THERMAL HISTORY AND DIFFERENTIATION OF ICE/ROCK PLANETESIMALS. L. L. Farrell[1], R. S. McGary[1] and D. W. Sparks[1]. [1]Department of Geology and Geophysics, Texas A&M University, College Station, TX 77843; email: farrell1@tamu.edu.

Introduction: The objects in the Asteroid Belt appear to have undergone a wide variety of thermal histories. The density of the dwarf planet Ceres indicates an ice volume fraction of up to 0.56. Thermal models [1] and geodetic studies [2] indicate that the water phase of Ceres has differentiated. Other asteroids, such as Vesta, contain no water, indicating either a hot accretion or a heating process that led to massive water loss.

The internal thermal structures of a planetesimal are affected by heat production, heat transfer processes, and mass transfer (differentiation). Long- and short-lived radioactive isotopes are found within the silicate part of the asteroids and provide internal heating energy. The heat is then transferred throughout the body by thermal conduction, fluid convection in liquid or partially liquid regions, and solid-state convection in ice layers. Phase changes such as melting /freezing of ice and rock, and hydration/dehydration will add or subtract heat. Gravity-driven differentiation of a mobile liquid phase redistributes heat and mass and changes the radial distribution of thermal properties. Liquid water at the surface of a planetesimal is lost to space by vaporization. We have developed a simple one-dimensional numerical thermal model of the evolution of a 200-1000 km planetesimal over the history of the solar system. Our objective is to determine the significant parameters that determine the internal structure and history and such bodies.

Formulation of 1-D Numerical Model: We solve the time-dependent energy conservation equation within a spherically symmetric planetesimal. We assume cold accretion, with a homogenous rock/ice ratio and initial temperature.

Heating occurs due to radioactive isotopes present in the silicate. The initial amounts of long-lived radioactive elements are fixed for all models (using the heating rates in [1]) whereas the short-lived element, ^{26}Al, is a variable. Because the half-life of ^{26}Al is short (0.7 My), the amount incorporated into a planetesimal will depend strongly on its accretion time. We model ^{26}Al heat production rates varying from 50 to 2000 times the initial rate from ^{238}U decay. Because the radioactive elements are found within the silicate rock, the heating rate varies with the original volume fraction of silicate, ν_0, which is parameter that can be varied. We include the latent heat of melting/fusion in the ice/water phase change. In our initial models we assume that the temperature increase due to the release of latent heat by silicate hydration hastens the onset of dehydration at higher temperatures, so this phase change has not yet been explicitly modeled.

There are three important dimensionless parameters that appear our conservation equation, and each is dependent on one or more of the controlling variables affecting our asteroid formation. The initial heat production parameter, D_{Al}, is a ratio of the ^{26}Al heating relative to the heat production rate of long-lived element ^{235}U. An increase in D_{Al} speeds up the initial melting of the ice allowing higher peak temperatures to be reached. The melting parameter, D_m, describes the heat required to melt ice, relative to the heat produced by long-lived isotopes. It relies most strongly on the initial silicate fraction, ν_0. A lower ν_0 means there is less radioactive heating, and the asteroid remains icy with overall cooler temperatures. The diffusion parameter, D_d, relates the heat loss by conduction to the heat production, and is most strongly dependant on the radius of the body. For larger planetesimals, the temperatures reached are higher.

Upon heating and melting, the water migrates upward, while the remaining solid compacts until the volume fraction of rock reaches a close-packing limit for solid grains (we assume this to be at a local silicate fraction of 0.7). The thermal parameters change as the local ice/rock/water mixture changes. The thermal conductivity of water and ice layers are enhanced to approximate the effect of convection. The upward migrating water may collect at the base of the unmelted ice/rock crust, forming a global liquid layer, or interior ocean. Because the frozen crust contains rock, it is denser than the water beneath it and is gravitationally unstable, and can be held in place only by its mechanical strength. We assume that this crust will founder and sink if it thins to 2% of the radius of the planetesimal. We model this by instantly extracting the accumulated liquid layer to the surface where it is vaporized into space, leaving behind a smaller, more silicate-rich asteroid.

Hydration reactions will convert a mixture of 70% (by volume) olivine and 30% water into serpentine and magnetite. At ~600 K, serpentine will break down, regrowing anhydrous silicates and releasing water which is expelled upward [3]. We approximate this process by allowing regions that have risen to 600 K to compact to 100% silicate. This results in a dense silicate core and a second stage of water migration and interior ocean formation.

Results: The thermal history of ice/rock planetesimals, including the formation and duration of an interior liquid water later, the amount of water lost to space, and the degree of serpentinization depend sensitively on three properties of the initial planetesimal: it's size, silicate volume fraction, and the amount of short-lived heat production.

The evolution of a typical modelled planetesimal begins with the melting of interior ice. Compaction of the silicate phase expels the melt water to a liquid layer beneath an overlying still-frozen shell. When the silicate phase has compacted to its close-packing limit, meltwater is no longer expelled. The liquid layer stops growing and begins to freeze from the surface. In areas that reach the dehydration reaction temperature, further compaction occurs that again contributes liquid water and heat to the water layer. This can remelt a largely frozen initial water layer, resulting in a "second-generation" interior ocean.

The thickness of the frozen shell (and therefore the depth of the liquid water layer) as a fraction of radius decreases with size of the planetesimal. In larger planetesimals, the shell can overturn, resulting in loss of the water in the liquid layer. This can happen during the formation of the initial liquid layer (within the first few million years) or the second ocean layer (after several hundred million years), or both.

The phase plots below show the results of two suites of numerical experiments conducted with a range of ^{26}Al heating (D_{Al}), and initial radius. The initial silicate volume fraction is 0.44 in Figure 1 and 0.62 in Figure 2. Each plot shows the present day (after 4.5 billion years of model time) silicate volume fraction (contours) and liquid water fraction. Water loss occurs if the planetesimal is larger than a critical radius (which depends on the initial silicate volume fraction), or if ^{26}Al heating is large enough. Note that the scales of the shading are different in the two figures. The initially ice-rich planetesimals in Fig. 1 are predicted to contain little or no liquid water today, distributed in pore space at depth. The largest and hottest of the initially ice-poor planetesimals of Fig. 2, have enough heat that significant layers of liquid water still exist (although they have driven off much of their initial water).

These results show that the thermal history and the modern-day structure of small ice-rock planetesimals depend in complicated ways on the interplay of several processes. With this model we hope to map out the array of possibilities that we can expect to exist in throughout the solar system.

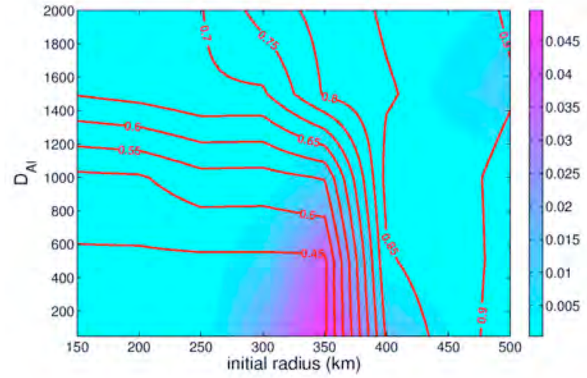

Figure 1. Present-day silicate fraction (contours) and liquid water fraction (shading) in model planetesimals in the Asteroid Belt ranging from 150 to 500 km in initial radius and with a varying amounts of ^{26}Al heating. The initial silicate fraction, ν_0, in each model is 0.44 (similar to Ceres).

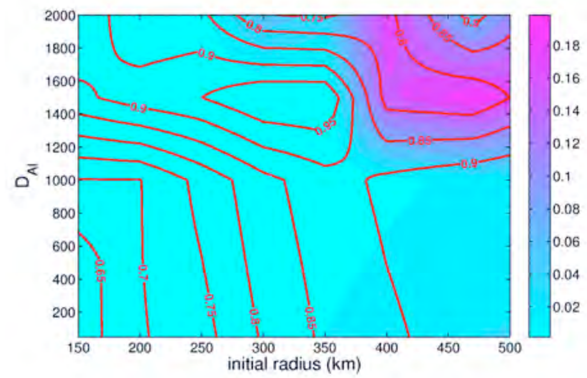

Figure 2. Same as Figure 1, except the initial planetesimal is more silicate rich, with ν_0 =0.62.

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