MECHANISMS OF HEAT AND MASS TRANSFER IN A DIFFERENTIATING EARTH. C.T. Herzberg, Department of Geological Sciences, Rutgers University, New Brunswick, N.J. 08903

The geochemical identity of the Earth's continental crust strongly suggests that some fraction of it first formed at about 3800 Myr ago and that new material was added to it episodically throughout the remainder of geological time (1,2,3, 4,5,6,7,8). This differs from models which advocate formation of most of the continental crust early in Earth history, with episodic thermal and tectonic reworking in areas identified as today's orogenic belts (9,10,11). Common to both models is the episodic nature of mass and/or heat transfer from the mantle, a characteristic of Earth history which is generally ignored or, if addressed, is related in nebulous terms to irregularities in mantle convection. The geological record appears to require a nonsteady state transfer of heat and mass; thus by implication it requires a nonsteady state convective flow pattern in the mantle.

Although it is generally acknowledged that the continental crust and some fraction of the mantle are complementary products of crystal-liquid fractionation operating during the differentiation of the Earth, models which explore the mechanism and consequences of phase separation in a gravity field have been essentially nonexistent. For terrestrial planets which have had a history of early heating, melting, and cooling, models of heat and mass transfer operating during the construction of new crust have focussed almost exclusively on the geometry and scale of mantle convection (12,13,14,15,16,4). Whether convection operates by mantle-wide cells or in two layers separated by an interface at 650 kilometers today provides a rigorous constraint on the thermal and mass evolution of the crust; however, the actual mechanism relating these internal process to surficial manifestation has remained obscur.

If the very existence of a continental crust (or lack of one) is primarily a consequence of a density contrast between separating silicate crystals and liquid in a planet which has been melted, we may anticipate that the mechanism of crustal evolution be understood in terms of these density relations. Fig. 1 illustrates the general form of these relations which do not vary significantly because of uncertainties in the T-P solidus location as well as the composition of magma buffered by mantle peridotite (16). Note that the largest density contrast between common mantle minerals and magma is located at the surface and 200 kilobars pressure (ie., 650 kilometers on Earth), and a minimal contrast occurs in the lowermost parts of the upper mantle.

A number of inferences can now be drawn. As long as the rate of supersolidus convection early in Earth history was large, convection rates may have been substantially greater than phase separation rates, and no continental crust would have been formed. Meteorite impact may have helped to keep the early Earth relatively homogeneous, but it may not have been the dominant process which retarded construction of the first continental crust until about 3800 Myr ago. The oldest rocks on Earth may have thus heralded the onset of phase separation of minerals and magma for the first time.

It is likely that the large mineral-magma density contrast in the uppermost part of the lower mantle would have resulted in strong fractionation and the evolution of decidedly nonchondritic compositions. However, the lower part of the upper mantle may not have undergone any significant fractionation because the densities of olivine and magma as well as pyroxenes would have been essentially the same. Thus the lower parts of the upper mantle may have

maintained a bulk Earth chemical composition (ie., minus some core forming elements) and later become a reasonable source for continental flood basalts with chondritic Nd and Sr isotopic ratios (16). The upper and lower mantles near the 650 kilometer interface would have evolved to distinct and unique chemical systems. This may have triggered a change from mantle-wide convection cells to those confined to two layers separated by today's 650 kilometer discontinuity about 3800 Myr ago. Clearly, the mineral-magma density relations provide a realistic mechanism for imposing strong and discontinuous chemical changes in the mantle at about 650 kilometers. Heterogeneous accretion or Fe variations with depth may not be required for sustaining a two layer convection system in the mantle. In the mechanism proposed here and in Herzberg (16), the geometry and scale of mantle convection are determined by these mineral-magma density relations.

Another consequence of two layer mantle convection is the possibility of catastrophic melting of the lower mantle because of the insulating effect of the upper-lower mantle boundary layer (17). This may have been followed by "destruction of the upper mantle convection pattern and enhanced surface tectonic activity" (17). Thus, the reality of two layer convection in the mantle may be fundamentally incompatible with steady state cooling of the Earth, and may possibly account for episodic crustal growth and/or tectonism with time. If these inferences are correct, nonsteady state thermal models of a cooling Earth would have to be able to quantify these episodic thermal pulses coinciding with the major crust forming and/or thermal events suggested by Moorbath (3) to cluster at 3800-3500, 2900-2600, 1900-1600, 1200-900, and 600-0 Myr.

REFERENCES (1) Hurley P.M. & Rand J.R. (1969) Science 164, 1229-1242. (2) Moorbath S. et al. (1973) Nature 245, 138-139. (3) Moorbath S. (1977) Chemical Geology 20, 151-187. (4) O'Nions R.K. et al. (1979) JGR 84, 6091-6101. (5) McCulloch M.T. & Wasserburg G.J. (1978) Science 200, 1003-1011. (6) Taylor S.R. & McLennan S.M. (1981) Phil. Trans. R. Soc. Lond. A301, 381-399. (7) Veiser J. & Jansen S.L. (1979) J. Geol. 87, 341-370. (8) Engel A.E.J. et al. (1974) Geol. Soc. Amer. Bull. 85, 843-858. (9) Hargraves R.B. (1976) Science 194, 363-371. (10) Bridgwater D. & Fyfe W.S. (1974) Geoscience Canada 1/3, 7-11. (11) Armstrong R.L. (1981) Phil. Trans. R. Soc. Lond. A301, 443-472. (12) Richter F.M. & Johnson C.E. (1974) JGR 79, 1635-1639. (13) DePaolo D.J. (1981) EOS 64, 137-140. (14) Schubert G. (1979) Ann. Rev. Earth Planet Sci. 7, 289-342. (15) Davies G.F. (1977) Geophys. J.R. Astron. Soc. 49, 459-486. (16) Herzberg C.T. (1982) JGR, submitted. (17) Schubert G. & Spohn T. (1981) Geophys. Res. Lett. 8, 951-954.

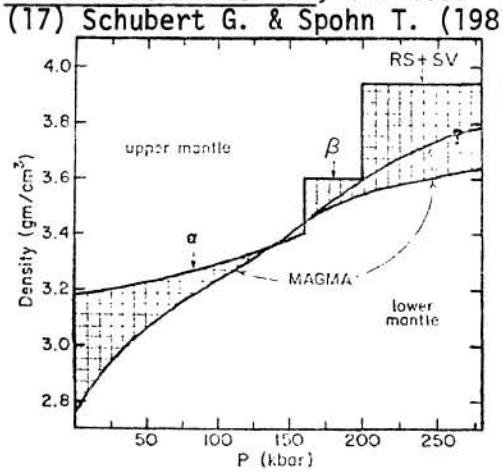

Fig. 1 Densities of α olivine (Fo_{90}), β modified spinel polymorph, RS + SV rocksalt + stishovite, and coexisting magma along the mantle solidus from (16).