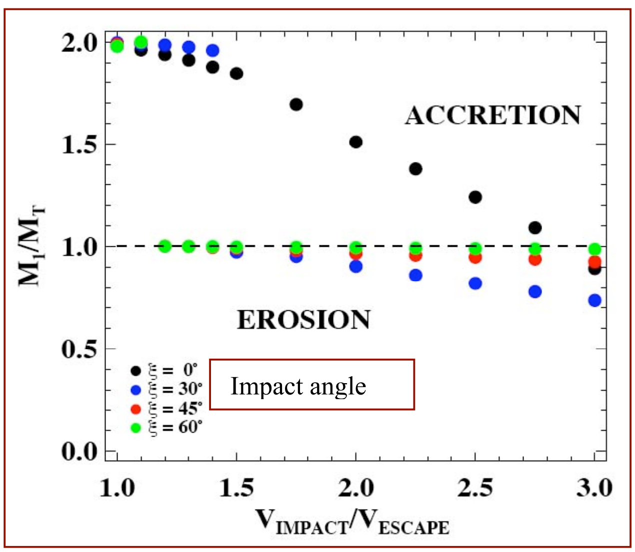
TIDAL FORCES AS DRIVERS OF COLLISIONAL EVOLUTION. E. Asphaug, C. Agnor and Q. Williams, Center for Origin, Dynamics & Evolution of Planets, Earth Sciences Dept. University of California, Santa Cruz CA 95060, asphaug@es.ucsc.edu

Impacts, Shocks & Tides: Planetary collisions are usually understood as shock-related phenomena, analogous to impact cratering. But at large scales, where the impact timescale is comparable to the gravitational timescale, collisions can be dominated by gravitational torques and disruptive tides.

Shock physics fares poorly, in many respects, in explaining asteroid and meteorite genesis. Melts, melt residues, welded agglomerates and hydrous and gas-rich phases among meteorites lead to an array of diverse puzzles [e.g. 1,2,3] whose solution might be explained, in part, by the thermomechanics of tidal unloading. Comet Shoemaker-Levy 9 disrupted in a process that is common in the present and ancestral solar system [e.g. 4], so here we consider specific effects tidal disruption had on the evolution of asteroids, comets and meteorites – the unaccreted residues of planet formation.

The Impactor. Having recently demonstrated that planetary collisions do not generally result in accretion [5; see Figure 1 below] we now show that disruptive tides and gravitational torques frequently do more work in a collision than impact shock. We focus upon the smaller of the encountering bodies – the "impactor" – since tidal stress in a two-body encounter, normalized to self-gravitation, scales with the 5th power of size. The smaller body suffers the most seriously (see Figure 2 next page), yet survives being accreted in most collisions.



Mass of largest remnant normalized to initial largest mass, versus impact velocity normalized to escape velocity, for impactors half the mass of Mars-sized 30 wt% Fe targets.

Warm Planets. Previous studies indicated that disruption of viscous planets does not occur [6]. These used smooth particle hydrodynamics (SPH) with an artificial damping coefficient, as rheological viscosity is not easily implemented in explicit schemes with short timesteps. We take a geological approach to the question, and conclude that an inviscid numerical approach is accurate. Tidal disruption requires deformational strain $\varepsilon_{def} \approx 10$ accruing over a few times τ_{grav}~$(G\rho)^{-1/2}$. The maximum viscosity $\eta_{\lim}$ allowing this deformation is approximately the stress that must be unloaded, $\sigma \approx G\rho^2 a^2$, divided by the required strain rate, $\dot{\varepsilon} \approx \varepsilon_{def}/\tau_{grav}$, where a is the radius of the disrupted object. This gives the result

$$\eta_{\lim} \approx \varepsilon_{def}^{-1} \sqrt{G} \rho^{3/2} a^2 \qquad (1)$$

Strains ε_{def}>10 can only occur if $\eta < \eta_{\lim}$~$10^{12} a_{km}2$ poise (g cm^{-1}s^{-1}), where $a_{km} = a/(1000\text{ km})$.

Young planetary embryos are vulnerable to viscous disruptive deformation because of temperature and stress dependence. Accretional heating and short half-life radionuclide decay lead to high temperature T, and viscosity decreases with $e^{-1/T}$. For power-law fluids, effective viscosity η_{eff} decreases with about the square of the applied stress (e.g. n~3 for cold ice and dry quartzite), and dissolved H_2O further lowers the viscosity. Melting and water dissolution during unloading (see below) further lowers the viscosity. Standard models of early-Earth convection adopt mid-mantle viscosities ~10^9 poise [7], and η~10^9-10^{13} poise is adopted [8] for Io's present asthenosphere.

Cold Planets: Once embryos or proto-asteroids cool below a threshold temperature so that τ_{grav} is shorter than the Maxwell time η_{eff}/σ, large gravitational deformation requires brittle fragmentation. Tidal disruption happens to cold, rocky monoliths larger than about 200 km grazing Earth-sized planets [9]. More generally, in any gravity-regime disruption event (whether caused by impacts or tides) a volume of mantle rock finds itself unloaded from $\sigma \approx G\rho^2 a^2$ to much lower pressure in the course of τ_{grav}. Release strain rates are relatively high (10^{-3} s^{-1} for a Vesta-sized planet unloading over τ_{grav}), so one can probably apply a dynamic fragmentation model [10] to evaluate the expected fragment size. Consider a Weibull distribution of active flaws $n(\varepsilon) = k\varepsilon^m$, where

$\varepsilon=\sigma/E$, σ is the far-field flaw activation stress, and E the elastic modulus. If a characteristic mantle stress $\sigma \approx G\rho^2 a^2$ unloads uniformly over τ_{grav}, then the unloading strain rate is

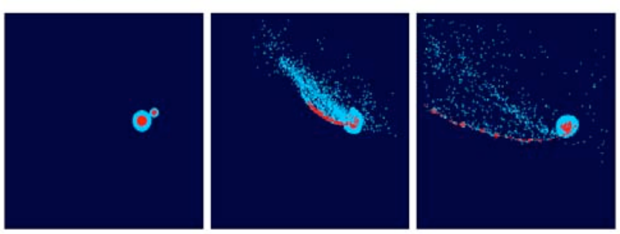
$$\dot{\varepsilon} = G^{3/2}\rho^{5/2}a^2/E \quad (1)$$

which gives $L \approx 6c_g \alpha^{-1/m+3} \dot{\varepsilon}^{-m/m+3}/(m+2)$ for the mean expected fragment size [10], where c_g is crack growth velocity, $\alpha = 8\pi c_g^3 k/[(m+1)(m+2)(m+3)]$, and m >~6 for most geologic materials.

By this analysis a 500 km diameter cold basalt sphere (k=4·10^{29}cm^{-3}, m=9) cracks into ~200 m fragments if unloaded, while a 1000 km sphere cracks into ~70 m fragments. No large monoliths survive, indicating instant rubble piles if the fragments do not disperse, and families of sub-km asteroids otherwise – a possible solution to the "missing mantle" paradox [see 3] for disrupted asteroids.

Gravitational Unloading: For large accreting embryos, the process of gravitational unloading is thermodynamically interesting, and is of relevance to meteorite petrogenesis and also for the mechanical evolution of an impact system. (In Moon formation, for example, an "extra kick" is sometimes invoked [11] to bring protolunar material into high orbit; unloading from hydrostatic pressure may do significant work.) . Consider a parcel of deep mantle on an unloading trajectory. The energy per unit mass of decompression is $\int dP/\rho$, which for constant density (e.g. up to the onset of vaporization) is $\sim P/\rho \approx G\rho r^2$ ~ 2·10^{10} erg g^{-1} for the base of a Mars-sized planet's mantle. The effect of this energetic release is dependent upon the equation of state. The fragments of SL9 were very active, probably owing to sudden exposure (dP) of a pristine interior. A disrupted planetary embryo unloads from pressures millions of times greater, and if rich in mantle volatiles might erupt [see e.g. 12]. The energetics of tidal disruption can easily rival the energetics of shock release.

$M_I : M_T$=1:10, $\left(\frac{V_{IMPACT}}{V_{ESCAPE}}\right) = 2.0$, $\xi = 30°$



Unaccreted moon-sized body M_I impacting and disrupting tidally near a Mars-sized body M_T at $2v_{esc}$. Red = iron, blue = rock. Target mass and composition is essentially unchanged by this collision. In contrast, the impactor is sheared out, its central pressure dropping to near-zero in the course of an hour. The disrupted impactor coalesces into three objects of mass M_2 = 0:16M_I , M_3 = 0:078M_I and M_4 =0:04M_I. The smallest of these two falls back to the target and the remaining two escape the collision all together. Each of these objects consists of >50% iron by mass and is rotating with periods <5.2h.

Fractionation: Even in grazing (non-impacting) collisions, we find that a Moon-sized differentiated embryo loses much of its outer mantle, and therefore most probably most of its crust and atmosphere, when encountering a Mars-sized embryo. Impacting non-accretionary encounters (e.g. Figure 2 below) have more severe effects. The lost outer materials from the smaller embryo are observed in simulations to drain down onto the larger embryo, increasing the volatile and atmophile inventory of growing bodies at the expense of the unaccreted ones. As tidal collisions are, in most velocity regimes, more common than accretionary collisions, embryo iron and volatile inventory should evolve so that unaccreted bodies become iron enriched and volatile depleted. In severe instances the interloper becomes a family of disrupted bodies fractionated from the same reservoir.

Degassing: As a disrupted embryo unloads to low pressure, the deep mantle crosses the water solubility pressure. An important question is whether there is adequate time for gas to segregate. For severe encounters such as Fig. 2, much of the impacting planet's dissolved gas will be efficiently pumped out of the disrupted interior, flushing mantle and crust materials in a hydrothermal event. Pressure release melting and effective mixing also appears likely. Even for non-impacting (grazing) collisions, we find permanent pressure unloading, to half of the original pressure, throughout the interior. Detailed petrologic effects of tidal unloading can be explored experimentally, using a pressure cell programmed to mimic the pressure release curves derived in the simulations.

References: [1] McCoy, T.J. et al. (1997). *Geochim. Cosmochim. Acta* 61, 623-637. [2] Keil, K. et al. *MAPS* 32, 349-363 (1997). [3] Burbine, T. H., Meibom, A. & Binzel, R. P. (1996). *MAPS* 31, 607-620. [4] Asphaug, E. and Benz, W. *Icarus* 1996 [5] Agnor, C. & Asphaug, E., *Ap. J.* 613, L157-L160 (2004). [6] Mizuno, H. & Boss, A. P. (1985). *Icarus* 63, 109-133. [7] Walzer, U. et al.; *Tectonophysics* 384, 55-90. [8] Tackley et al, *Icarus* 149, 79-93 (2001). [9] Jeffreys, H. (1947). *MNRAS* 107, 260-272 [10] Grady, D. E. & Kipp, M. E. (1980). *Int. J. Rock Mech. Min. Sci. Geomech*. 17, 147-157. [10] Melosh, H.J. & Sonnett, C.P. (1984). In Origin of the Moon, Houston (LPI), 621-624. [11] Wilson et al., *MAPS* 34, 541-557 (1999)

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