EARLY ARCHEAN ASTEROID IMPACTS ON EARTH:
STRATIGRAPHIC AND ISOTOPIC AGE CORRELATIONS
AND POSSIBLE GEODYNAMIC CONSEQUENCES

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8.5-1. INTRODUCTION

The heavily cratered surfaces of the terrestrial planets and moons testify to their long-term reshaping by asteroid and comet impacts and related structural and melting processes. Following accretion of Earth from the solar disc at ca. 4.56 Ga (see Taylor, this volume), the Earth–Moon system is believed to have originated by a collision between a Mars-size planet and Earth, followed by episodic bombardment by asteroids and comets, with a peak documented at ca. 3.95–3.85 Ga – the Late Heavy Bombardment (LHB) – recorded on the Moon, but not on Earth due to the paucity and high grade metamorphic state of terrestrial rocks of this age (Wilhelm, 1987; Ryder, 1990, 1991, 1997; see Iizuka et al., this volume). Major impact episodes on Earth are recorded from rocks dated at 3.46, 3.26–3.24, 2.63, 2.56, 2.48, 2.02 Ga (Vredefort), 1.85 Ga (Sudbury), 0.58 Ga (Acraman), 0.368–0.359 Ga (late Devonian), 0.214 Ga (late Triassic), 0.145–0.142 Ga (late Jurassic), 0.065 Ga (K–T boundary), and 0.0357 Ga (late-Eocene) (World Impact List GSC/UNB; Grieve and Shoemaker, 1994; Grieve and Pesonen, 1996; Grieve and Pilkington, 1996; French, 1998; Simonson and Glass, 2004). Some of these impact events coincided with, and are referred to as the direct cause of, mass extinctions and/or the radiation of species (Alvarez, 1986; Keller, 2005; Glikson, 2005a). These impacts represent a small part of the extraterrestrial impact record, as the retention of any impact-generated structure is a direct function of the preservation and/or destruction of crust. By analogy, the continuous re-accretion of sulphur on Io and ice on Europa obliterates their impact records, as does volcanic re-surfacing on Venus and plate tectonics on Earth. Estimates of the impact incidence on Earth is inferred from the lunar cratering record, and the current asteroid and comet flux, which suggests the formation of some 100–200 craters of diameter (D) D ≥ 100 km, and 5–10 craters of D ≥ 300 km since ca. 3.8 Ga (Neukum et al., 2001; Ivanov and Melosh, 2003; Glikson et al., 2004). Most of these craters would have impacted on hitherto subducted oceanic crust on Earth. The larger continental impacts are less likely to be destroyed by subduction, and resulted in deeper excavation of the continental crust and thicker ejecta layers, thereby being generally better preserved than their counterparts.
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that impacted oceanic crust. Glikson (2001) estimated that only between 8–17% of the larger impacts \( (D \geq 100 \text{ km}) \) is recorded as impact structures and/or ejecta. However, assuming a relatively smaller proportion of large projectiles, over 50% of the larger impacts have been recorded at this stage (Ivanov and Melosh, in Glikson et al., 2004).

Models suggest impact triggering of igneous, tectonic, or plate tectonic events (Green, 1972, 1981; Grieve, 1980; Alt et al., 1988; Jones et al., 2002; Elkins-Tanton et al., 2005; Glikson and Vickers, 2005), but this correlation has not been directly demonstrated on Earth. Ivanov and Melosh (2003) questioned impact-induced volcanic activity, whereas Glikson et al. (2004) state: ‘Indeed, a few giant impact events certainly occurred on Earth in post-LHB history, and a few might have encountered the fortuitous circumstances necessary for them to enhance magma production. This is an important topic for future study.’ Lowe et al. (1986, 1989) remarked on the potential tectonic significance of the co-occurrence between 3.26–3.24 Ga impact horizons recorded in the Barberton Greenstone Belt (Kaapvaal Craton) and the abrupt the break between the underlying mafic–ultramafic volcanic crust (3.55–3.26 Ga Onverwacht Group) and the overlying turbidite, felsic volcanic, banded iron formation and conglomerate-dominated succession (3.26–3.225 Ga Fig Tree and Moodies Groups), which includes the earliest observed granite-derived detritus (Fig. 8.5-1). This age period correlates broadly with a similar break in the Pilbara Craton, Western Australia, where the mafic–ultramafic volcanic rocks of the ca. 3.255–3.235 Ga Sulphur Springs Group are unconformably overlain by Soanesville Group clastic sedimentary rocks, including a local unit of olistostrome breccia (Glikson and Vickers, 2005; Van Kranendonk et al., 2006a). Widespread 3.27–3.24 Ga plutonic igneous rocks (dominantly granitic) occur in both the Kaapvaal Craton and Pilbara Craton (see Poujol, this volume, and Van Kranendonk et al., 2007a, this volume).

8.5-2. 3.47 GA IMPACT EVENTS

Microkrystite spherule-bearing lenses in chert and in the matrix of diamicite have been reported from the 3.47 Ga Antarctic Creek Member (ACM) of the Mount Ada Basalt, in the North Pole Dome area of the East Pilbara Terrane, Pilbara Craton, Western Australia (Lowe et al., 1989; Byerly et al., 2002; Van Kranendonk et al., 2006a). The diamicite, defined as ACM-S2, consists of a 0.6–0.8 m thick unit of spherule-bearing, chert-intraclast conglomerate, which is separated from the main ACM-S3 unit by dolerite and felsic volcanics/hypabyssals. Zircon dating from the spherule-bearing sandstone unit yielded an age of 3470 ± 2 Ma (Byerly et al., 2002). The microkrystite spherules are discriminated from angular to subangular detrital volcanic fragments by their high sphericities, inward-radiating fans of sericite, relic quench textures, and Ni–Cr–Co relations. SEM-EDS (scanning electron microscope coupled with electron-dispersive spectrometer) and LA-ICP-MS (laser induced coupled plasma mass spectrometer) analysis indicate high Ni and Cr in sericite-dominated spherules, suggesting a mafic composition of source crust. Ni/Cr and Ni/Co ratios of the spherules are higher than in associated Archaean tholeiitic basalts and high-Mg basalts, rendering possible contamination by high Ni/Cr and Ni/Co...
Fig. 8.5-1. Composite columnar section of the Barberton greenstone belt stratigraphy, showing the principal groups and formations and U-Pb zircon isotopic age determinations (after Byerly and Lowe, 1994; Byerly et al., 1999; Lowe et al., 2003).
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Chondritic components. The presence of multiple bands and lenses of spherules within chert, and scattered spherules in arenite bands within ACM-S3, may signify redeposition of a single impact fallout unit or, alternatively, multiple impacts. Controlling parameters include: (1) spherule atmospheric residence time; (2) precipitation rates of colloidal silica; (3) solidification rates of colloidal silica; (4) arenite and spherule redeposition rates; and (5) arrival of an impact-generated tsunami. The presence of spherule-bearing chert fragments in ACM-S3 may hint at an older spherule-bearing chert (?S1). Only a minor proportion of spherules is broken and the near-perfect sphericities of chert-hosted spherules and arenite-hosted spherules constrain the extent of shallow water winnowing of the originally delicate glass spherules. It is suggested that the spherules were either protected by rapid burial or, alternatively, disturbance was limited to a short term high energy perturbation such as may have been affected by a deep-amplitude, impact-triggered tsunami wave.

An impact spherule unit precisely correlated with spherules of the ACM occurs in a 30 cm to 3 m thick chert unit, H4c in the upper part of the Hooggenoeg Formation of the Onverwacht Group, Kaapvaal Craton (Lowe et al., 2003). Zircons in this unit have been dated at 3470 ± 6.3 Ma, identical in age to the Pilbara unit (Byerly et al., 2002). The spherules constitute a bed of medium- to coarse-grained, current-deposited sandstone, 10–35 cm thick and interbedded with fine-grained tuffs and black, or black-and-white, banded cherts. In sections where H4c is ~100 cm thick, the unit generally rests directly on, or just a few centimeters above, altered basaltic to komatiitic volcanic rocks. In the type section, H4c is 2.8 m thick, but varies from ~50 cm to 6 m thick within 500 m along strike. In this section, the unit is underlain by up to 160 cm, and overlain by 90 cm, of pale greenish to light gray sericitic chert and black carbonaceous chert.

8.5.3. THE 3.26–3.24 GA BARBERTON IMPACT CLUSTER

The uppermost, predominantly mafic–ultramafic volcanic sequence of the Onverwacht Group consists of an assemblage of komatiitic volcanics and their hypabyssal and altered equivalents, capped by ferruginous chert. This unit is known as the Mendon Formation (Fig. 8.5-1) and has been dated by U-Pb zircon from a middle chert unit as 3298 ± 3 Ma (Byerly, 1999). Unconformably overlying the Mendon Formation is the Mapepe Formation, the basal unit of the Fig Tree Group, consisting of a turbidite–felsic volcanic–ferruginous sedimentary rock association dated in the range of 3258±3 Ma to 3225±3 Ma (Lowe et al., 2003) (Fig. 8.5-1). The S2 impact fallout unit coincides with the base of the clastic Mapepe Formation of the Fig Tree Group, whereas the S3 and S4 impact fallout units occur within clastic sedimentary rocks and felsic pyroclastic rocks, 110–120 m above the basal contact. In places, the S3 unit occurs unconformably above deeply incised Mendon Formation komatiites.

Detailed documentation of the S2, S3 and S4 impact fallout units (Fig. 8.5-1) reveals a wide range of diagnostic field, petrographic, mineralogical, geochemical and isotopic criteria identifying extraterrestrial components mixed with volcanic and other detritus (Lowe and Byerly, 1986b; Lowe et al., 1989, 2003; Kyte et al., 1992, 2003; Byerly and Lowe, 2003).
8.5-3. The 3.26–3.24 Ga Barberton Impact Cluster

Fig. 8.5-2.
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Principal criteria include: (1) impact fallout units that can be correlated in below-wave base environments between basins and sub-basins, independently of facies variations of the host sedimentary rocks (Lowe et al., 1989, 2003); (2) microkrystite spherules that have inward-radiating quench textures and centrally offset vesicles (Glass and Burns, 1988) (Fig. 8.5-2), distinct from outward radiating textures of volcanic varioles, which may also contain microphenocrysts of feldspar and quartz (Glikson, 2005b, 2005c); (3) microkrystites that contain quench-textured and octahedral Ni-chromites (NiO < 24%), including a high Co, Zn and V variety unknown in terrestrial chromites (Byerly and Lowe, 1994); and (4) microkrystites that contain nickel and PGE nanonuggets (Kyte et al., 1992; Glikson and Allen, 2004; Glikson, 2005b). The Ni-rich chromites are distinct from terrestrial type Ni-chromites associated with sulphide (NiO < 0.3%) (Czamanske et al., 1976; Grove et al., 1977), and have low FeO/Fe2O3 ratios, possibly consequent on condensation under low-oxygen Archean atmospheric conditions.

Attempts at estimating the magnitude of the impact events and the composition of the target crust are based on inferences derived from spherule radii (O’Keefe and O’Hara, 1982; Melosh and Vickery, 1991) and geochemical mass balance calculations of asteroid size, based on PGE (Kyte et al., 1992; Byerly and Lowe, 1994) and 53Cr/52Cr isotope ratios (Shukloyukov et al., 2000; Kyte et al., 2003; Glikson and Allen, 2004). Early estimates based on spherule size distribution (mean 0.85 mm), and the iridium flux (~3.24 Ga S3 impact fallout unit, Mapepe Formation, Barberton greenstone belt. The spherules are dominated by chlorite, displaying inward radiating fans, central to offset vesicles and rims of microcrystalline quartz and are set in a micro-fragmental matrix containing angular clasts of volcanic rock and chert. Plane polarized light. A – spherule with chlorite rim, microcrystalline mantle and siliceous core; B – chlorite spherule with central siliceous vesicle, showing LA-ICPMS burn traces; C – spherule consisting of inward-radiating chlorite fans, interpreted as replacing quench textures of original pyroxene and olivine; D – detail of C; E – Spherule occupied by radiating palimpsest needles after ?olivine and/or pyroxene representing quench texture; F – spherule with inward-radiating chlorite sheafs and a central quartz-filled vesicle; G – spherule with inward-radiating ferruginous chlorite fans, light-coloured chlorite mantle and siliceous core; H – fragment of spinifex-textured komatiite.

Fig. 8.5-2. (Continued.) Microphotograph of microkrystite spherules in sample SA306-1 (A–D) and SA315-5 (E–H) from the ~3.24 Ga S3 impact fallout unit, Mapepe Formation, Barberton greenstone belt. The spherules are dominated by chloride, displaying inward radiating fans, central to offset vesicles and rims of microcrystalline quartz and are set in a micro-fragmental matrix containing angular clasts of volcanic rock and chert. Plane polarized light. A – spherule with chloride rim, microcrystalline mantle and siliceous core; B – chlorite spherule with central siliceous vesicle, showing LA-ICPMS burn traces; C – spherule consisting of inward-radiating chlorite fans, interpreted as replacing quench textures of original pyroxene and olivine; D – detail of C; E – Spherule occupied by radiating palimpsest needles after ?olivine and/or pyroxene representing quench texture; F – spherule with inward-radiating chlorite sheafs and a central quartz-filled vesicle; G – spherule with inward-radiating ferruginous chlorite fans, light-coloured chlorite mantle and siliceous core; H – fragment of spinifex-textured komatiite.
8.5-3. The 3.26–3.24 Ga Barberton Impact Cluster

Fig. 8.5-2. (Continued.)

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Fig. 8.5-3. Correlation between the flux of iridium (in units of \(10^{-4}\) mg cm\(^{-2}\)) and the diameter of chondritic projectile (Dp), based on mass balance calculations assuming mean unit thickness, mean Ir concentration and global distribution of fallout ejecta (after Glikson, 2005). Unit symbols - \(R_p\) (projectile radius) = \(\sqrt[3]{V_p/[(4/3)\pi]}\); \(V_p = M_p/d_p\); \(M_p = F_G/C_p\); \(F_G = (A_E/D_S \ast T_S)\).

\((A_E\) – measured element (E) abundance in fallout unit (ppb); \(D_S\) – mean specific density of fallout sediments (mg/cm\(^3\)) (assumed as 2.65 gr/cm\(^3\)); \(C_E\) – weight of element E in mg per cm\(^3\); \((C_E = A_E \ast D_S); T_S\) – mean stratigraphic thickness of spherule unit (in mm); \(F_E\) – local mean flux of element E in mg per 100 mm\(^2\) (cm\(^2\)) surface; \(F_G = A_E/D_S \ast T_S \ast \text{Earth surface area (}S_E\}; \(C_p\) – assumed concentration of element E in projectile (ppb) (\(C1\) – Ir-450 ppb; \(M_p\) – weight of projectile \((M_p = F_G/C_p); d_p\) – assumed specific density of the projectile (\(C1\) – 3.0 gr/cm\(^3\)); \(V_p\) – volume of projectile \((V_p = M_p/d_p)\).

A FeNi impactor would be 4–10 times smaller, but inconsistent with the maximum spherule sizes. From a bimodal size distribution of spherules in the SA306-1 sample, with a diameter range of 0.4–2.0 mm, Byerly and Lowe (1994) suggested early flux of small, Ir-rich, Ni spinel-bearing spherules (mean diameter \(\sim 0.65\) mm) and a late flux of larger, mostly Ir-poor, spinel-free spherules (mean diameter \(\sim 1.25\) mm), suggesting early ejection of projectile-rich components and a projectile 24 km in diameter. These authors showed the SA306-1 impact spinels have higher Ni/Fe and lower Fe\(^{+3}/\)Fe\(^{+2}\) than Phanerozoic impact spinels, hinting at a less oxidizing Archaean atmosphere. From Cr isotope-based estimates of the global thickness and mean Ir concentration of impact units \((S_2; \sim 20\) cm, 3 ppb; \(S_3; \sim 20\) cm, 300 ppb; \(S_4; \sim 10\) cm, 100 ppb), Kyte et al. (2003) suggested a parental asteroid 50–300 times the mass and 3–7 times the diameter of the \(\sim 10\) km-large K–T Chicxulub asteroid, consistent with estimate made by Glikson (2005b: Fig. 8.5-3).
Given projectile/crater diameter ratios in the range of 0.05–0.1, these impacts would result in impact basins at least 300–700 km large. The dominantly mafic to ultramafic composition of the impact ejecta requires that these basins were formed in mafic–ultramafic-dominated regions of the Archaean Earth, probably oceanic crust. Impacts of this magnitude can be expected to have had major consequences on the impacted crustal plates, greenstone–granite terrains, and underlying mantle regions.

Lowe et al. (1989) observed: ‘The transition from the 300-million-year-long Onverwacht stage of predominantly basaltic and komatiitic volcanism to the late orogenic stage of greenstone belt evolution suggests that regional and possibly global tectonic reorganization resulted from these large impacts.’ Evidence consistent with impact-induced effects in the Kaapvaal Craton is provided by U-Pb zircon ages of the Nelshoogte trondhjemite (3236 ± 1 Ma) and the Kaap Valley tonalite (3227 ± 1 Ma), suggesting plutonic emplacement within about ∼5–10 My after the S3 and S4 impacts (3243 ± 4 Ma). The tonalite-trondhjemite geochemistry of the granitoids suggests partial melting of mafic crustal sources.

Close temporal correlations are observed between the Onverwacht Group – Fig Tree Group transition and the transition between the Kelly Group (end of felsic magmatism at ca. 3290 Ma) and the 3255–3235 Ma Sulphur Springs Group in the East Pilbara Terrane of the Pilbara Craton (Fig. 8.5-4). In the Sulphur Springs Group, there is a change from ∼3255–3235 Ma ultramafic–felsic volcanic rocks and silicified epiclastic rocks of the Sulphur Springs Group and deposition of overlying sedimentary rocks (banded iron formation and turbidites) of the Soanesville Group (Fig. 8.5-5). Eruption of the Sulphur Springs Group has been variously interpreted in terms of back arc rifting (Brauhart, 1999), or mantle plume events, the latter including local caldera subsidence above a syn-volcanic laccolith (Van Kranendonk, 2000; Van Kranendonk et al., 2002; Pirajno and Van Kranendonk, 2005; Smithies et al., 2005b). Whereas in places the boundary above the marker chert (silicified epiclastic rocks) at the top of the Sulphur Springs Group appears conformable, it was clearly lithified prior to deposition of the overlying, local unit of olistostrome breccia (Hill, 1997) and more widespread turbiditic sandstones of the overlying Soanesville Group (Van Kranendonk et al., 2006a).

The following stratigraphic and age relations are outlined between the Barberton Greenstone Belt (BGB) and greenstone belts of the East Pilbara Terrane (PGB) (Fig. 8.5-4):

(A) BGB – S2 impact spherule unit directly underlying 3258 ± 3 Ma felsic tuff. PGB – sedimentary rocks of the Leilira Formation (ca. 3255 ± 4 Ma: Buick et al., 2002) at the base of the Sulphur Springs Group, but no impact spherules identified.
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Fig. 8.5-4. Isotopic age correlations between 3.28–3.22 Ga units in the Kaapvaal Craton and Pilbara Craton. Solid squares and error bar lines – U-Pb ages of volcanic and plutonic units; stars – impact ejecta layers; circled crosses – ferruginous sediments; MF – Mapepe Formation; UF – Ulundi Formation; NT – Nelshoogte tonalite; KPG – Kaap Valley granite; LF – Leilira Formation; KCF – Kangaroo Cave Formation; EPG – Eastern Pilbara granites; NRF – Nickol River Formation; KR – Karratha Granodiorite.

(B) BGB – the S3 and S4 impact spherule units of the Mapepe Formation, Fig Tree Group, directly overlie 3243 ± 4 Ma felsic tuff. PGB – 3235 ± 3 Ma felsic volcanic rocks in the Kangaroo Caves Formation, Sulphur Springs Group, but no impact spherules identified.

In the West Pilbara Superterrane of the Pilbara Craton, a similar lithological change is seen in the Roebourne Group, where komatiite and basalt of the >3270 Ma Ruth Well Formation are overlain by 3.27–3.25 Ga sandstone, shale, banded iron-formation, and felsic volcanic rocks of the Nickol River Formation, and accompanied by emplacement of the 3.27–3.26 Ga Karratha Granodiorite. Indeed, major plutonic activity at ~3.275–3.225 Ga (Fig. 8.5-4), defined as the Cleland Supersuite, is documented throughout much of the Pilbara Craton (Van Kranendonk et al., 2006).

The S2 impact ejecta unit in the south-eastern part of the Barberton greenstone belt is overlain directly by 20–30 m thick banded ferruginous chert, the Manzimnyama Jasplilite Member, of the basal Mapepe Formation (Lowe and Nocita, 1999). These widespread sedi-
8.5-4. Stratigraphic and Isotopic Age Correlations Between the 3.26–3.24 Ga Impacts

Fig. 8.5-5. Schematic section through the top of the Sulphur Springs Group (Kangaroo Caves Formation – 3235 ± 3 Ma), overlying olistostome, siltstone, ferruginous siltstone and felsic volcanics of the Pincunah Hill Formation (after Hill, 1997).
Figs. 8.5-6. Large blocks of felsic volcanics, chert and siltstone in the olistostrome, basal Pincunah Hill Formation, Sulphur Springs, Central Pilbara (after Hill, 1997). Pressure-temperature diagram portraying the position of the lithosphere-asthenosphere boundary for continental and oceanic geotherms and the position of dry and wet peridotite solidi. Downward arrow marks the ~30 km penetration depth of a 20 km asteroid, equivalent to decompression and rebound. Upward arrows mark alternative adiabatic melting paths of mantle diapirs (adiabatic cooling 0.33°C/km).
imentary rocks in the southern part of the Barberton greenstone belt include oxide facies BIF, jaspilite, and hematite-rich shale. The $S_3$ impact ejecta unit in the northern part of the Barberton greenstone belt underlies ferruginous sediments of the Ulundi Formation, including jaspilite, ferruginous chert and black chert deposited under quiet, deep-water conditions (Lowe et al., 2003). By contrast, deposition of $S_2$ and $S_3$ impact units along the Onverwacht anticline, which formed an antecedent rise located between deeper basins to the northwest and southeast, occurred under high-energy, shallow water conditions (Lowe et al., 2003).

In the Pilbara Craton, uppermost stratigraphic levels of the felsic volcanic Kangaroo Caves Formation contain BIF, as rafts in olistostrome and as a more widespread unit, up to 1000 meters thick (Van Kranendonk, 2000). The sequence is disconformably overlain by arenites and turbidites and ferruginous shale and silicified equivalents up to 100 m thick in some areas. The potential significance of a relationship between impacts and BIF is underpinned by similar observations in 3.46, 2.63 and 2.56 Ga impact units (Table 8.5-1; Glikson, 2006).

**8.5-5. POSSIBLE GEODYNAMIC CONSEQUENCES**

For a present-day mid-ocean geotherm ($25^\circ C \text{km}^{-1}$), assuming yet higher Archean geothermal gradients and possibly smaller-scale convection cells and plate dimensions (Lambert, 1983), thin ($<5$ km) crustal regions underlain by shallow asthenosphere ($<50$ km) can be expected to have occupied perhaps a quarter of the Archean oceanic crust. Low angle impacts in deep-water ocean will be partly to largely absorbed by the water column. A high-angle impact by a 20 km diameter projectile on thin oceanic crust underlying $<2$ km water depth would result in a $300 \pm 100$ km diameter submarine crater. As originally modeled by Green (1972, 1981), downward compression of the transient crater followed by rebound of asthenosphere originally located at $\sim 40–50$ km would result in intersection of the peridotite solidus by rising mantle diapers and in partial melting. Insofar as morphometric estimates after Grieve and Pilkington (1996) may apply to asthenospheric uplift, the amount of rebound may be similar to that of a central uplift $SU = 0.086D_s^{1.03}$, namely $\sim 30$ km. However, Ivanov and Melosh (in Glikson et al., 2004) point out that such large degrees of rebound, deduced from craters formed in high-viscosity crust, may not apply to the mantle. These authors state: ‘Indeed, a few giant impact events certainly occurred on Earth in post-LHB history, and a few might have encountered the fortuitous circumstances necessary for them to enhance magma production.’

Elkins-Tanton et al. (2005) modeled the magmatic consequences of very large impacts, suggesting that a 300 km radius crater excavating 75 km thick lithosphere will produce $10^6 \text{ km}^3$ magma through instantaneous in-situ decompression of the mantle and upwelling of the asthenosphere, triggering longer-term mantle convection and adiabatic melting, in particular under high geothermal gradients. These volumes are similar to those estimated by Ivanov and Melosh (2003). The short eruption spans of large volumes of some plateau basalts (Deccan: $2.10^6 \text{ km}^3$ in 0.3 Ma; Emeishan: $9.10^5 \text{ km}^3$ in 0.3 Ma; Siberian:
Table 8.5-1. Archaean and early Proterozoic asteroid impact fallout units and associated ferruginous sediments, Pilbara Craton, Western Australia, and Kaapvaal Craton (Barberton greenstone belt – BGB), South Africa

<table>
<thead>
<tr>
<th>Age/stratigraphy</th>
<th>Impact unit/s</th>
<th>Overlying ferruginous sediments</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>3470.1 ± 1.9 Ma: ACM-1, Antarctic Creek Member, Mount Ada Basalt, Warrawoona Group, Pilbara Craton</td>
<td>Silica-sericite spherules in ~1 m thick chert breccia/conglomerate</td>
<td>Overlain by felsic hypabyssal/volcanics</td>
<td>A</td>
</tr>
<tr>
<td>3470.1 ± 1.9 Ma: ACM-2, Antarctic Creek Member, Mount Ada Basalt, Warrawoona Group, Pilbara Craton</td>
<td>Silica-sericite spherules within ~14 m thick chert, arenite and jaspilite</td>
<td>~10 m thick jaspilite overlying spherule unit ACM-1</td>
<td>A</td>
</tr>
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<td>3470.4 ± 2.3 Ma: BGB–S1A and BGB–S1B, upper Hooggenoeg Formation, Onverwacht Group, Kaapvaal Craton</td>
<td>Two units of silica-chert spherules within 30–300 cm thick unit of chert and arenite</td>
<td>−</td>
<td>B</td>
</tr>
<tr>
<td>3258 ± 3 Ma: BGB–S2, base of the Mapepe Formation, Fig Tree Group, Kaapvaal Craton</td>
<td>≲310 cm thick silica-sericite spherules</td>
<td>MJM (Manzimnyama Jaspilite Member): BIF + jaspilite + ferruginous shale (&lt;20 m) and shale common above BGB–S2</td>
<td>B</td>
</tr>
<tr>
<td>3243 ± 4 Ma to 3225 ± 3 Ma: BGB–S3 and BGB–4, lower Mapepe Formation, Fig Tree Group, Kaapvaal, Craton</td>
<td>S3 – 10–15 cm-thick to locally 2–3 m thick silica-Cr sericite-chlorite spherules S4 – 15 cm thick arenite with chloride-rich spherules</td>
<td>BGB–S3 is overlain by ferruginous sediments of the Ulundi Formation in the northern part of the BGB</td>
<td>B</td>
</tr>
<tr>
<td>2629 ± 5 Ma: JIL, top Jeerinah Formation, Fortescue Group, Hamersley Basin.</td>
<td>Hesta – 80 cm thick carbonate-chlorite spherules and spherule-bearing breccia; 60 cm thick overlying debris flow</td>
<td>Marra Mamba iron-formation, immediately above ~60 cm thick shale unit overlying JIL</td>
<td>D</td>
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<tr>
<td>?2.63 Ga: Monteville Formation, West Griqualand Basin, west Kaapvaal Craton</td>
<td>5 cm thick spherule layer</td>
<td>Carbonate hosted</td>
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Table 8.5-1. (Continued)

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<th>Age/stratigraphy</th>
<th>Impact unit/s</th>
<th>Overlying ferruginous sediments</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>?2.56 Ga: Reivilo Formation, West Griqualand Basin, western Kaapvaal Craton</td>
<td>1.8 cm thick spherule unit</td>
<td>Carbonate-hosted</td>
<td>E</td>
</tr>
<tr>
<td>2562 ± 6 Ma: SMB-1, top of Bee Gorge Member, upper Wittenoom Formation, Hamersley Group, Hamersley Basin</td>
<td>&lt;5 cm thick K feldspar-carbonate-chlorite spherules in carbonate turbidite</td>
<td>Ferruginous siltstone (Sylvia Formation), banded iron formations (Bruno Member)</td>
<td>D</td>
</tr>
<tr>
<td>2562 ± 6 Ma: SMB-2, top of Bee Gorge Member, upper Wittenoom Formation, Hamersley Group, Hamersley Basin</td>
<td>&lt;20 cm thick K feldspar carbonate-chlorite spherules within turbidite</td>
<td>Ferruginous siltstone (Sylvia Formation), banded iron formations (Bruno Member)</td>
<td>D</td>
</tr>
<tr>
<td>2481 ± 4 Ma: S4, Shale Macroband, Dales Gorge Member, Brockman Iron-formation, Hamersley Group, Hamersley Basin</td>
<td>10–20 cm K-feldspar-stilpnomelane spherules at top of 2–3 m of ferruginous volcanic tuffs</td>
<td>Located 38 m above the base of the Brockman Iron-formation, Hamersley Basin</td>
<td>D</td>
</tr>
<tr>
<td>∼2.5–2.4 Ga: lower Kuruman Formation, West Griqualand Basin, west Kaapvaal Craton</td>
<td>1 cm thick spherule unit overlain by 80 cm breccia</td>
<td>Located 37 m above base of banded ironstones</td>
<td>E</td>
</tr>
<tr>
<td>1.85–2.13 Ga: Graensco, Vallen, Ketilidean, southwest Greenland</td>
<td>20 cm thick spherule unit</td>
<td>Carbonate-hosted spherules</td>
<td>E</td>
</tr>
</tbody>
</table>

3–4.10^6 km^3 in 0.4 Ma) are consistent with volumes deduced by these authors for excavation of large-radii impact craters. According to Elkins-Tanton et al. (2005), impacts of this magnitude will have world-wide effects for which ‘ample evidence could remain in the rock record.’

Ingle and Coffin (2004) suggested a possible impact origin for the short-lived ~120 Ma Ontong Java Plateau (OJP). Jones et al. (2002) conducted hydrocode simulations (Wünne- mann and Ivanov, 2003) to test whether impact-triggered volcanism is consistent with the OJP, using dry lherzolite melting parameters, a steep oceanic geotherm and a vertical incidence by a dunite projectile (diameter 30 km: velocity 20 km/s). The thermal and physical state of the target lithosphere is critical for estimates of melt production. Model calculations suggest impact-triggered melting in a sub-horizontal disc of ~600 km diameter down to >150 km depth in the upper mantle forms within ~10 minutes of the impact, whereas most of the initial melt forms at depths shallower than ~100 km. The volume of ultramafic melt would reach ~2.5 × 10^6 km^3, ranging from superheated melts within 100 km of ground zero, to partial melting with depth and distance. The total melt volume would reach ~7.5 × 10^6 km^3 of basalt if heat were distributed to produce 20–30% partial mantle melting.

Lowe et al. (1989) observed the possible tectonic significance of the S2–S4 Barberton impact layers in view of their juxtaposition with, and immediately above, the boundary between the ~3.55–3.26 Ga, predominantly mafic–ultramafic volcanic rocks of the Onverwacht Group and the overlying Fig Tree Group association of turbidite-felsic volcanic–banded iron formation. The evidence presented here and in earlier papers (Glikson and Vickers, 2005) shows that the 3.26–3.24 Ga impact cluster coincides with termination of the protracted evolution of greenstone–TTG (tonalite-trondhjemite-granodiorite) systems in the BGB and was accompanied by volcanic activity of the Sulphur Springs Group (Pilbara Craton) and by plutonic events in both the Kaapvaal Craton (Nelshoogte and Kaap Valley plutons) and the Pilbara Craton (Cleland Supersuite; Fig. 8.5-4). The full scale of the ~3.26–3.24 Ga bombardment may not have been identified to date, and the three recorded impact events may represent a minimum number. Iridium and chrome isotope-based mass balance calculations, assuming a global distribution of the impact layers – an assumption justified by the consistently tens of cm thick Achaean spherule layers as compared to the global ~3–4 mm thick K–T boundary layer (Alvarez, 1986). Spherule size analysis and the ferromagnesian composition of the impact ejecta suggest asteroids diameters of 20–50 km diameter, excavating impact basins 200–1000 km large and located in mafic–ultramafic crustal regions.

8.5-6. LUNAR CORRELATIONS

The Late Heavy Bombardment (LHB) of the Earth and Moon has been interpreted in terms of the tail-end of planetary accretion or, alternatively, a temporally distinct bombardment episode during 3.95–3.85 Ga (Tera and Wasserburg, 1974; Ryder, 1991). Some
of the largest lunar mare basins contain low-Ti basalt, which possibly represents impact-triggered volcanic activity, including Mare Imbrium (3.86 ± 0.02 Ga) and associated 3.85 ± 0.03 Ga K, REE, and P-rich basalts (KREEP; Ryder, 1991). Isotopic Ar-Ar dating of lunar spherules (Culler et al., 2000), when combined with earlier Ar-Ar and Rb-Sr isotopic studies of lunar basalts (BVTP 1981: Fig. 8.5-7), suggest the occurrence of post-LHB impact and volcanic episodes. Possible cause-effect relationships between impact and volcanic activity pertain in Oceanus Procellarum (3.29–3.08 Ga) and the Hadley Apennines (3.37–3.21 Ga). Such relationships gain support from laser 40Ar/39Ar analyses of lunar impact spherules (sample 11199; Apollo 14, Fra Mauro Formation: Fig. 8.5-7), showing a significant age spike at 3.18 Ga, near the boundary of the Late Imbian (3.9–3.2 Ga) and the Eratosthenian (3.2–1.2 Ga) as defined by the cratering record (Wilhelms, 1987). 34 lunar impact spherules yield a mean age of 3188 ± 198 Ma, whereas seven spherule ages with errors <100 My yield a mean age of 3178 ± 80 Ma. The small size of the sample and the large errors limit the confidence in these data. However, the combined evidence suggests that the period 3.24 ± 0.1 Ga experienced a major impact cataclysm in the Earth–Moon system, resulting in renewed volcanic activity in some of the lunar mare basins approximately at 3.2 Ga, as well as a Rb-Sr age peak about 3.3 Ga. As in the case of lunar impact spherules, large errors on the Ar-Ar and Rb-Sr isotopic ages preclude precise correlations.
Chapter 8.5: Early Archean Asteroid Impacts on Earth

8.5-7. SUMMARY

(1) The early Archean rock record from the Kaapvaal and Pilbara Cratons preserves evidence for two major impact clusters on Earth, one at 3.47 Ga, the other at between 3.26–3.24 Ga.

(2) A substantial body of petrological, geochemical, mineralogical and isotopic evidence for an extraterrestrial impact origin of the Barberton 3.26–3.24 Ga spherule units includes inward-radiating quench-textured and offset vesicles of microkrystite spherules, PGE abundance levels, PGE ratios, Ni/Cr, Ni/Co, V/Cr and Sc/V ratios, occurrence of quench-textured and octahedral nickel-rich chromites distinct from terrestrial chromites, and meteoritic $^{53}\text{Cr}/^{52}\text{Cr}$ anomalies (Shukloyukov et al., 2000; Kyte et al., 2003).

with terrestrial events. Further isotopic age studies of terrestrial and lunar materials are required in order to establish a $\sim$3.2 Ga bombardment in the Earth–Moon system.

Fig. 8.5-7. (Continued.)
8.5-7. Summary

(3) Mass balance calculations, based on the Ir flux and the $^{53}\text{Cr}/^{52}\text{Cr}$ anomalies (Byerly and Lowe, 1994) and on maximum spherule radii (Melosh and Vickery, 1991), suggest asteroid diameters in the order of 20–50 km, implying impact basins 200–1000 km large.

(4) Lowe and Byerly’s (1989) perception of potential significance of the 3.26–3.24 Ga impacts in view of their juxtaposition with, and position immediately above, the top of the >12 km thick, predominantly mafic–ultramafic volcanic sequence of the Onverwacht Group is corroborated by precise U-Pb ages.

(5) The occurrence of banded iron formation, jaspilite and ferruginous siltstone above the 3.26 Ga and 3.24 Ga impact units in the Barberton Greenstone Belt, and at equivalent stratigraphic levels in the Pilbara Craton, may indirectly suggest soluble Fe-enrichment of sea water closely following major impact events, possibly representing erosion of impact-triggered mafic volcanics under the low-oxygen fugacity conditions of the Archaean atmosphere, similar to other impact units in the Pilbara Craton (Glikson, 2006).