

DYNAMIC/THERMODYNAMIC SIMULATIONS OF THE NORTH POLAR ICE CAP OF MARS. R. Greve, *Institut für Mechanik III, Technische Universität Darmstadt, D-64289 Darmstadt, Germany, greve@mechanik.tu-darmstadt.de.*

Ice-sheet model SICOPOLIS

The present permanent north polar water ice cap of Mars is investigated with the dynamic/thermodynamic ice-sheet model SICOPOLIS (Simulation COde for POLythermal Ice Sheets), which was originally developed for and applied to terrestrial ice sheets like Greenland, Antarctica and the glacial northern hemisphere (Greve, 1997b, c; Calov et al., 1998; Greve et al., 1998). The model is based on the continuum-mechanical theory of polythermal ice masses (Hutter, 1982, 1993; Greve, 1997a), which describes the material ice as a density-preserving, heat-conducting power-law fluid with thermo-mechanical coupling due to the strong temperature dependence of the ice viscosity. It is further distinguished between *cold ice* with a temperature below the pressure melting point and *temperate ice* with a temperature at the pressure melting point, the latter being considered as a binary mixture of ice and small amounts of water. The influence of the considerable dust content of the ice cap on the mechanical properties of the ice is neglected.

The model computes three-dimensionally the temporal evolution of ice extent, thickness, temperature, water content and age as a response to external forcing. The latter must be specified by (i) the mean annual air temperature above the ice, (ii) the surface mass balance (ice accumulation minus melting and evaporation), (iii) the global sea level (not relevant for Martian applications) and (iv) the geothermal (areothermal) heat flux from below into the ice body.

Simulation set-up

The surface topography, h , of the permanent north polar water ice cap of Mars used here is based on the map constructed by Dzurlin and Blasius (1975), which was slightly filtered and digitized to a 40-km grid for this study, and complemented by the ice margin contour given by Budd et al. (1986). As for the ice thickness, H , a Gaussian distribution

$$H = H_0 e^{-r^2/r_0^2} \quad (1)$$

is used, where r is the distance from the position of maximum ice thickness, H_0 , assumed to be at 87°N, 0°W, and $r_0 = 400$ km. Based on the horizontal extent and observed surface undulations of the ice cap, Budd et al. (1986) argue that H_0 should be about 4 km. From the digitized surface topography and eq. (1), an estimated digitized bedrock topography, b , follows via $b = h - H$. Further, application of a local isostatic balance between ice load and lithosphere buoyancy yields the relaxed bedrock topography with no ice load, b_0 , as

$$b_0 = b + \frac{\rho}{\rho_a} H, \quad (2)$$

where ρ is the ice density (910 kg/m³) and ρ_a the mantle density, taken as 2350 kg/m³ (terrestrial value 3300 kg/m³

times mean-density ratio Mars/Earth). The bedrock response to changing ice loads is modelled by a delayed local isostatic balance with the time lag $\tau_V = 3000$ yr.

According to the data listed by Budd et al. (1986), the mean annual air temperature above the ice, T_{ma} , is described by a parameterization depending on elevation, h , and co-latitude, $\tilde{\phi}$ ($\tilde{\phi} = 90^\circ\text{N} - \phi$, where ϕ is the latitude),

$$T_{ma} = T_{ma}^0 + \gamma_{ma} h + c_{ma} \tilde{\phi}, \quad (3)$$

with $T_{ma}^0 = -90^\circ\text{C}$, the mean lapse rate $\gamma_{ma} = -2.5^\circ\text{C}/\text{km}$, and $c_{ma} = 1.5^\circ\text{C}/^\circ\text{lat}$. The accumulation of water ice on the surface of the ice cap is assumed to be spatially constant. Since the water vapour density in the Martian atmosphere is approximately 1/1000th the terrestrial value, and typical accumulation rates for terrestrial ice sheets are about 300 mm WE/yr (Greenland), the order of magnitude of the accumulation rate, S , can be estimated as 0.1 . . . 1 mm WE/yr. Surface melting/evaporation is parameterized by the standard degree-day method with terrestrial (Greenland) values for the snow- and ice-melt factors, $\beta_{snow} = 3$ mm WE/(d°C) and $\beta_{ice} = 12$ mm WE/(d°C), an amplitude of the annual temperature signal of 30°C and a standard deviation of additional temperature variations of 10°C. Further, the simulated ice cap is restricted to its present extent, and the areothermal heat flux is set to 33.5 mW/m² (Budd et al., 1986).

Results

A series of simulations were carried out, where the measured/estimated topography described above is used as initial condition, and the time-forward integration is conducted until the simulated ice cap is in steady state with the present climate forcing defined by the air temperature, the surface mass balance and the areothermal heat flux. For the initial ice volume, V_{init} , the three values $2 \cdot 10^6$ km³, $3 \cdot 10^6$ km³ and $4 \cdot 10^6$ km³ are used, corresponding to maximum initial ice thicknesses of $H_{init} = 4.26$ km, 6.38 km and 8.51 km, respectively. The accumulation rate is varied such that the initial ice volume is reproduced in the steady state as accurately as possible.

No.	V_{init}	H_{init}	S	t_{ss}	V_{ss}	H_{ss}
1	2.00	4.26	0.04	35	2.01	3.27
2	3.00	6.38	0.4	4	3.00	4.67
3	4.00	8.51	3.0	1	3.97	6.42

Table 1: Results of steady-state simulations no. 1-3 with varied ice volume. V_{init} , V_{ss} in 10^6 km³, H_{init} , H_{ss} in km, S in mm WE/yr, t_{ss} in Myr (quantities are explained in the main text).

Table 1 lists V_{init} , H_{init} , S , the time required to reach the steady state, t_{ss} , and the ice volume and the maximum ice thickness in the steady state, V_{ss} and H_{ss} , for the three

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simulations with optimum accumulation rate to reproduce the initial volume (referred to as simulations no. 1, 2, 3, respectively). Evidently, the accumulation rate required to maintain the initial volume varies by two orders of magnitude for the three simulations, and it is within the above estimated range of 0.1 . . . 1 mm WE/yr for simulation no. 2, which is therefore regarded as the most realistic reference simulation. Accordingly, t_{ss} (which is also a measure for the time necessary to build up the ice cap from ice-free initial conditions) varies by two orders of magnitude, and it is in any case much larger than for ice sheets on Earth [~ 100 kyr for Greenland, see Greve (1997b)].

Even though V_{init} and V_{ss} agree very well for the three simulations, the corresponding maximum thicknesses H_{init} and H_{ss} differ by about 25%. The reason for this is that the ice flow tends to redistribute the initial (measured) surface topography with steeper gradients in the interior and flatter slopes towards the margin to a more parabolic shape with the opposite behaviour. A reason for this shortcoming may be the tentative use of the terrestrial degree-day parameterization for surface melting/evaporation, which is not likely to describe the surface processes adequately under the very different Martian conditions.

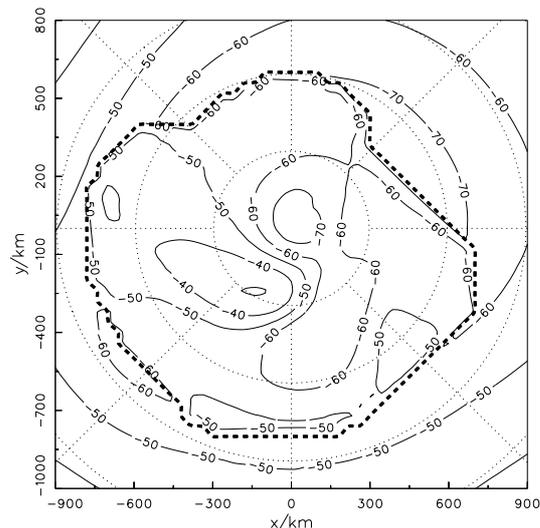


Fig. 1. Basal temperature relative to pressure melting for the steady state of simulation no. 2, in $^{\circ}\text{C}$. The dashed heavy line indicates the ice margin, latitude circles are spaced by 5° , the prime meridian 0°W points downward.

The ice-flow velocities are considerably smaller than for terrestrial ice sheets due to the reduced gravity acceleration and the low temperatures. For simulation no. 1 the maximum surface velocity in steady state is 0.058 m/yr, for simulation no. 2 it is 0.64 m/yr, and for simulation no. 3 it is 12.7 m/yr. If even smaller ice volumes/thicknesses than those of simulation no. 1 were assumed, virtually no ice flow would result, which is not very likely with regard to recent laser altimeter results of the Mars Global Surveyor (MGS) space probe which have revealed that large parts of the ice cap are very smooth (MGS

Press Conference, Spring Meeting of the American Geophysical Union, Boston, 1998), much like the flowing ice sheets on Earth.

It is further noticeable that for none of the three simulations the basal temperature reaches the pressure melting point. Fig. 1 depicts the basal temperature (relative to pressure melting) for the reference simulation no. 2, the maximum of -29.3°C being reached at 85.3°N , 30.3°W . For simulation no. 1, the maximum basal temperature is -46.9°C , for simulation no. 3 it is -4.4°C , already very close to pressure melting.

These results are strongly dependent on the applied areothermal heat flux, which is not very well known. In order to investigate under which circumstances pressure melting can be reached at the ice base, the reference simulation no. 2 was re-run with larger areothermal heat fluxes, namely 40, 50, 60 and 70 mW/m^2 . As a consequence, the maximum basal temperatures rise to -22.4°C , -12.9°C , -4.6°C and 0°C , respectively. Hence, for 70 mW/m^2 pressure melting is reached at the ice base, with an area at pressure melting of 4784 km^2 or 0.34% of the simulated total ice-covered area, and a basal melting rate of $3.3 \cdot 10^5 \text{ m}^3 \text{ WE/yr}$.

References

- Budd, W. F., D. Janssen, J. H. I. Leach, I. N. Smith and U. Radok. 1986. The north polar ice cap of Mars as a steady-state system. *Polarforschung*, **56** (1/2), 43-46.
- Calov, R., A. Savvin, R. Greve, I. Hansen and K. Hutter. 1998. Simulation of the Antarctic ice sheet with a three-dimensional polythermal ice sheet model, in support of the EPICA project. *Ann. Glaciol.*, **27** (in press).
- Dzurisin, D. and K. R. Blasius. 1975. Topography of the polar layered deposits of Mars. *J. Geophys. Res.*, **80** (23), 3286-3306.
- Greve, R. 1997a. A continuum-mechanical formulation for shallow polythermal ice sheets. *Phil. Trans. R. Soc. Lond., Ser. A*, **355**, 921-974.
- Greve, R. 1997b. Application of a polythermal three-dimensional ice sheet model to the Greenland Ice Sheet: Response to steady-state and transient climate scenarios. *J. Climate*, **10** (5), 901-918.
- Greve, R. 1997c. Large-scale ice-sheet modelling as a means of dating deep ice cores in Greenland. *J. Glaciol.*, **43** (144), 307-310; Erratum **43** (145), 597-600.
- Greve, R., K.-H. Wyrwoll and A. Eisenhauer. 1998. Deglaciation of the northern hemisphere at the onset of the Eemian and of the Holocene. *Ann. Glaciol.*, **28** (submitted).
- Hutter, K. 1982. A mathematical model of polythermal glaciers and ice sheets. *J. Geophys. Astrophys. Fluid Dyn.*, **21**, 201-224.
- Hutter, K. 1993. Thermo-mechanically coupled ice sheet response. Cold, polythermal, temperate. *J. Glaciol.*, **39** (131), 65-86.