

LETTERS

Dynamics of ice ages on Mars

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Unlike Earth, where astronomical climate forcing is comparatively small, Mars experiences dramatic changes in incident sunlight that are capable of redistributing ice on a global scale^{1–6}. The geographic extent of the subsurface ice found poleward of approximately $\pm 60^\circ$ latitude on both hemispheres of Mars^{7–9} coincides with the areas where ice is stable^{7,10,11}. However, the tilt of Mars' rotation axis (obliquity) changed considerably in the past several million years. Earlier work^{3,12} has shown that regions of ice stability, which are defined by temperature and atmospheric humidity, differed in the recent past from today's, and subsurface ice is expected to retreat quickly when unstable^{11–13}. Here I explain how the subsurface ice sheets could have evolved to the state in which we see them today. Simulations of the retreat and growth of ground ice as a result of sublimation loss and recharge reveal forty major ice ages over the past five million years. Today, this gives rise to pore ice at mid-latitudes and a three-layered depth distribution in the high latitudes of, from top to bottom, a dry layer, pore ice, and a massive ice sheet. Combined, these layers provide enough ice to be compatible with existing neutron and gamma-ray measurements^{7–9}.

The extensive subsurface ice deposits on Mars^{7–9} were anticipated on the basis of atmosphere–subsurface vapour exchange^{3,12,14}. Entirely unexpected, however, was the large amount of ground ice, 60% by weight or 70–85% by volume in the near-surface of the planet¹⁵, which is far greater than the porosity of typical soils. There are two possible mechanisms for the emplacement of the ice: (1) precipitation from the atmosphere^{2,4,5} or (2) diffusion and condensation of atmospheric vapour in regolith pores¹². The high ice content suggests that the former mechanism was at work. Indeed, climate simulations show that at high obliquity water ice from the north polar cap precipitates in the tropics^{1,2}, and that as the obliquity decreases, the low-latitude ice is redistributed to form the mid-latitude ice sheets^{4,5}. Subsequently, the ice sheets evolved; today the ice is buried and in its equilibrium position, where the vapour pressure of the ice balances the atmospheric humidity^{7,10,11}. When an ice sheet retreats, the dust it contains can form a lag deposit that protects the ice from rapid sublimation and retreat. But still, the time over which the subsurface ice is expected to adjust to drier or warmer conditions is comparable to that over which changes in Mars' obliquity and orbital parameters occur^{11–13}. Theoretical as well as experimentally measured diffusion coefficients of water vapour through soils¹³ are orders of magnitude too large to isolate ice from the atmosphere over millions of years.

Although there are two mechanisms for ice accumulation, the rate-limiting process for the loss of buried ice is always diffusion. A model of subsurface temperature and ice stability is designed (and described in the Methods). It is initiated with a cover of dirty ice, consisting of 85% ice and 15% dust. Subsequently, ice is lost by diffusion through the sublimation lag, which is assigned a realistic diffusivity¹³ of $4 \text{ cm}^2 \text{ s}^{-1}$. Ice can re-form by inward diffusion of atmospheric water, to fill the interstitial pores with a porosity of 40%. These climate simulations are computationally affordable because of a

new numerical approach that uses averaging methods for the diffusion equation, rendered nonlinear by phase transitions. Such averaging methods have been applied to the retreat of an ice table^{11,16} and can be extended to partially filled pores. The climate model does not keep track of the global H_2O balance, and every latitude is independently computed on a distributed computer cluster. It is assumed a global ice sheet formed by atmospheric precipitation five million years ago⁵, and for simplicity only a single ice sheet is considered. This climate model integrates history using time-varying orbital elements (ref. 17 and see its data at www.imcce.fr/Equipes/ASD/insola/mars/mars.html). Zonally averaged values of present-day albedo and thermal properties are used.

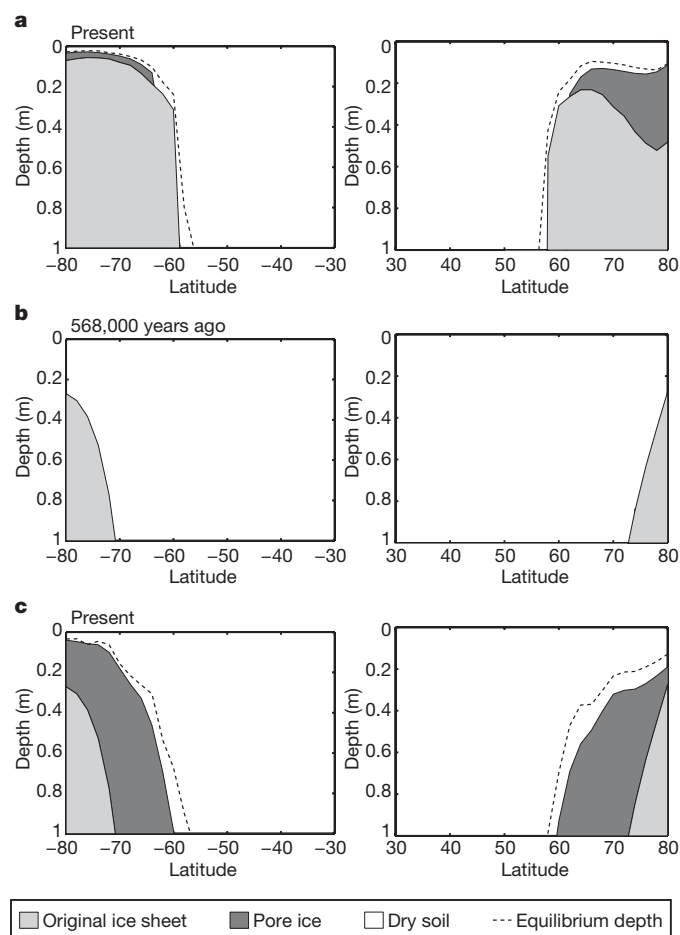


Figure 1 | Snapshots of the vertical ice distribution from model calculations. **a**, At present and with atmospheric humidity constant throughout history. **b**, For strongly varying atmospheric humidity, 568,000 years ago. **c**, For strongly varying atmospheric humidity, at present. Dashed lines show the present-day equilibrium depth.

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Figure 1a shows the resulting ice distribution, assuming constant atmospheric humidity, so that all ice movement is caused directly by temperature changes from orbital forcing. The partial pressure of water vapour on the surface is set to the present-day global average value of 0.13 Pa. There are three subsurface layers: (1) the original ice sheet remaining from the initial massive ice cover that formed by precipitation, (2) pore ice that forms by inward diffusion and constitutes at most 40% of the volume, and (3) dry soil. The depth to the ice closely follows the present-day equilibrium line, which attests to the rapidity of subsurface–atmosphere exchange. What is left from the original ice sheet closely matches the maximum equilibrium depth of the past five million years (not shown). Surprisingly, the latitudinal boundary is about as observed, $\sim 60^\circ$. The ice has not retreated further poleward, because despite large obliquity changes and large changes in annual mean insolation¹⁸, the annual mean temperature at this intermediate latitude changed little. Further poleward, temperature varied more over the last few million years, and the equilibrium ice table oscillated in depth. Trends of thermal conductivity of the soil with latitude account for the asymmetry between the southern and northern hemispheres.

Ice ages are not only driven directly by astronomical forcing via insolation changes, but indirectly through the supply of water from the polar cap. The amount of water that sublimates from the cap varies strongly with the planet's axis tilt, and as a result the atmospheric water vapour content is expected to vary greatly with time. The atmospheric vapour abundance may vary between zero, when the residual water cap is covered with carbon dioxide ice year round, and values estimated to be as high as 1,000 precipitable micrometres^{1,2}. Figure 1b and c shows the resulting ice distribution at two instances in time. Figure 1b is for a recent obliquity minimum, the last time all pore ice had disappeared. In Fig. 1c, the original ice sheet has retreated further poleward, while the pore-ice layer created by diffusive back-filling extends close to equilibrium. A comparison of Fig. 1a with Fig. 1c reveals that martian ice ages are to a large extent only indirectly controlled by orbital forcing, through the effect of insolation on atmospheric vapour content.

Figure 2a shows the evolution of southern hemisphere subsurface ice with a strongly varying atmospheric vapour content as a function of time. After an extremely rapid retreat from the tropics, ice reforms in pore spaces at high obliquity (dark grey). The retreat of the original ice sheet (light grey) proceeds during dry low-obliquity periods and is

independent of the model humidity at high obliquity. Ground ice forms and retreats with every major obliquity excursion; its maximum latitudinal extent depends sensitively on the maximum atmospheric humidity. The final stage of the evolution is that of Fig. 1c. Figure 2b shows the output and intake of water from the subsurface, summing latitudes 80°S to 80°N . The receding subsurface ice becomes a significant source of water vapour at low-obliquity periods that would otherwise be extremely dry. This could slow the retreat of the original ice sheet. At present, the subsurface contribution to the atmospheric humidity is nearly neutral, because the ice is close to equilibrium.

Existing inverse models of the elemental composition of the ground, from neutron and γ -ray spectroscopy, use a one- or two-layer vertical hydrogen distribution in the soil (in addition to the atmospheric layer), where each layer is assumed to have a homogeneous elemental composition. Figure 3 compares Gamma Ray Spectrometer Instrument Suite (GRS) measurements^{19,20} (one-layer model) with the predicted amount of ice left over from the original ice sheet or reformed recently by inward diffusion. Although neither of the two mechanisms by itself could simultaneously account for the mass fraction and latitudinal boundary of the observed ice, their combination provides just enough ice at the right places. This climate scenario simultaneously explains the extent and density of ground ice, with a realistically large soil diffusivity. The constant humidity scenario, on the other hand, is inconsistent with the GRS measurements (Fig. 3). Two-layer GRS results for the burial depth and ice content^{19,21} show that not only does the burial depth change with latitude, but so does the fraction of ice. Although the overall ice fraction is high, there is a latitude range where the ice content is small enough to be compatible with pore ice.

Two very different histories of atmospheric humidity have been considered: constant humidity (Fig. 1a) and strongly varying humidity from an exposed cap (Fig. 1b and c, and Fig. 2). These two histories bracket many other possibilities. The results depend on the history of atmospheric humidity (Fig. 3). Additional model calculations with different values for the diffusion coefficient and dust content give similar results, in part because the diffusive filling always closely follows the equilibrium depth.

The ice age scenario described here provides a comprehensive picture of how global scale glaciations may have evolved on Mars. It predicts that the top surface layer in the mid-latitudes has been

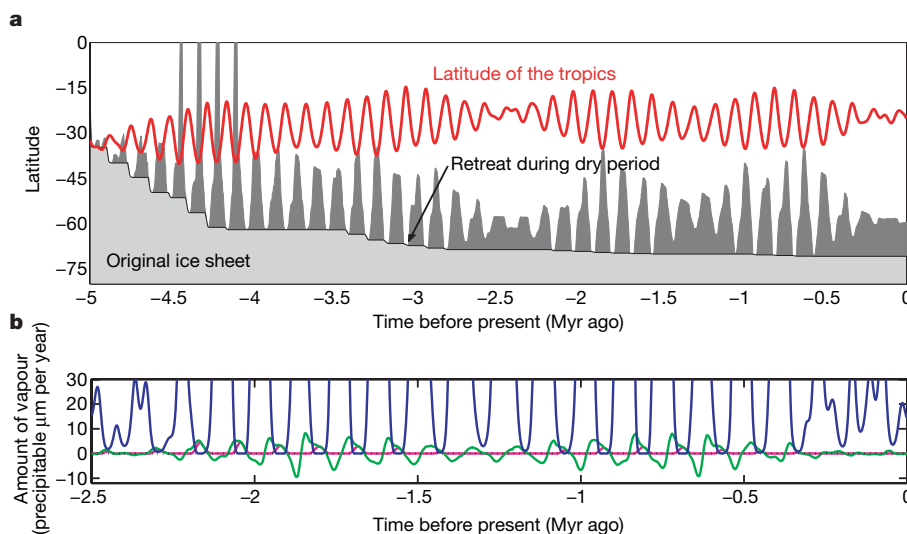


Figure 2 | Evolution of southern hemisphere ice over the past five million years for strongly varying atmospheric humidity. **a**, Latitudinal extent of ice in the topmost metre of soil. Shading as in Fig. 1. The red graph shows the latitude of the tropics, which equals (minus) the obliquity. Although only the obliquity is shown, the orbital eccentricity and longitude of perihelion

also vary. **b**, The blue trace indicates the atmospheric humidity (originating from the north polar cap), the green trace indicates the output or intake of water by the subsurface, and the magenta trace indicates the contribution to the output from the retreat of the original ice sheet. Column-integrated amounts are given in units of thickness of a global ice layer.

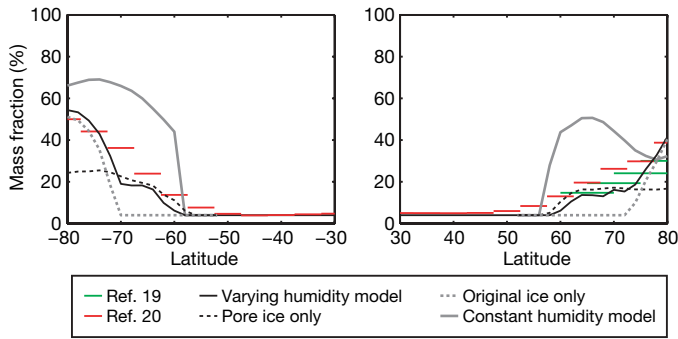


Figure 3 | Comparison with measurements by GRS onboard Mars Odyssey. Coloured bars show the ice (mass) fraction according to refs 19 and 20, in latitude bins, zonally averaged and assuming a homogeneous vertical distribution of hydrogen. The solid black and grey lines show the fraction of ice in the topmost metre of the soil according to the model described in the text and shown in Fig. 1a and c. The dry component is assumed to contain 4% water-equivalent hydrogen. The dashed and dotted lines pertain to the varying humidity scenario.

reworked numerous times during the past few million years^{22,23}, that much of the ice seen by GRS is diffusively formed pore-ice and less than half a million years old, but that an older ice sheet exists at higher latitudes. The pore-ice layers extend and retreat close to synchronously on both hemispheres. Over the last 2.5 million years the original ice sheets receded vertically by an estimated 60 cm (creating less than that in additional sublimation lag), considering only areas where it remains within the topmost metre today. Most of the ice loss, however, occurs at the retreating edge of the ice that lies below one metre, and the total ice volume change may be as high as 10^5 km³, an amount that is still small compared to the volume of the permanent polar deposits. These movements are probably the counterpart of the layers observed in polar regions²⁴.

The dynamic nature of the ice sheets makes Mars an ideal system in which to test and expand our knowledge of astronomical climate forcing. A great deal could be learned about terrestrial ice ages from the study of martian ice stratigraphy—a longer, cleaner and simpler record than Earth's.

METHODS SUMMARY

The climate model is composed of models of surface and subsurface temperature, atmospheric humidity, and the retreat and growth of ground ice. The thermal model¹¹ solves the one-dimensional heat conduction equation with a semi-implicit numerical method and resolves annual and diurnal temperature cycles. The atmospheric humidity is controlled by the emission of water vapour from the north polar cap and is modelled according to Toon *et al.*¹. The total atmospheric pressure is assumed to be constant.

Vapour transport calculations are intrinsically nonlinear, because of phase transitions, and hence require explicit numerical solvers for the diffusion equation, which are subject to stringent numerical stability requirements. Hence, explicit vapour diffusion models are orders of magnitude slower than a thermal model, a serious practical impediment for computations of ice evolution over long periods of Mars history. This limitation is overcome by employing net transport equations, as outlined in the following.

Previous work^{11,16,25} has established a method to compute the retreat of ground ice efficiently. The vapour flux from the ice to the atmosphere is given by the difference in mean annual vapour density between the ice table and the surface, as in equation (1) of the Methods. A similar averaging method can be developed for the growth of pore ice. Detailed microphysical calculations have shown¹¹ that condensation of atmospherically derived ice in soil pores leads to accumulation of ground ice below a sharply defined interface. Below this interface, the diffusive flux is governed by the saturation vapour pressure, as shown in

equation (4) in the Methods. The interface is located at a depth where the two inward fluxes (from the surface to the interface and from the interface downward) are equal, see equations (5) or (7). Instantaneous vapour density profiles are not required, and therein lies the computational efficiency.

Full Methods and any associated references are available in the online version of the paper at www.nature.com/nature.

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METHODS

The climate model focuses on atmosphere–subsurface exchange of H₂O and it incorporates a new accelerated method for pore-ice calculations.

Thermal model. The thermal model is described in detail elsewhere¹¹. It is run with a time step of 30 min in a layer 15 m thick. Temperatures are allowed to equilibrate over 14 Mars years and annual means are computed from the 15th Mars year. The model includes changes in thermal conductivity due to the presence of ground ice.

Fast numerical method for the retreat and growth of ground ice. Thermal models can take advantage of implicit or semi-implicit numerical methods that allow large time steps. Explicit vapour diffusion calculations are significantly slower than a thermal model, so here we exploit time-averaged equations for the retreat and growth of pore ice. The ice is assumed to recede vertically; internal deformations of the ice sheet are negligible, because its lateral extent is much larger than its depth.

Previous work^{11,16,25} has established a method of computing the retreat of ground ice efficiently. The vapour flux from the ice to the atmosphere is given by:

$$J = -D(\langle\rho_{sv}(z_0)\rangle - \langle\rho(0)\rangle)/z_0 \quad (1)$$

where D is the diffusion coefficient of dry soil, ρ_{sv} the saturation vapour density, z_0 the depth of the ice table, and $\rho(0)$ the vapour density on the surface. Angle brackets indicate annual means. The averaging procedure not only provides significant computational advantages, but also eliminates dependences on several microphysical parameters, such as adsorption^{11,25}. The rate of retreat of ground ice is given by $r = -J/\rho_{ice}$, where ρ_{ice} is the density of bulk ice, 927 kg m⁻³. When a layer of dirty ice retreats, the speed at which the dry layer grows is:

$$\frac{dz_0}{dt} = \frac{1 - \Phi_2}{1 - \Phi_1} r \quad (2)$$

where Φ_1 is the porosity of the dry layer and Φ_2 the ice content of the ice layer. The geometric factor arises because dust in the ice is converted to porous sublimation till. Over a time step of $\Delta t = 250$ years, the dry layer grows from an initial thickness z_0 to a new thickness:

$$\sqrt{z_0^2 + 2 \frac{1 - \Phi_2}{1 - \Phi_1} r z_0 \Delta t} \quad (3)$$

The retreat of pore ice can be handled similarly to the retreat of an ice layer.

A similar averaging method can be developed for the growth of pore ice. Detailed microphysical calculations have shown¹¹ that condensation of atmospherically derived ice in soil pores leads to accumulation of ground ice below a sharply defined interface. The depth z_0 of this interface can be determined as

follows. Above the interface, the flux is given by equation (1). Below the interface, the diffusive flux is governed by the saturation vapour pressure:

$$J = -D(1-f)\langle\partial\rho_{sv}/\partial z\rangle \quad (4)$$

where D is the diffusion coefficient with all pores open, $f(z)$ is the volume fraction of pore space filled with ice, and z is the depth below the surface. Because f changes little within one Mars year, it can be taken outside the average. $(1-f)$ corresponds to constriction from partially ice-filled pores. The interface is located where the two inward fluxes (1) and (4) are equal:

$$(\langle\rho_{sv}(z_0)\rangle - \langle\rho(0)\rangle)/z_0 = (1-f)\langle\partial\rho_{sv}/\partial z\rangle|_{z_0} \quad (5)$$

With a Clausius–Clapeyron expression for the saturation vapour density, one obtains:

$$\frac{\partial\rho_{sv}}{\partial z} = \frac{\rho_{sv}}{T} \left(\frac{H}{RT} - 1 \right) \frac{\partial T}{\partial z} \quad (6)$$

where T is the temperature, H is the enthalpy of sublimation, and R is the universal gas constant. With this expression, equation (5) becomes:

$$\frac{\langle\rho_{sv}(z_0)\rangle - \langle\rho(0)\rangle}{z_0} = (1-f) \left\langle \frac{\rho_{sv}}{T} \left(\frac{H}{RT} - 1 \right) \frac{\partial T}{\partial z} \right\rangle|_{z_0} \quad (7)$$

which is solved numerically for z_0 . The instantaneous vapour density profile $\rho(z)$ is not needed, because ρ_{sv} is determined by temperature and $\rho(0)$ is prescribed by the surface humidity. When pore spaces are completely filled with ice ($f=1$), equation (7) reduces to the equilibrium condition $\langle\rho_{sv}(z_0)\rangle = \langle\rho(0)\rangle$ and the flux vanishes. When the ice is stable but the pores are free of ice ($f=0$), the interface is located at a depth below the equilibrium depth and the flux is inward. With time, the interface moves up to the equilibrium position, if the ice is stable; otherwise it retreats.

Atmospheric humidity. The atmospheric humidity is controlled by the emission of water vapour from the north polar cap and is modelled according to ref. 1. The model reproduces the present-day atmospheric vapour partial pressure on the surface of 0.13 Pa. For the past, the calculations assume an exposed cap of the same albedo and area as today's. On present-day Mars, vapour abundance generally decreases from the poles towards the equator, an effect which is neglected, because the frost point temperature depends only logarithmically on humidity; a model calculation with a gradient in partial pressure produced almost identical results. The total atmospheric pressure is assumed to be constant, which is supported by the notion that most CO₂ is stored in the atmosphere rather than in the polar caps; variations in CO₂ pressure would have a minor influence via changes in soil vapour diffusivity, atmospheric absorption and the size of the seasonal CO₂ cap.