ice that sedimented at an earlier stage. This phenomenon is amplified when a tropical reservoir is present [17]. As described in the previous section, we performed simulations with initial conditions that assume ice deposits placed in the trop-

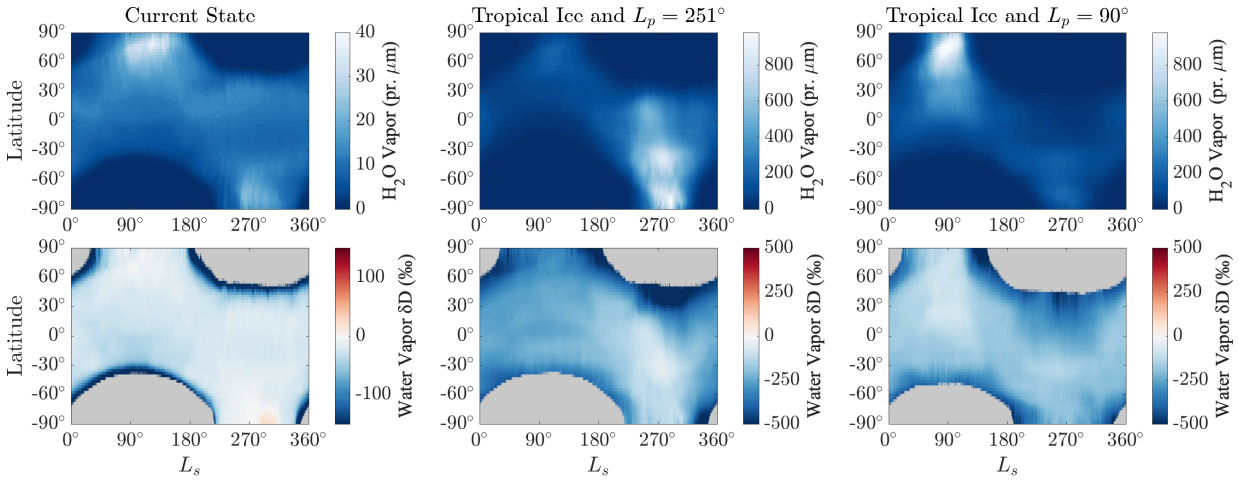

Figure 1: Zonal mean seasonal evolution of atmospheric water vapor and its D/H ratio for three different states: Current orbit and ice distribution (left column), current orbit and a thick tropical ice distribution (middle column), and an orbit with L_p=90° and a thick tropical ice distribution (right column).

ics. Figure 2a shows The NPLD growth rate for a thick tropical ice distribution as a function of obliquity, ϵ, and perihelion longitude, L_p, for present-day eccentricity. In addition to obliquity controlling the annual mean insolation as a function of latitude, and hence the sign and magnitude of polar accumulation, there is also a strong dependence on L_p. This phase controls the length and intensity of the high-humidity summer season, with a difference of a factor of ~ 2 in accumulation rate between a prolonged mild northern summer to a short intense one. We also calculate the mean D/H ratio of ice that condensed in the polar region and plot it in Figure 2b. The contours show D/H values in δD. The hydrogen isotopic anomaly depends on the orbital elements, and in particular L_p, with a difference of more than 100‰ over a precession cycle. A short, intense summer leads to higher D/H ratio of the accumulating ice. Notably, the polar δD value is negative for all cases. This occurs because as vapor transports from the tropics poleward, the water molecules first deposit in the intermediate latitudes and preferentially condense out the heavy isotope. The atmosphere becomes depleted in the HDO and therefore water reaching the pole is isotopically light. Under the assumption that the majority of the tropical source ultimately migrates to the polar cap, the depletion in HDO of the early deposits would be compensated by enrichment of subsequent deposits sourced from the intermediate latitudes. In order to test if the simulated ice core is affected by the history of the migration we considered an initial ice distribution with a smaller amount (thin tropical ice layer test case), placed in the equatorial region. We use this approach as an approximation, as the stages of the migration of this limited deposit over the simulation timescale reflects those of a much larger deposit, migrating over a much longer time (for more details on the model assumptions and the details of the ice migration see [17]). At first, as long as the tropical reservoir is present, ice migrates to latitudes higher than $\pm 45°$, but only a small portion of the ice that sublimated from the tropics reaches the polar caps. When ice is no longer available in the tropics, the planet's mean humidity drops and ice migrates from these quasi-stable regions towards higher latitudes, poleward of about $\pm 55°$. As the simulation advances and more ice migrates, the location of the source (the equator-most ice still available) migrates poleward. At the last stage, when the boundary has migrated to $\sim \pm 75°$, the ice accumulating on the caps is sourced only from these high latitude locations (either in the same hemisphere, or the opposite one). Thus, the upper layers of the cap are expected to represent a deposit that migrated from high latitudes (polewards of 70°), and not directly from the tropical reservoir. The migration stages mentioned above affect the D/H ratio of the condensed ice as can be seen in Figure 3. Different hydrogen isotopic sections appear along the profile due to the different migration stages and ice source locations mentioned above. The bottom portion of the profile (up to $\sim$0.2 ice fraction transported) is highly depleted in HDO relative to the equatorial source in both poles, due to the mid-high latitude reservoir fractionating the atmosphere as it grows. Once the tropical source is depleted, the enriched mid-high latitude deposit becomes the source, and the growing polar ice is therefore also enriched. As the polar ice grows at the expense of ever-higher latitude ice, its isotope ratio changes in accordance with the source, with a smaller modification due to heavy isotope trapping in intermediate latitudes between the source and the pole. In both hemispheres, and for all profiles examined, our simulations show that the upper layers of each PLD is enriched in HDO compared to the average D/H ratio of that PLD (dashed line). The effect is due to the HDO depletion

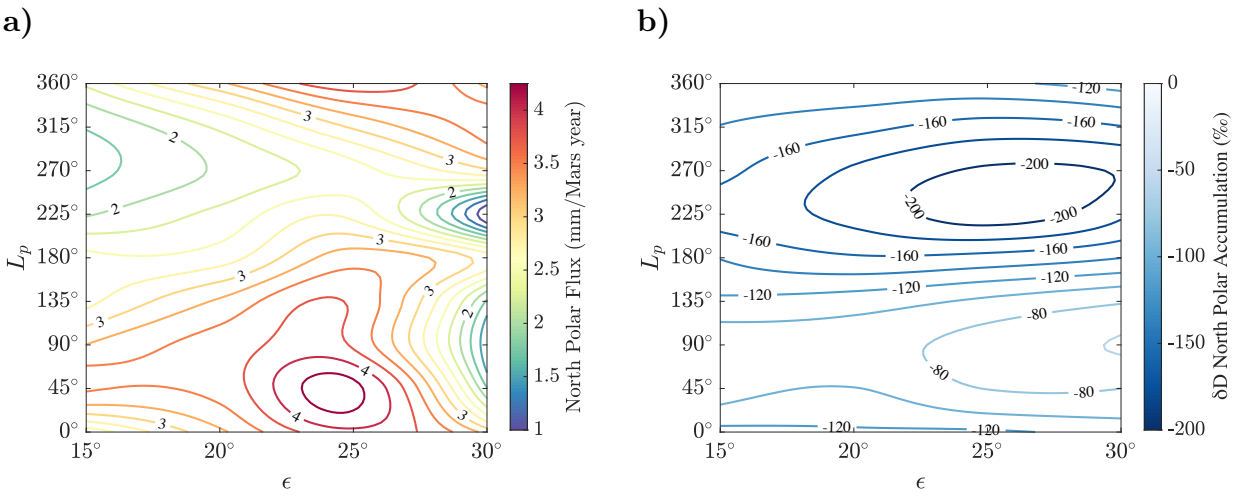

Figure 2: **a)** Simulation results of the North Polar ice flux as function of obliquity and perihelion longitude for eccentricity of 0.093. **b)** North polar ice accumulation D/H ratio for the same configuration as in a.

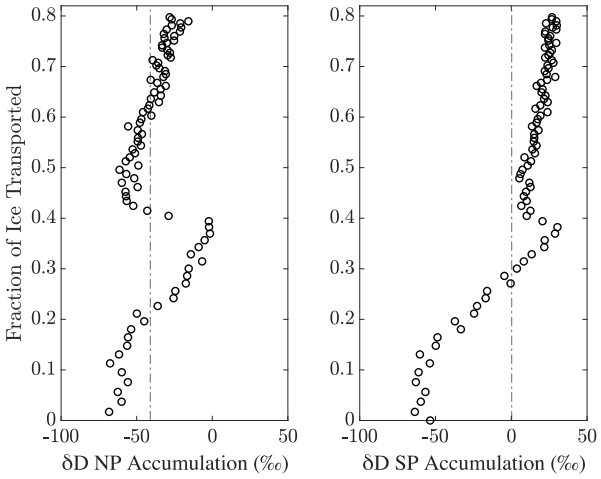

Figure 3: North and South Polar D/H profiles as a function of the fraction of ice transported for obliquity of 25°, eccentricity of 0.093 and L_p of 90° and an initial thin tropical ice distribution. The data points are separated by one Mars year, and the dashed line is the mean D/H ratio of each final ice cap.

of polar ice that accumulates early along with lower latitude enriched deposits. The simulations also show that the southern ice deposits are enriched in deuterium compared to the north. Subsurface ice can affect the NPLD physical stratigraphy via several mechanisms. It acts as a water source, as well as a heat reservoir altering the energy budget. Previous modeling work [9] that is supported by observations [7] showed that the subsurface ice distribution can extend to ∼30° or retreat to ∼ 60°, depending on the orbital configuration and atmospheric humidity. We tested how the position and distribution of the subsurface ice affects the simulated polar profile. We ran two sets of simulations, one with ground-ice at a depth of 15 cm extending to latitude 60° similar to at present, and one where the thermal inertia in the subsurface was set to mimic the presence of an ice sheet at a depth of 5 cm that extends to latitude 30° in both hemispheres, similar to past conditions. The difference between the two test cases as function of L_p is shown in Figure 4, where (a) shows the change in flux to the NPLD, and (b) the change in the D/H ratio of the condensed ice. Both the flux and the D/H ratio are higher in all L_p tested for the first case (margin at 60° latitude). This phenomenon occurs because when the subsurface extend further towards the equator and closer to the surface it acts to reduce the difference between seasons; as shown in Figure 2 a more intense summer leads to higher accumulation and a larger D/H ratio. This result supports the claim that the subsurface extent, depth, and ice volume (not shown here) will affect the physical and chemical stratigraphy of a simulated ice core. The results shown in this abstract and in previous published work [17] highlight the fact that in order to interpret the chemical stratigraphy of the PLDs a model that takes into account both the atmosphere evolution over seasonal timescales, and the ice reservoirs evolution over millennial timescales, is needed. In the conference we will discuss the development of such a model.

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a) **b)**

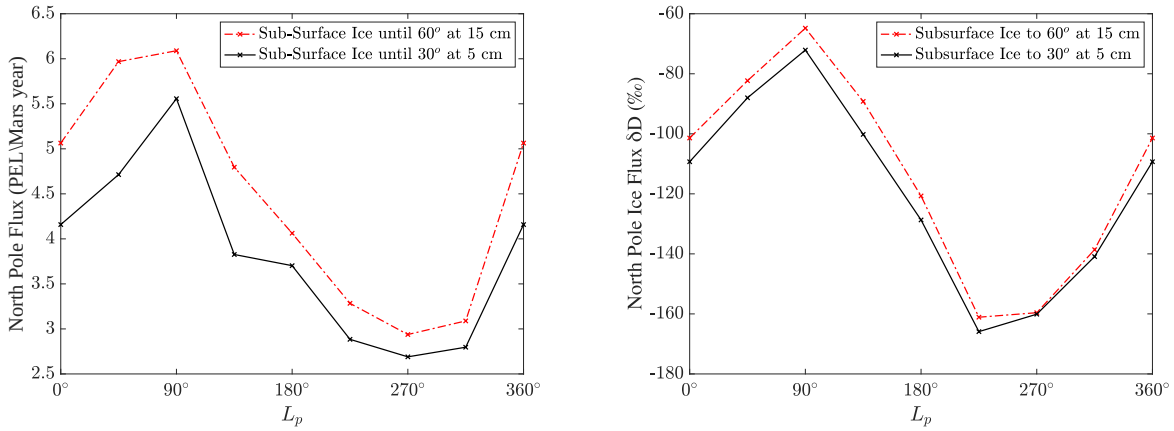

Figure 4: **a)** North polar flux as function of L_p for two subsurface ice cases, the first assumes ice from the pole to latitude 60° at a depth of 15 cm, and the second, to latitude 30° at depth of 5 cm. **b)** same as a, only the D/H ratio of the condensed ice.

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