Excerpt from the application

1 State of the art, preparatory work

1.1 State of the art

A lot of information on the processes operating in the protoplanetary accretion disk, that ultimately resulted in the formation of the small and large bodies of our Solar System, and on the earliest evolutionary history of these bodies has survived in meteoritic and cometary material. The main sources of information are meteorites because they have since long been recognized as ancient Solar System material, and ample material is available for laboratory investigations.

Chronological data used for input of thermal models  In recent years, studies using high resolution radioisotope chronology yielded significant advance in understanding formation and evolutionary time scales of meteorite parent bodies, i.e., the small precursor planetesimals that later formed the larger planets. One line of evidence demonstrated that the internal heat source for early differentiation and metamorphism was decay heat of short-lived nuclides like \(^{26}\text{Al}\) or \(^{60}\text{Fe}\) (Göpel et al. 1994; Trieloff et al. 2003; Kleine et al. 2008; Bouvier et al. 2007; McSween et al. 2002). Another line of evidence has shown that differentiated meteorite parent bodies are among the oldest objects, and formed more or less contemporaneously with CAIs (Kleine et al. 2005; Schersten et al. 2006; Bizzarro et al. 2005; Kleine et al. 2008), earlier than chondrite parent bodies that must have postdated individual chondrule formation, which in most cases occurred 1 – 4 Ma after CAIs (Kita et al. 2000; Kunihiro et al. 2004). These two lines of evidence are consistent with thermal models that describe asteroidal evolution due to internal \(^{26}\text{Al}\) heating (Miyamoto et al. 1981; McSween et al. 2002; Hevey & Sanders 2006): If planetesimals are larger than tens of km, the maximum degree of internal heating is given by the initial \(^{26}\text{Al}\) abundance, i.e., mainly formation time: early formed (0 – 1.5 Ma after CAIs) planetesimals will heat up strongly and differentiate, yielding iron meteorites and basaltic achondrites. Later formed planetesimals (>1 Ma after CAIs) heat up without differentiation, yielding thermally metamorphosed chondritic parent bodies (ordinary, enstatite chondrites, or more strongly heated Acapulcoites and Lodranites) or aqueously altered parent bodies (carbonaceous chondrites). Primitive chondrites (such as petrologic type 3) can survive in the outer cool layers of larger parent bodies (onion shell model Göpel et al. 1994; Trieloff et al. 2003), or in bodies that never grew larger than 10 – 20 km in size (Hevey & Sanders 2006; Yomogida & Matsui 1984), or just simply in bodies that formed relatively late. Such principally different scenarios (e.g., if mildly metamorphosed meteorites originate from early formed bodies and low layering depth or from more interior regions of lately formed bodies) can be distinguished via their cooling rates: low layering depths implies faster cooling than layering in larger depths of the parent body. This cooling paths can be evaluated by a variety of thermo-chronological methods, e.g., applying chronometers with different closure temperatures such as Hf-W (Kleine et al. 2005; Kleine et al. 2008), U-Pb-Pb (Göpel et al. 1994; Bouvier et al. 2007), Ar-Ar (Trieloff et al. 2003) or \(^{244}\text{Pu}\) fission tracks to individual meteorites of different petrologic type, in order to obtain a consistent data set for cooling curves for a distinct asteroid. For example,
in this way an onion shell structure could be derived for the H chondrite or A-L parent body, and modeling approaches provide constraints on their formation time (Trieloff et al. 2003; Kleine et al. 2005; Hevey & Sanders 2006).

Important parameters that controlled the different thermal histories of chondrite parent bodies include the time and duration of their accretion (and hence the amounts of short-lived radionuclides such as $^{26}$Al and possibly also $^{60}$Fe), their terminal size, and compositional variations such as the amounts of water ice. Information on these parameters can be obtained by dating chondrites and their components. Such ages provide two important constraints. First, ages for chondrules, which formed by transient heating events in the solar nebula prior to parent body accretion, give the earliest possible time of accretion. Second, ages for metamorphosed chondrites as well as ages for components formed during aqueous alteration (e.g., carbonates) provide information on the thermal evolution and internal structure of chondrite parent bodies. The chronology of chondrule formation has been studied intensively using the $^{26}$Al-26Mg system and earlier studies returned ages ranging from $\sim 1$ to $>4$ Myr after CAI formation. More recent studies showed that this range in ages in part reflects partial or complete resetting of the Al-Mg system during parent body metamorphism and, by considering only the most primitive chondrites, these studies provide the following range in chondrule formation ages: L and LL chondrules $\sim 1$ to $\sim 2.5$ Myr; CO chondrules $\sim 1.7$ to $\sim 3.0$ Myr; CR chondrules $\sim 2$ to $>4$ Myr. These data suggest that chondrules from carbonaceous chondrites (CO, CR) formed over a longer period of time than those from ordinary chondrites. It has been argued that the younger ages for some chondrules from carbonaceous chondrites cannot reflect resetting by parent body metamorphism because these samples have not been thermally metamorphosed. The effects of low-temperature alteration however were not studied in great detail yet but Kurahashi et al. (2004) showed that chondrules from the CO3.05 chondrite Yamato 80320 do not show any signs of alteration, yet their Al-Mg ages range from $\sim 1.7$ to $\sim 3.0$ Myr after CAIs. The age of chondrules has also been studied using the U-Pb and Mn-Cr systems. However, Pb-Pb ages are available for only a few chondrules and their interpretation in terms of formation intervals relative to CAIs is currently uncertain due to uncertainties in the absolute age of CAIs. Mn-Cr ages are only available for bulk chondrules and these ages may reflect volatile depletion of chondrule precursors rather than formation of individual chondrules. Moreover, the effects of parent body processes on the U-Pb and Mn-Cr systematics of chondrules have not been studied yet. Therefore, the current estimates for the duration of chondrule formation rely almost entirely on the Al-Mg ages. The thermal evolution of chondrite parent bodies has been quantified using various short- and long-lived isotope chronometers (e.g., the $^{182}$Hf-$^{182}$W, $^{207}$Pb-$^{206}$Pb and $^{39}$Ar-$^{40}$Ar systems). The ages obtained from each of these chronometers correspond to the time of cooling below the closure temperature ($T_c$) of a particular chronometer. This temperature, at which diffusional exchange of parent and daughter isotopes stopped, depends on the diffusivities of the elements of interest and thus is different for the various chronometers. Thus, the cooling history of a chondrite can be determined by applying to this sample several chronometers that have different closure temperatures. Alternatively, information on the cooling history can be obtained by applying one chronometer to a suite of co-genetic metamorphic rocks that derive from different burial depths (and hence have different cooling rates) in a single parent body. Göpel et al. (1994) presented Pb-Pb ages
for phosphates from L, LL, and H chondrites of petrologic types 4 to 6 and demonstrated that for H chondrites there is an inverse correlation of phosphate age with metamorphic grade, most consistent with an onion-shell structure of the H chondrite parent body. For the L and LL chondrites, however, such a correlation is not apparent, indicating a more complex cooling history. Combining the Pb-Pb data for H chondrites with $^{244}$Pu fission track and Ar-Ar ages, Trieloff et al. (2003) obtained well-defined cooling curves for a suite of unshocked H chondrites and demonstrated that these samples derive from a part of the H chondrite parent body that cooled more or less undisturbed. This interpretation is consistent with Hf-W ages for a range of H chondrites. In contrast to the previously described chronometers, the Hf-W system has a much higher closure temperature (ranging from $\sim 750^\circ C$ for H4 chondrites to $\sim 875^\circ C$ for H6 chondrites), which for H5 and H6 chondrites is similar to their peak metamorphic temperatures. As such, Hf-W ages provide essential information for constraining the thermal evolution of chondrite parent bodies. For the H chondrite parent body they reveal that peak temperatures in its center were reached at $\sim 10$ Myr after CAI formation. Metallographic cooling rates from a suite of H chondrites are inconsistent with cooling inside a parent body with an onion-shell structure, suggesting that parts of the H chondrite parent body may have been disrupted by impacts. Recently, the Mn-Cr system has also been used to constrain the thermal evolution of chondrite parent bodies. Trinquier et al. (2008) reported a Mn-Cr age for Ste. Marguerite (H4) that is indistinguishable from the Pb-Pb age of its phosphates. Similarly, Polnau & Lugmair (2001) reported a Mn-Cr age for Richardson (H5) that agrees with the Pb-Pb age of its phosphates. Shukolyukov & Lugmair (2004) reported Mn-Cr ages for type 4 and 6 enstatite chondrites ranging from $\sim 4565$ to $\sim 4560$ Ma. In the parent bodies of most carbonaceous chondrites, temperatures were too low to cause significant thermal metamorphism, but were sufficiently high for melting of ice. This led to aqueous alteration and the formation of several hydrous and anhydrous minerals including a variety of carbonates. Dating their formation provides important constraints on the thermal history of these bodies. The timing of carbonate formation in the CI and CM chondrites has mainly been studied using the Mn-Cr and Rb-Sr systems. There are several other carbonaceous chondrite groups that contain secondary carbonates, including the CR, CH, and CB chondrites, but no attempts have been made yet to determine the formation ages of these carbonates. The pioneering work of MacDougall et al. (1984) on carbonates from Orgueil (CI) revealed that aqueous alteration on the CI parent body started soon after accretion. More recently, Hoppe et al. (2008) presented Mn-Cr ages for several carbonate grains from Orgueil, which show that the aqueous activity started $\sim 3$ – $4$ Myr after CAI formation and lasted for several millions of years. The $^{53}$Mn-$^{53}$Cr system has also been used to date carbonates from other carbonaceous chondrites, especially the CM chondrites and also the polymict chondrite breccia Kaidun (containing CR-, CI- and CM-like materials). These studies obtained initial $^{53}$Mn/$^{55}$Mn ratios as high as $\sim 9 \times 10^{-6}$, corresponding to absolute ages of $\sim 4569$ Ma (as calculated relative to the angrite LEW86010). Consequently, these carbonates seem to be as old as CAIs, implying that the CM parent body accreted more or less instantaneously after condensation of the first solids. Such an early inferred accretion of the CM parent body is in marked contrast to the protracted accretion timescales for other chondrite parent bodies that must have accreted more than $\sim 2$ Myr after CAI formation (see above) and warrants further investigation.
It is expected that more data will become available in the near future (ETH Zürich, Kleine), also within the anticipated projects of the SPP, e.g., for enstatite chondrites. Constraints derived from modeling such high quality chronological data allow insight into the timing of planetesimal growth in various regions of the solar nebula (e.g., Trieloff & Palme 2006). However, the validity of conclusions derived from modeling approaches crucially depends on how realistic the applied models are.

For example, many studies applied simple analytical models (Miyamoto et al. 1981; Kleine et al. 2005). This model does not take into account an insulating regolith that could influence model outcome of layering depths significantly (e.g., Akridge et al. 1998, model of mega-regolith). Moreover, the initial state of planetesimals is likely not rocky but porous, which implies smaller heat conductivity and allow much smaller bodies to heat up significantly (Kleine et al. 2005). This could have had strong implications for parent body structure (e.g., raises the possibility that different petrologic types are derived from more than one parent body). Other effects that should be modeled in a more realistic approach are sintering of porous material during heating and transport of liquids and gases (e.g., water) in porous bodies.

In the following we outline the mathematical formulation of the problem and summarize what has been achieved by previous modeling approaches.

**Definition of the problem:** The thermal evolution of a spherically symmetric planetesimal is determined by the equation of energy conservation, that reads in spherical polar-coordinates as

\[
\frac{\partial c_v T}{\partial t} + \frac{p}{\rho^2} \frac{\partial \rho}{\partial t} = \frac{1}{\rho r^2} \frac{\partial}{\partial r} r^2 \rho c_v \kappa \frac{\partial T}{\partial r} + Q. \tag{1}
\]

Here \( T \) is the temperature, \( c_v \) the specific heat per unit volume at constant pressure, \( p \) the pressure, \( \rho \) the mass density, \( \kappa \) the thermal diffusivity, and \( Q \) the heat production per unit mass. The second term on the left hand side is the work done on the matter by compression or dilation, the first term on the right hand side describes heat flow by heat conduction and by radiative transfer. A transport of energy by convection is negligible in case of planetesimals.

The pressure stratification within a planetesimal is determined by the equation of hydrostatic equilibrium

\[
\frac{\partial p}{\partial r} = -\frac{G M_r}{r^2} \rho \quad \text{with} \quad M_r = 4\pi \int_0^r x^2 \rho(x) \, dx, \tag{2}
\]

where \( G \) is the gravitational constant and \( M_r \) the mass contained inside the sphere of radius \( r \). The system of the two basic equations (1) and (2) is closed by constitutive relations,

\[
\rho = \rho(T, p, X_i), \quad \kappa = \kappa(T, \rho, X_i), \quad c_v = c_v(T, \rho, X_i), \tag{3}
\]

the equation of state for the planetesimal material, and equations for calculating the specific heat \( c_v \) and thermal diffusivity \( \kappa \). Additionally one has to define the heat production \( Q(T, \rho, X_i) \), which may contain the release of gravitational energy by contraction if the body shrinks due to compaction.

Further one needs to specify the composition of the material, i.e., the different materials from which the planetesimal material is formed and their mass-fractions
In case of an inhomogeneous composition that evolves over time one has for each species an additional transport-diffusion-reaction equation for the temporal and spatial evolution of the mass-fractions $X_i$.

A planetesimal grows in size over time by colliding and merging with smaller planetesimals. This growth takes place by a series of discrete events that result in a jump-like increase in mass in each step. This process has in principle to be treated as a stochastic process, but presently it is generally simplified by treating growth as a continuous process with a specified average mass-infall rate $\dot{M}_{\text{inf}}$:

$$\frac{dM_r}{dt} \bigg|_{r=R} = \dot{M}_{\text{inf}} \quad (4)$$

where $R$ is the radius of the planetesimal.

Equation (1) is subject to boundary and initial conditions. Besides the trivial condition $\partial T/\partial r = 0$ at the centre one has at the outer boundary $R$ of a planetesimal an equation that describes the energy exchange of the body with its surrounding. The details of this depend strongly on (i) whether the planetesimal is still embedded in a protoplanetary accretion disk and the surface interacts with the gaseous surrounding, or if (ii) the disk is already dissipated and the planetesimal is subject to irradiation by the sun, and if (iii) there is still noticeable mass-infall. The position of the boundary $R$ varies with time if the planetesimal is growing.

**Formation of planetesimals** Many models for the thermal evolution of planetesimals have been published, that solve the basic set of equations in various approximations. The most simple approximation is that of Miyamoto et al. (1981) which solves equation (1) assuming constant coefficients $\kappa$ and $c_v$, heating by decay of $^{26}$Al and fixed boundary temperature. An improved version is Bennett & McSween (1996). This kind of model has usually been used to investigate the cooling history of planetesimals from laboratory data on chondrites.

A much more sophisticated model for ice-free planetesimals is Yomogida & Matsui (1984) that uses data on temperature dependence on $c_v$ and $\kappa$ derived by measurement made on chondrites. This model is also the sole model that considers the dependence of $\kappa$ on porosity and the evolution of porosity with heating on the basis of solving an equation for the evolution of pore size with applied temperature. The insulating influence of a regolith layer on the surface was discussed by Haack et al. (1990); Akridge et al. (1998). The influence of a temperature dependence of $c_v$ on models was studied by Ghosh & McSween (1999). Models of specific objects have also been constructed, e.g., for (4) Vesta (Ghosh & McSween 1998) and (6) Hebe (Ghosh et al. 2003). The latter discusses the importance of growth for heating models (see also Ghosh et al. 2006).

A lot of effort has been spent on developing models for icy planetesimals that consider melting and out-gassing (e.g., Grimm & McSween 1989; Prialnik & Podolak 1995; Cohen & Coker 2000; Prialnik 2002; Merk et al. 2002; Merk & Prialnik 2003, 2006; Rosenberg & Prialnik 2007; De Sanctis et al. 2007; Prialnik & Merk 2008). Equation (1) is solved in this case with an appropriate consideration of latent heats in the source term $Q$ and recipes for calculating $\kappa$ and $c_v$ for the ice-dust mixture. The main application for such models are comets and Kuiper-belt objects, but also carbonaceous chondrites. The way how porosity and heterogeneous composition can
be treated in such models beyond simple mixing rules is discussed in Sirono \& Yamamoto (1997); Shoshani et al. (1997); Shoshany et al. (2002); Leliwa-Kopytyński \& Kossacki (2000). Also the explosion of bodies by vapor overpressure was discussed (Wilson et al. 1999). The composition of the ice mixture has been modeled on the basis of accretion disk models by, e.g., Mousis et al. (2008).

Models considering the interplay of slow heating of bodies and growth by accretion of smaller bodies from a planetesimal swarm have been constructed by (Prialnik 2002; Merk et al. 2002; Merk \& Prialnik 2003, 2006) on the basis of a simplified coagulation model. This model is suited to directly being coupled with the thermal evolution model, providing the boundary condition Eq. (4). Growth and its influence on thermal evolution has been discussed also in Ghosh et al. (2006), based on models of planetary formation, but this cannot be directly integrated in thermal evolution models.

Some recent reviews on the whole problem are McSween et al. (2002); Ghosh et al. (2006); Prialnik et al. (2008).

**Transport of gaseous material ad water** The transport of gaseous material and the flow of water through the porous body of a planetesimal is important for its compositional evolution. This problem has been treated in a number of thermal evolution models (e.g. Grimm \& McSween 1989; Hashizume \& Sugiura 1998; Young 2001; McCord \& Sotin 2005)

**Typical questions that can be addressed** by a more sophisticated modeling approach are:

1. Meteorite parent bodies:

   - What are the minimum sizes for specific chondritic parent bodies derived from cooling time intervals measured by isotope chronology?
   - What is the quantitative relationship between formation time and internal heating due to short-lived nuclide decay heat of $^{26}$Al and $^{60}$Fe?
   - Are model-derived chondrite formation time scales consistent with formation ages of their individual chondrule populations?
   - Is it possible to achieve fast cooling of highly metamorphosed chondrites like Acapulco in a body that remained undifferentiated in interior regions? Do acapulcoites and lodranites originate from a small undifferentiated body, or an undifferentiated surface layer of an internally differentiated body?

2. Icy planetesimals in the outer solar system:

   - The standard accretion theory for Jupiter formation are core accretion models (Lissauer \& Stevenson 2007) in which a ten mass Earth core of icy and rocky planetesimals accreted the gaseous envelope from the surrounding solar nebula. Discussing dynamical arguments, Scott (2007) concluded that Jupiter was present after 3–5 Ma, within the “regular” lifetime of protoplanetary disks of about 3–6 Ma (Haisch et al. 2001).
The fast accretion of a ten earth mass core is thought to be enhanced by the larger mass density beyond the snow line, in the form of solid ices. However, if icy planetesimals form shortly after CAIs, these will likely melt and lose their water, annulling the mass density advantage beyond the snow line. Improved modeling of these precursor bodies put additional important constraints on the formation time scale of Jupiter and its core.

- Comets formed in the outskirts of our planetary system. When entering the inner solar system, they release hyper-volatile species such as CO and CO$_2$. If $^{26}$Al was homogeneously distributed in the early solar system (for which there is ample evidence, e.g., Thrane et al. 2006), parent body heating should have affected comets as well, meaning that their formation was so late that they were able to retain their volatile inventory. Improved modeling can constrain their formation time scales (Prialnik et al. 1987, modeled 6–7 Ma after CAIs) which is very close to current maximum lifetime estimates of protoplanetary disks. As comets needed solar gas to grow, this also constrains the lifetime of the solar nebula.

1.2 Preparatory work

Modeling of protoplanetary accretion disks: Protoplanetary accretion disks are the birthplaces of planetesimals. Their properties and composition define the properties of the environment and the chemical composition of the material from which the bodies growing in the region of the asteroid belt and closer to the proto-sun are formed. They also define the properties of the environment in which bodies with sizes of tens to hundreds of km diameter initially grew within the accretion disk during the first million years.

The structure and chemical and mineralogical composition of such disks has been
modeled since many years in stationary or time dependent models in the one-zone (Gail 1998, 2001, 2002, 2004; Wehrstedt & Gail 2002, 2003; Duschl et al. 1996) the (1+1)-dim approximation (Wehrstedt & Gail 2008) and in 2D hydrodynamic models (Keller & Gail 2004; Tscharnuter & Gail 2007). Additional modeling of disk evolution and chemical and mineralogical evolution of the disk material is presently performed within the frame of the Forschergruppe 759. The results and the existing codes can be used for the present project in order to determine (i) the radial and temporal variation of the mineral composition of the input material for the formation of planetesimals and larger bodies, and (ii) the local gas pressure and temperature conditions in the accretion disk as long as the disk is not yet dispersed. Figure 1 shows as an example for such model calculations the pressure-temperature stratification in the midplane of the Solar Nebula, resulting from a time dependent model of the evolution of the accretion disk around a solar type star.

Existing calculations show that there is a significant radial and temporal variation of the composition of the dust material in accretion disks. Though present model calculations are certainly not sophisticated enough to predict quantitatively correct the composition of the input material for planetesimal formation, the results of such calculations provide at least qualitatively an overview of the possible initial compositions of the material of planetesimals in dependence on the place and instant of their formation in the accretion disk. Figure 2 shows as an example the mass fractions of a number of minerals and of water ice in freshly formed planetesimals according to a calculation of time-dependent (1+1)-dim disk model that considers also the processing of dust (vaporisation/condensation, annealing of amorphous materials, oxidation of interstellar carbon dust) and radial mixing processes in the disk.

The thermal evolution of the growing bodies in a planetesimal swarm depends on the heating processes in their interior and on the heat exchange at their surface with their surrounding. During the first million years they are embedded in an accretion disk. Cooling of the surface of planetesimals during this initial phase is strongly coupled to the temperature of the accretion disk. The precise boundary conditions can be determined from our accretion disk models. If it should turn out that the biggest bodies become much hotter at their surface than the accretion disk material before disk dispersal, it may be necessary to model heat exchange between

Figure 2. Initial compositions of planetesimals (mass fractions) in the range of terrestrial planets and the asteroidal belt at two instants, $1 \times 10^6$ yr (left) and $3 \times 10^6$ yr (right). Results of a model calculation of the evolution of an accretion disk around a star with $1 \, M_\odot$. 

Figure 1 shows as an example for such model calculations the pressure-temperature stratification in the midplane of the Solar Nebula, resulting from a time dependent model of the evolution of the accretion disk around a solar type star.
Figure 3. Heat diffusivity of the porous material in ice-free planetesimals. The mixture of solids is taken from disk models, the porosity is assumed to be 20%. The heat diffusivity is calculated from recipes for calculating heat diffusivity of composed media. Full line: Calculated heat diffusivity by phonon and radiative heat conduction. Dotted line: Contribution of radiative transfer (Hofmeister 1999). Dashed line: Analytical approximation to laboratory measured values of meteoritic material with porosity of ≈ 20% (Yomogida & Matsui 1983).

Modeling of heat capacity and thermal conductivity: The thermal evolution of small bodies of up to a size of several hundred kilometres is determined by the heat capacity and thermal conductivity of porous materials with heterogeneous composition. For investigating the possibility of a construction of thermal evolution models that consider this crucial point, a preliminary model program for calculating these quantities from the properties of the individual components was developed. For heat capacities this is a trivial task since data for all relevant materials are available (e.g. Barin 1995). For heat conduction the concepts developed for technical purposes to calculate heat conductivity of composed porous and granular media from properties of components as reviewed in, e.g., Berryman (1995) can be applied. Properties of components can be derived by the method of Hofmeister (1999). Figure 3 shows as example the heat diffusivity of a porous mineral mixture at 3AU (left part of Fig. 2), taken from a model calculation of the Solar Nebula and measured heat diffusivity of chondritic material with about 20% porosity. The application of the mixing rules developed for technical purposes produces results of acceptable accuracy for modeling planetesimals.

Modeling of thermal evolution: Model calculations of the thermal evolution of planetesimals have up to now be done for the simplified case of solving the heat conduction equation (1) with constant coefficients and heating by decay of short lived radioactives (Trieloff et al. 2003; Kleine et al. 2005). Numerical calculations based on the full set of equations of hydrostatic equilibrium (2) and the energy equation (1) using realistic heat diffusivities and heat capacities have not yet been published, but there is a rich experience in the numerical solution of partial differential equations,
Figure 4. Thermal evolution of a growing planetesimal heated by $^{26}$Al decay. Matter composition as determined from an accretion disk model (essentially equal to chondritic composition), porosity assumed as 20%. The planetesimal is assumed to form at 3AU, to start growing $2 \times 10^6$ yr after formation of the proto-sun, and to grow to a maximum size of 100 km. The Solar Nebula is assumed to disappear between 5 and 5.5 Ma.

in particular of large sets of parabolic equations and hydrodynamic equations, from other projects (Gail 2002; Wehrstedt & Gail 2002; Tscharnuter & Gail 2007). A first version of a code for the model calculation of thermal evolution of growing planetesimals has already been developed and will form the starting point for the code development of this project. The numerical solution algorithm is based on the method of lines that proved to be well suited for treating growing bodies. An example of a solution is shown in Fig. 4

**Chronology:** The cooperating partners of this proposal have contributed part of the body of data that can - up to now - be used for thermal modeling.

The main area of expertise of Thorsten Kleine is in isotope geo- and cosmochemistry and, more specifically, in determining the timescales of early solar system processes. His main research topic in the past has been in the application of the Hf-W system to date a variety of processes in the evolution of planetary bodies. For instance, Hf-W chronometry provides the currently most precise determinations of the age of the Earth and its core, the age of the Moon and its magma ocean, and the timescale of core formation and mantle differentiation in Mars. Based on precise W isotope measurements of iron meteorites it could be shown that differentiation of asteroids predated the formation of chondrites. This has changed our view of how planetesimals first formed and differentiated. Hf-W chronometry also provides constraints on the timescales of magmatism on asteroids as well as on their metamorphic history, the main topic of this proposal. Current projects, which are also relevant for the research proposed here, includes the investigation of Mo isotope heterogeneities in the proto-planetary disk, the development and application of precise Sr isotope
measurements to early solar system processes, and the application of $^{146}\text{Sm}-^{142}\text{Nd}$ chronometry to meteorites and lunar samples. Ongoing research projects (funded by swiss SNF) that will be pursued in parallel to the research proposed here include the application of the Hf-W chronometer to a suite of metamorphosed chondrites from the ordinary (especially the L and LL), enstatite (EH and EL), and carbonaceous (CK) chondrite parent bodies. The interpretation of these results will be done using the thermal model developed in the proposed project.

The main expertise by M. Trieloff is noble gas geo- and cosmochemistry (Trieloff et al. 2000, 2002, 2003), incl. Ar-Ar dating of meteorites (Trieloff et al. 2003; Pellias et al. 1997; Korochantseva et al. 2005, 2007). Chronological data on H chondrites (Trieloff et al. 2003) confirmed the inverse correlation of cooling rate and petrologic type of undisturbed H chondrites based on $^{40}\text{Ar}-^{39}\text{Ar}$ and $^{244}\text{Pu}$ fission track ages (Trieloff et al. 2003) and U-Pb-Pb ages (Göpel et al. 1994; Bouvier et al. 2007). This demonstrated that at least part of the H chondrites cooled in an onion shell like structured asteroid that was internally heated by $^{26}\text{Al}$ decay heat and is the base for thermal modeling outlined here. Thermochronological data were also obtained for the A-L parent body (Pellias et al. 1997).

2 Goals

This project is an interdisciplinary cooperation linking isotope chronology of early solar system meteorite parent bodies and planetesimal modeling using methods common in theoretical astrophysics. Present data can be modeled with an existing preliminary code, but further code development is the primary task of this project. During runtime of this SPP, it is expected from accompanying proposals (ElGoresy, Bayreuth; Trieloff and Ott, Heidelberg and Mainz) and studies abroad (Kleine, ETH Zürich; Boyet, Clermont Ferrand; Bouvier, Tempe) that further data will be become available for modeling. Hence, intense cooperation inside and outside the SPP is envisaged.

This project aims to combine (i) modeling of the internal evolution and composition of parent bodies of chondrites with (ii) modeling of their initial composition as determined from evolutionary models of the protoplanetary accretion disk (Solar Nebula) and (iii) comparison with information from laboratory investigations of meteorites. This will enable much improved determinations of the cooling history and size determinations of parent bodies and more precise dating of processes in the Solar Nebula. Additionally it will allow to estimate properties and compositions of planetesimals and protoplanets in the region closer to the sun than the present day asteroid belt. From this it will be possible to check if there are scattered bodies from the region of terrestrial planet formation (0.1 AU to $\sim$ 2 AU) in the asteroid belt ($\sim$ 2 AU to $\sim$ 4 AU) that provide us with meteoritic material initially formed much closer to the young sun. It will also be possible to estimate properties and compositions of small bodies in the innermost solar system, that have not left direct successors in the present day Solar System, and it will be possible to predict the properties and composition of the building blocks of earth-like planets in extrasolar systems.

It is aimed to develop by close interaction between theoretical modeling and comparison with the empirical findings on the evolution of the early Solar System, inferred from its traces recorded in the meteoritic material, a reliable tool for further
cosmochemical investigations.

Theory

It is planned to develop a model program that follows the evolution of a growing body and its internal chemical, mineralogical and chemical evolution during heating up by decays of long lived radioactives, in particular $^{26}$Al and $^{60}$Fe, but also the more long-lived ones, and the subsequent cooling phase up to $10^8$ years.

The size of objects to be treated will be limited to bodies that do not become hot enough for melting of silicates and differentiation into a silicate mantle and iron core. Differentiation of protoplanets is treated in a project by Spohn & Breuer. Melting and vaporisation of water ice, however, will be included in the model calculation.

The location of the ice-front moves during the evolution of the protoplanetary accretion disk. It is locates at a distance of about 5 AU during the early evolution ($t < 10^6$ yr) and gradually moves inward as the disk loses mass to the star. At $t = 3 \times 10^6$ yr it is located at about 3 AU (Fig. 2). The planetesimals in the zone of the present day asteroid belt are formed in part with a considerable ice content, in part without any ice. The planetesimals formed closer to the sun are formed without any water ice. We will model the evolution in both cases: Bodies with and without initial ice content. The processes related to ice melting and water vapor transport will be treated similar as in Grimm & McSween (1989); Hashizume & Sugiura (1998); Young (2001); Merk et al. (2002).

Generally, the initial composition of the bodies will be derived from accretion disk models and from considerations on the composition of meteoritic material.

For bodies formed without or with only small quantities of ice, one has to consider, that they will contain some carbonaceous material from interstellar space, since about 60% of the carbon is bound in carbon dust and at least one half of it in a refractory component (Pollack et al. 1994; Zubko et al. 2004) that exists to much higher temperatures than ice (cf. Fig. 2). If this carbonaceous material reacts at elevated temperature with residual volatile oxygen-bearing material, all oxygen not bound in refractory minerals will be consumed and the interior of such bodies will attain very low oxygen fugacities and unusual mineralogy. It will be attempted to consider this by chemical equilibrium calculations. This initial presence of some carbonaceous material seems to have never been considered in previous calculations.

The planetesimals will begin their life as bodies formed from highly porous material, witnessed by the low density of cometary nuclei. As bodies gradually become heated, the material starts sintering. This process will be included in the same approximation as in Yomogida & Matsui (1984). This adds to the equations for the structure and thermal evolution an additional differential equation for the porosity, that will self-consistently be treated with the other equations of the internal constitution of the bodies. This is important for applications to geochronology since they heavily rely on closure temperatures for diffusion processes, and temperature depends on the heat conductivity of the material, which in turn depends strongly on porosity. Heat conductivity then will develop a strong radial variation between high (for compact material) and low values (for pulverized material or regolith) if one moves from the interior to the surface.

The coefficient of heat conduction will be determined from methods developed to estimate this from the properties of components for composite and porous media.
for technical applications (cf. the review Berryman 1995). For calculating the heat conductivity of the mineral compounds one can proceed as in Hofmeister (1999).

The consistent treatment of heat production, heat conduction of porous media, and evolution of porosity with heating is a central point of the project. It will allow for the first time to consider this important effect in determinations of the cooling history of meteoritic parent bodies.

With respect to the growth of the bodies from the initial planetesimals with sizes of $\approx 1 \text{ km}$ to bodies with sizes of hundreds of km we will have to introduce, for reasons of computing time requirements, a simplified treatment. It is planned to use both a simple growth equation approach (e.g., Merk et al. 2002) or a simplified coagulation approach as in Prialnik (2002); Merk & Prialnik (2006). The effect of mixing of bodies from different regions of the planetesimal swarm cannot be treated by such a model; here we have to use some information from other calculations (e.g., Petit et al. 2001).

The basic part of the model program will be a program for solving the hydrostatic equilibrium and the energy equation for a growing body in spherically symmetric geometry. More complicated geometries are not appropriate for this first step. The program will be constructed to treat bodies up to sizes of the order of 1000 km. In this case central pressures do not exceed 10 kbar and compressibility of the material is only a minor effect. Compressibility will therefore (optionally) either be neglected or treated in the Monaghan approximation for the equation of state.

The heat conduction equation (1), the equation of hydrostatic pressure stratification with self-gravity (2), and the equation for evolution of porosity (Yomogida & Matsui 1984) will be solved implicitly in Lagrangean coordinates on a staggered grid. First the problem will be solved by the method of lines. The growth of the planet will be treated by letting the outermost layer to increase in mass up to some prescribed maximum and adding a new layer if the maximum is exceeded. Test calculations have shown that this is well suited for treating growing bodies and will allow to obtain first results rather rapidly. This can be used for model calculations of planetesimals with no or small initial ice content.

In a second step a code will be developed that uses a moving grid according to the method of Dorfi & Drury, as described in Dorfi (1998). To this end a grid equation has to be added to the code and advection terms have to be introduced in the heat transport equation. A moving grid will be better suited to treat moving boundaries associated with compositional inhomogeneities and phase transitions. This will be more appropriate for model calculations of bodies with significant ice content.

With respect to the outer boundary condition for the heat transport equation the existence of an accretion disk during the first million years will be considered. The surface temperature will be higher in this case than for a body in empty space illuminated by a star. The properties of the disk will be taken from model calculations.

Model calculations will be performed with immediate application to experimental results performed by M. Trieloff and A. El Goresy in other SPP projects, as well as data of cooperating partners Th. Kleine (ETH Zürich) and other foreign scientists (M. Boyet, Clermont Ferrand, France; A. Bouvier, Tempe AZ).
Comparison with the meteoritic record

- Thermochronological data of H chondrites (already available) and enstatite chondrites (SPP projects) can be modeled by existing codes and the code to be developed.

- Parent body heating and cooling constrains the formation time scale which can be compared with the specific age of chondrule populations. This yields information on the relative timing of chondrule and chondrite formation (parent body growth slow or fast after chondrule formation?)

- Formation and evolution time scales of internally heated cometary bodies: how can comets retain their volatile inventory?

- Water loss and formation of icy planetesimals in the Jupiter formation region: Under which circumstances can water ice contribute to the fast growth of Jupiters core, or is it rather lost from internally heated planetesimals?

- Testing different starting materials of planetesimals: clues to enstatite chondrite formation – which circumstances imply reducing conditions?

- How did planetesimals look like in other regions of the early solar nebula, e.g., the terrestrial planet formation region?

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