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Ejecta-megaregolith accumulation on planetesimals and large asteroids

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Abstract — Megaregolith accumulation can have important thermal consequences for bodies that lose heat by conduction, as vacuous porosity of the kind observed in the lunar megaregolith lowers conductivity by a factor of 10. I have modeled global average ejecta accumulation as a function of the largest impact crater size, with no explicit modeling of time. In conjunction with an assumed cratering size-distribution exponent $b$ (cf. present-day asteroids), the largest crater implicitly constrains the sizes of all other craters that make significant contributions to a final megaregolith. For any given largest-impactor mass ratio of order 0.001 (at constant $b$ and impact velocity), globally averaged ejecta accumulation thickness is relatively constant over a wide range of $d$; e.g., for that specific mass ratio and $b = 2$, results range from 1.0 to 1.3 km for all $d$ between 50 and 800 km. The largest-impactor mass ratio is more likely some consistently major fraction of the catastrophic-disruption mass ratio, which (as a result of increasing gravity effects at larger scales) is a complex function of body size, with an anomalous high at $d \approx 100$ km, but in general implies the largest crater’s diameter is close to the $d$ of the target body. Total ejecta accumulation is then roughly proportional to $d$, and with conservative parameter assumptions (e.g., $b \approx 2$) will be 1-5% of the body’s radius. Global accumulations estimated by this approach are higher than in the classic Housen et al. (1979) study by a factor of roughly 10. This revision is caused mainly by higher (typical) largest crater size, caused in turn by higher estimated catastrophic-disruption mass ratio. For $b \approx 2$, the single largest crater will typically contribute close to 50% of the total of new (non-recycled) ejecta. For modeling of thermal implications, significant stochastic variations probably arise from two effects: concentration of ejecta mass into a relative few large fragments (although if formed relatively early these may be largely eroded by later cratering); and stochastically uneven distribution, especially on relatively small bodies. Megaregolith can be destroyed by sintering. The pressure sensitivity of the sintering process makes it effective at generally far lower temperature on larger ($>> 100$ km) bodies. Planetesimals $\sim 100$ km in $d$ may be surprisingly well-suited (about as well-suited as bodies 2-3 times larger, assuming equal heat production) for attaining temperatures conducive to widespread melting. A water-rich composition may be a significant disadvantage in terms of planetesimal heating, as the shallow interior may be densified by aqueous metamorphism, and will have a low sintering $T$. But development of a megaregolith thick and porous enough to have significant thermal evolution consequences is practically inevitable.
1. INTRODUCTION

The larger planetesimals, asteroids and rocky natural satellites are/were invariably blanketed by an accumulation of impact-crater ejecta (megaregolith). The extent of that accumulation is of interest for various reasons. The early bodies underwent rapid thermal evolutions that determined the course of their metamorphic/igneous modification, and ultimately influenced the origin and evolution of the planets. The thermal evolution was determined by a competition: heat generation, probably mainly by $^{26}$Al ($t_{1/2} = 0.72$ Ma), versus heat loss, which in small bodies occurs mainly by conduction. Megaregolith-ejecta accumulation can have important thermal consequences for such a body. The rate of loss is a function of the thermal conductivity $k$ of the outer layers. Vacuous porosity of the kind observed in the lunar megaregolith lowers conductivity by a factor of ~10 (Warren and Rasmussen, 1987). Recent thermal models make a range of extreme and, in terms of evolutionary implications, divergent assumptions about $k$. At one extreme are models (e.g., Ghosh and McSween, 1998; Merk et al., 2002; Wilson et al., 2008) that simply assume a solid-rocklike $k$. At another extreme, some (Hevey and Sanders, 2006; Sahijpal et al., 2007) assume a $k$ similar to that of lunar surface fines (i.e., lower by a factor of ~2000 compared to the solid-rock $k$; Langseth et al., 1976) prevails until, with rising $T$, sintering suddenly transforms the material to rocklike $k$. The megaregolith, for purposes of this work, is defined as the body’s layer of accumulated impact-crater ejecta, exclusive of material that may have become extensively modified (sintered) to a low porosity. This paper is concerned with megaregolith development on bodies of diameter $d_B$ between ~100 km and Moon-sized (3476 km); i.e., mass between $10^{18}$ and $10^{23}$ kg. Megaregolith is also important as the outer shell of porous, weak material into/through which later impacts transpire; as the usual context for spall-off of chunks that may become meteorites; and as the context for remote sensing observations.

The blast-out and accumulation cycle that produces megaregolith leads to an increase in porosity, by a factor estimated (e.g., Melosh, 1989; Richardson, 2009) to be ~ 20%. In the context of an atmosphereless planetesimal, this porosity probably tends to be vacuous, like the porosity of the Moon’s regolith and megaregolith. Vacuous porosity leads to a marked diminution of thermal conductivity $k$ (Wechsler et al., 1972; Horai and Winkler, 1980). Yomogida and Matsui (1984) found analogous effects with mildly porous chondritic materials. Models that assume gas pressure is not <<0.1 MPa within the pores (e.g., a model used in planetesimal modeling by Ciesla et al., 2009) vastly overestimate the $k$ of vacuous-porous materials. The lunar data (summarized in Fig. 1) show that vacuous porosity yields reduction in $k$ along a single exponential relationship, whether the material is a cohesive breccia or a loose soil. Even the igneous (non-fracture) porosity in unbrecciated mare basalt appears to have almost the same effect (regarding sample 70017, see the figure’s caption).

A low-$k$ planetary layer may be analogized to a resistor in an electrical circuit. The thermal
resistance of each layer is proportional to \((1/k_{i-o})(1/r_i-1/r_o)\), where \(r_o\) and \(r_i\) are the fractional radii of the outer and inner boundaries of the layer (e.g., Sucec, 1975). The resistances are additive. As a crude illustration of the potential importance of the insulating layer, the steady-state, internal heat generation neglected, heat flow \(q\) out of the body shown in Fig. 2 would be

\[
q = \frac{4\pi(T_1 - T_3)}{(1/k_{1-2})(1/r_1 - 1/r_2) + (1/k_{2-3})(1/r_2 - 1/r_3)}
\]  \(\text{(1)}\)

Results from (1) for the relative \(q\) implied by various assumptions regarding the megaregolith’s thickness and conductivity are shown in Fig. 2. For example, suppose a body of order 100-1000 km in diameter \(d_B\) has a 2-km megaregolith, with \(k = 0.1\)× solid rock, atop a 6-km conductive solid-rock layer, below which the body is approximately isothermal due to the rapidity of primordial heating in relation to thermal diffusivity. A planetesimal that undergoes rapid, uniform heating will develop an approximately isothermal deep interior (inward of \(r_1\) in Fig. 2), as the influence of radiative heat loss from the surface penetrates only to a “skin” depth that according to Wilson et al. (2008) is \(~ 8\) km (as will be discussed below, this seems an underestimation, unless the nominally “solid” interior material has a thermal diffusivity far lower than stipulated by Wilson et al.). Figure 2 indicates that a 2-km megaregolith layer, with \(k = 0.1\)× the deep (“solid”) interior \(k\), will reduce heat loss from the deeper interior by a factor of 3 in comparison to the rate with the higher \(k\) throughout. The same approximate total resistance would result if just 0.02 km of powdery regolith with \(k = 0.001\)× solid rock were separated from the near-isothermal core by a 7.98-km solid rock layer.

For the best known megaregolith-covered body, the Moon, based on an earlier version of Fig. 1, coupled with a compilation of porosities in lunar breccias (average: 17±10%), Warren and Rasmussen (1987) estimated that the \(k_{MR}\) of the megaregolith is \(~ 0.1\) times that of solid rock. In general, porosity is expected to be higher on smaller bodies, with their lower gravity and internal pressure, and lower potential for igneous activity and sintering after the main, late-accretionary era of impact cratering. The near-surface hydrostatic pressure-depth gradient \(dP/dz\) is directly proportional to \(d_B\), and the \(P\) at any given \(r/r_B\) scales as \(d_B^2\) (Fig. 3; density variations may alter this nominal \(P\) distribution, but only to a mild extent). On a small (say \(d_B < 100\) km) body the average megaregolith porosity could conceivably be as high as 40%, implying a reduction in \(k\) by a factor of \(10^2\) relative to the solid-rock \(k\) (Fig. 1).

Housen et al. (1979) authored the classic study of the development of asteroidal ejecta accumulations (cf. Housen and Wilkening, 1982). However, their approach focused on the issue of crater saturation for a “typical” surface region “exterior to sparsely scattered, large anomalous craters”; and on that region’s evolution, and in particular its elevation evolution, as a function of time (cf. Ward, 2002). An approach focused on time and saturation can be useful for application to powdery regolith (sensu stricto) on asteroids, especially asteroids that may have acquired fresh-rocky surfaces at some relatively recent date. But for the more basic purpose of constraining global impact
ejecta accumulation thickness, timing is an ancillary issue, and introduces unnecessary complication. It is precisely the few “large anomalous” craters that preponderate in contributions to a global megaregolith. Moreover, in some important respects, such as depth and volume of the excavation/ejection zone, the Housen et al. (1979) model has been superseded by modern cratering physics interpretation. My model builds from the simple premise that impactors, and the craters they produce, conform (approximately) to a power law size-frequency distribution; which implies that an estimate for the magnitude of the single largest crater implicitly constrains the sizes of all other craters large enough to be significant contributors to the final megaregolith. The key issue of the (typical or average) size of the largest crater is admittedly difficult to constrain. For now, suffice to note that 4 Vesta’s largest crater is a 460-km basin whose transient crater probably had a diameter of ~310 km, or 0.58 times the diameter of Vesta itself (Thomas et al., 1997; Asphaug, 1997).

2. MEGAREGOLITH ACCUMULATION MODEL

Volume and depth provenance of ejecta from an individual crater

Craters are complex and diverse. The approach of using any single equation to characterize the relationship between crater size ($D_r$) and the globally averaged thickness $z_1$ of the crater’s ejecta is justifiable, however, because our real aim is to evaluate the aggregate ejecta thickness $z_A$ from a multitude of craters on a multitude of bodies. Housen et al. (1979) modeled the excavation/ejection zone of a nascent impact crater as a spherical cap with depth/diameter ratio of 0.2. As discussed by, e.g., Melosh (1989, pp. 78-80, 119), pi-group scaling and observational constraints imply that the ejection/excavation zone may be better modeled as parabolic and with depth/diameter ratio ~1/10 to 1/8. Haskin et al. (2003) thus modeled the volume of ejecta from a crater as

$$V_1 = 0.09\pi R_t^3$$

(1a)

where $R_t$ is the radius of the ejection zone, which coincides with the radius of the transient crater. The factor of 0.09 (rather than the depth/diameter ratio of 0.10) is meant to compensate for a zone near the center-bottom of the transient crater that is pushed down (or vaporized) but not ejected (cf. Melosh’s Fig. 5.13; or Figs. 2 and 3 of Wada et al., 2004). Compared to the spherical cap (and 2× deeper) model of Housen et al. (1979), the $V_1$ implied by (1a) is 0.43 times less (or even smaller, considering that (1a) uses the transient crater radius, whereas Housen et al., 1979, drew little distinction between transient and final crater radius).

Maxwell’s (1977) analytical model of excavation flow suggests that the ejection zone is deeper toward its rim than in the paraboloid model. This model assumes that flow velocity falls as an inverse power of radial distance $r^{-Z}$ from the explosive center. The $Z$ model is not perfect; a precise fit to the excavation flow probably requires assuming that $Z$ is not a constant but a variable function of time (Anderson et al., 2003; cf. Yamamoto et al., 2009). Still, Maxwell’s (1977) model assuming $Z \sim 3$ gives a good fit to the excavation flow and the shape of the ejection volume, as constrained by a variety of observations (Melosh, 1989; Wada et al., 2004). Croft (1980) showed that the volume for
the ejection zone in this model is
\[ V_1 = \frac{2}{3} \pi R_t^2 \left( 1 - \frac{3}{Z+1} \right) \]  
(1b)
A depth/diameter ratio of 0.10, matching (1a), requires \( Z = 2.734 \). The factor \( 1 - \frac{3}{Z+1} \) then becomes 0.197, and with \( R_t \) set at unity (1b) implies an ejection volume 1.45 times greater than (1a).

A model for pi2group scaling developed by Holsapple (2003; cf. Housen et al., 1983, and Holsapple 1993) similarly implies a factor of 1.33 greater volume of excavation than (1a). In the absence of strong evidence for choosing among the (1a), (1b) and Holsapple models, for purposes of further discussion the volume of the ejection zone will be modeled as
\[ V_1 = 0.11 \pi R_t^3 \]  
(1c)

The contribution \( z_1 \) of this volume of ejecta to the average global thickness of megaregolith \( z_A \) in a spheroidal body may be approximated by dividing \( V_1 \) by the surface area, \( 4\pi r_B^2 \), where \( r_B \) is the radius of the body, which yields
\[ z_1 = 0.0275 \frac{R_t^3}{r_B^2} \]  
(2a)
or more conveniently
\[ z_1/r_B = 0.0275 \left( \frac{D_t}{d_B} \right)^3 \]  
(2b)

Some further complications in the modeling of \( z_1 \) will be evaluated in the Discussion section.

The relationship between a crater’s size and the depth provenance of its ejecta is an issue that transcends the narrow context of modeling megaregolith development. Planetary petrologists, especially lunar petrologists, often ponder the depth provenance of materials excavated by impact (e.g., Wilhelms, 1987; Warren, 2001b; Haskin et al., 2003). Yet the relationship between \( D \) and the statistical depth provenance of a crater’s ejecta has never, to my knowledge, been described in a detailed manner. This relationship was, implicitly, constrained long ago by Maxwell’s (1977) excavation model. For translating Maxwell’s (1977) model into the desired statistical assessment of the depth provenance of a crater’s ejecta, the first step is to calculate the depth of excavation \( \omega \) along the basal streamline, i.e., the streamline that intersects the ground level at \( r = R_t \). Croft (1980) showed that this streamline can be expressed by the following rectangular coordinates:
\[ r = R_t \sin \theta \left( 1 - \cos \theta \right)^{1/(Z-2)} \]  
(3a)
\[ \omega = R_t \cos \theta \left( 1 - \cos \theta \right)^{1/(Z-2)} \]  
(3b)
where \( \theta \) is the angle between a vertical line extending down from the point of impact and the streamline at the given \( r \) (e.g., with \( Z = 2.734 \) the maximum excavation depth \( \omega_{\text{max}} \) of 0.2\( R_t \) occurs at \( \theta = 65^\circ \) and \( r = 0.429 R_t \); at \( \theta = 90^\circ \), \( r = R_t \) and \( \omega = 0 \)). The resultant shape model can be converted into a depth-provenance spectrum by numerical integration, i.e., by counting the number of cells at a given depth when the ejection zone is modeled as a 3-dimensional grid space. In practice, instead of a fully 3-dimensional model, I employed a 2-d model with \( r^2 \) weighting of the cell volumes. Adequate resolution was achieved using 0.5\(^\circ\) increments for \( \theta \) and 0.001\( D_t \) vertical increments. Results are shown in Fig. 4. The curve shown assumes \( Z = 2.734 \) (i.e., depth/diameter of the ejection zone = 0.10). Note that as a consequence of the curve’s concavity, only 32 vol% of the ejection zone is
deeper than the depth of 0.5 (times the maximum), 11 vol% is deeper than the depth of 0.75, etc. For $Z = 3.00$ (where $\omega_{\text{max}} = 0.25R_t$) the curve becomes slightly more concave, but the difference in terms of Fig. 4 would be almost imperceptible.

**Populations of craters**

The anchor point for this model is the observed or assumed largest crater, whose transient crater diameter (of course, never precisely observable) is $D_L$. The size-frequency distribution of the rest of the population, on an average or typical body, is conventionally modeled by a power law

$$N_{\text{cum}}(D) = cD^{-b}$$

(4)

where $N_{\text{cum}}$ is the cumulative number of craters with diameter $D$ or larger, and in the ideal case of a single $b$ applying to the entire size spectrum, $c = 1/D_L^{b}$. This power law is usually applied to final crater rim diameters, but its form is equally germane to transient craters. Assuming the excavation crater shape is size-independent, a $b$ of 3 implies that the volume of excavation is size-independent, i.e., every size interval contributes an equal fraction of the total cumulative ejecta volume. A $b$ of 2 implies that the surface coverage of craters is size-independent, i.e., craters in every size interval represent the same areal fraction of the total surface, while the relative volume of cumulative cratering-excitation is proportional to $1/D$. A physically implausible $b$ of 1 would imply that all but the largest few craters contribute negligible ejecta volume. For any given $D_L$ as anchor point, the lower $b$ is, the thinner the final accumulation of ejecta will be.

As indicated by, e.g., Holsapple (2003), in the small-scale “strength” regime at any given impact velocity $v_i$ (and an impact angle not extremely far from 45°) the transient crater diameter $D_t$ will be in approximately fixed proportion to the impactor diameter $d_i$; e.g., if $v_i = 5$ km/s (the typical asteroid-asteroid encounter velocity), $D_t$ will be $\approx 10d_i$. As the influence of gravity $g$ increases for very large craters, the $D_t/d_i$ ratio tends to decrease. Still, we can constrain the $b$ of the crater-size power law (4) indirectly by constraining the exponent $\beta$ for the analogous impactor-size power law: $N_{\text{cum}}(d_i) = cd_i^{-\beta}$. In general, 2.5 is the canonical value for $\beta$ in a population that undergoes collision-fragmental selection (Dohnanyi, 1969). But the present-day asteroid population shows a complex distribution (e.g., Asphaug, 2009), probably as a result of various size-dependent, especially $g$-related, effects. Bottke et al. (2005a; cf. O’Brien and Greenberg, 2005) inferred that this population, although greatly reduced in numbers, probably has a size-frequency distribution similar in shape to the population during the late stages of major accretion. In this distribution, $\beta$ is $\sim 2.1$ overall, but 1.94 for the $d$ range of 1 to 50 km, $\sim 1.63$ for the range of 50-100 km, and it increases toward 3 for $d > 100$ km. The details of the size distribution have implications that are best evaluated after an assessment of constraints on $D_L$.

Dynamical models indicate that in general, especially during the late stage of accretion ($\sim 2$ Ma after its onset) when encounter velocities began to approach modern values even as multi-hundred-kilometer planetesimals became common (Weidenschilling and Cuzzi, 2006), which incidentally was at about the same time heat build-up from $^{26}$Al climaxed (Hevey and Sanders, 2006), planetesimals
probably had to endure impacts energetic enough to challenge their ability to survive. Beyond some impact-energy limit, ejected matter begins to escape more than it lands. Since for any given target-body size the transient crater diameter $D_t$ scales as the cube root of impact energy, the transition from growth to catastrophic disruption is quite abrupt, in terms of $D_t/d_B$. Thus, unless $b$ is much less than 2, $D_t$ is probably within a few tens of percent of the catastrophic disruption crater diameter $D_C$ (expressed in this work, like $D_L$, in terms of the transient crater diameter).

Housen et al. (1979) estimated that $D_C$ is of order 1/3 to 2/3 of $d_B$, depending on the size and mechanical strength of the target body. In contrast, Nolan et al. (2001) suggested that shock-induced fracture in advance of crater excavation flow reduces the potential for catastrophic mass loss, which results in a $D_C/d_B$ ratio of 1.3 for even a small asteroid (Gaspra, $d_B$ modeled as 12.6 km) being impacted at 5 km/s. As suggested by Bottke et al. (2005a,b), the disruption threshold can be modeled in terms of $d_C$, the diameter of the catastrophic impactor:

$$d_C/d_B = (2Q_C/v_i^2)^{1/3}$$

where $Q_C$ is the critical specific impact energy (units of erg/g). $Q_C$ is hard to constrain, especially for small bodies, which undergo strength-regime cratering. But for bodies greater than about 10 km in diameter, several different approaches (see reviews in O’Brien and Greenberg, 2005; and Asphaug, 2009) suggest a relationship not far from that shown in Fig. 5 (curve “10”) of Bottke et al. (2005b), which implies

$$Q_C = 0.318 d_B^{1.348}$$

The $d_C/d_B$ implied by (6) scales as $v_i^{-(2/3)}$. At the modern prevailing asteroid-asteroid encounter velocity of 5 km/s, $d_C/d_B = 0.024 d_B^{0.45}$ (for $d_B$ in km), and thus $d_C/d_B$ ranges from 0.07 to 0.53 for $d_B = 10$-1000 km. At a late-accretionary $v_i$ of say 2 km/s, these $d_C/d_B$ predictions shift to 0.12 and 0.99, respectively. Fig. 5 shows the $d_C/d_B$ implied by (6) translated into the critical impactor mass ratio $m_C/m_B$. Figure 5 also shows for comparison $m_C/m_B$ as implied by the “best fit” $Q_C(d_B)$ relationship of O’Brien and Greenberg (2005), which as reviewed by those authors is, for the $d_B$ range of interest, lower than most other estimates by a factor that is fairly representative (i.e., roughly 1-sigma below Bottke’s $Q_C$) of the overall scatter among such estimates in recent literature. In other words, the nominal uncertainty in $m_C/m_B$ is approximately a factor of 2.

For translating between $d_C/d_B$ and the catastrophic-destruction crater diameter ratio $D_C/d_B$ (and more generally between $d/d_B$ and $D_t/d_B$), I developed parameterizations (i.e., a series of polynomial fits) of $D_t$ results for various combinations of $d_t$ and $d_B$ using Holsapple’s (2003) implementation of pi-scaling for crater dimensions. Additional inputs were impact angle of 45 degrees; rocky physical characteristics for both impactor and target, i.e., densities of 3000 and 3200 kg m$^{-3}$, respectively; and $g$ and escape velocity calculated as a function of $d_B$ under the assumption of uniform density $\rho$ within the target; i.e., $g = (4/3)\pi G \rho_B r_B$ and $v_{esc} = (2Gm_B/r_B)^{1/2}$. Pi-scaling indicates that for impact velocity $v_i$ of 5 km/s the $D_t/d_t$ ratio is uniformly close to 10 in even the largest of craters on a $d_B << 100$ km body, but falls to, e.g., 7.4, 5.4, 3.7, 2.5 in $D_t/d_t \approx 1.0$ events for $d_B = 100, 200, 400$ and 800 km,
respectively.

Under the assumption that accretion was oligarchical (not runaway), so that the asteroids and planetesimals are/were stochastic survivors from a series of near-catastrophic collisions, Poisson statistics and the power-law size distribution, \( N_{\text{cum}}(d) = cd^\beta \), can be applied to estimate the probability of \( d_i \) being smaller than \( d_c \) by a given factor. The Poisson equation for probability of zero outcomes is simply \( p_0 = e^{-n} \), where \( n \) is the number expected from ideal sampling of the overall population. By this method (Fig. 6; with relative \( d \) translated into relative mass assuming simple \( d^3 \) proportionality), for \( \beta \sim 2 \), the most likely outcome is \( m_i/m_C \sim 0.45 \). This result varies as a function of \( \beta \); a \( m_i/m_C \) range of 0.35-0.59 is implied by varying \( \beta \) from 1.5 to 3. Fig. 7 shows the \( D_i/d_B \) ratios that result from assuming \( m_i/m_C = 0.25-0.75 \). As a rule of thumb, for bodies of the size range under consideration, \( m_i/m_C \sim 0.5 \) translates into \( D_i/d_B \sim 1.0 \).

Returning to the problem of constraining \( b \), the \( \beta(d_i) \) of the asteroidal size distribution (Bottke et al., 2005a, Fig. 1) can be translated into a \( \beta(m_i/m_C) \), by using (6) to derive \( d_c \), and thus \( m_C \), for any given target body size and impact velocity. Since \( \beta \) is a measure of slope, seemingly small bumps and dips on the size distribution become magnified, so that results (Fig. 8) for large \( m_i/m_C \) in the relevant \( d_B \) range are remarkably structured, with a peak at \( \sim 80 \) km and \( \beta \sim 2.5 \), a deep valley at \( \sim 220 \) km and \( \beta \sim 1.3 \), and then a gradual rise toward \( \sim 800 \) km and \( \beta \sim 3 \). One complication is that gravity’s effect of limiting the growth of large craters causes the ratio \( D_i/d_i \) to decrease (for any given \( v_i \)) with increasing \( d_i \), so the size-frequency exponent \( b \) for transient crater diameters is slightly greater than the corresponding \( \beta \) for impactor diameters (i.e., the distribution’s slope, for large, high-g bodies, is mildly but systematically steeper). I have not attempted to model the minor increase between \( \beta \) and the corresponding \( b \), except by taking \( m_i/m_C \sim 0.5 \), and \( D_i/d_B \sim 1.0 \) (rather than 0.45 and \( \sim 0.9 \)), as the most likely outcome implied by the Poisson-statistical approach at the end of the previous paragraph.

Another complication is that in late-accretionary times the prevailing \( v_i \) was lower. Weidenschilling and Cuzzi (2006) estimate that even after 2 Ma of accretion, typical impact velocities were still “a few tenths to \( \sim 1 \) km s\(^{-1} \).” The implied impactor diameter \( d_i \) to yield a given crater \( D_i \) scales as \( 1/v_i^{0.5} \). A lower \( v_i \) shifts the \( \beta \) spectrum’s features to smaller \( d_B \); e.g., with \( v_i = 2.5 \) km/s, the \( \beta \sim 1.3 \) valley shifts to \( \sim 160 \) km, and the two peaks shift to \( \sim 75 \) and 400 km.

In summary, \( D_i/d_B \) is unlikely to be much less than 1. For the near-largest impacts, \( b \sim 2 \) is probably conservatively low as a single value to represent the general populations of craters and target bodies considered in this work. For many of the largest craters on the largest bodies, particularly if the cratering occurred mostly during the late stages of accretion while the prevailing \( v_i \) was increasing but still much less than 5 km/s, \( b \) may have been closer to 3.

**Modeling ejecta accumulation**

For modeling purposes, the entire volume of the excavation/ejection zone is assumed to accumulate upon the surface of the target body (obviously this is not strictly correct, but it is justified...
as a simplification in the next section). The statistical crater size distribution is modeled per (4), for which the only input, other than $D_L$, is $b$. The statistical accumulated average thickness $z_A$ can be calculated by summing the individual ejecta volumes of all craters, starting from $D_L$, down to a size where the incremental increase in global ejecta layer thickness becomes insignificant. Models were constructed to include the largest 65,500 ($2^{16}$) craters on the body. This number is overkill for models assuming that $b$ is 3 or less. The only complication is that for each impact, the potential contribution of new ejecta, per (2), is reduced by the fraction $\zeta$ of the crater’s ejecta that is “recycled” from depths within the preexisting global ejecta layer. The magnitude of $\zeta$ is constrained as the fraction of the excavated volume that, per a precise 5th-order polynomial to the trend in Fig. 4 with the maximum depth of excavation $\omega_{\max}$ assumed = 0.1 $D_t$, is shallower than $z_A$. The Fig. 4 results were parameterized as

$$\zeta = -0.193\eta^5 + 0.312\eta^4 - 0.191\eta^3 - 0.608\eta^2 + 1.680\eta$$

where $\eta$ is the ratio $z_A/\omega_{\max}$. Here $z_A$ is the value immediately prior to the impact being evaluated; and both $z_A$ and $\omega_{\max}$ are treated in units of (i.e., normalized to) $r_B$, the radius of the target body. Of course, for any crater with $\omega_{\max} < z_A$, the ejecta is 100% recycled and no new growth of $z_A$ occurs.

Figure 9 shows the final results from this model: average accumulated ejecta layer thickness $z_A$ (in units of $r_B$, after all $2^{16}$ craters form) as a function of assumed $D_L/d_B$ ratio. The sequence of crater formation, although potentially marginally significant, is not crucial. The four main curves in the figure are each based on a set of 10 randomized sequences of formation for the $2^{16}$ craters. But as the two light-dashed curves indicate, even under the extreme assumption that the craters form in a sequence of size (either smallest to largest or largest to smallest), results are only marginally different from the average randomized-sequence result. The crater-formation sequence is slightly consequential for models assuming a high $b$; e.g., for $b = 3$ and $D_L/d_B$ in the range 0.4-1.0, $z_A$ could in principle vary, between the extremes of the decreasing crater size model and the increasing crater size model, over a factor of 1.4. Figure 10 shows the same model results translated, by straightforward conversion from units of $r_B$ for $z_A$ into units of kilometers, for a range of different target body diameters.

In Figure 11, the same results (Figs. 9 and 10) have been recast with the largest impactor mass ratio $m_I/m_B$ taking the place of $D_L/d_B$ for the x axis. Masses were derived from diameters assuming the same densities as employed in the modeling of $D_L/d_i$, i.e., Holsapple’s (2003) “rocky” densities. An interesting effect follows from the decrease in $D_i/d_i$ (for any given $v_i$) with increasing $d_B$. By (2b), ejecta yield ($z_i/r_B$) scales as the cube of $D_i/d_B$. Thus, the factor by which $D_i/d_i$ decreases with increasing $d_B$ (Holsapple, 2003) gets cubed in the evolution of $z_A$. The net effect is that even though the modeling implies a uniform ejecta accumulation $z_A/r_B$ ratio for any given $D_L/d_B$ (Fig. 9), in terms of absolute thickness (in kilometers) $z_A$ remains relatively constant for any given $d_i/d_B$ (or $m_I/m_B$) over a huge range in $d_B$ (Fig. 11). Readers with interest in specific target body sizes (and trusting the Bottke et al. (2005a) model for $Q_c$ plus their argument that the size-frequency distribution of
planetesimals is closely mirrored by the present-day asteroids) may want to fine-tune these results based on the $\beta$ variations shown in Fig. 8.

Assuming that the largest event involves a $m_L$ that is a large fraction of $m_C$ as calculated by the (extended) Bottke et al. (2005a,b) model (Fig. 5; for impact velocity $v_i$ of 5 km/s), we arrive at a simple plot of body diameter $d_B$ vs. expected thickness $z_A$ of accumulated ejecta (Fig. 12). Within the overall uncertainty of the modeling, the relationship is essentially linear at $z_A/d_B \sim 0.04(m_L/m_C)$ (valid for $m_L/m_C \sim 0.25$ to 0.75).

As illustrated in Fig. 13, the single biggest crater typically contributes a large fraction of the total ejecta accumulation, especially in cases of low $b$ combined with a high $D_L/d_B$ ratio. Consider, for the nominal $b$ of 2, the case of a largest crater with $D_L/d_B$ similar to the $D_t/d_B$ of Vesta’s great southern basin, $\sim 0.58$ (Vesta’s $d_B \sim 530$ km, and per Asphaug, 1997, Holsapple, 2003, etc., the $D_{rim}$ of $\sim 460$ km implies $D_t \sim 310$ km). This one crater will produce on average (depending, inter alia, on when it forms relative to other large craters) $\sim 40\%$ of the body’s total accumulation of otherwise unexcavated ejecta. However, in terms of the present surface layer about to be studied by Dawn (Russell et al., 2004), the basin’s ejecta probably so greatly churned the surface upon landing at distal locations (cf. Haskin et al., 2003) that the basin does not necessarily dominate the mix of surface debris except within $\sim 2R_t$ ($\sim 300$ km) of its rim.

Some caveats are in order. Realistically, for bodies near the smaller (100 km) end of the size range under consideration, $d_C$ is probably sensitive to the material properties of the target body, which means that the true uncertainty in $m_C$, as derived by extension of the Bottke et al. (2005a,b) model for $Q_C$ and $d_C/d_B$, is hard to even estimate. Moreover, for bodies other than growing planetesimals there is no assurance that $m_L$ would bear a strong relationship to $m_C$. If the Bottke et al. (2005a,b) model for estimating $d_C/d_B$ were applied to the Moon, the predicted diameter of a catastrophic impactor (assuming $v_i = 20$ km/s) would be $\sim 1270$ km. For comparison, the biggest definite lunar impact-basin, South Pole-Aitken ($D_t \sim 1100$ km), probably formed from an impactor with $d_i$ of roughly 300 km (Holsapple, 2003; of course, the Moon’s size and origin as a natural satellite probably led to an unusually prolonged history of resurfacing by igneous activity, in comparison to most planetesimals). The accuracy of the model probably diminishes (most likely tending to underestimate $z_A$; see next section) as $D_L/d_B$ increases past $\sim 1$. For a planetesimal evolving during the middle-late stages of accretion, when encounter velocities are still of order 2-3 km/s, $m_C/m_B$ is $\sim 4$ times greater, $d_C/d_B \sim 1.6$ times greater, and $D_C/d_B \sim 1.07$ times greater than the 5 km/s ratios (Fig. 5 and 7); and as a consequence of the higher $D_C/d_B$, for any given $m_L/m_C$ scenario, the implied total volume of ejecta is $1.07^3 = 1.23$ times greater than in the nominal 5 km/s model. Finally, if ever the ejecta is subjected to significant sintering-densification (see below), $z_A$ becomes an upper limit for megaregolith thickness.

**Additional single-crater complexities**

Equation (2) implicitly assumes that the volume of the ejection/excavation zone is identical to
the volume of the ejecta in its final state as debris strewn (mostly) onto the surface of the body (the stipulation sometimes called “Schröter’s Rule”). Here, two significant but offsetting aspects of approximation are involved. First, to the extent that the final debris is in general more porous and thus of higher volume than the pre-impact material, (2) tends to underestimate \( z_1 \). However, particularly on smaller bodies, a significant fraction of the ejecta may not land at all, and in that sense (2) tends toward overestimation of \( z_1 \). One way to gauge the magnitude of this effect is to apply the comprehensive crater scaling model of Holsapple (2003), which estimates the mass-velocity spectrum for the ejecta, with \( g \) and escape velocity (which scales approximately as \( d_B^{1/2} \)) among the input parameters. As summarized in Fig. 14, the fraction of ejecta launched off the body is in general roughly offset by the porosity-inflation factor (assumed to be \( \sim 5/4 \)). In actuality the porosity of the ejecta accumulation probably anticorrelates with \( d_B \) (i.e., with \( g \), which drives compaction). Other factors, nonspheroidal shape and the typically fast asteroidal rotation rate, probably slightly increase the launch-off proportion (Geissler et al., 1996). The modeling used for Fig. 14 conservatively assumes “rocky” strength for the target body. Pi-group scaling (Holsapple, 2003) indicates that for a given crater size, a stronger target results in a higher proportion of ejecta loss by launch-off. As an extreme example, during formation of a 5 km \( D_t \) on a small \( (d_B \) of order 10 km) asteroid, the proportion of launch-off is 5 times greater if the target is a strong “rocky” material than if it has strength equivalent to the powdery lunar surface regolith. (The mass of launch-off is about 2 times greater in the weak target for a given impactor mass, but the crater that forms, and the total volume of ejecta, is vastly larger in the weak target.) In terms of thermal-insulation implications, invoking a weakly cohesive target effectively preempts the megaregolith thickness issue, because weakly cohesive material will almost inevitably (in the context of atmosphereless bodies) be material rich in vacuous porosity, and thus constitute insulation equivalent to megaregolith even if it not formed mainly by aggregation of crater ejecta. In summary, simple application of (2) may overestimate \( z_A \), but probably not by an important factor unless \( d_B \) is <100 km.

The equation (2) model for \( z_1 \) does not agree well with results reported from hydrocode experiments by Nolan et al. (2001) for very large impacts on Gaspra, modeled as a sphere with \( d_B \sim 12.6 \) km. Nolan et al. (their Fig. 5) found \( z_1 \sim 16 \) m from formation of a single crater with \( D \sim 4.6 \) km. In their model some unspecified proportion of the ejecta is lost due to launch-off (however, the launch-off loss was probably low for this, the smallest crater they considered). In contrast, (2), assuming no ejecta loss by launch-off and that by “crater diameter” Nolan et al. meant \( D_t \) (and not a larger final \( D \)), implies for the 4.6-km crater a \( z_1 \) of only 4.3 m; a discrepancy of a factor of 3.7. A log-linear trend that Nolan et al. (2001) drew through their results indicates that a 3 km crater forms a \( z_1 \) of 13 m, which implies discrepancy versus (2) by a factor of 11. This discrepancy calls into question an inference by Nolan et al. (2001) that net ejecta deposition is “approximately linear in impactor size ... due to a combination of crater volume and fraction escaping” for large impacts. (Perhaps, despite other contextual indications, by “size” Nolan et al. meant mass.)
A complication not modeled by equation (2) may arise in cases where the $D_t/d_B$ ratio approaches or exceeds 1. A $D_t/d_B$ of ~ 1 implies the transient crater spans ~ 32% of the body’s circumference. In such a case, as Cintala et al. (1978) pointed out, the crater’s depth/diameter ratio (if depth is defined relative to the pre-impact surface, not to a chord across the crater rim) may increase markedly. Hydrocode modeling by Hammond et al. (2009; cf. Fig. 2 in Nolan et al., 2001) indicates that the maximum excavation depth $\omega_{\text{max}}$ is ~0.17±0.03 times $D_t$ in giant ($D_t/d_B \sim 0.2$-0.7) ~20 km/s lunar impacts. However, they also found that the $\omega_{\text{max}}/D_t$ ratio moderates at lower impact velocities; so the general implications of Hammond et al. (2009) for $\omega_{\text{max}}/D_t$ and the volume of ejecta $z_1$ are hard to gauge. However, if the giant impacts do excavate much deeper than 0.1 times $D_t$, they probably excavate higher proportions of “new” (as opposed to recycled) megaregolith than (3a) and (3b) imply. Comparison to the uppermost light-dashed curve in Fig. 9 (i.e., the ideal case of zero recycling) indicates that in scenarios where $D_t/d_B$ exceeds ~1, the nominal model may underestimate the total ejecta accumulation $z_A$ by 10-20 percent.

Other complications not modeled by equation (2) include: megaregolith destruction by compaction directly below the crater; the effects of secondary cratering (as the ejecta land, with compaction versus churning depending on, inter alia, the ejecta velocity); growth of the body (and thus its surface area) during accretion; and the transformation of a fraction of the ejecta into melt and even vapor. However, if the Moon is any guide, even impact-melt dominated ejecta tend to solidify into moderately porous, thermally insulating rock types (Fig. 1).

3. DISCUSSION

Temperature dependence of $k$

The conductivities shown in Fig. 1 are for 300 K, but temperatures within a megaregolith on a hot planetesimal or asteroid may extend from near 250 K at its surface to, near its base, the sintering $T$ of roughly (see below) 1000 K. At low-moderate $T$ the lattice-vibration component of thermal conductivity surely predominates, but the radiative component $k_{\text{rad}}$ is $T$-sensitive and insensitive to porosity. Early studies (Schatz and Simmons, 1972; Shankland et al., 1979) indicated a large $dk_{\text{rad}}/dT$ for olivine (which probably dominates the shallow interior of any not-yet extensively differentiated planetesimal or asteroid), such that $k_{\text{rad}}$ would reach, for typical mantle grain sizes, ~1.2 W m$^{-1}$ K$^{-1}$ at 1000 K. If this were accurate, the implication would be that regardless of porosity, a megaregolith’s $k$ becomes rocklike as $T$ approaches sintering. However, according to Hofmeister (1999, 2005), the early $dk_{\text{rad}}/dT$ estimates were grossly inaccurate, and olivine’s $k_{\text{rad}}$ only increases to ~0.2 W m$^{-1}$ K$^{-1}$ at 1000 K. Hofmeister (2005) also found that Fa content of olivine has little influence on $k_{\text{rad}}$. The deep megaregolith’s $k$ is probably moderated more by sintering-enhanced compaction, and by its relative immunity to gardening-pulverization (i.e., involvement in small cratering events), than it is by $dk_{\text{rad}}/dT$. 
Uneven ejecta deposition

Even in the low-g, high-surface-curvature, high spin-rate environment of a planetesimal or asteroid, ejecta probably land mostly within 2-3 radii of the (transient) crater’s rim (e.g., Nolan et al., 2001). Uneven megaregolith accumulation might be a significant limitation on the effectiveness of the megaregolith as a thermal insulation blanket. However, barring a flukish exception to power law model of size-frequency distribution, the biggest crater will usually be accompanied by a considerable number of comparably large craters formed at scattered random locations many other large craters. If $b = 2$, with average sampling and including craters formed both before and after the $D_l$ event, there will be 3 more craters with $D_t \geq 0.5 D_l$, an additional 12 with $0.25 D_l \leq D_t \leq 0.5 D_l$. Even if $b$ is as low as 1.5, with average sampling there will be 3 more craters with $D_t \geq 0.4 D_l$, and an additional 7 with $0.2 D_l \leq D_t \leq 0.4 D_l$. If $b = 2.5$, with average sampling there will be a total of 10 craters with $D_t \geq 0.4 D_l$. Divide these numbers by 2 to arrive at the numbers that on average will form at a later date than the $D_l$ crater. Areas that avoid ever being within a few radii of any major crater are probably minor. In terms of thermal implications, a moderating factor is that such a surface would generally at least undergo extensive gardening by small craters (i.e., develop a regolith; see “regolith within the megaregolith” section below), so its megaregolith, although thin, will tend to be more porous and have a lower conductivity than the global average megaregolith.

The Bottke et al. (2005a,b) model for relationships among $d_C/d_B$, $Q_C$ and $v_i$ implies that big $D_t/d_B$ craters are likely to be more widely scattered (in relative terms) on small-$d_B$ bodies, with their low $D_C/d_B$ (Fig. 5) and thus low $D_t/d_B$. However, on smaller bodies both seismic shaking (Cintala et al., 1979; Asphaug, 2008), and a momentum-transfer process that Nolan et al. (2001) call impact “jolting”, may lead to enhanced dispersal of the landed ejecta.

Very large ejecta

Ejecta that remain in the form of large intact fragments are in general much less porous, and thermal-conduction resistant, than accumulations of finely pulverized ejecta. The classic fragmentation model (Dohnanyi, 1969) produces a power-law size spectrum analogous to (4) with $\beta \sim 2.5$. Lab and field measurements confirm $\beta \sim 2.5$ for small-scale ejecta (Melosh, 1989). In any distribution with $\beta < 3$, mass is concentrated at the high-$d$ end. However, the distribution truncates at some high size, which depends largely on the abundance of mechanical defects in the target. A planetesimal or asteroid probably has mechanical defect abundance at least comparable to that of the deep lunar crust (Nolan et al., 2001). From observations of the rims of lunar craters (Moore, 1971) and of blocks on Ida, Lee et al. (1996) found a simple relationship between the largest ejecta block size $L$ (“size” here is the observable maximum dimension, which is probably a little larger than an equivalent diameter) and the crater diameter (final $D$, not $D_l$) for craters in these targets: $L \sim 0.25 D^{0.7}$. Based on this model, comparison with the total ejecta volume (1c) indicates that the craters of interest here yield largest fragments that have masses amounting to $\sim 0.0003$ ($D$ of order 500 km) to $0.02$ ($D$ of order 5 km) percent of the total ejecta mass. Assuming $\beta \sim 2.5$ for all smaller fragments,
only half of the total ejecta mass would be in fragments <0.044× as massive as the largest; only 25% would be in fragments <0.0036× as massive as the largest. Taken at face value, this model implies that survival of large blocks is a more important issue for smaller bodies. However, there may be an offsetting tendency for material toughness, i.e., large-fragment survivability, to decrease with decreasing \( d_B \).

The vestoid asteroids, which probably represent ejecta blocks from Vesta’s 460-km southern hemisphere basin (Asphaug, 1997), are larger in \( d \) (up to 14 km: Kelley et al., 2003) by a factor of 6 than predicted by the Lee et al. (1996) model. The timing of the biggest crater’s origin may be important for the large ejecta issue. If the largest crater forms early, then its biggest ejecta will be prone to demolition by subsequent cratering. If the largest crater forms after a history of bombardment by comparably big impactors, then the battered target interior will be more prone to break into small bits; unless in the mean time the interior has undergone both heating to igneous or near-igneous temperatures. In the igneous scenario, unless a very large impact occurs while the body is unusually susceptible to disruption (see, e.g., Warren and Huber, 2006), its interior, left undisturbed, will eventually reconsolidate into strong solids. Vesta may be an example of this scenario.

**The regolith within the megaregolith**

Regolith sensu stricto (especially fine-grained and porous surficial ejecta-debris) accumulates locally in area and time whenever small-\( D/d_B \) cratering (especially high-\( b \) cratering) persists through a stochastic lull in resurfacing events; i.e., high-\( D/d_B \) cratering and/or shallow magmatism. As noted by Robinson et al. (2002), the terminology of “megaregolith” versus “regolith” can be confusing. By historical accident, the seminal Shoemaker et al. (1969) work on extraterrestrial ejecta accumulations focused on the unusual context of mare lava plains, where the powdery lunar surface regolith is separated by only a few meters from near-intact bedrock (mare basalt, emplaced too recently to be thoroughly disaggregated). However, heavily cratered terrains such as the lunar highlands, where the regolith is underlain by megaregolith (the term was coined by Hartmann, 1973), do not fit as well into the original scheme of powdery regolith abruptly giving way to coherent bedrock. Asteroid geologists commonly draw little distinction between megaregolith and regolith, applying the latter term to any and all accumulated ejecta-debris (e.g., Housen et al., 1979). Nolan et al. (2001) extended megaregolith to include deep interior material that is impact-fragmented into coarse “rubble” without ever being ejected from a crater. As noted above, I prefer a definition that limits megaregolith to accumulated ejecta. In the lunar context, compared to megaregolith, regolith is distinctively finer in median grain size, much more porous, and richer in materials whose origins require exposure at the very surface, such as small impact-melt spheroids, implanted solar-wind noble gases, and agglutinates (Warren, 2001a; Wilhelms, 1987).

In terms of thermal implications, a regolith within a megaregolith is analogous to the megaregolith within the 8-km thermal “skin” modeled with Fig. 2 and equation (1). Results (again
assuming a steady state, neglecting internal heat generation; admittedly over-simplistic) for the relative $q$ implied by various assumptions regarding the regolith and megaregolith thickness and conductivity are shown in Fig. 15. For a regolith having $\sim 0.1$ times the megaregolith conductivity $k_{MR}$ (cf. Fig. 1; the 1 m deep soils probably have slightly lower $k$ than the overall several meters thick regolith), the regolith thickness must reach 11% of the overall megaregolith in order to cause a 50% increase in its overall thermal resistance. For comparison, the thickness of the lunar megaregolith is suspected to be $\sim 2$-3 km (Warren and Rasmussen, 1987); and the top $\sim 0.5$-1% of that thickness (in the maria, $\sim 0.2\%$) is regolith sensu stricto (Wilhelms, 1987). Fig. 15 thus suggests that the Moon’s rather thin regolith reduces heat flow through the overall megaregolith by a mild factor of $\sim 0.9$.

As discussed above, maximum crater excavation depth $\omega_{\text{max}}$ is never much greater than $0.1D_t$, so anywhere that a surface is exposed to a bombardment of tens-of-meters-scale and smaller cratering with a high $b$ (at these scales, from a variety of evidence, $b \sim 3.5$: Melosh, 1989) the upper few meters will grow increasingly pulverized, until some event resurfaces the area with rocky matter. Aspects of the impact process also “erode” regolith (Housen et al., 1979), but unless an impact is very large, while the global regolith is modified in detailed shape, its total volume is little changed. On the Moon’s maria-rich near side, the last resurfacing event was commonly basaltic lava extrusion. But more generally the resurfacing mechanism will be a mass of ejecta (new megaregolith) from one of what Housen et al. (1979) termed “scattered, large anomalous craters”. On a large body such as the Moon, the simple mass-addition effect is greatly enhanced by the jumbling that occurs (“ballistic sedimentation”: Oberbeck, 1975) when large distal ejecta land at velocities that are major fractions of the body’s escape velocity (2.4 km/s, for the Moon). Haskin et al. (2003) estimate that a single event $D_t/d_B \approx 0.21$ event, Imbrium, by a combination of churning and (at proximal locations) simple deposition, thoroughly reconstituted the megaregolith at practically all locations around the globe to a depth of at least 500 m. In detail, this estimate is sensitive to the size distribution and launch angle assumed for the distal ejecta, and to the strength-resistance of the megaregolith against secondary churning. But it seems clear that virtually no lunar regolith, sensu stricto, survived through this one 3.9 Ma (Wilhelms, 1987) event, except in the form of regolithic clasts within the megaregolith. Thus, it is naïve to assume (e.g., Wilson et al., 2008, p. 6158) that some fraction of the present few tens of meters of lunar highlands regolith is all that ever formed during the first 1/2 Ga on this body.

A volume of regolith that becomes thoroughly dispersed as a minor component within a jumble of megaregolith will marginally increase the porosity and thermal insulation of the megaregolith, but its effect on overall heat loss is not nearly as great as the effect when the regolith remains in place as a laterally continuous thermal barrier. The Haskin et al. (2003) model (cf. Petro and Pieters, 2008) implies that events with $D_t/d_B$ as small as $\sim 0.13$ are probably big enough to effectively recycle (i.e., churn-dilute by a factor of at least 5, which would leave the porosity, for example, only a few percent higher than the $\sim 17\%$ characteristic of the overall megaregolith) all regolith within 30 m of the surface on one hemisphere of the Moon; and a combination of 3 to 4 such events, at random
locations, would suffice to destroy virtually all regolith. Events of this $D_t/d_B$ are probably common
during the overall accumulation of a megaregolith. For example, if $b = 2$ and $D_t/d_B = 0.5$, with
average sampling there will be a total of 16 craters with $D_t \geq 0.125d_B$.

On smaller ($d_B$ not $>>100$ km) bodies, the churning when distal ejecta land is probably less
effective. Secondaries land at velocities never higher than $v_{esc}$, which for uniform density is directly
proportional to $d_B$. However, if small-body megaregoliths are weakly cohesive, as might be expected
from, inter alia, the expected lesser proportion of melt in small-body impacts (Melosh, 1989), that
lack of cohesion might largely offset the difference in landing velocity. Also, the small-body
enhanced process of launch-off will from time to time cause regional regolith destruction (cf. Housen
et al., 1979).

If this analysis is correct, the lunar-science custom of distinguishing between megaregolith and
regolith also has merit for asteroids and planetesimals, at least for those $>>100$ km in $d_B$. The
destruction of regolith in rare but almost inevitable large events means the distribution of material
types truly is to a considerable degree bimodal. Meteorite samples are of little help in constraining
this issue, because as lunar samples show (Warren, 2001a), only regolith breccias that are
extraordinarily tough will often survive the rigors of transit down to Earth’s surface. In any event, in
terms of global thermal insulation, the regolith (sensu stricto) is unlikely to ever become thick
enough to be important compared to the megaregolith.

**Porosity reduction: I. Simple compaction and aqueous metamorphism**

The porosity engendered by accumulation of ejecta into a megaregolith is incremental to
whatever residual accretionary porosity the body retains in its deeper interior. Realistically, most
planetesimals consist mainly of fractured rock, with at least slight porosity, from their very
beginnings. Planetesimal growth occurs through impact-agglomeration of countless former target
bodies. The net effect of each impact is usually an increase in the overall fracturing and porosity of
the target, albeit impact-compaction is locally effective with already porous targets (Housen and
Holsapple, 2003) during relatively gentle impacts, which were common during early stages of
accretion (Weidenschilling and Cuzzi, 2006), or through impact melting during end-stage impacts
between megameter-scale bodies. Wilson et al. (1999) modeled in a qualitative way the evolution of
porosity during generations of successively disrupted and reassembled planetesimals, and suggested
that 20-40% was probably the typical outcome. Nolan et al (2001), among many others, have also
discussed the development of deep porosity in small bodies by non-ejective impact processes.

Housen and Wilkening (1982) referred to such material as “accretionary megaregolith”, but
envisaged that it tended to be “destroyed by being converted into cohesive material by heating or
gravitational compaction.” Sintering will be discussed below. Near-surface, low-$T$ compaction
probably was roughly comparable in effectiveness on smaller bodies as within the Moon. In its upper
few meters, the lunar regolith is far less porous than would be expected on the basis of simple self-
weight compaction (Stesky, 1978). Carrier et al. (1991) suggested that an impact-seismic shaking
contributes to this enhanced compaction. Impact-seismic shaking is now appreciated as an important surface modification process on asteroids (e.g., Asphaug, 2008). Compaction must also, locally, be enhanced by impact-induced compression, including the gentle impacts of large secondaries; especially if the pre-impact porosity is very high (Housen and Holsapple, 2003).

As estimated by Carrier et al. (1991; their Fig. 9.16), the lunar regolith compacts from ~60% porosity at the very surface, to ~43% at depth of 1 m, where $P$ is still just 0.0029 MPa; to ~36% at 10 m where $P$ is 0.029 MPa (assuming regolith $\rho \approx 1800$ kg m$^{-3}$). For $\phi = 36\%$, the lunar porosity-conductivity relationship (Fig. 1) implies $k \approx 0.02$ W m$^{-1}$ K$^{-1}$, i.e., about 1/10 of the lunar megaregolith’s $k$. For comparison, assuming a shallow-interior density of 2000 kg m$^{-3}$, the pressure $P_{95}$ at which 95 vol% of the body is deeper (i.e., at $r/r_B = 0.95^{1/3}$) is $\sim 5 \times 10^{-12} d_B^2$ (for $d_B$ in m and $P_{95}$ in MPa); and thus the $d_B$ of a body with $P_{95} \approx 0.029$ MPa is 79 km. A 200 km ($d_B$) body has a $P_{95}$ ~6.5 times the $P$ at which compaction within the lunar regolith modifies porosity to ~36%, and $k$ to ~0.02 W m$^{-1}$ K$^{-1}$. This analogy between the Moon and much smaller bodies should not be overdrawn.

Many asteroids have bulk densities suggestive of greater than 36% porosity (Britt et al., 2002), and regoliths on smaller bodies may differ in material properties (most importantly grain size) from the lunar archetype (Asphaug, 2009). Intragrain porosity, insusceptible to reduction by low-$P$ compaction, may greatly augment the intergranular porosity that is susceptible. But the lunar analogy does suggest (a) that real-world vacuous planetary compaction, even at low, fixed $T$, is more complex and efficient than simple self-weight static compression; and (b) that planetesimal models (Hevey and Sanders, 2006; Sahijpal et al., 2007) assuming pre-sintering $k \approx 0.001-0.002$ W m$^{-1}$ K$^{-1}$, corresponding to ~60% porosity, may not be entirely realistic.

On warm and water-rich planetesimals, porosity might be further reduced by the formation of chemical precipitates, such as carbonates, sulfates, halides and oxyhydroxides, as well as hydrous silicates, such as serpentines and clay minerals, through aqueous metamorphism (Grimm and McSween, 1989; Brearley, 2006). Liquid water is potentially abundant for temperatures from 273 K to its critical point at 647 K and 22 MPa; and its vaporization $T$ is roughly $500 + 80 \log P$ (for $P$ in MPa). If liquid water is abundant and the body is sufficiently large (high-g) and permeable, hydrothermal convection will occur (Young et al., 2003); in which case convective energy transport will greatly augment conductive heat flow, and thereby hold temperatures, at least in the shallow interior, close to the $P$-governed H$_2$O vaporization $T$. If the body is permeable but too small to sustain hydrothermal convection (Young et al., 2003, estimate this condition would hold up to $d_B \approx 120$ km), water mobilized by warm-up will simply flow in a “single pass” up towards the shallow interior. The water tends to become concentrated in the shallow interior, most obviously in the single-pass scenario, but probably also eventually in the case of hydrothermal convection (which as it wanes must come to resemble single-pass flow), particularly if the body’s deeper interior undergoes continued warming into the temperature range of dehydration (~700-800 K: e.g., Grimm and McSween, 1989) and/or sintering-densification (see below). If the water collects mostly in the
shallow interior, eventually most of the final products of aqueous metamorphism will also be concentrated there.

The mineralogical changes associated with aqueous metamorphism would diminish porosity, both by directly filling pores (with chemical precipitates such as carbonates) and by replacing dense anhydrous mafic silicates with low-density hydrous derivatives. The reduction in porosity will be particularly drastic if the new hydrous silicates include expansive clays such as saponite, which is abundant in some carbonaceous chondrites (Zolensky, 1995). However, water mobilized near the surface of a small body is also prone to be lost by venting, evaporation and (after crystallization near the cold surface) sublimation, particularly if the shallow interior is both warmed and fractured by intensive impact gardening. A scenario of major aqueous-metamorphic densification probably requires not only that the body’s initial bulk composition be water-rich, but also that its surface happens to avoid large-$D/d_B$ cratering during the stage of aqueous flow.

**Porosity reduction: II. Sintering**

Sintering can destroy megaregolith from the bottom up, by a form of global thermal-burial metamorphism. As noted by Sahijpal et al. (2007), sintering may also work from the top down, in the event of massive extrusions of lava. However, to invoke that scenario is to obviate the main concern with megaregolith, i.e., whether the body can retain heat efficiently enough to become anatetic. Hevey and Sanders (2006) assumed that sintering occurs in a warming planetesimal as the temperature passes 700 K, and changes the material from lunar regolith-like in terms of porosity into solid rock, thus causing a sudden increase in $k$ by almost three orders of magnitude. Sahijpal et al. (2007) likewise assumed that sintering, and a factor of 1800 increase in $k$, occurs at 670-700 K. But these were mere assumptions, citing Yomogida and Matsui (1984) for derivation of the sintering temperature. Akridge et al. (1998) assumed that in small bodies sintering is negligible, and thus $k$ remains similar to that of lunar regolith, even at $T \sim 1200$ K.

Densification, as materials-scientists term the gradual elimination of porosity by sintering, is a complex function of temperature, time, pressure, and material properties (German, 1996; Kang, 2005). Materials-scientists recognize three distinct stages of sintering-densification. The initial stage involves growth of “necks” between loosely packed grains with minimal grain coarsening. The second stage involves moderate grain coarsening and, eventually, elimination of interconnective porosity. The third stage sees major coarsening and, usually, slight further densification. At temperatures less than about 2/3 of the melting $T$, each stage is of long duration. If near the second-third stage transition grain coarsening happens too fast relative to pore-size reduction, “breakaway” of pores from grain boundaries can effectively forestall the elimination of the final ~10% of porosity, so that full densification may require “precise manipulation of the initial powder microstructure and heating cycle” (German, 1996).

Poppe (2003) conducted a series of experiments for constraining the sintering behavior of extremely porous (95%) amorphous SiO$_2$ spherules 0.78 $\mu$m in $r$. He extrapolated his results to
estimate that 0.1-MPa sintering of this material takes ~1 Ma at 1000 K. It may be unwise to even attempt extrapolation from these results to the sizes, intragrain porosities, compositions and pressures relevant to this work.

Yomogida and Matsui (1984) inferred “600 ~ 650 K” as the most likely temperature “where sintering starts to become important.” However, this estimate was based on a single assumed lithostatic pressure $P$ (~ 1 MPa), albeit they explicitly modeled the sintered $k$ as sensitive to an “effective stress” of 10 MPa (more will be said about “effective stress” below). Even the deepest level to which ejecta accumulation extends (Fig. 12) will not have $P$ as high as 1 MPa unless the body is bigger than roughly 200 km in diameter (Fig. 3; lithostatic $P$ scales as $d_B^2$). Also, more recent measurements have shown that the rate of deformation of olivine is sensitive to the presence or absence of water, and the grain-boundary diffusion data used by Yomogida and Matsui (1984; from Schwenn and Goetze, 1978) appear suitable for “wet” olivine but fast relative to “dry” olivine, by roughly an order of magnitude in “effective diffusion constant” (Karato et al., 1986). As one gauge of the potential importance of water, Faul and Jackson (2007) note that the time needed for 1200°C growth to 1 mm grain size in olivine aggregates may vary from ~1 year in wet conditions to tens of Ma in otherwise equivalent dry conditions.

Although most megaregolith matter is probably crystalline, not glassy, sintering can be most readily constrained for glasses, where its rate and duration $t_s$ are determined by viscous flow. It seems unlikely that crystalline mafic silicates would sinter-densify faster than mafic silicate glass, so the glass rate may be viewed as an upper limit on the rate for crystalline matter. Simonds (1973) applied a viscous-flow model to the sintering of lunar-basaltic glass. Assuming spheroidal grains, the governing equation (cf. German, 1996) is

$$ t_s = \frac{2}{3} \left( \frac{\eta}{\gamma} \right) \left( \frac{X^2}{r} \right) $$

where $\eta$ is the $T$-dependent viscosity, $\gamma$ is the surface-interfacial energy (surface tension), $r$ is the grain radius and $X$ is the assumed radius of the neck between touching grains. For modeling $t_s$ at the late-initial stage of sintering, I assume $X = r/5$ (Simonds, 1973) and $\gamma = 0.5$ J m$^{-2}$ (Cooper and Kohlstedt, 1982). To represent the $T$-dependent $\eta$, I employ the results of Cukierman et al. (1973) for the composition of lunar olivine basalt 15555 (similar $\eta$ is implied by a wide variety of basaltic compositions; the viscosity data set used by Simonds was never explicitly published). With these parameters, (8) indicates that the $T$ for the late-initial stage of sintering, assuming a grain radius of order 0.1 mm and a $t_s$ of order 1-5 Ma, is ~785 K (± 15 K for any factor of 10 variation in the grain size; given the likely crucial importance of $^{26}$Al in primordial heating, the duration of sintering in all but the largest of planetesimals was probably limited to <<10 Ma). Zagar (1979) extended this model into one that explicitly addresses the diminution of porosity $\phi$. In German’s (1996) slightly simplified form, Zagar’s equation is

$$ t_s = -2r \frac{\eta}{\gamma} \ln \left( \frac{\phi}{\phi_i} \right) $$

where $\phi_i$ is the initial, presintering porosity. Results, based on the same parameter assumptions as
used for (8), are shown for a range of porosity reduction factor and \( r \) in Fig. 16. Assuming \( r \) of order 0.1 mm, Fig. 16 implies that basaltic glass a would have gone through the intermediate stage of sintering at \( \sim 820 \) K. But this merely represents a 0.1-MPa result. The process would go faster (or in a given time period, at lower \( T \)) at the \( P \) of the base of a megaregolith on a large asteroid or planetesimal (Fig. 3), and faster still in a large body’s deep interior. Still, it is noteworthy that at low \( P \), sintering of basaltic glass occurs at a considerably high \( T \) than Yomogida and Matsui (1984) estimated for sintering in general.

During pressure-sintering, ambient pressure is amplified into an “effective stress” or pressure \( P_{E} \) that develops at contact areas between grains. If \( P \) is high enough to be a controlling factor (German, 1996, indicates this condition begins at a \( P \) of order 0.1 MPa), the time \( t_s \) for a given extent of sintering-densification scales as \( P_{E}^{-1} \) (Kang, 2005, equation 5.23). German’s (1996) equation 7.8 indicates that \( P_{E} \) may be approximated well enough for present purposes (assuming vapor pressure even in closed-off pores remains low) as a function of porosity:

\[
P_{E} \sim P \phi^{(6.7\phi)}
\]  

(10)

This approximation is valid for the \( \phi \) range 0-\( \frac{1}{3} \). If \( \phi \) exceeds \( \frac{1}{3} \), the approximation begins to significantly underestimate \( P_{E}/P \). As some examples, as \( \phi \) approaches zero, the \( P_{E}/P \) ratio approaches 1, but \( \phi = 25\% \) implies \( P_{E}/P = 5 \); and \( \phi = 33\% \) implies \( P_{E}/P = 13 \). In short, for the \( \phi \) range (say 10-25\%) relevant to the early stages of megaregolith sintering \( P_{E}/P \) is roughly 3, but for reduction of the first half of the porosity from a powdery (Moon-style) regolith \( P_{E}/P \) is >>10.

Summing up, “the” temperature of sintering-densification \( T_s \) is a function of, inter alia, both pressure and the order of magnitude abundance of water, which by speeding diffusion enhances sintering. If \( T_s \) is considered to represent a reduction in porosity to final value of order 10\%, then estimating \( dT_s/d(\log t_s) \) to be \( 146 \pm 34 \) K (German, 1996; Kang, 2005; Akechi and Hara, 1979; Alister et al., 1979) and assuming grain size \( (r) \) of order 0.1 mm, as a rough approximation we arrive at

\[
T_s \sim A - 146 \log_{10} (10P_{E})
\]  

(11)

(NB: only valid for \( P_{E} > 0.1 \) MPa; at lower \( P \), \( T_s \sim A \)), where \( T_s \) is in K, \( P_{E} \) is in MPa, and A ranges from \( \sim 900 \) K for water-rich conditions (which makes for agreement with Yomogida and Matsui, 1984; who assumed \( P_{E} = 10 \) MPa) to roughly 1300 K (constrained mainly by tenuous extrapolation from the grain-growth observations of Faul and Jackson, 2007) for anhydrous conditions.

The abundance of water is difficult to constrain. Most igneous meteorites, representing several tens of separate parent bodies (many different types of irons; several varieties of primitive acondrites; ureilites; aubrites; and HEDs, including some anomalous eucrites (Nyquist et al., 2009)), contain Fe-metal, which is an almost certain indication of anhydrous origin. Water is prone to react with Fe-metal to form FeO (+ ultra-fugacious H₂; McSween and Labotka, 1993). Even in these cases, however, water may have been abundant at some stage of the parent body’s warm-up.

The sintering temperature \( T_s \) implied by (11) is shown in Fig. 17; and also, with \( P_{E} \) translated
into combinations of $d_B$ and depth, in Fig. 18 (for two values of porosity, 5-15%). The reduction in porosity from 15 to 5% has about the same effect on $T_S$ as decreasing depth within any given body (i.e., lithostatic pressure) by a factor of two.

Near the surface, assuming equivalent density, the pressure-depth gradient $dP/dz$ (Fig. 3) is directly proportional to $d_B$. Pressure within the ejecta accumulation zone will range from zero at the surface to $P(z_A)$ at the zone’s basal depth. If $m_L$ tends to be in some consistent proportion to the critical impactor mass $m_C$, so that the final ejecta accumulation thickness (in kilometers) scales roughly as $d_B$ (Fig. 12), then $P(z_A)$ and the other pressures within the zone, e.g., $P(0.5z_A)$, will be proportional to $d_B^2$. Thus, as the bodies heat, sintering will densify the lower ejecta accumulation zone at a much lower $T_S$ in a big body than in a small one (Fig. 19). Except for the uppermost ~10% of the ejecta accumulation zone (and only then in large bodies), the range in $T_S$ within any given ejecta accumulation zone is mild in comparison to the $T_S$ diversity that arises as a function of $d_B$. Fig. 19 shows results for only one assumed $m_L/m_C$, but the effect of a different $m_L/m_C$ is also comparatively mild; e.g., the $T_S$ of the bottom of the zone varies by ~50 K as the assumed $m_L/m_C$ ratio is varied (for any given $d_B$) from 0.25 to 0.75.

The sensitivity of $T_S$ to $d_B$ means that destruction of megaregolith by sintering is more efficient on larger bodies. As heating proceeds, smaller bodies end up maintaining almost full insulation from their comparatively thin ejecta accumulations at temperatures where on larger bodies the megaregolith’s thermal resistance is reduced by a large factor. For example, at 1100 K, ejecta-zone sintering is still, according to (11), negligible for porosity up to ~16% on a body 150 km in $d_B$, in which the bottom of the ejecta zone, for $m_L/m_C = 0.5$ (as assumed for Fig. 19) is ~4.6 km deep. By contrast, in a 1000 km ($d_B$) body at 1100 K, sintering will have densified to <5% porosity (i.e., $k$ within a factor of two of solid rock) everything deeper than ~1.3 km. The same relationships would apply at ~700 K in a water-rich body.

Even a few hundred meters of megaregolith will constitute a significant thermal resistance. The thermal evolution will be determined by a complex interaction (beyond the scope of this work) between heat generation and heat loss, as moderated by the megaregolith’s insulation. But these considerations indicate that for the crucial transition from a hot but still solid interior, to an interior that undergoes (beginning at roughly 1400 K) extensive melting, size of the body is not so all-important as it would be in the absence of megaregolith insulation. The metamorphic vs. igneous-differentiation fate of a body may depend almost as much on the stochastic (although undoubtedly $d_B$-correlated) $m_L/m_C$ result, which determines its ejecta accumulation $z_A$, as on its sheer size and initial heat-source ($^{26}$Al) content.

Comparison with past work, known asteroids and the Moon

As reviewed by Housen and Wilkening (1982), early work on the issue of asteroidal regolith development typically assumed that most ejecta undergo launch-off during impacts onto hard, solid-rock surfaces, leaving only thin accumulations except on the largest asteroids. In their time-keyed
modeling, Housen et al. (1979) still predicted that only a thin ejecta accumulation ("regolith", in their simple terminology) develops on most asteroids; e.g., over $10^9$-to-$10^{10}$ years “a few hundred meters” thickness develops on a 100-km ($d_B$) asteroid, and 0.9 km on a 300-km asteroid. Those results are lower by a factor of ~10 than results derived here for $m_l/m_c \sim 0.5$ (Fig. 12; Table 1). The main reason for this discrepancy is that Housen et al. assumed much smaller $m_c$ and $D_C$ ($m_t$ and $D_t$ in their terminology), and thus much smaller $D_L$. My results show agreement with Housen et al. (1979) if I assume $D_L/d_B$ is fixed at ~0.5.

Housen et al. (1999) and Housen and Holsapple (2003) revisited impact cratering on porous asteroids, but their studies were focused on extremely porous (34-96%) targets (cf. Benz and Jutzi, 2007; Ciesla et al., 2009). As discussed above, analogy with the lunar regolith suggests that porosity $>>40\%$ is probably seldom sustainable on even a cold body $>>50$ km in diameter, except very near the surface where $P < 0.003$ MPa (cf. Fig. 3; as mentioned in the figure’s caption, adjustment for bulk density is required). In any event, to assume such high target porosity in the largest events obviates the question of whether or not the megaregolith becomes thick enough to have important thermal consequences.

Ward (2002) also gave estimates for the ejecta accumulation thickness for three different assumed target body sizes; and from those, it is possible to interpolate to various other sizes (Table 1). Like Housen et al. (1979), Ward chose to estimate cratering effects as a function of an assumed duration and rate. My choice, for the comparison in Table 1, of 5 Ga and a rate of 10 times the present-day (Earth) value is arbitrary. Ward’s modeling did not consider effects of different target body size, such as the role of $g$ in determining the $d/D_t$ ratio.

Figure 10 shows largest observable crater diameters $D_{LO}/d_B$ for several large rocky asteroids (Asphaug, 2008) and the Moon. These $D_{LO}/d_B$ are lower by factors of ~2-3 than the $D_L/d_B$ (~1) suggested by modeling the largest impactor mass as roughly 0.5 times the catastrophic-disruption mass $m_c$ (Fig. 7). However, these observations are best viewed as lower limits on the largest impacts experienced by these bodies. Impact-seismic shaking is effective at smoothing over the surfaces of smaller bodies, so much so that Asphaug (2008) suggests that this seismic process, rather than survival against impacts with mass $\sim m_C$, limits the observed asteroidal $D_{LO}/d_B$ ratios; and the size-dependency of the process leads to the decrease in $D_{LO}/d_B$ with decreasing $d_B$. With larger bodies, resurfacing occurs by magmatism. The Moon and Vesta were probably hot enough for long enough (a “magma ocean” is often invoked for both) that many of the largest impacts left no manifestation on the present surfaces.

Asteroids have by definition experienced a longer, colder evolution than typical planetesimals. After asteroids cooled to the point where sintering became insignificant, they continued to undergo a long history of reduced-intensity but nonetheless cumulatively important impact cratering. Thus, overall porosity is expected to be generally much higher within a large asteroid than in a typical similar-sized warm planetesimal. As reviewed by Britt et al. (2002), asteroid porosities, inferred for
~20 bodies (including Phobos and Deimos) from observed density and estimated “grain” density, are diverse but generally high. Even for the 15 largest \( (d_B > 50 \text{ km}) \) bodies, average porosity is 35%. Only the three most massive asteroids show clear evidence of size-related diminution of porosity (Pallas, Vesta and Ceres, average \( \phi = 5\% \)). Asteroids as massive as \( 2 \times 10^{19} \text{ kg} \) \( (d_B \sim 270 \text{ km}) \) show no such evidence. Unfortunately, no constraints are available for asteroids with \( d_B \) between \( \sim 270 \) and 520 km.

Asteroids much smaller than 50 km \( (d_B) \) are outside the scope of this work, and undergo a different style of cratering (strength regime) that in combination with their low escape velocity is unfavorable to retention of ejecta. Observations suggest their (mega)regoliths are indeed thin (Chapman et al., 2002; Robinson et al., 2002; Sullivan et al., 2002), albeit ejecta retention is far more efficient than most pre-1991 (Galileo’s visit to Gaspra) models predicted. The largest constrained asteroids are Mathilde (equivalent \( d_B \sim 53 \text{ km} \)), and to some extent Vesta. Mathilde has a whole-body porosity of \( \sim 52\% \) (Britt et al., 2002), making demarcation of a megaregolith within its shallow exterior impossible, given the limited observational evidence. The Dawn mission (Russell et al., 2004) will soon reveal much about Vesta. As discussed above, Vesta appears to have undergone an unusual evolution. After extensive melting, much of its interior was annealed, which probably imparted an extreme “monolithic” strength. Much later, its present megaregolith was produced (along with the vestoids) largely through the formation of one exceptionally large and late crater. For both Vesta and especially the Moon, the present megaregolith thickness probably reflects only a fraction of the total cratering, as the earliest surfaces were overprinted by magmatism. For the Moon, based on seismic and crater-morphologic data (Head, 1976), the geometries of large lunar grabens (Golombek, 1979), and radar data indicating that as lunar craters exceed \( \sim 20 \text{ km} \) in diameter \( D \) they begin to excavate a more cohesive type of material (Thompson et al., 1979), the megaregolith thickness appears to be roughly 1.5-3 km; increasing (possibly to well over 3 km) with proximity to the South-Pole Aitken basin (Thompson et al., 2009). Within uncertainty, this agrees with modeling (Fig. 10) if South-Pole Aitken is the largest impact since crustal genesis and sintering has not destroyed (densified) a large fraction of the total ejecta accumulation. Sintering might yield a notably gradational megaregolith/solid crust transition, but the available constraints on the transition are far from conclusive in this respect.

**The thermal “skin” issue, revisited**

As discussed in the Introduction, a planetesimal that undergoes rapid heating by a uniform-distribution heat source (such as, before differentiation \(^{26}\text{Al}\)) will tend to evolve into an approximately isothermal interior beneath a thermally graded “skin”. According to Wilson et al. (2008), a complex equation on page 207 of Carslaw and Jaeger (1947) implies the skin thickness \( z_{\text{skin}} \) is \( \sim (kt)^{1/2} \), where the thermal diffusivity \( \kappa \) is \( k \) divided by the heat capacity and the bulk density (for solid silicate rock, \( \kappa \sim 8 \times 10^{-7} \text{ m}^2 \text{ s}^{-1} \)); which implies, if \( t \) is limited to a few Ma, the depth to the approximately isothermal deep interior, even for solid-rock \( \kappa \), will be \( \sim 8 \text{ km} \). Using a lower \( \kappa \), as
appropriate for megaregolith, \( z_{\text{skin}} \) would be yet lower (scaling as \( \kappa^{1/2} \)). However, this simple formula appears to considerably underestimate \( z_{\text{skin}} \). There is a factor of 3 or 4 discrepancy between \( z_{\text{skin}} \) by this formula and the \( z_{\text{skin}} \) shown for the exact same scenario in Fig. 32 of the revised and extended edition of Carslaw and Jaeger (1959; also cf. Figs. 2-4 of Hevey and Sanders, 2006). I have implemented the full Carslaw and Jaeger (1947, 1959) equation and obtain perfect agreement with that Carslaw and Jaeger (1959) figure. Results for some pertinent scenarios are shown in Fig. 20. The three thick curves show results from models using the same \( \kappa \left(10^{-6} \text{ m}^2 \text{s}^{-1}\right) \) as assumed by Wilson et al. (2008). A good match to the thermal profiles shown in Fig. 4 of Wilson et al. (2008) is obtained, but with \( (\kappa t)^{1/2} = 2 \text{ km}, \) not 8 km; i.e., the “skin” is 4 times thicker than \( (\kappa t)^{1/2} \); and for the given \( \kappa, t \) must be shorter, by a factor of \( (2/8)^2 = 1/16 \), than stated by Wilson et al. (2008). The specific “Wilson et al.” simulation shown in Fig. 20 is derived using \( t = 0.127 \) Ma. The other two thick curves show correct results with \( (\kappa t)^{1/2} = 8 \) km. Note that the size of the body makes very little difference in terms of the thermal profile that develops, in the rapid-uniform heating scenario, in the outer few kilometers (on this important point, Wilson et al., 2008, were certainly correct).

Fig. 20 also shows (the set of six thin curves) results assuming a \( \kappa \) lower by a factor of 10 than the value used for solid rock by Wilson et al. (2008). This is reasonably apt for a small-body megaregolith, because a factor of 10 diminution in \( \kappa \) implies (cf. Fig. 1; the associated increase in porosity implies a diminution in density by \~4/3\) diminution in \( k \) by a factor of \~13. The relevant time for peak temperature is probably, assuming low average \( k \) and negligible \(^{60}\text{Fe}\), about 2-4 Ma (3-6 half-lives of \(^{26}\text{Al}\)) after accretion of the body (Hevey and Sanders, 2006; cf. age constraints from igneous meteorites: Nyquist et al., 2009). For 3 Ma, Fig. 20 implies \( z_{\text{skin}} \) is \~11 km. For much of the range of body size under consideration, the \( z_{\text{skin}} \) found in Fig. 20 is thicker, even at 2 Ma, than the estimated ejecta accumulation \( z_A \) (Fig. 12). As \( d_B \) passes \~400 km, a \( m_L/m_C \) of 0.5 may suffice to make \( z_A > z_{\text{skin}} \). This comparison assumes uniform \( k_{\text{MR}} \). It seems likely that gravity-pressure effects result in a rough anticorrelation between megaregolith porosity and \( d_B \); which would imply a correlation between \( k_{\text{MR}} \) and \( d_B \), and thus between \( z_{\text{skin}} \) and \( d_B \). Even so, \( d_A/d_B \) is probably greater than \( dz_{\text{skin}}/d_B \). In cases where \( z_{\text{skin}} \) is thinner than \( z_A \), the thermal advantages associated with body largeness will be to some extent moderated.

The limiting factors for planetesimal heat loss will be the lesser of \( z_{\text{skin}} \) and the megaregolith thickness; along with the \( k \) (i.e., the porosity) of the shallow interior above that level. The megaregolith thickness will generally be close to \( z_A \). But it may be significantly less than \( z_A \) if the ejecta accumulation has been sintered to an important extent, and/or has not yet grown to its approximate final extent. Even for planetesimals, the largest crater that ever forms on the body does not necessarily form before the body reaches its peak temperature. For thermal modeling, \( D_L \) is the largest crater that (in a statistical sense) has formed as the body has evolved. For that matter, the body’s size will not be static as it evolves. However, so long as accretion is oligarchical, the general expectation is that at any given stage \( m_L/m_C \) will be of order 1/4 to 1/2 (Fig. 6), and thus (albeit the
surface may be renewed by magmatism) $D_L/d_B$ will be close to 1.

4. CONCLUSIONS

1. A difference in approach versus previous ejecta accumulation studies is that here the model is keyed to assumptions for the largest impact crater size ($D_L$), with no explicit modeling of time. In conjunction with assumed cratering size-distribution exponent $b$ (constrained by comparison with present-day asteroids), $D_L$ implicitly constrains the approximate sizes of all other craters large enough to be significant contributors to the final megaregolith. Another noteworthy difference is that the typical $D_L$ is constrained (albeit still only loosely) using a more sophisticated model (Bottke et al., 2005a,b) for estimating the impactor mass $m_C$ that, at a given velocity, results in catastrophic disruption.

2. Globally averaged ejecta accumulation thickness $z_A$ is relatively constant over a wide range of body diameter $d_B$ for any given largest impactor mass ratio $m_L/m_B$ of order 0.001 (along with constant $b$ and impact velocity $v_i$). In general, for $b = 2$ and $v_i = 5$ km/s the relationship is $z_A$ (in km) $\sim 300(m_L/m_B)^{0.8}$. As a specific example, $m_L/m_B = 0.001$ results in $z_A$ ranging from 1.0 to 1.3 km for all $d_B$ between 50 and 800 km.

3. For planetesimals, the largest impactor mass $m_L$ is more likely some consistently major fraction of the catastrophic-disruption mass $m_C$. The total ejecta accumulation $z_A$ (assuming $b \sim 2$ and $v_i = 5$ km/s) is then roughly proportional to $d_B$, with $z_A/d_B \sim 0.04(m_L/m_C)$; i.e., assuming $m_L/m_C$ is between 1/10 and 2/3, $z_A$ will be 1-5% of the body’s radius $r_B$.

4. However, ejecta accumulations on bodies with $d_B \sim 100$ km may be significantly (roughly a factor of 1.6) higher than implied by the nominal model. This diameter is the “sweet spot” for high $m_C/m_B$ (and thus $D_L/d_B$). A 1000-km body, for example, is implied to have lower $D_L/d_B$ by (on average) a factor of $\sim 0.86$. The main advantage is from simple proportionality to $(D_L/d_B)^3$ (2b), but craters with very high $D_L/d_B$ ($> \sim 1$) also excavate to significantly deeper (lower $r/r_B$) levels, so they enhance the yield of fresh, as opposed to recycled, ejecta.

5. Global ejecta accumulations estimated by this approach are higher than in the classic Housen et al. (1979) study by a factor of roughly 10. This revision is caused mainly by higher (typical) $m_L$ that are suggested by higher estimated $m_C$.

6. For $b \sim 2$, the single largest crater will generally contribute close to 50% of the total of new (non-recycled) ejecta. Considering the stochastic nature of the cratering process, there are probably many cases where the single largest crater contributes $>>50%$.

7. For modeling of the thermal effects of ejecta accumulation among different bodies, significant stochastic variations probably arise from two effects: concentration of ejecta mass into a relative few large fragments (although if formed relatively early these may be largely eroded down by later cratering); and stochastically uneven ejecta distribution, especially on relatively small bodies.

8. In cases of stochastically uneven ejecta distribution, the lower $k$ associated with a thin layer of
regolith (sensu stricto; i.e., uncommonly fine-grained and porous ejecta-debris produced by thorough small-scale, high b gardening of the surface) probably to some extent compensates for low regional accumulation thicknesses. Otherwise, however, regoliths thick enough to have major thermal consequences probably seldom develop, because regolith tends to be destroyed by churning-dilution during the landings of much greater volumes of ejecta from the biggest handful of craters.

9. For any given radial position r/rB, the pressure sensitivity of the sintering process makes it effective at far lower temperature TS on larger (dB >> 100 km) bodies.

10. Planetesimals ~ 100 km in diameter may be surprisingly well-suited (comparable to bodies 2-3 times larges in dB, assuming equal heat production) for attaining temperatures associated with widespread melting, for four reasons: (1) the ~100 km “sweet spot” for high DL/dB; (2) the value of b for impactors that have near-critical mass (m/mC of order 1/3 to 1) declines markedly as dB increases from ~ 80 to ~220 km; (3) generally higher TS; and (4) on such relatively small planetesimals, the thickness zskin of the thermally graded layer during rapid whole-body heating is likely to exceed zA, whereas in large bodies zskin may sometimes be less than zA.

11. A water-rich composition may be a significant disadvantage in terms of planetesimal heating. The shallow interior may be densified by aqueous metamorphism. Also, if the deep megaregolith is water-rich it will have a lower TS.

12. Development of a megaregolith thick and porous enough to have important thermal evolution consequences is practically inevitable. However, the cratering process that generates megaregolith is stochastic enough to leave great scope for diversity of outcome.

13. More work is needed, especially on two issues: heat loss with a thick but uneven coverage of megaregolith, and pressure-aided anhydrous sintering.

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REFERENCES


Table 1. Global ejecta accumulation results ($z_A$, in km) compared with previous work.

<table>
<thead>
<tr>
<th>$d_B$ (km)</th>
<th>100</th>
<th>200</th>
<th>300</th>
<th>800</th>
<th>duration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Housen et al. [1979]</td>
<td>~ 0.3</td>
<td>---</td>
<td>0.9</td>
<td>---</td>
<td>&quot;10^8-10^{10}$ y&quot;</td>
</tr>
<tr>
<td>Ward [2002]*</td>
<td>2.1*</td>
<td>2.5*</td>
<td>2.7*</td>
<td>3.3*</td>
<td>5 Ga</td>
</tr>
<tr>
<td>This work, $b \equiv 2$, $D_L/d_B = 0.5$</td>
<td>0.36</td>
<td>0.7</td>
<td>1.1</td>
<td>2.8</td>
<td>N/A</td>
</tr>
<tr>
<td>This work, $b \equiv 2$, $m_L/m_C = 0.5$</td>
<td>4.1</td>
<td>6</td>
<td>8</td>
<td>18</td>
<td>N/A</td>
</tr>
<tr>
<td>$b \equiv 2$, $m_L/m_C = 0.25$</td>
<td>1.8</td>
<td>3.6</td>
<td>5</td>
<td>11</td>
<td>N/A</td>
</tr>
<tr>
<td>${\text{variable } b}$, $m_L/m_C = 0.5$</td>
<td>(2.3 →) 4.4</td>
<td>(1.4 →) 4.8</td>
<td>(1.6 →) 6.5</td>
<td>(3.0 →) 27</td>
<td>N/A</td>
</tr>
<tr>
<td>${\text{variable } b}$, $m_L/m_C = 0.25$</td>
<td>(2.3 →) 2.0</td>
<td>(1.4 →) 2.7</td>
<td>(1.6 →) 4.2</td>
<td>(3.0 →) 18</td>
<td>N/A</td>
</tr>
</tbody>
</table>

* For Ward [2002] results, I have very arbitrarily selected 10x present-day cratering rate; the present-day rate would imply accumulation thicknesses ~ 0.14x those shown.

Ratio $m_L/m_C$ is probably anomalously high at $d_B \sim 100$ km (see text). The choices for variable $b$, shown in {brackets}, conservatively equate the $\beta$ implied in Fig. 7 with $b$. 

References:


Figures

1. Thermal conductivity ($k$) at 300 K, plotted as a function of vacuous porosity for lunar rocks and soils. Data are from Horai and Winkler (1980) and other sources cited in Warren and Rasmussen (1987), most notably, for the 1 m deep soils, $k$ from Langseth et al. (1976; their Fig. 6) and porosity from Carrier et al. (1991). The breccias are mostly impact-melt breccias, except for 77017 (porosity = 15%) which is a fragmental breccia, and 10065 (porosity 24%), which is a regolith breccia. The exponential equation for the line between solid rock and (average) surface soil is $k = 2e^{-0.1246\phi}$. The one clear deviant from the trend, mare basalt 70017, is unusually coarse-grained, and Horai and Winkler (1976) used an uncommonly small sample for their 70017 measurements.
2. Results from application of equation (1) to cooling of a megaregolith-covered body (shown in schematic cross section in the inset): relative steady-state heat flow $q$ as a function of the relative conductivity $k_{MR}$ of the megaregolith, shown for a range of assumed megaregolith thickness. The models shown assume the depth to the ~isothermal deep interior (depth to $r_1$) is 8 km (Wilson et al., 2008; see text) and that the body is ~500 km in diameter $d_B$. (Results are not very sensitive to $d_B$; e.g., for megaregolith thickness of 1 km and $k_{MR}/k_{rock} = 0.1$, relative $q$ varies by only a factor of 1.074 as $d_B$ ranges from 100 to 1000 km.) The actual situation is never as simple as modeled here, as the body is being heated from within by, e.g., (in the case of an early planetesimal) $^{26}$Al.
3. Pressures within bodies of uniform 3000 kg m\(^{-3}\) density, calculated using equation 2-64 of Turcotte and Schubert (1982). To scale to a different density, multiply the indicated \(P\) times (density/3000) squared. Grey region indicates estimated range of \(P\) for deepest, most sinter-densification prone portion of megaregolith as modeled (assuming \(b = 2\) and \(D_L/d_B = 0.6-1\)) in this work.
4. Depth provenance for ejecta from an individual crater, as implied by Maxwell’s (1977) Z model assuming $Z \sim 2.9$ (a depth/diameter ratio for the excavation zone of precisely 0.1 implies $Z = 2.734$; however, as explained in the text, the precise choice of $Z$ is not important for this diagram).
5. The critical impactor mass for catastrophic disruption $m_C$, expressed as the ratio $m_C/m_B$, calculated as a function of target-body diameter $d_B$ for a range of impact velocities, by extension of the $Q_C$ model of Bottke et al. (2005a,b). Also shown for comparison is the 5 km/s $m_C/m_B$ implied by the lower $Q_C$ estimate of O’Brien and Greenberg (2005).
6. Poisson-statistical probability for the largest mass of impactor $m_L$ in terms of $m_L/m_C$ ratio, assuming that the body is a fortunate survivor among many that have been catastrophically disrupted, and that the population of impactors conforms to the power-law size distribution with slope $\beta$ (the $\beta$ shown refers to the diameters of the impactors, albeit this chart shows diameter translated into mass).
7. The largest (transient) crater $D_L/d_B$, calculated as a function of target-body diameter $d_B$ assuming that the largest crater is formed with $m_L/m_C = 0.5$; i.e., the impactor mass $m_L$ is 50% as massive as the catastrophic-disruption mass $m_C$, as calculated by extension of the model of Bottke et al. (2005a,b). Translation from that mass $m_L$ of the largest impactor into (for an assumed impact velocity) its $D_L$ is modeled based on Holsapple (2003), assuming “rock” impacts into “hard rock” targets (for further description, see text) at 45 degrees and the indicated velocity. Light dashed curves show results assuming $m_L/m_C = 0.25-0.75$. 
The size-frequency exponent $\beta$ implied by the modern asteroids (Bottke et al., 2005a) shown in relation to the target body diameter $d_B$ and four values of the impactor mass ratio, $m_i/m_C$, i.e., the mass of an impactor ratioed to the catastrophic-disruption impactor mass (at $v_i$ of 5 km/s) for the given $d_B$. 

impact velocity = 5 km/s
1. Basic Results

9. Results for relative thickness of the global ejecta accumulation $z_A$ as a function of the size of the largest crater ($D_L$) and the crater size-frequency exponent ($b$). The heavy-continuous curves represent averages from 10 different models (for each $b$ value in the sequence 1.5, 1.75, 2, 2.25 … 3), each with 65,000 model craters forming in a different random sequence. The thin-dashed curves indicate, for the $b = 2$ model, results based on extreme variants of crater formation order: The curve on the high-$z_A$ side of the main, random-order curve represents a model with craters forming (implausibly) in sequence from small to large; the curve on the low-$z_A$ side of the main, random-order curve is based craters forming (implausibly) in sequence from large to small.
Global ejecta accumulation thickness \( z_A \) expressed in kilometers, as a function of the size of the largest crater \( (D_L) \) and two different assumed values for the crater size-frequency exponent \( (b) \), as calculated for a range of target body diameter \( d_B \). These results were derived from the same averaging of ten 65,000-crater models as described for Fig. 3. Assumptions include target body consisting of Holsapple’s “hard rock” \( (\rho_B = 3200 \text{ kg m}^{-3}) \), impactor 3000 kg m\(^{-3}\) hitting at \( v_i \) of \(~5\ \text{ km s}^{-1}\) and 45 degree impact angle. Also shown on the assumption that \( b = 2 \) are the sizes of the largest craters on asteroids Vesta (V), Amalthea (A), Mathilde (M), Ida (I) and Gaspra (G), along with the Moon (crescent symbol) and Phobos (P), as compiled by Asphaug (2008). The Moon is plotted under the assumption that its largest impact basin is South Pole-Aitken, 2500 km in \( d \). Under the hypothesis that Procellarum Basin is also an impact structure whose rim (or main ring) is 3200 km in diameter (Wilhelms, 1987), the Moon’s \( D_L/d_B \) would shift to \(~0.42\).
11. Global ejecta accumulation thickness $z_A$ expressed in kilometers, as a function of the mass of the largest impactor ($m_L$) and two different assumed values for the crater size-frequency exponent ($b$), as calculated for a range of target body diameter $d_B$. These results were derived from the same averaging of ten 65,000-crater models as described for Figs. 8 and 9. Five color-filled squares indicate (for $b = 2$ curves) the catastrophic disruption mass $m_C/m_B$ as estimated using the Bottke et al. (2005a,b) approach (Fig. 5; assuming $v_i = 5$ km/s). Assumptions include target body consisting of Holsapple’s “hard rock” ($\rho_B = 3200$ kg m$^{-3}$), impactor 3000 kg m$^{-3}$ hitting at $v_i$ of $\sim 5$ km s$^{-1}$ and 45 degree impact angle.
12. Results for the mean global ejecta accumulation thickness $z_A$ for a growing planetesimal as a function of the diameter of the body, shown for a range of assumed values for mass of the largest impactor $m_L$ in relation to the catastrophic disruption mass $m_C$; with $m_C$ estimated using the approach of Bottke et al. (2005a,b) for $v_i = 5 \text{ km/s}$. 
13. Fraction of accumulated ejecta contributed by the single largest crater, shown as a function of $b$ for three different values of $D_L/d_B$ (the three corresponding approximate $m_L/m_B$ ratios for impacts at 5 km/s into a 100 km ($d_B$) “rocky” target body are $3.6 \times 10^{-6}$, $9 \times 10^{-5}$, and $2.3 \times 10^{-3}$). These results were derived from the same averaging of 65,000-crater models, with the craters formed in 10 different randomized sequences, as described for Fig. 3. As noted in the figure, these results are not corrected for the greater proportion of launch-off associated with large events, especially important for small bodies.

NB: Uncorrected for launch-off, which diminishes relative contribution from the largest crater by roughly 5 to 10%, especially on smaller ($d_B < 100$ km) bodies.
The proportion of the ejecta lost by off-launch as constrained by pi-group scaling (Holsapple, 2003), shown as a “correction factor” for the implied final volume of the landed ejecta, as a function of transient crater diameter $D_t$ and target body diameter $d_B$. The correction consists of multiplying $z_1$ times the fraction of ejecta that lands (i.e., does not undergo launch-off) times a factor of 5/4 as (approximate) correction for increased porosity in comparison to the target material. Assumptions include target body consisting of Holsapple’s “hard rock” ($\rho_B = 3200$ kg m$^{-3}$), $v_i$ of order km s$^{-1}$ (the calculations were done assuming 5 km s$^{-1}$), and 45 degree impact angle.
Relative heat flow $q$ as a function of the regolith/(total megaregolith) thickness ratio, shown for a range of assumed relative conductivity $k$ of the regolith and megaregolith. These results are applicable to bodies of all sizes, and a wide range (0.001-0.01) of assumed megaregolith thickness/$r_B$ ratio (varying megaregolith thickness/$r_B$ within this range has negligible effect on the relative $q$).
16. Time-temperature relationships for intermediate-stage sintering of spheroids of basaltic glass, modeled as an isothermal process by the method of Zagar (1979) for grain size ($r$) of 0.01-1 mm and porosity/initial porosity ratio ($\phi/\phi_i$) of 0.3-0.7.
17. Sintering temperature $T_{S}$, for reduction of porosity to roughly 5-10%, calculated as a function of “effective” (grain-contact) pressure $P_{E}$ using equation (11).
18. Sintering temperature $T_s$, calculated as a function of target-body diameter $d_B$ and depth (i.e., $P$). Results are calculated for sintering through two porosities: 15% (heavy dashed curves) and 5% (light continuous red curves).
19. Sintering temperature $T_S$, calculated as a function of target-body diameter $d_B$ assuming that the largest crater is formed with $m_L/m_C = 0.5$; i.e., the impactor mass $m_L$ is 50% as massive as the catastrophic-disruption mass $m_C$, as calculated by extension of the model of Bottke et al. (2005a,b). Results are calculated for three levels within the ejecta accumulation zone (whose bottom depth is $z_A$), and for sintering through two porosities: 15% (heavy continuous curves) and 5% (light dashed curves).
20. Thermal profiles showing the thermally graded outer “skin” that envelops the approximately isothermal interior of a rapidly heated planetesimal, as modeled per Carslaw and Jaeger (1959). The y-axis shows the temperature normalized to the $T$ at the very center of the body for the individual model (the diagram should not be taken to imply that a 100-km body gets virtually as hot as a 1000-km one). The three thick curves show results from models using $\kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}$ (i.e., apt for solid rock), but $(\kappa t)^{1/2} = 2$ km for the upper “Wilson et al.” (2008) model, and 8 km for the other two. The set of six light curves are for $\kappa = 10^{-7} \text{ m}^2 \text{ s}^{-1}$ (i.e., roughly apt for a megaregolith) and $t$ ranging from 2 to 7 Ma.