Thermal Evolution Models of Asteroids

Harry Y. McSween Jr.
University of Tennessee

Amitabha Ghosh
University of Tennessee

Robert E. Grimm
Blackhawk Geoservices

Lionel Wilson
Lancaster University

Edward D. Young
University of California at Los Angeles

Thermal evolution models for asteroids that experienced metamorphism (ordinary chondrites), aqueous alteration (carbonaceous chondrites), and melting and differentiation (HED achondrites) are compared. These models, based on decay of ^{26}Al, can be used to study a variety of asteroidal processes such as the insulating effect of regolith, the buffering effect of ice and fluid flow, and the complications arising from redistribution of heat sources during differentiation. Thermal models can also account for an apparent relationship between peak temperature and heliocentric distance of asteroids in the main belt. Thermal evolution models using other heat sources (electromagnetic induction, collisions) are poorly constrained at this point and have been used primarily for simple plausibility calculations.

1. INTRODUCTION

Many asteroids and the meteorites derived from them have been heated, as manifested in metamorphism, aqueous alteration, melting, and differentiation. Almost half a century ago, Harold Urey recognized that decay of long-lived radioactive isotopes (K, U, Th), the primary heating mechanism for planets, was not an effective heat source for asteroids, because the timescale for energy release is long compared to that for conductive heat loss from small bodies. *Urey* (1955) suggested decay of the short-lived radionuclide ^{26}Al and performed a back-of-the-envelope calculation of the heat produced — a precursor to the first asteroid thermal evolution model. During the next several decades, thermal models were used as plausibility tests for various proposed heat sources. More recently, thermal models have been used to describe quantitatively the geologic evolution of asteroids, thereby linking their formation to measurable parameters in meteorites.

The case for ^{26}Al heating of asteroids has become increasingly robust. Live ^{26}Al in the early solar system was widespread (*MacPherson et al.,* 1995; *Huss et al.,* 2001), and its decay product has been found in most classes of chondrites (*Lee et al.,* 1976; *Russell et al*., 1996; *Kita et al.,* 2000) and several achondrites (*Srinivasan et al.,* 1999; *Nyquist et al.,* 2001). Reasons why evidence for ^{26}Al might be obscured in other achondrites have been given (*LaTourrette and Wasserburg,* 1997; *Ghosh and McSween,* 1998). This heat source appears capable of explaining the full range of temperature excursions of asteroids within the main belt (*Grimm and McSween,* 1993). Although nebular heterogeneity of ^{26}Al has been suggested (*Ireland and Fegley,* 2000), the consistency of ^{26}Al/^{27}Al ratios in calcium-aluminum-rich inclusions (CAIs) and in chondrules, regardless of chondrite class, implies broad nebular homogeneity and indicates that differences in initial ratios reflect formation time (*Huss et al.,* 2001).

A competing hypothesis that asteroids were heated by electromagnetic induction (*Sonnett et al.,* 1968) is based on resistance to flow of electric currents induced by outflows from the young Sun. However, studies of T-Tauri stars have found that solar winds are focused at high latitudes, avoiding the nebular disk where planetesimals form (*Edwards et al.,* 1987), and mass losses, the rates of which govern magnetic fields, have been revised downward significantly (*DeCampli,* 1981). Induction models thus hinge on the choice of reasonable parameters where, as noted by *Wood and Pellas* (1991), most parameters are unconstrained. Nevertheless, several recent thermal models (*Herbert,* 1989; *Shimazu and Terawawa,* 1995) suggest that electromagnetic induction heating could melt asteroids, so in the absence of other information this heat source cannot be ruled out.

Numerous authors (e.g., *Mittlefehldt,* 1979; *Wasson et al.,* 1987; *Rubin,* 1995) have appealed to impact heating to explain metamorphism and melting in meteorite parent bodies. However, this process, by itself, cannot account for global thermal effects in meteorite parent bodies. The global temperature rise from near-disruptive collisions is no more than a few degrees, even for high-porosity asteroids with greater impact strength (*Keil et al.,* 1997). This stems from the fact that collisional energy is proportional to gravitational potential energy, which is negligible in bodies of asteroidal dimensions (*Melosh,* 1990). The high relative abundance of chondrites heated to high temperatures argues that asteroid metamorphism was a global process, unlike the low proportion of metamorphic target rocks in impact craters. However, a correlation between metamorphic grade and shock stage in chondrites (*Rubin,* 1995) may support collisional heating. Although partial melting of phases with low melting points or low shock impedance has been suggested to have produced some achondrites and iron meteorites, shock experiments and studies of impacted materials demonstrate that impact produces either total melts or localized incomplete melts on a microscopic scale that cannot segregate into pools of substantial size (*Keil et al.,* 1997).

The heat transfer equation is the basis for most model calculations (for a detailed discussion, see *Ghosh and McSween,* 1998). Three methods exist for its numerical solution: the classical series solution, the finite difference method, and the finite element method, with the latter being most accurate. By necessity, asteroid thermal models must make assumptions that address uncertainties in initial conditions (e.g, asteroid temperature at the beginning of the simulation), boundary conditions (e.g., nebular ambient temperature, asteroid emissivity), and model parameters (e.g., specific heat capacity, thermal diffusivity, presence of regolith, voids, or ice). Initial temperatures are usually constrained from nebular models (e.g., *Wood and Morfill,* 1988), and many thermal models assume asteroid accretion was instantaneous. Boundary conditions are implemented in two ways: The Dirichlet boundary condition forces the asteroid surface temperature to that of the ambient nebula, and the radiation boundary condition calculates a heat flux depending on temperature difference between the asteroid surface and the nebula. Although the radiation boundary condition is numerically unstable, it is probably more realistic. Model parameters are constrained, to the extent possible, using meteorite and asteroid data (e.g., peak temperatures, cooling rates, closure ages, ^{26}Al contents, asteroid sizes). Published asteroid thermal evolution models are briefly summarized in Table 1.

2. ORDINARY CHONDRITE ASTEROIDS AND THE EFFECT OF A REGOLITH

Construction of thermal models for the parent asteroids of ordinary chondrites (Oc) is relatively straightforward, because heat movement through these asteroids is domi-

TABLE 1. Chronological summary of published asteroid thermal evolution models.

Reference	Model
Urey (1955)	First feasibility calculation of ^{26}Al as an asteroid heat source
Sonnett et al. (1968)	First proposal for electromagnetic induction heating of asteroids
Herndon and Herndon (1977)	Feasibility study of ^{26}Al as an asteroid heat source
Fujii et al. (1979)	Comparison of internal and external heating models for asteroids
Minster and Allegré (1979)	^{26}Al heating model for the H-chondrite parent body
Wood (1979)	Model to reproduce metallographic cooling rates of iron meteorites
Miyamoto et al. (1981)	^{26}Al heating model to constrain sizes of Oc parent bodies using cooling rates, isotopic closure ages, and fall statistics
Yomogida and Matsui (1984)	^{26}Al heating model for small, unsintered asteroids
Grimm (1985)	Model of asteroid metamorphism with fragmentation and reassembly
Grimm and McSween (1989)	^{26}Al heating model of ice-bearing planetesimals, to account for aqueous alteration in Cc
Herbert (1989)	Model of electromagenetic induction heating that causes melting
Haack et al. (1990)	Thermal model of a differentiated asteroid based on decay of long-lived radionuclides
Miyamoto (1991)	^{26}Al heating model to account for aqueous alteration in Cc asteroids
Grimm and McSween (1993)	Explanation of inferred thermal stratification of the asteroid belt based on heliocentric accretion and ^{26}Al heating
Shimazu and Terasawa (1995)	Model of electromagnetic induction heating
Bennett and McSween (1996)	Updated ^{26}Al heating model for Oc asteroids, using revised chronology and thermophysical properties
Akridge et al. (1998)	Model for ^{26}Al heating of Oc asteroid (6 Hebe) with a megaregolith
Ghosh and McSween (1998)	^{26}Al heating model of HED parent body 4 Vesta
Wilson et al. (1999)	Overpressure and explosion resulting from heating Cc asteroids
Young et al. (1999)	^{26}Al heating model of Cc asteroids with fluid flow, to explain O-isotopic fractionations
Cohen and Coker (2000)	Short- and long-lived radionuclide heating model of Cc parent bodies used to study racemization of amino acids
Wilson and Keil (2000)	Thermal effects of magma migration in 4 Vesta
Ghosh et al. (2001)	Effect of incremental accretion on inferred thermal distribution of asteroids in the main belt

nated by conduction (only minor fluids were present and rock fabrics indicate no solid-state convection occurred) and rigorous model constraints are provided by meteorite data. Ordinary chondrite metamorphism occurred at temperatures ranging up to ~1175 K (*McSween et al.,* 1988), i.e., below the melting point for a eutectic mixture of metal and sulfide. Peak temperatures for highly metamorphosed (type 6) chondrites are estimated from geothermometry based on pyroxene compositions (*Olsen and Bunch,* 1984) and on crystallographic ordering in plagioclase (*Nakamura and Motomura,* 1999), and those for the least-metamorphosed (type 3) chondrites are based on thermoluminescence sensitivity (*Sears et al.,* 1980). Meteorite cooling rates are determined from measurements of the temperatures and times at which specific radiogenic isotope systems ceased to equilibrate and fission tracks ceased to anneal. The derived chondrite cooling curves (*Pellas and Storzer,* 1981) show that heating commenced at the time of asteroid accretion (consistent with ^{26}Al decay as the heat source), cooling was rapid (in a small body), and chondrites at higher metamorphic grades cooled more slowly than less-metamorphosed chondrites (implying that the asteroid interior was hotter than the near-surface regions). The thermal structure of such a body resembles an onion, with each successive layer representing a limited interval of temperature corresponding to a particular metamorphic grade.

The thermal model of *Miyamoto et al.* (1981), which incorporated ^{26}Al heating and an extensive set of thermophysical data from Oc, described a 100-m.y.-long thermal evolution of several asteroids with onion-shell stratigraphy. This thermal model was updated (*Bennett and McSween,* 1996) by incorporating refined thermophysical properties of chondrites and a shortened thermal history of 60 m.y. based on Pb-Pb isotope chronology (*Göpel et al.,* 1994). The revised H-chondrite asteroid model (the L-chondrite model is similar) is illustrated in Fig. 1. The initial chondritic ^{26}Al/^{27}Al ratio requires an interval of ~2 m.y. between the formation of CAIs (the earliest formed nebular materials) and asteroid accretion, in conformity with constraints on the timing of asteroid formation from radiogenic isotope systematics (*Lugmair and Shukolyukov,* 2001). Higher metamorphic grades in the asteroid interior reach peak temperatures later than low-grade chondrites that were closer to the surface. The bulk of the asteroid is composed of highly metamorphosed type 6 chondrites, with only thin veneers of less-metamorphosed material. A test of this model is that

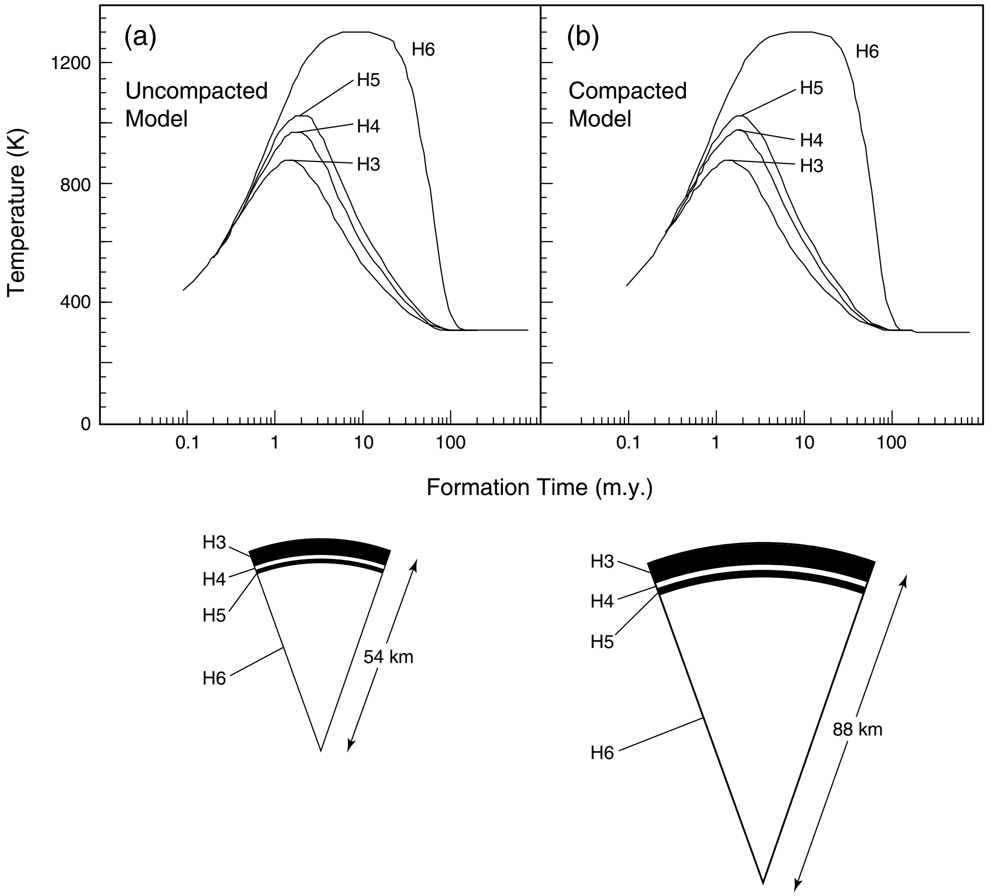

Fig. 1. Time-temperature curves plotted at various depths in the **(a)** uncompacted and **(b)** compacted Oc (H-chondrite) parent bodies of *Bennett and McSween* (1996). Sketches illustrating the corresponding volume proportions of petrologic types are plotted below for each case.

it approximately reproduces the cooling histories of H4, H5, and H6 chondrites (*Bennett and McSween,* 1996). The calculated radius of the H-chondrite parent body (88 km) is similar to the measured radius of asteroid 6 Hebe (~93 km), thought to be the probable source of H chondrites (*Gaffey and Gilbert,* 1998).

Particulate materials have much lower thermal conductivity than consolidated rock, and their effects on thermal models are appreciable. *Wood* (1979) and *Yomogida and Matsui* (1984) considered asteroids to be composed originally of powder that became sintered as temperatures rose during the calculations. *Bennett and McSween* (1996) used measured thermophysical data for high-porosity chondritic breccias to model uncompacted asteroids, and *Akridge et al.* (1998) and *Ghosh and McSween* (2001) modeled asteroids having particulate regoliths of varying thickness. The insulation afforded by even 120 m of regolith (the thickness threshold for insulation has not yet been established) results in a nearly isothermal asteroid interior with a large thermal gradient in the unconsolidated regolith. In effect, this increases the proportion of highly metamorphosed chondrite and moves the metamorphic boundaries (the onion shells) closer to the asteroid surface. Another consequence is that chondritic asteroids must be smaller, to preclude protracted thermal histories and melting. For example, the uncompacted H-chondrite parent body of *Bennett and McSween* (1996) has a radius of only 54 km, relative to the compacted model of 88 km (Fig. 1). Based on their thermal calculations, *Yomogida and Matsui* (1984) even suggested that each metamorphic grade of ordinary chondrite might have been derived from a different, small body. However, H chondrites of different metamorphic grade share the same (8 Ma) cosmic-ray exposure age, implying that they were parts of the same asteroid when launched by impact.

Metallographic cooling rates, determined from measured Ni diffusion profiles in taenite, in some Oc regolith breccias show extreme variations of as much as 1000 K/m.y. (*Williams et al.,* 1999). These cooling rates correspond to burial depths spanning the interval from the asteroid surface to ~100 km (the approximate asteroid radius) and are independent of metamorphic grade. It is inconceivable that an asteroid could survive an impact that sampled its center. The existence of breccias that sample such a depth interval implies that the parent body was disrupted and gravitationally reassembled, producing a rubble-pile structure (*Taylor et al.,* 1987). *Grimm* (1985) reasoned that asteroids shattered during accretion would reaccrete promptly (on the free-fall timescale) and therefore metamorphic grades would be set by initial position within the body but cooling rates would be determined by position following reassembly.

To facilitate calculation, thermal models for Oc parent bodies have generally assumed that asteroid accretion was instantaneous. This approximation can introduce errors, since it ignores the period during which ^{26}Al was most potent as a heat source. *Wood* (1979) and *Yomogida and Matsui* (1984) followed the progressive thermal evolution of small bodies of accreted dust that sintered into rock at a specific temperature. *Ghosh and McSween* (2000) devised a thermal model for the H-chondrite parent body that accreted incrementally, based on a constant growth rate. Peak temperatures in instantaneous accretion models must be reached, by definition, after accretion is complete. However, model runs with long duration of accretion (>2 m.y. from the time accretion starts) can reach peak temperature in the asteroid center while accretion is happening (Fig. 2a).

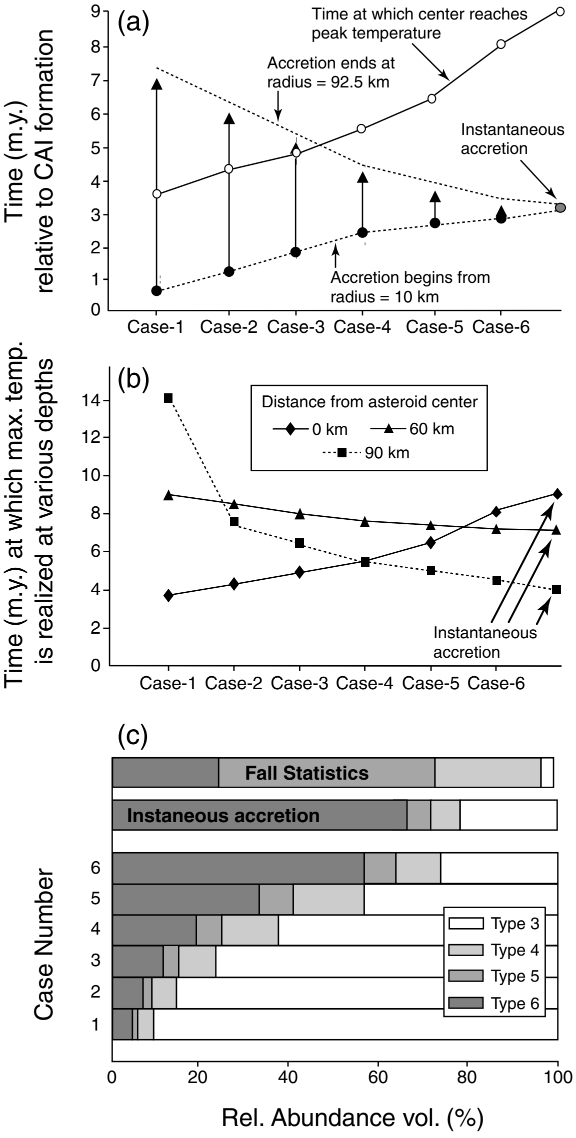

Fig. 2. **(a)** Timelines for the thermal evolution of asteroid 6 Hebe are shown for six cases with different accretion times and durations. Arrows represent periods of asteroid growth. The time (relative to CAI formation) at which peak temperature is attained at the asteroid center is indicated. Note that in cases 1, 2, and 3, peak temperature at the asteroid center is attained before accretion ends. **(b)** Time (relative to CAI formation) at which peak temperature is attained at various distances from the asteroid center. **(c)** Volume proportions of petrologic types obtained in cases 1–6 compared with results for an instantaneous accretion model. After *Ghosh and McSween* (2000).

Instantaneous accretion models also consistently underestimate the time at which the peak temperature is realized, because they fail to account for heating during accretion. The time at which the peak temperature is achieved decreases from the center to the surface of the asteroid in instantaneous accretion models. However, for incremental accretion with long duration, the opposite relationship is observed (Fig. 2b). Finally, the volumetric proportions of metamorphic grades may differ considerably. Instantaneous accretion models overestimate the amount of highly metamorphosed chondrite in the asteroid interior (Fig. 1), but neither incremental or instantaneous accretion models are able to match the observed chondrite fall statistics (Fig. 2c).

3. CARBONACEOUS CHONDRITE ASTEROIDS AND EFFECTS OF WATER ICE AND FLUID FLOW

Aqueous alteration is characteristic of many carbonaceous chondrites (Cc). Alteration produced secondary minerals that either contain water or hydroxyls (phyllosilicates) or formed by precipitation from hydrous fluids (carbonates and sulfates) (*Zolensky et al.,* 1989). Petrographic (*Brearley,* 1997) and kinetic (*Prinn and Fegley,* 1987) arguments support the assumption that melting of H_2O-rich ice incorporated into Cc parent bodies caused the aqueous alteration, although some hydrous alteration has been suggested to have occurred prior to asteroid accretion (*Metzler et al.,* 1992). The presence of free water profoundly influenced the thermal and chemical evolution of Cc parent bodies.

Oxygen-isotopic partitioning in CM and CI chondrites indicates that temperatures within many Cc parent bodies were within ~50° of the melting temperature of water ice during aqueous alteration (*Clayton and Mayeda,* 1984, 1999, *Leshin et al.,* 1997; *Young et al.,* 1999). *Grimm and McSween* (1989) first suggested that the large fusion heat of ice, the high heat capacity of water, and the ability of circulating water to enhance heat loss all may have contributed to thermal buffering of primordial heat sources in Cc parent objects. This fundamental difference in Cc and Oc initial composition led to low-temperature aqueous alteration instead of high-temperature metamorphic recrystallization.

Detailed modeling of hydration reactions — which liberate large amounts of heat — has been more difficult, as the cooling effect of endothermic melting of water ice is insufficient to negate the larger exothermic enthalpies of hydration. Where reaction rates are rapid compared to rates of thermal dissipation, temperatures of hundreds of degrees in excess of the constraints imposed by O-isotopic data would have resulted throughout large portions of Cc parent bodies (*Grimm and McSween,* 1989; *Cohen and Coker,* 2000). Low temperatures associated with aqueous alteration therefore imply either slow hydration-reaction rates or dissipation of heat by mechanisms more efficient than conduction. Reaction times must effectively exceed the conductive cooling time of the body for the former to hold.

Large temperature excursions could have been mitigated by hydrothermal convection. Hydration reactions as fast as 10^4 yr can be hydrothermally buffered with permeabilities comparable to those of fractured crystalline rocks and unconsolidated sands. Such permeabilities are comparable to the upper limit suggested previously by *Grimm and McSween* (1989), but are still far smaller than the maximum permeabilities of basaltic lavas. Hydrothermal convection is likely to have been important for parent bodies larger than several tens of kilometers in diameter. Flowing water in Cc parent bodies is supported by apparent water/rock volume ratios approaching or exceeding unity from oxygen-isotopic data (*Leshin et al*., 1997; *Clayton and Mayeda,* 1999). The convective model of *Grimm and McSween* (1989) produced uniformly low temperatures and pervasive alteration throughout the asteroid interior, or allowed alteration within a surficial regolith when water was introduced from below.

Young et al. (1999) reinterpreted the O-isotopic data in terms of progress of moving reaction fronts caused by flow of water. They reasoned that the trend of Cc O isotopes upward (toward higher $^{18}O/^{16}O$) along a mass fractionation line could best be explained by progressive partial reequilibration of aqueous fluid as it flowed down a thermal gradient. A monotonic thermal gradient is obtained in the presence of fluid flow by allowing "exhalation" of water under internal gas pressure. Isotopic exchange in both silicates and carbonates is tied to the kinetics of aqueous alteration in the exhalation model, which in turn depends on thermal history. The model successfully explains patterns of variation in Cc O-isotopic ratios and is consistent with the hypothesis that different Cc classes are samples of various horizons within asteroid precursors that had similar geological histories.

A Cc thermal model for a body thought to be too small for convection of water (radius = 9 km) is shown in Fig. 3. The model is based on the approach of *Young et al.* (1999) and uses a chondritic concentration of Al with an initial $^{26}Al/^{27}Al$ of 1×10^{-5} (corresponding to accretion at 1.6 m.y. after CAIs). The results are summarized using two-dimensional time vs. radius plots (the solutions are spherically symmetrical and thus one-dimensional in space). The Cc parent body was considered to be composed initially of forsterite olivine and water ice in these calculations. Forsterite was converted to secondary hydrous minerals (represented by talc) and carbonate minerals (represented by magnesite). Progress of the hydration and carbonation reactions was driven by the amount of CO_2 in the fluid rather than by temperature alone. It is envisaged that CO_2 would have come from oxidation of C within the parent body and/or from the ice itself.

Important features of the thermal evolution of small icy bodies like that in Fig. 3 are the short time span associated with geological evolution (<1 m.y.) and the presence of protracted temperature gradients that permit coexistence of metamorphosed rocks deep in the interior and aqueously altered rocks toward the surface (Fig. 3). The rocks exposed to intensive aqueous alteration are spatially removed from

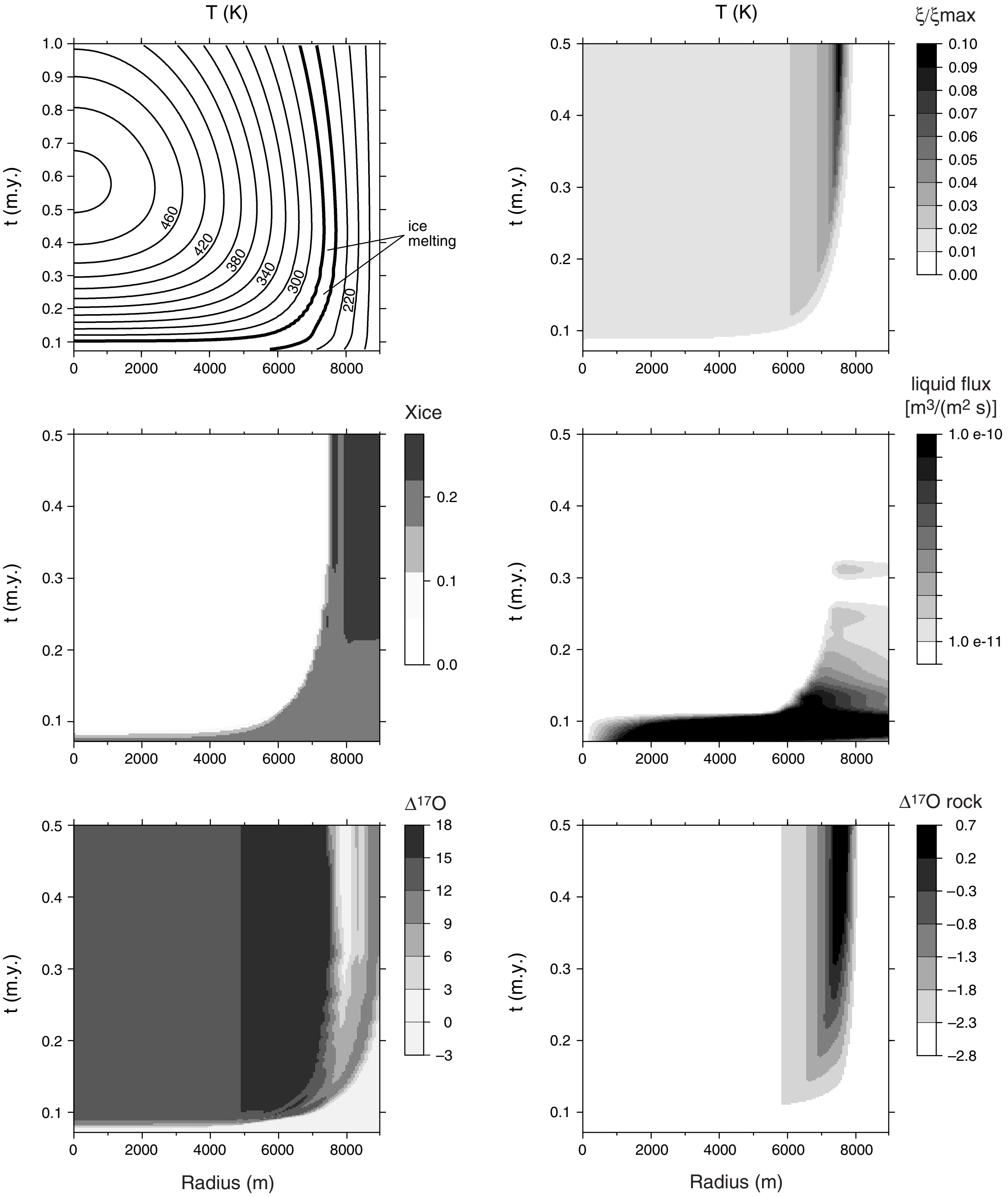

Fig. 3. Plots of time vs. radius for a small Cc parent body (radius = 9 km). Upper left shows the temperature history up to 1 m.y.; all other panels show history up to 0.5 m.y. Upper right panel shows progress of the model hydration and carbonation reaction relative to the maximum progress in mol. units. Middle left shows distribution of water ice in vol. fraction with time and radial position. Middle right shows flux of liquid water as a function of time and position (note the episodic nature of the flux at radial positions beyond ~7 km). Lower left shows changes in $\Delta^{17}O$ in liquid water with time and position. Lower right shows evolution of rock $\Delta^{17}O$ with time and position. Note that the zone of maximum mineralogical alteration coincides with the zone of maximum shift in $\Delta^{17}O$. Initial conditions for the model (*Young et al.,* 1999) included 0.2 vol. fraction water ice, 0.1 vol. fraction empty pore space, surface temperature of 180 K (a simple approximation to radiation to space), bulk ice mol. fraction CO_2 of 0.2, rock $\delta^{18}O$ and $\delta^{17}O$ values of –3.6 and –4.6 respectively, and water $\delta^{18}O$ and $\delta^{17}O$ values of 35.0 and 34.0 respectively (corresponding to a water $\Delta^{17}O$ of 15.8). Ice $\Delta^{17}O$ values substantially greater than those of rock are consistent with other studies (e.g., *Clayton and Mayeda,* 1999).

those subjected to thermal metamorphism (Fig. 3). The suggestion (*Brearley,* 1999) that matrixes of some largely anhydrous Ccs (CVs) could be dehydrated equivalents of intensively altered Ccs (CMs and CIs) may not be consistent with these models.

An analogous model for a large Cc asteroid, e.g., having the size of 1 Ceres, must invoke a smaller initial ^{26}Al/^{27}Al of 6.8×10^{-7} (accreting 4.4 m.y. after CAIs) in order to avoid driving peak temperatures well above the maximum recorded in Cc. Important features of large-body models are the long time span prior to aqueous alteration (>5 m.y. after accretion), lack of temperature gradients in the interior where aqueous alteration can occur, and absence of aqueous alteration where temperatures are sufficient for metamorphism. In addition, in the absence of convection, large bodies with heat production sufficient for metamorphism displace and expel water too rapidly for fluid-rock reaction to occur using realistic reaction rates.

High vapor pressures associated with ice melting may also have profoundly affected the geological evolution of Cc asteroids. Consideration of vapor permeabilities appropriate for chondrites and vapor pressures within icy planetesimals suggests that Cc parent bodies may have fractured and vented gases (*Grimm and McSween,* 1989) and could have exploded due to vapor overpressures once water ice began to melt (*Wilson et al.,* 1999; *Cohen and Coker,* 2000). Observations that Cc clasts are common in other meteorite groups (*Zolensky et al.,* 1996) and that highly altered Ccs are brecciated (*Wilson et al.,* 1999) may suggest that explosive disaggregation was an integral part of the evolution of Cc parent bodies.

Although different in fundamental ways, the Cc thermal models of *Young et al.* (1999) and *Cohen and Coker* (2000), as well as the regolith alteration model of *Grimm and McSween* (1989), suggest that low-temperature aqueous alteration was restricted to relatively narrow horizons within the asteroids. The depth of the alteration zone and the timescale for alteration depend upon the size of the body and the rate of heat production. Icy bodies with radii ≤50 km would have experienced aqueous alteration and metamorphism within ~1 m.y. of accretion. Aqueous alteration on much larger bodies would have been delayed by ~5 m.y. or more relative to the time of accretion. The model of *Young et al.* (1999) suggests that a single small body could have produced both metamorphosed and aqueously altered Cc rocks. The same may not be true of larger bodies. In the absence of convection of water, rapid heating of larger bodies to metamorphic temperatures drives water outward with such speed that no aqueous alteration can occur. Recent suggestions that aqueous alteration in Ccs occurred over intervals on the order of 8 m.y. (e.g., *Hutcheon et al.,* 1999) may be consistent with diachronous aqueous alteration within large parent bodies with radii of hundreds of kilometers.

While there has been considerable progress in thermal modeling of Cc parent bodies, there is still no self-consistent model that incorporates reaction heat, isotopic exchange, and fluid flow. Hydrothermally convective interiors are consistent with gross isotopic water/rock ratios, relatively uniform compositions of Cc, and heat loss (*Grimm and McSween,* 1989), but recirculating water may not satisfy isotopic constraints. The "exhalation" model precisely matches the isotopic constraints (*Young et al.,* 1999), but as presently formulated may not produce sufficient alteration, nor is it likely to be able to extract heat without very slow reaction kinetics. Better knowledge of the rates of hydration and carbonation reactions at low temperatures would be useful for judging the relative importance of convection (recirculation) vs. exhalation (single-pass flow) in the evolution of Cc parent bodies.

4. DIFFERENTIATED ASTEROID 4 VESTA AND THE EFFECT OF REDISTRIBUTING HEAT SOURCES

The eucrites and closely related diogenites and howardites (collectively called HED achondrites) are basalts, pyroxenites, and regolith breccias thought to have been extracted from asteroid 4 Vesta (*Consolmagmo and Drake,* 1977; *Binzel and Xu,* 1993; *Farinella et al.,* 1993; *Drake,* 2001). Unlike models of chondrite parent bodies, thermal calculations for achondrite parent bodies require incorporation of complexities introduced by melting and differentiation. *Ghosh and McSween* (1998) modeled the thermal history of Vesta from instantaneous accretion to cooling, using decay of short-lived radionuclides (primarily ^{26}Al, although ^{60}Fe was included) as heat sources. Achondrites and iron meteorites demonstrate that many other differentiated asteroids existed, and a thermal model for a differentiated body based on long-lived radionuclide decay has also been formulated (*Haack et al.,* 1990).

Although Vesta's radius is known (*Thomas et al.,* 1997), the mass of Vesta as determined by its gravitational effect on a nearby asteroid has considerable uncertainity (*Standish and Hellings,* 1989), which introduces a corresponding uncertainty in bulk density. This, in turn, makes it impossible to reliably estimate the size of the core or the asteroid's metal content. *Ghosh and McSween* (1998) preferred H chondrite as the starting composition, which has a metal content similar to Vesta estimates by *Dreibus et al.* (1997). Initial compositions of L and LL chondrites produce slightly higher temperatures for the same parameter set, due to increases in the relative amounts of ^{26}Al. Bulk compositions of H, L, and LL chondrites yield core radii of 123, 108, and 90 km respectively.

Jones (1984) estimated the mantle composition of the HED asteroid based on olivine-melt partition coefficients for Sc, Mg, and Si. He concluded that the undifferentiated mantle could be approximated by a mixture of 25% eucrite and 75% olivine. In the absence of a better model, *Ghosh and McSween* (1998) assumed the crust composition to be eucrite and the depleted mantle composition to be pure olivine. The degree of partial melting of Vesta's mantle was assumed to be 25% based on experimental studies of eucrites (*Stolper,* 1977*; Grove and Bartels,* 1992; *Jurewicz et al.,* 1995). A competing model based primarily on trace-element abundances suggests a much larger degree of melting, producing a magma ocean (*Righter and Drake,* 1997).

The mechanisms that lead to sulfide or silicate melt segregation in asteroids, and thus the formation of cores and crusts, are poorly understood. There exist two schools of thought about the degree of melting required for separation of metal-sulfide liquids from a silicate matrix: one requiring extensive melting (*Stevenson,* 1990; *Taylor,* 1992), and the other limited melting (*Larimer,* 1995). Neither approach takes into account the rate of melt generation. In addition to physical properties of the melt and enclosing rock, the rate of melt migration depends on how fast melting takes place, which in turn depends upon the rate of heat generation by ^{26}Al. When the eutectic temperature of the Fe-FeS system is reached at a particular depth, a melt of eutectic composition is generated. Separation of the metal-sulfide liquid promotes further melting, because the residue has a higher Al content than the melt plus residue. Thus, migration of metal-sulfide liquid results in a positive feedback mechanism: The greater the amount of metal-sulfide melt drained away, the greater will be the melting of the residue, and hence the amount of melt generated will increase. *Ghosh and McSween* (1998) reasoned that if melt migration were somehow triggered, thermal considerations point to rapid core separation.

The timeframe of crust formation on Vesta is difficult to constrain. In regions of the upper mantle where upward movement and decompression of rocks during solid-state convection allow partial melting, melt segregation occurs initially by percolation along grain boundaries. Deformation of the matrix allows melt to be concentrated (*Richter and McKenzie,* 1984; *Barcilon and Lovera,* 1989). However, the region in which melt is concentrated itself rises buoyantly by deforming the surrounding rocks (*Marsh,* 1989). In both cases the timescale is controlled by the viscosity of the matrix, the size of the concentration zone, and the gravitational acceleration (and is therefore slower in an asteroid than on Earth). However, at some stage in the upward segregation process, the rheological response of the surrounding rocks changes from plastic to elastic and a liquid-filled fracture, i.e., a dike, forms (*Sleep,* 1988). The propagation speed of the dike is controlled by the viscosity of the fluid rather than the viscosity of the enclosing rocks, and the melt rise speed is therefore likely to increase by many orders of magnitude. As soon as dikes dominate the process, transfer of melt to shallow depths or to the surface is essentially instantaneous (*Wilson and Keil,* 1996). For convenience in coding, *Ghosh and McSween* (1998) assumed temperature "windows" for both metal-sulfide and silicate melting, and assumed instantaneous formation of core and crust.

Ghosh and McSween (1998) divided the evolution of Vesta into three stages (Fig. 4): (1) radiogenic heating of a homogeneous asteroid until core separation, (2) subsequent heating of the mantle until crust formation, and (3) subsequent heating and cooling of the differentiated asteroid. Two end members, which assumed that either all or no melt erupted, were evaluated since it is not known what proportion of the silicate magma generated at depth eventually erupts. The model places instantaneous accretion of Vesta at 2.9 m.y. after CAI formation. Core formation occurs at 4.6 m.y., and crust formation at 6.6 m.y. The model ages compare favorably with constraints on the timing of core and crust formation from ^{182}Hf-^{182}W (*Lee and Halliday,* 1997), ^{26}Al-^{26}Mg (*Srinivasan et al.,* 1999), and ^{53}Mn-^{53}Cr (*Lugmair and Shukolyukov,* 2001) isotope systematics in HED meteorites. This model illustrates the thermal effect of redistributing ^{26}Al during differentiation. After core formation, the core contains no ^{26}Al and its abundance of ^{60}Fe is too low (*Shukolyukov and Lugmair,* 1996) to contribute significant heat. Thus, the heat engine in the core is shut off, whereas the temperature in the overlying mantle increases (Fig. 4b). This gives rise to a reverse thermal gradient where temperature decreases with increasing depth. In terms of cooling history, this means that not only is heat loss from the core inhibited, but some heat in fact flows into the core by thermal diffusion from the overlying mantle. This reverse gradient persists for ~100 m.y., and is responsible for minimizing heat loss from Vesta's interior during this time interval. Interestingly, this phenomenon is not observed in planets, where core formation takes place long after ^{26}Al decay, but should be observed in small planetesimals that underwent metal-sulfide melting and segregation at a time when ^{26}Al was still potent. A similar reversed thermal gradient is observed in one model end member (Fig. 4c) after ^{26}Al is sequestered in the crust, causing the crust to attain higher temperatures than the underlying mantle.

This study may provide answers to several longstanding problems with the hypothesis of heating by ^{26}Al. The rarity of excess ^{26}Mg, the decay product of ^{26}Al, in eucrites can be explained because the timing of volcanism is such that the ^{26}Al concentration would commonly fall below detectable limits. Excess ^{26}Mg has since been detected in several eucrites (*Srinivasan et al.,* 1999; *Nyquist et al.*, 2001). Chronologic data suggest a time interval of ~100 m.y. between the formation of noncumulate and cumulate eucrites (*Tera et al.,* 1997). Since ^{26}Al is not potent beyond a few million years after the solar system formed, the long time interval was thought to be problematic (*Wood and Pellas,* 1991). A combination of factors — the reverse thermal gradients in the core and crust after metal segregation and crust formation, respectively, and the low thermal diffusivity — produced a prolonged cooling history for Vesta. Figures 4c,d show that temperatures in the mantle stay hot enough after 100 m.y. to prevent geochemical closure in cumulate eucrites.

Ghosh and McSween (1998) suggested the possibility that chondritic percursor rocks, present in the outer layer of Vesta before development of a crust, may still exist. The radiation boundary condition ensures that the temperature in near-surface layers remains low. The thickness of the unaltered carapace decreases with increasing degrees of

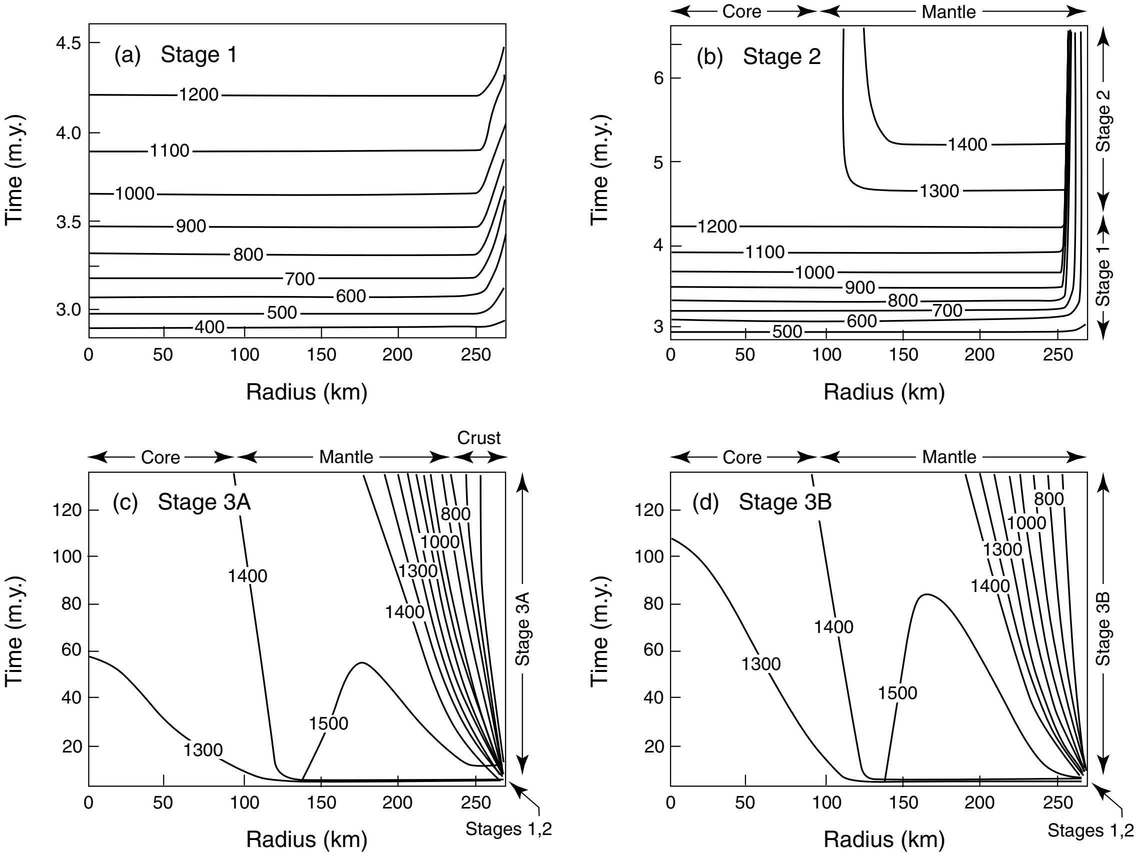

Fig. 4. Temperature contours for 4 Vesta, on plots of time elapsed since CAI formation and radial distance from the asteroid center, after *Ghosh and McSween* (1998). **(a)** Stage 1 is the interval from accretion to core separation. **(b)** In stage 2, core formation has redistributed ^{26}Al, causing heat generation in the core to stop. Mantle temperatures continue to rise, causing silicate melting for production of the crust. Comparison of stages 3A and 3B illustrates the difference in heat transfer between a configuration **(c)** where the entire melt generated is extruded onto the surface and **(d)** where the melt entirely solidifies as plutons.

melting and, for 25% partial melting, the outer 10 km of the asteroid never achieves melting temperatures, although parts of the layer are metamorphosed. However, eruptions of silicate melt, or intrusions of dikes or sills at shallow depth, must cause local metamorphism (*Yamaguchi et al.*, 1997; *Wilson and Keil,* 2000). Further work is needed to establish whether all the unmelted carapace will be destroyed by igneous crust formation or by increased melting in the mantle as in the magma ocean scenario (*Righter and Drake,* 1997). As in chondrite parent bodies, small impacts are not capable of widespread melting (*Melosh,* 1990). Large impacts can cause some melting, but the effect is restricted to the hemisphere that is impacted, leaving the other hemisphere unaltered or at most slightly metamorphosed (*Williams and Wetherill,* 1993).

5. THERMAL STRUCTURE OF THE ASTEROID BELT

The heliocentric distribution of asteroid spectral types (*Gradie and Tedesco,* 1982) has been interpreted to indicate high peak temperatures appropriate for melting or metamorphism for bodies closer to the Sun, with mildly heated or unaltered bodies at greater distances (*Bell et al.,* 1989). This pattern persists, despite some subsequent dynamical stirring of asteroid orbits and ejection of bodies from the main belt. *Grimm and McSween* (1993) devised a quantitative model to explain this radial thermal structure. Because accretion time increases with heliocentric distance (*Wetherill,* 1980), objects that accreted at greater distances had smaller proportions of live ^{26}Al available to drive heating.

The results, expressed as contours of peak temperature on a plot of asteroid size vs. semimajor axis (the latter is equivalent to accretion time relative to CAI formation), are shown in Fig. 5. In this diagram, bodies inward of 2.7 AU are anhydrous (90% rock, 10% voids), whereas those farther from the Sun contain ice (60% rock, 30% ice,10% voids). The vertical bar at 2.7 AU marks the approximate distance for the transition from melted or metamorphosed asteroids to those that experienced aqueous alteration, and the bar at 3.4 AU denotes the transition to unaltered asteroids in which ice was never melted. The accretion times at the top of Fig. 5 produce appropriate peak temperature

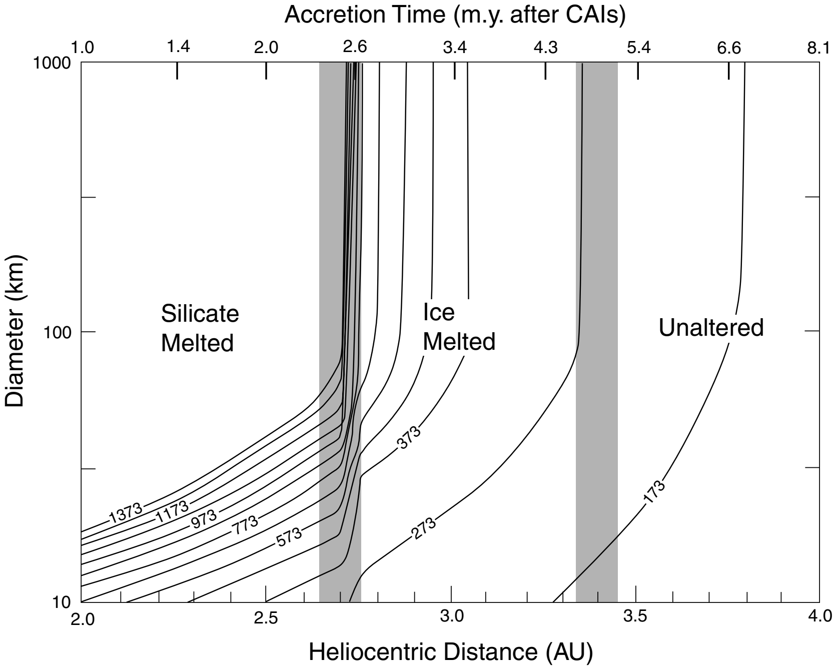

Fig. 5. Contours of peak temperature in asteroids as functions of size and semimajor axis (or accretion time, relative to CAI formation). Shaded bands mark major divisions in the asteroid belt based on interpretation of spectra. Modified from *Grimm and McSween* (1993).

contours (1375 K for silicate melting, 273 K for ice melting) for ~100-km-diameter bodies at these heliocentric distances. The accretion times in Fig. 5 have been slightly increased from that published by *Grimm and McSween* (1993), to correct a coding error in the fusion heat of water.

Ghosh et al. (2001) formulated a more complex model that incorporates incremental rather than instantaneous accretion. The multizone accretion model (*Weidenschilling et al.,* 1997) allows accretion to begin simultaneously (as 0.5-km planetesimals) throughout the belt, but growth rates still vary with swarm density and semimajor axis. Although accretion in the inner asteroid belt is faster than in the outer belt, the difference in accretion rates by itself is not sufficient to produce thermal stratification in a model of ^{26}Al heating. The buffering effect of ice in the outer belt lowers peak temperatures for bodies in this region. Other factors that may contribute to the thermal stratification are differences in accretion temperature between the inner and outer belt, and the accretion of planetesimals that are unsintered and hence capable of achieving higher temperatures for smaller asteroid sizes. Bodies that are too small to sustain metamorphic temperatures comprise most of the mass of the multizone accretion code. Thus, unmetamorphosed small bodies dominate the inner belt. The thermal distribution can be made to conform approximately with the observed distribution of asteroids if these small bodies are destroyed by mutual collisions (*Davis et al.,* 1989).

These calculations demonstrate that heliocentric thermal zoning of the asteroid belt can be achieved by ^{26}Al heating with realistic accretion scenarios. This Sun-centered pattern might also be consistent with solar electromagnetic induction heating, but that mechanism is not sufficiently constrained to allow a similar computation. Neither heating mechanism, by itself, provides an obvious explanation for why Vesta is differentiated while Ceres, at double its size, is not. The thermal histories of individual asteroids must reflect complex interactions between their sizes, accretional timescales, physical states, and chemical compositions.

6. CONCLUSIONS AND FUTURE WORK

Thermal evolution models using ^{26}Al as a heat source have been used to address a spectrum of problems, including metamorphism of Oc parent bodies, aqueous alteration of Cc parent bodies, and melting of differentiated asteroids. Models based on ^{26}Al heating and either instantaneous accretion varying with heliocentric distance or stochastic, incremental accretion appear to be broadly consistent with the thermal stratification of the asteroid belt inferred from the taxonomic distribution of asteroids.

However, ^{26}Al heating requires a longer time interval (~2 m.y.) for accretion to match asteroid peak temperatures than is allowed by most nebular accretion models. This may imply that metamorphism and melting occurred in smaller bodies than currently envisioned, or that ^{26}Al was heterogeneously distributed so that its overall abundance was less than the canonical value. Although electrical induction heating of asteroids is plausible, the hypothesis is difficult to test quantitatively because it hinges on the choice of parameters that are largely unconstrained. Collisional heating appears to be insufficient to account for global thermal metamorphism or significant partial melting in bodies of asteroidal size.

The most straightfoward asteroid thermal models are for metamorphosed Oc parent bodies. An added complexity in these models is the presence of a regolith during heating, which effectively insulates the asteroid interior and profoundly affects its thermal evolution. Carbonaceous chondrite parent bodies originally contained ice, the melting of which acts as a thermal buffer to limit temperature excursions. Fluid flow was also apparently important in controlling heat loss, but still must be fully reconciled with stable-isotopic data. Thermal models for asteroids that experience partial melting and differentiation are more complex, because heat sources migrate within the body during the simulation. These models also have many unconstrained parameters, including the extent of melting and the depth range of melt emplacement.

Many parameters in existing thermal models need revision or refinement. Better theoretical estimates of the initial temperatures of originally accreted materials, as well as a way to anchor timescales in nebular models to CAI formation, are required. Improved constraints from meteorites are also needed. For example, additional measurements of specific heat capacity and diffusivity, as well as accurate peak temperatures from geothermometry and more precise ages from high-resolution chronometers, would improve chondrite thermal models. Overprinted shock effects must be disentangled from meteorite cooling rates. Spacecraft missions to asteroids will hopefully provide data on regolith thicknesses, thermal properties, and ages. Thermal models for differentiated bodies require better constraints on the relative timing of core separation and mantle melting, and existing models do not yet adequately account for heat loss by convection. Also, it is critical to tie whole-asteroid thermal models to magma migration models.

The essence of thermal evolution models is knowing what can be simplified without sacrificing accuracy. The most common simplifying assumption in existing models is that accretion happened instantaneously. However, preliminary attempts to account for the heat budget during asteroid growth show that the rate of accretion can profoundly affect thermal evolution. Incorporation of realistic, incremental accretion scenarios for both chondritic and achondritic asteroids would be a major step forward in thermal modeling.

REFERENCES

Akridge G., Benoit P. H., and Sears D. W. G. (1998) Regolith and megaregolith formation of H-chondrites: Thermal constraints on the parent body. *Icarus, 132,* 185–195.

Barcilon V. and Lovera O. (1989) Solitary waves in magma dynamics. *J. Fluid Mech., 204,* 121–133.

Bell J. F., Davis D. R., Hartmann W. K., and Gaffey M. J. (1989) Asteroids: The big picture. In *Asteroids II* (R. P. Binzel, T. Gehrels, and M. S. Matthews, eds.), pp. 921–945. Univ. of Arizona, Tucson.

Bennett M. E. and McSween H. Y. Jr. (1996) Revised model calculations for the thermal histories of ordinary chondrite parent bodies. *Meteoritics & Planet. Sci., 31,* 783–792.

Binzel R. P. and Xu S. (1993) Chips off of asteroid 4 Vesta: Evidence for the parent body of basaltic achondrite meteorites. *Science, 260,* 186–191.

Brearley A. J. (1997) Disordered biopyriboles, amphibole, and talc in the Allende meteorite: Products of nebular or parent body aqueous alteration? *Science, 276,* 1103–1105.

Brearley A. J. (1999) Origin of graphitic carbon and pentlandite in matrix olivines in the Allende meteorite. *Science, 285,* 1380–1381.

Clayton R. N. and Mayeda T. K. (1984) The oxygen isotope record in Murchison and other carbonaceous chondrites. *Earth Planet. Sci. Lett., 67,* 151–161.

Clayton R. N. and Mayeda T. K. (1999) Oxygen isotope studies of carbonaceous chondrites. *Geochim. Cosmochim. Acta, 63,* 2089–2104.

Cohen B. A. and Coker R. F. (2000) Modeling of liquid water on CM meteorite parent bodies and implications for amino acid racemization. *Icarus, 145,* 369–381.

Consolmagmo G. and Drake M. J. (1977) Compositional evolution of the eucrite parent body: Evidence from rare earth elements. *Geochim. Cosmochim. Acta, 41,* 1271–1282.

Davis D. R., Weidenschilling S. J., Stuart J., Farinella P., Paolicchi P., and Binzel R. P. (1989) Asteroid collisional history — Effects on sizes and spins. In *Asteroids II* (R. P. Binzel, T. Gehrels, and M. S. Matthews, eds.), pp. 805–826. Univ. of Arizona, Tucson.

DeCampli W. M. (1981) T Tauri winds. *Astrophys. J., 244,* 124–146.

Drake M. J. (2001) The eucrite/Vesta story. *Meteoritics & Planet. Sci., 36,* 501–513.

Dreibus G., Brückner J., and Wänke H. (1997) On the core mass of asteroid 4 Vesta. *Meteoritics & Planet. Sci., 32,* A36.

Edwards S., Cabrit D., Strom S. E., Heyer I., Strom K. M., and Anderson E. (1987) Forbidden line and Hα profiles in T Tauri star spectra: A probe of anisotropic mass outflows and circumstellar disks. *Astrophys. J., 321,* 473–495.

Farinella P., Gonczi R., Froeschlé Ch., and Froeschlé C. (1993) The injection of asteroid fragments into resonances. *Icarus, 101,* 174–187.

Fujii N., Miyamoto M., and Ito K. (1979) The role of external heating and thermal metamorphism of chondritic parent body. *Planet. Sci., 1,* 84.

Gaffey M. J. and Gilbert S. L. (1998) Asteroid 6 Hebe: The probable parent body of the H-type ordinary chondrites and the IIE iron meteorites. *Meteoritics & Planet. Sci., 33,* 1281–1295.

Ghosh A. and McSween H. Y. Jr. (1998) A thermal model for the differentiation of asteroid 4 Vesta, based on radiogenic heating. *Icarus, 134,* 187–206.

Ghosh A. and McSween H. Y. Jr. (2000) The effect of incremental accretion on the thermal modeling of asteroid 6 Hebe (abstract). *Meteoritics & Planet. Sci., 35,* A59.

Ghosh A., Weidenschilling S. J., and McSween H. Y. Jr. (2001) Thermal consequences of the multizone accretion code on the structure of the asteroid belt (abstract). In *Lunar and Planetary Science XXXII,* abstract #1760. Lunar and Planetary Institute, Houston (CD-ROM).

Göpel C., Manhes G., and Allegré C. J. (1994) U-Pb systematics of phosphates from equilibrated ordinary chondrites. *Earth Planet. Sci. Lett., 121,* 153–171.

Gradie J. C. and Tedesco E. F. (1982) Compositional structure of the asteroid belt. *Science, 216,* 1405–1407.

Grimm R. E. (1985) Penecontemporaneous metamorphism, fragmentation, and reassembly of ordinary chondrite parent bodies. *J. Geophys. Res., 90,* 2022–2028.

Grimm R. E. and McSween H. Y. Jr. (1989) Water and the thermal evolution of carbonaceous chondrite parent bodies. *Icarus, 82,* 244–280.

Grimm R. E. and McSween H. Y. Jr. (1993) Heliocentric zoning of the asteroid belt by alumimum-26 heating. *Science, 259,* 653–655.

Grove T. L. and Bartels K. S. (1992) The relation between diogenite cumulates and eucrite magmas. *Proc. Lunar Planet. Sci., Vol. 22,* pp. 437–445.

Haack H., Rasmussen K. L., and Warren P. H. (1990) Effects of regolith/megaregolith insulation on the cooling histories of differentiated asteroids. *J. Geophys. Res., 95,* 5111–5124.

Herbert F. (1989) Primoridal electrical induction heating of asteroids. *Icarus, 78,* 402–410.

Herndon J. M. and Herndon M. A. (1977) Aluminum-26 as a planetoid heat source in the early solar system. *Meteoritics, 12,* 459–465.

Huss G. R., MacPherson G. J., Wasserburg G. J., Russell S. S., and Srinivasan G. (2001) Aluminum-26 in calcium-aluminum-rich inclusions and chondrules from unequilibrated ordinary chondrites. *Meteoritics & Planet. Sci., 36,* 975–997.

Hutcheon I. D., Weisberg M. K., Phinney D. L., Zolensky M. E., Prinz M., and Ivanov A. V. (1999) Radiogenic ^{53}Cr in Kaidun carbonates: Evidence for very early aqueous activity (abstract). In *Lunar and Planetary Science XXX,* abstract #1722. Lunar and Planetary Institute, Houston (CD-ROM).

Ireland T. R. and Fegley B. Jr. (2000) The solar system's earliest chemistry: Systematics of refractory inclusions. *Intl. Geol. Rev., 42,* 865–894.

Jones J. H. (1984) The composition of the mantle of the eucrite parent body and the origin of eucrites. *Geochim. Cosmochim. Acta, 48,* 641–648.

Jurewicz A. J. G., Mittlefehldt D. W., and Jones J. H. (1995) Experimental partial melting of the St. Severin (LL) and Lost City (H) chondrites. *Geochim. Cosmochim. Acta, 59,* 391–408.

Keil K., Stöffler D., Love S. G., and Scott E. R. D. (1997) Constraints on the role of impact heating and melting in asteroids. *Meteoritics & Planet. Sci., 32,* 349–363.

Kita N. T., Nagahara H., Togashi S., and Morishita Y. (2000) A short duration of chondrule formation in the solar nebula: Evidence from ^{26}Al in Semarkona ferromagnesian chondrules. *Geochim. Cosmochim. Acta, 64,* 3913–3922.

Larimer J. W. (1995) Core formation in asteroid-sized bodies. *Meteoritics, 30,* 552.

LaTourette T. and Wasserburg G. J. (1997) Mg diffusion in anorthite: Implications for the formation of early solar system planetesimals. *Earth Planet. Sci. Lett., 158,* 91–108.

Lee D.-C. and Halliday A. N. (1997) Core formation on Mars and differentiated asteroids. *Nature, 388,* 854–857.

Lee T., Papanastassiou D. A., and Wasserburg G. J. (1976) Demonstration of ^{26}Mg excess in Allende and evidence for ^{26}Al. *Geophys. Res. Lett., 3,* 41–44.

Leshin L. A., Rubin A. E., and McKeegan K. D. (1997) The oxygen isotopic composition of olivine and pyroxene from CI chondrites. *Geochim. Cosmochim. Acta, 61,* 835–845.

Lugmair G. W. and Shukolyukov A. (2001) Early solar system events and timescales. *Meteoritics & Planet. Sci., 36,* 1017–1026.

MacPherson G. J., Davis A. M., and Zinner E. K. (1995) The distribution of aluminum-26 in the early solar system — a reappraisal. *Meteoritics & Planet. Sci., 30,* 365–386.

Marsh B. D. (1989) Magma chambers. *Annu. Rev. Earth Planet. Sci.,17,* 439–474.

McSween H. Y. Jr., Sears D. W. G., and Dodd R. T. (1988) Thermal metamorphism. In *Meteorites and the Early Solar System* (J. F. Kerridge and M. S. Matthews, eds.), pp. 102–113. Univ. of Arizona, Tucson.

Melosh H. J. (1990) Giant impacts and the thermal state of the early Earth. In *Origin of the Earth* (H. E. Newsom and J. H. Jones, eds.), pp. 69–84. Oxford Univ., New York.

Metzler K., Bishoff A., and Stöffler D. (1992) Accretionary dust mantles in CM chondrites: Evidence for solar nebula processes. *Geochim. Cosmochim. Acta, 56,* 2873–2897.

Minster J. F. and Allegré C. J. (1979) ^{87}Rb-^{87}Sr chronology of H chondrites: Constraint and speculations on the early evolution of their parent body. *Earth Planet. Sci. Lett., 42,* 333–347.

Mittlefehldt D. W. (1979) The nature of asteroid differentiation processes: Implications for primordial heat sources. *Proc. Lunar Planet. Sci. Conf. 10th,* pp. 1975–1993.

Miyamoto M. (1991) Thermal metamorphism of CI and CM carbonaceous chondrites: A internal heating model. *Meteoritics, 26,* 111–115.

Miyamoto M., Fujii N., and Takeda H. (1981) Ordinary chondrite parent body: An internal heating model. *Proc. Lunar Planet. Sci. 12B,* pp. 1145–1152.

Nakamuta Y. and Motomura Y. (1999) Sodic plagioclase thermometry of type 6 ordinary chondrites: Implications for the thermal histories of parent bodies. *Meteoritics & Planet. Sci., 34,* 763–772.

Nyquist L. E., Reese Y., Wiesmann H., Shih C.-Y., and Takeda H. (2001) Live ^{53}Mn and ^{26}Al in an unique cumulate eucrite with very calcic feldspar (An-98). *Meteoritics & Planet. Sci., 36,* A151–A152.

Olsen E. J. and Bunch T. E. (1984) Equilibration temperatures of the ordinary chondrites: A new evaluation. *Geochim. Cosmochim. Acta, 48,* 1363–1365.

Pellas P. and Storzer D. (1981) ^{244}Pu fission track thermometry and its application to stony meteorites. *Proc. R. Soc. Lond., A374,* 253–270.

Prinn R. G. and Fegley B. J. (1987) The atmospheres of Venus, Earth and Mars: A critical review. *Annu. Rev. Earth Planet. Sci., 15,* 171–212.

Richter F. M. and McKenzie D. (1984) Dynamical models for melt segregation from a deformable matrix. *J. Geol., 92,* 729–740.

Righter K. and Drake M. J. (1997) A magma ocean on Vesta: Core formation and petrogenesis of eucrites and diogenites. *Meteoritics & Planet. Sci., 32,* 929–944.

Rubin A. E. (1995) Petrologic evidence for collisional heating of chondritic asteroids. *Icarus, 113,* 156–167.

Russell S. S., Srinivasan G., Huss G. R., Wasserburg G. J., and MacPherson G. J. (1996) Evidence for widespread ^{26}Al in the solar nebula and constraints for nebula time scales. *Science, 273,* 757–762.

Sears D. W., Grossman J. N., Melcher C. L., Ross L. M., and Mills A. A. (1980) Measuring metamorphic history of unequilibrated ordinary chondrites. *Nature, 287,* 791–795.

Shimazu H. and Terasawa T. (1995) Electromagnetic induction heating of meteorite parent bodies by the primordial solar wind. *J. Geophys. Res., 100,* 16923–16930.

Shukolyukov A. and Lugmair G. W. (1996) ^{60}Fe-^{60}Ni systematics in the eucrite Caldera. *Meteoritics, 31,* A129.

Sleep N. H. (1988) Tapping of melt by veins and dikes. *J. Geophys. Res., 93,* 10255–10272.

Sonnett C. P., Colburn D. S., and Schwartz K. (1968) Electrical heating of meteorite parent bodies and planets by dynamo induction from a premain sequence T Tauri "solar wind." *Nature, 219,* 924–926.

Srinivasan G., Goswami J. N., and Bhandari N. (1999) ^{26}Al in eucrite Piplia Kalan: Plausible heat source and formation chronology. *Science, 284,* 1348–1350.

Standish E. M. and Hellings R. W. (1989) A determination of the masses of Ceres, Pallas and Vesta from their perturbation upon the orbit of Mars. *Icarus, 80,* 326–333.

Stevenson D. J. (1990) Models of the Earth's core. *Science, 214,* 611–619.

Stolper E. M. (1977) Experimental petrology of eucritic meteorites. *Geochim. Cosmochim. Acta, 41,* 587–611.

Taylor G. J. (1992) Core formation in asteroids. *J. Geophys. Res., 97,* 14717–14726.

Taylor G. J., Maggiore P., Scott E. R. D., Rubin A. E., and Keil K. (1987) Original structures, and fragmentation and reassembly histories of asteroids: Evidence from meteorites. *Icarus, 69,* 1–13.

Tera F., Carlsson R. W., and Boctor N. Z. (1997) Radiometric ages of basaltic achondrites and their relation to the early history of the Solar System. *Geochim. Cosmochim. Acta, 61,* 1713–1731.

Thomas P. C., Binzel R. P., Gaffey M. J., Storrs A. D., Wells E. N., and Zellner B. H. (1997) Impact excavation of 4 Vesta: Hubble Space Telescope results. *Science, 277,* 1492–1495.

Urey H. (1955) The cosmic abundances of potassium, uranium, and thorium and the heat balances of the Earth, the Moon and Mars. *Proc. Natl. Acad. Sci. U.S., 41,* 127–144.

Wasson J. T., Rubin A. E., and Benz W. (1987) Heating of primitive, asteroid-size bodies by large impacts. *Meteoritics, 22,* 525–526.

Weidenschilling S. J., Spaute D., Davis D. R., Mazari F., and Ohtsuki K. (1997) Accretional evolution of a planetesimal swarm. *Icarus, 128,* 429–455.

Wetherill G. W. (1980) Formation of terrestrial planets. *Annu. Rev. Astron. Astrophys., 18,* 77–213.

Williams C. V., Keil K., Taylor G. J., and Scott E. R. D. (1999) Cooling rates of equilibrated clasts in ordinary chondrite regolith breccias: Implications for parent body histories. *Chem. Erde, 59,* 287–305.

Williams D. R. and Wetherill G. W. (1993) Equilibrium models of mass distribution and collisional lifetime of asteroids (abstract). In *Lunar and Planetary Science XXIV,* pp. 1523–1524. Lunar and Planetary Institute, Houston.

Wilson L. and Keil K. (1996) Volcanic eruptions and intrusions on the asteroid 4 Vesta. *J. Geophys. Res., 101,* 18927–18940.

Wilson L. and Keil K. (2000) Crust development on differentiated asteroids (abstract). In *Lunar and Planetary Science XXXI,* abstract #1576. Lunar and Planetary Institute, Houston (CD-ROM).

Wilson L., Keil K., Browning L. B., Krot A. N., and Bourcher W. (1999) Early aqueous alteration, explosive disruption, and reprocessing of asteroids. *Meteoritics & Planet. Sci., 34,* 541–557.

Wood J. A. (1979) Review of the metallographic cooling rates of meteorites and a new model for the planetesimals in which they formed. In *Asteroids* (T. Gehrels, ed.), pp. 849–891. Univ. of Arizona, Tucson.

Wood J. A. and Morfill G. (1988) A review of solar nebular models. In *Meteorites and the Early Solar System* (J. F. Kerridge and M. S. Matthews, eds.), pp. 329–347. Univ. of Arizona, Tucson.

Wood J. A. and Pellas P. (1991) What heated the parent meteorite planets? In *The Sun in Time* (C. P. Sonnett and M. S. Giampapa, eds.), pp. 740–760. Univ. of Arizona, Tucson.

Yamaguchi A., Taylor G. J., and Keil K. (1997) Metamorphic history of the eucritic crust of 4 Vesta. *J. Geophys. Res., 102,* 13381–13286.

Yomogida K. and Matsui T. (1984) Multiple parent bodies of ordinary chondrites. *Earth Planet. Sci. Lett., 68,* 34–42.

Young E. D., Ash R. D., England P., and Rumble D. III (1999) Fluid flow in chondrite parent bodies: Deciphering the compositions of planetesimals. *Science, 286,* 1331–1335.

Zolensky M. E., Bourcier W. L., and Gooding J. L. (1989) Aqueous alteration on the hydrous asteroids: Results of EQ3/6 computer simulations. *Icarus, 78,* 411–425.

Zolensky M. E., Weisberg M. K., Buchanan P. C., and Mittlefehldt D. W. (1996) Mineralogy of carbonaceous chondrite clasts in HED achondrites and the Moon. *Meteoritics & Planet. Sci., 31,* 518–537.