IMPACT DECOMPRESSION MELTING: A POSSIBLE TRIGGER FOR IMPACT INDUCED VOLCANISM AND MANTLE HOTSPOTS ?

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Abstract. We examine the potential for decompression melting beneath a large terrestrial impact crater, as a mechanism for generating sufficent quantity of melt to auto-obliterate the crater. Decompression melting of the sub-crater mantle may initiate almost instantaneously, but the effects of such a massive melting event may trigger long-lived mantle up-welling or an impact plume (I-plume) that could potentially resemble a mantle hotspot. The energy released is largely derived from gravitational energy and is outside (but additive to) the conventional calculations of impact modelling, where energy is derived solely from the kinetic energy of the impacting projectile, be it comet or asteroid; therefore the empirical correlation between total melt volume and crater size will no longer apply, but instead be non-linear above some threshold size, depending strongly on the thermal structure of the lithosphere. We use indicative hydrocode simulations (AUTODYNE-2D) to identify regions of decompression beneath a dynamic large impact crater, (calculated as P-Lithostatic P) using SPH and Lagrangian solvers. The volume of melting due to decompression is then estimated from comparison with experimental phase relations for the upper mantle and depends on the geotherm. We suggest that the volume of melt produced by a 20 km iron projectile travelling at 10 km/s into hot oceanic lithosphere may be comparable to a Large Igneous Province (LIP ~10^6 km^3). The mantle melts will have plume-like geochemical signatures, and rapid mixing of melts from sub-horizontal sub-crater reservoirs to depths where garnet and/or diamond is stable is possible. Direct coupling between impacts and volcanism is therefore a possibility that should be considered with respect to global stratigraphic events in the geological record. Maximum melting would be produced in young oceanic lithosphere and could produce oceanic plateaus, such as the Ontong Java plateau at ~120 Ma. The end-

Permian Siberian Traps, are also proposed to be the result of volcanism triggered by a major impact at ~250 Ma, onto continental or oceanic crust. Auto-obliteration by volcanism of all craters larger than ~200 km would explain their anomalous absence on Earth compared with other terrestrial planets in the solar system. This model provides a potential explanation for the formation of komatiites and other high degree partial melts. Impact reprocessing of parts of the upper mantle via impact plumes is consistent with models of planetary accretion after the late heavy bombardment and provides an alternative explanation for most primitive geochemical signatures currently attributed to plumes as originating from the deep mantle or outer core.

Introduction

Researchers have already suggested that several larger geological features had an impact origin, but have auto-obliterated the traditional evidence of impact by subsequent large-scale igneous activity. Examples of such suggestions include the Bushveld Complex (Hamilton 1970, Rhodes 1975), the Deccan Traps (Rampino 1987; Negi et al. 1993), the break up of tectonic plates (Seyfert and Sirkin 1979; Price 2001), the formation of oceanic plateaus (Rogers 1982) and catastrophic mantle degassing from volcanism triggered by oceanic impact (Kaiho et al. 2001, but see Koeberl et al. 2002). An alternative mechanism relating volcanism to giant impacts proposed by Boslough et al. (1986) concerned the potential for antipodal focussing of energy transmitted through the Earth to trigger volcanism on the other side of the Earth to the impact itself, although the physics of this specific mechanism have recently been questioned by Melosh (2000). These suggestions have usually been rejected on the grounds that an impact model is less plausible than the widely accepted plume model (Mahoney and Coffin 1997; Richards et al. 1989). In the case of the Deccan traps, an iridium-rich layer between flows is taken by Bhandari et al. (1995) to indicate that this volcanism was already active before the K/T bolide event, as an argument against impact volcanism. Similarly, convincing evidence for impact in rocks from the Bushveld Complex have not been found (e.g., Buchannan and Reimold 1998). The present paper is an attempt to demonstrate more rigorously the plausibility of an impact model for the initiation of a large-scale igneous event. In addition, Glikson (1999) pointed to the planetary-scale role of mega-impacts in the history of development of the Earth's crust, and drew attention to the likely preferential melting efficiency of mega-impacts in oceanic lithosphere due to their higher geothermal gradients and thinner crust. Many of Glikson's ideas and fundamental implications are substantiated by our results for decompression melting, as predicted both by Glikson and ourselves (Price 2001).

Central to this paper is our contention that the phenomenon of pressure-release melting, or decompression melting, described in detail later, is the key to understanding the volumes of melt generated during large impacts and that in part this process has been overlooked or wrongly de-emphasised (Melosh 1989; Pierazzo et al. 1997). Melosh (2000) contends that there is no firm evidence that impacts can induce volcanic activity in the impact crater region, and he presents strong arguments, based on the amount of energy available, against the proposal that an impact could trigger volcanism at a distance. He notes that suggestions of impact-induced volcanism have often been based on observations of the large basalt-filled basins on the lunar nearside, but these are undermined by the discovery of large unfilled farside basins, and by the evidence that nearside volcanism apparently postdated basin formation by as much as 1 Gyr. Melosh concluded that pressure-release melting was highly unlikely on the Moon and he discounted the possibility of presure-release melting on the Earth. We agree with Melosh that pressure-release melting is unlikely on the Moon. If the temperature profile (temperature vs. pressure) were similar to the Earth's oceanic profile, excavation to a depth of approximately 500 km would be necessary to trigger pressure-release melting on the Moon. The largest verified terrestrial craters (Vredefort, Sudbury, Chicxulub; all ~200 km crater size) are all continental, and may be too small to have triggered pressure-release melting in a continental shield with low geothermal gradient; or if they did generate decompression melts, these have not yet been recognized. However, we do not agree that decompression melting can be ignored; our indicative simulations imply that a Sudbury-scale impact crater (~200 km diameter crater) would trigger instantaneous pressure-release melting if it occurred on oceanic lithosphere where geothermal gradients are high. Somewhat larger impacts on continental lithosphere would be required to trigger volcanism, which we propose to be the case for the Siberian Traps. The potential energy range available from very large impactors is vast. To put this into context, the largest conjectured terrestrial impact was the Moon-forming event, when an impactor 10% of the mass of the Earth (a true ‘mega-impact’) apparently ripped away part of the entire mantle possibly briefly exposing the Earths already-differentiated core (Canup and Asphaug 2001). We focus on less extreme large impacts likely to generate terrestrial craters in the range of ~>200 km, which may have been relatively common during the early part of the Earth’s history, and are still dwarfed by potential projectiles available in the (upper) size range of known near-earth crossing objects.

In this paper we address the traditional objections to an impact-related origin of major terrestrial igneous features and will conclude (1) that the plume hypothesis may not explain all of the features to which it is currently applied, (2) the generally dismissed process of pressure-release melting does provide a mechanism for larger impacts to generate large volumes (~10^6km^3) of melt and (3) the flux of larger impactorsis sufficient to explain the number of large igneous provinces (LIPs; ~10^6km^3 of melt) seen on Earth. We propose that a candidate oceanic LIP generated by impact volcanism might be the Ontong Java Plateau and a candidate continental LIP might be the Siberian Traps; we suggest a range of features by which this hypothesis may be tested. We propose that mantle hotpots triggered by large impacts offer a plausible upper mantle alternative to deep rooted lower mantle plumes, and will be associated with a comparable array of igneous, geochemical and metasomatic features. We recognize that this concept of reducing very large energetic geological processes to very short timescales and extraterrestrial triggers will require a substantial shift in approach by many traditional Earth scientists, but we believe that the underlying arguments are unavoidable.

Large Igneous Provinces (LIPs)

Large igneous provinces are widely thought to be produced by mantle melting resulting from a plume. Two main hypotheses have been proposed to explain the relationship between mantle plumes and flood basalts. In the plume-head hypothesis, Campbell and Griffiths (1990) consider that a large plume head, with a diameter of ~1000 km, originates at the core-mantle boundary and rises to form, beneath the lithosphere, an oblate circular disk with a diameter of ~2000 km. This leads to uplift of the overlying lithosphere of 0.5-1.0 km, and the development of volcanic activity. Plume-head melting occurs as the consequence of adiabatic decompression when the top of the plume reaches the top of the asthenosphere. Melting, they contend, will start at the hot leading edge of the top of the plume, where the plume can melt to produce high MgO magmas. As the plume head continues to rise and flatten, the cooler entrained-mantle edge of the plume may start to melt if it rises to sufficiently low pressures at shallow depths. In the second model, White and McKenzie (1989) assumed a much smaller plume, with an unspecified origin. They emphasized that it is the production of melt material, which is of paramount importance, and note that the potential temperature of the plume is only 100-300^{o}C higher than the surrounding mantle. Only in the low velocity zone (LVZ) are the P and T conditions such that the mantle is close to melting. As the increase in temperature caused by the plume is modest, the plume will only give rise to melting in a relatively narrow depth zone immediately beneath the LVZ. Consequently, they conclude that the depth of the stem of the plume is immaterial. But vital to their model is the coincident development of lithospheric thinning, which determines the volume of melt produced.

Although the role of plumes and hot spots in the development of volcanic chains such as Hawaii is widely accepted, there are some, however, who question whether such plumes can be responsible for all LIPs. Thus, for example, Saunders et al. (1992) maintain that the relatively short period between the initial contact from below, to the generation of melt is likely to be less than 10 Ma. Despite the heat transfer that may take place between plume and continental lithosphere, they argue that large volumes of melt material are unlikely to be generated, and even that the melt that occurs may freeze in-situ as heat is lost to the lithosphere. Campbell and Griffiths (1990) point out the shortcomings in the White and McKenzie hypothesis, while Anderson (1998) questions both plume models, and suggests that sources of geochemical anomalies and melting processes may occur instead at shallow depths in the mantle (we agree). Other authors suggest detailed field evidence in some large igneous provinces does not support either mantle plume model. Thus, in a recent review Sharma (1997) observes; "Collectively the [cited field] observations suggest that the Siberian Traps eruption cannot be linked directly either to lithospheric stretching in the absence of a plume or to hotspot initiation. Yet there appears to be consensus supporting a plume origin among those working on the Siberian Traps. Two pieces of evidence have engendered such a confluence of opinion: (i) the large volume ($>2x10^6$ km^3) of magma emplaced and (ii) the short duration ~1 Ma of eruption." Subsequent geological and geophysical papers are also incompatible with a conventional mantle plume as the cause of the Siberian Traps (Czamanske et al. 1998; Elkins-Tanton and Hager 2000).

In the following section, we indicate how decompression melting resulting from a large impact might generate large volumes of melt, which could be emplaced very rapidly, and might offer an alternative explanation to the mantle plume model for the Ontong Java Plateau, the Siberian Traps, and perhaps other LIP's.

Impact Melting

There is a well-established correlation between observed terrestrial crater size and the total volume of impact melt (Fig. 1, after Cintala and Grieve 1994). However, the observed craters are all in continental crust and perhaps the largest, Sudbury, is ~200 km diameter, with a lower bound estimated melt volume of ~ 8000 km^3. Studies of the largest known terrestrial craters, Sudbury, Vredefort, and Chicxulub, indicate similar rim diameters (~200-250 km). We concentrate on a detailed discussion of the well-studied Sudbury crater. Stöffler et al, (1994) summarized the results of an eight-year research project on the Sudbury structure. On the basis of textural, chemical, and isotopic evidence, they concluded that the Sudbury Igneous Complex (SIC) represents a differentiated impact melt with no significant deep-sourced magmatic or volcanic contribution. They also cite geophysical evidence that the SIC is not funnel-shaped with an extension to deeper levels of the crust. Their revised estimate of the original crater rim diameter is 220 km, and they estimate that the original volume of Sudbury impact melt was about 12,500 km^3. The depth of the transient cavity, the maximum depth of excavation and the maximum depth of melting are estimated to be in the ranges of 28 to 37 km, 15 to 21 km, and 25 to 35 km, respectively. These latter estimates are derived from a combination of constraining field observational data with the heuristic scaling relations presented in Melosh (1989). Assuming a 20 km/s impact of a projectile with density of 3 g/cm^3, Pi-group scaling relations predict a projectile diameter of about 14 km, corresponding to an impact energy of 8.6 x 10^{23} J. As noted by Melosh (1989), depending on one's choice among the proposed scaling relations, the uncertainty in prediction of impact energy from crater diameter could be as high as a factor of forty, for very large impact craters. We, therefore, have no qualms in comparing Stöffler's results with hydrocode calculations of impact events that differ by as much as a factor of three in impact energy. Pierazzo et al. (1997) calculated melt production for the 20 km/s impact of a 10 km diameter dunite projectile (3.3 x 10^{23} J) on various targets. In this context, the calculated volume of granitic melt, 8900 km^3, is in good agreement with Stöffler's estimated volume. The cited calculation was concerned only with melt production and did not run long enough to determine other crater parameters, such as transient crater diameter. Roddy et al. (1987) calculated the 20 km/s impact of a 10 km diameter quartz projectile (2.6 x 10^{23} J) on a layered continental site. The maximum transient crater depth was ~37 km, the upper limit of Stöffler's estimate, but the maximum transient crater diameter was ~80 km, substantially less than the 110 km estimated by Stöffler. One may note that the transient crater calculated by Roddy et al. (1987) has a much greater depth-to-diameter ratio than predicted by more generic calculations of O'Keefe and Ahrens (1993). These generic calculations generally supported Pi-group scaling. However, they do not appear to have modelled silicate phase transitions accurately and they were restricted to an impact velocity of 12 km/s. Roddy et al. (1987) explicitly modelled the effects of silicate phase transformations, which are known to have a major effect on wave propagation (Swegle 1990).

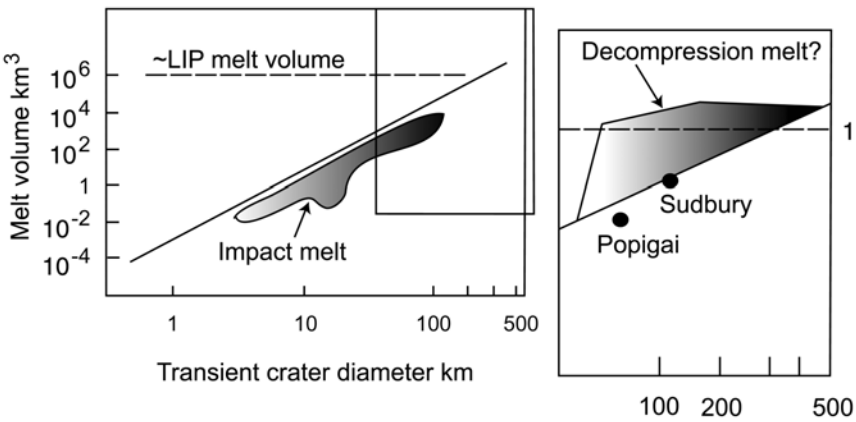

Fig. 1. Correlation of observed volume of impact melt versus crater diameter for terrestrial impact craters (eg: Cintala and Grieve 1994) compared with melt volume required for a Large Igneous Province (LIP ~10^6 km^3). Enlargement schematic shows the hypothetical increase in melt volume, due to decompression melting of lithospheric mantle, resulting in a non-linear relationship with crater diameter. Decompression melting triggered by impact might produce sufficient magma to feed a LIP.

We note that the estimated original volume of Sudbury melt, 12,500 km^3, is substantially less than the ~10^6 km^3 volume of a large igneous province. The calculations of Pierazzo et al. (1997) indicate that production of 10^5 km^3 of melt corresponds to a 20 km/s vertical impact of a 22.4 km diameter dunite projectile (4 x 10^{24} J). Using Pi-group scaling, the predicted transient crater diameter is 145 km, leading to a final diameter of about 300 km. For the 12 km/s vertical impacts modelled by O'Keefe and Ahrens (1993), the maximum depth of excavation is a constant fraction, about 0.05, of the final diameter. This implies that the predicted depth of excavation would be only 15 km, the lower bound of Stöffler's estimate for Sudbury (derived from Lakomy's (1990) geological study of the footwall breccia). The contradictions between Sudbury ground truth and the results of applying generic scaling relations imply that additional detailed modelling is needed. We are particularly interested in the behaviour of a heated target, such as the Earth, whose geothermal gradient is well understood.

Decompression Melting

Partial melting of the mantle occurs wherever the ambient temperature exceeds the mantle solidus temperature. Under adiabatic conditions in the upper mantle this situation arises during uplift or decompression of hot mantle, since the melting temperature for mantle peridotite increases with pressure (positive dT/dP). The mantle potential temperature is the temperature at any depth on the mantle solidus intersected by the adiabatic ascent path of a known melt temperature at the surface; this is adjusted for the additional thermal loss associated with latent heat of melting. McKenzie and Bickle (1988) correlated the total 2-D thickness of melt that can be extracted with the mantle potential temperature and degree of lithospheric thinning. Thus, the uniform thickness of oceanic crust (~7 km) is consistent with the volume of melt produced if the mantle has a potential temperature of ~1280^{o}C. We now consider how decompression melting may be induced by a large impact, where lithospheric thinning is effectively instantaneous, as required by McKenzie and Bickle (1988).

Decompression melting has not been encountered in laboratory shock experiments, nor is it expected, since it is a phenomenon restricted to large-scale impacts. It is well understood however, and is the main process, advocated by geophysicists for melting on Earth. It is seen in mantle xenoliths rapidly decompressed by rising volcanic magmas (Jones et al. 1983), and can be simulated in sacrificial solid-media experiments (Langenhorst et al. 1998). Therefore, it should be seriously considered whenever an impact is sufficiently large to cause the transient crater depth to excavate a substantial fraction of the local crustal thickness, and thereby cause a sudden drop in lithostatic pressure beneath the crater. This is because the temperature interval between ambient geotherm and lithological melting closes rapidly with increasing depth. By contrast, decompression of most crustal melts, causes freezing, since these generally have negative melting curves at low pressures (Wyllie 1979). There is thus an increasingly likelihood for decompression melting with increasing transient crater depth (H_t). Terrestrial geotherms are fixed at depths of approximately 400 and 660 km by the olivine to ß-phase and spinel to perovskite phase transitions respectively (Poirier 2000). At much shallower depths, geotherms are superadiabatic and vary according to lithospheric structure. Variations in the shallow geotherms represent exactly the region of interest for impacts. For oceanic lithosphere, geotherms vary with age from hot and young to cold and old. Geotherms for continental crust extend from the coolest gradients typical of stable cratons to those that overlap with lower oceanic values during active regional metamorphism.

The volume of decompression melt can be estimated by combining calculations of the pressure drop beneath an impact crater with mantle melting behaviour from published experimental data (as recently compiled, for example, by Thompson and Gibson 2000). For mantle peridotite, the degree of partial melting is, to a first order, related directly to the excess temperature above the solidus, for any given pressure. For example (Fig. 2), a pressure reduction of 15 kbar (1.5 GPa) is equivalent to raising the temperature by up to ~150^{o}C and, in peridotite previously at solidus temperature (T), leads to 20-40% melting. This simple observation is the crux of our argument, it represents an enormous potential for substantially melting the mantle <u>beneath</u> an impact crater, and has profound consequences for the geological history of the Earth. Melt compositions will vary according to the degree of melting and correspond approximately to komatiite (>30%), picrite (up to 30% melting) and basalt (circa 10-20% melting) respectively. At pre-impact depths shallower than ~75 km and for lower degrees of partial melting, there is also a compositional dependance on pressure for various varieties of basalt. In the most favourable case, thermally active oceanic lithosphere is already in a partially molten state at shallow depth.

Melting is not a kinetically hindered process because it is entropically so favourable, and so decompression melting will occur virtually instantaneously in hot mantle wherever there is sufficient reduction in pressure beneath a large impact, including reduction of lithostatic load by excavation of crater material, massive central uplift or lithostatic modification during formation of multi-ring structures.

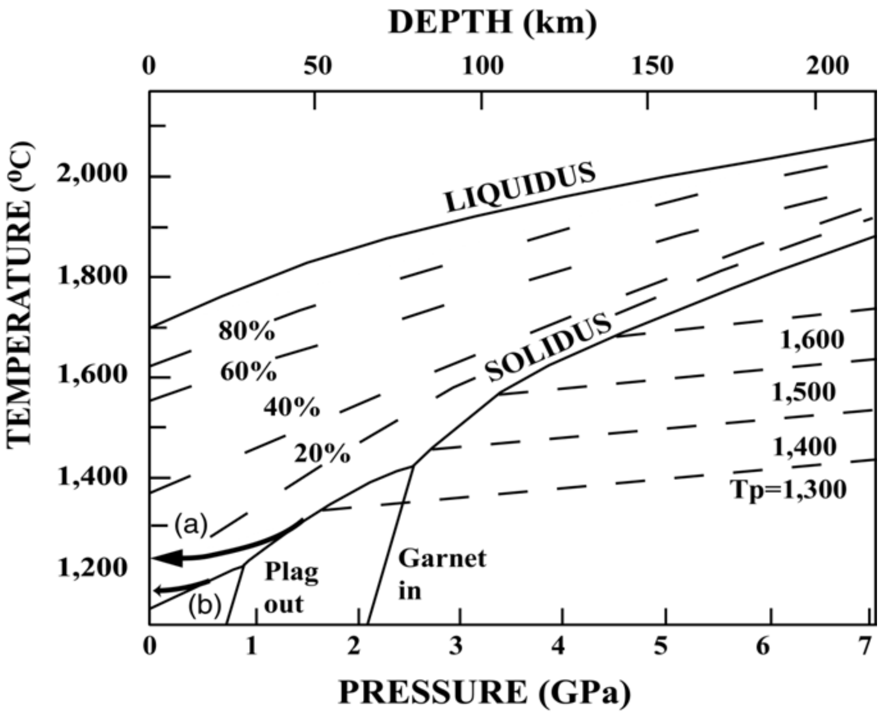

Fig. 2. Phase relations for mantle peridotite, showing degrees of melting at temperatures above the solidus, and curves for mantle potential temperatures (T_p) in upper mantle peridotite (after Thompson and Gibson, 2000). Melt compositions vary with the degree of melting and correspond to basalt (~10-20% melting), picrite (up to 30% melting) and komatiite (>30%). Two examples of decompression melting are shown, (corrected for latent heat of melting, but uncorrected for impact heating, or adiabatic uprise). A pressure decrease of (a) - 1.5 GPa is similar to raising the temperature by up to ~150°C and, in peridotite previously at solidus temperature, leads to 20-30% melting (picritic). (**b**) -0.5 GPa causes ~10% melting (basaltic). Any contribution to heating from impact would increase the degree of melting. We propose that decompression melting is important for hot target lithosphere (Earth) and may trigger large-scale volcanism. The mantle thermal anomaly could be long-lived and may superficially resemble a hotspot, but with no lower mantle root.

Hydrocode model

To quantify the instantaneous stress drop resulting from impact crater formation, we have performed indicative hydrodynamic simulations using the hydrocode AUTODYNE-2D (version 4.1) similar to that described by Hayhurst and Clegg (1997). The AUTODYNE-2D code has been well validated by data from small-scale hypervelocity experiments with a variety of target and impactor materials ((Hayhurst et al, 1995). The impact parameters were not intended to represent the complexities of a real impact, but were chosen so that most of the calculation would take place in a regime where Hugoniot uncertainties were small. The model "lithosphere" has a pre-determined pressure gradient to simulate the effects of lithostatic load, similar to the global geophysical model for the Earth called PREM (Primitive Earth Reference Model, Poirier 2000). There was no pre-impact thermal gradient employed in this simulation, but the self-compression density and thermal effects of gravitational and shock compression, were included. Lithostatic pressure and total pressure were calculated separately and integrated at the end of each run to quantify the pressure change, and specifically to determine regions of negative pressure, or decompression. The target dimensions are a 2-D box 300 km by 300 km, mirrored along the vertical axis of the crater to give a model space 600 km by 300 km. The lower boundary (300 km depth) was chosen to avoid back reflections in the model, but still caused noise in the data at the end of each run; this could be extended in future models or amended using a different solver, to a boundary transparent to shock. The target material selected was basalt, (SESAME EOS number 7530) using a no-strength model. Obviously future models could incorporate layers to represent crust, and peridotite to represent mantle. The pure iron impactor (SESAME EOS number 2410), was modelled as a sphere of 10 km radius with initial contact velocityof 10 km/s. The model symmetry used normal incidence, with idealised cylindrical symmetry. The SPH solver (Smoothed Particle Hydrodynamics) was filled with 22,500 particles, and progress in terms of (for example) velocity, density, temperature and pressure of each, being calculated in each step of the run.

After 40 seconds, the simulation produces a transient crater which has proportions of depth to diameter close to 1:1, which greatly excedes the 1:3 ratio of conventional impact crater assumptions (Melosh 1989); however, high aspect transient craters have been found in previous simulations. Thus, Roddy et al. (1987) using a 10 km quartz, 10% porosity impactor at 20 km/s find at maximum depth of 39 km ~ 30 seconds after impact, the diameter of the cavity is only 62 km (aspect ratio ~1:2). Also, Pierazzo et al. (1997) calculate for a similar impact with 10 km dunite moving at 20 km/s at the same time after impact, a crater diameter of ~60 km and depth of ~35 km (aspect ratio again ~1:2). Our calculation shows that, as expected on the basis of simple analytical considerations, the calculated depth-to-diameter ratio does indeed depend on relative shock and release properties of both impactor and target. We used a larger impactor-target density ratio (iron:basalt) that the simulations referred to above, and this led to the 1:1 depth-to-diameter ratio in our model. The calculations and analyses are validated by data from small-scale hypervelocity impact experiments with a variety of target and impactor materials (Pond and Glass 1970). The aspect ratio will directly influence the depth of mantle impact.

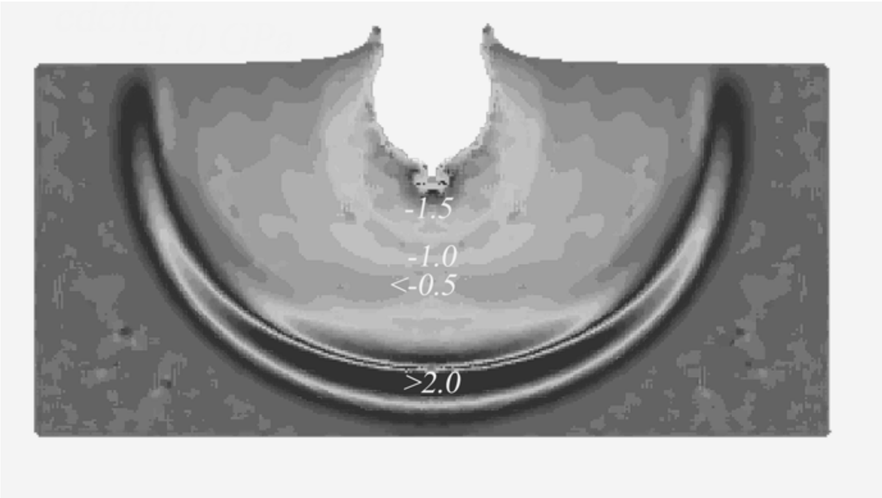

Fig. 3. Indicative hydrocode model of a simulated impact designed to show regions where decompression melting should occur. Model conditions: 300 x 300 km cell, impactor = 10 km radius iron, velocity 10 kms^{-1}, orthogonal impact, target = basalt (homogeneous), pressure gradient = PREM (Poirier 2000). Labelled are pressure zones relative to lithostatic load, for –1.5, -1.0, <-0.5 Gpa. If these zones occured in hot (young) oceanic lithosphere, decompression partial melting should occur. In the short term, this could ~instantaneously generate the volume of melt required, (basaltic, picritic or komatiitic) to form a large igneous province (LIP ~$10^6 km^3$); in the longer term, the thermal signature and could resemble a mantle hotspot, or impact plume (I-plume).

The results for the simulation show that after 40 seconds, there is a virtually spherical transient crater ~ 100 km in diameter (Fig 3), below which there are clearly identified zones of decompression. Figure 4 plots pressure versus depth below the transient crater, and shows three curves, for (a) lithostatic load (starting condition), (b) pressure induced by impact and (c) pressure difference (b-a). It can be seen that at 40 seconds after impact, a zone of decompression with magnitude ~ 1.0 GPa extends over a large interval from 120 to 180 km depth. Comparison of this information with Fig 2 shows that in the Earth's mantle, this decompression would occur in garnet peridotite and overlap with the stability field for diamond (pressures higher than ~5 GPa); if melting were initiated at this depth, and erupted, the geochemical signature of any resultant volcanic lavas should reflect the influence of garnet.

There are two causes of decompression – one long term, the other more transient. The latter transient effect is due to the interaction of rarefaction waves originating at the free surface. A longer-lasting zone of decompression occurs directly beneath the crater produced by the excavation of the crater material and the resultant loss of lithostatic load. The amount of melting generated by these processes can be estimated by direct comparison of the decompression values calculated in the simulation (as in Fig. 3) with the mantle melting relations shown in Fig. 2. Melting will occur virtually instantaneously over a range of depths during the course of the impact. We calculate that, for young oceanic lithosphere, the integrated volume of rock to experience super-solidus conditions is ~ 2 x 10^7 km^3 during the course of the shock event. This would lead to the production of ~ 3 x 10^6 km^3 of melt as the depressurised volume of mantle experiences an average of 15% partial melting. In a real impact, the melt extraction process would be complicated by, for example, gravitational instability of newly formed low-density

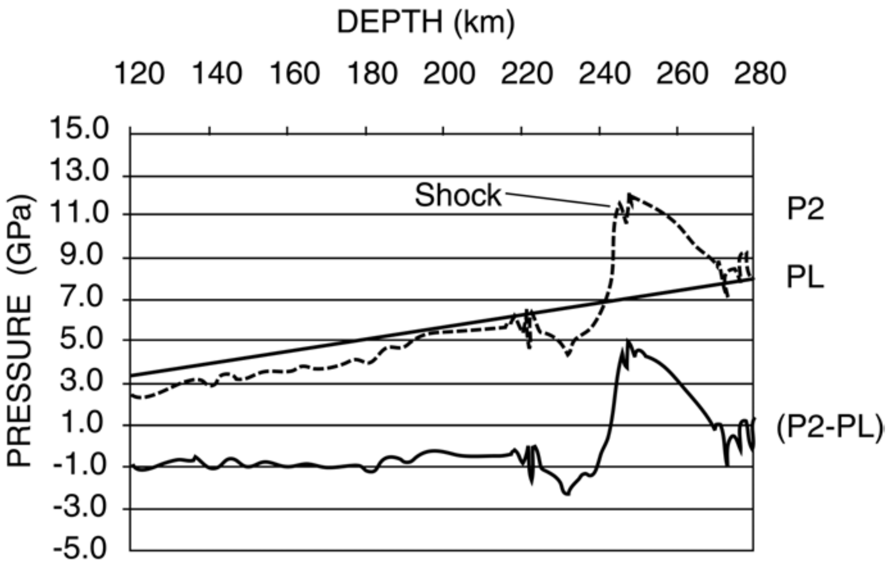

Fig. 4. Pressure versus depth at 40 seconds after impact for indicative model (AUTODYNE-2D). PL = lithostatic load, P2 = impact pressure, (P2-PL) = pressure difference (negative values = decompression). Decompression of ~ -1.0 GPa extends from a depth of ~120 to 180 km, and at lesser values to ~200 km depth, where some melt may exist everywhere in the Earth's mantle (the low velocity zone). The model indicates that decompression melting might be a significant process triggered by large impact craters on Earth, and is expected to be most effective in oceanic lithosphere, where geothermal gradients are high.

melts beneath the impact crater, melt viscosity, foundering of crustal rocks, variations in porosity and permeability in shattered rocks, and explosive interaction with water. Withdrawal of a large volume of melt from the mantle, previously unsupported by, for example, a deep rising conventional plume, could lead to further mass up-flow of the upper mantle during a secondary stage of dynamic flow or collapse into the vacated "space" with resultant further melting

(Price 2001). For simplicity, we therefore assume delivery of only ~30% of the melt to the surface. Thus, our results provide an estimate of ~ 1 x 10^6 km^3 of basaltic melt, comparable to the characteristic volume of LIP's.

There are two main caveats, which we point out about our simulation. Firstly, our simulation uses materials with no inherent strength, and treats the target as a fluidized material. Justification for this is provided by the observation of asthenospheric doming beneath large lunar craters (Neumann et al. 1996), where lunar mantle (thought to be broadly similar to Earth's silicate mantle) may have flowed as a liquid due to shock (Elkins-Tanton, pers. comm., 2002). However, if friction in the model is increased, and the material treated as a cold brittle solid, then the zone of decompression beneath the crater attenuates much more quickly with depth, diminishing the potential for decompression melting. High-friction models may be more appropriate for shallow crustal impacts, where the rocks may fail under shear and tensile loading, but even there, friction during impact is apparently dramatically reduced during impacts to very low levels, perhaps due to acoustic fluidization (Ivanov 1998). Secondly, the time slice we have chosen at 40 seconds coincides approximately with the maximum depth dimension of the transient crater. This time may not represent the state of the mantle after crater formation is completed, although it apparently does for strength-free or friction-free mantle. We expect the full crater to develop in these simulations in about ~200-400 seconds, and we have run decompression volume versus depth profiles as a function of time. Our simulation shows similar results at 60 and 90 seconds, after which the model degraded due to undesirable interaction with the 300 km depth limit.

We have compared our results from AUTODYNE-2D with calculations provided by Boris Ivanov (pers. comm. 2002) using a different hydrocode (SALES; see also Ivanov et al. 1997, Ivanov and Deutsch 1999) and dunite in place of basalt target, but using similar impactor dimensions. He estimates a volume of 2 to 4 x 10^6 km^3 mantle decompressed to >~ 0.1 GPa in a fluidized mantle, which is less than about half the amount in our model. Given the differences in hydrocodes, number of data points, geothermal gradients (he uses continental) and materials, this is actually rather similar. Both models show that the decompressed mantle volume is more than an order of magnitude larger than the total excavated crater volume (~3 x 10^5 km^3). Our initial model has therefore succeeded in demonstrating the potential for melting due to decompression, in contrast to previous impact melt studies which have concentrated on comparing shock heating with geothermal gradients (e.g., Pierazzo et al. 1997; Turtle and Pierazzo 2000). We are currently refining the model towards a more complex (e.g., layered) lithosphere target, including the melt extraction process.

Flux of Impactors

Having shown that a large impact into hot lithosphere could potentially generate large volumes of melt, we need to consider whether the probability of this occurring is large enough to be significant in the Phanerozoic history of the

Earth. These arguments have been well rehearsed in discussion of the striking coincidence of timing between emplacement of flood lavas (LIP's) and at least 5 major extinction events at stratigraphic boundaries throughout the Phanerozoic (Rampino 1987, Rampino and Stothers 1988, Courtillot 1992). Recent calculations imply formation of >450 terrestrial craters of D> 100 km since the late heavy bombardment, and cratering rate estimates solely for oceanic impacts (crater >30km) suggest that a large 200 km crater may occur every 150 Ma, and a 500 km crater every 450 Ma (Glikson 1999; Shoemaker et al. 1990; Koulouris et al. 1999). Examination of the terrestrial impact record over the last ~100 Ma shows that a crater with diameter ~100 km or more has occurred on average once every 35 Ma (Popigai 100 km, 35 Ma; Chesapeake Bay 85 km, 35 Ma; Chicxulub 180 km, 65 Ma). Similar impact rates are inferred independently from studies of comets and for the combined probabilities of comets and asteroids; Weissman (1997) indicated that the impact probability of long period comets large enough to produce craters >10 km is about 1 per million years, and estimates an interval of 1.7×10^7 yrs between potentially catastrophic long period comet impacts. Both comets and asteroids cause impacts, but comets can have much higher velocities. If one assumes that this flux has remained constant since the end of the late heavy bombardment (at ~3.8-4 Ga), then the derived flux is very similar to previous recent estimates (Grady et al. 1998). There are perhaps ~1000 craters of diameter >10 km "missing " from the geological record in the last 3000 Ma. More significantly, the expected number of craters > 200 km diameter is ~25 and there should also be 1 to 5 craters of diameter > 500 km; these have not yet been identified. Our contention is that the larger craters would have been auto-obliterated by impact volcanism, now represented by some LIP's, and that they will appear very different to conventional craters.

Impact Signatures

Oceanic impacts would generally be devoid of mineralogical indicators like shocked quartz, and although the oceanic crust contains other minerals potentially susceptible to shock effects, these are generally dominated by plagioclase feldspar, which transforms to glass (maskelynite) and is unlikely to survive even modest hydrothermal alteration. Potential mineral indicators of oceanic impact derived from the target oceanic crust, could include spinels, perhaps also including nickel-bearing and chromium-bearing varieties (Robin et al. 2000), and spherules. Most mantle minerals (olivine, pyroxenes, garnet) when shocked, transform to metastable phases or glass, or more likely just melt due to their higher initial temperatures (Fig. 5). Any hot minerals and glass would then be susceptible to seawater alteration, to secondary hydrous minerals. Thus, the likelihood of resistant minerals with distinctive shock features to survive over geological time from an oceanic impact, is substantially lower than for a continental impact. Geologically old fractured impacted oceanic terrains might show extensive hydrous mineral development. In general, we agree with the overview presented

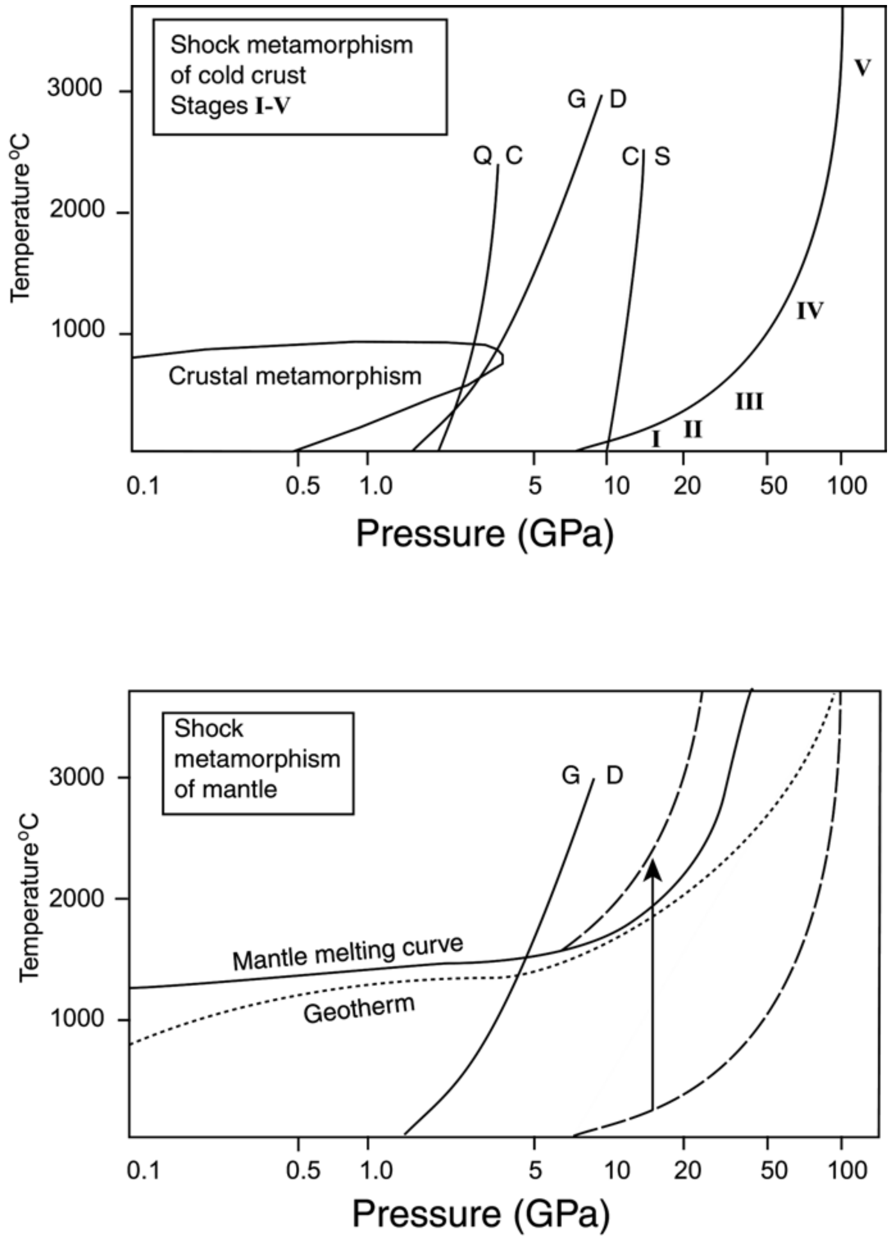

Fig. 5. Shock metamorphism of (a) continental crust with initial low ambient temperature and (b) hypothetical melting of lithosphere/mantle, due to ambient much higher temperature (geotherm). –Mineral equilibrium phase transitions shown for G-D (graphite-diamond; Bovenkerk et al 1959) Q-C (quartz-coesite), and C-S (coesite-stishovite; Fei and Bertka, 1999).

by Elston (1992), that in large impacts, “smoking gun” shock phenomena are likely to be lacking, largely because heat effects overwhelm shock effects.

In the hypothetical scenario of an impact sufficiently large to auto-obliterate, by definition the traditional proximal indicators will be obscured except or until erosion or other geological processes remove the lava deposits. Distal deposits of glassy materials should be lower in silica compared to continental impact glasses. Large craters might still be identified in plan view by circular structures reflecting crustal or tectonic deformation (Price 2001) sometimes mirrored in remote sensing geophysical data, or by radial distribution of the igneous sequence. Thus, the Sudbury Igneous Complex (conventional impact melt) is located centrally within the eroded impact crater, but the circular structure of the buried Chicxulub crater has been determined largely from geophysical methods. Geophysical data for Chicxulub show substantial modifications to the vertical crustal structure. These vary depending on the scale of observation and resolution of data (Morgan et al. 1997), but can include large-scale (50-100 km) regional mantle penetrating faults, low angle faults (possibly associated with melting), displaced and centrally uplifted Moho, and local (~1-10 km) scale displaced and rotated fault blocks. We emphasise that all criteria established for large impacts are restricted to craters in continental crust, which have undergone brittle failure and not penetrated the crust. We have no comparable criteria for a large impact, which punctured oceanic crust and mantle, though this must have occurred. Furthermore, we do not know how the morphology of such a large oceanic impact crater might be further modified through the massive melting event and the transfer of these melts to the surface.

Impact-plumes (I-plumes)?

As for lunar melt extraction (Wilson and Head 2001), we hope in future to model the distribution and extraction of melts from beneath the crater floor. Our indicative model develops saucer-like sub-horizontal sill-like bodies at different depths. This reflects conventional impact melts within craters except that these decompression melts are far below the crater itself. We conjecture that melt extrusion would start with highly energetic eruption of low viscosity peridotitic melts, which would be bouyant compared to solid surrounding lithosphere. Interaction of these hot fluid melts with surface water would be likely to produce ultramafic and mafic pyroclastic rocks (cf. Siberian traps). Extraction of such large volumes of melt could lead to secondary mantle flow at ever decreasing rates due to bulk increasing viscosities with secondary melting, and associated metasomatism. These regions of zoned partially molten mantle represent a massive thermal perturbation resembling a conventional hotspot, and share a number of characteristics with mantle plumes. Such impact-plumes (or "I-plumes") could produce similar magmatic and geochemical signatures, but differ from traditional hot-spot plumes (or "H-plumes") in that; I-plumes neither require pre-magmatic thermal doming (see, e.g., Siberian traps) nor would they be related to a deep geophysical fingerprint. I-plumes may thus offer a possible alternative to H-plumes and are linked to shallow enrichment and depletion events restricted to the upper mantle, as an alternative to the widely perceived involvement of the D" layer at the core-mantle-boundary (e.g., Thompson and Gibson 2000).

Komatiites

The conclusion that high degrees of partial melting or even complete melting of mantle peridotite are possible following a large oceanic impact (Jones et al. 1999) strongly supports an old suggestion that komatiites (MgO > 18 wt%) can be generated by impacts (Green 1972); high-Mg lavas also occur in many LIPs including the Siberian traps. It avoids the problem of storage of high degrees of komatiitic melt and it does not constrain their petrogenesis to either wet or dry varieties. If this view is correct, then komatiite is unlikely to be a unique magma type but instead represents geochemical snapshots of mantle melting, or perhaps mixtures of multiple melting zones (subhorizontal layers in our models). Impact derived decompression melting may have been particularly effective during higher impact fluxes and periods of higher heat flow, as presumably during the early Archaean. Geologically young komatiites occur as spinifex-textured glassy flows of Mesozoic/Tertiary age from Gorgona Island (Gansser et al. 1979; Echeverria 1980; Kerr et al. 1997; Storey et al. 1991), and komatiites of Permian-Triassic age, have recently been described from northwestern Vietnam (Glotov et al. 2001). A feature of the Gorgona komatiites is their preservation of a large volume chaotic to stratified ultramafic breccia (23-27 wt% MgO), with glassy picritic blocks in a fine-grained matrix of plastically deformed high-Mg glassy globules (Echeverria and Aitken 1986). Conventional petrological and geochemical modelling requires a separate magmatic source for the komatiites, compared with associated basalts and picrites. The glassy breccias have been interpreted as evidence for violent submarine eruptions. We postulate instead, that the Gorgona komatiites might have resulted from decompression melting following an oceanic impact, and the ultrabasic breccias record violent interaction between variously melted peridotite and seawater. The classic Barberton komatiite sequence also indicates deep submarine eruption (Dann 2000) and is associated with enigmatic spherule beds with distinctive extraterrestrial Cr isotope ratios providing evidence of at least two major impacts at ~3.24 Ga from projectiles >20 km in diameter (Shukulyukov et al. 2000), suggesting that impacts might be reconsidered (Jones 2002). Lastly, very rapid extraction of komatiite melts formed by decompression partial melting of the deep mantle where diamond is stable, is perhaps the only way to preserve mantle diamonds in some komatiites (Capdevilla et al. 1999).

Candidates for Impact Volcanism

Our indicative model demonstrates the potential for large impact craters (~200 km) to trigger volcanism through decompression melting at any depth extending down to the low velocity zone (~200 km), with volumes of melt comparable to LIP's. The translation of released gravitational energy into melting depends on the geothermal gradient of the target region. Young oceanic lithosphere is most susceptible to this process (geotherm >~17^{o}C/km), but in principle it could happen anywhere, including "cold" continental lithosphere (geotherm ~13^{o}C/km), but with a proportionately larger impactor or higher velocity required. We have not yet determined the minimum size of event to initiate decompression melting, but we take an intuitive guide from the geological record. Since there are no known terrestrial impact craters greater than ~200 km diameter, we conjecture that this may be the lower size limit and larger craters in continental crust have auto-obliterated. Very little is known about oceanic impact craters, but these would require smaller impacts to trigger decompression melting, with the optimum target being an active ridge system with active volcanism before impact. Larger impacts produce more melt in a similar short time, with no upper volume limits; this is in contrast to mantle plumes where melting and melt delivery to the surface is a rate-controlled process related to mantle rheology. Here we present the case for two LIPs, one oceanic and one thin crusted-continental (or oceanic), which might represent impact-generated LIP's. Whether or not they are, remains to be tested.

Ontong Java Plateau

The Ontong Java Plateau is the largest and thickest oceanic plateau on Earth thought to have been formed by the coincidence of two plumes: a major mantle plume or superplume at ~120 Ma and a secondary plume at ~90 Ma (Phinney et al. 1999). It is not associated with major global mass extinctions (Coffin and Eldholm 1994; Wignall 2001). Geophysical data shows much greater and irregular crustal thickness (15 – 38 km) compared with normal oceanic crust (6-10 km) and a low velocity seismic "root" extending down to 300 km (Richardson et al. 2000). However the unexpectedly small subsidence history of the OJP lead Ito and Clift (1998) to rule out cooling of a large plume head; instead they suggested substantial magmatic underplating. Remnant surrounding seafloor magnetic anomalies show that the OJP formed in young oceanic crust perhaps only 10 Ma old, and may have formed very close to an active spreading ridge (Gladczenko et al. 1997). These fundamental indicators are sufficiently close to our model conditions (maximum melting in young oceanic lithosphere) that we suggest a large oceanic impact at around ~120 Ma, could have triggered this LIP; further details of this candidate for impact volcanism and the large scale effect of the impact on plate motions are presented elsewhere (Price 2001). In this case, the impact site is now represented by a massive layer of volcanic rock, which forms the oceanic plateau itself.

Siberian Traps

The Siberian Traps represents the single largest eruption of “continental” flood lavas. A somewhat larger impact would be required for our model to operate in continental crust. However, recent plate tectonic reconstructions constrained by seismic tomography indicate that Siberia may actually have been an oceanic environment with micro-continents and subduction zones (Van der Voo et al 1999). The lavas are dated at the end of the Permian (e.g., Campbell et al. 1992; see also Reichow et al 2002), where a double extinction event may have occurred (Wignall 2001). Up to one third of the lower succession is represented by pyroclastic rocks, with individual tuff units covering up to 30,000 km^2; it was initially marine and developed in a massive subsiding basin that rules out a conventional mantle plume (Czamanske et al. 1998). Elkins-Tanton and Hager (2000) endorsed Sharma’s view (1997) that the Siberian Traps cannot be the result of a traditional form of mantle plume. There is some independent global evidence that an impact occurred at the P-Tr boundary, although the evidence is by no means as convincing as for the K/T boundary. A weak Ir-anomaly together with possible shocked quartz were found both in Antarctica and Australia (Retallack et al. 1998). Chinese strata at Meishan placed the boundary at 251.4 +/- 0.3 Ma and record rapid addition of isotopically light carbon over a time interval of 165,000 years, or less (Bowring et al. 1998), but problems with dating at this site have emerged (Mundil et al. 2001). Investigation of the marine faunal extinction including the same Meishan outcrops, lead Jin et al. (2000) to conclude that “a predicted true extinction level [occurred] near 251.3 Ma (94% of genera are included in a 0.1-Ma interval spacing). A more reasonable conclusion...is a sudden extinction at 251.4 Ma, followed by the gradual disappearance of a small number of surviving genera over the next 1 million years”. An impact event is also supported by controversial evidence from extraterrestrial noble gases in fullerenes recovered from P-Tr boundary beds in China, Japan and Hungary (Becker et al. 2001), although the reliability of such techniques is seriously questioned (I. Gilmour, pers. comm., 2001; Farley and Mukhopadhyay 2001). Although the evidence for impact at the P/Tr boundary is much less clear than for bolide impact at the K-T boundary (Alvarez et al. 1980), there is a similar duality of signals between likely volcanic and impact sources. Therefore, it would seem important to test our hypothesis that the Siberian Traps could have been caused by decompression melting at the impact site, and that impact volcanism can uniquely explain the dual signals in the geological record. The geological record may be consistent with this idea, but we are not aware of any literature concerning the critical volcanic-sedimentary interface at the base of the Siberian traps. However, the onset of volcanism is everywhere an unconformity marked by tuffs uniformly above folded and variably missing palaeozoic strata (Czamanske et al. 1998). The thickest volcanic sequence is in the northern part (4,000 metres, Maymecha-Kotuy; 3,500 metres Norilsk) where massive Ni-sulphide mineralisation is related to mantle-dissecting faults (Hawkesworth et al. 1995). The large-scale occurrence of native nickel- iron (Oleynikov et al. 1985) in intrusive rocks related to the extrusive lavas, (including Pt-bearing nickel-rich iron; Ryabov and Anoshin

1999), is consistent with impact geochemical models that predict native iron and nickel iron (Gerasimov et al. 2001; Miura et al. 2001), and is reminiscent of native iron at the base of the flood lavas in west Greenland (Klöck et al. 1986). Also, the regional geology of the wider Siberian craton and bounding mountain fold belts (Baikal, Verkhoyansky, Taymyr) should be reconsidered in terms of the possible major plate tectonic effects of an impact, as confirmed by changing plate vectors at 250 Ma (Price 2001). The large-scale foundering of continental Siberian lithosphere at this time, recently proposed on the basis of geophysical data (Elkins-Tanton and Hager 2000) is consistent with our impact volcanism hypothesis. The recent recognition that the Siberian traps may have been double the volume than previously assumed, extending west as far as the Urals (Reichow et al. 2002), is easily accommodated in an impact volcanism model by relatively small changes in impactor parameters. The end-Permian event is complicated by the possible double epicentre implications required to produce the slightly older Emeishan flood-lava province in south China (Lo et al. 2002), which, if the dating is reliable (Mundil et al. 2001), were erupted a few million years earlier in a "marine" environment. This is not a problem for an impact volcanism explanation, simply requiring two impacts (Shoemaker-Levy 9 showed us that multiple impacts can occur; furthermore about 10% of known terrestrial craters >20 km are pairs, similar to the recent prediction (16%) that many near-Earth orbiting asteroids are double systems; Margot et al 2002); however, it may require extraordinary pleading to explain two separate mantle superplumes. The Emeishan traps basal ash layers are characterized by concentrations of microspherules, whose origin is not fully understood (Yin et al. 1992) and earlier thought to have derived from the Siberian traps >2000 km away (Cambell et al. 1992). On the basis of exotic "impact metamorphosed" metallic Fe-Ni grains with up to 30% Ni (Kaiho et al. 2002) within the spherules (Miura et al. 2001) and an absence of shocked quartz, it has been suggested that an oceanic impact was the source of the Emeishan volcanism (Kaiho et al. 2001), but this work has been strongly criticised as being inconclusive (Koeberl et al. 2002). If subsequent investigations can demonstrate that the exotic grains are extraterrestrial (as for the K/T boundary) this would be the first direct evidence for impact at the base of the Emeishan traps, and would dramatically strengthen the claims of Kaiho et al. (2001) that the volcanism was triggered by an oceanic impact, as predicted by our model.

Discussion and Conclusions

Our indicative model shows that it is possible for the volume of decompressed mantle beneath a large ~200 km sized crater to greatly exceed the excavated volume of the impact crater itself, primarily due to reduction of lithostatic load. Under suitable conditions of geothermal gradient, this would lead to near instantaneous melting with volumes of the order of 10^6 km^3, similar to the characteristic volumes of LIP's. Optimum target conditions are represented by young oceanic lithosphere, close to or at an active ridge system and could be triggered by a smaller impact; the same process can operate in continental targets, perhaps requiring a somewhat larger impact depending on geothermal gradient and crust/lithosphere architecture. Our model ~200 km impact crater is formed by an initial transient crater, ~80-100 km deep, much deeper than the total crust, whether it is oceanic (~10 km) or continental (~30 km). The melting would take place under the entire crater, deep in the upper mantle where garnet is stable, and can extend down to the zone of stable diamond and the low velocity zone (~200 km). Initial melting may occur at various depths as sub-horizontal, saucer- or sill-like bodies, suggesting that mixing of melts from different depths (reservoirs) would be possible during the melt extraction process (volcanism). By comparison with conventional plume models, this would instantaneously trigger massive volcanism, with geochemical signatures dominated by a garnet-peridotite source mantle, and possible mixing of geochemical reservoirs.

The resultant thermal anomaly in the mantle could be long-lived, and the induced large-scale vertical and horizontal thermal gradients are expected to have a long-term effect on secondary mantle flow, leading to secondary mantle melting which may also be voluminous (see "Impact plumes" above). A secondary pulse of melting, from longer-term asthenospheric flow is currently being investigated by a group at MIT to reinvestigate the origin of lunar mare as post impact melts (Elkins-Tanton et al. 2002). Although this secondary melting is unlikely to approach similar volumes to the intial decompression melting, such adiabatic melting in convection currents nonetheless offers an attractive mechanism for sustaining volcanic activity at the impact site for up to 10 million years after the initial impact (Elkins-Tanton, pers. comm., 2002). We have demonstrated that the previously suggested but generally dismissed mechanism of pressure-release melting should indeed enable large impacts to generate excess volumes of mantle melt. The flux of large impactors expected is sufficient to explain many of the large igneous provinces seen on Earth, and might generally be considered where conventional plume explanations are untenable. As a result, this decompression melt may contribute more melt than conventional shock melting, and the cumulative melt volume may not scale linearly with crater dimension. We have presented arguments to support LIP candidates as impact volcanism-derived. Thus, the Ontong Java Plateau is a promising candidate for the auto-obliterated site of an oceanic impact crater, but has not yet been studied in sufficient detail, largely due to the inherent problems of studying submarine plateaus. We propose that the Siberian Traps, which are accessible and currently under considerable scrutiny, may be better explained by a large impact than by a conventional mantle plume. The closure of a former ocean between Siberia and Mongolia, as well as amalgamation with north and south China blocks may also have been occurring during Permian-Triassic times, and the impact target region may have been oceanic, with a mixture of micro-continents and subduction zones (Van der Voo et al 1999).; this would be much easier to fit with our model. If the end-Permian extinction requires two events separated geographically by 2-3000 km (Siberian traps, Emeishan traps China) this is no problem for impact models. A Siberian impact could explain, for example, the lack of thermal doming, their extreme osmium isotope geochemistry (Walker et al. 1997), and also the occurrence of cliftonite-bearing (cubic graphite) metallic nickel-iron in the intrusive traps

(Oleynikov et al. 1985) as metamorphosed relics or products of meteoritic iron (as at Meteor Crater; Brett and Higgins 1967). Impact volcanism must have obscured the evidence for the original impact crater, and may at least partly explain why recognition of a global impact anomaly at the P/T boundary has been so difficult. The current day Moho topography beneath Siberia is variable but segmented, and has been interpreted as a series of mantle ridges and rifts (Kravchenko et al. 1997); seismic velocity structure shows a continuous substantial lateral velocity inversion (8.0 versus 8.4 km/s) at ~100 km depth underlying the entire Siberian platform (Mooney 1999). All of these are consistent with being relict impact features, albeit on a larger scale than commonly observed.

Our estimate of the amount of meteoritic material added to the Earth by large impacts since the end of the late heavy bombardment using cratering rate models is $\sim 10^9$ km^3. This represents only about 1% of the volume of the Earths crust, but could, for example, account for the entire PGE budget of the crust, and agrees with enrichment after core segregation supported by recent experimental models (Holzheid et al. 2000). Some of the largest impact craters in continental crust are associated with economic mineralisation, such as the nickel-rich massive sulphides at Sudbury (<200 km) (e.g., Molnar et al. 1999). The Witwatersrand gold deposits are concentrically zoned around the Vredefort structure and may be related to the acknowledged impact-driven hydrothermal activity (Gibson and Reimold 2000). Large impacts are expected to propagate significant hydrothermal activity, aided by intense rock fracturing and the thermal energy deposited in the crust (Kring 2000); they are potential mineral exploration targets, both terrestrial and extraterrestrial (Norman 1994).

There may be additional mantle signatures related to impact, such as the energy to drive large-scale mantle metasomatism and mineralisation. We have suggested that because of the combined effects of decompression and impact heating, complete melting of the mantle, and high degree partial melts are easily achievable, consistent with the production of ultrabasic melts like komatiites from large impacts. Komatiite Ni-PGE-sulphide ore systems typically have high Os concentrations, low Re/Os ratios, and near-chondritic Os isotope compositions, from which Lambert et al. (1998) concluded that large scale dynamic processes, including major lithospheric pathways, are critical to the development of these massive magmatic systems. The long-term effects of sustained melt extraction might result in rootless mantle hotspots, or impact plumes, which will require further modelling. The global consequences for plate tectonics throughout Earth history have recently been explored further by one of us (Price 2001).

We have concentrated on perhaps the most potent melting process in the Earth's mantle, specifically that triggered by decompression beneath a large impact crater, and agree with Boslough et al. (1986), who stated "the impact-produced flood basalt hypothesis is attractive because it is potentially testable on the basis of predictions of features that have not yet been discovered...unlike current plume models for flood basalts and hotspots". In conclusion, we assert that the concept of impact-induced volcanism has not been adequately examined and may offer a new framework for the interpretation of large-scale igneous and geological processes.

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