Cooling of the Kärdla impact crater: II. Impact and geothermal modeling

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Abstract—Impact and geothermal modeling was performed to explain hydrothermal alteration in a 4 km marine complex crater at Kärdla, Estonia. The impact modeling was used to simulate the formation of the crater and the post-impact temperature distribution in crater environment. The geothermal modeling accounted for coupled heat transfer and multi-phase fluid flow in a variably saturated medium. The modeling results suggest that strong convective fluid flow was initiated. During the first stage, the cooling was rapid due to the effect of the latent heat of vaporization, which efficiently decreased the temperature to the boiling point. The modeling results are consistent with geological observations.

INTRODUCTION

Meteorite impacts and nuclear explosion cratering are events with large amounts of energy being transformed into heat. The physics of the energy release during the formation of these two types of craters are similar. Because military studies of the cratering process were started about 60 years ago, advances in understanding its mechanisms have been considerable. As a result, numerical elastic-plastic modeling tools, which are also applicable in meteorite impact cratering studies, have been developed (e.g., Amsden et al. 1980; Thomson 1979). In recent years, interest in the thermal aspect of cratering has been growing. Temperature is one of the key parameters in equations of state (EOS) and thus also in impact modeling. Although a lot of work is still needed, especially experimental work on the behavior of porous materials, new EOS for different materials allow, for example, estimates of melt production in large craters (O’Keefe and Ahrens 1977; Pierazzo et al. 1997). However, much less attention has been paid to dissipation of heat in craters. A proper understanding of processes in the cooling crater contributes to our knowledge of how ores formed (e.g., in Sudbury, Ivanov and Deutsch 1999a) and hypotheses on the origin and early evolution of life (e.g., Cockell and Lee 2002).

Most of the early attempts to estimate thermal history of craters or the lifetime of possible impact-induced hydrothermal systems used the approximation of a laterally infinite sheet representing a melt or hot suevite layer. Usually, the upper and lower boundaries of the sheet are kept at a constant temperature, allowing diffusion of heat both upward and downward. Naturally, this only provides a rough approximation that holds for the outer part of the crater, where during crater modification hot material is emplaced on top of rocks that have not been heated much by the shock wave. Such an assumption is oversimplified because it describes cooling only in a limited area. In the central part of the crater, the rocks underneath the melt/suevite layer are also heated and, consequently, the cold lower boundary condition is obviously not valid. In addition, even if the heat diffuses downward, which it certainly does, it accumulates at greater depths and it still has to pass the upper layers and escape from the system through the surface at a later time, thus extending the cooling time.

Turtle et al. (2003) made an integrated model for the Vredefort structure, South Africa. As first step, the impact heating was numerically simulated to get the post-impact temperature distribution within the crater. These temperatures were then used as a basis for conductive heat transfer modeling. Unfortunately, the authors provide only a few details on the course of cooling by presenting temperatures immediately after collapse, 0.3 Ma after the impact, at which point the highest temperatures near the surface have disappeared, and 30 Ma years after the collapse, when an equilibrium temperature profile has been achieved. It is difficult to evaluate the time of equilibration because it has not been defined. It could be a few percent of the initial heating (tens of degrees) or equilibrium in the numerical sense, meaning, for example, a small change in temperature during each transient time-step (e.g. 1 µK over 100,000 years). Due to the diffusive nature of conductive heat transfer,
the cooling rate becomes smaller and smaller with time and different definitions of thermal equilibrium may be achieved over a very long period of time. Nevertheless, the assumption of conduction as the only heat transfer type is likely to be valid for a crustal-scale thermal model because convection at great depth is probably inhibited due to the low permeability of rocks. Furthermore, radiative heat transfer can be easily masked by uncertainty in thermal conductivity and its pressure-temperature dependencies.

Although convection is suppressed at great depth,
geological evidence indicates that hydrothermal systems have developed in many terrestrial craters that vary in size from large (Sudbury, diameter $D = 250$ km; Allen et al. 1982) to relatively small (Lonar, $D = 1.8$ km; Hagerty and Newsom 2003). A recent paper by Abramov and Kring (2004) presents the results of numerical modeling of impact-induced hydrothermal convection at Sudbury. The model takes into account the heat of crystallization for the melt layer, heat transfer by conduction, and convective heat transfer by single- or two-phase fluid flow in the saturated porous medium. It also takes into account the reduction of permeability at temperatures above the brittle/ductile transition ($360$ °C). Indeed, the transition from convective to conductive heat transfer has been observed in present-day geothermal fields (e.g. Muroka et al. 1998), but further investigation is needed to determine whether these conditions (collapse of pores and fractures) are applicable to impactites, which often still preserve shock-produced large cavities and caverns. The permeability value of $10^{-11}$ darcies ($10^{-23}$ m$^2$) assumed by Abramov and Kring (2004) for rocks at temperatures above $500$ °C is at least two to three orders of magnitude lower than those usually reported for ductile crustal rocks (Manning and Ingebritsen 1999; Ingebritsen and Manning 1999).

Their model results might suffer from poor resolution, which is well demonstrated for the case of reduced breccia thickness. If the thickness of the high-permeability layer equals a single grid cell, then it cannot properly accommodate both the basal thermal boundary layer and the vertical flow part of the convection cell. As a result, the heat removal from the top might be underestimated because of numerical reasons. Nevertheless, Abramov and Kring (2004) demonstrate convincingly that convective heat transfer plays an important role in the cooling of the Sudbury impact crater, and the long-lived hydrothermal system causes hot upwellings and overall cooling from the outside inward.

Rathbun and Squyres (2002) attempted to model impact-induced hydrothermal systems in hypothetical Martian craters where environmental conditions differ from conditions on Earth because the availability of water for maintaining long-period hydrothermal circulation systems is questionable on Mars. Nevertheless, their models might have underestimated the post-impact temperatures, as it is difficult to agree with their assumption of impact heating of only about few tens of degrees at the center of a $7$ km diameter crater. Low heating might also be the likely reason why the simulated flow rates of free convection in case of simple crater are about $4$ orders of magnitude lower than those of forced convection.

Geological evidence of hydrothermal alteration is observed in the Kärdla impact crater, Estonia (Kirsimäe et al. 2002, Versh et al. 2005). The aims of this paper are to study the time scale and contribution of different thermal processes involved in the cooling of the Kärdla crater by means of numerical heat and multi-phase fluid transfer modeling. In order to estimate heating by the impact and the post-impact temperature distribution, numerical impact modeling of the Kärdla crater has been performed.

**GEOLOGICAL SETTINGS**

The Kärdla meteorite impact structure, $4$ km in diameter, formed in an Ordovician shallow sea shelf about $455$ million years ago. The target was composed of granitic basement overlain by $120$ m of Cambrian porous siliciclastic rocks and $20$ m of Ordovician limestones. The structure is practically uneroded and hosts a complete sequence of impactites and post-impact sediments in an annular trough. Although extensively drilled at the NE rim, the inner structure of the crater is based on only three deep boreholes, which extend to the allochthonous breccia layer (Fig. 1). Two of them (K-18 and K-1 at crater radial distances of $250$ and $450$ m, respectively) prove the existence of a central uplift and, thus, the complex nature of the crater. The seawater backsurge created at least two gullies (Suuroja et al. 2002) and deposited a sediment layer at least $150$ m thick (previously partially interpreted as slumped breccia) that primarily consists of redeposited sedimentary rocks originating outside of the crater. It also contains blocks of allochthonous breccia and fractured basement, which were derived from the rim and probably from the central uplift as well.

Impact-heated rocks in Kärdla were hydrothermally altered as described by Versh et al. (2005) in detail. A parasequence of at least eight hydrothermal phases has been recognized that covers a full range from initial high-temperature to late stage low-temperature associations. The alteration sequence includes secondary K-feldspar, occurring in two morphological generations; Fe-chlorite/corrensite; secondary quartz; three morphologically, chemically, and structurally distinctive generations of calcite; dolomite; chalcopyrite/pyrite; hematite and goethite.

Mineralogical data allow for estimates of the post-impact temperatures. The data, however, are limited because there are only three boreholes inside of the crater that extend to allochthonous breccia layer or deeper. The post-impact temperature estimates are therefore available only for radial distances of $250$, $450$, and $700$ m (boreholes K-18, K-1, and K-12, respectively). They show both vertical and lateral variations. Vertically the highest concentration of hydrothermal minerals is found in the allochthonous breccias in the upper part of the structure, whereas radially the alteration intensity decreases away from the center. In general, the main aluminosilicate phase of the hydrothermal alteration (chloritization) occurred at temperatures from $150$ to $300$ °C (Kirsimäe et al. 2002). It was probably preceded by a precipitation of cryptocrystalline K-feldspar I at somewhat higher temperatures and followed a K-feldspar II at temperatures below $150$ °C. The homogenization temperatures of fluid inclusions in the quartz fall into the
same range with the chloritization, but also show temperatures in excess of 400 °C. At later times of hydrothermal circulation, calcite/dolomite and sulfide/Fe-oxyhydrates precipitated at temperatures below 75 °C down to ambient conditions (Versh et al. 2005).

**IMPACT MODELING**

The formation and modification of the Kärdla impact crater was numerically simulated using the modified version (Ivanov et al. 1997; Ivanov and Deutch 1999b) of the two-dimensional hydrocode SALE (Amsden et al. 1980). The target in Kärdla consisted of four layers as described above, but in the simulations we assumed only a non-layered granite target because 1) no usable equation of state exists for high-porosity sandstone and/or siltstone, and 2) the limestone layer (about 20 m) and water layer (about 100 m) were too thin to essentially affect the simulation results. An analytical equation of state (ANEOS; Thompson and Lauson 1972) for granite was used to describe the pressure and temperature versus density and internal energy of the projectile and the target over a wide range of parameter values. In its central part, the model was discretized to a 10 × 10 m grid.

The size of the projectile and the acoustic fluidization (AF) parameters were determined by trial-and-error to fit the simulated crater diameter and central uplift height to the observed features of the Kärdla structure. A granite projectile of 300 m diameter with vertical velocity of 15 km s⁻¹ was assumed. This vertical velocity corresponds to an actual velocity of 21.2 km s⁻¹ at the statistically most probable impact angle of 45°. In order to simulate the formation of a central peak by vertical uplift of basement rocks, a relatively low effective kinematic viscosity (lim⁰ = 1.6 × 10⁴ m² s⁻¹) and decay time of the block oscillations (T_so = 8 s) were required. An oscillation period of 0.1 s was used. More details about scaling of the AF model parameters are presented in Melosh and Ivanov (1999), and in Wünnemann and Ivanov (2003).

According to the modeling results, the maximum crater depth (1.2 km) was reached in 6–7 s, after which the central uplift started to rise (Fig. 2). However, at the same time the excavation in the radial direction continued, and the final crater diameter was achieved about 20 s after the impact. The central uplift continued to rise until 30–35 s after the impact and then collapsed to some extent.

The simulation results suggest that impact-heated rocks are found at two positions. First, a hot layer of brecciated and probably partly melted rocks that are influenced by impact to different degrees. It is formed near the free surface during the crater excavation and modification stages and it extends outside the crater as ejecta blanket. In Kärdla, this layer corresponds to allochthonous breccias lying directly on top of autochthonous breccia. This allochthonous breccia layer is the only location where PDFs in quartz have been found so far (Suuroja 1999; Preeden 2002; Puura et al. 2004). Second, material of the central mound and its closest neighborhood are heated up during shock wave passage and subsequent decompression.

During the first 1/1000th of a second after the impact, pressures up to 200 MPa and temperatures of 14,000–15,000 K occurred at the projectile-target contact. Such extreme conditions were very limited both in space and time. Most of the rocks in the crater area experienced more moderate conditions. The highest simulated post-impact...
temperatures, which slightly exceed 800 °C, were found in the near-surface part at the center of the uplift (Fig. 3). However, at a radial distance of only 250 m, the temperatures in the uppermost 80 m (corresponding to present day depths of 350–430 m in the borehole K-18) are in the range of 630–650 °C. At a distance of 450 m, temperatures decrease from 480 °C near the surface to about 250 °C at a depth corresponding to the lowermost part of borehole K-1. Temperatures of about 300 °C were simulated at the location of borehole K-12, which penetrates only about 20 m to allochthonous breccias. At the rim the near-surface temperatures were about 100 °C, but further away from the crater temperatures up to 200 °C may have existed in the shocked ejecta fragments and the corresponding deposits.

**HEAT AND MASS TRANSFER SIMULATIONS**

**Governing Equations and Numerical Procedure**

Coupled multi-phase flow and heat transfer were solved using a fully implicit 2D axi-symmetric finite difference code. For each component k (water, air) in phase ψ (liquid, gas) the Darcy flow \( U \) (m s\(^{-1}\)) is described by (de Marsily 1986; Pruess et al. 1999):

\[
U_\psi = \frac{k_k}{\mu_\psi}(\nabla p_\psi + \rho_\psi g \nabla z) = -\frac{k_k\rho_\psi g}{\mu_\psi}(\nabla h_\psi + \rho_r \nabla z) \tag{1}
\]

where \( k \) is permeability (m\(^2\)), \( k_r \) is the unitless relative permeability, \( \mu \) is viscosity (Pa s), \( P \) is pressure (Pa), \( \rho \) is density (kg m\(^{-3}\)), \( g \) is the acceleration of gravity (m s\(^{-2}\)), \( P_\psi \) is the pressure head of reference density fluid \( \rho_0 \), and \( \rho_r = (\rho_\psi - \rho_0) / \rho_0 \). The mass flux \( q_\psi \) (kg m\(^{-2}\) s\(^{-1}\)) is:

\[
q_\psi = \rho_\psi U_\psi = -\rho_0 K_\psi (\nabla h_\psi + \rho_r \nabla z) \tag{2}
\]

with \( K_\psi = k_k \rho_\psi g / \mu_\psi \) as a hydraulic conductivity tensor (m s\(^{-1}\)). The mass accumulation term in a compressible porous medium is

\[
\rho_0 S_{\psi \psi} X_\psi^r \frac{\partial h_\psi}{\partial t} = -\nabla q_\psi X_\psi \tag{3}
\]

where

\[
S_{\psi \psi} = \rho_\psi g (\alpha + \varphi \beta_\psi) \tag{4}
\]

is the specific storage coefficient (m\(^{-1}\)), \( t \) is time (s), \( X \) is a unitless mass fraction of fluid phase ψ, \( s \) is a unitless fraction of porosity occupied by fluid, \( \varphi \) is unitless porosity, and \( \alpha \) and \( \beta \) are compressibilities (Pa\(^{-1}\)) of rock and fluid, respectively.

After defining fluid pressure head in phase \( \psi \) as \( h_\psi = h_0 + h_{\psi \psi} \) with \( h_0 \) and \( h_{\psi \psi} \) as the pressure heads of reference fluid and the capillary pressure head, respectively, the total mass balance is obtained by summing accumulations over all components and phases. The capillary pressure head of the water phase is calculated using the van Genuchten function (van Genuchten 1980; Pruess and Garcia 2002):

\[
h_c = \frac{1}{\alpha_{vG}} (s^{1/\beta_{vG}} - 1) \tag{5}
\]

where \( s \) is taken equal to water saturation by assuming full and irreducible saturations of 1 and 0, respectively. The fitting parameters of the van Genuchten function \( \lambda_{vG} = 0.55 \) and \( \alpha_{vG} = 2.5 \) have been assumed. The relative permeability is dependent on saturation of phases and was calculated using the Brooks-Corey-Burdine model (Chen et al. 1999).

The change in saturation accounting for accumulations of all phases is:

\[
\phi \rho_0 \frac{\partial s_\psi}{\partial t} = -\nabla q_\psi X_\psi \sum s = 1 \tag{6}
\]

Thermal equilibrium is described by

\[
\frac{\partial Q}{\partial t} = \nabla [(\lambda \nabla T - \sum c_\psi q_\psi X_\psi s_\psi) T] + L_h (q_l s_l + q_g s_g) X_1 + H \tag{7}
\]

where \( \lambda \) is thermal conductivity tensor (W m\(^{-1}\) K\(^{-1}\)), \( T \) is temperature (°C), \( c \) is specific heat capacity (J kg\(^{-1}\) K\(^{-1}\)), \( L_h \) is pressure-dependent latent heat of vaporization (J kg\(^{-1}\)), and \( H \) is radiogenic heat production (W m\(^{-3}\)). Subscripts \( l \) and \( g \) denote liquid and gas phases, respectively. The amount of energy per unit volume (J m\(^{-3}\)) is

\[
Q = \left[ \sum (\phi \rho_\psi c_\psi X_\psi) + (1 - \phi) \rho_s c_s \right] T \tag{8}
\]

with subscript \( s \) standing for solid rock. If phase change of the fluid in a grid cell occurs, the temperature in this cell is forced to the boiling point temperature until a single phase remains.

Water and steam properties (density, viscosity, compressibility, thermal conductivity, heat capacity, and enthalpy) were calculated as functions of pressure and temperature using “IAPWS Industrial Formulation 1997 for the Thermodynamic Properties of Water and Steam.” Thermal conductivity is the porosity-weighted arithmetic mean of that of rock matrix and fluid. In addition, the thermal conductivity of rock matrix is assumed to be temperature-dependent (Kukkonen and Jõeleht 1996):

\[
\lambda_m = \frac{\lambda_r}{1 + B(T - T_r)} + C(T + 273.15)^3 \tag{9}
\]

where \( \lambda_r \) is matrix thermal conductivity at room temperature.
The highest porosity and permeability values were applied to resurge deposits \((20\%, 10^{-12} \text{ m}^2)\), whereas the allochthonous breccia and fractured basement were assumed to have gradual decrease of permeability and porosity with depth to 1% and to \(10^{-17} \text{ m}^2\).

\(T_r = 20 \degree C\). Parameter \(B = 0.0015\) accounts for the decrease of lattice conductivity with an increase of temperature due to thermal expansion. Parameter \(C = 5 \times 10^{-11} \text{ K}^{-3}\) accounts for the radiative heat transfer component, which starts to contribute significantly at temperatures exceeding 800 \degree C.

To avoid numerical instability, the II’ in scheme (Clauser and Kiesner 1987; Clauser 2003) was applied to the diffusion-convection Equation 7. The II’ in scheme uses local upwinding and is second-order accurate and unconditionally stable.

The fluid flow and heat transport equations are solved in a double-iterative manner. First, a solution of the fluid flow is iteratively obtained. Then, the updated hydraulic head and fluid flow values are used in solving the heat transfer. Updated temperature values are used in the next cycle for updating hydraulic parameters. For steady-state calculations, the hydraulic-thermal cycle is repeated until the maximum temperature change during the cycle is below 1 mK. The convergence criteria for head and temperature fields are 10 \(\mu\)m and 10 \(\mu\)K, respectively. For transient calculations these convergence criteria are 1 \(\mu\)m and 1 \(\mu\)K, respectively. Both fluid flow and heat transfer equations are solved using the “strongly implicit procedure” (Stone 1968). The length of the following time step in the transient calculations is adjusted automatically similarly to the HST3D code (Kipp 1987), so that the maximum change of head, temperature or fluid phase would not exceed 0.1 m, 0.1 K, and 5%, respectively.

**Model Space**

The model space, 3.0 km in radius and 3.08 km in depth, was discretized with a 20 × 20 m grid and assumed axial symmetry. For each grid cell porosity, permeability, thermal conductivity, and radiogenic heat production values were applied. Porosity and permeability distributions (Fig. 4) are based on available petrophysical data (Plado et al. 1996; 2002; Jõeleht 1995). Relatively high porosity (12%) was assumed for the allochthonous breccia layer. In the deeper, autochthonous part, the porosity decreases gradually from 5% underneath the allochthonous breccia layer to 1% in more distant areas.

The post-impact water saturation in the breccia layers and fractured basement is linked to porosity. The impact process causes fracturing and enhances porosity, but in our view, it does not bring additional water to the rocks either during the shock wave passage or the excavation stage. Thus, rocks have the amount of water corresponding to pre-impact porosity of crystalline rocks (here assumed to be 1%) and the saturation in impactites is inversely proportional to the post-impact porosity (up to 12%). The rest of the pore space is filled by air. The resurge sediments are assumed fully saturated. The post-impact steady-state hydrostatic hydraulic head was used at the start of transient heat and mass transfer modeling.

Compared to porosity, the permeability of the impacted rock units is less well constrained, as there are only a few pumping test data available for Kärdla. The tests in the fractured basement rocks in the north-eastern rim (Kala et al. 1976) yielded permeabilities in the range of \(10^{-12}−10^{-12} \text{ m}^2\), which is less than an order of magnitude lower than that of overlaying Cambrian siliciclastic rocks. Pumping in the allochthonous and autochthonous part of the borehole K-18 (at depths of 350–430 m) gave similar results \((10^{-12} \text{ m}^2\), Tassa and Perens 1984). Because other deep permeability data are not available, we assumed that permeability decreases gradually with depth similarly to porosity, from \(1 \times 10^{-13} \text{ m}^2\) at the allochthonous breccia to background values of the basement \((1 \times 10^{-17} \text{ m}^2)\) in the more distant areas.

The impact modeling code is not able to account for the resurge flow, yet. The flow, however, did take place and eroded the crater rim and seabed outside of the crater. Resurge sediments are at least 170 m thick (in borehole K-1) and are mainly composed of sedimentary rocks with blocks of allochthonous breccia from the rim and central uplift. Resurge deposits were included in the model by assuming their initial post-impact temperature to be that of seawater (20 \degree C) and assigning a permeability of \(1 \times 10^{-12} \text{ m}^2\), which is typical for the Cambrian sedimentary rocks in the surrounding area.

The thermal conductivity of the allochthonous breccia \((2.6 \text{ W m}^{-1} \text{ K}^{-1})\) is lower than that of the autochthonous breccia and basement \((3 \text{ W m}^{-1} \text{ K}^{-1})\) due to increased porosity (Jõeleht and Kukkonen 1996; Plado et al. 2002). A radiogenic
Fig. 5. Simulations suggest that the borehole sites were dominated by liquid water already during the first decade after the impact, but at the central uplift above, boiling point temperatures existed for almost 100 years. The hydrothermal flow continued for several thousands of years and began to cease 9,000–10,000 years after the impact. Temperature isolines are drawn every 50 K, with the outermost contour of 50 °C. Mass flow rates of liquid water (black arrows) and vapor (white arrows) are drawn at the same lognormal scale (bottom right of each figure), with the largest arrow corresponding to $10^{-3}$ kg m$^{-2}$ s$^{-1}$ and the shortest to $10^{-7}$ kg m$^{-2}$ s$^{-1}$; $r$ and $z$ indicate radial and vertical dimensions of the model space, respectively.
heat production of 1 \( \mu \text{W m}^{-3} \) was assumed for all rock units (Jõeleht 1995).

A constant temperature (20 °C) and hydraulic head (100 m above the pre-impact seabed) were applied to grid cells corresponding to the seabed. A constant heat flow density (50 \( \mu \text{W m}^{-2} \)) was applied to the lower boundary and a no-flow condition was used at lateral sides.

**Cooling History**

The cooling history of the Kärdla crater can be divided into three periods that were successively dominated by 1) boiling heat effects, 2) convection, and 3) conduction.

The first and the shortest period of cooling was strongly influenced by convective heat transfer with boiling heat effect. The diameter of the vapor-dominated zone decreased to half of its initial 1.6 km size in about 5 years (Fig. 5). The drop of temperature to the boiling point (250–290 °C depending on hydrostatic pressure) took place in a few tens of years in the upper part of the central uplift (permeable allochthonous breccia and upper portion of autochthonous breccia) and lasted for less than a hundred years in the deeper part.

During the first period, the driving force of the convective flow is the large density contrast between liquid water and air/vapor. The simulated mass flow rates of liquid water in the upper part of the section are as high as 10\(^{-3}\)–10\(^{-5}\) kg m\(^{-2}\) s\(^{-1}\). Although these flow rates indicate a strongly convecting environment, the effect of boiling heat overrides all other thermal effects. The latent heat of vaporization of fresh water at hydraulic pressure conditions in Kärdla is 1.3–1.7 MJ kg\(^{-1}\). The vapor. The simulated mass flow rates of liquid water in the upper part of the section are as high as 10\(^{-3}\)–10\(^{-5}\) kg m\(^{-2}\) s\(^{-1}\). Although these flow rates indicate a strongly convecting environment, the effect of boiling heat overrides all other thermal effects. The latent heat of vaporization of fresh water at hydraulic pressure conditions in Kärdla is 1.3–1.7 MJ kg\(^{-1}\), which corresponds to about 500 K temperature change in the same volume of typical rocks. As a rough estimate, for a rock with 5% porosity that is initially filled with dry air, only the amount of about three pore volumes of water is required to be vaporized in order to decrease the bulk temperature by 100 K. However, in rocks with 12% porosity (such as the allochthonous breccia), only about 1.2 pore volumes of water is needed to be vaporized for the same change in temperature. The high-enthalpy steam flows upwards and discharges energy to the near-surface layers and sea. Since relatively small flow rates can move the vaporization front, the gas phase in the central part of the crater remains poorly saturated with respect to water vapor and thus the rocks were probably not much affected by alteration.

The second period, which started after the most efficient heat removal process has stopped, lasted considerably longer. Nevertheless, heat transfer in the central part of the crater is still dominated by convection. Flow rates continued to remain high because the crater rocks were not fully saturated yet and there was much trapped air that flowed buoyantly toward the surface. Even in the liquid phase saturated medium, the 20% density difference between the recharging cold water and hot water that is close to boiling point causes strong convection as well. The influence of relative permeability, which reduces flow rates in partially saturated media, decreased gradually with time as rocks became more and more saturated with liquid water.

The transition from a system dominated by convective heat transfer to a conduction-dominated system took place 3000–4000 years after the impact. This transition occurs when the flow system Peclet number (the ratio of heat transferred convectively over that transferred conductively) decreases below 1. From a numerical point of view, this means that the sum of the vertical grid Peclet numbers (Clauser and Kiesner 1987) decreases below 1. Although convective heat transfer stops dominating at this point, groundwater flow still continues for a long time and thus may produce low-temperature hydrothermal mineralization.

After some 9000–10,000 years, the impact-induced flow system has ceased. However, thermal equilibration takes a much longer time. Even though the maximum temperature disturbance after 10,000 years is only about 15 K, the heat flow density is two times higher than in surrounding areas and the anomaly could have been easily detected with modern geothermal techniques. The heat flow density anomaly becomes instrumentally undetectable 40,000–50,000 years after the impact.

The above results mainly describe the cooling of the central uplift and its vicinity. The cooling rates and the lengths of respective periods, however, vary in different parts of the crater. Figure 6 shows a family of curves representing the temperature above steady state versus time in grid cells of the topmost one kilometer of the model at different radial distances. At these distances (excluding 0-distance i.e., the crater’s center) borehole data are available for comparison. It is evident that cooling to ambient temperatures at the radius of the crater depression (r = 700 m), which in Kärdla has been the main recharge area for the hydrothermal system, was very rapid (Fig. 6). At the slope of the central uplift (the approximate locations of boreholes K-18 [r = 250 m] and K-1 [r = 450 m]) cooling to the boiling point was also very quick (a few years to a few tens of years), but then the temperatures remained more or less constant for a while. This can be attributed to strong convective upward flow of fluids. The flow velocity is so high that partial boiling may occasionally occur as the pressure and boiling point temperature decrease upwards. At the crater’s rim (r = 2000 m), the bulk temperatures have not been high, but cooling to ambient conditions took at least few hundred years.

**DISCUSSION AND CONCLUSIONS**

Numerical models are only approximations due to ambiguity in the parameters and processes involved. These uncertainties exist both in impact and thermal models. In impact modeling, for example, we were not able to incorporate water and sediment layers. Possibly, addition of
the water layer to the model will not change the post-impact temperature distribution much. However, there is no consensus on the sea depth at the crater site at the time of impact. Different authors estimate sea depth ranging from 20 m (Puura and Suuroja 1992; Ormø and Lindström 2000), to over 50 m (Ainsaar et al. 2002; Suuroja et al. 2002), and up to 100 m (Lindström et al. 1992; Suuroja 1996). Suuroja et al. (2002) admit that the particular type of bioclastic argillaceous-calcareous sedimentation that occurred before and continued after the impact might have taken place at a depth interval from tens to about 200 m. Nevertheless, even in the latter case, the sea depth was less than the size of the projectile, and thus it was not capable to decrease the velocity of the projectile significantly (Shuvalov 2002).

According to laboratory experiments, the impact heating of dry porous sandstones is usually higher than that of non-porous rocks due to compaction-related frictional heating (Ahrens and Gregson 1962; Kieffer et al. 1976). However, it is difficult to predict the thermal behavior of water-saturated porous rock because of massive pore-water vaporization. A

Fig. 6. Temperature above steady state versus time. Each line shows the change of temperature in grid cells of the uppermost 1000 m. The data represents the uppermost 1000 m of the model at radial distances of 0, 250, 450, 700, and 2000 m.
better post-impact temperature estimate requires 3D modeling with proper EOSs and the resurge flow included. This task extends beyond the aims and capabilities of this work.

The seawater depth has only a small effect on thermal modeling. Even for the shallowest estimates, there was still at least 200 m of water column inside of the crater. Compared to the obtained results above, a 100 m lower water column decreases the boiling temperature by a few tens of degrees and the latent heat of vaporization increases by about 10%, whereas the effects due to a 100 m higher water column are even smaller. Although sea depth variations have a minor effect inside the crater, they might be more important at the rim. If the highest parts/peaks of the rim rose above sea level, they became groundwater recharge areas instead of being places where hot water discharged to the sea. However, there

Fig. 7. The same as in Fig. 6, but for the model without convection.
is no evidence to either support or reject the hypothesis of flow direction reversal.

Permeability is the key parameter in hydrothermal convection systems because it controls fluid flow rates in the rocks. In Kärdla, the permeability data are insufficient for a full description of the crater rocks. The values we have used provide only a qualitative picture of the actual system. Furthermore, permeabilities used in the model characterize the present-day situation, which incorporates permeability reduction due to compaction and closure of pores/fractures by mineral precipitation. We, however, do not account for a temporary decrease of permeability at temperatures above brittle/ductile transition as it is described by Abramov and Kring (2004), because the main part of the pores/caverns and fractures of the allochthonous and autochthonous breccias are due to impact deformation of the rocks and have not been erased by the ductile behavior of rocks. This ambiguity in permeability is still great due to the relative permeability that reflects obstructed flow conditions in an air/liquid-saturated medium. Consequently, the flow rate depends on the initial post-impact pore pressure and water saturation of the rocks, factors that we could only guess. The modeling, however, suggests that an exact knowledge of the initial saturation is not crucial to the final results. Initially, the highest liquid water flow rates occur at the hydrothermal systems’ recharge area in the upper part of the section, where the intrinsic permeability and water saturation are also highest. Although the suction due to the air/water density contrast is large, the relative permeability forces the liquid water flow to be more intrusion-like because the flow rates within the intrusion are larger than at its edges. From the thermal point of view, the permeability variation can change the timing of transitions from the vaporization-dominated stage to convection-dominated, and to conduction-dominated stages, but it cannot cancel the existence of these stages.

Nevertheless, a proper knowledge of the permeability structure is important because it can modify the shape of the entire flow system. The present results are based on the assumption that there is a quite strong gradual decrease of permeability with depth (Fig. 4). Therefore, the upper part of the section cools down faster than the deeper section. However, additional test models suggest that if the decrease of permeability with depth is smaller, then the flow system becomes deeper. This implies that the deeper part of the central uplift becomes vapor-free earlier than the upper part, which is heated by rising steam. Supposedly, these two cases should be mirrored by differences in the intensity of hydrothermal alteration and the isotopic composition of the precipitates, but the geological information gathered so far is insufficient to prove it.

Two initial conditions in the cooling modeling, the water saturation and hydraulic head, have been guessed. The water saturation is based on an assumption that no additional water is brought into the impact-created pores during the crater formation and the rocks contain the same amount of water that existed in the rocks before the impact. In our view, it is unlikely that water enters into the pores not only during the shock wave passage but also during the excavation stage because, according to impact modeling in shallow marine conditions, the cavity expands faster in water than in rocks (Shuvalov 2002) and no source for additional water exists. We assumed that the rest of the pore space is filled by air. Although the latter is probably true for the near-surface layers, it is questionable for the deeper part of the crater because there might not have been enough time for air to flow to the deeper sections during the craters’ formation and modification stages. It could mean that a limited amount of air still existed in pores of the deeper part of the structure but pore fluids were in underpressure conditions. This is the opposite of our transient modeling assumption of hydrostatic hydraulic head as a starting condition that more likely suggests overpressure conditions in the imprints. The hydrostatic pressure is, however, valid for the laterally distant areas and probably also for the deep sections that were not affected by the impact.

The modeling results indicate that the hydraulic quasi-equilibrium in the deeper part of the model is achieved in about 1–1.5 years. This is the time required to get the excess air out of the model and represent the uncertainty associated with the saturation process. Although there might not be air that must escape from the pores before water can intrude, the water phase is discontinuous and a relative permeability effect is applicable to such an environment. We estimate that the error due to an improper hydraulic starting condition is much smaller than the ambiguity due to other parameters such as permeability or relative permeability.

In order to demonstrate the importance of convective heat transfer, a model with only conductive heat transfer (Fig. 7) has been carried out. In this case, the simulated cooling period is much longer, and within the depth range of the autochthonous breccias temperatures above 300 °C could have lasted for thousands of years. It is, however, noticeable that the initially hottest part of the section (corresponding to the allochthonous breccia) cooled by several hundreds of degrees during the first century. The cooling of the allochthonous layers would be even more rapid if resurgence deposits did not cover them.

Regardless of the ambiguity in parameters, the model results are broadly consistent with mineralogical observations (Versh et al. 2005), although, at the first sight, the simulated post-impact temperatures are too high at the center. At the radial distance of 250–450 m, simulated temperatures of up to 500–650 °C would allow formation of a high-temperature hydrothermal mineral assemblage made of garnet, actinolite and/or epidote, but so far, these minerals have not been identified in drill cores K-18 or K-1. Kirschmäe et al. (2002) proposed that the absence of these mineral assemblages is because the initial high-temperature stage (>300 °C) was too
short for alteration to achieve equilibrium for high temperature phases. Numerical simulations support this hypothesis, but also suggest that the low water saturation of the air phase may have played an important role. Initially, the rocks were filled with dry air, which probably inhibited the hydrothermal alteration until the air became more saturated by water vapor. This happened only in the vicinity of the boiling front, which moved quickly towards the center and also caused a rapid drop in temperature. Even near the boiling front, the air tends to be undersaturated because the steam is mixed with dry air that escapes from greater depths.

Based on the impact simulations, the rim is relatively unaffected thermally, and post-impact temperatures reaching the boiling point can be associated only with the allochthonous breccia either in the form of injection dikes or the ejecta layer. These results are in agreement with mineralogical observations indicating that, for example, the K-feldspar that might be related to the near-boiling conditions occurs only in allochthonous injection dikes, whereas the rest of the rock is barely affected by any hydrothermal alteration. Moreover, high temperature mineral alteration at the rim is restricted to fracture planes and grain boundaries, which also implies a very short duration of these processes. However, it is obvious that mineralogical observations at individual locations may reveal much higher (maximum) temperature estimates compared to the numerical modeling, which produces only bulk average estimates over the 20 × 20 m grid spacing used.

The cooling of crater rocks is site-specific, depending on distance from the center and depth. The fastest cooling occurs at the groundwater recharge area, which in the case of Kärdla is at the crater depression. The central uplift works like a chimney that keeps the fluid flowing toward the center and upward. This also implies that the precipitation of different hydrothermal minerals/generations occurs simultaneously at different locations. At the time when Fe oxyhydrates, calcite, or other low-temperature alteration products are formed at the crater depression area, the central part may still be at vapor-dominated conditions with high-temperature alteration products.

In conclusion, we suggest that post-impact cooling of the Kärdla crater was rapid and driven by convective heat transfer during the early phases of the development. The highest post-shock temperatures in the central area of the structure decreased by 50% from their initial values in 1200–1500 years and by 80% in less than 4000 years after the impact. The transition from a system dominated by convective heat transfer to a conduction-dominated system took place about 3000–4000 years after the impact, whereas it took about 9000–10,000 years for the impact-induced hydrothermal flow system to cease. However, the thermal equilibration of the impact-heated rocks took a much longer time and the heat flow density anomaly became instrumentally undetectable only 40,000–50,000 years after the impact.

The validity of the numerical heating-cooling model for Kärdla structure is supported by geological observations, which are well constrained both for maximum temperature and temperature field distribution. This emphasizes the strong need for linking between complex numerical modeling and geological studies.

The numerical modeling example of the Kärdla structure demonstrates the importance of convective heat transfer in the development of Earth and probably also Mars-based impact structures. The cooling rates of impact-heated rocks in a water-saturated environment can be much higher than believed so far.

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REFERENCES


