

**TRACE ELEMENT FRACTIONATION IN DIFFERENTIATED PLANETESIMALS BY NON-EQUILIBRIUM PARTIAL MELTING.** H. Palme and S. Chakraborty, Universität zu Köln, Mineralogisch-Petrographisches Institut, Zülpicher Strasse 49b, 50674 Köln, Germany.

A comparatively large number of meteorite parent bodies has undergone partial or complete melting. Eucrites, for example, are either partial melts or residues of fractional crystallization of a chondritic parent body. Lodranites and ureilites are solid refractory residues of partial melting processes. As melting must have occurred shortly after accretion, the decay of  $^{26}\text{Al}$  has been suggested as appropriate heat source. It is shown here that partial melting in a sufficiently large body heated by  $^{26}\text{Al}$  will lead to unusual trace element fractionations due to incomplete equilibration between melt and residual solid. Although major elements are at all times in equilibrium between melt and solid residue trace elements are not. The effect of non-equilibrium melting is most dramatically seen in the behaviour of Eu. The first phase to melt in a chondritic parent body is plagioclase, the major carrier of Eu (more than 80 %). This leads to a strong positive Eu-anomaly in the early-formed melt. During and after partial melting Eu and other trace elements will continually reequilibrate with the solid residue. A major fraction of Eu will re-partition into the minerals of the solid residue attempting to reduce the Eu-anomaly of the melt. If the melt separates immediately after formation, there is not enough time to establish trace element equilibrium between the partial melt and the residue. In a chondritic planetesimal, heated by  $^{26}\text{Al}$  decay, the formation of a 10 % partial melt will take several thousand years. However, because of the slow diffusion of REE in pyroxene, equilibration of the melt with the residue requires much longer times (at least 50 000 years assuming 1 mm grain size in the residue). Negative Eu-anomalies and other trace element characteristics in lodranites and ureilites are readily explained by non-equilibrium melting. If the eucrite parent body was heated by decay of  $^{26}\text{Al}$  trace equilibrium between eucritic melts and residual solid is unlikely, at least at small degrees of partial melting. The effects of non-equilibrium melting critically depend on the diffusivities of trace elements in residual minerals, the grain size of residual minerals, the phases that contain the trace elements and the time scale for formation and separation of the melt.

*Formation of partial melts by  $^{26}\text{Al}$  heating:* The presence of life  $^{26}\text{Al}$  in the early solar system is well established [1]. The  $^{26}\text{Al}/^{27}\text{Al}$  ratio at the beginning of the formation of the solar system 4.6 by ago was  $5 \cdot 10^{-5}$ . Heating by decay of  $^{26}\text{Al}$  would be most effective if accretion of a planetesimal began immediately after the closure of the solar system from further supply of fresh interstellar material. With an Al-content of 1.5 %, typical of type 3 carbonaceous chondrites, and an assumed heat capacity ( $c_p$ ) of 1.25 J/g it would take 125 years to raise the temperature in perfectly insulated material by one degree. The silicate solidus temperature (1100°C, e.g. [2]) of an early accreted body could thus be reached within about 170 000 years. By this time only 17 % of the  $^{26}\text{Al}$  would have decayed. Once the solidus is reached silicates will begin melting. Plagioclase will be the first phase to melt. Since chondrites contain about 10 % of albitic plagioclase (An<sub>10</sub>) the first 10 % of melt will to a large extent consist of plagioclase with minor amounts of other phases. The heat of fusion of anorthite is 291 J/g compared to 226 J/g for albite at the respective melting points [3]. Assuming a conservative value of 300 J/g, 10 % of partial melting, predominantly of plagioclase would require some 3000 years. The effective time of equilibration between melt and crystalline residue is, however, shorter, since the melt fraction increases gradually from zero at the beginning to, for example 10 %, at the time of separation of the magma. As is shown below, the time the partial melt is in contact with the solid is by at least one order of magnitude below the time required for establishing equilibrium partitioning of REE between melt and pyroxene, the main carrier of most REE in these rocks.

*Equilibration of melts with residues:* Non-equilibrium melting has been discussed by several authors primarily to explain unusual trace element patterns, e.g. [4,5]. Recent calculations by Chakraborty and Lasaga [6] provided a more quantitative basis for understanding the effects of non-equilibrium melting. In this model equilibrium distribution of trace elements between a melt and all residual phases is calculated as a function of time. For a simple test case we assumed a chondritic source with ol, opx, cpx and plag in appropriate ratios. Trace element contents were calculated assuming equilibrium distribution among phases and bulk uniform CI-normalized enrichment factors of 1.3. In such an assemblage about 85 % of Eu, 40 % of Sm and 0.4 % of Yb are contained in plagioclase. It is then assumed that plagioclase melts completely, accommodating all elements that were earlier contained in plagioclase. The time to reequilibrate this melt with the residue is then calculated, assuming grains of 1 mm diameter and conventional mineral/liquid equilibrium partition coefficients. Eu will then move back into cpx and Yb will diffuse out of cpx into the melt. Some of the results of the calculation are shown in the Fig. It can be seen that nearly 100 000 years are required to decrease the Eu in the melt by 50 % which is still far from

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equilibrium. As indicated in the Fig. the decrease of Eu in the melt is accompanied by a simultaneous increase in cpx, resulting in strongly zoned cpx crystals. Yb has to go the opposite way, leading to extremely inhomogeneous cpx crystals (The Fig. shows cpx profiles of Eu and Yb after 200 000 years). Equilibration will thus tend to flatten the originally steep plagioclase REE-pattern and at the same time remove the Eu-anomaly. All calculations were done with diffusivities of  $10^{-15}$  cm<sup>2</sup>/sec. This is in the range of diffusivities of Sm for natural diopside and two orders of magnitude above that of synthetic diopside [7].

*Discussion:* Lodranites are the best example for residues produced by non-equilibrium partial melting. Lodranites have negative Eu-anomalies, depletion of light REE and extremely low Al-contents [8]. Ureilites are similarly low in Al but have more complex REE-patterns. Many ureilites have negative Eu-anomalies and many are depleted in light REE [9]. It requires several steps to explain a negative Eu-anomaly in a refractory residue (lodranites, ureilites) with conventional models assuming equilibrium distribution of trace elements between melt and solid [9]. Non-equilibrium melting provides a much simpler explanation. Additional enrichment of light REE is, however, required to explain the V-shaped REE-pattern found in several ureilites [9].

The enrichment of light REE and the negative Eu-anomaly in some eucrites (Stannern and Bouvante) have been explained by 5 % equilibrium partial melting of a chondritic source with some plagioclase left in the residue [10]. This requires equilibrium between melt and residue which cannot be achieved if the eucrite parent body was heated by the decay of <sup>26</sup>Al, making an origin by fractional crystallization more plausible.

Differentiation of the Moon must have occurred very early in the lunar history. It is possible that non-equilibrium melting played some role. Early non-equilibrium partial melting of the outer parts of the Moon could lead to a plagioclase rich crust and would thus impose a negative Eu-anomaly on the residue, a potential source region for mare basalts. Complete melting of major parts of the Moon would not be required.

Lit.: [1] Lee T. (1977) *Ap. J. Lett.* **211**, L107; [2] Takahashi E. (1983) *Proc. 8th Symp. Ant. Met.*, Tokyo, p.168; [3] Robie R. et al. (1979) *Geol. Surv. Bull.* 1452; [4] Prinzhofer A. and Allegre C.J. (1985) *EPSL*, **74**, 251; [5] Bedard H. (1989) *EPSL*, **91**, 359; [6] Chakraborty S. and Lasaga A.C. (in prep.); [7] Sneeringer M. et al. (1984) *GCA*, **48** 1589; [8] Zipfel J. and Palme H. (1993) *LPSC XXIV*, 1579; [9] Spitz A. and Boynton W.V. (1991) *GCA*, **55**, 3417; [10] Consolmagno G.I. and Drake M.J. (1977) *GCA*, **41**, 1271.

