3D Modeling of Landscape-Modifying Processes on the Galilean Satellites

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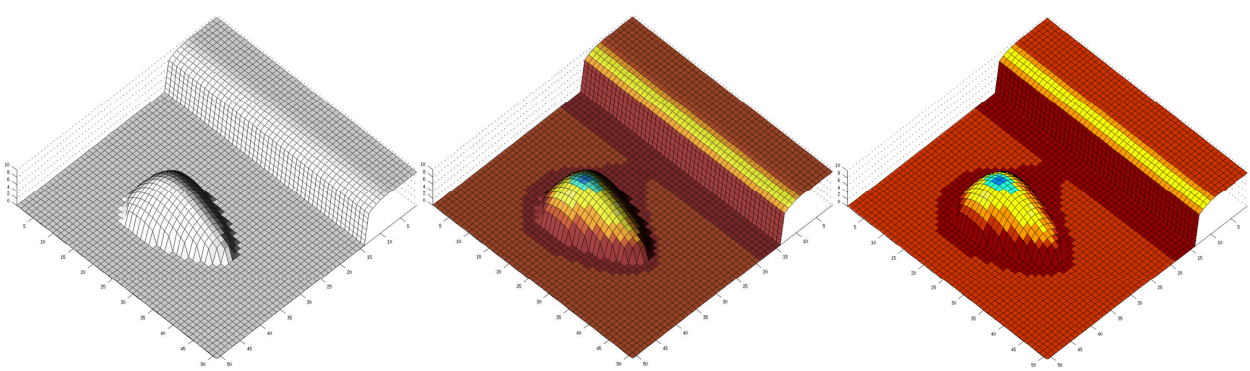
We evaluate the sequence and extent of volatile redistribution and erosional processes that have shaped the topography of the Galilean satellites by applying a physics-based 3-D landform evolution model to Galilean satellite erosional landforms. Model simulations of icy satellite landform erosion and volatile redistribution are guided and constrained by iteratively comparing model results with the general statistics of erosional landform classes derived from DEMs. In some cases, DEMs of pristine landforms (e.g. fresh craters, fault blocks) are used as the "initial conditions" in model runs. The model includes the following coupled processes: (a) sublimation and re-condensation of surface volatiles; (b) subsurface heat conduction; (c) direct solar heating and radiative cooling; (d) indirect solar and thermal radiation from other surfaces (reflected and emitted); (e) shadowing by topography; (f) subsurface sublimation/condensation and vapor diffusion; (g) development of a "sublimation lag" of non-volatile material; and (h) disaggregation and downward sloughing of surficial material.

In comparison to previous topographic thermal models for airless icy bodies [Spencer, 1987; Colwell *et al*., 1990; Vasavada *et al*. 1999], our model includes several important additions and improvements:

(a) Dynamic topography - We model the changes in landform predicted to result from sublimation/condensation and gravitational mass wasting. Previous studies have speculated about scenarios for how the surface might evolve as a result of these processes, but given the many potential feedbacks between topography, temperature, albedo, the amount of ice present, and the stability of slopes, it is very difficult to predict without using a model that can track the time evolution of these variables in a self-consistent way.

(b) Shadowing and radiative scattering/emission for asymmetric surfaces with non-zero thermal inertia - Previous models that included these effects were restricted to symmetric geometries such as idealized craters and trenches, and assumed surface temperatures were always in radiative equilibrium. Our modeling uses realistic topography that includes exposures of high thermal inertia material, which can respond differently to being shadowed during the morning versus the afternoon.

(c) Treatment of temperature-dependent thermal properties - The thermal conductivity of particulate minerals in vacuum has the functional form $k = A+BT^3$ [Cremers and Birkebak, 1971], while the conductivity of solid water ice is $k_{ice} = 567/T$ [Squyres *et al.*, 1985], and for both materials the heat capacity increases linearly with temperature over the range found on the Galilean satellites (80-160K). These dependencies can cause significant diurnal variations in the efficiency of subsurface heat conduction.



Model Components and Implementation

Sublimation: This component is based on a new 3-D finite-element sublimation model developed to study the long-term evolution of ice deposits on Mars. The sublimation rate of an ice surface is determined almost entirely by its temperature, and on planetary bodies with little or no atmosphere, the surface temperature is largely controlled by the topography (e.g. slope, azimuth, and shadowing). But the temperature varies on a much more rapid timescale than the topography, so one of the key modeling considerations is how to link the calculations of z(t) and T(t). In order to accurately capture the variations in surface temperatures we need to use a time step less than 1/10 of a diurnal cycle, whereas the timescale for significant topographic changes is likely to be on the order of $10^{5\text{-}7}$ years. For the case of the Galilean satellites, we can take advantage of the fact that they do not experience significant seasonal variations in their insolation patterns (obliquity < 1 deg.), so that one day is pretty much like any other. That means that we can simply multiply the results obtained for one day by the number of days required to exceed some criteria of minimum change in the topography (or surface albedo).

The model surface is a rectangular grid of points with fixed horizontal spacing ($\Delta x = \Delta y$) and variable height (z). At each time step, the vertical displacement of each grid point due to sublimation is given by

$$\Delta z_{sub} = \Delta t * (q_{sub} / \cos\theta) * (1/\rho_{ice})$$

where q_{sub} is the sublimation rate, θ is the average local slope, and ρ_{ice} is the ice density. The factor of $(1/\cos\theta)$ accounts for the fact that sublimation occurs perpendicular to the surface slope. For numerical stability, we have implemented this finite difference equation using the Lax method which has the additional advantage of being robust to the slope discontinuities that tend to develop.

Subsurface Heat Conduction: Our model includes one-dimensional (vertical) heat conduction below each surface grid point. Rather than prescribing the thickness of each layer, we prescribe the depth of the interface(s) between regions with different thermal properties, and use a fixed number of layers within each region. We then use a semi-implicit finite-difference scheme, which is unconditionally stable and accurate to second-order, to solve the diffusion equation within each region and match the heat flux across each interface. This model makes it much easier to account for changes in layer thickness or composition, and allows us to use the largest possible time steps without sacrificing vertical resolution.

Subsurface Sublimation and Lag Deposits: Thin layers of non-volatile material on the surface can increase sublimation by lowering the albedo, but thicker layers can dramatically reduce sublimation: by damping the subsurface thermal wave and by physically impeding the vapor flux. The thermal effects of the lag are easily simulated by adjusting the thermal properties of the upper model layers. The vapor diffusion effects are modeled by replacing the surface ice sublimation term with an expression for subsurface sublimation: $q_{sub} = F_v = D_k\ (\varepsilon/\tau)\ N_{ice}/z_{ice}$, where N_{ice} is the equilibrium vapor density at the depth of the ice (z_{ice}), D_K is the Knudsen diffusion coefficient, ε is porosity, τ is the tortuosity of the diffusive path (e.g., Moore *et al*. [1996]).

References

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