GLACIATION OF MARS FROM 10 MILLION YEARS AGO UNTIL 10 MILLION YEARS INTO THE FUTURE SIMULATED WITH THE MODEL MAIC-2. R. Greve[1], B. Grieger[2] and O. J. Stenzel[3]. [1]Institute of Low Temperature Science, Hokkaido University, Kita-19, Nishi-8, Kita-ku, Sapporo 060-0819, Japan (greve@lowtem.hokudai.ac.jp), [2]European Space Astronomy Centre, P.O. Box – Apdo. de Correos 78, 28691 Villanueva de la Cañada, Madrid, Spain, [3]Max-Planck-Institut für Sonnensystemforschung, Max-Planck-Straße 2, 37191 Katlenburg-Lindau, Germany.

Introduction: The Mars Atmosphere-Ice Coupler MAIC-2 is a simple, latitudinal model, which consists of a set of parameterizations for the surface temperature, the atmospheric water transport and the surface mass balance (condensation minus evaporation) of water ice [1]. It is driven directly by the orbital parameters obliquity, eccentricity and solar longitude (L_s) of perihelion [2]. The main purpose of MAIC-2 is to simulate the glaciation of Mars over time.

Model MAIC-2: In the following a brief outline of the model MAIC-2 is given. For more details see Ref. [1].

Surface Temperature. In order to derive a parameterization for the daily mean local surface temperature $T(\phi,t)$ (depending on latitude ϕ and time t), the radiation balance

$$\sigma T^4 = (1-A)F \qquad (1)$$

is used, where σ is the Stefan-Boltzmann constant ($\sigma = 5.67 \times 10^{-8}\ \mathrm{W\,m^{-2}\,K^{-4}}$), A is the surface albedo (globally $A = 0.3$ assumed) and F is the local daily mean insolation as a function of the orbital parameters. In the absence of seasonal CO_2 frost, Eq. (1) provides reasonable results for the surface temperature. However, the equation does not account for the fact that at $T = T_{\mathrm{cond}} \approx 148$ K condensation of atmospheric CO_2 sets in. In this case, the temperature is kept at $T = T_{\mathrm{cond}}$, and the growth and decay of the seasonal CO_2 ice cap is modeled by a simple energy balance.

The results obtained with this scheme agree well with the data given in the Mars Climate Database that are based on output of a General Circulation Model for the Martian atmosphere [3].

Saturation Pressure of Water Vapour. The water-vapour saturation pressure P_{sat} is obtained from the Clausius-Clapeyron relation, which can be integrated only approximately. Different approximations are available; in MAIC-2 the Magnus-Teten formula for water vapour over ice [4] is used.

Evaporation. Ingersoll [5] discussed the water vapour mass flux in the Martian carbon dioxide atmosphere. The evaporation rate E of water from the surface, expressed as a mass flux in kg $\mathrm{m^{-2}\,s^{-1}}$, is

$$E = E_0 \times 0.17 \Delta\eta\, \rho D \left(\frac{(\Delta\rho/\rho) g}{\nu^2} \right)^{1/3}, \qquad (2)$$

where $E_0 < 1$ is the dust insulation factor, $\Delta\eta$ the concentration difference at the surface and at distance, ρ the atmospheric density, $\Delta\rho$ the CO_2 density difference at the surface and at distance, D the diffusion coefficient of water in CO_2 ($1.4 \times 10^{-3}\ \mathrm{m^2\,s^{-1}}$), g the acceleration due to gravity ($3.72\ \mathrm{m\,s^{-2}}$) and ν the kinematic viscosity of CO_2 ($6.93 \times 10^{-4}\ \mathrm{m^2\,s^{-1}}$) [6]. For the detailed forms of the terms $\Delta\eta$, ρ and $\Delta\rho/\rho$ see [1].

Sears and Moore [6] state that the evaporation rate of ice is probably about half that of liquid water. In addition, any significant evaporation of the dusty ice will lead to an enrichment of dust at the surface, thus producing an insulating layer which hampers further evaporation. These effects can be accounted for by setting the evaporation factor E_0 in Eq. (2) to a value less than unity, and we will use a standard value of $E_0 = 0.1$ [1].

Condensation. The water content W in the atmosphere is is expressed as an area mass density, i.e., in units of kg/m^2. Multiplied with the acceleration due to gravity g, this becomes equivalent to the partial pressure of water vapour at the surface. Thus we compare this pressure to the water vapor saturation pressure P_{sat} and condense all excessive water, i.e.,

$$C = \frac{1}{\Delta t}\left(W - \frac{P_{\mathrm{sat}}(T)}{g} \right), \quad \text{if } gW > P_{\mathrm{sat}}(T), \qquad (4)$$

where Δt is the model time step.

Transport. Atmospheric transport of water vapour is approximated by instantaneous mixing (on a time scale of several sols = Martian days).

Ice evolution. With the condensation C and the evaporation E, the net mass balance a_{net} of the water ice deposits, expressed in m ice equiv. $\mathrm{a^{-1}}$, is

$$a_{\mathrm{net}} = \frac{C-E}{\rho_{\mathrm{ice}}}, \qquad (5)$$

where $\rho_{\mathrm{ice}} = 910\ \mathrm{kg\,m^{-3}}$ is the density of ice. Since glacial flow is neglected, the evolution of the ice thickness, H, is simply

$$\frac{\partial H}{\partial t} = a_{\mathrm{net}}. \qquad (6)$$

Note that MAIC-2 allows for negative ice thicknesses ($H < 0$). Such as situation is interpreted as ground ice under an ice-free regolith layer.

Simulation Set-up: We carry out a simulation from 20 million years ago ($t = –20$ Ma) until 10 million years into the future ($t = +10$ Ma), driven by the orbital parameters obliquity (axial tilt), eccentricity and solar longitude (L_s) of perihelion computed by Laskar and others [2]. The obliquity is shown in Fig. 1. It shows two distinct periods; *stage 1* prior to ~ 4 Ma ago with a high average obliquity (~ 35°), and *stage 2* since ~ 4 Ma ago and into the future with a low average obliquity (~ 25°). The cycles with periods of 125 ka and 1.3 Ma are prominent during both stages.

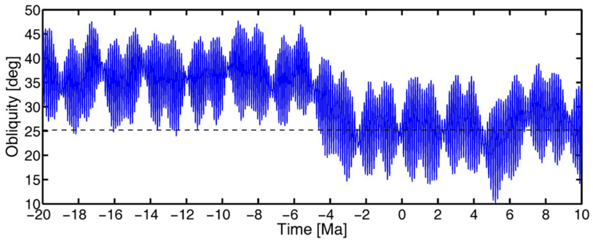

FIG. 1. Martian obliquity from 20 Ma ago until 10 Ma into the future [2].

The simulation starts at –20 Ma with an initial ice layer of 19 meters thickness on the entire Martian surface as initial condition. The numerical time-step is 0.02 a (≈ 7 sols), and the first 10 Ma are interpreted as spin-up. Following Greve and others [1], the dust insulation factor is chosen as $E_0 = 0.1$.

Results: The maximum ice thickness and its position on the planet from –10 Ma until +10 Ma are shown in Fig. 2. The simulation produces a variable glaciation with two distinctly different regimes. Stage 1, the period of high average obliquity prior to –4 Ma, is characterized by ice thicknesses less than 400 m and a very mobile glaciation all over the planet. By contrast, during stage 2, the period of low average obliquity from –4 Ma until today and into the future, the position of maximum thickness changes much less rapidly, flip-flops between the poles only and remains at the south pole from –0.6 Ma on. The polar ice deposits grow almost monotonically to their present-day thicknesses and beyond that, only interrupted by episodic, moderate shrinking events at times when maximum amplitudes of the main obliquity cycle of 125 ka occur (most pronounced at approximately –3.2, +3.0, +7.6 and +8.8 Ma; the +7.6 Ma event is marked by a blue circle in Fig. 2). During such episodes, the polar deposits continue in thin, very mobile ice deposits that extend into the mid and sometimes even into the low latitudes. The latter result agrees with the findings by Head and others [7] who report evidence for past "ice ages" on Mars during periods when the obliquity regularly exceeded 30°.

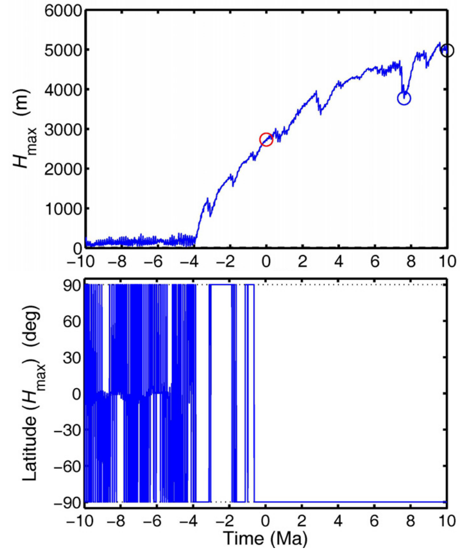

FIG. 2. Top: Maximum thickness H_{max} of the simulated water ice deposits as a function of time. Bottom: Latitude where the maximum ice thickness occurs as a function of time.

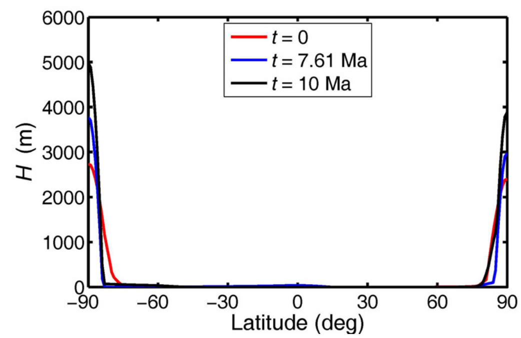

FIG. 3. Ice thickness H for three selected time slices (which correspond to the marks in Fig. 2).

For the three times indicated by circles in Fig. 2, the ice thickness distributions are shown in Fig. 3. For the present ($t = 0$), a very good agreement with the observed polar deposits is achieved (volumes match within ~ 3% and thicknesses at the poles within ~ 20%). The simulated deposits at +7.61 Ma show a metres-thick ice ring between 52°S and 14°N in addition to the polar ice. At +10 Ma, ice thicknesses reach ~ 5 km at the south pole and ~ 4 km at the north pole.

References: [1] Greve R., B. Grieger and O. J. Stenzel (2010) *Planet. Space Sci. 58,* 931–940. [2] Laskar J. and 5 others (2004) *Icarus 170,* 343–364. [3] Lewis S. R. and 8 others (1999) *J. Geophys. Res. 104,* 24177–24194. [4] Murray F. W. (1967) *J. Appl. Meteorol. 6,* 203–204. [5] Ingersoll A. P. (1970) *Science 168,* 972–973. [6] Sears D. W. G. and S. R. Moore (2005) *Geophys. Res. Lett. 32,* L16202. [7] Head J. W. and 4 others (2003) *Nature 426,* 797–802.