MODELLING WATER FLOW WITH BEDLOAD ON THE SURFACE OF MARS. A. S. Bargery and L. Wilson, Environmental Science Department, Lancaster University, Lancaster, LA1 4YQ, UK. (a.bargery@lancaster.ac.uk).

Introduction: Fracture systems on the surface of Mars (e.g. Olympica Fossae, West Elysium Graben, Memnonia Fossae, Cerberus Fossae) are thought to have been formed by the intrusion of shallow dykes and sills into the cryosphere [1]. An injection of magma into an ice-rich layer would melt the ice locally and collapse would provide a means of escape to the surface for the melted cryospheric ice. The escape may include water from an aquifer existing directly beneath the cryosphere.

Although the mechanisms for cracking the cryosphere [1] and water release through these fractures [2] have been addressed, there is a need for studies into the behaviour of the resulting fluid flows on the surface, with estimation of how far a flow of water could travel before it froze completely and came to a standstill. The speed of the flood would also have implications for the sediment load that it could carry and its erosive power.

The study of *Wilson and Head* [3] includes models for both water release through cryosphere fractures and subsequent water flow on the surface. Although their model considers both the thermodynamics and fluid mechanics involved, the entire evolution of a solid-liquid mixture is not considered (for example with sediment or after ice crystals begin to form). Their model describes the water evaporation rate, which is one of the main controls, but here we also consider the simultaneous effects of water freezing. This is important because the average global surface temperature on Mars is about 210K, so that a flood of water would cool much more rapidly than on Earth. When ice crystals begin to form, the bulk viscosity increases and the fluid becomes non-Newtonian.

We propose that the flow becomes a solid-liquid mixture with entrainment of sediment and formation of ice. Generally, suspensions are non-Newtonian fluids exhibiting a yield stress which needs to be overcome for a flow to take place. A number of rheometrical studies of clay-water mixtures have been made [e.g., 4, 5, 6] but essentially for moderate solid fractions or at high shear rates (say, $>10\ s^{-1}$). In general, a Bingham model has been fitted to simple shear behaviour data. However, tests made in a wide shear rate range show that these fluids are shear thinning; i.e., their apparent viscosity decreases with shear rate, and a Herschel-Bulkley model [7], for example, is more appropriate [6]. For the moment we model flows as the most simple non-Newtonian fluid, a Bingham fluid and advocate the use of the empirical equations derived by *Malin* [8] and *Pinkerton and Stevenson* [9]. We assume that the groundwater and cryosphere ice are pure H_2O and do not contain any salts. The model of *Wilson and Head* [3] includes a relationship between the fissure width and the flow depth. Our model uses the flow depth as an input, the values of which are estimates from numerous previous studies of outflow channels on Mars. We model a turbulent, water-dominated phase and a prolonged ice-covered, cooling phase. The emphasis is on the heat balance within the ice-dominated phase of possibly many years duration, before the entire system freezes.

Cooling effect of bedload: We assume that erosion of both the fracture walls and the surface topography results in a sediment load that includes ice, which comprises up to 20 % of the cryosphere at less than 1 km depth [10]. Our model assumes that the fraction of eroded material that is ice is everywhere 20 %. If the temperature of the water is above 273 K, we assume that all the ice is melted by the water. By considering the heating of the entrained bedload rock and ice to 273 K, the heat required to melt the ice, and the subsequent heating up to a new equilibrium water temperature of the entrained rock and melted ice, we find

$$\delta\,\theta_w = (\theta_w (d_r\,\rho_r\,c_r + 1) + 273\,(d_i\,\rho_i\,c_i - 1) - \theta_c (d_r\,\rho_r\,c_r + d_i\,\rho_i\,c_i) - d_i\rho_i L_i\,) / (2\,c_w\,d_w\,\rho_w + d_r\,\rho_r\,c_r) \qquad (1)$$

where θ_w is the temperature of the water (decreasing all the time), θ_c is the initial temperature of the eroded material, L_i is the latent heat of fusion of ice, c_w, c_r, c_i are specific heats, ρ_w, ρ_r, ρ_i are densities, and d_w, d_r, d_i are thicknesses, of water, rock and ice, respectively. It is uncertain at what point erosion ceases, so we assume that erosion would continue until 100 % of the flow was solid (including ice crystals). Alternatively we can chose a value for the limit of the capacity of the flow to carry sediment.

When the water temperature reaches 273 K, entrained ice does not melt, and the temperature of the water does not decrease until all of its latent heat of fusion has been lost to warming the entrained rock to 273 K. This loss of latent heat from the water causes further ice crystal formation. The rate of formation of ice crystals due to this heat loss from the water can be calculated by consideration of the latent heat of freezing. The volume fraction of ice produced from this heat loss is

$$v_{fi} = (273 - \theta_c)(\,v_i\,\rho_i\,c_i + v_r\,\rho_r\,c_r)/(\,\rho_i\,L_i) \qquad (2)$$

The increase in ice fraction is at the expense of the fraction of water, but ice is less dense than water so the

decrease in water fraction is slightly less than the increase in ice fraction.

In the model, we assume that the erosion rate, *e*, is constant over the length, width and course of the flow. We use a test value for the erosion rate of 1 mm per second. The total volume of water in the flow is *hWut* where *t* is the time after release of the water on to the surface and *h* is the depth of the water, and the total volume of sediment at any one point in time is *WLet*. The volume fraction of bedload, v_b, is the ratio of these two volumes; hence, if we assume constant flow velocity,

$$v_b = et / h \qquad (3)$$

In each time increment, the bedload added to the system cools the flow by the calculated amount (from equation (1)). This removes the need to calculate the conductive heat loss rate from the water to the regolith (as these are equivalent). The volume fractions are then regarded as volume fraction rates and there will be an accumulated fraction of bedload rock.

Evaluation: In a first model we deal with water flowing in an environment at temperatures well below the triple point, with an upper surface exposed to the atmosphere and this provides a mechanism for estimating the maximum travel distance. The model proposed by *Allison and Clifford* [11] for ice-covered water volcanism on Ganymede has been modified for Mars and the original model for ice-water flows on Mars of *Wallace and Sagan* [12] has been reconsidered. What we consider to be the best aspects of both (with respect to Mars) have been combined and extended upon within a second model.

The 'outflow channels' of Mars may have formed as a consequence of more than a single type of event. The sources of the outflow channels vary from overflowing from the Chasmata and large gorges, such as Valles Marineris, to regions of chaos and also graben associated with magmatic intrusion and collapse features [e.g., 3]. Therefore, the resulting flows may differ in gross physical characteristics, such as the duration of the flow. Catastrophic outbursts may have short durations and will be water dominated. These flows may deposit the semi-circular 'moraines' seen, for example, in the high-resolution MOC images E2100112 and R0900475 from southern Elysium Planitia [13]. Long duration flows filling topographic depressions are likely to be ice cover dominated. Two separate models are required, which will be formulated by modification of the analysis described here. This may be accomplished by including the energy balance with or without a temporally permanent and spatially continuous ice cover.

A range of possible applications are envisaged for our existing model: 1) to study water flowing in environments where the temperature is well below the triple point, with an upper surface exposed to an atmosphere [13]; 2) to provide a method for predicting theoretically the maximum flow length; 3) to study flowing water released from any melting ice layer (e.g., an ice cap or an ice sheet); 4) to model sub-glacial flow and subsequent release of groundwater during sub-glacial eruptions; 5) to evolution of cryosphere fractures [1]; 6) to investigating possible mechanisms of rapid flooding of the northern lowlands [3, 14, 15].

Current limitations of the model include assuming the groundwater is pure H_2O and does not contain any salts, and the inherent modelling of water as a Newtonian fluid and a water/ice mixture as a Bingham fluid, but these simplifications can be systematically remedied.

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