A MODEL FOR CRUSTAL SUBDUCTION BY LARGE IMPACTS; R. W. Wichman and P. H. Schultz, Dept of Geological Sciences, Brown University, Providence, R.I. 02912.

Introduction: In the standard model of crater excavation, ballistically ejected material represents only about half the volume of the transient cavity (1). The other half corresponds to downwardly displaced, shock-compressed material driven beneath the transient cavity (e.g., 1, 2, 3, 4, 5). In large craters, the final crater then forms by the collapse of this transient cavity with uplift and inward flow of the shocked, displaced material comprising the crater floor (6). Although this idealized model of crater formation fits most well-preserved planetary impact structures, it implicitly assumes an elastic halfspace beneath the target surface which may be inappropriate for modeling the largest early basin-forming impacts. For these very large impacts, the depth of the transient cavity may excede the lithospheric thickness and, at least on Mars, such cavities apparently can interact with underlying viscous mantle regimes during basin formation (7). The extreme size of these cratering events also challenges some of the assumptions concerning cavity growth and collapse extrapolated from smaller structures (8). In this abstract, we propose that viscous deformation beneath very large impacts can allow emplacement of vertically displaced crustal material in the mantle, and we speculate on the implications such "impact subduction" might have for subsequent mantle evolution.

Subduction Model: Although material flow fields result in ballistic trajectories for most near-surface regions of the transient cavity, a full target section is preserved under the center of the impact. In the case of an elastic half space, compression of this section against undeformed rocks at greater depth enhances lateral flow, disrupting the column and spreading it across the base of the transient cavity. If viscous flow occurs beneath the impact, however, downward displacement of this crustal section is partly accomodated by lateral flow in the mantle below. This process transfers impact deformation from the lithospheric section to the mantle and results in the emplacement of shocked crustal material into rocks of the upper mantle or asthenosphere. Although later dynamic rebound might limit the depths such material could reach, rebound uplifts the region beneath the crater as a whole (not the displaced elements of the transient cavity alone) and the sub-impact crustal section initially should stay in the mantle.

Viscous deformation of the mantle during impact is thus a necesary condition for the emplacement of crustal sections at depth. The probability of such deformation can be evaluated by comparing the duration of the impact event to the Maxwell time (T_m) of the mantle: $T_m = \tau/2\mu\dot{\epsilon}$ where τ is applied shear stress, μ is shear modulus (~ 10^6 MPa) and $\dot{\epsilon}$ is strain rate of deformation. The Maxwell time is defined as the time required for viscous creep under stress to equal elastic strain (9); consequently, viscous behavior occurs when deformation times are greater than T_m. Deformation is essentially elastic for timescales less than T_m (9). If an impact generates shear stresses of over 10 kilobars (10^3 MPa), strain rates in an olivine mantle range from 10^{-2} to 10^{-4} /s for mantle temperatures of 800–1000°C (10). These values translate to Maxwell times on the order of 1–100 seconds. Because mantle flow requires that the duration of impact exceed T_m, only large, low-velocity (5–6 km/s) impactors (which have impactor penetration times of several tens of seconds (8)) are likely to induce such a viscous mantle response.

The extent of viscous deformation beneath an impact depends on the impact angle and the duration of the impact relative to T_m. If we define d as the depth of the transient cavity (roughly the penetration depth of the projectile into the target), for near-vertical (>60°) impacts, d can be approximated by the projectile diameter (D_P) but d decreases significantly as the impact angle is then reduced to 5° (11). While rare, near-vertical impacts are not improbable and are the most likely to emplace material at depth in the mantle. For a near-vertical impact with a duration equal to T_m, therefore, we expect viscous mantle deformation comparable to the size of the impact cavity extending to depths of ~D_P beneath the base of the transient cavity. Since the thickness of the down-driven core is of the same scale as the mantle deformation, the crustal section remains near the base of the transient cavity and is likely to be embedded in the basin floor after dynamic rebound. If the duration of the impact is significantly greater than T_m, however, lithospheric material can penetrate the mantle to depths of several D_P. We propose that mantle flow will engulf this displaced crustal section outright with depths of crustal burial in the range of ~ $0.5D_P$ to $2D_P$. This burial of crustal material in the mantle is reminiscent of terrestrial plate subduction and, for a projectile 200 km in diameter, such "subduction" could bury crust to depths of between 100 and 400 km.

The requirement of large, low-velocity impacts for this subduction mechanism limits the extent of this process in planetary history. First, impacts of sufficient size are restricted to the period of basin-forming impacts before ~3.7 Ga. The low impact velocity, however, is a stronger constraint on the occurrence of impact subduction, since only a few planetary impactor populations permit impacts at 5–6 km/s. The planet most likely to have experienced such collisions is Mars, where impact velocities range down to ~5 km/s for co-orbiting, hellocentric objects (12). The coincidence in age of these basins with a time of predicted high mantle temperatures (13) also favors subduction by this mechanism. On the Earth and Venus, impact velocities range from 15–40 km/s and 17–44 km/s, respectively (8,14) and impact subduction is much less likely. Although viscous deformation beneath the impact is still possible at velocities of 15 km/s (for mantle temperatures on the order of

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1300–1400°), the increased extent of of vaporization and melting beneath the impact may preclude preservation of a lithospheric section in the mantle during cavity collapse. Some contamination of the mantle by projectile or crustal components might still be expected, however. On Mercury, the predicted impact velocities for comets and earth–crossing asteroids (~34–44 km/s (8)) probably preclude impact subduction. Nevertheless, impactors from the postulated population of Vulcan asteroids (15) should have had much lower velocities, so impact subduction is also possible for Mercury. The probability of subduction on the Moon (impact velocities down to ~6 km/s (12)), depends more strongly on the state of the lunar interior. Thinner lithospheres associated with a magma ocean might permit early subduction events, but the growth of the lithosphere over time would inhibit mantle flow and prevent later subduction.

Implications: Unlike the more continuous subduction of oceanic plates observed on Earth, impact subduction is randomly located and episodic on a global scale. Hence only random and isolated regions of a planetary mantle can be modified by this process. Long–term effects of oceanic subduction such as repeated passage of melts into island arcs or convection beneath back arc basins are thus unlikely to occur for impact subduction. In further contrast, mantle cooling associated with subduction of cold lithospheric plates should not occur with impact subduction due to both shock heating and shear deformation in the subducted section. To first order, this section is presumed to be in thermal equilibrium with normal mantle temperatures at the time of emplacement. Although such a view is over simplistic, the mantle evolution sequence presented below can be regarded as indicative of the relative time scales required to achieve various mantle states.

Impact subduction can potentially influence mantle evolution in two ways. Injection of crustal radiogenic elements into the mantle could affect the long–term thermal history, whereas the introduction of crustal volatiles could affect the melting sequence. In the first case, a crustal block would begin to melt in ~4–6 Ma for subduction to 10 kb pressure with an initial post–impact temperature of 1000°C. Total equilibrium melting then would occur in 9–11 Ma. Significant mantle melting is unlikely to result in this time, since ~16 Ma are needed to achieve lherzolite melting temperatures outside a subducted crustal block. The melting of embedded crustal material, however, will influence the long term thermal evolution of the surrounding mantle. Since a fraction of melt is trapped along grain boundaries during porous flow, we can approximate the mantle composition after crustal melting by mixing a disseminated crustal component into the mantle. For mantle–crust ratios of 100:1, such mixing can double the abundance of heat producing elements in the mantle and these added heat sources eventually can induce mantle partial melts some 100–500 Ma after the subduction event.

The subduction of volatile concentrations could produce mantle melts on much shorter time scales. Although water is unlikely to be a major constituent of the crustal section as a whole, water or ice may be concentrated in near surface regions. For a volatile–rich regolith 500 m thick with 25% porosity, subduction under a 100–km radius projectile can subduct over 1000 km^3 of water. If the projectile caps the subducted section and drives it into the mantle, this volatile phase may not escape into the transient cavity and will be trapped instead near the top of the crustal section. Addition of such a vapor phase to surrounding mantle compositions significantly reduces the solidus temperatures and, at 1000°C and 10 kb, can initiate immediate mantle melting. If carbonates are present in the martian regolith, the associated fluid–rich and volatile phases could possibly acheive a kimberlitic character.

Conclusions: Large, low velocity impacts may inject significant crustal sections into a planetary mantle, but this process will be most efficient if the mantle yields viscously around impact–driven subsidence. Such behavior is most likely before 3.7 Ga on Mars, but also may have occurred on Mercury or the early Moon. The depth of subduction is dependent on the relative scale of impact and mantle flow regimes, but can achieve depths of over 100–200 km for projectiles over 100 km in radius. The effects of such subduction on mantle evolution are unlike those observed in terrestrial subduction zones and primarily reflect the effects of subducted volatile and radiogenic isotope concentrations. Escape of vapor into the mantle should produce kimberlite–like mantle melts soon after impact. Crustal melts develop some 5–10 Ma after impact and enrich higher mantle regions in radiogenic isotopes. Finally, isotopic heating of this enriched mantle may lead to renewed mantle melting several hundred million years after the original impact event. Such a mantle melt sequence may fit the general sequence of highland volcanism observed on Mars where explosive, patera volcanism evidently preceded formation of most of the basaltic shields and ridged plains (16).

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