Martian post-impact hydrothermal systems incorporating freezing

Charles J. Barnhart a,*, Francis Nimmo a, Bryan J. Travis b
a Department of Earth and Planetary Sciences, University of California, Santa Cruz, CA 95064, USA
b Earth and Environmental Sciences Division, Los Alamos National Laboratory, Los Alamos, NM 87545, USA

ARTICLE INFO

Article history:
Received 11 June 2009
Revised 8 December 2009
Accepted 14 January 2010
Available online 25 January 2010

Keywords:
Impact processes
Mars
Mars, Interior
Mars, Surface
Geological processes

ABSTRACT

We simulate the evolution of post-impact hydrothermal systems within 45 km and 90 km diameter craters on Mars. We focus on the effects of freezing, which alters the permeability structure and fluid flow compared with unfrozen cases. Discharge rates, total discharge and water–rock ratios increase with permeability. Systems with permeabilities of \(10^{-12}\) m² or higher exhibit convection in the hydrosphere, allowing them to derive heat from greater depths. Surface discharges persist for \(\sim 10^3–10^5\) years under freezing surface conditions, with higher permeabilities permitting longer lifetimes. Maximum discharge rates and total discharges range from 0.1 to 10 m³ s⁻¹ and \(10^6\) to \(10^{12}\) m³, respectively, for systems with permeabilities between \(10^{-14}\) and \(10^{-12}\) m². Near-surface water–rock ratios range from <1 for low permeability, frozen cases to \(\sim 10^3\) for high permeabilities and/or unfrozen cases. Propagation of the freezing front radially inwards focuses flow towards the center of the crater resulting in a diagnostic increase in water–rock ratios there. This process may explain the phyllosilicate assemblages observed at some crater central peaks.

© 2010 Elsevier Inc. All rights reserved.

1. Introduction

Atmospheric pressure and temperature conditions at Mars have apparently conspired against the presence of liquid water on the surface for much of its history (Carr, 1996). The surface, however, tells a different story. It records a diverse collection of geochemical and geomorphic signatures left behind by the presence of liquid water (Craddock and Howard, 2002; Mustard et al., 2008). Exploring how various climatic and geological processes permit water on the surface, and tying those processes to observable data, can contribute to our understanding of the hydrological evolution of Mars. One likely process involves the interaction of high velocity bolides with Mars's volatile-rich crust. A kilometer-scale bolide traveling at several kilometers per second delivers enough energy to melt ice, heat subsurface water and drive hydrothermal circulation (Newsom, 1980). This post-impact hydrothermal (PIH) circulation can lead to surface discharge of water and chemical alteration of rock—both of which are potentially detectable.

Here we show how hydrothermal systems initiated by bolide impacts can deliver liquid water to the surface even under current atmospheric conditions. Our analysis focuses on how freezing can affect PIH systems in ways that can be remotely sensed. Specifically, we employ a numerical model to determine geophysical quantities—discharge rate, total discharge, near-surface temperature evolution, and water-to-rock ratios—that govern observable geochemical and geomorphological surface features. We model the evolution of PIH systems as multi-phase, convective flow through a saturated, heterogeneous, porous medium in which both advection and conduction contribute to temporal and spatial changes in temperature and ice content. The code we employ is MAGHNUM (Mars Global Hydrology Numerical Model) (Travis and Rosenberg, 1991; Travis et al., 2003). MAGHNUM solves the governing equations for mass and energy transport in a saturated porous medium. Our models differ from previous efforts (Newsom, 1980; Newsom et al., 1996; Rathbun and Squyres, 2002; Abramov and Kring, 2005) in that we incorporate freezing aquifers.

In this section we provide background on the martian cryosphere, previous Mars-based groundwater studies, and terrestrial PIH systems. In Section 2, we explore theoretical considerations and describe our model setup. In Sections 3 and 4 we present and discuss results, and in Section 5 we summarize our conclusions.

1.1. Are PIH systems possible on ‘cold–dry’ Mars?

PIH systems require water in either solid or liquid form. Mars’s crust likely contains significant quantities (e.g. Clifford, 1993). Energetic neutron counts collected by the Neutron Spectrometer aboard 2001 Mars Odyssey leave little doubt that the near-subsurface of Mars holds water ice, especially in high latitude regions (Boynton et al., 2002; Feldman et al., 2002). Discharge estimates for large outflow channels (e.g. Kasei Vallis), argue for \(\sim 7 \times 10^7\) km³—or up to 500 precipitable (pr) m—of water in the...
past (Carr, 1996). Crater softening studies argue for layers of ice-rich regolith up to 150 m deep in the mid-latitudes, which yields a total ice volume of $10^7 \text{km}^3$ (Squyres and Carr, 1986; Parsons and Nimmo, 2009). The volume of the upper 10 km of the martian crust is roughly $1.4 \times 10^9 \text{km}^3$. Assuming a depth-averaged porosity of 0.01, the crust could hold $\sim 1.4 \times 10^7 \text{km}^3$ at saturation. These volumes are comparable to estimates for the polar caps; $1.2-1.7 \times 10^5 \text{km}^3$ and $2-3 \times 10^5 \text{km}^3$ of ice reside in the northern and southern polar caps, respectively (in total 22–33 pr m) (Smith et al., 1999).

Presently, water is not thermodynamically stable on the surface of Mars. Mean annual surface temperatures range from $\sim 154 \text{K}$ at the poles to $\sim 218 \text{K}$ at the equator ($\pm 5 \text{K}$) (Clifford, 1993). The mean surface pressure is 6.1 mbar with a partial pressure of water $\text{pH}_{2}\text{O} = 10^{-10}$ bars (Carr, 2006). At such low humidities, the frost point, as measured by the Viking Mars Atmospheric Water Detectors (MAWD), is a mere 200 K (Farmer and Doms, 1979).

Surface temperatures and theoretically derived background geothermal fluxes imply a region of subfreezing temperatures (i.e. a cryosphere), that is at least hundreds of meters thick and likely averages a few kilometers (Clifford and Hillel, 1983; Clifford, 1993; Mellon et al., 1997; Travis et al., 2003). Kilometer-scale bolide impacts deliver enough energy to locally eliminate the cryosphere by melting and/or vaporizing the subsurface ice (Wohletz and Sheridan, 1983; Mouginis-Mark, 1987; Newsom et al., 1996). Some fraction of the bolide’s kinetic energy becomes residual heat that can drive hydrothermal flow (Newsom et al., 1996; Rathbun and Squyres, 2002; Abramov and Kring, 2005).

Given the evidence for volatiles within the crater-riddled crust of Mars, PIH systems are likely commonplace. Their capacity to create and preserve detectable geochemical and geomorphic signatures under a range of climate and geological conditions remains an open question and is the focus of this study. Mineral assemblages (Ehmann et al., 2008) and geomorphic features (Moore and Howard, 2005) associated with complex craters motivate our decision to model PIH systems in 45 km and 90 km diameter craters.

1.2. Previous Mars-based hydrothermal studies

Several studies, including simple analytic models and 3-D numerical porous-flow models, have investigated the behavior of PIH systems on Mars. Early work focused on potential association with crater lakes. When in communication with a large subsurface aquifer, energy balance calculations show that crater lakes will remain at least partially unfrozen for millions of years due to sublimation removal of ice from the top and release of latent heat as ice plates to the base of the ice sheet (McKay et al., 1985; McKay and Davis, 1991). Newsom et al. (1996) demonstrated that the additional heat supplied by a post-impact melt sheet extends lake lifetimes by tens of millions of years.

Rathbun and Squyres (2002) was the first study to employ numerical models. Their work showed that PIH systems associated with complex craters 180 km in diameter could support a lake up to 1 km thick. Abramov and Kring (2005) modeled PIH systems associated with 30, 100, and 180 km diameter craters exposed to 5 °C surface temperatures and showed that host rock permeability is the main factor affecting fluid circulation, and that these systems provide hospitable conditions for microbial life for $\sim 100$ kyr. The code that these studies used, HYDROTHERM, does not simulate freezing. Freezing has important effects on the thermohydrologic properties of the aquifer including latent heat release and drastic changes in permeability.

Other sources of heat can drive hydrothermal circulation. Under specialized circumstances, magmatic intrusions may drive enough flow to generate fluvial valleys (Gulick, 1998). Harrison and Grimm (2002) showed that magma chamber-induced hydrothermal circulation and surface discharge increase linearly with permeability and that total discharge volumes span a range comparable to the inferred discharges of valley networks in volcanic provinces on Mars.

Travis et al. (2003) explored the effect of a background geothermal flux on putative sub-cryospheric aquifers, and found that for sufficiently high permeabilities, background geothermal heat flux alone could drive hydrothermal convection. These buoyant convective plumes are capable of thinning the martian permafrost to a thickness of 1 km or less.

1.3. Terrestrial impact structures

On Earth, water saturates the upper crust, and any hypervelocity impact capable of generating a crater will disrupt the local hydrosphere (Oskiniski and Spray, 2001). Hydrothermal systems have been inferred at many terrestrial impact structures including: Haughton (Oskiniski and Spray, 2001), Vredefort (Grieve and Garvin, 1984), Sudbury (Ames et al., 1998), Chicxulub (Kring, 2000), Ries (Newsom et al., 1986), Chesapeake Bay (Sanford, 2005), Manson (McCarville and Crossey, 1994), and Lockne (Sturkell et al., 1998).

Studies of terrestrial impact structures add to our general understanding of their internal structure. For example, Haughton crater—a 23 km diameter structure located on Devon Island, Nunavut, Canada—demonstrates how permeability controls PIH system plumbing, cooling, and surface mineralogical expression. Pipe structures, characterized by pronounced Fe-hydroxide alteration of the carbonate country rocks, are concentrated throughout the faulted annulus on crater wall slumps, but have not been found in the impact brecias or in the central uplifted area (Oskiniski and Spray, 2001). This observation indicates that relatively high permeabilities are concentrated in the pipes and that the brecciated floor has relatively low permeability (Oskiniski and Spray, 2001). Concentric listric faults and interconnecting radial faults provide pathways for hot fluid and steam.

1.4. Mars mineralogical observations

The TES, CRISM and OMEGA spectrometers orbiting Mars have detected a variety of clay alteration minerals (Christensen and Ruff, 2004; Bibring et al., 2006; Mustard et al., 2008). Clay alteration mineralogy depends on both the peak fluid temperature experienced by the rock and the water/rock ($W/R$) ratio (Griffith and Shock, 1997; Schwenzer and Kring, 2008). Recent work that simulates equilibrium chemical reactions shows that higher $W/R$ ratios ($\sim 1000$) occurring at modest temperatures ($< 250 \text{°C}$) will produce nontronite and hematite with lower percentages of chlorite, and that low $W/R$ ($\sim 1-10$) will produce chlorites, smectites, mica, amphibole and garnets (Schwenzer and Kring, 2008, 2009). Moreover, depending on host-rock conditions, serpentine or feldspar may also form (Schwenzer and Kring, 2009). The type and spatial distribution of these phyllosilicates associated with PIH systems will thus be affected by thermal and mass transport characteristics central to our models. We discuss the implications of our results with respect to these detections and geochemical models in Section 4.2.

2. Theory and model setup

Here, we discuss the thermal and hydrological properties of Mars’s crust. Next we use a simple theoretical analysis to illuminate PIH system behavior and to justify model assumptions. Then, we establish the model’s initial conditions and temperature field as prescribed by an analytical calculation of impact shock-heating.
Finally, we present the numerical code we employ and the parameter space explored by this study.

### 2.1. Thermohydrological properties

The following section details various thermal and hydrological parameters important to PIH system evolution. Table 1 lists parameters used in this study.

#### 2.1.1. Porosity

Porosity, ϕ, is given by:

$$\phi = \frac{V_{\text{void}}}{V_{\text{total}}}$$  \hspace{1cm} (1)

where $V_{\text{void}}$ and $V_{\text{total}}$ are the volume of void space and total rock, respectively. The porosity structure of the martian surface is poorly constrained. Measurements are limited to soil samples, yielding higher porosities than expected at depth (Clark et al., 1976; Toon et al., 1980; Clifford and Hilleg, 1983). Of course, the bulk porosity of the regolith is likely to display considerable heterogeneity. Mars images show complex stratigraphy and planet-wide layering (Edgett and Malin, 2004). If the crust is composed of a megaregolith, the hydraulic properties will depend primarily upon the abundance of breccia and the compressional state of fractures (Hanna and Phillips, 2005). To complicate porosity estimates further, impact shock waves may be affected by interactions at material boundaries and, in turn, change the porosity by crushing pores (Pierazzo et al., 2005). For the Moon, seismic properties suggest that there is an exponential decline in porosity with depth with a best fit surface porosity, $\phi_0$, of 0.2 and a decay constant, $H_0$, of 6.5 km (Toksoz, 1979):

$$\phi(z) = \phi_0 e^{-z/H_0}$$  \hspace{1cm} (2)

When scaled to martian gravity, porosity becomes $<0.01$ at $\sim$8.5 km depth and the decay constant, $H_0$, is 2.8 km (Clifford, 1981). This is the formalism used in several previous studies (e.g., Clifford, 1993; Clifford and Parker, 2001), and we adopt it for our work as well.

#### 2.1.2. Permeability

The permeability, $k$, characterizes a porous medium’s resistance to flow through it (Turcotte and Schubert, 2002, 89-2). Previous studies (Rathbun and Squyres, 2002; Abramov and Kring, 2005) show that permeability governs fluid flow in PIH systems. Like porosity, the bulk permeability of the martian regolith is unknown and is likely to exhibit strong spatial variability following an impact. We test a wide range of surface permeabilities, $10^{-16}$, $10^{-14}$, $10^{-12}$, and $10^{-10}$ m$^2$, in an attempt to explore the consequences of reasonable geologic materials. We assume that porosity is primarily in the form of interconnected fractures, allowing us to express permeability through a porosity-dependent cubic flow law (Travis et al., 2003), which we assume can be modified to account for ice content to yield:

$$k(z) = k_0(1 - \sigma)^3 e^{z/z_0}$$  \hspace{1cm} (3)

where $\sigma$ is the fraction of pore or fracture space filled with ice, $H_0$ is the same decay constant used in the porosity function and $k_0$ is the surface permeability (m$^2$)—a free parameter in this study.

The permeability of geologic materials spans many orders of magnitude. The permeability of breccias in known terrestrial impact craters depends on the geologic setting of the impact crater. For example, the Chesapeake Bay impact crater formed in Cenozoic sediments that are generally characterized by sands, silts and clays deposited by marine transgressions and regressions (Sanford, 2005). Based on these materials and expected time ranges required to flush out the original water in the breccia, permeability likely ranges from $10^{-16}$ to $10^{-14}$ m$^2$, but was likely higher at the onset of PIH system development before subsequent compaction (Sanford, 2005). Higher permeabilities, $10^{-12}$ to $10^{-10}$ m$^2$, might be reasonable for plains made up of lava flows on Mars (Berman and Hartmann, 2002; Manga, 2004).

#### 2.1.3. Background heat

Assuming conductive equilibrium, prior to an impact event the crust has a background temperature gradient and water/ice distribution established by the crustal geothermal heat flux and the ambient surface temperature. The basic relation for conductive heat transport is given by:

$$q_{\text{geo}} = k_1 \left( \frac{dT}{dz} \right)$$  \hspace{1cm} (4)

where $q_{\text{geo}}$ is the geothermal heat flux (W m$^{-2}$), $k_1$ is the thermal conductivity (W m$^{-1}$ K$^{-1}$), and $\frac{dT}{dz}$ is the temperature gradient with respect to height (K m$^{-1}$).

The thermal properties of the martian crust have been approximated using both frozen soil and basalt as analogs (Clifford, 1993). Thermal conductivities will cover a wide range on Mars but the column-averaged thermal conductivity over the cryosphere, a length-scale comparable to the modeling domain of this study, is likely to vary between $\sim$1.0 and 3.0 W m$^{-1}$ K$^{-1}$ (Travis et al., 2003). Thermal conductivity increases with ice content and decreases with increasing temperature. Here we calculate the effective thermal conductivity as a volume-weighted mean of the bulk rock thermal conductivity (2.5 W m$^{-1}$ K$^{-1}$) and ice and/or

---

**Table 1**

A list of hydrologic, thermal, and material parameters that govern PIH systems including parameters used in the governing equation for energy conservation, Eq. (11). The base of the table lists the four parameters varied in our numerical simulations of PIH systems.

<table>
<thead>
<tr>
<th>Hydrological, thermal and material parameters</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi_0$ Surface porosity</td>
<td>0.2</td>
<td>Dimensionless</td>
</tr>
<tr>
<td>$\rho_r$ Density of rock</td>
<td>2600</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$\rho_w$ Density of water</td>
<td>Variable $\sim$ (10$^3$)</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$\rho_{ice}$ Density of ice</td>
<td>917</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$c_r$ Specific heat of rock</td>
<td>1000</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$c_{ice}$ Specific heat of ice</td>
<td>E.O.S. dependent</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$q_{geo}$ Latent heat of fusion of ice</td>
<td>~2050</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$\gamma$ Liquid fraction</td>
<td>Node water fraction</td>
<td>Dimensionless</td>
</tr>
<tr>
<td>$\beta$ Latent heat of fusion of ice</td>
<td>$3.33 \times 10^5$</td>
<td>J kg$^{-1}$</td>
</tr>
<tr>
<td>$T_m$ Melt temperature range</td>
<td>25</td>
<td>K$^{-1}$</td>
</tr>
<tr>
<td>$T$ Temperature</td>
<td>273.15 pure H$2$O</td>
<td>K</td>
</tr>
<tr>
<td>$I$ Internal energy of water</td>
<td>E.O.S. dependent</td>
<td>J kg$^{-1}$</td>
</tr>
<tr>
<td>$E$ Enthalpy of liquid water</td>
<td>E.O.S. dependent</td>
<td>J kg$^{-1}$</td>
</tr>
<tr>
<td>$u$ Darcy velocity of liquid</td>
<td>Node dependent</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$k_T$ Thermal conductivity of ice-</td>
<td>Rock: 2.5, ice:</td>
<td>W m$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td></td>
<td>rock matrix</td>
<td>temp. dep.</td>
</tr>
<tr>
<td>$g$ Gravitational acceleration</td>
<td>3.71</td>
<td>m s$^{-2}$</td>
</tr>
<tr>
<td>$\alpha$ Thermal expansivity of water</td>
<td>E.O.S. dependent</td>
<td>k$^{-1}$</td>
</tr>
<tr>
<td>$\mu$ Dynamic viscosity of water</td>
<td>E.O.S. dependent</td>
<td>Pa s</td>
</tr>
<tr>
<td>$H_s$ Hydrological scale height</td>
<td>2800</td>
<td>m</td>
</tr>
<tr>
<td>$q_{geo}$ Background geothermal flux</td>
<td>32.5</td>
<td>mW m$^{-2}$</td>
</tr>
<tr>
<td>$T_{geo}$ E.O.S. temperature limits</td>
<td>0–900</td>
<td>°C</td>
</tr>
<tr>
<td>$P_{geo}$ E.O.S. pressure limits</td>
<td>$10^{2}$–$10^{9}$</td>
<td>Pa</td>
</tr>
<tr>
<td>$D$ Final crater diameter</td>
<td>45, 90</td>
<td>km</td>
</tr>
<tr>
<td>$R_f$ Impactor radius</td>
<td>1900, 4700</td>
<td>m</td>
</tr>
<tr>
<td>$T_0$ Ambient surface temperature</td>
<td>5, 53.15</td>
<td>°C</td>
</tr>
<tr>
<td>$k_0$ Surface permeability</td>
<td>$10^{-16}$, $10^{-14}$, $10^{-12}$, $10^{-10}$</td>
<td>m$^2$</td>
</tr>
</tbody>
</table>
water content for each model node throughout the domain at a given time. We obtained expressions for the thermal conductivities of ice and liquid water from Grimm and McSween (1989).

Although no direct measurements of \( q_{geo} \) have been made for Mars, estimates based on compositional (Fanale, 1976) and geophysical arguments (Davies and Arvidson, 1981; Schubert et al., 1992; Nimmo and Tanaka, 2005) generally cluster around \( \sim 30\text{ mW m}^{-2} \) during the early to mid-Noachian Period, higher concentrations of radiogenic elements would have fueled a higher geothermal flux of \( \sim 60\text{ mW m}^{-2} \) (Nimmo and Tanaka, 2005). This study will assume a constant background geothermal flux of 32.5 mW m\(^{-2}\) that enters through the base of the domain. A heat flux of 32.5 mW m\(^{-2}\) is equivalent to the temperature gradient of 13 °C km\(^{-1}\) used by Abramov and Kring (2005), assuming a thermal conductivity of 2.5 W m\(^{-1}\) K\(^{-1}\). The crustal heat flux of Mars, like Earth, is likely spatially nonuniform (Clifford and Hillel, 1983; Squyres et al., 1992). However, given the considerable uncertainty in the global average value, the addition of higher order spatial variations is unwarranted.

### 2.2. Analytical expressions

Following an impact event, the distribution of subsurface ice and the surface discharge of liquid water depend on heat sources and permeability. The impact event itself is the most important initial condition; the energy delivered by the impact and the redistribution of shock-heated material and rock at depth sets the stage for any subsequent hydrothermal activity. Before exploring results from simulated evolution of PIH systems, it is informative to develop simple analytic expressions that approximate the governing physics.

#### 2.2.1. Cooling timescales

An order of magnitude estimate of the thermal conductive cooling time of an impact structure is:

\[
t_{\text{cond}} \sim \frac{l^2}{k}
\]

where \( l \) is a characteristic length scale, and \( k \) is the thermal diffusivity. An \( l \) of 1 km and a \( k \) of \( \sim 10^{-6} \text{ m}^2 \text{ s}^{-1} \) (i.e. basalt) yields a conductive cooling time of \( \sim 30,000 \) years.

The advection of fluid reduces cooling timescales. For advection, we use the Dupuit approximation (Turcotte and Schubert, 2002), to obtain a time scale for the flow as a function of permeability, \( k \), in the absence of heat:

\[
t \sim \frac{l^2 \mu}{k \rho gh} \sim t_{\text{cond}} \frac{\kappa \mu}{k \rho gh}
\]

Here, we have introduced a distance along the direction of flow, \( l \), over which the head changes by an amount \( h \). Thus, \( h \) is technically a head difference, and the ratio \( h/l \) is the head gradient driving flow. Other variables are defined as follows: density, \( \rho \), gravity, \( g \), and dynamic viscosity, \( \mu \).

To first order, the shock-heated region of an impact structure occupies a hemisphere that will be desiccated as a result of the shock. Assuming a saturated region with a permeability of \( 10^{-12} \text{ m}^2 \) exists outside the zone of desiccation and that \( l \sim h \sim 1 \text{ km} \) due to the shocked zone’s spherical geometry, Eq. (6) shows that for a fluid viscosity of \( \sim 10^{-3} \text{ Pa s} \), the time required for fluid to inundate the shocked region is \( \sim 10 \text{ years} \)—over four orders of magnitude shorter than the conductive cooling time (Eq. (5)). This means that the hottest and driest portion of the PIH system experiences negligible cooling before water and steam begin to interact with it.

Permeability exerts the strongest control on cooling time. Higher permeabilities allow greater advection of heat and cool hydrothermal systems more rapidly than low permeabilities. Assuming 1-D steady state-flow driven by the buoyancy of hot water, it is possible to derive the Darcy velocity, \( u \), of a rising plume and hence a thermally driven advective timescale, \( t_{\text{adv}} \) (cf. Turcotte and Schubert, 2002):

\[
t_{\text{adv}} = \frac{\mu l}{(k/\rho g)} \left( \frac{1}{\Delta T} \right)
\]

where \( \Delta T \) is the temperature contrast across a characteristic length scale, \( l \). Fig. 1 plots \( t_{\text{adv}} \) against the surface permeability and shows that bigger \( \Delta T \) results in quicker cooling, because velocity scales with \( \Delta T \) and heat flow \( \left( \dot{q} = \rho C_p u \Delta T \right) \) scales as \( \Delta T^2 \). For likely surface permeabilities, \( 10^{-14} - 10^{-10} \text{ m}^2 \), systems will advect heat to the surface in \( \sim 10^3 - 10^8 \) years. The timescales generated by this analytical calculation are similar to results obtained by more sophisticated numerical calculations (see Section 3).

#### 2.2.2. Geothermal flux

The geothermal heat flux, \( q_{geo} \), plays two important roles in the behavior of a hypothetical hydrosphere residing below the cryosphere (Travis et al., 2003). Together with permeability, \( q_{geo} \) determines whether or not the hydrosphere transfers heat conductively or by convection. The Rayleigh number \( (Ra) \) characterizes the vigor of convection:

\[
Ra = \frac{g \beta (k) \Delta \theta q_{geo}^2}{\mu k T}
\]

where \( (k) \) is the domain averaged permeability and \( \Delta \theta \) is the domain height. For \( Ra \) values greater than 20–40 meta-stability gives way to
fluid convection in an open porous medium with a porosity that decays exponentially with depth (Nield and Bejan, 1992). Fig. 2 shows that conductive or convective driven heat transfer depends primarily on permeability—$k_0 \gg 10^{-11}$ m$^2$ assures background convection in absence of impacts. The geothermal flux also determines the depth of the cryosphere. PIH systems that have enough initial heat to penetrate through the cryosphere to an aquifer below will have much more water available for hydrothermal flow.

2.3. Energy delivered by bolide impact

The bolide impact delivers heat and redistributes material. We estimate the post-impact subsurface temperature field by combining an analytic calculation for the change in temperature as a function of shock-emplaced heat with the effects of excavation and geothermal uplift (Fig. 3). A short description of our method for generating the initial temperature fields is provided below. The Appendix contains calculation details. Our temperature fields combine aspects of previous analytical calculations for shock-heating (Kieffer and Simonds, 1980; Melosh, 1989; Rathbun and Squyres, 2002; Abramov and Kring, 2005) with analytical calculations for crater formation (Melosh, 1989; Cintala and Grieve, 1998). Initial temperature fields agree well with temperature fields generated by hydrocode calculations (Pierazzo et al., 2005; Senft and Stewart, 2008).

Initial temperature distributions are established for two final crater diameters (45 km and 90 km), two surface temperatures (5°C and −53.15°C) and a background geothermal flux of 32.5 mW m$^{-2}$. The 5°C surface temperature was chosen to match previous PIH studies (Rathbun and Squyres, 2002; Abramov and Kring, 2005); the −53.15°C or 220 K temperature (hereafter rounded to −53°C) was chosen to approximate present-day mean annual surface temperatures at the equator of Mars. If the surface permeability we specify yields a convective hydrosphere (see Fig. 2), then we obtain an initial average geothermal temperature profile from numerical simulations run to a dynamic equilibrium state in the absence of impact-induced heating (cf. Appendix). In all cases we assume an initially static aquifer. For the 45 km crater we assume a bolide with velocity, radius ($R_p$) and density of 7 km s$^{-1}$, 1900 m, and 2600 kg/m$^3$, respectively. The same velocity and density are assumed for the 90 km diameter crater; we simply assume a larger bolide diameter of 4700 m. Using an analytical expression for the amount of waste heat derived from the Murnaghan equation of state (Kieffer and Simonds, 1980), the amount of shock energy initially imparted by the bolide is 66% of the kinetic energy (cf. Appendix). Excavation and vaporization removes a portion of the shock energy bringing final waste heat budgets into agreement with hydrocode studies (Gisler et al., 2006).
Fig. 3 is a plot of our initial temperature distribution for a PIH system occupying a 45-km crater exposed to subfreezing surface temperature. It includes contributions from shock-heating, the background geothermal temperature profile, and advected heat from the centrally uplifted region. The shock-heating is strongly concentrated at the top center of the crater and falls off rapidly with distance. Fig. 3 shows how this steep falloff leaves most of the crater’s floor and its rim frozen. Hotter material brought from depth as the crater collapses from its transient to its final form supplies additional heat to the PIH system as indicated by the concave down temperature contour lines in Figs. 3 and 11.

2.4. The numerical model: MAGHNUM

We simulate the evolution of PIH systems using the Mars Global Hydrology Numerical Model (MAGHNUM) (cf. Travis et al., 2003). MAGHNUM iteratively provides a finite difference numerical solution of the governing partial differential equations for 1-, 2-, or 3-D, time-dependent transport of non-isothermal water and heat through a porous, permeable medium. MAGHNUM addresses the dynamics of vapor, water and ice in extraterrestrial environments (Travis et al., 2003; Travis and Schubert, 2005; Schubert et al., 2007). In this application, we do not treat vapor dynamics. Fluid flow is governed by the thermohydrologic properties of the martian subsurface and modified by the melting or freezing of ice clogging the pores.

2.4.1. Governing equations

The mass balance equation for the water component is:

$$\nabla \cdot (\rho \vec{u}) = 0$$  \hspace{1cm} (9)

Darcy’s law determines the fluid velocity:

$$\vec{u} = -\frac{k}{\mu} \nabla P + \rho g z$$  \hspace{1cm} (10)

where additional variables are pressure ($P$), and the height ($z$) above some reference ($z_0$). The use of Darcy’s law assumes relatively low velocities so that flow is entirely laminar. During the earliest stages of the PIH circulation, flow is likely to be turbulent in the near surface, in the hottest regions, where our results may slightly overestimate fluid velocities.

Incorporating freezing complicates the energy balance equation (Travis et al., 2003), yielding the following equation—variables defined in Table 1:

$$\frac{\partial}{\partial t} \left[ (1 - \phi) \rho_c c_T + \phi \rho p l + (1 - \sigma) \rho_{c_e} c_e \cdot \left( \frac{c_e}{c_e T} \right) \right] + \nabla \cdot (\rho \vec{u} E) = \nabla \cdot (k \nabla T)$$  \hspace{1cm} (11)

The left-hand side of Eq. (11) quantifies the time rate of change of energy for five processes: the rate of change of energy in the three materials (rock, water, and ice); the melting/freezing process; and the advection of enthalpy. These are balanced on the right-hand side by the diffusion of heat. The integral term represents cumulative release and absorption of latent heat around temperature $T_m$. The ice fraction’s dependence on the capillary pressure is omitted; its effect will be minor compared to the temperature dependence for many permeable materials, and it can be mimicked to some degree by choice of $\beta$ (Travis et al., 2003). $\beta$ controls how smoothly the phase change occurs and is chosen here to be 25 K$^{-2}$.

The equation of state tables of Harr et al. (1984) determine the pressure and temperature dependencies of density, viscosity, heat capacity, thermal expansivity and enthalpy for liquid water. The equation of state limits for pressure and temperature in these tables are between $10^5$ and $2 \times 10^8$ Pa (1–2000 bars) and 0°C and 900°C. Although the average surface pressure on Mars, $\sim 10^5$ Pa, is below the table’s limits, a hydrostatic pressure of $10^5$ Pa is reached at a depth of $\sim 30$ m under Mars’s gravity. This depth is well within the first grid row. The change of state between liquid water and water ice occurs over a sharp temperature interval $\sim 0.1$–0.2°C (Travis et al., 2003). Ice fraction within pores controls permeability (Eq. (3)) which, in turn, controls the Darcy velocity (Eq. (10)) and affects the energy balance (Eq. (11)).

2.4.2. Numerics

MAGHNUM discretizes the governing equations in space using an integrated finite difference algorithm on a grid, and discretizes in time using the fully implicit backward Euler method. MAGHNUM has 1-D, 2-D, or 3-D capability, on irregular (but orthogonal) grids. In this application, we use 2-D cylindrical geometry and a uniform grid. The resulting non-linear system of flow equations is solved at each time-step using the Newton–Raphson iterative method. At each time-step, a sequential process first solves for pressure; this allows calculation of fluid velocities. Then transport of heat updates energy and mass in each grid cell. Finally, temperature and ice fraction are determined using the equation of state.

2.5. Domain setup and boundary conditions

Taking advantage of the crater’s radial symmetry, we model PIH systems in a 2-D axisymmetric domain with 100 radial nodes and 50 vertical nodes. We place the center of the crater’s surface at the top left-hand corner of the domain. The right half of Fig. 3 represents the dimensions of the 2-D domain; the left half is simply its reflection. Depending on crater diameter, 45 km or 90 km, the domain spans 30 km or 60 km radially with respective node widths of 333 m or 666 m. We set the domain height to the depth of assumed pore closure, 10 km, yielding a node thickness of 200 m. Our work differs from Rathbun and Squyres (2002) and Abramov and Kring (2005) in that we ignore topographic relief—an assumption we discuss further in Section 5.3.

Although energy and mass are conserved across all cell interfaces, the domain’s boundary is open, allowing heat and fluid to exit or enter. We assume that the domain is saturated at all times; any mass exiting the domain must be balanced by mass influx elsewhere. The following boundary conditions control the four sides of the domain. Along the base, we prescribe a constant geothermal flux and allow free fluid flow (although permeability is very low). At the axis of symmetry (left-hand side) the temperature gradient is zero, temperature is variable, and no horizontal fluid flow is permitted. We assume the right-hand side to be far from the PIH system, fix the temperature at values assigned by the geothermal profile, and permit fluid flow.

The top boundary condition allows fluid to exit and enter the domain and is designed to simulate the physical effects imposed by subfreezing surface temperatures. At each time step, the boundary condition of each surface cell is determined by whether or not positive fluid flow out of the cell is occurring. This binary condition controls the behavior of a row of dummy cells that, conceptually, reside above the surface cells. If a particular surface cell has positive upwards flow then its corresponding dummy cell is set to the surface cell’s temperature, neither cooling nor heating that surface cell. Otherwise, the dummy cell is set to the ambient temperature value and conductively cools the surface cell. The row of dummy cells is essential, because no surface melting would occur in their absence. Our approach attempts to recreate the physics of fluid entering a frozen environment—while flow is occurring the cell is well out of equilibrium with its environment and when flow ceases the cell cools conductively toward thermal equilibrium with the atmosphere.
2.6. Parameter space

In this study we focus on how surface temperatures and freezing affect geochemical and geomorphological characteristics of PIH systems. Given this aim, we vary three parameters: crater diameter, ambient surface temperature and surface permeability. Crater diameter, set to 45 or 90 km, determines the amount of heat available to drive hydrothermal flow. The surface temperature affects discharge and W/R ratios. By setting the temperature to 5 °C we test hypothetical warm periods of Mars’s past and compare our results with the work of Abramov and Kring (2005). A surface value of −53 °C simulates PIH behavior throughout most of Mars’s past including the present epoch. The permeability of Mars’s subsurface is unknown and values for reasonable geological materials span many orders of magnitude. We prescribe a wide range of surface permeability values (10^{-10}, 10^{-14}, 10^{-12}, and 10^{-16} m^{2}) in order to explore the effects of both conductive and convective hydrospheres (Fig. 2). These variable parameters are listed in Table 1.

3. Results

3.1. Discharge rate and total discharge

We record the Darcy velocity throughout the domain at each time step and assume surface discharge to be the positive vertical component of the velocity vector for each surface node. We calculate the surface discharge rate as the product of Darcy velocity and area for the entire PIH system by summing the discharge rates for all surface nodes as follows:

\[ q = \sum_{i=1}^{n_{\text{nodes}}} \max(0, v_i) (r_i^2 - r_{i-1}^2) \]

(12)

where \( r_i \) is the radius at each node, and \( v_i \) the magnitude of the positive vertical component of velocity determine \( q \), the discharge rate at a particular time. The total discharge is an integration of the entire system’s instantaneous discharge rate, \( q \), over time; we assume that no steam or fluid is lost to the atmosphere.

Fig. 4A shows how discharge rate varies with permeability for 5 °C surface temperature runs. Fig. 4A shows that discharge rates for a system with an especially low surface permeability, 10^{-16} m^{2}, are negligible. Higher surface permeabilities, 10^{-12} and 10^{-10} m^{2}, produce higher initial discharge rates of \(~3 \) and \(~60 \) m^{3} s^{-1}, respectively. Although discharge still scales with permeability, these systems continue to produce low discharge rates of about 1 m^{3} s^{-1} for millions of years as heat from the central uplift reaches the surface beginning around 100 kyr after the impact event.

Systems with a permeability of 10^{-14} m^{2} produce initial discharge rates on order of 0.1 m^{3} s^{-1} that decay to negligible values over 10^{6} years. These discharge rates are comparable to discharge rates we infer for a smaller, 30 km diameter PIH system simulated by Abramov and Kring (2005). To make this comparison we converted three reported peak flux values (kg s^{-1} m^{-2}) simulated by Abramov and Kring (2005) to discharge rate (m^{3} s^{-1}) by assuming that the peak flux occurred uniformly at the surface from the center of the crater to a radius of 1 km, an area of \( A = \pi(10^{3})^2 \) m^{2}, and applying the following conversion: \( q_{\text{flux}} = A \frac{k_{0}}{\text{perm}} \). Inferred discharge rates are 0.05, 0.02, and 0.002 m^{3} s^{-1} occurring at 10^{3}, 10^{5}, and 10^{6} years, respectively. Fig. 4A shows that these inferred discharge rates (diamonds) are in reasonable agreement with our results.

Fig. 4B shows that for PIH systems exposed to frozen surface temperatures (−53 °C), surface discharge rates again scale with permeability, but cease when the entire surface refreezes. Frozen PIH systems with low surface permeabilities, 10^{-16} m^{2}, freeze in only \(~1700 \) years and produce negligible discharge rates and total discharge. Greater surface permeabilities, 10^{-14} and 10^{-12} m^{2}, yield greater initial discharge rates and increase the length of time that the surface remains unfrozen. Systems with high permeability, 10^{-10} m^{2}, behave in a fundamentally different way. They harbor a pre-existing vigorously convecting hydrosphere capable of mining heat from the central uplift and delivering it to the surface before the surface can freeze. This additional heat yields discharge rates and total discharges that are an order of magnitude greater and allow the system to maintain contact with the surface for >50 kyr.

A bolide capable of generating a 90 km diameter crater contains roughly 10 times as much kinetic energy as a bolide that forms a 45 km diameter crater: \(~10^{22} \) versus \(~10^{21} \) J. Fig. 5 shows that this additional shock-heating and a more extensive central uplift produce surface discharge rates and total surface discharges that are roughly 10 times as great as the 45 km case for systems exposed to frozen temperatures. Low permeability systems (k_{0} = 10^{-16}, 10^{-15} m^{2}) freeze within 3000 years and produce low total discharges of 10^{7} and 10^{9} m^{3}, respectively. The larger shocked region of the 90 km crater and greater central uplift allows heat mined from the central uplift to advect to the surface before it freezes for systems with surface permeabilities of 10^{-12} and 10^{-10} m^{2}. Fig. 5 shows that, ignoring loss to the atmosphere, the total discharge from these systems would fill their 90 km craters to a depth of \(~1 \)–8 km.
permeability evolution with time, we did not include chemical effects in our model. We recorded the cumulative $W/R$ ratio by time integrating the incremental $W/R$ ratio measured at each node throughout PIH system evolution. The incremental $W/R$ ratio is a dimensionless measure of the water mass that flowed through a given rock mass (Taylor, 1977; Fisher and Narasimhan, 1991). We calculated the incremental $W/R$ ratio as follows:

$$\frac{W}{R} = \Delta t \left( \frac{\rho_{H_2O}}{\rho_{\text{rock}}(1 - \phi)} \right) \left[ \frac{\bar{u}_r + \bar{u}_s}{\Delta r} \right]$$

where $\Delta t$, $\Delta z$, and $\Delta r$ are the time step, vertical node size and radial node size, respectively. The radial and vertical components of the Darcy velocity vector, $\bar{u}$, are defined as $u_r$ and $u_z$. Alternatively, Eq. (13) can be written as:

$$\frac{W}{R} = \Delta t \left( \frac{\rho_{H_2O}}{\rho_{\text{rock}}(1 - \phi)} \right) \left[ \bar{u} \times \bar{W} \right]$$

In this formulation it is clear that $W/R$ ratio depends on the $w/A$ ratio (Fisher and Narasimhan, 1991), where $A$ is node area, and $\vec{w} = (\Delta r, -\Delta z)$. As $\vec{w}$ goes to zero so will $A$ and the $W/R$ ratio will go to an asymptotic value. Like temperature or pressure, $W/R$ is a point function and we approximate its value on the domain grid.

Node area is uniform throughout the domain and for all simulations of a particular crater size but the node size in the radial direction is twice as great in the 90 km versus 45 km cases (666 m versus 333 m). This reduces the accuracy of the contribution from the horizontal component of velocity in the 90 km cases. However, PIH systems and bottom heated convective domains principally produce vertical flow so we assume that a direct comparison of cumulative $W/R$ ratios between runs with different crater sizes is permissible in this study. The ‘absolute’ $W/R$ ratios calculated here should exceed geochemical estimates because (1) residence times may be too short to allow geochemical equilibrium, and (2) water often passes through previously altered rock (Fisher and Narasimhan, 1991). In any case, relative differences between $W/R$ ratios at different locations or between different runs are more important

**3.2. W/R ratios and temperature evolution**

The chemical alteration of rock by hydrothermal circulation depends on host-rock chemistry, temperature, peak pressure, exposure time, and the amount of water that flowed through the rock (Griffith and Shock, 1997). $W/R$ ratios strongly influence the chemistry and mineralogy of hydrothermal alteration products in basalt (Fisher and Narasimhan, 1991). Although Osinski (2005) shows that host-rock chemistry may play an important role in porosity-

**Fig. 5.** Similar to the 45 km simulations, the permeability of 90 km PIH systems controls discharge rates and total surface discharge. Low permeability systems, $10^{-15}$-$10^{-14}$ m$^2$ produce geomorphologically insignificant discharge rates and are incapable of advecting enough heat to remain unfrozen at the surface much longer than a few 1000 years. The $10^{-17}$ m$^2$ system exhibits much larger discharge rates, total discharge and system lifetime when compared to the 45 km system. It behaves like the 45 km, $10^{-18}$ m$^2$ case—the additional, more deeply deposited, heat present in the 90 km crater allows the surface of the $10^{-17}$ m$^2$ system to remain unfrozen long enough for heat from the central uplift to advect to the surface. In general, the 90 km systems produces greater discharge rates and total discharges when compared to the 45 km systems but permeability plays the primary role in determining how quickly the surface freezes.

**Fig. 6.** Profile plots of $W/R$ ratios for the upper 200 m of rock (top panels) and contour plots at depth (bottom panels) as a function of radius for 45 km PIH systems with a surface permeability of $10^{-18}$ m$^2$ exposed to ambient surface temperatures of 5 °C (left) or -53 °C (right). Surface permeabilities $\geq 10^{-15}$ m$^2$ yield rapid advection of fluid and enable convection in the hydrosphere fueled by geothermal heat alone. These two effects produce high $W/R$ ratios, >100 at depth, especially in convective plumes. In planar view, convective plumes may produce a diagnostic ‘bulls-eye’ pattern of geochemical alteration due to alternating annuli of high and low $W/R$ ratios (left). For PIH systems under frozen conditions, the cryosphere prevents convective plumes from reaching the surface near the rim of the crater (right); discharge and $W/R$ ratios are zero. Frozen PIH systems actually produce higher $W/R$ ratios at the center of the crater than clement systems. As the PIH system freezes, the remaining unfrozen area of the surface becomes increasingly constricted toward the center of the crater, focusing flow there.
than the absolute values. They indicate at which locations geo-
chemical alteration is more likely to have occurred.

\( W/R \) ratios at all depths increase with permeability. Figs. 6–8 show profiles for the \( W/R \) ratios for the uppermost node (upper 200 m of rock) and contour plots of the \( W/R \) ratios for rocks in the subsurface to a depth of 10 km. Fig. 9 tracks the evolution of temperature and cumulative \( W/R \) ratios for the upper 200 m of the centermost node of the crater. Fig. 6 shows that systems with higher permeabilities (\( 10^{-10} \) m\(^2\)) allow convection, rendering them capable of mining heat from the central uplift before the surface freezes. Convective systems subjected to surface temperatures below freezing are particularly interesting because an impermeable front—the surface ice/water interface that marches inwards from mid-floor to the central peak—forces flow towards the center of the crater. This effect prolongs modest near-surface temperatures (0–100 °C for 50 kyr) at the crater center and yields \( W/R \) ratios >1000 (Fig. 9D). Vigorous convective upwelling plumes enhance \( W/R \) ratios far from the crater’s center in the non-frozen case but are unable to melt through the cryosphere in the frozen case (Fig. 6). The profile of \( W/R \) ratios of surface-exposed rock in Fig. 6

**Fig. 7.** Profile plots of \( W/R \) ratios for the upper 200 m of rock (top panels) and contour plots at depth (bottom panels) as a function of radius for 45 km PIH systems with a surface permeability of \( 10^{-10} \) m\(^2\) exposed to ambient surface temperatures of 5 °C (left) or –53 °C (right). Modest permeabilities allow advection of heat and convective plumes in the shock-heated region and above the central uplift. For these systems, ambient surface temperatures increase \( W/R \) ratios. Systems exposed to element ambient temperatures continue to advect heat to the surface from the shock-heated region and the central uplift for ~10\(^6\) years producing \( W/R \) ratios of ~300 at the crater’s center and of ~100 in areas of the surface above convective plumes. When exposed to subfreezing ambient surface conditions, the upper 200 m of the system freezes in <10,000 years. \( W/R \) ratios are zero for much of the surface. Flow never occurs at the rim or in much of the floor; surface exposed \( W/R \) ratios do not exceed ~10 at the central peak. PIH systems that produce modest \( W/R \) ratios of ~10 may explain spectral detection of alteration minerals present on the central peak of mid-sized craters.

**Fig. 8.** Profile plots of \( W/R \) ratios for the upper 200 m of rock (top panels) and contour plots at depth (bottom panels) as a function of radius for 90 km PIH systems exposed to an ambient surface temperatures of –53 °C with a surface permeability of \( 10^{-10} \) m\(^2\) (left) or \( 10^{-12} \) m\(^2\) (right). Ninety kilometer craters harbor PIH systems that produce \( W/R \) ratios that are ~10 times greater than respective 45 km systems. Systems exposed to subfreezing ambient surface temperatures produce a surface \( W/R \) ratio profile that is similar in character but larger in magnitude than the 45 km system. \( W/R \) ratios are concentrated at the center because that location is the last to freeze. Convective plumes in the system exposed to element surface conditions produce an alternating pattern of high (>1000) and modest (~100) \( W/R \) ratios. In the subfreezing case, these convective plumes fail to carve through the cryosphere and do not produce discharge at the rim.
shows that the cryosphere prevents flow and maintains zero near-surface \( W/R \) ratios at the rim and across much of the crater floor.

For 45 km diameter craters with lower permeability, \( 10^{-12} \) m\(^2\), exposed to frozen surface temperatures, the upper 200 m of rock at the crater’s center experience fluid temperatures between 0 and 100 °C for 20,000 years and \( W/R \) ratios of \( \sim 50 \) (Figs. 7 and 9C). All flow subsequent to surface freezing occurs in deeper parts of the domain, 2–6 km below the surface, where permeability is roughly 1% of the surface value. Fig. 7(right) shows how reduced permeability at depth restricts flow and limits \( W/R \) ratios to values \( \ll 1 \).

With surface temperatures above freezing, the discharge rate slowly decays as heat is advected from the system for \( 10^5 \)–\( 10^6 \) years (Fig. 9A). This prolonged surface contact and flow in areas of low permeability yields greater \( W/R \) ratios of \( \sim 300 \) at the central peak and \( W/R \) ratios of 1 in convective plumes extending to the surface in regions near the rim (Fig. 7, left). By the time the convective plumes reach the surface far from the crater’s center, the fluid has cooled to the surface temperature, 5 °C.

Larger, 90 km, frozen PIH systems produce \( W/R \) ratios \( \sim 2\)–\( 5 \) times greater than their respective 45 km counterparts at the central peak in simulations with surface permeabilities of \( 10^{-11} \) and \( 10^{-12} \) m\(^2\) (Fig. 8). In the \( 10^{-12} \) m\(^2\) case, convective plumes create alternating annuli with high \( W/R \) ratios (\( \sim 1000 \)) and modest \( W/R \) ratios (\( \sim 400 \)) with discharge temperatures near the surface value. (Fig. 8, left) shows a strong \( W/R \) ratio peak at the center of the crater and a second one at a radius of \( \sim 13 \) km. The second peak is generated by fluid, heated by the shocked region and the central uplift, being directed by an overlying cryosphere. The \( 10^{-12} \) m\(^2\) case generates a surface \( W/R \) ratio profile that is similar to, but larger than, the frozen 45 km case. The most significant difference is that the initially unfrozen surface extends radially out to 20 km versus 10 km in the 45 km crater. This increases the surface area that experiences \( W/R > 0 \) by a factor of four to \( 10^9 \) m\(^2\). Again, as the surface freezes, discharge is forced toward the crater’s center, enhancing \( W/R \) ratios there. Similar to the 45 km case, Fig. 8 shows that convective plumes present in highly permeable simulations (\( 10^{-12} \) m\(^2\)) fail to erode the cryosphere at radial distances beyond about 5 km. Consequently, even the most vigorous frozen PIH systems do not produce discharge at the crater wall or rim.

In summary, permeability, crater diameter and surface temperature affect surface discharge rates, surface discharge duration, \( W/R \) ratio values and \( W/R \) ratio distribution. Total surface discharge and discharge rate increase with permeability and crater size. The background geothermal heat flux allows the hydrosphere to convect for surface permeabilities \( \geq 10^{-11} \) m\(^2\). Convection

![Fig. 9](https://example.com/fig9.png)

Fig. 9. These four plots show the time-evolution (x-axis) of temperature (solid line, left y-axis) and \( W/R \) ratios (dashed line, right y-axis) in the crater’s centermost 333 m and surface-exposed upper 200 m of rock for four 45 km PIH system simulations. (B and C) \( W/R \) ratios cease to accumulate once temperatures fall below 0 °C (shaded region). If the surface remains unfrozen long enough, advective plumes originating from the central uplift reach the surface, elevate surface temperatures and increase cumulative \( W/R \) ratios (A and D). As the system cools, the surface freezes beginning with more radially distant regions and concluding with the crater’s center. This forces flow toward the center of the crater which greatly increases cumulative \( W/R \) ratios there (C and D).
significantly increases surface discharge rates and duration of PIH systems. Frozen surface temperatures reduce surface discharge duration and affect the distribution of surface-exposed rock W/R ratios. As PIH systems freeze, ice directly and focuses flow to unfrozen regions. This greatly increases surface discharge W/R ratios in the last place on the surface to freeze, the central peak. Fig. 10 shows the cumulative W/R ratios at the central peak as a function of surface permeability for the subfreezing 45 km and 90 km cases as well as the unfrozen, 10^{-12} m^2, 45 km case. At low permeabilities W/R ratios are negligible at the central peak. For surface permeabilities of 10^{-12} m^2 crater size becomes important. The frozen 90 km case is capable of delivering heat from the central uplift to the surface before it freezes, yielding high W/R ratios of ~300. At high permeabilities, 10^{-10} m^2, Fig. 10 shows that frozen systems have higher W/R ratios at the central peak than unfrozen systems. This occurs because of the focusing effect of the advancing freezing front.

4. Discussion

Here we will examine the extent to which PIH hydrothermal systems associated with 45–90 km diameter craters on Mars are capable of generating detectable geochemical and geomorphological features. Three principal aspects of the PIH system drive hydrothermal behavior and determine the scale and character of these features: crater size, permeability, and surface temperature. The crater size determines the PIH system’s heat budget. Permeability controls convective vigor, surface discharge, and W/R ratios. Surface temperatures determine the presence and extent of a cryosphere, which focuses flow towards the center of the crater, which, in turn, modulates surface discharge and W/R ratios.

### Table 2

<table>
<thead>
<tr>
<th>Feature</th>
<th>Max. discharge (m^3 s^{-1})</th>
<th>Duration (years)</th>
<th>Total discharge (m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>45 km PIH system</td>
<td>0.1–10</td>
<td>10^3–10^5</td>
<td>10^9–10^{12}</td>
</tr>
<tr>
<td>90 km PIH system</td>
<td>0.1–100</td>
<td>10^5–10^6</td>
<td>10^9–10^{13}</td>
</tr>
<tr>
<td>Outflow Channels</td>
<td>10^{-5–10}^{(a,b)}</td>
<td>Uncertain</td>
<td>Uncertain</td>
</tr>
<tr>
<td>Valley Networks</td>
<td>10^{-5–10}^{(c–e)}</td>
<td>&lt;10^3–10^6 years</td>
<td>Uncertain</td>
</tr>
<tr>
<td>Alluvial fans</td>
<td>10^{-10}^{(f)}</td>
<td>100–300 days</td>
<td>10^9–10^{13}</td>
</tr>
</tbody>
</table>

Comparisons of discharge rates and durations of martian fluvial morphologies. PIH system discharge rates are capable of producing gullies but are likely incapable of generating alluvial fans even if permeability structures favored their formation. In all cases, a permanently frozen region prevents discharge near the rim. PIH systems associated with 45 km craters are incapable of supporting valley networks or larger fluvial morphologies.

#### 4.1. Geomorphic signatures

Several craters are associated with features suggestive of past surface water including gullies, deltas, meanders and lake deposits (e.g. Cabrol and Grin, 2001). Table 2 compares model PIH system durations and discharges with those estimated to be responsible for martian features. Outflow channels require massive discharge rates that dwarf rates associated with PIH systems. Valley networks, too, require discharge rates that are somewhat too large. A few craters exhibit alluvial fans with alcoves perched on the crater’s wall and debris aprons deposited on the crater’s floor (Moore and Howard, 2005). Kraal et al. (2008) suggest that a large-scale fan present in a 120 km crater formed in only a few decades and that it required discharge rates on order of 10^{-1}–10^{5} m^3 s^{-1}. It is possible that smaller-scale fans might be associated with PIH systems, but the location of their alcoves, high up on the crater’s wall, argue against a hydrological relationship to PIH systems unless heterogeneous permeability structures are involved. In all simulations of PIH systems exposed to frozen ambient surface conditions, a permanently frozen region prevents discharge near the crater wall. Alluvial fans are therefore likely not directly caused by PIH systems subjected to freezing surface temperatures.

Groundwater discharge has been suggested for the formation of some gullies (Heldmann et al., 2007; Mellon and Phillips, 2001). Gullies ~100 m to 1 km in length on crater wall slopes of ~20° require discharge rates of ~10^{-5} m^3 s^{-1} to form (Parsons and Nimmo, 2010). These discharge rates argue for PIH systems with surface permeabilities on order 10^{-12}–10^{-10} m^2. However, the total discharge from these PIH systems would bury low-lying gullies. This would bias gully preservation to topographic highs: the central peak, peak ring, or crater wall. Numerous craters exhibit gullies on their walls (Heldmann et al., 2007) and a few craters, such as Lyot, Mojave, and Zunil exhibit gully-like, dendritic features incised into their central peaks (Tornabene et al., 2007; Williams and Malin, 2008). Permeability structures, including vertical faults within the central peak and listric faults extending radially from beneath the crater’s floor to the slumps at the wall, could provide fluid conduits. Like alluvial fans, gullies at the crater wall, however, are unlikely to be formed by PIH systems exposed to frozen surface temperatures. Again, this is simply a consequence of the permanently frozen region occupying nearly half the radial extent of the crater—directing subsurface flow to the center and preventing
flow near the wall. Although PIH systems modeled here are capable of forming gullies it seems unlikely that they are responsible for most gully formation on Mars.

Very large-scale impactors—bolides 100–300 km in diameter—produce global atmospheric effects (Segura et al., 2002). Smaller scale PIH systems may generate local-to-regional atmospheric effects. Like Icelandic geysers, steam and vapor may readily condense out near the PIH system, contributing to “pasted-on terrain” development on the rim of the crater. Orbit-induced climate variations (e.g. Laskar et al., 2002) or subsequent neighboring impact events may potentially lead to the melting of snowpacks and ice deposits, which is a suggested mechanism for gully formation (Christensen, 2003).

4.2. Geochemical alteration

The OMEGA (Bibring et al., 2006) and CRISM (Mustard et al., 2008) spectrometers indicate that phyllosilicates and hydrated silicates are widely distributed in the southern highlands. Impact craters saturate the southern highlands and indicate that they are among the oldest surfaces on Mars (Hartmann and Neukum, 2001). The crater age of the southern highlands corresponds with the Phyllosian era during which the deep phyllosilicates and layered phyllosilicates were apparently formed (Mustard et al., 2009).

Low-grade metamorphism of basaltic rock depends on temperature, pressure, host-rock chemistry and W/R ratios (Griffith and Shock, 1997). W/R ratios for PIH systems simulated here vary in magnitude and location depending on crater size, ambient surface temperature, and surface permeability. PIH systems exposed to 5 °C surface temperatures yield elevated W/R at the central peak for all simulated surface permeabilities. Systems with high surface permeabilities allow the hydrosphere to convect. Figs. 6–8 show that this convection generates spatially-varying convective plumes. Given our axisymmetric domain, in planar view this behavior would generate a bull’s-eye pattern of high (>100) and moderate (10) W/R ratios in surface-exposed rock across the entire floor of the crater. Such a pattern might generate a spatial footprint that might be detectable from orbit. Reality, however, is unlikely to be so simple: heterogeneous permeability structures would inhibit the formation of such patterns. Systems with lower surface permeabilities (10⁻¹⁴–10⁻¹² m²) exhibit modest W/R ratios at the central peak (1–100) and uniformly low W/R ratios of ~1 in exposed rocks across much of the floor.

PIH systems exposed to frozen surface temperatures produce elevated near-surface W/R ratios that peak at the center of the crater and decay to zero at the initial unfrozen extent of the crater’s floor. Systems with a high surface permeability yield central peak W/R ratios >1000. Systems with a surface permeability of 10⁻¹² m² yield lower W/R ratios. Fig. 9C shows that temperatures within the upper 200 m of rock at the crater’s center are above 100 °C for 2000 years following an impact event and remain above freezing for 20,000 years. As mentioned above (Section 1.4), low W/ R ratios and modest temperatures (~150 °C) are likely to produce prehnite, chlorites, smectites, mica, amphibole and garnets. Some of these clays and minerals, including prehnite, are detected in the central peak of a 45 km crater in Nili Fossae (Ehlmann et al., 2008). Prehnite, indicative of low-grade metamorphism of basaltic rock, is particularly compatible with PIH systems modeled here because it can form at low pressure and requires formation temperatures below ~300 °C (Liou et al., 1985; Digel and Ghent, 1994). Our model results for cases with permeabilities in the range of 10⁻¹⁴–10⁻¹² m² and frozen surface temperatures are consistent with the formation of prehnite, particularly at the central peak.

PIH systems create an environment conducive to forming mineral assemblages. Perhaps numerous and successive PIH systems occupying a wide range of crater sizes along with repeated brecciation and impact excavation of the southern highlands could have produced most of the phyllosilicates and hydrated silicates detected there. Subsequent reworking of the highlands (Craddock and Howard, 2002) and enhanced late-Noachian fluvial activity (Howard et al., 2005) involved low temperature fluvial erosion that redistributed sediments from topographic highs to low-standing basins. This redistribution would contribute to the masking of mineralogical signatures by later sedimentation.

A few other effects might be associated with PIH systems. Reported spectroscopic observations (Mumma et al., 2009) suggest that Mars may exhibit localized abundances of methane in its atmosphere. Similar to clay mineralogy, abiogenic methane production depends on the W/R ratio and the water temperature (Lyons et al., 2005; Mumma et al., 2009). Although PIH systems might be related to methane production they are unlikely to explain the reported ongoing production of methane: that would require a large scale (~45 km) crater impact event within the last 10⁶ years. PIH systems may foster a hypothetical abode for life (Abramov and Kring, 2005). Simple organisms may potentially find refuge from the harsh vagaries of the martian surface deep under the cryosphere (Sleep and Zahnle, 1998). Even under frozen surface temperatures, PIH systems provide a chemically dynamic and water-rich bridge that links the surface to these refugia for 100,000s of years. Craters that show evidence of geochemical alteration and fluvial features related to PIH systems would make a strong candidate location for micro-fossils.

4.3. Limitations, assumptions, and caveats

4.3.1. Homogeneous permeability

The importance of PIH system permeability is second only to the system’s heat budget in determining system behavior and its observable consequences. The permeabilities and porosities in our simulations decay smoothly with depth and exhibit no horizontal heterogeneity. Terrestrial hydrothermal systems have complex permeability structures and heterogeneities that greatly influence flow and heat transport (Spinelli and Fisher, 2004). The smooth convective rolls and plumes present in the 5 °C simulations might be an artifact of the axisymmetric domain and probably would not arise naturally if the domain had a complex permeability distribution (i.e. faulting, layering; Nield and Bejan, 1992). Sub-surface structures such as lava flows, faults, and buried impact craters (cf. Edgett and Malin, 2004) would likely be disrupted by the impact event but may still affect permeability. Impact craters, buried and exposed, and their associated fracture systems likely dominate the basement structural fabric of the ancient highlands (Rodriguez et al., 2005).

If anything, a homogeneous permeability and porosity distribution is the one distribution we can surely assume does not exist on Mars. But for the sake of systematic first order parameter space exploration, we incorporated such a distribution with the important complication that freezing affects permeability. We can, however, make inferences on how heterogeneous permeability structures would affect PIH behavior. Regions of high permeability may include highly fractured brecciated zones radially distributed from the center of the crater beneath the melt sheet; a central peak composed of poorly consolidated material uplifted up and through the melt sheet; and listric slump faults extending to the crater wall and rim. Conversely, a melt sheet pooled on the crater’s floor will likely steer flow to the center or toward the wall. Pore-clogging precipitation of minerals will cause flow to migrate from initial regions of high flow and hot temperatures to unclugged regions. Finally, fractures may extend well beyond our domain depth of 10 km. These fractures are unlikely to experience much geochemical alteration, but heat conducted or mined via fractures from this
depth will enter the domain during the lifetime of the hydrothermal system.

This work also ignores topographic relief. There are two reasons for doing so: (1) under Mars’s gravity, hydrothermally driven flow can easily overcome a few hundred meters of relief (Harrison and Grimm, 2002), and (2) post-impact permeability structures may limit flow where a complex crater’s natural topography would otherwise direct it. That is, where highly fractured and faulted zones occurring at relative topographic highs at the crater’s rim and central peak would enhance flow, low-permeability melt deposits on the crater’s floor would reduce flow. Previous workers, (Rathbun and Squyres (2002) and Abramov and Kring (2005)) did include topography. However, these studies assumed saturated domains which necessitated crater lake levels that were at least as high as the central peak or peak ring. This setup is essentially the same as ours in that it maintains a zero head-gradient across the surface that, in turn, prevents topography from altering surface discharge. However, Abramov and Kring (2005) show that the heat located in the central peak or peak ring does affect subsurface flow routing by wicking water up and into these areas of relatively high topography.

4.3.2. Heat budget

We have made many assumptions in the construction of our temperature distribution. Our approach produces shock-heated regions that agree well with hydrocode simulations (Pierazzo et al., 2005) and with prior modeling efforts (Abramov and Kring, 2005), but contain roughly seven times more energy than shock-heated regions generated by the Gault–Heitowit approach described in Rathbun and Squyres (2002). Consequently, our work may be compared to Abramov and Kring (2005), but not necessarily to Rathbun and Squyres (2002). We ran additional simulations with temperature distributions generated using the same method as Rathbun and Squyres (2002) and found that discharge rates were lower by less than an order of magnitude and that the fundamental behavior remained unchanged: permeability still controlled discharge rates and the time required for the system to freeze; and flow was concentrated to the center of the crater as the system froze.

We have ignored the role of volatiles and porosity in determining shock-induced heating. The low strength of ice and the shock impedance contrast between icy and rock layers may further modify the temperature distribution and ejection volumes (Senft and Stewart, 2008). Other assumptions are detailed in the Appendix including the decision to ignore contributions from a melt sheet. Finally, the background geothermal flux determines the depth of the cryosphere. Especially low background geothermal fluxes of ~10 mW/m² would support a cryosphere thick enough to prevent shock-heat penetration into the hydrosphere for craters of the size considered here. PIH systems under this scenario would be water-starved and should be a topic of future study.

4.3.3. Saturated domain

PIH systems require water to influence geomorphology and geochemistry. We assumed fully saturated model domains. Regions of Mars, especially those on the equator, are likely unsaturated (Boynton et al., 2002). Drier, partially saturated substrates would, of course, produce lower surface discharge rates and lower W/R ratios. The wide variety of crater morphologies (i.e. rampart craters, lobed ejecta) not related to subsequent modification (i.e. aeolian and periglacial effects) on Mars may be due to varying amounts of water in the crust at the time of impact.

5. Conclusion

Post-impact hydrothermal systems are a likely mechanism for delivering hot water to Mars’s cold, dry surface. The geomorphic and geochemical signatures of PIH systems occurring in 45–90 km craters on Mars depend on rock permeability and the surface temperature. If the PIH system occurs during a time in which surface temperatures are below freezing, discharge will not occur at the rim of the crater. Instead, all geomorphic activity and mineralogical alteration will occur near the center of the crater. Discharge rates, total discharge and W/R ratios increase with permeability. Systems with higher surface permeabilities (10⁻¹⁰ m²) allow convection, rendering them capable of mining heat from the central uplift before the surface freezes. Convective systems subjected to surface temperatures below freezing are particularly interesting because an impermeable freezing front, the surface ice/water interface, marches from mid-floor to the central peak and focuses flow towards the center of the crater. This effect prolongs modestly high temperatures (0–150 °C for 25 kyr) and yields W/R ratios >1000 for systems with high surface permeabilities of 10⁻¹² m² (Fig. 9D). For systems with modest surface permeabilities, 10⁻¹⁰ m², the upper 200 m of rock at the crater’s center experience fluid temperatures between 0 and 180 °C for 3500 years and W/R ratios of ~10 (Fig. 9C). These results imply that different surface permeabilities should produce spatially diagnostic mineral alteration patterns. PIH systems with surface permeabilities of 10⁻¹¹–10⁻¹² m² exposed to frozen ambient temperatures may explain mineral assemblages and fluvial features observed at central peaks of craters. In particular, the low-grade metamorphic mineral prehnite may be particularly diagnostic of the low temperature (0–300 °C), low pressure hydrothermal activity predicted by our models. Discharge rates and total discharges are capable of producing gullies, ponds and lakes, but not alluvial fans, valley networks or outflow channels. Future work should focus on the effect of heterogeneous permeability structures, specific case studies capable of tying theory to observations, and potential linkages with the regional hydrosphere and atmosphere.

Acknowledgments

We thank Don Korycansky and Andy Fisher for their advice and consultation. We acknowledge Keith Harrison, Oleg Abramov and Robert Lowell for their constructive reviews and valuable insight. The NASA Mars Fundamental Research Program Grant NNX07AU77G and the NASA Ames Graduate Student Researcher Program Grant 8254 funded this research.

Appendix A. Input temperature field generation

A.1. Heat sources

Three heat sources contribute to the initial temperature profile that drives PIH hydrothermal activity: impact-induced shock-heating, a melt sheet pooled on the crater floor, and an uplifted geothermal gradient. Unfortunately, there is not a good understanding of how large-scale impacts deposit their heat. Current theory is highly dependent upon the equation of state of impactor/target material, assumptions of how shock travels through the medium, and how that shock is converted into waste heat. Porosity plays a major role in impact heat deposition, as do pore fluids. However, hydrocode simulations of impacts into a porous planet-scale target with, for example, an exponentially decaying porosity, have yet to be investigated. That said, simplified models of impact-induced shock-heating can be constructed using
analytical formulations (e.g. Melosh, 1989) or hydrocode results (e.g. Pierazzo et al., 2005).

A.2. Shock-heating

We estimate the distribution of shock-heating produced by an impactor of radius, \( R_p \), density, \( \rho_p \), and velocity, \( v_p \), by combining an analytic expression for the change in internal energy as a function of distance from the impact point with pi-scaling arguments that quantify the morphological change from the transient to final crater. The relationship between final crater diameter, \( D \), and transient crater diameter, \( D_t \), is given by Melosh (1989): \( D_t = 0.6D \).

Once \( D_t \) has been calculated we use the following scaling relationship to determine the diameter of the bolide as a function of the bolide's density and velocity (rearranged from Schmidt and Housen (1987) and Cintala and Grieve (1998)). Note: variables describing projectile density, \( \rho_p \), target density, \( \rho_t \), bolide diameter, \( D_p \), velocity, \( v_p \), transient crater diameter, \( D_t \), and gravity, \( g \), must be expressed in cgs units:

\[
D_p = \left( 0.862D_t \left( \frac{\rho_t}{\rho_p} \right)^{\frac{1}{3}} v_p^{-0.44} g^{0.22} \right)^{1.3281}
\] (15)

In both the 45 km and 90 km case we assume an asteroid-like impact event and set \( \rho_p \) equal to \( \rho_t \) at 2600 kg m\(^{-3} \) and we set \( v_p \) to 7 km s\(^{-1} \). These are the same values used in both Rathbun and Squyres (2002) and Abramov and Kring (2005). There are, however, caveats concerning Eq. (15)’s derivation and application. First of all, it is scaled by gravity, \( g \), from empirical relationships derived from laboratory experiments and lunar morphologies (Schmidt and Housen, 1987). Furthermore, a non-unique combination of projectile and target densities, projectile diameters, and projectile velocities can generate the transient crater diameter required to fit the desired final crater diameter. Finally, hydrocode modeling indicates that low density comets impacting at high velocities will generate a final crater of comparable size compared with higher density slower asteroids, but will produce ~8 times as much melt (Pierazzo et al., 2005). Such craters may harbor very different PIH systems from those considered here and further study is needed.

Given a bolide diameter, we determine the change in temperature as a function of distance from the impact point. We follow a modified approach of the one described in Abramov and Kring (2005) which uses an expression for specific waste heat, \( \Delta E_w \), derived from the Murnaghan equation of state by Kieffer and Simonds (1980):

\[
\Delta E_w = \frac{1}{2} \left( P V_0 - \frac{2K_0 V_0}{n} \right) \left[ 1 - \left( \frac{\rho_p}{\rho_t} + 1 \right)^{-1/n} \right]
\]

\[+ \frac{K_0 V_0}{n(1 - n)} \left[ 1 - \left( \frac{\rho_p}{\rho_t} + 1 \right)^{-1/(1/n)} \right]
\] (16)

where \( K_0 \) is the adiabatic bulk modulus at zero pressure, \( n \) is the pressure derivative of the bulk modulus, \( V_0 \) is the specific uncompressed volume (\( 1/\rho_0 \)) and \( P \) is the peak shock pressure (Kieffer and Simonds, 1980; Abramov and Kring, 2005). We use the same values for \( K_0 = 19.3 \) GPa and \( n = 5.5 \) as Abramov and Kring (2005) as obtained from Gauld (1963), Kieffer and Simonds (1980) and Melosh (1989). We deviate from Abramov and Kring (2005) by choosing to use a steeper exponent, \( f \), for peak pressure falloff:

\[
P = A \left( \frac{r}{R_p} \right)^f
\] (17)

where \( r \) is the distance from the impact and \( A \) is the pressure at \( r = R_p \) and \( f \) is the pressure falloff exponent. For distances less than \( R_p \), \( P = A \). Melosh (1989) and Abramov and Kring (2005) estimate \( A \) as \( A = \rho_p [C_r + S_r u_0] u_0 \), where \( C_r \) and \( S_r \) are parameters in the linear shock velocity–particle relation (Melosh, 1989; Appendix II). For basalt, \( C_r = 2600 \) m s\(^{-1} \) and \( S_r = 1.58 \) (dimensionless); the initial shock velocity is \( u_0 = 0.5 v_p \). The change in temperature is calculated by dividing the specific waste heat by the heat capacity of basalt. Motivated by temperature distributions generated by a numerical hydrocode model for a 30-km martian crater by Pierazzo et al. (2005) we found that temperature contours with depth were replicated reasonably well analytically with a value for \( f \) of 3.0. Fig. 11B shows the shocked temperature distribution added to the background geothermal temperature profile (Fig. 11A).

Modern hydrocode studies suggest that more than 50% of the impactor’s kinetic energy can be converted into the internal energy of the target (e.g. Gisler et al., 2006; Abramov, personal communication). One can integrate Eq. (16) over a hemisphere with a radius set to the final crater radius to obtain an estimate for the amount of internal energy deposited into the target. Internal energy estimates for the 45 and 90 km craters are \( 1.3 \times 10^{21} \) J and \( 1.9 \times 10^{22} \) J, respectively. These values are roughly 67% of their respective bolides’ kinetic energies: \( 1.9 \times 10^{21} \) J and \( 2.8 \times 10^{22} \) J. The shallower falloff value of \( f = 2.016 \) used by Abramov and Kring (2005) deposits roughly twice as much waste heat as the bolide’s kinetic energy. Of course some shock-heated material would be ejected or vaporized during the impact event.

We also generated initial post-impact temperature distributions following the Gault–Heitowit approach detailed in Rathbun and Squyres (2002) and found that the shock-emplaced heat was 1% of the impactor’s kinetic energy. However, simulations using...
the Gault–Heitowit–Rathbun and Squyres (2002) approach instead of the Kieffer–Simonds–Abramov and Kring (2005) approach did not change the fundamental behavior of PIH systems exposed to subfreezing temperatures: (1) permeability governs discharge rate, total discharge, and W/R ratios, (2) an advancing freezing front concentrates flow to the center of the crater and (3) a permanently frozen subsurface region prevents discharge at or beyond the rim of the crater. For these reasons, we chose to use the Kieffer–Simonds analytic calculation of the initial temperature distribution as a tool to replicate hydrocode results, but acknowledge that is not an exact description of the physics of shock-heating due to impact.

A.3. Excavation and geothermal uplift

The impact event excavates a significant portion of the shock-heated region and compresses the remaining material hydrodynamically. This forms a transient crater. We describe the volume of material that is excavated by a parabola that extends to a depth of \( \frac{1}{10} \) the diameter of the transient crater, \( h_{\text{exc}} \) (cf. Melosh, 1989, §5.5.4) and intersects the surface at the radius of the transient crater, \( r_p \). The crosshatch pattern in Fig. 11C shows the excavation parabola. The transient crater is defined by a second parabola centered at a depth of \( \frac{1}{10} \) the diameter of the transient crater. Here we introduce three assumptions: (1) all material within the excavation parabola is removed; (2) that the final crater has no topography (i.e. the surface is flat) and that consequently; (3) the transient crater rebounds all the way to the surface and that material at pre-impact temperatures established by the background geothermal profile are brought closer to the surface by a height, \( z(r)_{\text{exc}} \) defined as a function of radius by the excavation parabola:

\[
z(r)_{\text{exc}} = \left( \frac{h_{\text{exc}}}{r_p} \right)^2 - h_{\text{exc}}
\]

Fig. 11D shows the resultant temperature distribution following excavation and rebound. This approach only moves material in the vertical direction. In reality, material would move in from the sides too. Our domain depth, however, is only 10 km deep, allowing us to assume that material from depth is moved primarily in the vertical direction.

A.4. Further temperature adjustments

The temperatures at specific nodes are adjusted further if the vapor pressure at the node is above the hydrostatic pressure (i.e. the water is boiling), or if the temperature is outside the range defined by the equation of state look-up tables. We assume that vapor advects heat and escapes on a short timescale compared to other timescales. The vapor pressure for each node is calculated as a function of temperature (Hardy, 1998). If the vapor pressure exceeds the hydrostatic pressure, \( P_{\text{hydro}} = \rho g z \) plus a background surface pressure of 10 KPa, in a particular node, then that node's temperature is iteratively reduced until it is no longer the case. Adjustments of this kind occur in the hottest and shallowest areas of the domain—the upper shock-heated region at the center of the crater—and are made to a depth of 800 m and 1800 m in the 45 km and 90 km case, respectively.

MAGHNUM assigns the viscosity, density and expansivity of water within each node using an equation of state look-up table (see Section 2.6). The equation of state look-up table assigned to this study does not go beyond temperatures of 900 °C. Therefore any node with a temperature above 900 °C is initially set to 900 °C. This adjustment primarily affects nodes in the uniformly shocked hemisphere of radius \( R_p \)—a shallow region from which heat is assumed to rapidly advect out of the system.

In reality, the melt volume produced by the impact event contributes additional heat to PIH systems. Melt volume has been found to depend on crater size and this relationship is consistent when scaled by gravity for terrestrial and lunar craters (Melosh, 1989; Cintala and Grieve, 1998). Cintala and Grieve (1998) use scaling laws and the lunar cratering record to relate the volume of melt to the transient-cavity diameter (Eq. (19)):}

\[
V_m = c D_t^d
\]

where \( V_m \) is the melt volume (km^3), \( D_t \) the transient crater diameter (km), and \( c \) and \( d \) are constants that for a chondritic impactor traveling at 10 km s\(^{-1}\) assume the values 1.08 \times 10^{-4} \text{ and 3.85, respectively.} \ The 45 km and 90 km craters have respective transient crater diameters of ~27 km and ~54 km which yield melt volumes of ~35 km\(^3\) and ~504 km\(^3\).

We do not include a treatment of melt heating in our simulations because these volumes would fill the respective floors of the 45 km and 90 km craters to a depth of 22 m and 79 m—a depth that would conductively cool in less than 20 and 300 years, respectively, and are shallower than the node height (\( \Delta z = 200 \text{ m} \)) used in our simulations. Abramov and Kring (2005) did not include melt in their simulation of a 30 km diameter crater but they did include a melt sheet in simulations of 100 km craters. In Abramov and Kring (2005) simulations of 100 km craters, the melt sheet conductively supplied heat to the PIH system and remained impermeable for 20,000 years. Relative to system evolution, this is a long period of time. However, a time of 20,000 years is likely a consequence of an overestimation of melt production due to the vertical resolution of their model, 500 m, which has a conductive cooling time of at least 10,000 years. By not including the additional heat supplied by the melt sheet our simulations may underestimate the time it takes for the system to freeze. However, our geochronically significant result, that flow is likely enhanced at the central peak, remains viable. Because a melt sheet pools around the central peaks of complex craters, an initially impermeable melt sheet may further enhance flow at the central peak early in system evolution.

Figs. 11 and 3 plot temperature contours which define the shock-heating and central geothermal uplift generated by an impactor with a radius of, \( R_p = 1900 \text{ m} \), a velocity of, \( v_s = 7 \text{ km s}^{-1} \), and density of, \( \rho_i = \rho_o = 2600 \text{ kg m}^{-3} \). This bolide carries \( 1.5 \times 10^{21} \text{ J} \) of kinetic energy and produces a transient crater diameter of 27 km, an excavation depth of 3333 m and a final crater diameter of 45 km. For the 90 km case, bolide density and velocity remains the same but the impactor radius is set to 4700 m. This increases the kinetic energy to \( 2.1 \times 10^{22} \text{ J} \), \( D_t \) to 54 km, and \( h_{\text{exc}} \) to 6666 m.

References


