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The role of impacts on Archaean tectonics

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ABSTRACT

Field evidence from the Pilbara craton (Australia) and Kaapvaal craton (South Africa) indicate that modern tectonic processes may have been operating at ca. 3.2 Ga, a time also associated with a high density of preserved Archaean impact indicators. Recent work has suggested a causative association between large impacts and tectonic processes for the Hadean. However, impact flux estimates and spherule bed characteristics suggest impactor diameters of <100 km at ca. 3.5 Ga, and it is unclear whether such impacts could perturb the global tectonic system. In this work, we develop numerical simulations of global tectonism with impacting effects, and simulate the evolution of these models throughout the Archaean for given impact fluxes. We demonstrate that moderate-size (~70 km diameter) impactors are capable of initiating short-lived subduction, and that the system response is sensitive to impactor size, proximity to other impacts, and also lithospheric thickness gradients. Large lithospheric thickness gradients may have first appeared at ca. 3.5–3.2 Ga as cratonic roots, and we postulate an association between Earth's thermal maturation, cratonic root stability, and the onset of widespread sporadic tectonism driven by the impact flux at this time.

INTRODUCTION

Solar system impact flux models indicate an extended tail to the accretion process, with basin-forming impacts extending into the Archaean (Bottke et al., 2012; Marchi et al., 2014; Nesvorný et al., 2017; Morbidelli et al., 2017). Comparisons with post–late heavy bombardment, late Imbrian lunar impact rates suggest the production of ~70 craters with diameters, *D*, >150 km on Earth after 3.7 Ga (e.g., Bottke and Norman, 2017).

The terrestrial record of these events is incomplete due to limited preservation of Archaean crust. Impact-related spherule beds have been identified in the Barberton greenstone belt in the Kaapvaal craton, South Africa (Lowe and Byerly, 1986; Lowe et al., 2014), and the Pilbara craton, Australia (e.g., Glikson et al., 2016). Such layers form as vaporized impactor and rock mass condense to form small spherules, which, for large impacts such as Chicxulub (Mexico), are expected to fall out as globally contiguous deposits, which are preserved in favorable sedimentary environments. The Kaapvaal craton and Pilbara craton spherule layers suggest at least nine major impacts in the period 3.5–3.2 Ga (Lowe et al., 2014; see Table 1), many associated with iridium and chromium isotope anomalies. Modeling of the spherule layer thickness and spherule size distribution suggests that they ranged in projectile size from ~30 up to 70 km, with impact velocities between 18 and 22 km/s (Johnson and Melosh, 2012). Dating of these spherule beds suggests that many large impacts cluster at ca. 3.46–3.47 Ga (Glikson et al., 2016) and 3.2 Ga, with three major events, including the largest estimated impactor (41–70 km in diameter), occurring within 17 m.y. of each other.

The Barberton greenstone belt hosts most of the recognized pre–3.0 Ga spherule beds, with one (layer S1) correlated across the Pilbara craton (Byerly et al., 2002; see also Glikson and Vickers, 2006). Additionally, the Marble Bar chert in the Pilbara craton hosts two distinct spherule horizons (Glikson et al., 2016). Recent work on the ca. 3 Ga Maniitsoq structure, West Greenland (Garde et al., 2012), has suggested an impact origin. Despite the features of this structure being buried at 20–25 km at the time of formation, an impact origin is argued based on regional circular deformation associated with an aeromagnetic anomaly, modified planar deformation features, widespread fracturing, brecciation, and microstructural deformation features. If true, this suggests that the periods 3.41–3.47 and ca. 3.2 Ga preserve a remarkable record of intense impacting during the waning stages of accretion.

Lowe et al. (2003, 2014) noted that the formation of the spherule beds in the Barberton greenstone belt at ca. 3.2 Ga marked a transition in tectonic style. The underlying Onverwacht Group represents a typical Paleoarchean anorogenic volcanic regime dominated by komatiitic and basaltic volcanism and chemo-biological sedimentation. 3.2 Ga represents the onset of uplift, deformation, and terrigenous clastic sedimentation in the Fig Tree Group, representing the first major orogeny. Lowe et al. (2014) suggested a causative link between this orogenesis and the preserved impact events. 3.2 Ga also marks the onset of major lateral tectonics in the Pilbara craton (Van Kranendonk et al., 2007), including the rifting of the Karratha and Kurrana terranes and the possible onset of the first Wilson cycle. Van Kranendonk et al. (2007) suggested that this may mark the onset of plate tectonic processes.

Many recent estimates for the initiation of plate tectonics concur with the 3.2 Ga Pilbara craton record, albeit with an uncertainty range from ca. 700 Ma (Stern et al., 2016) to >4.4 Ga (Harrison et al., 2005). A tectonics transition at ca. 3.0 Ga has been inferred from geochemical models of MgO in mafics through time (Tang et al., 2016); inflections in MgO and Ni in mafic lithologies, and apparent percent melt changes, from statistical geochemistry (Keller and Schoene, 2012); a shift in juvenile Rb/Sr from primarily mafic, thin (~20 km), pre–3 Ga crust, to higher Rb/Sr ratios from thicker crust (Dhuime et al., 2012); a shift at 3.0 Ga in the source material of Archaean black shales from juvenile to differentiated material, from Hf systematics (Nebel-Jacobsen et al., 2018); and increased felsic volcanism from 3.5 Ga from Ti isotopes in shales (Greber et al., 2017).

Shirey and Richardson (2011) noted that eclogitic inclusions in diamonds appear at 3 Ga in Kaapvaal kimberlites, and suggested

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TABLE 1. AGE, SIZE, AND VELOCITY CONSTRAINTS ON PRE–3.0 Ga IMPACTORS FROM KAAPVAAL CRATON (SOUTH AFRICA) AND PILBARA CRATON (AUSTRALIA) SPHERULE BEDS

Layer	Age (Ga)	Minimum size (km)	Maximum size (km)	Minimum velocity (km/s)	Maximum velocity (km/s)
S1*	3.470	29	53	18.8	21.2
MBCM	3.454	48	89	19.5	23.3
S7	3.416	26	100	15.5	24.4
S8	3.298	53	101	16.9	24.4
S6	3.270	46	93	19.7	24.4
S2	3.260	37	58	17.7	25.6
S3	3.243	41	70	20.6	22.8
S4	3.243	33	53	18.2	22.2
S5	3.240	38	71	16.4	25.3

Note: S1–S8 ages are from Lowe et al. (2014); sizes and velocities are from Johnson and Melosh (2012) (S1–S4), or derived empirically from Melosh (2012) (S5–S8 and MBCM). MBCM—Marble Bar Chert Member, Pilbara craton (Glikson et al., 2016; containing two spherule horizons).

*Correlates with the Antarctic Creek Member, Pilbara craton (Byerly et al., 2002; Glikson, 2014).

a subduction origin, inferring plate tectonics from this time. Smart et al. (2016) argued that the nitrogen abundance in Archaean diamonds imply that they formed from an oxidized fluid and inferred its introduction into the mantle, via subduction, before ca. 3.2 Ga.

Archaean geodynamics simulations have largely shown the proclivity of hot early-Earth systems to enter a hot, stagnant volcanic regime, transiting to plate tectonics only as the system cools (see O'Neill et al., 2015, 2018, and references therein). The transition from pre–plate tectonics to plate tectonics in these models is very nonlinear, and may exhibit many false starts, consistent with geological observations (O'Neill et al., 2018), during which the system may be sensitive to external factors, including impacts.

Previous modeling of the geodynamic effects of large impacts in the Hadean (O'Neill et al., 2017) showed that extremely large impacting bolides (>~700 km diameter) directly initiate active tectonics and subduction due to the thermal buoyancy of impact-heated mantle. It also demonstrated that much smaller impacts could act as triggers for subduction if they occurred on lithosphere that was already primed for subduction. However, the proposed initiation of plate tectonics at 3.2 Ga would have occurred on a planet in a vastly different thermal regime to that of the Hadean. It is not clear whether (1) the proposed size of the Mesoarchean impacts could have initiated subduction at this time, or (2) subduction could have been self-perpetuating, and in fact started ongoing and continuing plate tectonic processes.

The purpose of our study is to assess whether the proposed size and flux of impacting bodies in the Mesoarchean could have initiated subduction events, assess the geodynamic factors favorable to tectonics, and determine if such events could have developed into self-perpetuating global plate tectonics, or whether they failed (Moyen and van Hunen, 2012; O'Neill et al., 2018).

MODEL AND RESULTS

Estimates of the size of the impactors that generated the Archaean spherule beds come from Johnson and Melosh (2012), who argued that spherule size distributions are related to impact velocity, and the reduced spherule layer thickness (with sedimentary effects removed) is related to impactor size. We have extrapolated their approach to more recently identified Archaean spherule beds of Lowe et al. (2014), and the resulting estimate ranges are shown in Table 1. The age of these spherule beds is more difficult to quantitatively constrain, because while they occur in a tight sedimentary succession, absolute ages are provided only by bounding volcanic units, or in some cases by youngest detrital zircons. Consequently, we have adopted a generic standard deviation of ± 10 m.y. on the most likely age estimates of Lowe et al. (2014).

We test whether the impacts in Table 1 can induce Archaean subduction by adopting an approach for modeling the thermal effect of impacts on the mantle (O'Neill et al., 2017; Roberts et al., 2009). The mantle convection simulations (Kronbichler et al., 2012) are in a spherical annulus, and employ a temperature-, pressure-, and strain rate–dependent Arrhenius viscosity, and plastic yielding. The models include evolving heat-production sources and a dynamic cooling core (Zhang and O'Neill, 2016). This thermal field was imported from a previous simulation, which evolved from 4.5 to 3.5 Ga from a post–magma ocean condition, incorporating waning heat production and impact flux (O'Neill et al., 2017), and was designed specifically to match that of Archaean mantle at 3.5 Ga. The models generally exhibit long-lived lid stability and volcanism, which is entirely consistent with many Eoarchean terranes (O'Neill et al., 2018).

Above a certain size threshold (~200–300 km, depending on the system), large projectiles are generally able to cause subduction events (Fig. 1).

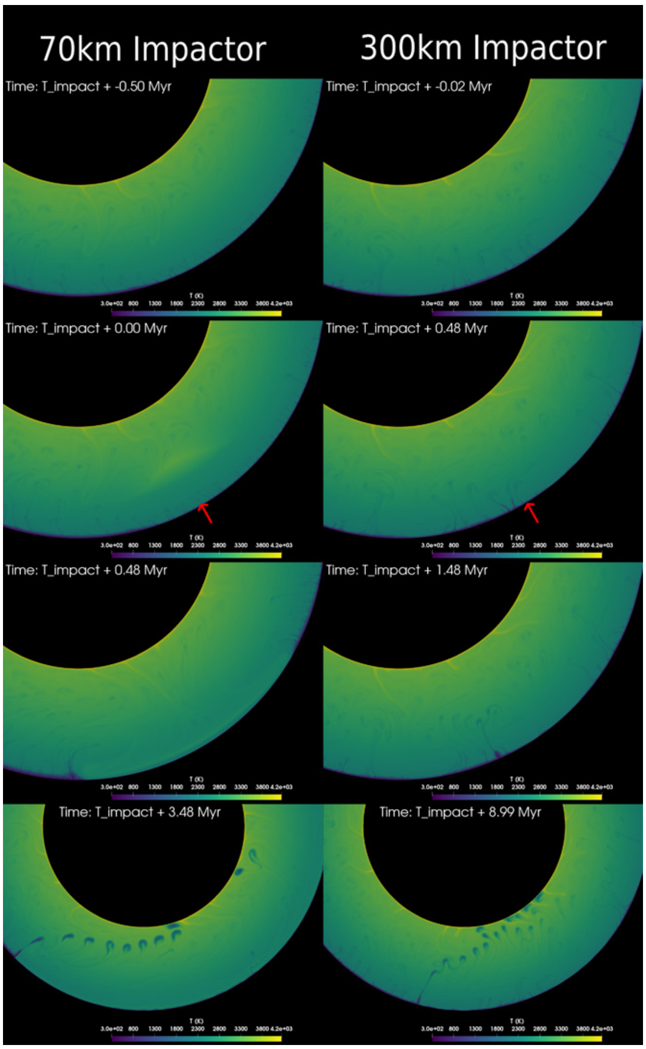

Figure 1. Series of modeled mantle temperature snapshots through time for a series of four impacts (S1–S4, from Johnson and Melosh, 2012; figures are shown from the time of impact S3 [T_impact = 3.243 Ga] onward), illustrating subduction initiation coincident with impact S3. Impactor sizes were held constant in this example, at either 300 km (to demonstrate effect of extreme example) or 70 km (maximum size of impactor S3 estimated from Johnson and Melosh [2012]). Red arrows show point of S3 impact. Ages of impacts are provided in Table 1.

In Figure 1, we show an illustrative example of an extreme 300-km-diameter projectile, which is beyond Archaean estimates, impacting at 3.2 Ga (after 300 m.y. of model evolution). The impact causes a large thermal anomaly in the mantle, which upwells, causing melting and spreading in the lithosphere above it. This consequently results in overthickening at the edge of the upwelling, and prolonged convergence initiates subduction at the edge. Below this size threshold, smaller impacts (projectiles ~50–100 km diameter) generate a much smaller and weaker thermal anomaly. They *may* still initiate subduction (Fig. 1), but commonly do not, contingent on factors related to the local lithospheric structure (i.e., thickness and gradient) and proximity to other recent impacts.

We unpack the effect of these factors in Figure 2. The primary metric we use to gauge subduction success is mobility—the ratio of internal root-mean-square velocities to those at the surface (Tackley, 2000). Values greater than one are generally associated with plate tectonic regimes, and models that show subduction here exhibit high (>1.1) values of mobility. To account for some impacts occurring off-slice, the models shown in Figure 2 employ either a subset of the impactors on the simulation plane (impactors S1–S4 [Table 1], for either constant impact sizes [Fig. 2A], or impact spacings [Fig. 2B]), or the impact suite (S1–S8; Figs. 2C and 2D). Figure 2A shows the maximum value of mobility for simulations using impactors S1–S4, assuming different impactor sizes in each simulation. All impacts assume an impact velocity of 20 km/s.

Size is a key determinant in mobility for large impacts (Fig. 2A), with average mobility decreasing for smaller impactors. Both 50 km (all impactors 50 km in size) and 100 km examples exhibited no subduction. Interestingly, the 75 km example exhibited subduction due to interaction of the associated thermal anomaly with the local lithospheric structure—highlighting the complex interactions between these systems.

The geographic positions of the spherule bed–producing impacts are unconstrained. Johnson and Melosh (2012) noted that the spherule falls are likely global. As a result, we vary the spacing between each subsequent impact between 0° (all impacts coincident) through to 90° (each impact 90° from the previous; Fig. 2B). The proximity of previous impacts has a large effect on mobility, with locally proximal impacts building on previous responses, promoting convergence. Figures 2C and 2D show the effect of lithospheric factors in the area of the impacts. For these, we use age estimates of layers S1–S8 (Table 1), assuming in all cases an impactor diameter of 100 km and a velocity of 20 km/s, but incorporating variable spacing (from Fig. 2B). The lithospheric thickness of the target (Fig. 2C) has very little effect on the mobility exhibited by the simulations. However, the local lithospheric thickness gradient (Fig. 2D) at the point of impact is positively correlated with mobility. Impact-induced upwellings in regions of strong lithospheric gradients can drive convergence and overthickening on the one hand, and induce thinning and critical lithospheric weaknesses on the other, in the zone surrounding the impact. This critical imbalance between thick negatively dense lithosphere and thin weak lithosphere has been shown to be significant in subduction initiation in simple convecting systems (Moresi and Solomatov, 1998).

Large uncertainties surround the distribution of impactors on the early Earth. To address this, we have run a small Monte Carlo suite of simulations, starting from the same initial conditions, incorporating impactors S1–S8 (Table 1). However, each simulation varies the age, size, and velocity of each impact, drawing a random sample from the uncertainty distribution of each parameter (Table 1), and the impactor's longitude from a uniform distribution (between 0° and 360°). The results for 30 simulations are shown in Figure 3. The blue shaded region in Figure 3 envelops the output of the 30 individual simulations, plotting mobility versus time. A number of individual simulations exhibited subduction and significant mobility (33% with mobility >1.07; Earth's present-day mobility is ~1.05–1.1). There are substantial mobility clusters near concentrations of impact flux. However, while these impact events can initiate subduction, it is not self-sustaining and does not develop into proper plate tectonics, as the tectonic system cannot sustain itself in the absence of external forcings.

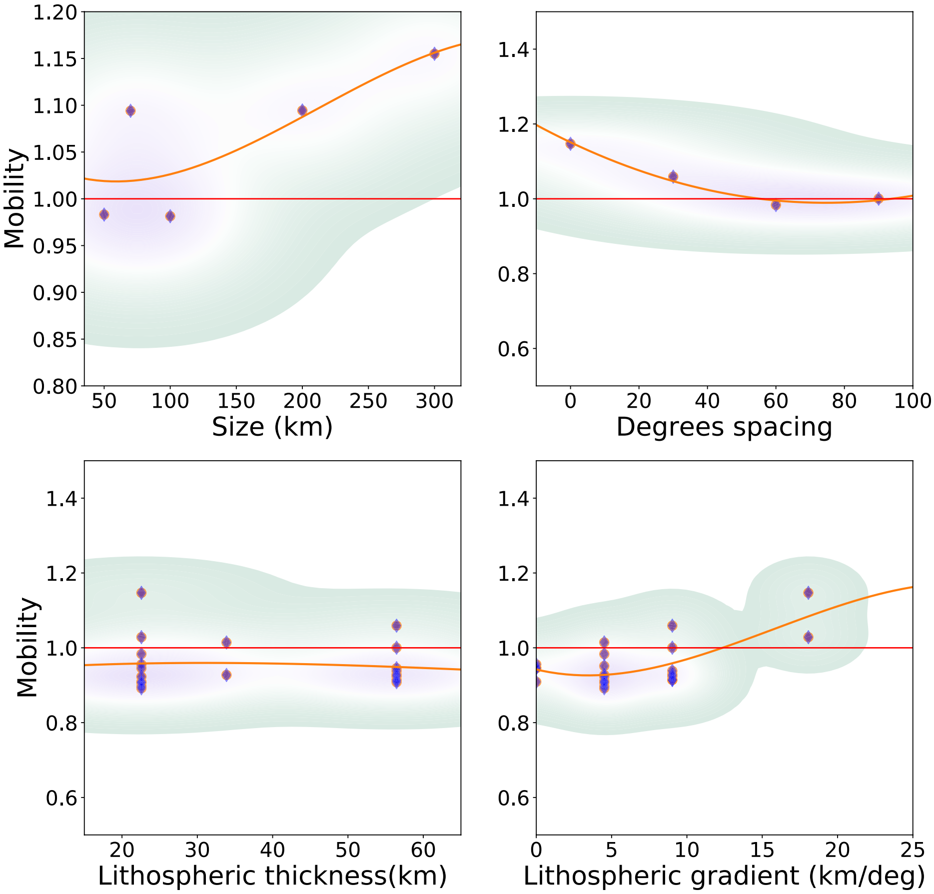

Figure 2. Modeled variation in mobility (ratio of internal root-mean-square velocities to those at surface) versus impact size (A), impact spacing (B; four impacts, either coincidental or 30°, 60°, or 90° apart; impact ages as per Fig. 1), lithospheric thickness for a Monte Carlo spread of impact sizes based on uncertainties shown in Table 1 (so random velocities and longitude) (C), and lithospheric thickness gradient (averaged within a few degrees of impact) (D). Shaded regions are kernel density estimations (KDEs) of results (purple indicating high probability densities, green bracketing uncertainty projections).

DISCUSSION AND CONCLUSIONS

The impact flux preserved in the Archaean record is likely a minimum estimate, though the size distribution inferred from spherule beds is consistent with lunar impacting rates (see Fig. DR1 in the GSA Data Repository[1]). Demonstrating an association between plate dynamics and

[1]GSA Data Repository item 2020049, supplementary model information, input, and characteristics, is available online at http://www.geosociety.org/datarepository/2020/, or on request from editing@geosociety.org.

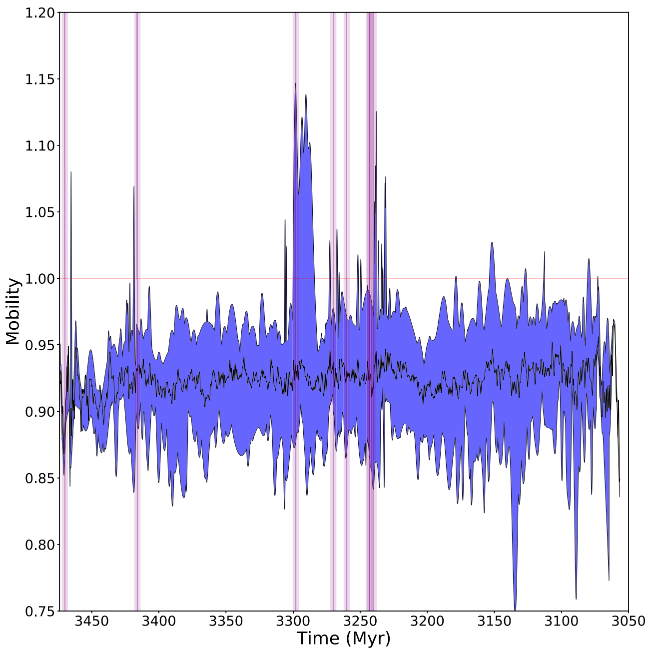

Figure 3. Evolution of mobility (ratio of internal root-mean-square velocities to those at surface) from all models over time. Exact age, size, velocity, and longitude of impacts S1–S8 (Table 1) are randomly selected from distribution based on uncertainties in these factors for each simulation (shown as purple vertical lines; mean age is darker purple, age uncertainty is lighter purple). Blue shaded area shows range of mobility demonstrated from all models throughout their evolution, as well as mean (thick black line) and maximum and minimum values (thin black lines). A number of runs had high mobility (and subduction) ca. 3.3–3.2 Ga. All models exhibited surface mobility values >1 (shown as horizontal red line), but not necessarily subduction, which was present in models with mobility >1.07 (33% of models).

impacts, however, requires multiple geodynamic constraints on plate behavior, such as paleomagnetics and volcanic fixity. Examples may include the association between long-lived volcanic fixity (e.g., ~100 m.y. of volcanism in the eastern Pilbara craton; Van Kranendonk et al., 2007) and apparent paleomagnetic fixity (i.e., overlapping poles; Strik et al., 2003)—which would support plate fixity and stagnant-lid tectonics. Alternatively, diverging apparent polar wander (APW) paths (ideally) or a single rapidly moving APW path with either fast volcanic progressions or the development and rapid evolution of plate-boundary indicators (such as arc and/or subduction hallmarks; e.g., Pearse, 2008) might indicate plate motion. As of now, such tests for 3.2 Ga are not—yet—diagnostic.

The Archaean spherule beds offer a unique insight into impacting effects on the early Earth. The change in tectonic behavior in the Kaapvaal craton and the onset of (micro)plate motions in the Pilbara craton suggest a period of plate mobility, requiring subduction, coincident with impacting events. The long tail of large impacts into the Mesoarchean is coincident with the first observed development of significant lithospheric thickness variations, as observed in cratonic diamonds in the Kaapvaal craton at 3.5–3.2 Ga (Shirey and Richardson, 2011). The development of significant lithospheric gradients is a critical factor for subduction initiation, and the temperatures of the early Earth may have precluded these from developing naturally—intrinsically weak lithosphere may have flowed rather than developed significant gradients. The onset of plate tectonics—either through impacts or through internal forces—may have been retarded until the thermal state of the Earth permitted the generation of significant lithospheric heterogeneity. The dynamics of the transition of the Earth from an earlier archaic regime into modern plate tectonics was likely protracted (O'Neill et al., 2018). Our work shows there is a strong association between tectonics, cratonization, and impacts, and the timing of the transition to active tectonics may have been very sensitive to the evolving solar system environment.

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