ICE FLOW, ISOSTASY AND GRAVITY ANOMALY OF THE PERMANENT NORTH POLAR H_2O ICE CAP OF MARS. R. Greve, *Institut für Mechanik III, Technische Universität Darmstadt, D-64289 Darmstadt, Germany (greve@mechanik.tu-darmstadt.de)*, V. Klemann, *Institut für Planetologie, Universität Münster, D-48149 Münster, Germany*, D. Wolf, *Kinematik und Dynamik der Erde, GeoForschungsZentrum Potsdam, D-14473 Potsdam, Germany*.

Modelling Approach: The flow of the permanent north polar H_2O ice cap of Mars and the isostatic depression of the underlying bedrock are investigated with the 3-d dynamic/thermodynamic ice-sheet model SICOPOLIS [1] coupled to a two-layer visco-elastic model for the lithosphere/mantle system [2,3]. SICOPOLIS describes the ice as a density-preserving, heat-conducting power-law fluid with thermo-mechanical coupling due to the strong temperature dependence of the ice viscosity [4], and 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 heat flux from below into the ice body. However, owing to the now well-known surface topography [5] on the one hand, but the shortage of information about the surface mass balance on the other, here the inverse strategy of prescribing the topography and computing the surface mass balance required to sustain the topography is pursuited [4]. Following further the approach of [4], we use a conceptional, paraboloid-like ice cap, growing and shrinking between the present minimum extent within $80.5^\circ\mathrm{N}$ and an assumed past maximum extent southward to $75^\circ\mathrm{N}$ with a period of 1.3 Myr (first modulation of obliquity cycle), vary the surface temperature with the same period between its measured present distribution and a $30^\circ\mathrm{C}$ warming coinciding with the maximum ice extent, and apply a geothermal heat flux of $35\,\mathrm{mW\,m^{-2}}$.

The lithosphere/mantle model displace comprises an elastic lithosphere of constant thickness, underlain by a Maxwell-viscoelastic half-space mantle. Both layers are treated as incompressible, and we apply terrestrial standard values for the rheological parameters: density of the lithosphere and of the mantle $\rho_\mathrm{l} = \rho_\mathrm{m} = 3380\,\mathrm{kg\,m^{-3}}$, shear modulus of the lithosphere $\mu_\mathrm{l} = 64\,\mathrm{GPa}$, shear modulus of the mantle $\mu_\mathrm{m} = 145\,\mathrm{GPa}$, viscosity of the mantle $\eta_\mathrm{m} = 10^{21}\,\mathrm{Pa\,s}$ [3]. The thickness of the lithosphere, H_l, which is the most crucial parameter of the lithosphere/mantle system, is varied between 50 and 400 km. The field equations of displacement, stress and gravity are solved in the Hankel-wavenumber, k, and Laplace-frequency, s, domain, where they are simply a system of ordinary differential equations in the remaining vertical coordinate, z, and the results are transformed back to the space-time domain by computing the inverse Laplace and Hankel transformations.

Simulations: We carry out four simulations with the coupled SICOPOLIS/displace model and lithosphere thicknesses of 50, 100, 200 and 400 km, referred to as displace (50, 100, 200, 400 km). Another simulation assumes a simple local-lithosphere-relaxing-asthenosphere (LLRA) model with an isostatic time lag of 3 kyr [6], representing maximum isostatic response, and the other extreme of no isostasy at all is investigated by a further simulation with rigid lithosphere. The model time is from 6.5 Myr before present until today, so that five 1.3-Myr cycles are covered and the results are not influenced by arbitrary initial conditions. Here only results for the simulated present ice cap will be discussed.

Results: Table 1 shows the vertical lithosphere displacements (positive downward) in the centre of the ice cap at the north pole, w_max, and at the ice margin at $80.5^\circ\mathrm{N}$, w_marg, as well as the maximum surface-accumulation rate, S_max, for the six simulations. Evidently, w_max is decreasing in the order of the six simulations due to the increasing lithosphere rigidity, whereas w_marg has a maximum for $H_\mathrm{l} = 100\,\mathrm{km}$ due to the counteracting increasing non-local response of the lithosphere. As discussed extensively by [4], surface accumulation and ablation rates are likely of the order of $0.1\,\mathrm{mm\,w.e.\,yr^{-1}}$ (w.e.: water equivalent), which, by comparison with the S_max values of Table 1, favours the last three simulations, that is, $H_\mathrm{l} \geq 200\,\mathrm{km}$. This finding is in good agreement with independent estimates of the lithosphere thickness [7].

Simulation	w_max [m]	w_marg [m]	S_max [$\mathrm{mm\,w.e.\,yr^{-1}}$]
LLRA	1142	0	4.246
displace (50 km)	848	158	1.810
displace (100 km)	568	171	0.734
displace (200 km)	278	137	0.333
displace (400 km)	124	80	0.268
Rigid lithosphere	0	0	0.244

Table 1: Results of simulations for the present ice cap. w_max is maximum vertical lithosphere displacement (taken at the north pole), w_marg vertical lithosphere displacement at the ice margin, S_max maximum surface-accumulation rate.

In Fig. 1 the velocity field for simulation displace (200 km) is depicted for a vertical transect across the north pole. As it is expected, the ice flow is downward and outward in the inner accumulation zone, and points towards the surface in the outer ablation zone. Flow velocities are of the order of some millimetres per year, so that the present ice cap is virtually stagnant, as was already found by [4]. The reason for the conspicuous upward-pointing velocity vectors close to the base is that the modelled ice cap was larger in the past, so that the lithosphere still experiences some uplift which raises the basal ice. Compared to a fully local compensation of the ice load like in simulation LLRA, the bedrock depression is rather small, but the increase in ice thickness is still sufficient to produce a ca. 50% larger ice flow as in the case of a rigid lithosphere.

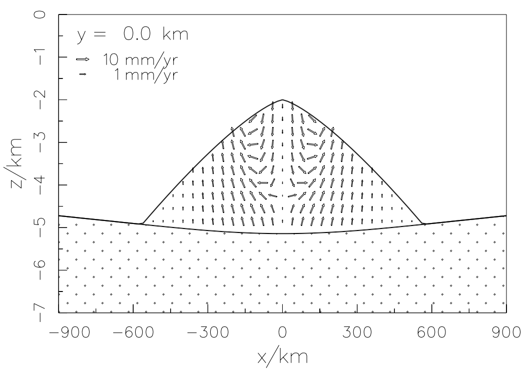

Figure 1: Transect across the north pole at $x = y = 0$ (x, y span the horizontal plane) for present ice cap of simulation displace (200 km): Topography of ice and lithosphere surface, ice-velocity field.

The computed spectrum of relaxation times of the lithosphere/mantle system for simulation displace (200 km) given in Fig. 2 shows two different modes, M0 and L0, due to the two-layered model. M0 is the buoyancy mode, mainly responsible for vertical displacements of the lithosphere, whereas the viscoelastic mode L0 is less significant [2,3]. It is most important that the relaxation times in both modes for any wavenumber (inverse size) of the ice load do not exceed 10 kyr, whereas the period for changes in the ice extent is 1.3 Myr in the simulation. Significant changes on shorter time scales were found to be very unlikely [4]. As a consequence, the lithosphere/mantle system is nearly in steady state with the ice load at any time of the cycle. Therefore, the viscoelastic properties of the mantle (ρ_m, μ_m, η_m) do not influence the results, and the only important parameters of the ground model are the lithosphere thickness, H_l, and, to a lesser extent, the density, ρ_l, and the shear modulus, μ_l, of the lithosphere. This finding, even though demonstrated only for simulation displace (200 km), remains equally valid for the other simulations.

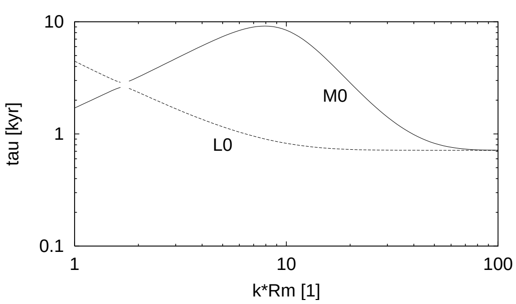

Figure 2: Relaxation time, τ, vs. Hankel wavenumber of the ice load, k, normalized with the mean Martian radius, R_m, for simulation displace (200 km).

The ice cap and the corresponding isostatic depression of the ground go along with a gravity anomaly, which can in principle be detected by orbiting space probes. Fig. 3 displays the vertical displacement of the geoid, h_g (positive upward), for the four simulations displace (50, 100, 200, 400 km). As the signal vanishes in the limit case of fully local isostatic compensation like in simulation LLRA, it is the more pronounced the less isostatic depression takes place, that is, the thicker the lithosphere is. For 200 km thickness, the geoid displacement is 45 m in the centre of the ice cap (north pole), reaches zero approximately at the ice margin, and takes negative values of up to -5.5 m in the adjacent ice-free region due to the non-local lithosphere depression. Compared to the gravity anomalies of major features of the Martian surface, such as the Tharsis rise with $h_\mathrm{g} \sim 1400$ m and the Argyre basin with $h_\mathrm{g} \sim -400$ m [8], the signal of the ice cap is rather small, but should still be discernible by satellite gravimetry.

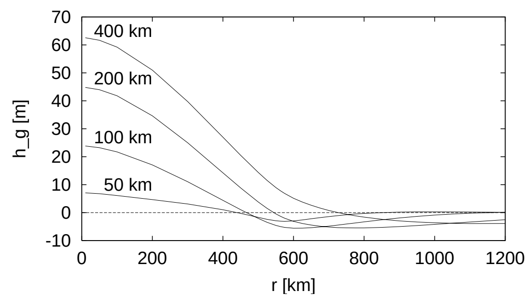

Figure 3: Vertical geoid displacement, h_g, vs. radial distance from the north pole, r, for simulations displace (50, 100, 200, 400 km).

Conclusions: Comparison of the simulated flow of the north polar permanent ice cap with accumulation rates feasible from the point of view of atmosphere physics indicates a thickness of the Martian lithosphere of at least 200 km. For 200 km, the isostatic depression in the centre of the ice cap is approximately 25% of the fully local isostatic compensation, and maximum ice-flow velocities are of the order of some millimetres per year. Relaxation times for the lithosphere/mantle system are less than 10 kyr, much smaller than significant changes of the ice load, so that the ground depression follows the ice-load history in a quasi-static fashion. The vertical geoid displacement due to the combined ice load and ground depression is of the order of some 10 m.

References: [1] Greve R. (1997) *J. Climate*, **10** (5), 901-918. [2] Wolf D. (1985) *J. Geophys.*, **57**, 106-117. [3] Klemann V. and Wolf D. (1999) *Geophys. J. Int.*, **139**, 216-226. [4] Greve R. (2000) *Icarus*, Mars Polar Science Special Issue (in press). [5] Zuber M. T. et al. (1998) *Science*, **282**, 2053-2060. [6] LeMeur E. and Huybrechts P. (1996) *Ann. Glaciol.*, **23**, 309-317. [7] Schubert G. et al. (1992) in *Mars*, eds. H. H. Kieffer et al., Univ. of Arizona Press, Tucson, 147-183. [8] Esposito P. et al. (1992) in *Mars*, eds. H. H. Kieffer et al., Univ. of Arizona Press, Tucson, 209-248.