**Introduction:** Recent observations of Martian ice-related features and theoretical work on the dynamical history of Mars have illuminated how obliquity variations led to the periodic deposition of ice-rich layers, ranging from 10 meters to 1 kilometer thick, on the surface [1, 2]. Because icy layers are thought to be globally distributed and have formed throughout Martian history, the impact crater record provides a powerful tool to investigate the properties and evolution of these layers. To do so, however, we must first understand how the presence of such layers affects the impact cratering process and the final crater morphology. Furthermore, icy layers have been proposed as a contributing factor to the layered ejecta morphology. Furthermore, icy layers have been proposed as a contributing factor to the layered ejecta structures that are found around the majority of well-preserved Martian impact craters [e.g. 3].

Our goal is to use numerical simulations to (1) describe and quantify the effects of icy surface and subsurface layers on crater formation processes and (2) use these results to probe the properties and history of such layers on Mars. Here, we discuss the numerical method and present results for cratering in targets with variations of a single layer of ice.

**Method:** Cratering simulations are conducted using CTH, an Eulerian shock physics code developed at Sandia National Laboratories [4]. CTH utilizes adaptive mesh refinement [5], meaning smaller-scale features (such as fine layers at the surface, the ejecta curtain, and the shock wave) can be resolved without expending too much computational energy elsewhere in the problem domain. Shock physics codes work by solving three parts: (1) the equations for the conservation of mass, momentum, and energy, (2) an equation of state, and (3) a constitutive (or strength) model. Both (2) and (3) are material dependent and the validity of a numerical calculation relies strongly on the quality of the equation of state and constitutive model used.

**Strength Model** – The strength model describes the response of a material to deviatoric stresses. Since existing strength models in CTH for geologic materials were either too simple or too specialized, we have recently implemented the rock strength model of Collins et al. [6] into the code and have validated it over a large size range (from laboratory to planetary scale) [7]. The model degrades shear strength as a function of pressure, temperature, and damage; tensile strength is degraded as a function of temperature and damage. Damage is a dimensionless quantity between 0 (completely intact material) and 1 (completely fragmented material) that represents the amount of fracturing that a material has undergone. Shear and tensile damage are tracked separately, and under tensile failure, void space is added to a computational cell to simulate fracture. Acoustic fluidization, used to assist complex crater collapse, is included as an option in the calculation [8, 9]. We choose strength parameters by fitting to laboratory test data and adjusting for scale and strain rate effects as described in Senft and Stewart [7]. For basalt, we used the fits given in [10] and data from [11]. For ice, we used data from [12, 13, 14]. Acoustic fluidization parameters for basalt were fitted to depth versus diameter and rim height versus diameter trends for well-preserved Martian craters as measured by [15]. As a first order approximation, ice is assumed to have the same acoustic fluidization parameters as basalt.

**Equation of State (EOS) –** CTH uses Sesame tables, which are gridded equations of state to allow for fast look-up and interpolation of the EOS. Pressure, energy, and entropy are tabulated on a density-temperature grid. For the basalt equation of state, we use the Sesame table for basalt developed by Kerley; this table is a density scaled version of the EOS for α-quartz and includes two solid phases, a liquid phase with dissociation, and vapor [16]. We constructed a new Sesame table for H2O. This table includes three solid phases (ice Ih, VI, and VII), liquid, and vapor. The EOS of the phases and phase boundaries are experimentally determined [17, 18, 19, 20], and the tabulated EOS reproduces measured shock Hugoniot.

**Description of Simulations –** We conducted simulations of the vertical impact of 200-m, 400-m, and 800-m diameter basaltic asteroids onto the Martian surface at a velocity of 10 km/s. These impact conditions create craters with nominal final rim-to-rim diameters of 4.9, 8.5, and 16.0 km, respectively, in a hard rock target [21]. Simulations of targets with varying pure ice layer thicknesses and depths to the ice layer were conducted and compared to simulations in homogeneous basalt targets. All calculations were initialized with lithostatic pressure and a geotherm of 15 K/Km. The pressure and temperature at the surface were assumed to be 6 mbar and 210 K, respectively. All calculations were conducted in 2D under cylindrical symmetry, and the maximum resolution was 40 cells across the diameter of the projectile.

**Results:** A range of icy layer effects is illustrated using time series from simulations of a 200-m diameter projectile onto the Martian surface for different target configurations (Fig. 1). In Fig. 1, brown represents
Figure 1: Impact of a 200 m diameter projectile onto the Martian surface vertically at 10 km/s for different target configurations. Cross sections of cylindrically symmetric calculations are shown, where brown represents basalt and blue represents H$_2$O. In A the target is homogenous basalt; in B there is a 100 m surface ice layer; in C there is a 100 m ice layer buried at 200 m depth; in D there is a 200 m surface ice layer; in E there is a 200 m ice layer buried at 200 m depth; in F there is a 400 m surface ice layer; and in D there is a 400 m ice layer buried at 100 m depth. Time increases downwards and the scale is the same in all panels. Note that a time of “end” refers to the last simulation output. For simulations A through E this is when the material has stopped moving appreciably; however, for runs F through G, material is still moving (the runs were not completed due to time constraints). In F, ice is still flowing into the crater, and in G, ice is still flowing outwards (away from the crater center).
basalt, blue represents \( \text{H}_2\text{O} \), and mixed cells are colored by the material with the largest mass fraction. Small volumes of water and basalt are vaporized, but vapor was removed from the simulation for computational efficiency and will be studied in future work.

In Fig. 1a, the target is homogenous basalt (no ice layer). Crater formation proceeds as expected, with the ejecta blanket forming a smooth inverted cone that sweeps outward across the surface (20 s) and the formation of a bowl shaped transient cavity whose walls collapse slightly, shallowing the crater (“end”).

In Fig. 1b, a thin (100 m) surface ice layer is added. Crater formation proceeds as in the homogenous case, but when the basalt ejecta is laid down near the rim, it compresses the ice layer underneath. This leads to horizontal, non-ballistic motion of the near-surface ice, which ultimately thins the icy layer near the rim and thicken it at greater distances.

When the thickness of the surface ice layer is increased to 200 m (Fig. 1d), the proportion of ice relative to basalt in the ejecta blanket increases. The ejection of ice at higher angles than basalt creates a curved profile to the ejecta blanket, which will change the final ejecta distribution.

If the thickness of the surface ice layer is large enough (400 m, Fig. 1f), than the ejecta blanket of the icy layer separates (or detaches from) the basalt ejecta blanket (2 s). This two-layer ejecta blanket structure has also been observed in simulations of marine targets [22, 23]. Also, the icy rim appears to be unstable, and thus flows into the crater at late times (“end”).

Note that tensile failure produces the observed void spaces within the icy layers in the simulations shown in Figures 1e, 1f, and 1g.

Burying the ice layer under a basalt layer produces further morphological variations. Figs. 1c and 1e show a 100 and 200 m ice layer, respectively, buried under a 200 m thick basalt layer. In both cases, the top basalt layer tears away from the underlying surface (along the weaker ice layer) at early times (2 s). The ejecta trajectories are modified by wave reflections between the layers. The high ejection angles result in a hinge-like evolution of the ejecta curtain instead of the standard inverted cone structure. The hinge area then collapses back towards the crater cavity (20 s). Finally, as the hinge slumps, it squeezes the ice layer, resulting in a late-stage icy extrusion into the crater. This ice behaves in a fluid manner because it is warm (thermal softening); however, it is largely still ice (unmelted) (see Fig. 2, which displays the final temperature contours around the crater for the simulation shown in Fig. 1e).

If the thickness of a buried icy layer becomes large enough (Fig. 1g: a 400 m thick ice layer buried under a 100 m basalt layer), then the actual crater (in the underlying basalt layer) becomes very small and the amount of ice being extruded into the crater at late times becomes very large. As the ice being extruded from all sides of the crater meets, it creates a large central uplift which collapses back down in on itself and flows outward. This outward flowing ice is near the melting point.

Simulations of the formation of 8.5 and 16.0-km rim diameter craters in the same targets show similar features, but their occurrence depends on the thickness of the ice layer and its burial depth relative to the crater size.

**Comparison with Observations:** Our simulations suggest that several of the features associated with Martian impact craters may be a result of surface or near subsurface icy layers, including:

- **“Dewatering” features:** Tornabene et al. [24] has recently documented flow features associated with young impact craters of a large size range (~3 to 60 km) and suggest that they are a result of the flow of water into the crater cavity. Our simulations (Figs. 1c, 1e, 1f, and 1g) show warm, thermally weakened ice flowing away from the crater rim and into the crater (as late-stage icy extrusions), supporting this interpretation.

- **Rim Moats:** Our simulations show non-ballistic, horizontal flow of ice away from the crater rim, which may produce observed circum-rim moats (Figs. 1b and 1d) [15].

- **Layered Ejecta Structures:** Non-ballistic trajectories modify the radial distribution of ejecta from a simple power law. Terminal ramparts cannot be directly observed in the simulations because the scale is too small and the physics of debris flows are different from the physics of large-scale impact cratering events. However, simulations can provide the initial conditions (flow velocities, ice/rock ratios, and temperatures of the ejecta) for debris flow models. Note that complete interpretation of layered ejecta blankets must include the effects of the atmosphere.

- **Lack of Secondary Craters:** Boyce and Mougins-Mark [24] observe a lack of secondary craters around some double layer ejecta craters. Our simulations show that when there is a buried ice layer (Figures 1c, 1f, and 1g), the ejecta flow can be somewhat impeded, leading to most of the ejecta being deposited close to the crater rim. This could lead to a deficit of secondary craters.

- **Paleolakes:** A number of possible paleolakes in Martian craters have been identified [e.g. 26]. Our simulations show warm ice ponded on crater floors during the impact event (Figures 1c, 1e, and 1g).
**Central Pit Formation:** The thermal evolution of the liquid water and ice deposits in the crater floor can be used to study the possible formation of central pits by dewatering [27].

**Natural Variations in Crater Morphologies:** Martian impact craters display large variations in depths, rim heights, and amounts of ejected and uplifted material versus crater diameter [e.g. 15]. Our simulations produce a large range of these measures. Note, however, that some of the observed features need not be due to ice layers necessarily, but may be attributed to target strength variations in general.

**Conclusions:** We have performed simulations of Martian impact cratering events into layered icy terrains in order to understand the effect that such layers can have on impact cratering processes and the final crater morphology. The effects can be significant and may explain many of the features seen around Martian impact craters.


![Figure 2: Cross section of the final crater, with temperature contours, for the simulation shown in Figure 1e. The target has a 200-m thick ice layer buried by 200-m of basalt. Blue represents H2O and brown represents basalt.](3309.pdf)