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Regolith-atmosphere exchange of water in Mars' recent past

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Abstract

We investigate the exchange of water vapour between the regolith and atmosphere of Mars, and how it varies with different orbital parameters, atmospheric dust contents and surface water ice reservoirs. This is achieved through the coupling of a global circulation model (GCM) and a regolith diffusion model. GCM simulations are performed for hundreds of Mars years, with additional one-dimensional simulations performed for 50 kyr. At obliquities $\varepsilon = 15^\circ$ and 30° , the thermal inertia and albedo of the regolith have more control on the subsurface water distribution than changes to the eccentricity or solar longitude of perihelion. At $\varepsilon = 45^\circ$, atmospheric water vapour abundances become much larger, allowing stable subsurface ice to form in the tropics and mid-latitudes. The circulation of the atmosphere is important in producing the subsurface water distribution, with increased water content in various locations due to vapour transport by topographically-steered flows and stationary waves. As these circulation patterns are due to topographic features, it is likely the same regions will also experience locally large amounts of subsurface water at different epochs. The dustiness of the atmosphere plays an important role in the distribution of subsurface water, with a dusty atmosphere resulting in a wetter water cycle and increased stability of subsurface ice deposits.

Keywords: Mars, Mars, atmosphere, Mars, climate, Mars, surface

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1. Introduction

Studies of possible regolith-atmosphere exchange of water on Mars date back many years (e.g. Smoluchowski, 1968; Flasar and Goody, 1976; Fanale and Jakosky, 1982; Fanale et al., 1986; Clifford, 1993; Mellon and Jakosky, 1993; Mellon et al., 1997), but it was not until observations by the Gamma Ray Spectrometer (GRS) suite of instruments aboard Mars Odyssey that the presence of large subsurface reservoirs of water was revealed (Boynton et al., 2002; Feldman et al., 2004, 2007, 2008; Maurice et al., 2011). Since this time, additional observations of water from the surface have been made by the Phoenix lander (Mellon et al., 2009; Cull et al., 2010). The trenches dug by the Phoenix lander exposed ice at a mean depth of 4.6 cm, with the ice found to be mostly pore ice, though thin and relatively pure ice deposits were also observed near the surface (Mellon et al., 2009). Observations using the ChemCam instrument on the Curiosity rover have revealed a hydrogen emission peak in some soils, which may be the result of adsorbed water or hydration of the amorphous component of the soil (Meslin et al., 2013). New impact craters have revealed evidence for relatively pure subsurface ice in the mid-latitudes (Byrne et al., 2009).

In order to understand the regolith-atmosphere exchange of water, and determine the origin of the observed water distribution, numerous laboratory experiments and modelling studies have been undertaken. Laboratory studies have focused on calculating the diffusion coefficient of water vapour in the regolith, and the adsorption of water onto regolith grains (Zent and Quinn, 1997; Chevrier et al., 2007; Hudson et al., 2007; Bryson et al., 2008; Chevrier et al., 2008; Hudson and Aharonson, 2008; Sizemore and Mellon, 2008; Beck et al., 2010; Siegler et al., 2012). As well as revealing information on likely rates of diffusion and ice stability in different materials at different conditions, these studies also provide useful constraints for models of regolith-atmosphere interaction.

Previous modelling studies have focused on understanding the subsurface water distribution and the stability of subsurface ice in both the past and present epochs. These studies involve either (i) the explicit calculation of water vapour diffusion between the regolith and atmosphere (Mellon and Jakosky, 1993; Tokano, 2003; Böttger et al., 2005; Schorghofer and Aharonson, 2005; Schorghofer, 2007; Schorghofer and Forget, 2012; Williams et al., 2015), or (ii) the determination of the equilibrium ice table depth from subsurface temperatures and near-surface water vapour values (Mellon and

38 Jakosky, 1995; Mellon et al., 1997, 2004; Aharonson and Schorghofer, 2006;
39 Chamberlain and Boynton, 2007; Zent, 2008).

40 For present-day conditions, studies find that subsurface ice is mostly sta-
41 ble polewards of $\pm 50^\circ$, with the regolith thermal inertia and albedo playing
42 a role in the stability of subsurface ice, in agreement with the GRS ob-
43 servations (Mellon and Jakosky, 1993; Mellon et al., 1997, 2004; Tokano,
44 2003; Böttger et al., 2005; Schorghofer and Aharonson, 2005; Aharonson
45 and Schorghofer, 2006; Chamberlain and Boynton, 2007). The mid-latitudes
46 are largely ice free, though regions with high surface roughness may con-
47 tain detectable quantities of ice (Aharonson and Schorghofer, 2006), while
48 shallow subsurface ice is required to account for the distribution of CO_2 ice
49 on pole-facing slopes in the southern mid-latitudes (Vincendon et al., 2010).
50 The formation of ice lenses has been investigated in order to understand the
51 Phoenix observations of shallow, relatively pure water ice that cannot be
52 explained by vapour diffusion alone (Sizemore et al., 2015), and it has been
53 proposed that some subsurface ice in the present day may be the remnants
54 of buried surface ice deposits from a past climate (e.g. Jakosky and Carr,
55 1985; Mischna et al., 2003; Levrard et al., 2004).

56 For past epochs, studies have focused on determining the equilibrium
57 ice table depth, and how it varies with orbital parameters (Mellon and
58 Jakosky, 1995; Chamberlain and Boynton, 2007; Zent, 2008), and the time
59 integration of the subsurface ice content in a model initialized with an ice
60 sheet (Schorghofer, 2007; Schorghofer and Forget, 2012). Unlike the present
61 study, these simulations do not consider the feedback between the sub-
62 surface water content and the global circulation, and are instead run as
63 one-dimensional models for specific locations on the surface (Mellon and
64 Jakosky, 1995; Chamberlain and Boynton, 2007; Zent, 2008), or consider
65 zonal averages (Schorghofer, 2007; Schorghofer and Forget, 2012). Addition-
66 ally, the simulations of Mellon and Jakosky (1995), Schorghofer (2007) and
67 Schorghofer and Forget (2012) were run for thousands or millions of years,
68 and as they are interested in the long-term behaviour of subsurface water,
69 large time steps are used. Mellon and Jakosky (1995) use time steps of 5
70 minutes for the thermal model and 10 sols for the diffusion model, while
71 Schorghofer (2007) and Schorghofer and Forget (2012) use time steps of 30
72 minutes for the thermal model and 100–250 years for changes in subsurface
73 ice content. Because of the one-dimensional nature of the models, various as-
74 sumptions are made, such as how the atmospheric water content and pressure
75 vary in past epochs, which will affect the diffusion calculation.

In this paper we investigate how the regolith-atmosphere interaction of water vapour via diffusion varies with different orbital parameters, atmospheric dust contents and surface ice reservoirs. We use a global circulation model (GCM) coupled with a regolith diffusion model, and run simulations for hundreds of Mars years. This removes the need to make assumptions about the near-surface atmospheric water content, and allows us to study the spatial distribution of subsurface water over the whole globe, and how it is affected by the atmospheric circulation. One-dimensional studies using the GCM output are also performed over longer time periods (50 kyr) for various locations at different obliquities.

2. Model description

2.1. Global circulation model

The GCM used for this study results from collaboration between the Laboratoire de Météorologie Dynamique (LMD), the University of Oxford and The Open University. The model combines the most recent LMD physical schemes with a spectral dynamical core, an energy and angular-momentum conserving vertical finite-difference scheme and a semi-Lagrangian advection scheme for tracers (for further details see Forget et al., 1999; Lewis et al., 2007). For the surface properties, we use data from the Thermal Emission Spectrometer (TES). Albedos are from Christensen et al. (2001), while thermal inertias are from Mellon et al. (2000), with corrections made to account for the effect of clouds in the initial dataset (Wilson et al., 2007). Small-scale topographic parameters used by the gravity wave drag scheme, along with the resolved topography, are obtained from MOLA data (Zuber et al., 1992; Smith et al., 1998). These data are stored as $1^\circ \times 1^\circ$ global maps, which are smoothed to the required model resolution.

Due to the spectral nature of the model, different resolutions are referenced using the label Tx , where ‘T’ represents triangular spectral truncation, and x represents the horizontal wavenumber the model is truncated at. The results presented in this paper are obtained using two different model resolutions. To enable study of the global regolith-atmosphere interaction over many hundreds of Mars years, the majority of simulations are run at T5 spectral truncation, which corresponds to a grid resolution of 22.5° in latitude and longitude for physical processes. Test simulations using present-day conditions were compared to observations and previous modelling studies (e.g. Smith, 2004; Navarro et al., 2014; Steele et al., 2014b) which showed that

the T5 resolution could reproduce the present-day water cycle (see section 4). Additional simulations are performed at T31 spectral truncation (5° resolution) to gain a better understanding of the spatial distribution and stability of surface and subsurface ice. In the vertical there are 20 levels in sigma coordinates, extending to an altitude of ~ 85 km.

Dust is not transported in the model because of the difficulty in accounting for surface lifting at low resolutions. Instead, we set the global visible dust optical depth, τ_{vis} , to either 0.3 or 3, in order to account for both clear and dusty conditions (these values are comparable to the optical depths observed in the present day during clear periods and global dust storms respectively). The vertical profile of dust follows a modified Conrath distribution (Lewis et al., 1999), with the altitude of the dust top being dependent on the dustiness of the atmosphere. For the clear simulations ($\tau_{\text{vis}} = 0.3$) the dust top is taken to be $z_{\text{top}} = 30$ km, while for the dusty simulations ($\tau_{\text{vis}} = 3$) we set $z_{\text{top}} = 60$ km. These values are in agreement with present-day observations (e.g. Määttänen et al., 2013; Smith et al., 2013).

Water vapour and ice mass mixing ratios are transported as tracers, using the microphysics scheme of Montmessin et al. (2004) to account for the formation and sedimentation of ice particles. Clouds in the model are not radiatively active. While it has been shown that clouds in the present-day climate can influence the temperature structure of the atmosphere and strengthen the overturning circulation (e.g. Wilson et al., 2008; Madeleine et al., 2012; Steele et al., 2014a; Navarro et al., 2014), their inclusion in models has also led to some inconsistencies compared with spacecraft observations. Additionally, as we are not transporting dust we cannot account for the complex coupling between the dust and water cycles (e.g. Kahre et al., 2015). Ice can sediment anywhere on the surface, and if more than $5 \text{ pr-}\mu\text{m}$ of water ice is deposited onto the surface, the albedo is changed from that of the regolith (derived from TES data) to that of water ice (0.4). Surface water ice can only sublime if there is no covering layer of CO_2 ice.

2.2. Regolith model

The regolith model is an updated version of that used by Böttger et al. (2004, 2005), which is based on the one-dimensional model of Zent et al. (1993). Diffusion is calculated implicitly on 30 unevenly-spaced levels extending to around 20 m below the surface. (The first four levels are within 1 mm of the surface, with the first level at 0.1 mm.) While the GCMs physical

time step is 30 minutes (used for calculating radiative transfer, tracer transport, turbulence, convection etc.), the regolith model operates on a shorter time step of 1 minute. This reduced time step is needed to avoid numerical instabilities which can occur when large amounts of vapour diffuse quickly at high obliquity, and is the same value used by Zent et al. (1993).

In the regolith model, the concentration of water in a volume of regolith, σ , is decomposed into three states: vapour contained within the pore spaces (ϕn), vapour adsorbed onto regolith grains (α) and pore ice (ζ), all measured in kg m^{-3} . Thus,

$$\sigma = \phi n + \alpha + \zeta, \quad (1)$$

where ϕ is the porosity of the regolith and n is the vapour density. The model uses the adsorption isotherm of Fanale and Cannon (1971), which has been used extensively in previous studies of regolith diffusion (e.g. Zent et al., 1993; Mellon and Jakosky, 1993, 1995; Mellon et al., 1997; Böttger et al., 2005). Unlike the scheme of Böttger et al. (2004, 2005) which used the σ value in the diffusion calculation, we use the $\sigma - \zeta$ value. This removes the assumption from the original scheme that all the ice in a grid box turns to vapour and adsorbed water before diffusion is calculated. (See Appendix A for a description of the diffusion model.)

Böttger et al. (2004, 2005) used a constant diffusion coefficient specified in advance, but here we allow the diffusion coefficient to vary in time and space, and consider both Fickian and Knudsen diffusion. In this case, the effective diffusion coefficient, D , is given by the Bosanquet relation, $D^{-1} = D_{\text{F}}^{-1} + D_{\text{K}}^{-1}$ (where D_{F} and D_{K} are the Fickian and Knudsen diffusion coefficients respectively). For Fickian diffusion in a porous medium, the diffusion coefficient is that given by Mason and Malinauskas (1983):

$$D_{\text{F}} = \frac{\phi}{\tau} D_{12}, \quad (2)$$

where D_{12} is the mutual diffusion coefficient, τ is the tortuosity (which characterizes the convoluted nature of the porous pathways) and the ratio ϕ/τ is the ‘obstruction factor’. The porosity is parameterized as in Mellon and Jakosky (1993), giving

$$\phi = \phi_0(1 - \zeta/\rho_{\text{ice}}\phi_0), \quad (3)$$

where ϕ_0 is the ice-free porosity. We take $\phi_0 = 0.4$, which falls within the range determined at the Viking Lander 1 site (Moore and Jakosky, 1989;

179 Boynton et al., 2002). The tortuosity is parameterized as in Hudson (2008),
 180 giving

$$\tau \approx \tau_0(1 - \zeta/\rho_{\text{ice}}\phi_0)^{-1}, \quad (4)$$

181 where τ_0 is the ice-free tortuosity, parameterized as $\tau_0 = \phi_0^{-1/3}$ (Millington,
 182 1959). This parameterization produces results in agreement with both
 183 simulations (Zalc et al., 2004) and measurements using Mars-analogue soils
 184 (Sizemore and Mellon, 2008). Combining Equations 3 and 4, the obstruction
 185 factor becomes

$$\frac{\phi}{\tau} = \phi_0^{4/3} \left(1 - \frac{\zeta}{\rho_{\text{ice}}\phi_0}\right)^2. \quad (5)$$

186 The coefficient of mutual diffusion for the case of water vapour diffusing
 187 into CO_2 is that of Wallace and Sagan (1979):

$$D_{12} = D_{\text{H}_2\text{O},\text{CO}_2} = 0.1654 \text{ cm}^2 \text{ s}^{-1} \frac{p_{\text{ref}}}{p(z)} \left[\frac{T(z)}{T_{\text{ref}}} \right]^{3/2}, \quad (6)$$

188 where $T(z)$ and $p(z)$ are the temperatures and pressures at a distance z
 189 below the surface, and $T_{\text{ref}} = 273.15 \text{ K}$ and $p_{\text{ref}} = 1013 \text{ hPa}$ are reference
 190 temperatures and pressures. Since the regolith model only extends down to
 191 $\sim 20 \text{ m}$ the pressure change is negligible, so we take $p(z) = p_{\text{surf}}$. Finally,
 192 combining Equations 2, 5 and 6 gives

$$D_{\text{F}} = 0.1654 \text{ cm}^2 \text{ s}^{-1} \phi_0^{4/3} \frac{p_{\text{ref}}}{p_{\text{surf}}} \left(1 - \frac{\zeta}{\rho_{\text{ice}}\phi_0}\right)^2 \left[\frac{T(z)}{T_{\text{ref}}} \right]^{3/2}. \quad (7)$$

193 The Knudsen diffusion coefficient is proportional to the mean velocity of
 194 the diffusing molecules. We use the parameterization given by Evans et al.
 195 (1961):

$$D_{\text{K}} = \frac{\pi}{8 + \pi} \frac{\phi}{1 - \phi} \frac{\bar{\nu} \bar{r}}{\tau}, \quad (8)$$

196 where $\bar{r}$ is the average pore size in the regolith and $\bar{\nu}$ is the mean velocity of
 197 the diffusing molecules, given by $\bar{\nu} = (8k_{\text{B}}T/\pi m_{\text{w}})^{1/2}$, with k_{B} the Boltzmann
 198 constant and m_{w} the molecular mass of water.

199 The value of $\bar{r}$ will vary with location. Soil properties have been studied at
 200 the locations of the Viking landers, Pathfinder, Spirit, Opportunity, Phoenix
 201 and Curiosity. A recent overview of soil properties is given by Demidov
 202 et al. (2015). The spatial variation of soil properties is difficult to account
 203 for in a GCM. Flasar and Goody (1976) used a value of $\bar{r} = 1 \mu\text{m}$ for their

204 calculations of Knudsen diffusion, while Fanale et al. (1986), Mellon and
 205 Jakosky (1993) and Mellon et al. (1997) use a value of $10\ \mu\text{m}$. In studying
 206 the pore sizes in JSC Mars-1 simulant, Sizemore and Mellon (2008) found
 207 that grains in the $38\text{--}63\ \mu\text{m}$ and $250\text{--}500\ \mu\text{m}$ size ranges had peaks in their
 208 pore-size distribution at around $10\ \mu\text{m}$ and $100\ \mu\text{m}$ respectively. Using a value
 209 of between $10\text{--}100\ \mu\text{m}$ in our scheme results in a diffusion coefficient during
 210 the daytime of $4\text{--}7\ \text{cm}^2\ \text{s}^{-1}$, which is in agreement with previous studies of the
 211 diffusion coefficient (e.g. Chevrier et al., 2007; Sizemore and Mellon, 2008;
 212 Hudson et al., 2007; Hudson and Aharonson, 2008; Hudson et al., 2009). A
 213 smaller value, e.g. $1\ \mu\text{m}$, would result in a diffusion coefficient $\sim 0.7\ \text{cm}^2\ \text{s}^{-1}$,
 214 which is comparable to the value found for heavily salt-encrusted soils, but
 215 not the general regolith (Hudson and Aharonson, 2008). Since the majority
 216 of landers determined grain sizes $\lesssim 100\ \mu\text{m}$, we use a pore size of $\bar{r} = 10\ \mu\text{m}$,
 217 in line with the results of Sizemore and Mellon (2008).

218 From Equations 7 and 8 it can be seen that the diffusion coefficient gets
 219 larger as temperature increases and pressure decreases. For example, in the
 220 Hellas basin at night we have $D \approx 1.7\ \text{cm}^2\ \text{s}^{-1}$, while on the Tharsis Plateau
 221 and on top of Olympus Mons during the daytime the diffusion coefficients are
 222 around $5\ \text{cm}^2\ \text{s}^{-1}$ and $6\ \text{cm}^2\ \text{s}^{-1}$ respectively. (The scheme of Böttger et al.
 223 (2005) used a constant diffusion coefficient of $1\ \text{cm}^2\ \text{s}^{-1}$, so diffusion operates
 224 more quickly in our simulations.) Thus, the flux of water into or out of the
 225 regolith will be affected by the altitude of the surface and the temperature,
 226 along with the amount of water in the near-surface layer. This flux is de-
 227 termined via a balance of the fluxes at the regolith–atmosphere boundary.
 228 For the regolith, the flux is $F_{\text{reg}} = \kappa_{\text{reg}}(n_1 - n_b)$, while for the atmosphere
 229 the flux is $F_{\text{atm}} = \rho\kappa_{\text{atm}}(q_1 - q_b)$, and we require $F_{\text{atm}} = -F_{\text{reg}}$. (Here the
 230 subscript ‘1’ represents the first regolith or atmosphere layer, the subscript
 231 ‘b’ represents the regolith–atmosphere boundary, q is the water vapour mass
 232 mixing ratio in the atmosphere, and ρ is the atmospheric density at the sur-
 233 face.) The coefficients in the flux terms for the regolith and atmosphere are
 234 given by $\kappa_{\text{reg}} = D/z_{0.5}$ and $\kappa_{\text{atm}} = C_d|\mathbf{u}|$, with $z_{0.5}$ the depth to the first
 235 regolith layer midpoint, C_d the drag coefficient and $|\mathbf{u}|$ the magnitude of the
 236 near-surface wind. The presence of surface CO_2 ice or water ice shuts off
 237 the regolith-atmosphere interaction, though vertical redistribution of water
 238 already in the regolith can still occur through diffusion and phase changes
 239 (lateral transport is not modelled).

240 The presence of subsurface ice affects the regolith thermal inertia, which
 241 in turn affects the regolith temperature. While models of thermal conductiv-

242 ity have suggested that even small amounts of pore ice can rapidly increase
 243 the thermal conductivity of the regolith (e.g. Mellon et al., 1997; Piqueux
 244 and Christensen, 2009), recent laboratory experiments have shown that the
 245 thermal conductivity of an ice-regolith mix increases approximately linearly
 246 with ice content (Siegler et al., 2012). As such, when ice is present in the
 247 subsurface we model the thermal conductivity, k , following Equation 36 of
 248 Siegler et al. (2012):

$$k(f, T) = 0.8\phi_0 f k_{\text{ice}}(T) + k_{\text{dry}}, \quad (9)$$

249 where k_{dry} is the thermal conductivity of the ice-free regolith (horizontally
 250 varying but constant with depth) which is derived from TES thermal in-
 251ertia data (Mellon et al., 2000), $f = \zeta/\rho_{\text{ice}}\phi_0$ is the filling fraction, and
 252 $k_{\text{ice}}(T) = 488.19/T + 0.4685$ is the thermal conductivity of ice at tempera-
 253 ture T (Hobbs, 1974). As an example, taking a dry regolith conductivity of
 254 $k_{\text{dry}} = 0.04 \text{ W m}^{-1} \text{ K}^{-1}$ (corresponding to a thermal inertia of around 250 tiu)
 255 and taking all the pore space to be filled with ice, the thermal conductivity for
 256 the ice-rich layer varies from $0.77 \text{ W m}^{-1} \text{ K}^{-1}$ (at 270 K) to $1.23 \text{ W m}^{-1} \text{ K}^{-1}$
 257 (at 150 K). This is lower than the values of $2.5\text{--}2.8 \text{ W m}^{-1} \text{ K}^{-1}$ used in previ-
 258 ous studies (Mellon et al., 1997, 2004; Chamberlain and Boynton, 2007; Zent,
 259 2008; Williams et al., 2015), but comparable to values used by Schorghofer
 260 and Aharonson (2005) and Schorghofer (2007).

261 3. Simulations performed

262 Calculations show that the obliquity of Mars over the last 5 million years
 263 has varied between around $15^\circ\text{--}35^\circ$, while beyond this time (up to 20 million
 264 years ago) the obliquity varied between around $25^\circ\text{--}45^\circ$ (Laskar et al., 2004,
 265 and see also Figure 1). Each obliquity cycle lasts roughly 10^5 yr and, com-
 266 bined with changes to eccentricity and precession (varying on time-scales of
 267 around 10^5 yr and 50 kyr respectively), a variety of different orbital configu-
 268 rations are possible. GCM simulations cannot be performed over an entire
 269 obliquity cycle due to the computation time required. Thus, in order to
 270 study the effects of various orbital parameters and dust conditions on the
 271 distribution of subsurface water (in particular the locations of stable sub-
 272 surface ice deposits), 24 simulations were performed. These simulations are
 273 essentially ‘snapshots’ at various points throughout Mars’ history. The pa-
 274 rameters varied were the obliquity ($\varepsilon = 15^\circ, 30^\circ$ and 45°), eccentricity ($e = 0$

Figure 1: Variation of obliquity and eccentricity over the past 20 million years. Solar longitude of perihelion varies much more regularly, and is not shown. Data are from Laskar et al. (2004).

and 0.1), solar longitude of perihelion ($L_p = 90^\circ$ and 270°) and visible dust optical depth ($\tau_{\text{vis}} = 0.3$ and 3). For validation purposes, a simulation was also performed for present-day conditions.

Table 1 displays each of the simulations performed and the combination of parameters used, while Figure 2 shows the obliquities and surface ice reservoirs used. Simulations 1–18 have surface ice reservoirs defined at the poles, as they represent Mars as it moves from low obliquity to higher obliquity, and at low obliquity ice forms in the polar regions (e.g. Levrard et al., 2004). Ice caps are placed at both poles for the lowest obliquity, while at other obliquities the ice caps are placed over the pole experiencing summer during aphelion (the ice caps extend from the poles to $\pm 67.5^\circ$ latitude, i.e. all of the first or last rows of the model grid). Simulations 19–24 represent Mars having been at its highest obliquity, and beginning to move towards lower obliquity. We assume that the polar ice caps have sublimed away completely, and a tropical ice source is defined over the Tharsis region, as in previous modelling work (e.g. Jakosky and Carr, 1985; Levrard et al., 2004; Forget et al., 2006; Madeleine et al., 2009). While the models are initialised with pre-defined infinite reservoirs of ice, surface ice can, and indeed does, form elsewhere due to the transport of water and the sedimentation of ice particles.

Figure 2: Overview of the obliquities and surface ice reservoirs (white shading) for the simulations performed. The grid spacing corresponds to T5 resolution. In (a) there is also a south polar ice cap. In (b) and (c) the ice cap can either be at the north or south pole, but not both. In (d) there is an ice reservoir in the Tharsis region, but no permanent polar ice caps are specified. Coloured shading denotes surface topography at 1° resolution.

294 This ice can then sublime if temperatures are high enough. The atmospheric
 295 mass and solar luminosity are kept unchanged, as these simulations model
 296 the relatively recent past.

297 For each simulation, initial regolith temperatures were obtained from ‘spin
 298 up’ runs with no water cycle, performed for 100 Mars years. The regolith
 299 was initialised at each grid point with 2 kg m^{-3} of water, existing primarily
 300 as adsorbed water, which is the same value used by Zent et al. (1993) and
 301 Böttger et al. (2005). This is because if the regolith is completely dry, any
 302 vapour sublimed from the surface ice reservoirs preferentially diffuses into
 303 the nearby surrounding regolith, resulting in an excessively dry water cycle
 304 at lower obliquities. (It is worth noting that because of the increased vapour
 305 abundances at high obliquity, simulations initialised with either a completely
 306 dry regolith, or with 2 kg m^{-3} of water, produce the same results after a few
 307 hundred Mars years.) The value of 2 kg m^{-3} is not large enough to form any
 308 subsurface ice without further diffusion of water from the atmosphere.

309 When discussing the geographic distribution of water, we will generally
 310 refer to three regions: the tropics ($\pm 22.5^\circ$ latitude), the mid-latitudes (22.5° –
 311 67.5° in each hemisphere) and the polar regions (67.5° – 90° in each hemi-
 312 sphere). When discussing the number of years simulations have been run for
 313 we are referring to Mars years, but note that the time periods of 5 Myr and
 314 5–20 Mya correspond to Earth years. For brevity, we will hereafter refer to
 315 ‘water ice’ and ‘water vapour’ as simply ‘ice’ and ‘vapour’.

Name	ε	L_p	e	τ_{vis}	Ice reservoir
run01	15	90	0.1	0.3	Both poles
run02	15	90	0.1	3	Both poles
run03	15	270	0.1	0.3	Both poles
run04	15	270	0.1	3	Both poles
run05	15	-	0	0.3	Both poles
run06	15	-	0	3	Both poles
run07	30	90	0.1	0.3	South pole
run08	30	90	0.1	3	South pole
run09	30	270	0.1	0.3	North pole
run10	30	270	0.1	3	North pole
run11	30	-	0	0.3	North pole
run12	30	-	0	3	North pole
run13	45	90	0.1	0.3	South pole
run14	45	90	0.1	3	South pole
run15	45	270	0.1	0.3	North pole
run16	45	270	0.1	3	North pole
run17	45	-	0	0.3	North pole
run18	45	-	0	3	North pole
run19	45	90	0.1	0.3	Tropics
run20	45	90	0.1	3	Tropics
run21	45	270	0.1	0.3	Tropics
run22	45	270	0.1	3	Tropics
run23	45	-	0	0.3	Tropics
run24	45	-	0	3	Tropics

Table 1: Overview of the parameters used for each simulation: obliquity (ε), solar longitude of perihelion (L_p), eccentricity (e) and visible dust optical depth (τ_{vis}). When the eccentricity is zero, the value of L_p is no longer relevant. Simulations 1–18 represent Mars moving from low to higher obliquity, while simulations 19–24 represent Mars moving from high to lower obliquity.

316 4. Model validation

317 Before performing simulations for past climates, the model was validated
 318 for the present-day atmosphere (with $\varepsilon = 25.19^\circ$, $L_p = 251^\circ$, $e = 0.093$,
 319 and initially 2 kg m^{-3} of water at each gridpoint in the regolith). The dust
 320 distribution used in the model is obtained from the dust maps of Montabone
 321 et al. (2015). Figure 3a shows the vapour column distribution resulting from
 322 the assimilation of TES temperature profiles and vapour columns at T31
 323 resolution (see Steele et al., 2014b). The first half of the period shown ($L_S =$
 324 $180^\circ\text{--}360^\circ$) corresponds to Mars year 24, with the second half corresponding
 325 to Mars year 25. Figure 3(b–d) shows the results of a T5 simulation after
 326 800 years.

327 Comparing the vapour columns, it can be seen that the main features of
 328 the water cycle are captured by the T5 simulation, with vapour peaks in the
 329 polar regions during summertime, dry polar regions and mid-latitudes in win-
 330 ter, and a relatively dry region between $0^\circ\text{--}30^\circ\text{S}$ around southern hemisphere
 331 summer. The time of peak vapour abundance in the southern hemisphere
 332 occurs later in the T5 simulation than in the assimilation. This is due to
 333 increased dust cover in the south polar region of the T5 simulation (caused
 334 by smoothing of the dust maps to T5 resolution). There were no observations
 335 to assimilate for a period around $L_S = 15^\circ$, resulting in the water cycle in
 336 the assimilation being slightly dry around this time. In general the differ-
 337 ences in the vapour columns between the assimilation and T5 simulation are
 338 within the $5 \text{ pr-}\mu\text{m}$ uncertainty of the TES data (Smith, 2004). In terms of
 339 ice clouds (Figure 3c), the model captures the aphelion cloud belt, the north
 340 polar hood (which extends to the poles) and the south polar hood (which
 341 forms an annulus around the south pole). These are in good agreement with
 342 observations (e.g. Smith, 2004; Benson et al., 2010, 2011).

343 The subsurface water content is shown in Figure 3d. Comparing with
 344 MONS observations (e.g. Feldman et al., 2004; Maurice et al., 2011), the
 345 model agrees in terms of the spatial distribution of subsurface water, with a
 346 dry area circling the globe in the southern mid-latitudes, and relatively large
 347 subsurface water contents around the Tharsis, Arabia Terra and south polar
 348 regions. (The north polar region is dry, as we defined this to be the location
 349 of a permanent ice cap, so no diffusion into the regolith occurred.) While the
 350 spatial distribution shows broad agreement, the subsurface water abundances
 351 at high latitudes have not reached the values observed by MONS, as diffusion
 352 is a slow process and the simulation time of 800 years is relatively short.

Figure 3: Validating the water cycle at T5 resolution for present-day conditions. Panels (a–c) show zonal averages as a function of latitude and time, and are (a) the vapour column distribution from a T31 assimilation (Steele et al., 2014b), (b) the vapour column distribution from a T5 simulation, and (c) the ice optical depth from a T5 simulation. Panel (d) shows the subsurface water content (adsorbed water, ice and vapour) between 5 cm–2 m below the surface after 800 years of simulation. In panel (d) the black contours show topography and the grey shading shows the location of the polar ice cap.

353 A one-dimensional simulation was also performed at the location of the
 354 Phoenix lander (68.2°N, 125.7°W) in order to validate the diffusion model
 355 used in the GCM. The required inputs (surface and subsurface temperatures,
 356 surface pressure and near-surface vapour) were taken every 30 minutes from
 357 the GCM simulation of the present-day atmosphere. Near-surface vapour
 358 amounts are negligible for the majority of the year due to cold temperatures.
 359 Interaction between the atmosphere and regolith only occurs between $L_S =$
 360 68°–126° when surface ice deposits sublime. During this time, the average
 361 near-surface vapour mass mixing ratio is $\sim 2 \times 10^{-4}$, with a peak value of
 362 $\sim 5 \times 10^{-4}$. The regolith was initially dry, and the simulation was performed
 363 for 1000 years.

Figure 4: Zonally- and temporally-averaged water vapour columns from year 800 of each of the 24 simulations performed. The obliquities and ice cap locations are labelled in each plot; see Table 1 for full details.

364 The results show subsurface ice to be stable at a depth of ~ 9.5 cm, with
 365 pore ice temporarily forming in a layer a few centimetres deep on top of
 366 the ice layer between $L_S = 75^\circ$ – 90° . The depth of 9.5 cm is comparable to the
 367 deepest depth to ice observed by Phoenix, but deeper than the mean depth of
 368 4.6 cm (Mellon et al., 2009). However, considering the variability in ice table
 369 depths observed over a relatively small location (with one trench excavated
 370 to 18.3 cm showing no ice), and the variations in thermal inertia and slope
 371 angle which we do not take into account here but which can affect ice stabil-
 372 ity (Aharonson and Schorghofer, 2006), our model results are comparable to
 373 the observations. The adsorbed water values above the ice table in our model
 374 are ~ 2 – 8 kg m^{-3} , which corresponds to ~ 0.1 – 0.5% by mass. These are in line
 375 with adsorbed water amounts measured on martian regolith analogs (Pom-
 376 merol et al., 2009), and are consistent with regolith observations by Phoenix,
 377 where it was inferred that adsorbed water was responsible for variabilities in
 378 the nature and cohesive strength of the soils (Arvidson et al., 2009).

379 5. Regolith-atmosphere interaction over the last 5 Myr

380 5.1. 15° obliquity

381 Considering first the $\varepsilon = 15^\circ$ simulations (with ice caps in both polar
 382 regions), there is little difference in the water cycle when changing the ec-
 383 centricity or solar longitude of perihelion (see Figure 4a). Changing the
 384 dust distribution does have an effect, however, with the atmosphere in the

Figure 5: Atmospheric and subsurface water contents from two simulations at $\varepsilon = 15^\circ$ with different atmospheric dust contents. (a–b) The mean annual vapour column distribution, scaled to a reference pressure of 610 Pa to remove the effects of topography. (c–d) The subsurface water content (adsorbed water, ice and vapour) between 5 cm–2 m below the surface after 800 years of simulation. Black contours show topography, and grey shading in the lower panels shows the locations of the polar ice caps.

385 dusty simulations being around 2–4 pr- μm wetter than the clear simulations.
 386 This can be seen in Figure 5(a,b), which shows the mean annual vapour
 387 column distribution from run03 and run04 (the other simulations show sim-
 388 ilar trends). The dustier simulations are wetter due to warmer atmospheric
 389 temperatures. In the tropics, daytime temperatures in the dusty simulations
 390 are 10–20 K warmer than in the clear simulations at altitudes below 5 km,
 391 and 30–40 K warmer at altitudes above 15 km. In the polar regions dur-
 392 ing summer, the temperatures below 15 km are 5–10 K warmer in the dusty
 393 simulations, and above 25 km they are 30–40 K warmer. These warmer tem-
 394 peratures strengthen the meridional circulation, allowing more vapour to be
 395 transported away from the subliming polar ice caps in the summer.

396 The increased transport from the polar ice caps, combined with cooler
 397 surface temperatures in the dusty simulations, results in the regolith at the
 398 edges of the ice caps having a larger water abundance than in the clear

399 simulations. This can be seen in Figure 5(c,d), which shows the subsurface
400 water content (in the adsorbed and vapour phases, as no ice has formed)
401 between 5 cm–2 m. After 800 years of simulation, the water content in these
402 relatively wet regions of the subsurface is only increasing slowly (by around
403 $0.1\text{--}1\text{ g m}^{-2}\text{ yr}^{-1}$). Considering that the mean annual temperature of the
404 subsurface in these regions is $\sim 180\text{ K}$, it would take many thousands of years
405 for the amount of water adsorbed to be in equilibrium with the atmosphere.
406 In the tropics there are two dry areas in regions with relatively high thermal
407 inertia.

408 5.2. 30° obliquity

409 At $\varepsilon = 30^\circ$, sublimation from the polar ice caps is increased compared to
410 at $\varepsilon = 15^\circ$, hence the water cycle is globally wetter. Peak vapour columns are
411 $\sim 100\text{ pr-}\mu\text{m}$, compared to $\sim 15\text{ pr-}\mu\text{m}$ at $\varepsilon = 15^\circ$, and mean vapour columns
412 are around 1.5–2 times as large as at $\varepsilon = 15^\circ$ (see Figure 4b). In terms
413 of the spatial distribution of subsurface water, Figure 6 shows the global
414 subsurface water content between 5 cm–2 m below the surface. To enable a
415 better spatial resolution, the results of the T5 simulation at year 800 were
416 used to initialise a T31 simulation which ran for an additional 20 years. The
417 stability of subsurface ice reacts very quickly to changing temperatures, so 20
418 years of simulation is long enough to gain meaningful results from the higher
419 resolution simulations. (Stable subsurface ice is defined as ice which does
420 not lose mass over the course of the 20-year simulation.) Figure 6 ignores
421 the dry upper $\sim 5\text{ cm}$ where little water exists due to the warm temperatures
422 limiting the amount of vapour adsorbed on to the regolith grains. The water
423 content is calculated to a depth of 2 m as this is the depth to which vapour
424 had diffused at the end of the simulation.

425 It is clear that there are certain locations where subsurface water pref-
426 erentially accumulates. The polar regions contain the most water, with up
427 to 20 kg m^{-2} existing in both the adsorbed and ice phases after 800 years of
428 simulation. Away from the polar regions there are locally large water abun-
429 dances in the Arabia Terra region (centred at around 20°E near the equator),
430 south-east of Elysium Mons (on the dichotomy boundary between Elysium
431 Planitia and Terra Cimmeria) and around Alba Mons, though no stable sub-
432 surface ice is present. In the northern hemisphere, three relatively dry areas
433 are observed: two surround the relatively wet Arabia Terra region, with a
434 third to the north-east of Elysium Mons. In the southern mid-latitudes the

Figure 6: Total subsurface water content (adsorbed water, ice and vapour) between 5 cm–2 m below the surface after 820 years of simulation (800 years at T5 resolution and 20 years at T31 resolution). Black contours show topography, while blue contours highlight areas of increased subsurface water (at each pole and in two regions in the tropics) observed by MONS (Feldman et al., 2004). Polar ice cap locations are shown with grey shading.

435 water distribution is dependent upon the location of the ice cap. For a south-
 436 polar ice cap as in Figure 6(a,b) the southern mid-latitudes do not have any
 437 particularly dry areas, while for a north-polar ice cap as in Figure 6(c,d) this
 438 region is relatively dry. The subsurface water distribution away from the
 439 polar regions is little affected by changes to the solar longitude of perihelion,
 440 the eccentricity or the dust content of the atmosphere, even though these
 441 lead to differences in the distribution and peak values of vapour. Instead,
 442 the distribution is largely determined by the regolith properties, with regions
 443 of relatively little water corresponding to regions of high thermal inertia and
 444 low albedo.

445 Comparing with MONS observations (Feldman et al., 2004; Maurice et al.,
 446 2011), the simulated and observed subsurface water distributions show good
 447 spatial agreement, with the simulations with a north-polar ice cap (Fig-
 448 ure 6c,d) displaying the best agreement. The similarity is because while
 449 the water cycle at $\varepsilon = 30^\circ$ is wetter than in the present day, the regolith

450 temperatures in the tropics and mid-latitudes are broadly the same, and
 451 there still isn't enough near-surface vapour to form ice from diffusion. Thus,
 452 the simulation results suggest that the subsurface water distribution revealed
 453 by MONS is likely to occur for a range of obliquities close to that of today,
 454 when the atmospheric water content is supplied by a north polar ice cap.

455 The MONS observations of the polar regions (e.g. Feldman et al., 2007;
 456 Maurice et al., 2011) show increased subsurface water abundances and de-
 457 creased burial depths centred roughly at 135°W, 70°N and 90°E, 70°S. From
 458 the GCM results, these areas correspond to regions of increased atmospheric
 459 water content due to transport by topographically-steered flows to the west
 460 of Tharsis during northern hemisphere summer, and stationary waves to the
 461 east of Hellas during southern hemisphere summer (Forget et al., 2006; Steele
 462 et al., 2014b). During autumn, the cooling surface temperatures and locally
 463 large vapour amounts mean the flux of vapour into the regolith is greater
 464 in these regions compared to elsewhere. For example, at 90°E in run12, the
 465 flux of vapour into the regolith is greater at 70°S ($\sim 6 \text{ g m}^{-2} \text{ sol}^{-1}$) than it
 466 is at 60°S or 80°S ($\sim 3 \text{ g m}^{-2} \text{ sol}^{-1}$). Similar values occur in run08 at 135°W
 467 and 60–80°N. As these regions of increased subsurface water are the result
 468 of vapour transport due to topographic features, it is likely they will also ex-
 469 perience increased subsurface water at different obliquities. (Feldman et al.
 470 (2005) have previously linked regions of decreased subsurface water to the
 471 east and west of Arabia in the present day with the location of western
 472 boundary currents.)

473 While the spatial distribution of subsurface water agrees well after 820
 474 years of simulation, the water abundances have not yet reached the values
 475 observed by MONS. For example, the wet region in Arabia Terra is $\sim 0.8\%$
 476 water by mass (compared to $\sim 8\%$ in the observations), while the polar re-
 477 gions are around 2% (compared to values up to $\sim 50\%$ in the observations).
 478 However, the wet regions in the tropics are still slowly gaining water by
 479 around $20\text{--}40 \text{ g m}^{-2} \text{ yr}^{-1}$ (in the form of adsorbed water), and subsurface ice
 480 is increasing at the polar regions (but the small areal extent means it has
 481 little impact on the global mass calculation). However, even if the wet re-
 482 gions in the tropical subsurface were in equilibrium with the atmosphere,
 483 then the subsurface water abundance in the upper 2 m of regolith would not
 484 reach the 10% value observed by MONS (which would require $\sim 170 \text{ kg m}^{-3}$
 485 of subsurface water). It has been proposed that hydrous minerals may be
 486 responsible for some of the observed hydrogen near the equator (e.g. Bish
 487 et al., 2003; Wang et al., 2013; Karunatillake et al., 2014).

Figure 7: (a–c) Variation of the filling fraction ($f = \zeta / \rho_{\text{ice}} \phi_0$) with depth and time for three different longitudes at 78°N . Coloured shading shows the ice evolution over 50 kyr. Grey shading shows the result after initialising the regolith with ice abundances of $f = 0.5$ from the surface down to the layer with the largest ice abundance after 50 kyr. (d) Variation of the near surface water vapour abundance over part of the year at the three locations in panels (a–c).

488 To investigate the build-up of ice over the polar regions further, a series
 489 of one-dimensional simulations were performed with latitudes of $\pm 78^\circ$ and
 490 longitudes of $\pm 45^\circ\text{E}$ and $\pm 135^\circ\text{E}$. The simulations were run for 50 kyr, with
 491 the relevant data (surface and subsurface temperatures, surface pressure and
 492 near-surface vapour) taken every 30 minutes from year 820 of run07–run12.
 493 (After 820 years these values in the global model have reached equilibrium,
 494 and repeat each year.) The results for three locations from run08 are shown in
 495 Figure 7(a–c), with the corresponding near-surface water vapour abundances
 496 shown in panel (d). The results from the other simulations are similar, so
 497 are not shown.

498 Subsurface ice begins to form when vapour abundances reach the sat-
 499 uration value. Due to the initially low vapour abundance in the regolith,
 500 this generally corresponds to the depth where the coldest temperatures are
 501 found, though in reality will be dependent on the nature of the pore spaces
 502 (e.g. Hudson et al., 2009). Over time the ice builds up and the regolith poros-
 503 ity is reduced, which limits the transport of vapour deeper into the regolith
 504 below the ice layer. For example, for a filling fraction of $f = 0.5$, the Fickian
 505 diffusion coefficient is 25% of the ice-free value. The vapour values above
 506 the ice layer then increase, eventually reaching saturation and allowing ice to
 507 form. Thus, the ice table migrates upwards over time. After 50 kyr, the top
 508 of the ice layer has reached depths of $\sim 30\text{--}70\text{ cm}$ (depending on location),
 509 with the ice extending to depths just below 1 m. To see to what depth ice

would be stable at these three locations given more time, we took the results after 50 kyr and filled the regolith with ice from the surface down to the layer with the largest ice abundance. We then ran the simulations until the ice distributions reached equilibrium. (Ice in the upper few centimetres of the regolith is unstable and rapidly sublimates.) The results are shown by the grey shading in Figure 7(a–c). As can be seen, ice is stable at depths of ~ 20 cm, which is similar to the MONS observations (e.g. Maurice et al., 2011).

5.3. *Equilibrium ice table depth*

The previous two sections have looked at the subsurface water distributions resulting from the explicit calculation of vapour diffusion in the regolith. However, it is also possible to determine the depth at which buried ice is in equilibrium with the atmosphere from the regolith temperatures and near-surface vapour values. We take this depth to be the point in the regolith at which the mean annual saturated vapour pressure, $\bar{e}_{\text{sat}}$, is equal to the mean annual atmospheric vapour pressure in the near-surface layer, $\bar{e}_{\text{vap}}$. This same procedure has been used in other studies, such as Mellon and Jakosky (1995); Mellon et al. (2004); Schorghofer and Aharonson (2005); Chamberlain and Boynton (2007). We perform equilibrium depth calculations at T31 resolution using the orbital parameters and dust optical depths appropriate for the simulations at $\varepsilon = 15^\circ$ and $\varepsilon = 30^\circ$ (see Table 1). The $\bar{e}_{\text{vap}}$ values are obtained from year 820 of the relevant simulations (after 800 years at T5 resolution and 20 years at T31 resolution).

The regolith thermal inertias initially correspond to ice-free values, except in the lowest model layer where they are amended to those of an ice-regolith mix using Equation 9 (with $f = 1$, taking all the pore space to be filled with ice). The model was then run for a year, and the regolith temperatures were used to find $\bar{e}_{\text{sat}}$ in the model layer above the ice-regolith mix. If $\bar{e}_{\text{sat}} < \bar{e}_{\text{vap}}$, then the thermal inertia of this layer was also amended to that of an ice-regolith mix. The model was then run for another year with this new thermal inertia distribution, and the procedure was iterated (increasing the thermal inertia of the layer above) until $\bar{e}_{\text{sat}} = \bar{e}_{\text{vap}}$.

The result of one $\varepsilon = 15^\circ$ simulation is shown in Figure 8a. The other simulations at this obliquity follow similar trends in the polar regions, but there are differences in the tropics and mid-latitudes. Figure 8b shows the result for the present-day. At $\varepsilon = 15^\circ$ it can be seen that polewards of $\sim 60^\circ$ in both hemispheres, ice can be stable at shallow depths below the surface. In the region around Olympus Mons and Alba Mons, ice can be stable at

Figure 8: Maps of the ice table stability depth for simulations at (a) $\varepsilon = 15^\circ$, (b) $\varepsilon = 25.19^\circ$ and (c) $\varepsilon = 30^\circ$. The $\varepsilon = 25.19^\circ$ simulation uses present-day orbital parameters and dust maps. The $\varepsilon = 15^\circ$ and 30° simulations have $L_p = 270^\circ$, $\tau_{\text{vis}} = 3$ and $e = 0.1$. The surface ice reservoirs are shown with grey shading. White shading indicates regions where no stability occurred down to a cut-off depth of 2 m. (We use this cut-off depth as below ~ 2 m the regolith temperatures are not affected by the seasonally varying surface temperatures, and remain as initialised.) Black contours show topography.

547 depths of ~ 0.5 – 2 m in the dusty simulations, but this region of stability is
 548 not present in the clear simulations, or the simulations with zero eccentricity
 549 (as vapour values are lower and/or surface temperatures are warmer). In the
 550 present-day simulation, the ice table in the polar regions is slightly deeper
 551 than at $\varepsilon = 15^\circ$ due to the warmer mean annual temperatures. The region of
 552 ice stability around Olympus Mons and Alba Mons is larger than at $\varepsilon = 15^\circ$,
 553 as more vapour is sublimed from the polar ice cap and hence mean annual
 554 near-surface vapour amounts are larger.

555 Similar calculations at low obliquity were performed by Mellon and Jakosky
 556 (1995) and Chamberlain and Boynton (2007), although using one-dimensional
 557 models. At low obliquity, they determined that ice would not be stable any-
 558 where between $\pm 60^\circ$ latitude, while our results show ice can be stable in the
 559 region around Olympus Mons and Alba Mons. The reason for this difference
 560 is likely due to the fact that in our $\varepsilon = 15^\circ$ simulations the vapour columns in
 561 this region can reach ~ 15 pr- μm , while Mellon and Jakosky (1995) assumed a
 562 vapour column of 2.62 pr- μm for $\varepsilon = 19.55^\circ$, and Chamberlain and Boynton
 563 (2007) assumed a value of 0.33 pr- μm for $\varepsilon = 15^\circ$. Thus, our simulations
 564 have larger near-surface vapour values, meaning ice stability can occur in
 565 locations with higher regolith temperatures.

566 The same procedure can be carried out for the $\varepsilon = 30^\circ$ simulations. The

567 result of one simulation is shown in Figure 8c. Again, the other simulations
 568 follow similar trends in the polar regions, with differences around the Tharsis
 569 region. Compared to the previous results, there is a slight reduction in the
 570 areal extent of stable subsurface ice between $\sim 50\text{--}60^\circ\text{N}$, and ice is stable at
 571 depths ranging from 1 cm to 2 m in the Tharsis region. The reduced areal
 572 extent of ice in the northern hemisphere is because the increased obliquity
 573 results in both warmer mean annual surface temperatures at these latitudes,
 574 and reduced mean annual near-surface atmospheric vapour pressures. The
 575 latter is caused by the warmer summertime temperatures increasing vapour
 576 transport away from the polar regions at higher altitudes, reducing the near-
 577 surface vapour values.

578 The stable ice over the Tharsis region is related to cloud formation. As
 579 the water cycle at $\varepsilon = 30^\circ$ is wetter than in the present day, thick clouds form
 580 around the Tharsis Montes. Ice which sediments onto the surface at night
 581 sublimates during the day, increasing the near-surface vapour values compared
 582 to the surrounding areas, and allowing for the stability of subsurface ice.
 583 There is no stable subsurface ice in the mid-latitudes to the north of Alba
 584 Mons, unlike at $\varepsilon = 25.19^\circ$. This is due to the warmer mean annual sur-
 585 face temperature and reduced mean annual near-surface atmospheric vapour
 586 pressure, as discussed earlier. As in the $\varepsilon = 15^\circ$ simulations, the regions of
 587 stable ice in the tropics and mid-latitudes are not present in the simulations
 588 with a clear atmosphere or with zero eccentricity. Thus, when determining to
 589 what depth ice may be stable, particularly away from the polar regions, it is
 590 important to consider the circulation and dustiness of the atmosphere, along
 591 with cloud formation, as these can have a large effect on the near-surface
 592 vapour values.

593 At $\varepsilon = 31.1^\circ$, Mellon and Jakosky (1995) predicted subsurface ice to be
 594 stable to depths of a few centimetres over almost the whole globe, while at
 595 $\varepsilon = 30^\circ$, Chamberlain and Boynton (2007) predicted no stable subsurface
 596 ice between around 15°N – 35°S . Chamberlain and Boynton (2007) noted that
 597 they predicted less extensive subsurface ice deposits due to (i) their smaller
 598 global vapour column value of $100\text{ pr-}\mu\text{m}$, compared to $232\text{ pr-}\mu\text{m}$ in Mellon
 599 and Jakosky (1995), and (ii) their use of a near-surface vapour depletion
 600 scheme. Our results in Figure 8c show less extensive stable subsurface ice
 601 deposits than predicted by Mellon and Jakosky (1995), as we have smaller
 602 vapour columns in our simulations. Compared to Chamberlain and Boynton
 603 (2007), our results show a reduction in the amount of stable subsurface ice
 604 in the northern hemisphere mid-latitudes, which is also likely due to the

605 smaller vapour columns in our simulations. We also predict extensive stable
 606 subsurface ice in the Tharsis region, unlike Chamberlain and Boynton (2007),
 607 which is related to cloud formation as discussed earlier.

608 **6. Regolith-atmosphere interaction between 5–20 Mya**

609 Throughout the time period 5–20 Mya, Mars’ obliquity ranged from
 610 around 25–45° (Laskar et al., 2004, and see also Figure 1). As we have
 611 already studied the regolith-atmosphere interaction at $\varepsilon = 30^\circ$, here we con-
 612 centrate on $\varepsilon = 45^\circ$. Again we vary L_p , e and τ_{vis} , but we also consider two
 613 distinct surface ice reservoirs: polar ice caps and tropical ice reservoirs. This
 614 is because while ice caps are likely to remain at the poles as Mars approaches
 615 $\varepsilon = 45^\circ$, previous modelling work has shown that at this obliquity stable ice
 616 forms at tropical latitudes (e.g. Mischna et al., 2003; Forget et al., 2006).
 617 The polar ice caps are again defined at the pole which experiences summer
 618 during aphelion (or at the north pole for simulations with a circular orbit).
 619 The tropical ice reservoir is placed in the Tharsis region between 100°–125°W
 620 and 22°N–22°S (two grid points at T5 resolution; see Figure 2).

621 *6.1. Polar ice reservoirs*

622 Considering first the simulations with polar ice caps, there is a large
 623 difference in water abundances between the clear and dusty simulations (see
 624 Figure 4c). The increased atmospheric temperatures in the dusty simulations
 625 act to transport more water away from the ice caps, allowing more ice to
 626 sublime. This results in a water cycle that is globally wetter than for a
 627 clear atmosphere, with peak vapour columns of around 1500 pr- μm over the
 628 subliming ice caps and 200 pr- μm in the tropics. Additionally, the lower
 629 daytime surface temperatures in the dusty simulations limit the amount of
 630 vapour diffusing back out of the regolith. Thus, both surface and subsurface
 631 ice forms much more extensively in the dusty simulations.

632 In the clear simulations, subsurface ice forms at the opposite pole to that
 633 where the ice cap was defined (initially at a depth of ~ 1 m), though the values
 634 after 820 years (800 years at T5 resolution and 20 years at T31 resolution)
 635 are only small ($\sim 2 \text{ kg m}^{-2}$). In the dusty simulations the ice is not limited to
 636 the polar regions, and exists at all latitudes, though not all longitudes (see
 637 Figure 9). Ice is nearest to the surface in the Tharsis region, with the top
 638 of the ice layer varying between 2–5 cm, depending on location. In the polar
 639 regions, ice is stable at depths between around 0.1–1 m (here, ice stability is

Figure 9: Depth to stable subsurface ice deposits for the three dusty simulations at $\varepsilon = 45^\circ$. Red shading at the poles denotes the locations of the specified ice caps, while red shading away from the poles denotes the locations where stable surface ice with a depth > 1 cm has formed in the simulations. Grey shading shows locations where there is seasonal subsurface ice but no stable ice. White shading shows locations where there is no subsurface ice at any time throughout the year. Results are after 820 years of simulation (800 years at T5 resolution and 20 years at T31 resolution), not from an equilibrium ice table depth calculation. Black contours show topography.

640 determined from the diffusion/condensation model, not the equilibrium ice
 641 table depth model). In all simulations at $\varepsilon = 45^\circ$, regions centred around
 642 30°W and 100°E are free of stable subsurface ice due to the combination of
 643 higher regolith thermal inertia and the atmospheric circulation transporting
 644 less water to these regions. These results are similar to those of Jakosky
 645 et al. (2005) when they assumed an atmospheric water content of $100\text{ pr-}\mu\text{m}$,
 646 hypothesised to result from the southern ice cap losing its CO_2 covering
 647 relatively recently, allowing more water to sublime.

648 The results from the clear simulations at $\varepsilon = 45^\circ$ are in agreement with
 649 those of Chamberlain and Boynton (2007), who found that subsurface ice is
 650 not stable close to the equator at $\varepsilon = 45^\circ$. However, the dusty simulation
 651 results, with stable subsurface ice at various locations in the tropics, fall
 652 between those of Chamberlain and Boynton (2007) and those of Mellon and
 653 Jakosky (1995), who found that subsurface ice is stable everywhere at $\varepsilon >$
 654 32° . The reason for the increased abundance of subsurface ice compared to
 655 Chamberlain and Boynton (2007) is likely because the vapour columns in the
 656 tropics in our simulations are larger ($\sim 200\text{ pr-}\mu\text{m}$ compared to $\sim 100\text{ pr-}\mu\text{m}$).
 657 Thus, in order to fully understand the location and stability of subsurface

Figure 10: The maximum ice concentration around each latitude circle, as a function of latitude and depth, for three dusty simulations at $\varepsilon = 45^\circ$. Results are from year 820 (with 800 years at T5 resolution and 20 years at T31 resolution). Ice caps were defined at the south pole in run14 and the north pole in run16 and run18.

ice during past epochs, it is clear that we need an understanding of the dust and water cycles, in particular how the atmospheric circulation affects the near-surface vapour distribution.

While Figure 9 shows the depth to the top of the ice layer, Figure 10 shows the range of depths subsurface ice exists over, in terms of the maximum ice concentrations as a function of latitude and depth. While this is a snapshot of the ice distribution after 820 years, with ice abundances still increasing, it highlights the spatial heterogeneity of the ice deposition. In the mid-latitudes, ice occurs at depths ranging from around 0.5–2.5 m, and, as the poles are approached, the range of depths decreases and the ice exists closer to the surface. In the tropics there is a large variation in ice depth, varying between 1 cm and 2.5 m. This is due to the variation in topography encountered, as well as variations in soil properties and vapour amounts. Subsurface ice does not exist as close to the surface in run16 (Figure 10b), as the orbital parameters result in lower vapour abundances, and hence not enough vapour diffuses into the regolith to exceed the saturation value in the relatively warm tropics. Subsurface ice amounts in the tropics peak at around 50–100 kg m^{-3} depending on the simulation. Larger subsurface ice amounts form in the polar regions, becoming almost pore-filling in the southern polar region in run18 (Figure 10c). These large ice values decrease the porosity of the regolith (which falls to around 0.07 for 300 kg m^{-3} of ice), reducing the

Figure 11: Total subsurface water content (adsorbed water, ice and vapour) over (a–c) the upper 10 cm of regolith, and (d–f) the 10 cm–1.5 m depth range, after 820 years of simulation (800 years at T5 resolution and 20 years at T31 resolution). Results are shown for (a,d) run14, (b,e) run16, and (c,f) run18. Polar ice cap locations are shown with grey shading. Black contours show topography, and white crosses in panels (a,d) show the locations of the 1D simulations in Figure 12.

679 diffusion coefficient and limiting the transport of vapour below the ice layer.
680 In terms of the spatial distribution of subsurface water (not just ice)
681 at high obliquity, Figure 11 shows the global subsurface water content from
682 three simulations (run14, run16 and run18) in the top 10 cm of regolith (upper
683 panels) and between 10 cm–1.5 m (lower panels). The water content is given
684 as a percentage of the total mass, i.e. $\text{wt}\% = 100\sigma/(\sigma + \rho_{\text{reg}})$. It can be seen
685 that the subsurface water distributions are different over the two depth ranges
686 considered. In the upper 10 cm, run14 and run18 (panels a,c) have increased
687 water abundances in the Tharsis region and Arabia Terra, while run18 also
688 has increased water around Elysium Mons and on the dichotomy boundary
689 between Elysium Planitia and Terra Cimmeria (centred around 20°S, 150°E).
690 These locations correspond to areas surrounding stable surface ice deposits
691 (see Figure 9) where the atmospheric water content is large at times when
692 the surface ice is subliming. Little stable surface ice was deposited in run16,

693 hence the upper 10 cm of regolith are relatively dry (panel b).

694 Lower down in the regolith, between 0.1–1.5 m, subsurface ice forms at the
695 pole opposite that where the ice cap was defined. Thus, run14 has subsurface
696 ice at the north pole, while run16 and run18 have subsurface ice at the
697 south pole (Figure 11, panels d–f). Ice amounts are larger in run16 and
698 run18 (up to $\sim 15\%$ water by mass) compared to run14 ($< 5\%$ water by
699 mass), as the peak atmospheric vapour abundances are larger (~ 1000 pr- μm
700 compared to ~ 500 pr- μm). Away from the polar regions, all three simulations
701 show relative increases in water to the north and south-west of the Tharsis
702 region, and around the Argyre and Hellas basins. This is because these
703 regions experience increased atmospheric water content due to transport by
704 topographically-steered flows around Tharsis, and stationary waves around
705 the Argyre and Hellas basins. Peak vapour columns range from 500–1000 pr-
706 μm in run14, and 500–2000 pr- μm in run16 and run18. Additionally, run14
707 and run18 have increased subsurface water around Arabia Terra (resulting
708 from diffusion close to the stable surface ice deposits) and to the north-west
709 of Elysium Mons.

710 Compared to the results from $\varepsilon = 30^\circ$ (Figure 6) it can be seen that
711 the $\varepsilon = 45^\circ$ results are quite different. The reasons for these differences
712 are partly due to circulation changes at high obliquity, which result in a
713 vapour distribution different to that at lower obliquity, but are also due to the
714 increased water abundances. This means that saturation can be reached and
715 ice can form even in regions where the subsurface temperatures are relatively
716 warm. Also, the formation of stable surface ice in various locations can
717 lead to the surrounding regions having relatively large values of subsurface
718 water. Thus, while regolith properties strongly determine the subsurface
719 water distribution in the present day and at $\varepsilon = 30^\circ$, they have less of an
720 impact at high obliquity when the vapour abundance is larger.

721 As these results are after only 820 years of simulation, we chose three lo-
722 cations from run14 to perform further one-dimensional simulations for 50 kyr,
723 as we did for the $\varepsilon = 30^\circ$ case. We chose one location in the north polar
724 region (78°N , 135°W), one to the east of Alba Mons (33°N , 90°W) and one
725 to the south west of the Tharsis region (33°S , 155°W), in order to obtain
726 a variety of different conditions. The results are shown in Figure 12. The
727 simulation in Figure 12c uses a time step of 30 s (as opposed to 1 minute at
728 the other locations), as the large near-surface vapour values, and the long pe-
729 riod of time diffusion is occurring over (see Figure 12d), can cause numerical
730 instabilities with longer time steps.

Figure 12: Variation of (a–c) the filling fraction, and (d) the near surface water abundance as in Figure 7, but for three locations at $\varepsilon = 45^\circ$.

Stable ice in the north polar region (Figure 12a) exists deeper in the regolith than at the other two locations. This is because the high obliquity results in warmer peak subsurface temperatures. For example, at a depth of 30 cm, peak temperatures at 78°N are ~ 255 K, compared to ~ 210 K at the other two locations. Subsurface temperatures are similar at 33°N and 33°S, but near-surface vapour abundances are larger at 33°S (average value of 11.4 mg m^{-3} compared to 4.6 mg m^{-3} at 33°N), and so ice is more extensive and exists closer to the surface. Similar one-dimensional simulations were also performed for the corresponding clear simulation (run13) but the results are not shown as no subsurface ice formed, except at 33°N, where the results are the same as in Figure 12b.

After 50 kyr, the top of the ice table ranges from a depth of ~ 90 cm at 78°N to ~ 4 cm at 33°S. As for the $\varepsilon = 30^\circ$ case, to see to what depth ice would be stable given a longer period of time, we filled the regolith with ice from the surface down to the layer with the largest ice abundance, and ran the simulations until the ice distribution reached equilibrium. The results are shown by the grey shading in Figure 12. As can be seen, the top of the ice table ranges from ~ 50 cm at 78°N to ~ 2 cm at 33°S. The stability depth of ice at 78°N is greater than at $\varepsilon = 30^\circ$, where it was stable at ~ 20 cm (see Figure 7a). This is because the subsurface temperatures in the polar region are warmer at high obliquity.

6.2. Tropical ice reservoir

As in the previous results with polar ice caps, dusty simulations, with a tropical ice reservoir acting as an unlimited source of water, result in in-

Figure 13: Depth to stable subsurface ice deposits for two dusty simulations at $\varepsilon = 45^\circ$, with an ice reservoir defined in the Tharsis region (red rectangle). Red shading elsewhere denotes the locations where stable surface ice with a depth > 1 cm has formed in the simulations. Grey shading shows locations where there is seasonal subsurface ice but no stable ice. White shading shows locations where there is no subsurface ice at any time throughout the year. Results are after 820 years of simulation (800 years at T5 resolution and 20 years at T31 resolution), not from an equilibrium ice table depth calculation. Black contours show topography.

creased formation of subsurface ice. Figure 13 shows the depth to stable subsurface ice for two dusty simulations after 820 years of simulation (the results of run24 are similar to run22). It can be seen that stable subsurface ice forms in both polar regions, initially at depths below ~ 1 m. However, with time this ice moves closer to the surface as in Figure 12a. Stable subsurface ice close to the surface is found around Tempe Terra and Alba Mons (centred $\sim 30^\circ\text{N}$) and around Solis Planum and Argyre Planitia (centred $\sim 40^\circ\text{S}$). This is a result of the transport of large vapour values from the nearby Tharsis ice sheet, with vapour columns in the region reaching peak values of around $1000 \text{ pr-}\mu\text{m}$ (annual mean values are around $30\text{--}100 \text{ pr-}\mu\text{m}$; see Figure 4d). As this transport is mainly meridional rather than zonal, the majority of the tropics are free of stable subsurface ice. Stable surface ice deposits also build up, which are shaded red. In Figure 13a these exist in similar regions to those in Figure 9(a,c), i.e. around the Tharsis region. Due to the large vapour values caused by the subliming Tharsis ice sheet, ice is accumulating at a rate of $\sim 1 \text{ m yr}^{-1}$ over Olympus Mons, but around $0.01\text{--}0.2 \text{ m yr}^{-1}$ at the other locations. In Figure 13b there is only one stable surface ice deposit away from the Tharsis ice sheet, located at around 30°S , 90°W . This reduction in stable surface ice is due to the warmer summertime temperatures.

774 7. Discussion

775 We find that the subsurface water content is affected by both the proper-
 776 ties of the regolith (thermal inertia and albedo) and the atmospheric trans-
 777 port of vapour, with the relative importance of each depending upon the
 778 orbital parameters. At low to moderate obliquities ($\varepsilon = 15^\circ\text{--}30^\circ$), the re-
 779 golith properties have more control on the subsurface water distribution in
 780 the tropics and mid-latitudes than changes to the eccentricity, e or solar lon-
 781 gitude of perihelion, L_p . This is because changes to e and L_p only have a
 782 small effect on the amount of water sublimed from the polar ice cap, and
 783 as such the atmospheric water content remains relatively low (compared to
 784 at high obliquity). Thus, the amount of water diffused into the regolith in
 785 the tropics and mid-latitudes is not large enough to reach saturation and
 786 form ice, so the water exists mostly in its adsorbed state. The amount of
 787 adsorbed water decreases as temperatures increase (as the water is mobilised
 788 into vapour which then diffuses out of the regolith), so low thermal inertia re-
 789 gions (which have greater day-night temperature variability but lower mean
 790 temperatures) tend to have locally higher subsurface water contents. Con-
 791 versely, at high obliquity ($\varepsilon = 45^\circ$) atmospheric vapour abundances become
 792 much larger, and more vapour can diffuse further into the regolith, allowing
 793 subsurface ice to form at all latitudes (but not all longitudes).

794 Compared to observations of the present-day global subsurface water dis-
 795 tribution by the MONS instrument (Feldman et al., 2004, 2007, 2008; Mau-
 796 rice et al., 2011), the simulations for $\varepsilon = 30^\circ$ with a north polar ice cap show
 797 the best agreement. This suggests the subsurface water distribution revealed
 798 by MONS is likely to occur for a range of obliquities close to that of today,
 799 when the atmospheric water content is supplied by a north polar ice cap.
 800 The circulation of the atmosphere is also important in producing the ob-
 801 served subsurface water distribution. The MONS observations of increased
 802 subsurface water abundances and decreased burial depths centred roughly
 803 at 135°W , 70°N and 90°E , 70°S (Feldman et al., 2007; Maurice et al., 2011)
 804 are the result of increased atmospheric water content due to transport by
 805 topographically-steered flows to the west of Tharsis during northern hemi-
 806 sphere summer, and stationary waves to the east of Hellas during southern
 807 hemisphere summer. As these circulation patterns are due to topographic
 808 features, it is likely the same regions will also experience increased subsurface
 809 water at different obliquities, and shows the importance of considering the
 810 effect of the atmospheric circulation on the near-surface vapour values.

811 While there is agreement with the spatial distribution of subsurface water,
 812 our model results at $\varepsilon = 30^\circ$ tend to have lower water masses than observed
 813 by MONS. One reason for this may be the choice of adsorption isotherm,
 814 with the amount of water adsorbed differing by a factor of ~ 100 between
 815 different isotherms (Zent and Quinn, 1997). In the polar regions, a match to
 816 the MONS measurements of $\sim 60\%$ water by weight would only be possible if
 817 the regolith porosity was $\sim 80\%$, with all the pore space being filled with ice.
 818 However, this is larger than the porosity determined at the Viking Lander
 819 1 site (Moore and Jakosky, 1989; Boynton et al., 2002), and larger than the
 820 maximum predicted porosity of 65% (Prettyman et al., 2004). A more likely
 821 reason for the large water values observed by MONS in the polar regions
 822 would be the combination of pore ice from vapour diffusion along with ice
 823 lenses (Sizemore et al., 2015) or buried surface ice (Jakosky and Carr, 1985;
 824 Mischna et al., 2003; Levrard et al., 2004). In the equatorial region, the high
 825 water-equivalent hydrogen abundances of $\sim 10\%$ by weight (which cannot be
 826 matched by the GCM) are likely due to the presence of hydrous minerals
 827 (e.g. Bish et al., 2003; Wang et al., 2013; Karunatillake et al., 2014). Our
 828 drier regolith is also likely related to the fact that our GCM simulations have
 829 run for only ~ 800 years, and diffusion of significant amounts of water is a
 830 relatively slow process (the additional one-dimensional simulations show that
 831 large amounts of pore ice can accumulate over thousands of years in various
 832 locations).

833 While subsurface ice in the tropics and mid-latitudes did not form in our
 834 simulations at $\varepsilon = 15^\circ$, and only formed over the polar regions at $\varepsilon = 30^\circ$,
 835 we can use the near-surface vapour values and regolith temperatures to de-
 836 termine the equilibrium ice table depth. At $\varepsilon = 15^\circ$ we predict stable sub-
 837 surface ice in the region around Olympus Mons and Alba Mons, while none
 838 was predicted in previous studies by Mellon and Jakosky (1995) and Cham-
 839 berlain and Boynton (2007). This difference is likely related to the vapour
 840 values used, as Mellon and Jakosky (1995) assumed a value of $2.62 \text{ pr-}\mu\text{m}$
 841 and Chamberlain and Boynton (2007) assumed a value of $0.33 \text{ pr-}\mu\text{m}$, while
 842 our simulations have values of $\sim 15 \text{ pr-}\mu\text{m}$. At $\varepsilon = 30^\circ$ we predict less ex-
 843 tensive stable subsurface ice deposits than Mellon and Jakosky (1995), as we
 844 have lower vapour columns in our simulations of $\sim 60\text{--}80 \text{ pr-}\mu\text{m}$ (compared to
 845 $232 \text{ pr-}\mu\text{m}$). We also predict stable subsurface ice in the Tharsis region, while
 846 Chamberlain and Boynton (2007) do not. This is because the near-surface
 847 vapour values in the Tharsis region are increased due to the sedimentation
 848 (and subsequent sublimation) of ice particles from the thick clouds which

849 form. Thus, it is important to consider the atmospheric circulation when de-
850 termining to what depth ice may be stable, particularly away from the polar
851 regions, as this can have a large effect on the near-surface vapour values

852 At $\varepsilon = 45^\circ$, subsurface ice forms only in the polar regions in the clear sim-
853 ulations, while in the dusty simulations a large amount of stable subsurface
854 ice exists at all latitudes (though not all longitudes). When polar ice caps
855 are defined, the subsurface ice is extensive around the Tharsis region, and
856 around Arabia Terra and north of Elysium. When an ice reservoir is instead
857 defined over the Tharsis Montes, subsurface ice is much less abundant in the
858 tropics, and exists mainly to the north and south of the ice reservoir (as the
859 circulation preferentially transports the sublimed vapour meridionally rather
860 than zonally). Indeed, in all simulations a dusty atmosphere leads to in-
861 creased subsurface water contents compared to a clear atmosphere. There
862 are two reasons for this. Firstly, the dustier atmosphere is warmer and acts
863 to transport more water away from the subliming ice reservoirs. Secondly,
864 the dusty simulations have lower daytime surface temperatures, so less of
865 the adsorbed water diffuses back into the atmosphere, and vapour can reach
866 saturation as regolith temperatures are lower. As shown by Newman et al.
867 (2005), dust lifting by both near-surface wind stress and dust devils increases
868 with increasing obliquity, so Mars at high obliquity was likely dustier than
869 it is today. Thus it is important to gain a better understanding of the dust
870 cycle at past epochs, as it has a large effect on subsurface ice stability.

871 It should be noted that the simulations performed here do not include the
872 radiative impact of clouds. Recent modelling and data assimilation studies
873 have shown that clouds in the present-day can influence the temperature
874 structure of the atmosphere and strengthen the overturning circulation (e.g.
875 Wilson et al., 2008; Madeleine et al., 2012; Steele et al., 2014a; Navarro
876 et al., 2014), potentially bringing modelling results into better agreement
877 with observations. Madeleine et al. (2014) have also shown that, at 35°
878 obliquity, cloud radiative effects can increase the atmospheric vapour content
879 by more than an order of magnitude. The effect is similar to that in our
880 dusty simulations, in which increased atmospheric temperatures strengthen
881 the meridional circulation and allow more vapour to be transported away
882 from the subliming polar ice caps in the summer. We chose not to include
883 cloud radiative effects in our simulations for a number of reasons. Firstly,
884 the GCM was run at a low resolution with grid boxes of 22.5° , to enable
885 many hundreds of years to be simulated. As such, the horizontal extent of
886 any clouds in the tropics will be larger than would occur in reality, which will

887 affect the scale of any heating/cooling. Secondly, there is a complex coupling
888 between the dust and water cycles, and here we are only considering either
889 clear or dusty conditions, and not transporting dust particles (again, because
890 of the low resolution of the GCM, but also due to a lack of knowledge of the
891 dust cycle in past epochs). Thirdly, while cloud radiative effects have led to
892 improvements in model predictions of the present-day climate, they also lead
893 to some inconsistencies (e.g. Madeleine et al., 2012; Navarro et al., 2014),
894 and for past epochs we cannot check with observations to see which features
895 are likely correct or incorrect. Thus, we note that it is possible that in
896 reality the water cycle may have been wetter than represented in our $\varepsilon = 45^\circ$
897 simulations, which would result in larger subsurface ice values than predicted,
898 and possibly increased rates of surface ice accumulation in the tropics.

899 8. Conclusions

900 We have used a GCM with a regolith diffusion model in order to study
901 the regolith-atmosphere exchange of vapour in Mars' recent past. We have
902 run GCM simulations for hundreds of years, and one-dimensional simulations
903 for thousands of years, and considered various orbital parameters (obliquity,
904 eccentricity and the solar longitude of perihelion), atmospheric dust contents
905 and surface ice reservoirs. Our main findings are as follows:

- 906 1. It is important to consider the circulation of the atmosphere when
907 determining the subsurface water distribution. The simulations per-
908 formed here show that circulation patterns forced by topography lead
909 to increased near-surface (and hence increased subsurface) water abun-
910 dances in regions where MONS observations show increased subsurface
911 water abundances and decreased burial depths. Due to the nature of
912 the topographic forcing, it is likely the same regions will also have ex-
913perienced increased subsurface water in the past, and will do so in the
914 future.
- 915 2. The dustiness of the atmosphere plays an important role in the dis-
916 tribution of subsurface water. As the dustiness increases, the warmer
917 atmospheric temperatures allow more water to sublime and be trans-
918 ported away from surface ice deposits. The colder surface temperatures
919 in a dusty atmosphere mean subsurface ice can be stable closer to the
920 surface and closer to the equator, particularly at high obliquity where
921 atmospheric humidities can be very large. Newman et al. (2005) showed

922 an increase in dust lifting with increasing obliquity, so Mars at high
 923 obliquity was likely dustier than it is today, and extensive subsurface
 924 ice may have been prevalent.
 925 3. At $\varepsilon = 45^\circ$, large regions of stable subsurface ice form in the midlati-
 926 tudes and tropics when a permanent polar ice cap is defined. However,
 927 when defining a tropical ice sheet in the Tharsis region instead, the ma-
 928 jority of the tropics are free of stable subsurface ice. This is because the
 929 atmospheric circulation preferentially transports the subliming vapour
 930 away from the ice sheet meridionally rather than zonally.

931 Appendix A. Regolith diffusion model

932 The one-dimensional diffusion calculation operates on the water vapour
 933 plus adsorbed water component of each grid box, e.g. on $m = \phi n + \alpha$ (see
 934 Equation 1). The regolith scheme thus solves the equation

$$\frac{\delta m}{\delta t} = \frac{\delta J}{\delta z}, \quad (\text{A.1})$$

935 with $J = D\delta n/\delta z$, where δz is the depth below the surface.

936 Adsorption of water vapour onto regolith grains uses an empirical relation
 937 from Fanale and Cannon (1971):

$$\alpha(n, T) = \frac{\rho_r \beta \sqrt{n}}{e^{\delta/T}} \left(\frac{k_B T}{m_w} \right)^{0.51} = F(T) \sqrt{n}, \quad (\text{A.2})$$

938 where ρ_r is the regolith density, $\beta = 2.043 \times 10^{-8} \text{ Pa}^{-1}$, $\delta = -2679.8 \text{ K}$,
 939 T is the temperature, k_B is the Boltzmann constant and m_w is the mass
 940 of a water molecule. Thus, at time t in regolith layer k we have $m_k^t =$
 941 $\phi_k^t n_k^t + F(T_k^t) \sqrt{n_k^t} = \phi_k^t n_k^t + F_k^t \sqrt{n_k^t}$. Solving for n_k^t gives

$$n_k^t = \frac{m_k^{t^2}}{F_k^{t^2}} \left(1 - \frac{2m_k^t \phi_k^t}{F_k^{t^2}} \right). \quad (\text{A.3})$$

If ice is present, Equation A.3 will result in $n_k^t = n_{\text{sat}}$. Equation A.3 is used
 to obtain a relation for the vapour increment:

$$\begin{aligned} \delta n_k^t &= n_k^t - n_k^{t-dt} \\ &= \frac{m_k^{t^2}}{F_k^{t^2}} \left(1 - \frac{2m_k^t \phi_k^t}{F_k^{t^2}} \right) - \frac{m_k^{t-dt^2}}{F_k^{t-dt^2}} \left(1 - \frac{2m_k^{t-dt} \phi_k^{t-dt}}{F_k^{t-dt^2}} \right). \end{aligned} \quad (\text{A.4})$$

942 To remove the dependence of Equation A.4 on m_k^t , we substitute in $m_k^t =$
 943 $m_k^{t-dt} + \delta m_k^t$, and ignore terms with powers of δm_k^t (since $\delta m_k^t \ll m_k^t$). This
 944 results in the equation $\delta n_k^t = A_k^t \delta m_k^t + B_k^t$, where

$$A_k^t = \frac{2m_k^{t-dt}}{F_k^{t2}} - \frac{6m_k^{t-dt2}\phi_k^{t-dt}}{F_k^{t4}} \quad (\text{A.5})$$

945 and

$$B_k^t = \left(\frac{1}{F_k^{t2}} - \frac{1}{F_k^{t-dt2}} \right) m_k^{t-dt2} - 2\phi_k^{t-dt} \left(\frac{1}{F_k^{t4}} - \frac{1}{F_k^{t-dt4}} \right) m_k^{t-dt3}. \quad (\text{A.6})$$

946 Thus, we have

$$\frac{\delta m_k^t}{\delta t} = \frac{\delta n_k^t - B_k^t}{A_k^t \delta t}, \quad (\text{A.7})$$

947 and hence Equation A.1 becomes

$$\frac{n_k^t - n_k^{t-dt} - B_k^t}{A_k^t \delta t} = \frac{J_{k+0.5}^t - J_{k-0.5}^t}{z_{k+0.5} - z_{k-0.5}}, \quad (\text{A.8})$$

948 where

$$J_{k+0.5}^t = D_{k+0.5}^t \frac{n_{k+1}^t - n_k^t}{z_{k+1} - z_k}. \quad (\text{A.9})$$

949 The $k \pm 1$ subscripts denote terms evaluated at layer midpoints, with the
 950 $k \pm 0.5$ terms evaluated at layer boundaries. Combining Equations A.8 and
 951 A.9 gives

$$C_k^t (n_k^{t-dt} + B_k^t) = \Pi_k^t n_k^t - d_{k-0.5}^t n_{k-1}^t - d_{k+0.5}^t n_{k+1}^t, \quad (\text{A.10})$$

where

$$\begin{aligned} C_k^t &= (z_{k+0.5} - z_{k-0.5}) / A_k^t \delta t, \\ \Pi_k^t &= C_k^t + d_{k-0.5}^t + d_{k+0.5}^t, \\ d_{k+0.5}^t &= D_{k+0.5}^t / (z_{k+1} - z_k). \end{aligned}$$

We require a solution to Equation A.10 of the form $n_k^t = \gamma_{k-0.5}^t n_{k-1}^t + \beta_{k-0.5}^t$. Thus, in order to remove the n_{k+1}^t term from Equation A.10, we

substitute with $n_{k+1}^t = \gamma_{k+0.5}^t n_k^t + \beta_{k+0.5}^t$. This results in

$$\begin{aligned} n_k^t &= \left(\frac{d_{k-0.5}^t}{\Pi_k^t - d_{k+0.5}^t \gamma_{k+0.5}^t} \right) n_{k-1}^t + \frac{C_k^t (n_k^{t-dt} + B_k^t) + d_{k+0.5}^t \beta_{k+0.5}^t}{\Pi_k^t - d_{k+0.5}^t \gamma_{k+0.5}^t} \\ &= \gamma_{k-0.5}^t n_{k-1}^t + \beta_{k-0.5}^t. \end{aligned} \quad (\text{A.11})$$

952 Once n_k^t has been calculated for each regolith layer using Equation A.11,
 953 the increments $\delta n_k^t = n_k^t - n_k^{t-dt}$ are calculated, leading to the new total
 954 water content $\sigma_k^t = \sigma_k^{t-dt} + (\delta n_k^t - B_k^t)/A_k^t$. Once σ_k^t is known, partitioning
 955 of the water content into the three states is carried out. If $n_k^t \leq n_{\text{sat}}$ (with
 956 n_k^t calculated from Equation A.3, though now using the total water content,
 957 σ_k^t , rather than the ice-free content m_k^t) then $\alpha_k^t = F(T_k^t) \sqrt{n_k^t}$ and $\zeta_k^t = 0$. If
 958 $n_k^t > n_{\text{sat}}$ then $n_k^t = n_{\text{sat}}$, $\alpha_k^t = F(T_k^t) \sqrt{n_{\text{sat}}}$ and $\zeta_k^t = \sigma_k^t - \phi_k^t n_{\text{sat}} - F(T_k^t) \sqrt{n_{\text{sat}}}$.
 959 Thus, the diffusion calculation uses the ice-free water content in each grid
 960 box to calculate the vapour flux. This is to avoid unrealistically large values
 961 of the A_k^t and B_k^t coefficients, which would result from large values of σ_k^t .
 962 The vapour change in each grid box is then added to the previous time step's
 963 total water content, and this is used in the condensation-sublimation step.

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