Impact origin of Archean cratons

Vicki L. Hansen[1,*]
[1]DEPARTMENT OF EARTH AND ENVIRONMENTAL SCIENCES, UNIVERSITY OF MINNESOTA–DULUTH, 1114 KIRBY DRIVE, DULUTH, MINNESOTA 55218, USA

ABSTRACT

Archean cratons consist of crustal granite-greenstone terrains (GGTs) coupled to roots of strong, buoyant cratonic lithospheric mantle (CLM). Although this association is unique to the Archean and formed from ca. 4.0 to 2.5 Ga, the origins of terrestrial cratons are debated. I propose that crustal plateaus, quasi-circular craton-like features (~1400–2400 km diameter, 0.5–4 km high), on Earth's sister planet Venus might serve as analogs for Archean cratons. Crustal plateaus, which are isostatically supported by a compositionally controlled low-density root, host a distinctive surface called ribbon-tessera terrain. Ribbon-tessera also occurs as arcuate-shaped inliers in the Venus lowlands, widely interpreted as remnants of rootless crustal plateaus. Within each crustal plateau, surface ribbon-tessera terrain comprises a vast igneous province analogous to terrestrial GGTs, and the plateau root is analogous to CLM. Crustal plateaus and ribbon-tessera terrain collectively represent Venus' oldest preserved features and surfaces, and they formed during an ancient period of globally thin lithosphere. To explain the linked features of crustal plateaus, a bolide impact hypothesis has been proposed in which a large bolide pierces ancient thin lithosphere, leading to massive partial melting in the sublithospheric mantle. In this model, melt escapes to the surface, forming an enormous lava pond, which evolves to form ribbon-tessera terrain; mantle melt residue forms a strong, resilient buoyant root, leading to plateau support and long-term stability of an individual crustal plateau. Building on the similarity of GGT–CLM and Venus crustal plateaus, I propose an exogenic hypothesis for Archean craton formation in which a large bolide pierces thin Archean lithosphere, causing localized high-temperature, high-fraction partial melting in the sublithospheric mantle; melt rises, forming an igneous province that evolves to form a GGT, and melt residue develops a complementary CLM. By this mechanism, Archean cratons may have formed in a spatially and temporally punctuated fashion at a time when large bolides showered Archean Earth.

LITHOSPHERE; v. 7; no. 5; p. 563–578 | Published online 24 August 2015 doi:10.1130/L371.1

INTRODUCTION

There is little consensus with regard to processes by which terrestrial Archean cratons formed. Cratons, which refer to old and stable parts of the continental lithosphere, include both crust and cratonic lithospheric mantle (CLM). Archean cratons, in contrast to their Proterozoic successors, are characterized by crustal granite-greenstone terrains (GGTs) and unusually strong and buoyant CLM (e.g., Condie, 2007; Griffin and O'Reilly, 2007). GGTs, which are uniquely Archean in age, include keel-shaped greenstone packages (basalt, komatiite, and sedimentary strata) deformed between ovoid to elongate granitoid bodies (tonalite-trondhjemite-granodiorite [TTG]). Archean CLM, distinguished by high-Mg composition, is dry, refractory, low density, high viscosity, and has a high brittle yield stress (Griffin and O'Reilly, 2007). The preservation of GGTs over 2.5 b.y. is credited, in part, to the uniquely strong and buoyant character of the underlying CLM, which has protected individual GGTs from being recycled (Jordan, 1975, 1981; Davies, 1979; Abbott et al., 1997; Moser et al., 2001; Poudjom Djomani et al., 2001; Lenardic et al., 2003; Lee et al., 2011). Thus, GGTs owe their preservation to CLM, and a growing body of age and geochemical data indicates that GGTs and associated CLM are also genetically related (Groves et al., 1987; Pearson, 1999; Griffin et al., 2004, 2009a; Carlson et al., 2005; Bédard, 2006; Lee, 2006; Percival and Pysklywec, 2007; Pearson and Wittig, 2008). Models for GGT formation include crustal density inversion, convergent-margin plate-tectonic processes, plume-driven processes, and hybrid plate-tectonic and plume mechanisms. Models for CLM formation include ocean ridge–type magmatism, convergent plate-boundary magmatism, plume-type magmatism, and extensive melting of hot ambient mantle in ocean ridge–like settings. Essentially all of the currently considered models for GGT, CLM, or coupled GGT–CLM formation call on endogenic processes.

This contribution presents a brief review of Archean cratons and models of coupled GGT–CLM formation. It is proposed that Venusian crustal plateaus—large ancient quasi-circular features that formed during the earliest era of recorded history on Venus (e.g., Phillips and Hansen, 1994, 1998; Ivanov and Head, 1996)—represent analogs of terrestrial Archean cratons. An adaptation of the bolide impact hypothesis for crustal plateau formation (Hansen, 2006) is discussed to address coupled GGT–CLM formation. In this model, Archean GGTs and associated CLM formed as a result of the impact of a large bolide (order of ≥30 km diameter) with relatively thin lithosphere; such a bolide would pass through the lithosphere into the sublithospheric mantle, resulting in high-fraction, high-temperature partial melting; melt would then escape upward to form a large igneous province, which would ultimately evolve into a GGT, whereas mantle melt residue would form the associated CLM root. A bolide impact hypothesis for GGT–CLM formation describes a process that would be uniquely Archean, requiring both thin lithosphere and frequent large bolides.

ARCHEAN CRATONS AND COUPLED GGT–CLM FORMATION

Archean cratons are characterized by GGTs and unusually strong and buoyant CLM. There is an extremely rich literature on the nature of GGTs and CLM and proposed models to explain their formation (e.g., Bleeker, 2002; Griffin and O'Reilly, 2007; Lee et al., 2011). This contribution is not intended to provide a complete review of these topics. Rather, the focus

*vhansen@d.umn.edu

© 2015 Geological Society of America | For permission to copy, contact editing@geosociety.org

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

here is on those critical aspects that are relevant to a comparison of terrestrial and Venusian features, providing context for a new hypothesis for Archean craton formation.

Key Features of Granite-Greenstone Terrains

GGTs, a hallmark of Archean crust, provide the most complete record of early terrestrial crust formation. Individual GGTs are typically quasi-circular provinces that are distinctly different from linear Proterozoic and Phanerozoic orogenic belts (Anhaeusser at al., 1969; Ayres and Thurston, 1985; Bickle et al., 1994; Goodwin, 1996; Choukroune et al., 1995, 1997; Hamilton, 1998; Condie, 2007). Globally, GGTs display similar geological characteristics, marked by distinctive assemblages of domal to elongate granitoid bodies (tonalite-trondhjemite-granodiorite, or TTG) with intervening greenstone septa or synclinal keels consisting of supracrustal units that variably wrap around the domes. Supracrustal units include interlayered metasediments, komatiite flows, and so-called non-arc basalt (Herzberg et al., 2010). Structurally, the supracrustal units display variably developed L-S tectonite fabrics, typically characterized by downdip (to oblique) radial elongation lineation patterns and shear criteria that indicate dip-slip movement, commonly with upward displacement of the TTGs relative to enveloping greenstones (Chardon et al., 1996, 2002; Choukroune et al., 1997; Collins et al., 1998; Lin, 2005; Lin et al., 1996, 2013; Parmenter et al., 2006). Midcrustal parts of the supracrustal assemblages commonly show concentric contact metamorphic zonation, whereas shallow crustal levels record a consistent regional low-grade metamorphism (Bédard et al., 2003). The shallow-level metasedimentary units may also contain detritus from older surrounding rocks or from eroded TTGs. Supracrustal strata generally young away from granitoid bodies and lack evidence of stratigraphic repetition, as would be expected in the case of structural imbrication; rather, the strata comprise a coherent, areally extensive, autochthonous stratigraphic package (Thurston et al., 2008). In general, GGT basalt and komatiite units have juvenile isotopic signatures, although some felsic units contain inherited zircon providing evidence of local older crust.

TTG, Komatiite, and Non-Arc Basalt

The types and characteristics of igneous rocks in GGTs—TTG, komatiite, and non-arc basalt—collectively impose constraints on crust formation mechanisms. TTGs are distinctly different from Paleozoic and younger volcanic arc granitoids; their formation cannot be accommodated within a modern-type convergent-margin environment (Bédard et al., 2003, 2013; Herzberg et al., 2010). TTGs within some provinces record evidence of crustal inheritance (e.g., Superior Province; Bédard, 2003; Thurston et al., 2008), and elsewhere dacite is intermingled with komatiite (e.g., Yilgarn craton; Trofimovs et al., 2004). Archean peridotitic komatiites have the highest MgO contents and eruption temperatures of any terrestrial volcanic rock (Bickle, 1986); their formation requires high-temperature (T) partial melting (~1700–1800 °C) and a high melt fraction of the (primitive) mantle (Herzberg et al., 2010). GGT non-arc basalt, so named because its formation is incompatible with convergent plate-margin settings, has model primary magma that exhibits mantle potential temperatures of 1500–1600 °C (Herzberg et al., 2010). In addition, the mean Mg numbers of residues associated with Archean non-arc basalt are 92–94 (higher than the mean Mg number of residue produced during formation of modern-day mid-ocean-ridge basalt; Herzberg and Rudnick, 2012). Interlayered komatiite and basalt flows indicate temporal overlap of flow formation (Green, 1972; Trofimovs et al., 2004).

Models of Granite-Greenstone Terrain Formation

Genetic models for GGT include: (a) early ideas of density-driven crustal overturn (i.e., sagduction and diapirism; McGregor, 1951; Anhaeusser et al., 1969; Mareschal and West, 1980; Goodwin, 1982; Choukroune et al., 1997); (b) plate-tectonic models involving accretion of arcs and sedimentary and/or accretionary prisms (Windley, 1984; Percival and Williams, 1989; Percival et al., 1994; Calvert et al., 1995; Cawood et al., 2006, 2009; Polat et al., 2009); (c) hybrid plate-tectonic models of oceanic plateau accretion coupled with arc and plume magmatism (Davis et al., 1988; Card, 1990; de Wit, 1998; Kusky, 1998; Kusky and Polat, 1999; Chown et al., 2002; Daigneault et al., 2002; Wyman et al., 2002; Percival et al., 2004; Benn and Moyen, 2008); (d) crustal convective overturn models, conceptually similar to sagduction/diapirism models (Chardon et al., 1996, 1998, 2002; Collins et al., 1998; Rey and Houseman, 2006; Bédard 2003, 2006; Bédard et al., 2003, 2013; Van Kranendonk et al., 2004; Sandiford et al., 2004; Robin and Bailey, 2009); and (e) hybrid models that include both crustal overturn and convergent plate-margin processes (Choukroune et al., 1997; Rey et al., 2003; Lin, 2005; Van Kranendonk, 2010; Lin et al., 2013).

Key Features of Cratonic Lithospheric Mantle

The roots of Archean cratons are variably referred to as subcontinental lithospheric mantle, subcratonic lithosphere, and cratonic lithospheric mantle. The term cratonic lithospheric mantle, or CLM, is used herein, and discussion is limited to Archean CLM. CLM comprises a thick (≥200 km), sharply defined root of dehydrated, refractory, highly depleted mantle peridotite that, owing to its composition, is uniquely buoyant and strong despite its relatively cold thermal state (Jordan, 1981, 1989; Pollack, 1986; Boyd, 1989; Carlson et al., 2005; Griffin et al., 2003, 2009b). CLM composition, known from mantle xenoliths and inferred though model calculations, is enriched in MgO and depleted in FeO, CaO, and Al_2O_3 (Herzberg et al., 2010; Herzberg and Rudnick, 2012, and references therein). Archean CLM is thicker (200 km vs. 100 km), cooler, more depleted, and more Mg-rich than younger continental lithospheric mantle (Griffin and O'Reilly, 2007). Once formed, CLM is difficult to recycle into the deep mantle because of its high buoyancy (i.e., compositional low density), high viscosity, and high brittle yield strength (Jordan, 1975, 1981; Griffin and O'Reilly, 2007; Lee et al., 2011), although hydration of CLM might ease mantle recycling in some cases (Lee, 2006). It is well accepted that CLM represents highly depleted mantle melt residue, with inferred melt fractions on the order of 25%–45% or greater, with initial melting taking place at 3–6 GPa and final melting at 1–3 GPa (Herzberg et al., 2010, and references therein).

Models of Cratonic Lithospheric Mantle Formation

The composition and character of CLM are well known, yet its origin is controversial. Models can be broadly grouped into four end-member scenarios in which the melt residue results from: (f) melting and metasomatism of accreted oceanic lithosphere; (g) melting and metasomatism of accreted arc lithosphere; (h) high-degree melting in an ocean-ridge setting within a high-temperature ambient mantle environment; or (i) high-degree melting in a high-temperature plume head (Lee, 2006; Lee et al., 2011; Herzberg and Rudnick, 2012). These models are here designated f–i in order to differentiate them from models for GGT formation (a–e) noted earlier.

GENETIC MODELS OF GGT–CLM COEVOLUTION

Although formation mechanisms of GGTs and CLM have commonly been historically

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

treated separately—that is, many models of GGT origin do not link their evolution to CLM formation, and vice versa—several studies, noted earlier herein, indicate that GGTs and Archean CLM are genetically related, and that the strong buoyant CLM root protected and preserved the associated GGT for over 2.5 b.y. Several geodynamic models that address formation of either GGT or CLM also provide explanation of linked mantle-crust processes, even if the linkage was not explicitly emphasized. Increasingly, recent models have proposed a coupled GGT–CLM coevolution (e.g., Bédard, 2006; Robin and Bailey, 2009; Herzberg and Rudnick, 2012; Campbell and Griffiths, 2014).

The following discussion reviews geodynamic models and mechanisms for GGT and CLM coevolution that account both for their key features and mechanisms of coupled formation. GGT model a, and to a certain extent model d, involves the tectonomagmatic evolution of igneous provinces or crustal packages driven by density inversion, yet this process does not address formation of an associated CLM. With the exception of GGT model a, the remaining models b–i can be grouped into four classes of crust-mantle interaction: accretionary models, involving the accretion of ocean lithosphere and/or volcanic arcs along convergent plate margins (b, f, and g, and possibly c); ocean-ridge–hot-ambient-mantle models (h); plume models that call for a major role by deep mantle plumes (d and i, and possibly c); and hybrid models involving plume and accretion tectonics (c and e, and possibly b and f). The first three classes are briefly discussed in light of requirements imposed by key features of both GGT and CLM. Hybrid models are not discussed, individually or as a group, although discussion of the other models may likewise apply to some hybrid models.

Accretionary Models

Accretionary processes are commonly invoked to explain the origin of volcanic crust, yet accretionary models fail to explain: (1) the nonlinear nature of GGTs; (2) the lack of evidence for imbrication and repetition of strata across individual GGTs; (3) the apparent autochthonous or in situ development of GGT supracrustal units; (4) the widespread occurrence of GGT non-arc basalt; and (5) the interlayered nature of non-arc basalt and komatiite. The eruption of high-*T* lavas in GGTs and formation of strong anhydrous CLM are inconsistent with melting in a hydrous, convergent-margin environment; rather, peridotitic komatiite and non-arc basalt magmatism requires dry, high-*T*, high-fraction melting, as does the formation of anhydrous CLM (Herzberg et al., 2010; Herzberg and Rudnick, 2012). Accretionary models are therefore not discussed further here.

Ocean-Ridge and Plume Models: Coupled GGT–CLM Formation

Ocean-ridge and plume models account for petrologic data that indicate the GGT and CLM are genetically related, and that they are linked by high-*T*, high-fraction partial melting of the mantle; in these scenarios, melt rises to form a large igneous province that evolves into a crustal GGT, whereas melt residue forms the associated CLM (Herzberg et al., 2010; Herzberg and Rudnick, 2012). Although these two types of models both invoke high-*T* melting and have similar outcomes, they address the problem from different perspectives (Fig. 1). Based on the high-MgO values of GGT non-arc basalt, thermodynamic arguments require high-*T*, high-fraction melting to form both the basalt melt and CLM melt residue (Herzberg et al., 2010; Herzberg and Rudnick, 2012). The high temperatures required to form Mg-rich melts is implicit in both ocean-ridge and plume models, but the environment invoked to explain high-*T* partial melting and high-fraction melting of the mantle differs greatly. The debate centers on where melting took place and what the high-MgO values of non-arc basalt

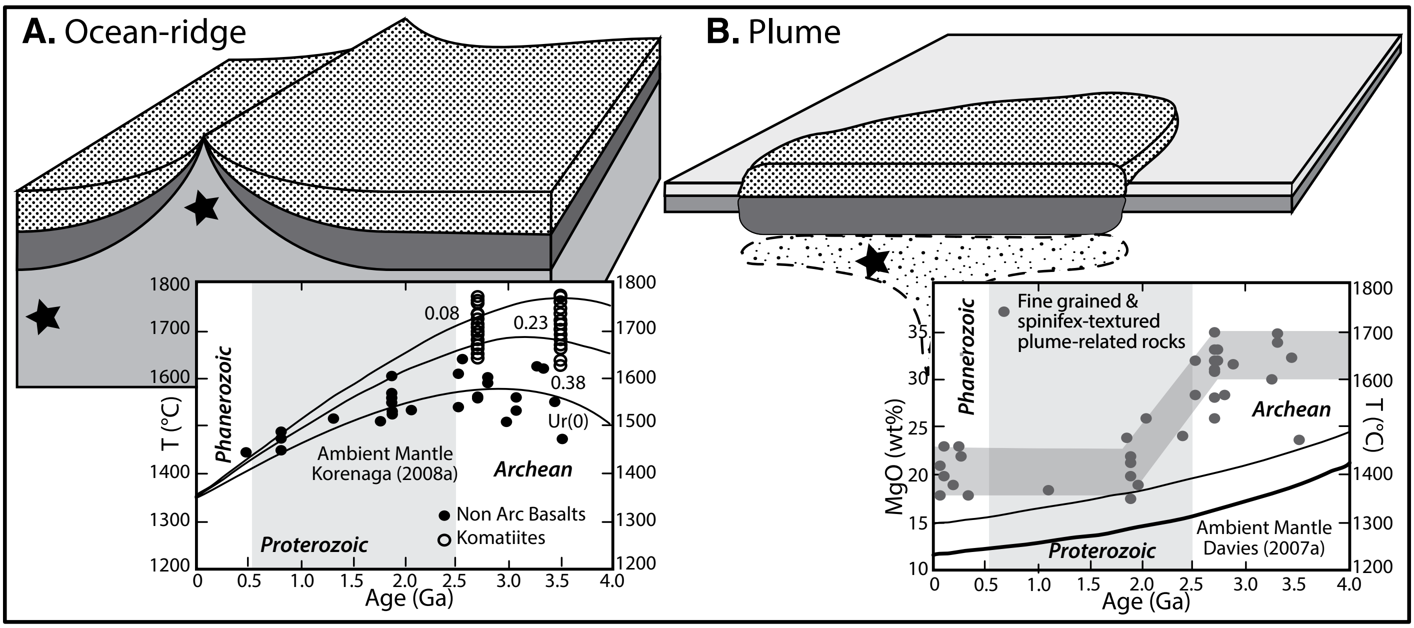

Figure 1. Block diagrams of (A) ocean-ridge and hot-ambient-mantle model and (B) plume model for GGT–CLM formation. For each model, block diagrams show inferred geodynamic setting of melting (stars) based on interpreted temperature proxies shown in accompanying graphs. Graphs show inferred temperatures of melting as a function of age, derived from MgO contents of magmatic products, and model secular mantle cooling curves. Graph in A is adapted from Herzberg et al. (2010, their Fig. 1) and shows MgO content of non-arc basalt (filled dots) and komatiite lavas (open dots), and mantle cooling models with present-day Urey ratios (Ur) ranging from 0.08 to 0.38 (Korenaga, 2008a). Only temperatures are plotted here, following presentation of Herzberg et al. (2010). Graph in B is adapted from Campbell and Griffiths (2014, their Fig. 2) and shows MgO contents of plume-related lavas superimposed on mantle cooling models (curved lines) of Davies (1999, 2007a).

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

represent, both of which depend, at least in part, on model assumptions.

In an ocean-ridge or hot-ambient-mantle model (Herzberg and Rudnick, 2012), GGT non-arc basalt and CLM form together as a result of ocean-ridge magmatism within a hot-ambient-mantle environment (Fig. 1A). Decompressional, high-*T*, high-fraction melting results in the formation of 30–45-km-thick oceanic crust along a divergent boundary resembling modern ocean ridges; melt residue forms complementary CLM. In this model, GGTs preserved today evolved from the thick mafic crust, and this survived >2.5 b.y., residing atop host CLM. Because MgO values of GGT non-arc basalt are interpreted as a proxy for Archean ambient mantle temperature, the values of which are consistent with thermal model estimates of Korenaga (2008a), this class of model assumes a high Archean ambient mantle temperature (~1600 °C). The ocean-ridge model does not directly address formation of GGT komatiites.

Plume models (Fig. 1B; e.g., Campbell and Griffiths, 2014) also address coupled GGT–CLM formation, and the thermodynamic and petrologic modeling employed in ocean-ridge models applies equally well in plume models (Herzberg et al., 2010; Herzberg and Rudnick, 2012). However, in a mantle plume model, the high-MgO values of basalt and komatiite are interpreted as a proxy for plume temperature rather than that of the ambient mantle. Plume models assume a relatively low temperature Archean mantle, consistent with thermal model estimates of Davies (1999, 2007c); in this case, mantle plumes represent elevated thermal anomalies responsible for partial melting.

Both ocean-ridge and plume models start with constraints imposed by the composition of non-arc basalt; in general, proponents of each mechanism agree that these volcanic rocks record high-*T* and high-fraction melting of the mantle, leading to formation of material that will ultimately form GGT (melt) and CLM (residue). A key difference lies in the interpreted ambient temperature of the Archean mantle. In ocean-ridge models, MgO values of GGT non-arc basalts serve as a proxy for the Archean ambient mantle, and, by extrapolation, evidence of a high-*T* ocean-ridge environment (Fig. 1A); thus, the MgO-derived temperature tracks an Archean global norm. In contrast, in a plume model, MgO values of both basalt and komatiite record mantle plume temperature; MgO-derived temperatures record mantle thermal anomalies (Fig. 1B), and thus the inferred global norm (i.e., ambient mantle temperature) is lower. There are, at least, two important unknowns in these models. We do not know, a priori, the geodynamic environment of high-*T*, high-fraction melting that led to coupled GGT–CLM formation, or the ambient temperature of Archean mantle. The former embodies model hypotheses, and the latter is a topic of debate, with values ranging from ~1400 °C to 1750 °C (Korenaga, 2008b). The ocean-ridge and plume models are, therefore, equally valid given the current state of understanding of mantle thermal evolution, yet their different consequences may provide a basis for evaluation.

VENUS CRUSTAL PLATEAUS AS AN ANALOG FOR GGTs–CLM

Earth's sister planet Venus might provide a fruitful environment in which to discover analogs for GGT–CLM and therefore provide fresh ideas for their formation. Specifically, Venusian crustal plateaus are ancient continent-like features characterized by a unique tectonic terrain (so-called ribbon-tessera terrain; Hansen and Willis, 1996, 1998) that might serve as an analog for terrestrial Archean cratons. Moreover, a hypothesis by Hansen (2006) for the formation of crustal plateaus by bolide impact and formation of large lava ponds, which emerged from detailed structural and geologic analysis of ribbon-tessera terrain, might also serve as a viable hypothesis for coevolution of terrestrial GGT–CLM in the Archean. To develop the concept of a Venusian analog, the discussion begins with a brief description of Venus crustal plateaus and ribbon-tessera terrain, followed by highlights of the lava-pond–bolide-impact hypotheses for genetically-linked ribbon-terrain–crustal plateau formation.

Venus Crustal Plateaus and Ribbon-Tessera Terrain

Venus and Earth are commonly referred to as sister planets, due to similarities in size, density, inferred composition, heat budget, and (inner) solar system location (Solomon et al., 1991). Given these similarities, Venus may provide clues about processes operative on early Earth that we otherwise cannot obtain. Venus and Earth were likely most similar in their early years (Hansen, 2007). Of relevance to the discussion here are Venus' continent-like features called crustal plateaus and their related lowland ribbon-tessera-terrain inliers, which together might serve as analogs for GGT–CLM on Earth. Crustal plateaus, large quasi-circular features (~1400–2400 km diameter; 0.5–4.0 km high) characterized by a distinctive deformation fabric called ribbon-tessera (Hansen and Willis, 1996, 1998), are isostatically compensated by thick crust or low-density upper mantle (Smrekar and Phillips, 1991; Simons et al., 1997; Phillips and Hansen, 1998; Hansen, 2006). Spatial correlation of plateau topography with tessera suggests a genetic relationship between ribbon-tessera terrain at the surface and formation of a low-density root (Bindschadler et al., 1992a, 1992b; Bindschadler, 1995; Brown and Grimm, 1997; Ghent and Hansen, 1999; Hansen et al., 1999). Inliers of distinctive ribbon-tessera terrain that outcrop in the lowlands are widely accepted as remnants of rootless crustal plateaus (Bindschadler et al., 1992b; Phillips and Hansen, 1994; Bindschadler, 1995; Ivanov and Head, 1996; Ghent and Tibuleac, 2002), an interpretation borne out by regional trends of ribbon-tessera fabric across Venus' lowland (Hansen and López, 2010). Crustal plateaus and ribbon-tessera terrain formed during the earliest era of recorded history on Venus (Bindschadler et al., 1992a; Bindschadler, 1995; Phillips and Hansen, 1994; Ivanov and Head, 1996), marking a time when Venus hosted a globally thin lithosphere (e.g., Solomon, 1993; Grimm, 1994; Solomatov and Moresi, 1996; Schubert et al., 1997; Phillips and Hansen, 1998; Brown and Grimm, 1999).

Tessera terrain refers to a distinctive surface characterized by two suites of near-orthogonal lineaments, first recognized as "parquet terrain" in low-resolution *Venera* radar images (Barsukov et al., 1986; Basilevsky et al., 1986). High-resolution radar imagery from the National Aeronautics and Space Administration (NASA) *Magellan* mission has allowed for more complete description and characterization of this surface (Hansen and Willis, 1996), including the definition of ribbon-tessera terrain characterized by orthogonal suites of contractional and extensional structures (Hansen and Willis, 1998). Formation of ribbon-tessera, and the crustal plateaus that host it, remains a topic of debate (see Hansen, 2006, and references therein). Reminiscent of the debates about GGT–CLM formation, many models address coupled ribbon-tessera terrain and crustal plateau (root) formation, whereas others do not. A complete review of this debate is beyond the scope of the current contribution (see Hansen et al., 2000; Hansen, 2006). This discussion highlights the results of the most detailed structural study of ribbon-tessera and the evolutionary model for coupled ribbon-tessera and crustal plateau formation (Hansen, 2006).

Ribbon-tessera terrain constitutes Venus's locally oldest surface unit (Basilevsky and Head, 1998; Head and Basilevsky, 1998), although globally not all ribbon-tessera formed at the same time (Bindschadler et al., 1992a; Bindschadler, 1995; Hansen and Willis, 1996; Hansen et al., 2000). Ribbon-tessera fabric includes short- to long-wavelength folds (1–50 km) that

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

record layer shortening, and orthogonal structures that record layer extension, including periodically spaced ridges and troughs (ribbons) and isolated graben complexes. Although there is some debate about whether folds dominantly predate or postdate ribbon structures, most workers agree that: (1) fold and ribbon formation broadly overlapped in time; (2) graben complexes formed late; and (3) magmatic activity (local flooding by lava) accompanied ribbon-tessera fabric formation (see Hansen, 2006, and references therein). The characteristic ribbon character and spacing across 10^6 km^2 (Ghent and Tibuleac, 2002) require a high local geothermal gradient (minimum heat flow similar to, or greater than, that of terrestrial mid-ocean ridges) during ribbon structure formation (Gilmore et al., 1998; Ruiz, 2007). Calculated heat-flow values for ribbon formation and the estimated depth to the crustal solidus require either extremely thin crust, or a very shallow crustal magma reservoir across areas of ribbon-tessera formation (Ruiz, 2007).

Detailed geologic mapping and structural analysis indicate that ribbon-tessera fabrics record broadly synchronous contraction, extension, and local flooding of an initially strong, thin layer (≤100 m to <1 km thick) above a low-viscosity layer (melt), developed across the entire areal extent of individual crustal plateaus (~1400–2400 km diameter; Hansen, 2006). A remarkable feature of ribbon-tessera terrain is its association with a vast melt layer during progressive deformation and cooling, resulting in the formation of longer-wavelength structures as the layer thickened over time. Extensional structures (both ribbons and graben complexes) developed orthogonal to contractional structures (folds). Short-wavelength structures, both ductile folds and brittle ribbon fabrics, formed early with structural wavelength tracking layer thickness (Hansen, 2006, her Table 1 and Fig. 7). As the initially thin, regionally extensive layer was deformed, low-viscosity lava flooded local structural lows. Simultaneously with progressive deformation, the layer thickened by: (1) deformation-driven layer shortening and thickening; (2) lava flooding of surface structural lows; and (3) layer thickening from below as a result of secular cooling and crystallization. As layer thickness increased, so too did the wavelength of evolving structures. Early-formed shorter-wavelength structures and lava-filled valleys were carried piggyback on later-formed, longer-wavelength folds (Hansen, 2006, her Fig. 10). Periodically spaced ribbon structures formed earlier during progressive deformation when the layer was thin; more widely spaced, less periodically developed graben complexes formed later, typically cutting the crests of long-wavelength folds. The deformation history places strict limits on the rheological properties of the evolving crustal layer, requiring an exponential decrease in viscosity with depth. This sharp decrease in viscosity requires a transition from a solid surface layer to a subsurface liquid, which also served as the source of low-viscosity material that flooded local topographic lows (for analysis and discussion, see Hansen, 2006). The picture that emerges is one of a vast lava pond on an extensive but relatively thin lithosphere (e.g., tens of kilometers thick, as opposed to ~100 km thick), with ribbon-tessera terrain forming the "scum" of the pond during progressive solidification, consistent with thermal model constraints (e.g., Gilmore et al., 1998; Ruiz, 2007).

To summarize, ribbon-tessera terrain represents lava pond "scum"—a mechanically competent surface layer that congealed across a vast lava pond (i.e., large igneous province) with an areal extent similar to that of individual present-day crustal plateaus. Folds formed by selective amplification of instabilities in the congealed surface as a result of layer shortening, likely driven by convection in the pond magma. Extensional structures formed with trends normal to fold crests, tracking maximum horizontal extension. Pond magma locally leaked to the surface, flooding local lows in the ever-changing structural topography; solidification of flood lava helped preserve a record of the progressive deformation of the surface layer. The surface layer thickened as a result of deformation, flooding, and cooling. As the surface layer thickened, longer-wavelength structures carried earlier-formed structures "piggyback," preserving the earlier record. During the mechanical evolution of the ribbon-tessera terrain fabric, ductile strike-parallel shear zones developed locally (e.g., Hansen, 1992; Tuckwell and Ghail, 2003; Kumar, 2005; Romeo et al., 2005; Hansen and Tharalson, 2014). This model is referred to as the lava pond hypothesis for ribbon-tessera terrain formation, and it accounts for the entire structural and geologic history gleaned from detailed geologic mapping.

Bolide impact hypothesis for Coupled Ribbon Tessera–Crustal Plateau Formation

Formation of ribbon-tessera terrain as lava-pond scum does not address the fact that ribbon-tessera characterizes crustal plateaus, which stand up to 4 km above their surroundings, nor does it account for lowland inliers of ribbon-tessera. These relationships require a mechanism to: (1) form a lava pond with the areal extent matching that of individual crustal plateaus; (2) form and support regional-scale plateaus; and (3) address apparent plateau collapse. These additional criteria can be accommodated by a bolide impact hypothesis for combined ribbon-tessera–crustal plateau formation (Hansen, 2006).

Figure 2 illustrates a bolide impact hypothesis for coupled ribbon-tessera–crustal plateau evolution. Ribbon-tessera fabric developed at the top of a vast, near-surface melt volume formed as a result of a large bolide that pierced ancient thin lithosphere, causing massive high-*T*, high-fraction partial melting in the sublithospheric mantle (Fig. 2, time steps 1–3). Melt rose to the surface, forming a large-diameter lava pond (large igneous province scale), the surface of which evolved into ribbon-tessera terrain; mantle melt residue formed a low-density root responsible for plateau uplift (Fig. 2, time steps 2–3). If a solidified lava pond (ribbon-tessera) lost its root of melt residuum (Fig. 2, time steps 4a–6a), then the ribbon-tessera would subside to a mean planetary elevation, and subsequent surface flows would result in its progressive burial. If, however, ribbon-tessera remained coupled to its melt-residue root (Fig. 2, time steps 4b–6b), it could be preserved as a high-standing crustal plateau, escaping later burial (Fig. 2, time step 5b). Secular cooling and resulting thickening of the mechanical lithosphere would "lock" the low-density root in place, assuring preservation of that crustal plateau (Fig. 2, time step 6b). Thus, ribbon-tessera fabrics represent the progressively solidified "scum" of massive lava ponds, which individually represent igneous provinces recording the crustal component of crustal plateau formation; the mantle melt residue forms the plateau root. Even if detached, melt-residue root material would not be recycled into the deep mantle due to its low density.

The lava-pond–bolide-impact hypothesis for crustal plateau formation summarized here accounts for plateau scale, topography, gravity/topography signature, and documented deformation patterns and deformation and magmatic histories. Additionally, this model provides a mechanism for spatially distributed and temporally punctuated steady-state resurfacing (resurfacing refers to the destruction, or removal, of impact craters from a planet surface) at a time when Venus had a thin lithosphere that could be pierced by large bolides. Steady-state resurfacing by this mechanism can explain the critical constraints imposed by Venus' impact crater density, making moot any need for catastrophic resurfacing (Bjonnes et al., 2012). As Venus' lithosphere thickened, presumably as the result of secular cooling, large bolides that impacted the surface would no longer penetrate the lithosphere, but instead form large

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

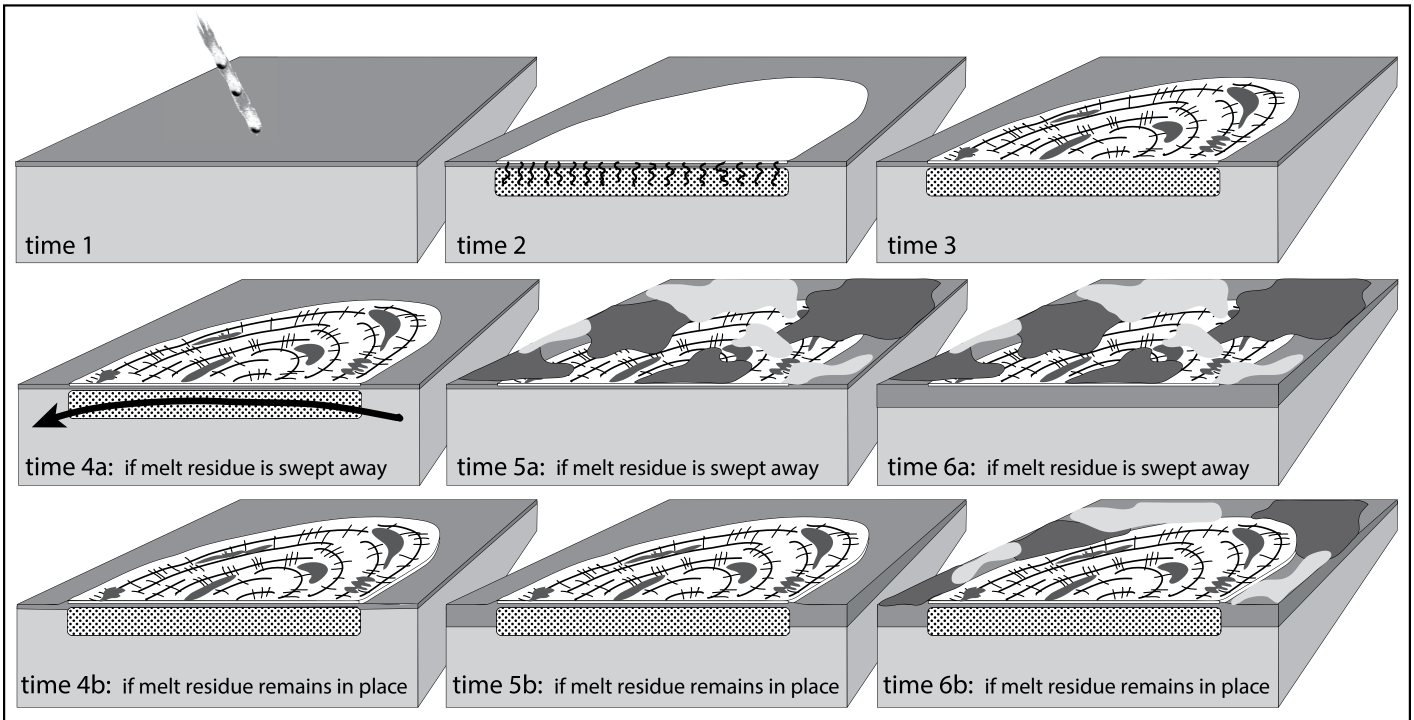

Figure 2. Cartoon sequence of block diagrams illustrating crustal plateau formation and collapse following a hypothesized bolide impact (Hansen, 2006). Top row, times 1–3: Block diagrams showing mantle and thin lithosphere—a global condition of ancient Venus. Time 1—A large bolide pierces through the lithosphere into the ductile mantle. Time 2—Massive partial melting occurs in the shallow mantle, induced by bolide impact; a lens of low-density melt residue (stippled) remains in the mantle as the melt escapes upward (wavy lines) to form a vast lava pond at the surface. Time 3—Convection in the lava pond results in surface deformation and local leaking of lava into topographic lows (dark gray), resulting in the formation of distinctive ribbon-tessera terrain tectonic fabric. Middle row, times 4a–6a: If the melt residue is swept away by mantle convection, the solidified lava pond (ribbon-tessera terrain) lies at mean planetary elevation; with time, the ribbon-tessera terrain becomes variably buried by younger surface flows. Note that continuity of ribbon-tessera fabric patterns between isolated exposures reveals the presence of shallowly buried ribbon-tessera terrain. Bottom row, times 4b–6b: If the low-density melt residue remains in place, the solidified lava pond (ribbon-tessera terrain) is elevated to crustal plateau stature (time 4b), and with secular cooling and thickening of the global lithosphere, the melt-residue root becomes "locked" in place, assuring long-time preservation of the ribbon-tessera terrain (time 5b); younger surface flows embay the edges of the high-standing crustal plateau, but the ribbon-tessera is protected given its high elevation (time 6b). If plate-tectonic processes had developed, the lowland lithosphere with ribbon-tessera terrain (time 6a) would have likely been recycled to the mantle; however, ribbon-tessera terrain of a high-standing crustal plateau, supported by a locked-in melt-residue root (time 6b) would not be subducted to the mantle, although it might become dismembered, translated, and reassembled. However, Venus did not evolve plate tectonics; therefore, both lowland ribbon-tessera terrains and crustal plateaus (times 6a and 6b) are preserved.

impact basins within the lithosphere, such as Mead crater (270 km diameter), the largest currently recognized impact basin on Venus.

Comparison of Venus Crustal Plateaus and Terrestrial GGTs–CLM

Venus crustal plateaus and terrestrial GGT–CLM (Archean cratons) share many similarities, offering the possibility that insight gained from one planet can inform us about its neighbor. Both crustal plateaus and Archean cratons represent ancient regional-scale features with unique and characteristic crustal components and low-density mantle roots. Venus lacks water-driven erosion processes, which are active on our own planet; as a result, ribbon-tessera terrains preserve a *surface* record of crustal plateau formation, whereas terrestrial GGTs preserve an eroded view within the crustal component of Archean cratons. Therefore, a comparison of these features must consider the currently exposed architectural level of present-day crustal plateaus and GGTs. Similarities between crustal plateaus and GGTs include the following features.

(1a) Both formed early in their respective planet's recorded history, and (1b) both are unique features that only formed during a limited (early) time frame, or eon. Thus, it is plausible that they each record a unique geodynamic process that was restricted, for whatever reason, to early planet evolution.

(2) Both features record extremely rich geologic histories, recording a progressive evolution of the specific terrain through time.

(3) The number of crustal plateaus and cratons with GGTs are on the order of ten or less on each planet.

(4) These features are likely of similar size and geometry. Although the original areal extent of individual GGTs is unknown, due at least in part to more recent dissection and dispersal, the Superior Province of northern Minnesota and Canada is broadly similar in size to that of a Venus crustal plateau.

(5) GGTs and crustal plateaus display gentle, long-wavelength folds (50–200 km; Abitibi Subprovince of the Superior Province—Thurston et al., 2008; Bédard et al., 2013, their Fig. 3; Yilgarn craton—Van Kranendonk et al., 2013, their Fig. 3; crustal plateaus—Ghent and Hansen, 1999, their Figs. 9–10; Hansen et al., 1999, their Figs. 2–4; Banks and Hansen, 2000, their Plate 1, Fig. 2; Hansen, 2006).

(6) Stratigraphic facing is preserved across GGTs and crustal plateaus (GGT—Thurston et al., 2008; crustal plateaus—Banks and Han-

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

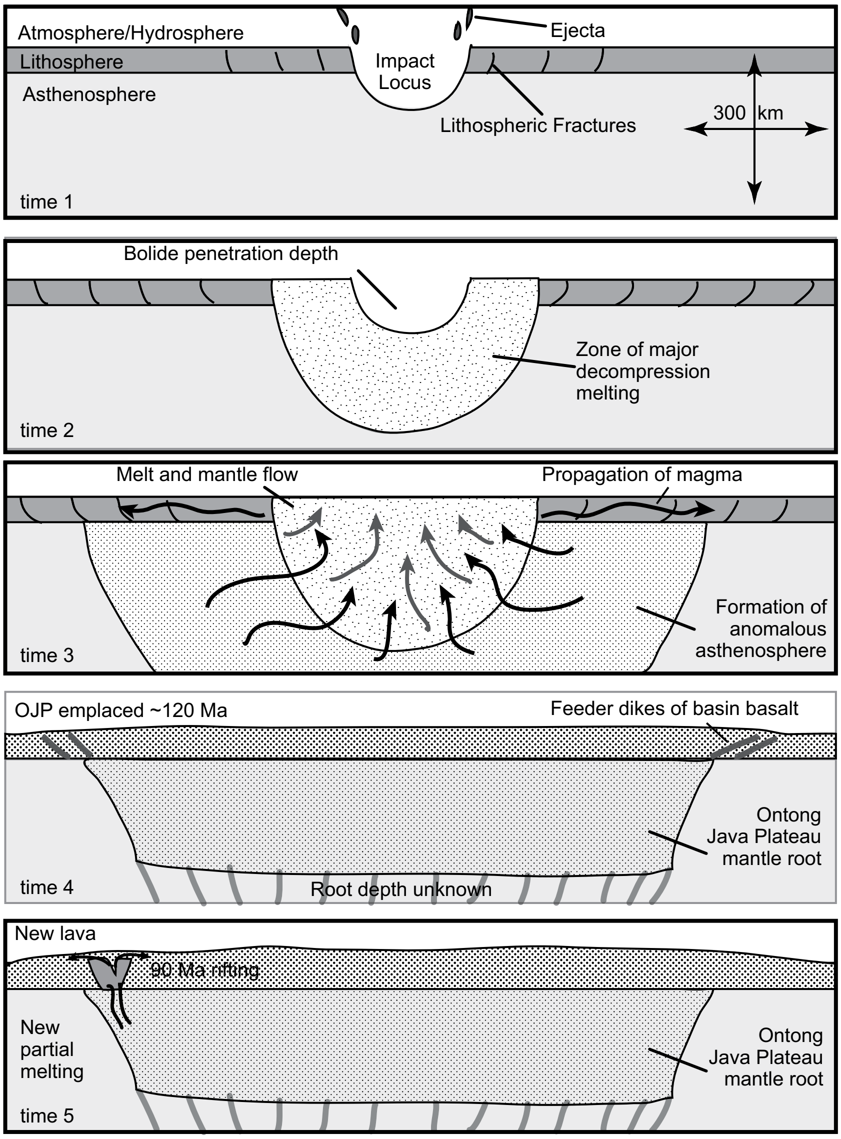

Figure 3. Time-step model for the formation of the Ontong Java Plateau (OJP) as a result of large-bolide impact on thin lithosphere (modified from Ingle and Coffin, 2004). Time 1—At the moment of impact, the uppermost asthenosphere is penetrated, the water column is vaporized, 20-m.y.-old oceanic lithosphere at the impact site is obliterated, and the surrounding lithosphere fractures. Time 2—During the moment of maximum penetration, the crater is completely formed, and the region of decompression melting is focused. Time 3—The void becomes in-filled from the bottom and sides; melt migrates along fractures in the lithosphere; refractory surrounding mantle fills space vacated by outflowing magma. Time 4—In the final stage of plateau formation, the ~35-km-thick crust lies above the deep, dry, refractory mantle root, representing the melt residue. Time 5—Later tectonism might result in localized volcanism.

sen, 2000; Hansen, 2006; Graupner, 2013; Slonecker, 2013).

(7) GGTs and crustal plateaus both preserve evidence of high-strain, strike-parallel ductile shear zones that commonly terminate along strike and/or merge with one another (GGTs—Bédard et al., 2013, their Fig. 3; Van Kranendonk et al., 2013, their Fig. 3; crustal plateaus—Hansen, 1992; Tuckwell and Ghail, 2003; Ignacio et al., 2005; Kumar, 2005; Romeo et al., 2005; Graupner, 2013; Hansen and Tharalson, 2014).

(8) Deformation accompanied magmatic activity.

(9) Although GGTs and crustal plateaus formed early in their planet's evolution, each survived, at least locally, to modern time.

(10) Coupled GGT–CLM formation and coupled ribbon-tessera–crustal plateau formation both call on high-*T*, high-fraction melting of the sublithospheric mantle, with melt rising to form the surface igneous province, and the melt residue forming a strong, buoyant root, which ultimately preserves the overlying surface igneous province (that is, GGT, or ribbon-tessera terrain). In the case of ribbon-tessera terrain, if its root were lost, it may be variably buried; in the case of GGT–CLM, if the CLM root were lost, the GGT would be recycled to the (deep) mantle by younger plate-tectonic processes.

Together, these features can be successfully explained by extrapolation of a bolide impact origin on Venus or Archean Earth.

BOLIDE IMPACT MODEL FOR COUPLED TERRESTRIAL GGT–CLM FORMATION

Petrologic modeling indicates that GGT–CLM could form as a result of high-*T*, high-fraction melting of the Archean mantle, in which melt rises to form large igneous provinces that evolve into GGTs, whereas melt residue forms the CLM root (Herzberg and Rudnick, 2012). I propose a new hypothesis of bolide impact to explain high-*T*, high-fraction melting in the sublithospheric Archean mantle and the coevolution of coupled GGT–CLM. In this bolide impact model, as in plume models, the high-*T* signatures of GGT non-arc basalt and komatiite are interpreted as an indicator of a hot thermal anomaly rather than as reflecting ambient mantle temperature. However, in the bolide impact hypothesis, the thermal anomaly driver is exogenic, rather than endogenic. Coupled GGT–CLM formation builds on the bolide impact hypothesis for Venus crustal plateaus presented earlier, in which a large bolide (≥20–30 km diameter) pierces the lithosphere into the sublithospheric mantle, causing high-*T*, high-fraction melting; the melt rises to ultimately form the crustal component of an Archean craton, the GGT, and the melt residue forms the associated CLM root.

Large-Bolide Impact on Thin Lithosphere

The bolide impact hypothesis requires a large bolide to pierce the Archean lithosphere and cause massive melting in the sublithospheric mantle. The physics of large impact events, in which a bolide travels into a planet's sublithospheric mantle, is not well understood (Reese et al., 2002); however, recent studies have begun to address what happens in such a scenario (e.g., Ingle and Coffin, 2004; Jones et al., 2002, 2005; Elkins-Tanton and Hager, 2005). A bolide's ability to puncture a planet's lithosphere depends on bolide mass and lithosphere thickness, with lithosphere thickness being a key factor. In the case

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

of thick lithosphere, even a very large bolide will not travel through the lithosphere—it will instead impact within the lithosphere, forming an enormous impact basin (e.g., Ivanov et al., 2010). However, a large bolide might penetrate relatively *thin* lithosphere, as is considered here.

The potential effects of large-bolide impact on thin lithosphere are illustrated by considerations of the Ontong Java Plateau, which has been compared to GGTs in both size and basalt composition. Numerous authors have postulated that the greater Ontong Java Plateau represents the surface signature of a deep mantle plume (e.g., Tarduno et al., 1991; Bercovici and Mahoney, 1994; Coffin and Eldholm, 1994; Farnetani and Richards, 1994; Farnetani et al., 1996; Ito and Clift, 1998; Ito and Taira, 2000). However, Ingle and Coffin (2004) cited a host of geochemical and geophysical data to argue that the Ontong Java Plateau formed as the result of massive partial melting in the mantle caused by impact of a large bolide on relatively young (hence thin) oceanic lithosphere. Following earlier suggestions (Rogers, 1982; Price, 2001), Ingle and Coffin (2004, p. 123) stated that the "absence of an obvious hotspot source or track, minor crustal uplift associated with emplacement, minor total subsidence compared with normal oceanic crust or other oceanic plateaus and submarine ridges, high degrees of melting at shallow, upper mantle depths, low water contents of basalts, enrichment of platinum group elements in basalts, and a ~300 km deep, seismically slow mantle root are more consistent with the consequences of an impacting bolide." In their model (reproduced here as Fig. 3), a large bolide (20–30 km in diameter) pierces 20-m.y.-old (50-km-thick) ocean lithosphere, causing high-*T*, high-fraction partial melting; melt rises to form anomalously thick (35 km) basalt-rich crust, whereas melt residue forms a buoyant mantle root, which supports the large plateau. Primary magma compositions of the Ontong Java Plateau are indistinguishable from Archean and Proterozoic non-arc basalt (Herzberg and Rudnick, 2012), and the extraction of such melt from the mantle would generate a dry, low-density, highly depleted mantle melt residue, comparable to Archean CLM (Ingle and Coffin, 2004).

Figure 3 illustrates possible stages in the evolution of a large-bolide impact on thin lithosphere. Although much is unknown about the effects of a large bolide on thin lithosphere, existing modeling is consistent with a proposition that the Ontong Java Plateau might represent such an event (Jones et al., 2002, 2005; Elkins-Tanton and Hager, 2005). What is critical to the discussion here is not whether the Ontong Java Plateau actually formed by bolide impact, but rather that it serves as an example by which to consider the possible effects of a large bolide that has penetrated thin lithosphere. Notable features apparent in Figure 3 include: (1) the location of bolide penetration is small compared to the regional extent of the later-formed igneous province; (2) the area involved in high-*T*, high-fraction melting is large compared to the bolide impact locus and the zone of major decompression melting; (3) the area over which mantle melt residue occurs is large compared to the impact locus, and it essentially underlies the region covered by the "final" surface igneous province; (4) much of the "original" lithosphere would remain intact, with the mantle-derived melt flowing onto pre-impact lithosphere (time 2 and 3); (5) the surface igneous province and the mantle root are genetically related and spatially correlated; and (6) the lithosphere distal to the surface igneous province would remain unchanged.

Critical factors that contribute to melt formation include: (1) lithosphere thickness, where thinner lithosphere leads to greater melt production; (2) potential temperature of the mantle, with higher temperature increasing melt production; (3) bolide size, in which larger bolides yield greater total melt volume; (4) bolide impact velocity; and (5) assumptions regarding melt formation and melt crystallization (Jones et al., 2002, 2005; Elkins-Tanton and Hager, 2005). Modeling by Elkins-Tanton and Hager (2005) indicated that a 20-km-diameter bolide, 70–50-km-thick lithosphere, and 1450 °C mantle potential temperature resulted in high-fraction melting (51%–64%), a maximum melt depth equivalent to 3.5–4 GPa, in situ melt volume of $2.0–2.6 \times 10^6$ km^3, and total melt volume of $3.1–5.1 \times 10^6$ km^3. A 30-km-diameter bolide with similar lithosphere thickness and mantle temperature resulted in even higher-fraction melting (63%–75%), greater maximum melt depth (4.0–4.5 GPa), and larger in situ ($7.4–9.4 \times 10^6$ km^3) and total ($1.0–1.5 \times 10^7$ km^3) melt volume—comparable to the crustal volume of the Ontong Java Plateau. Mantle potential temperature had the greatest effect on melt volume. Melt volumes increased by a factor of 2–3 with an increase of 50 °C, holding other parameters constant. Again, whether or not the Ontong Java Plateau formed as a result of large-bolide impact is not critical to the discussion here; its formation remains a topic of debate (Tejada et al., 2012). What is relevant here is that a plausible bolide impact model for the Ontong Java Plateau can serve as a conceptual starting point for what might happen when a large-bolide impacts thin lithosphere. Furthermore, Ingle and Coffin's (2004) model of Ontong Java Plateau formation illustrates a first-order sequence of events during such an impact.

Large-Bolide Impact during the Archean

To apply lessons from modeling of a large-bolide impact on thin lithosphere to the formation of Archean cratons, we must consider plausible values for the critical factors of mantle potential temperature, lithosphere thickness, and bolide size (and/or mass) and velocity.

Temperature of the Archean Mantle

First, Archean mantle potential temperature is debated, as indicated earlier; however, it is agreed that mantle potential temperature was higher in the Archean than today. Plausible estimates range up to 1450–1750 °C (e.g., Korenaga, 2008a; Davies, 2007b; Lee et al., 2011), or 100–300 °C greater than the temperatures explored in the impact models discussed here.

Thickness of the Archean Lithosphere

Second, Archean lithosphere thickness is unknown. Lithosphere thickness depends on mantle temperature, mantle viscosity, and mechanisms of heat transfer, all of which are unknown for the Archean. Lithosphere thickness is, to a first-order, a function of mantle temperature, with thinner lithosphere at higher temperatures (Parsons and McKenzie, 1978). The potential temperature of the Archean mantle, as noted earlier, is debated. However, even relatively low estimates for the Archean ambient mantle temperature result in a plausible lithosphere thickness of ~70 km (Korenaga, 2008b). Mantle viscosity, a function of mantle temperature, also contributes to Archean lithosphere thickness; modeling that considers the effect of viscosity yields plausible average values of 20–35 km, many times thinner than contemporary oceanic lithosphere (Davies, 2006, 2007b). Although the crust may comprise a large portion of the resulting lithosphere, the critical factor here is lithosphere thickness (not crust thickness) over regional-scale tracts of the Archean Earth surface. Proposed heat-transfer mechanisms for the Archean range from plate-tectonic convection, to episodic plate tectonics, to lithosphere conduction, to heat-pipe models, among others (e.g., O'Reilly and Davies, 1981; Davies, 2006; Korenaga, 2006; O'Neill et al., 2007; Stern, 2008; Moore and Webb, 2013), but they are not mutually exclusive. Stated another way, different mechanisms could have operated simultaneously. Different regional tracts of the Archean lithosphere might have transferred heat via different mechanisms, and thus lithosphere thickness could have varied significantly across the planet. Furthermore, proposed values for lithosphere thickness can vary within the context of a proposed single mechanism; for example, assuming a plate-tectonic mechanism of heat transfer can yield quite different values for litho-

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

sphere thickness, including lithosphere thicker or significantly thinner (~20–25 km thick) than contemporary lithosphere (e.g., Korenaga, 2008b; Davies, 2007a, 2007b). Even within the context of a single suite of assumptions, Archean lithosphere thickness likely varied significantly (Davies, 2007b). In summary, there is no consensus with regard to Archean lithosphere thickness; plausible values vary from as low as 20 km to as much as 50–70 km.

Nature of Bolide Impacts

Bolide impact events were both larger and more frequent in the past than they are today, with a peak in the impact cratering record known as the Late Heavy Bombardment (LHB; Cohen et al., 2000). This spike—also referred to as the lunar cataclysm, because it was initially proposed to address the formation of young, large (>300-km-diameter) lunar basins—is a hypothetical event in which an anomalously large number of large bolides (≥20–50 km diameter) collided with the terrestrial planets of the inner solar system from 4.1 to 3.8 Ga. The cause, and even the occurrence, of the LHB is debated (e.g., Koeberl, 2006); data from NASA's *GRAIL* mission indicate that impact bombardment of the early solar system likely warrants reevaluation because crust of the near side of the Moon, which hosts more large impact basins than the far side, is enriched in heat-producing elements relative to the far side (Miljković et al., 2013). Therefore, near-side crust and upper mantle were hotter during impact basin formation; as a result, bolide impact studies based on near-side data overestimate the size of ancient large impact basins, and therefore also overestimate bolide size.

Regardless of these uncertainties, two recent studies postulate that the terrestrial effect of the LHB lasted through the Archean Eon (Fig. 4).

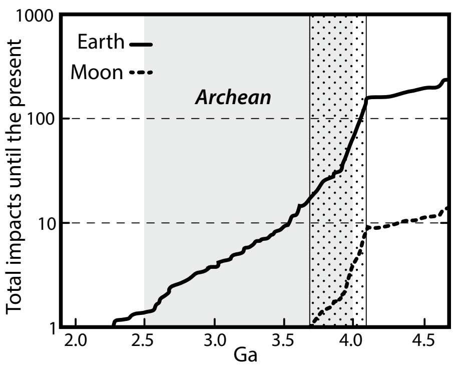

Figure 4. The number of large-bolide basin-forming impact events through time for the Moon and Earth based on modeling of Bottke et al. (2012). The conventional Late Heavy Bombardment, as derived from lunar basins, is shown in the dotted pattern. Figure is modified from Thompson (2012).

On the basis of computer simulations, Bottke et al. (2012) argued that most of the large late impactors came from the so-called E belt, located 1.7–2.1 AU (astronomical units) from the Sun (now mostly extinct). Their model results indicate that E-belt bolides colliding with the Moon resulted in the formation of ~10 lunar basins between 4.1 and 3.7 Ga, and four Chicxulub-sized basins (~180 km diameter) between 3.7 and 1.7 Ga. By comparison, their model predicts that Earth was struck by 15 ± 4 large bolides (>20–30 km diameter) between 3.7 and 2.5 Ga, and ~70 Chicxulub-sized (~10 km diameter; Covey et al., 1994) or larger bolides between 3.7 and 1.7 Ga. Johnson and Melosh (2012) independently estimated the sizes and impact velocities of bolides responsible for the formation of spherule layers preserved in Archean GGT strata, formed due to bolide impact events (Lowe et al., 2003; Simonson and Glass, 2004; Glass and Simonson, 2012). The age of the spherule layers, and thus the bolide impact events, are well constrained. It is evident, therefore, that large bolides impacted Archean Earth. In total, 11 Archean spherule layers (ages ranging from 3.47 to 2.46 Ga) were modeled to estimate bolide diameter and impact velocity (Johnson and Melosh, 2012). Estimates of bolide diameter range from 11 to 70 km; the smallest bolide (11–18 km diameter; 22.3–23.8 km/s velocity) corresponds to the youngest event at 2.46–2.52 Ga; the largest bolide (41–78 km diameter; 20.6–22.8 km/s velocity) correlates with a 3.24 Ga event. Estimates of impact velocity are similar to those assumed in the bolide impact models described previously. These results, together with modeling of Bottke et al. (2012), indicate that the LHB affected Earth well beyond the conventional time frame postulated from lunar studies, and furthermore that Earth experienced a flux of large bolides lasting through the Archean Eon (Fig. 4). This result is not particularly surprising given differences in mass and surface area of Earth and the Moon. In short, it is plausible that Earth was showered by large bolides during the whole of the Archean, at the same time Earth's global lithosphere was likely to have been relatively thin.

Effects of Large-Bolide Impact on Thin Archean Lithosphere

We can consider the amount of melt produced within the context of published models of large-bolide impact given the Archean parameters discussed earlier herein. Recall that mantle potential temperature has the greatest effect on the volume of melt. Given an Archean lithosphere on the order of 20–40 km thick (Davies, 2006, 2007b) or even 70 km thick (Korenaga, 2008b), impact by a large bolide, on the order of 20–40 km in diameter, with a mantle potential temperature of 1450 °C would produce the melt and melt residue as seen in GGT–CLM provinces. Elkins-Tanton and Hager (2005) did not explore the effects of a >30 km bolide on lithosphere thinner than 50 km; however, if we consider a 30 km bolide on 50-km-thick lithosphere and mantle potential temperatures of 1500 °C, 1550 °C, 1600 °C, and 1650 °C, expected total melt volumes would be 3–4.5 × 10^7 km^3, 6–13.5 × 10^7 km^3, 12–40 × 10^7 km^3, and 24–122 × 10^7 km^3, respectively. A 50-km-diameter bolide impacting 70-km-thick lithosphere, over the same range in mantle temperature, results in total melt volumes ranging from 2–3 × 10^7 km^3 (1500 °C) to 16–81 × 10^7 km^3 (1650 °C). Mantle temperature of 1750 °C would yield even higher melt volumes.

Thus, it is plausible that large bolides might have significantly impacted the lithosphere during the Archean Eon, and it is further possible that such events could produce surface igneous provinces on the scale of GGTs and associated sublithospheric melt-residue roots. High-T partial melting, high-fraction melting, and maximum melt depths are consistent with petrologic constraints (e.g., Herzberg and Rudnick, 2012). The bolide impact hypothesis for coevolution of GGT–CLM that formed Archean cratons therefore meets the same criteria defined by both the ocean-ridge and plume models. In short, a bolide impact model requires large bolides and thin lithosphere, both of which are expected during the Archean Eon.

COMPARISON OF MODELS FOR COUPLED GGT–CLM FORMATION

Three different comprehensive geodynamic models for coupled GGT–CLM formation—ocean-ridge, plume, and bolide impact hypotheses—account for high-T, high-fraction partial melting of the Archean mantle indicated by thermodynamic and petrologic modeling (Herzberg and Rudnick, 2012). Each model makes predictions, or has implications, based on the interpreted mechanism of coupled GGT–CLM formation (Table 1; Figs. 1 and 2). A brief comparison of these models, with particular emphasis on their differences, is listed in Table 1.

Ocean-Ridge and Hot-Ambient-Mantle Model

In the ocean-ridge and hot-ambient-mantle model (Fig. 1A; Table 1), GGT–CLM result from ocean-ridge magmatism within a hot-ambient-mantle environment. Thick (30–45 km) mafic crust forms along divergent-plate bound-

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

TABLE 1. COMPARISON OF THREE GEODYNAMIC MODELS FOR COUPLED GGT–CLM FORMATION

Model	Non-arc basalt formation	GGT komatiite flow formation	GGT felsic unit formation	Addresses GGT inheritance	GGT built in part on felsic crust	Plan view of GGT	Areal extent of GGT	Addresses limited modern GGT occurrence	Is Archean ambient mantle T important?	"Required" Archean ambient mantle T (°C)	Requires Archean plumes	Requires Archean plate tectonics	Can account for density inversion of crust
Ocean ridge/hot ambient mantle	Yes	Not directly, but yes, if plumes	Would require special setting?	No	No	Planar "plate"	Near global	No	Yes, higher is better	~1600	Yes, for komatiites	Yes	No?
Plume	Yes	Yes	Yes	Yes	Yes	Quasi-circular	~1000–2500 km diameter*	Yes	Yes, lower is better	~1490†	Yes	No	Yes
Bolide impact	Yes	Yes	Yes	Yes	Yes	Quasi-circular	~500–>2500 km diameter§	Yes	No	>1450#	No	No	Yes

Note: Accretionary (I) and hybrid (IV) models are not considered here; CLM—cratonic lithospheric mantle; GGT—granite-greenstone terrain; T—temperature.
*Diameter is a function of plume-head radius, which depends on thermal buoyancy of the plume relative to the adjacent mantle.
†May not work well at high ambient mantle temperatures given a required thermal difference of 100–300 °C between plume and adjacent mantle.
§The lower limit depends on lithosphere thickness, bolide size, impact velocity, and ambient mantle temperature; a wide range of sizes could result.
#Could accommodate a wide range of temperatures, from <1450 °C with very large bolides, to >1650 °C.

aries, forming regions of areally extensive GGT and complementary CLM. The supracrustal non-arc basalt would be grossly homogeneous in character, similar conceptually to modern oceanic lithosphere. This model is attractive in that it calls on familiar geologic processes, although within an elevated thermal environment. However, the model does not account for inherited crustal components in some GGTs, those built on felsic crust. Crust formed at an ocean ridge would be juvenile and free of crustal inheritance. In addition, evolution of GGT felsic units might require special petrologic circumstances, given their rarity in ocean-ridge environments.

Despite the appeal of this model for explaining coupled GGT–CLM formation, it appears problematic based on the distribution of Archean cratons signified by GGTs. If these features formed at anomalously warm ocean ridges, a near global distribution of 30–45-km-thick non-arc basaltic crust would be expected, yet this is not observed, even after long stability of cratonic lithosphere. A fundamental question, highlighted by the authors (Herzberg and Rudnick, 2012), is what became of the rest of this thick basaltic crust? If formed globally, how could such extensive tracts of GGT be destroyed? Addressing these questions becomes critical to the viability of the model.

An ocean-ridge model also might not adequately address synchronous formation of non-arc basalt and komatiite flows in GGTs. GGT non-arc basalt forms within an extremely hot ocean-ridge setting, yet mantle plumes are invoked to generate interlayered komatiite flows. Given that mantle plumes are driven by thermal buoyancy as a function of a thermal difference compared to the adjacent mantle (nominally ~100–300 °C; Griffiths, 1986; Campbell and Griffiths, 1990; White and McKenzie, 1995), the hotter the ambient mantle, the less likely it would be for plumes to develop. Therefore, this model requires a mechanism to generate extremely hot Archean mantle plumes, with temperatures significantly above the interpreted high ambient-mantle temperature. In addition, calling on Archean mantle plumes raises the question: How might Archean plumes and ocean-ridge magmatism interact in order to preserve interlayered komatiite and non-arc basalt flows? As a related consequence, if plumes are required for komatiite flows, then within the context of this model, plumes must have occurred everywhere along Archean ocean ridges where GGTs formed. Alternatively, the model requires a mechanism by which preserved GGTs (Archean ocean crust that escaped widespread destruction) were somehow the same GGT crust that intersected with locally formed plumes. The ocean-ridge model therefore requires the coincident occurrence of both Archean plate tectonics and deep mantle plumes.

Plume Models

Plume models (Fig. 1B; Table 1) result in local, spatially isolated formation of individual GGT–CLM features. These models successfully address interlayered GGT non-arc basalt and komatiite flows. Given that a mantle plume could form beneath previously formed crust, this model can also account for: GGTs built on felsic crust, providing evidence of GGT crustal inheritance, and formation of GGT felsic units. The plume model can also explain limited preservation of coupled GGT–CLM because GGT–CLM forms as isolated features. Therefore, a CLM root could be stripped away by mantle convection, leaving its GGT rootless and subject to subsequent subduction. Plume models may, however, have limiting weaknesses. Lee (2006) pointed out that the high-degree melt extraction required for continental mantle lithosphere currently occurs only in localized regions of contemporary plumes; therefore, CLM formation by plumes would require larger and hotter Archean plumes. Indeed, plume advocates argue for high-T Archean plumes, so this alone is not problematic (Fig. 1B; Campbell and Griffiths, 2014). However, the occurrence of hot Archean plumes could be limited if ambient mantle temperature was indeed much higher in the Archean (≥1600 °C). A plume model does not require Archean plate tectonics, nor is it incompatible with Archean plate tectonics; it does, of course, require the occurrence of Archean mantle plumes.

Bolide Impact Model

The bolide impact model (see Figs. 2 and 3; Table 1) is similar to plume models in that it calls on a local mantle thermal anomaly as a driver for high-T and high-fraction partial melting to form GGT and its associated CLM. However, in the bolide impact hypothesis, the thermal anomaly results from a large exogenic bolide that pierces thin lithosphere. By this model, partial melting might take place at lower pressure in the mantle than that predicted by a plume model, which would influence the resulting petrology. In the bolide impact model, parts of the mantle could experience extremely high-fraction partial melting (≥50%) to form komatiite melts, and elsewhere lower degrees of melting (30%–45%) could lead to formation of non-arc basalt melts. The resulting melt residues would collectively form the CLM. Depths of melting predicted by bolide impact models (e.g., Elkins-Tanton and Hager, 2005) are consistent

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

with depths estimated for CLM (see references cited earlier herein). In addition, melting would be extremely fast and, as a result, the process(es) of melt removal might lead to diapiric rise of a crystal/melt mixture, in turn resulting in further decompression melting (e.g., Green, 1972). Details of the process(es) associated with high-*T*, high-fraction melting due to large-bolide impact are currently unknown, and they are beyond the scope of this contribution. Indeed, as noted by Reese et al. (2002), the physics of large impact events in which the bolide travels into a planet's sublithospheric mantle is not well understood, and the effect on melt extraction would be further complicated by the physics of impact. However, we can surmise that the catastrophic generation of ≥45% melt and rapid escape of this melt could lead to the formation of a large igneous province at the surface that might undergo further melt evolution and differentiation, as well as development of a corresponding root of melt residue.

Because an incoming bolide could impact existing felsic or mafic crust, this model can effectively explain GGTs built on felsic crust, local GGT crustal inheritance, and formation of GGT felsic units. This model may be quite successful at forming a wide range of igneous units within GGTs, but that would require additional geodynamic and petrologic modeling well beyond the scope of this work. As with the plume model, spatially and temporally punctuated formation of GGT–CLM would allow for possible decoupling of GGT and CLM, and therefore a mechanism by which some GGTs could be recycled yet preserve the complementary CLM within the mantle.

Formation of Archean GGTs by bolide impact also explains their intrinsic structural and petrologic features, interpreted to be the result of density inversion–driven tectonics, or so-called crustal convective overturn (McGregor, 1951; Anhaeusser et al., 1969; Mareschal and West, 1980; Goodwin, 1982; Chardon et al., 1996, 1998, 2002; Choukroune et al., 1997; Collins et al., 1998; Bédard, 2003, 2006; Sandiford et al., 2004; Rey and Houseman, 2006; Van Kranendonk et al., 2004; Robin and Bailey, 2009). As with plume models, bolide impacts provide a mechanism by which GGT crust may have developed an inverted density structure, with lower-density, more felsic-rich material at depth, and higher-density basalt and komatiite occurring at higher eruptive/stratigraphic levels.

Implications of Archean Bolide Impact

A bolide impact model is appealing in its plausibility of both mechanism and outcomes. It integrates basic parameters of higher bolide flux, thinner lithosphere, and a warmer mantle geotherm. It is worth considering how such a proposed mechanism might have played out on the other terrestrial planets—Mars and Mercury. First, it is notable that continent-like features are unknown on either Mars or Mercury, much less features that might be conceptually correlative to Archean cratons. Although this might change with further planetary exploration, future recognition of continental lithosphere on these planets is unlikely given the resolution of data currently available. Second, a key requirement of the bolide impact hypothesis is the occurrence of thin lithosphere covering a sufficiently large fraction of a planet surface so as to provide a sizeable target region for large bolides and subsequent development of crustal plateau–like features, which on Venus are on the order of 1400–2400 km in diameter. Mars and Mercury are both significantly smaller than Earth and Venus, and as such these planets lose heat much more quickly, resulting in the formation of thick lithosphere early in their history. Although much remains unknown about Mercury, its surface is highly cratered and appears of generally homogeneous character. In the case of Mars, it is clear that Mars possessed a thick lithosphere early in its evolution. As one example, the Hellas impact basin (~2300 km diameter, ~7.1 km deep) formed on thick, rather than thin, lithosphere (Ivanov et al., 2010). Mars does offer clues, however, in the form of the Tharsis Province, a vast volcanic plateau. Although the formation of Tharsis is debated, it has been proposed that Tharsis marks the surface expression of a large bolide that impacted through ancient Mars lithosphere into the mantle, ultimately generating a mantle-scale plume with an extended geologic history on the order of ~1 b.y. or more (Reese et al., 2002, 2004; Golabek et al., 2011; Roberts and Arkani-Hamed, 2014). One possible lesson from Mars is that although initial impact of a large bolide might be a rapid catastrophic event, a long and complex evolutionary history might ensue from a bolide that penetrates through the lithosphere into the mantle.

A bolide impact model for terrestrial GGT–CLM formation can account for observations in the rock record that the ocean-ridge and plume models might not address. (1) An impact model considers the possible effect of the LHB on early Earth. (2) This model specifically addresses why Archean GGT–CLM features are unique (requiring both thin lithosphere and large bolides). (3a) This model can explain why high-*T* komatiites are limited to the Archean, requiring extreme high-fraction melting of the mantle. (3b) Furthermore, it is possible that high-*T* komatiite-flow formation requires catastrophic melting of the mantle (e.g., Green, 1972), as addressed by this model. An origin by bolide impact is consistent with (4) evidence that the Archean Eon was a time of major production and preservation of continental lithosphere as compared to the Hadean and the Proterozoic; (5) a lack of evidence for Archean ocean spreading; and (6) inferences that the Archean mantle was thermally heterogeneous (e.g., Kamber, 2015). (7) This model can provide a mechanism by which heat-producing elements are extracted from the mantle and concentrated in the crust by a uniquely Archean process. (8) A bolide impact model might address why many GGTs are unusually rich in mineral resources (Begg et al., 2010), resulting from extreme melt conditions triggered by the energy input of a large extraterrestrial object, which could introduce a highly enriched, unique heterogeneity to the "local" mantle environment. (9) This model can account for the proposed sharp dichotomy between Archean GGT–CLM and younger terrains (Griffin et al., 2009a, 2013), given that large-bolide flux waned into the Proterozoic. (10) The hypothesis presents a mechanism by which Archean cratons could develop remotely from one another (that is, outside isotopic influence from other crustal packages), and/or how early continental crust could have developed in a spatially limited fashion, rather than as regionally extensive tracts, thereby eliminating the requirement to destroy large volumes of early-formed crust, given the limited ancient rock record (Kemp et al., 2015). (11) This hypothesis can account for the lack of shocked minerals in the Hadean and Archean, as discussed later.

Herzberg and Rudnick (2012) highlighted two essential observations with regard to mantle-melt residue over time. First, there has been a significant reduction in the Mg number of residues produced over the past 3 b.y., reaching a maximum at 2.5–3.5 Ga, when melt temperature and melt fractions were at a maximum. Second, residue density reached a minimum at the same time. The bolide impact model can address both of these secular changes because melts produced by large-bolide impact during the LHB, which waned on Earth at ca. 2.5 Ga (e.g., Bottke at al., 2012; Johnson and Melosh, 2012), would form as a result of extreme high-*T*, high-fraction melting of the mantle, with melting occurring in a spatially and temporally punctuated fashion throughout the Archean.

The bolide impact hypothesis might also explain the lack of shocked minerals in Hadean and Archean rocks. Shocked minerals, such as zircon and quartz, are a strong indication of large-bolide impact (French and Koeberl, 2010; Koeberl et al., 2012; Langenhorst and Deutsch, 2012), yet shocked zircons have not been identified in detrital Hadean populations (e.g., Ca-

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

vosie et al., 2007) or in Archean spherule deposits (Byerly et al., 2002). In contrast, shocked zircons are recognized in Proterozoic deposits (e.g., 2.02 Ga Vredefort and 1.87 Ga Sudbury impact structures; Cavosie et al., 2010; Thomson et al., 2014). The oldest known shocked mineral is a single tiny (~100 µm) quartz grain in a 2.64 Ga spherule layer in Australia (Rasmussen and Koeberl, 2004). Although seeming to the contrary, large-bolide impacts on thin Archean lithosphere should not yield widespread formation of shocked minerals. The formation of shocked minerals indeed requires a large bolide; however, shocked minerals form from the target material, not the bolide (French and Koeberl, 2010; Langenhorst and Deutsch, 2012). Shocked mineral formation also requires a thick target lithosphere, and that target lithosphere should contain abundant mineral species of the appropriate type. For example, the formation of shocked quartz requires a quartz-rick target crust built on thick lithosphere, representing continental rather than oceanic (i.e., mafic) lithosphere (Rasmussen and Koeberl, 2004). Given these requirements, the lack (or paucity) of shocked minerals of Archean age might be predicted by the bolide impact hypothesis, which requires large bolides and thin lithosphere. The impact of a large bolide on thick lithosphere forms a vast impact crater basin, similar to that of Vredefort, Sudbury, the large lunar impact basins, or Mars' Hellas basin (Ivanov et al., 2010), in which shocked minerals form in the target crust/lithosphere. However, the mechanism presented herein calls for a large bolide that pierces through thin lithosphere into the sublithospheric mantle. Thin lithosphere is a key component of the hypothesis. As a large bolide penetrated into the sublithospheric mantle, the bolide would fail to form a vast impact crater basin (Fig. 3, time 1). The scar in the lithosphere would be relatively small, and if shocked minerals were formed, their abundance would be quite small compared to a similar-sized bolide impacting thick lithosphere. Furthermore, such a situation would result in high-*T*, high-fractional partial melting of the sublithospheric mantle, producing a melt that rises to the surface and buries the bolide entry point (Fig. 3, compare times 1 and 4). Because the target in this case is sublithospheric mantle, not crust, the resulting melting process would not produce shocked minerals. Thus, it is unlikely that shocked minerals would form during large-bolide impact on thin lithosphere, and if limited parts of the crust formed shocked minerals, such minerals would rarely be preserved.

Small impact events, in which a bolide forms a classic impact basin in a thin lithosphere, might indeed result in the formation of shocked minerals. However, smaller impact events would also yield relatively small amounts of shocked minerals that, in turn, would be much less likely to be preserved. The probability of producing shocked quartz and zircon is further reduced if bolides impacted mafic lithosphere. In addition, mafic lithosphere that lacked a CLM root would likely be recycled to the mantle once plate tectonics became a major mechanism of heat transfer on Earth.

We can turn the prediction around and conclude that the lack of Hadean and Archean shocked minerals may provide clues about early Earth. As noted earlier, the thickness of the Archean lithosphere is unknown; Archean lithosphere could have been thick, or thin, or marked by gradations in thickness across the planet. We simply do not know. However, it is noteworthy that, to date, there are few examples of Archean-age shocked minerals, in contrast with a relative abundance of Proterozoic shocked minerals. Shocked minerals might therefore serve as a proxy for lithosphere thickness, such that formation and preservation of shocked minerals requires both thick lithosphere and large bolides. The thinner a planet's lithosphere, the smaller is the bolide required to produce an impact basin. Thus, in the case of globally thin lithosphere, the occurrence of shocked minerals should be low. In short, the frequent occurrence of Proterozoic-age shocked minerals could reflect the presence of a thick, relatively stable continental lithosphere, contrary to that in the Archean Eon.

Earth and Venus Comparison

Venus and Earth were likely most similar during the early years, each hosting relatively thin global lithospheres and peppered with large bolides (Fig. 5). Extremely large bolides would likely completely disrupt any global lithosphere, whereas small bolides would have little lasting

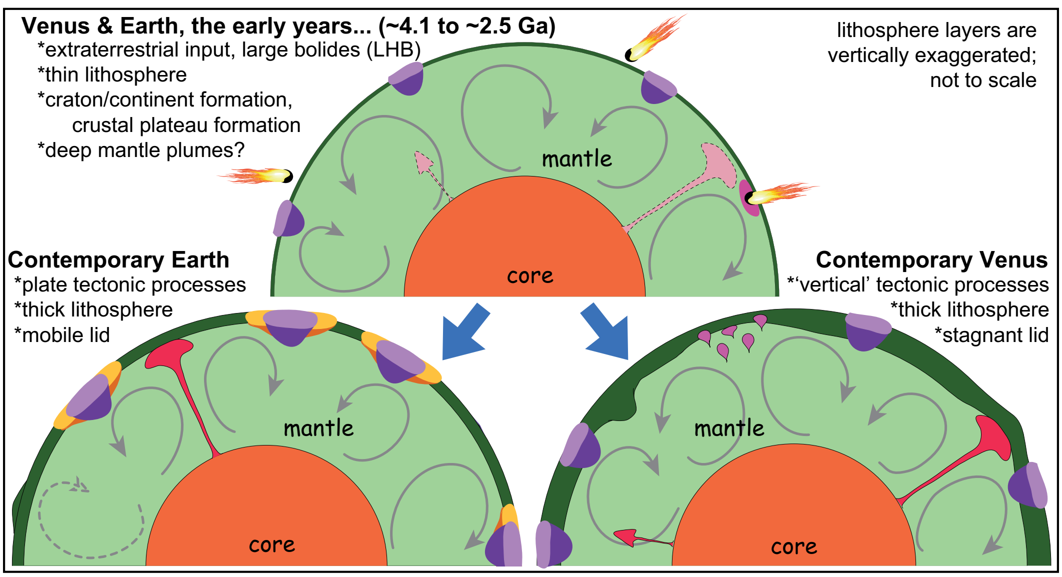

Figure 5. Cross-sectional cartoons of Venus and Earth illustrating endogenic and exogenic geodynamic processes through time. From 4.1 to 2.5 Ga, Earth and Venus were most similar; both planets could have been host to a relatively thin global lithosphere and showered with large bolides during the Late Heavy Bombardment (LHB); this could have led to the formation of crustal plateaus on Venus and analogous GGT–CLM on Earth. After 2.5 Ga, the large-bolide influx waned, and each planet cooled. Venus and Earth followed different evolutionary paths; Earth developed plate tectonics, and Venus developed a thick stagnant lid. Crustal plateaus (Venus) and GGT–CLM features formed between 4.1 and 2.5 Ga were variably preserved, depending on whether or not the melt-residue root became decoupled from its overlying terrain (see Fig. 4). In the terrestrial case, if a GGT became decoupled from its CLM root, the GGT would likely be subducted and recycled to the mantle during subsequent development of plate-tectonic processes. Alternatively, it might be dismembered and partially accreted to coupled GGT–CLM, forming the cratonic seeds of Earth's continents. On Venus, any ribbon-tessera terrains that lost their melt-residue root would reside at mean planetary elevation (not shown in the cartoon).

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

effect. Yet, the nexus of bolide-lithosphere interaction might result in a truly first-order effect on the host planet—the formation of crustal plateaus (Venus) or Archean cratons (Earth). Let us consider that GGT and CLM form by large-bolide impact on relatively thin Archean lithosphere. The number of GGT–CLM terrains that formed would be a function of the number of large bolides that showered Earth in this critical time period. Similarly, the number of crustal plateaus that formed on Venus would depend on bolide flux. If the LHB did in fact occur, this impact flux would have affected both Earth and Venus. Therefore, it is reasonable to apply Archean bolide sizes and velocities derived for Earth to Venus. Large-bolide impacts on Venus could have resulted in the formation of crustal plateaus and lowland ribbon-tessera terrain inliers, and the impact of large bolides on Earth, as proposed here, could have resulted in the formation of GGTs–CLM. Although coupled GGT–CLM formation would allow for preservation of a GGT, such preservation is only assured if a GGT remained coupled to its CLM, which would be, in a sense, its unique buoyant life preserver. If decoupled from its corresponding CLM, GGT crust would be subjected to later subduction. If on Venus a ribbon-tessera terrain remained coupled with its melt-residue root, it would survive as a crustal plateau, escaping burial; however, if a ribbon-tessera terrain became detached from its root, it would be subjected to local burial. Venus preserves a record of this partial loss or modification process for the simple reason that plate tectonics never evolved on Venus. If plate tectonics had evolved on Venus, then the lowland ribbon-tessera terrains, buried or not, would almost certainly have been recycled into the mantle; in addition, Venus' coherent high-standing crustal plateaus would likely have been dismembered, dispersed, or amalgamated. However, plate tectonics did not evolve on Venus, and, as a result, this planet preserves potentially critical clues of terrestrial planet evolution that are lost forever on Earth. On both planets, detached sublithospheric melt-residue roots could accumulate elsewhere, forming deep long-lived roots, such as the continental tectosphere on Earth (Jordan, 1975, 1981) or beneath Ishtar Terra on Venus (Hansen and Phillips, 1995). It is plausible that Venus and Earth followed similar paths in terms of lithosphere thickness, with neither planet able to preserve large impact basins (and in the case of Earth, shocked minerals) due to relatively thin lithosphere during what is known on Earth as the Archean Eon. As lithosphere thickened, large impact basins could be preserved, such as Venus' Mead crater (~270 km diameter; Phillips et al., 1992; Schaber et al., 1992; Herrick et al., 1997) and the ca. 2 Ga Vredefort impact basin on Earth (250–300 km diameter). The paucity of large impact basins on Venus, and the lack of shocked minerals in Archean and Hadean rocks on Earth as noted herein, might be expected if these sister planets each hosted thin lithosphere during the time period of the Archean Eon. This timing for thin lithosphere on Venus is consistent with the number, spatial distribution, and apparent pristine character of impact craters on Venus (Bjonnes et al., 2012).

CONCLUDING IDEAS

There is no clear consensus with regard to the processes that formed Archean cratons or, by extrapolation, Earth's continents. There is, however, growing evidence that these features are marked by a unique pairing of structural/petrologic elements—crustal GGTs and strong, low-density mantle roots (CLM). GGTs owe their preservation to CLM, despite the common observation that both have been variably modified by postformational processes. Yet, GGT and CLM together preserve a record of geodynamic processes that occurred uniquely during the Archean. Myriad models have been proposed to explain the formation of GGTs and CLM, either separately or collectively, that refer to contemporary endogenic mechanisms based largely on uniformitarian principles. However, early Earth was likely quite different from contemporary Earth; contemporary Earth is the evolved product of early Earth. In our quest to understand early Earth processes, therefore, it is useful to consider the potentially consequential role of large-bolide impacts, as also suggested by others (e.g., Green, 1972; Elkins-Tanton and Hager, 2005). The bolide impact hypothesis outlined here provides an alternative explanation to other hypotheses currently considered for Archean craton formation. The bolide impact model does not require Archean plate tectonics, or Archean plumes, yet this model is also not incompatible with the other mechanisms.

ACKNOWLEDGMENTS

I thank J.W. Goodge for extremely helpful discussion, comments, and guidance; four anonymous reviewers; A. Davatzes and E. Kirby for reviews that helped to clarify the ideas presented herein; and editorial guidance from K. Stüwe. A. Davatzes helped me to appreciate aspects/challenges unique to the Archean of which I would have otherwise been unaware. I gratefully acknowledge the McKnight Foundation and the National Aeronautics and Space Administration (grant NNX06AB90G) for support of this work. The McKnight Foundation support, in particular, allows for exploration of ideas that might be considered outside the mainstream.

REFERENCES CITED

Abbott, D.H., Drury, R., and Mooney, W.D., 1997, Continents as lithological icebergs: The importance of buoyant lithospheric roots: Earth and Planetary Science Letters, v. 149, p. 15–27, doi:10.1016/S0012-821X(97)00065-4.

Anhaeusser, C.R., Mason, R., Viljoen, M.J., and Viljoen, R.P., 1969, A reappraisal of some aspects of Precambrian shield geology: Geological Society of America Bulletin, v. 80, p. 2175–2200, doi:10.1130/0016-7606(1969)80[2175:AROSAO]2.0.CO;2.

Ayres, L.D., and Thurston, P.C., 1985, Archean supracrustal sequences in the Canadian Shield: An overview, *in* Ayres, L.D., Thurston, P.C., Card, K.D., and Weber, W., eds., Evolution of Archean Supracrustal Sequences: Geological Association of Canada Special Paper 28, p. 343–370.

Banks, B.K., and Hansen, V.L., 2000, Relative timing of crustal plateau magmatism and tectonism at Tellus Regio, Venus: Journal of Geophysical Research, v. 105, p. 17,655–17,667, doi:10.1029/1999JE001205.

Barsukov, V.L., Basilevsky, A.T., Burba, G.A., et al., 1986, The geology and geomorphology of the Venus surface as revealed by the radar images obtained by *Veneras* 15 and 16: Journal of Geophysical Research, v. 91, p. D378–D398, doi:10.1029/JB091iB04p0D378.

Basilevsky, A.T., and Head, J.W., 1998, The geologic history of Venus: A stratigraphic view: Journal of Geophysical Research, v. 103, p. 8531–8544, doi:10.1029/98JE00487.

Basilevsky, A.T., Pronin, A.A., Ronca, L.B., Kryuchkov, V.P., Sukhanov, A.L., and Markov, M.S., 1986, Styles of tectonic deformations on Venus: Analysis of Veneras 15 and 16 data: Journal of Geophysical Research, v. 91, no. B4, p. D399–D411.

Bédard, J.H., 2003, Evidence for regional-scale, pluton-driven, high-grade metamorphism in the Archaean Minto block, northern Superior Province, Canada: The Journal of Geology, v. 111, p. 183–205, doi:10.1086/345842.

Bédard, J.H., 2006, A catalytic delamination-driven model for coupled genesis of Archaean crust and sub-continental lithospheric mantle: Geochimica et Cosmochimica Acta, v. 70, p. 1188–1214, doi:10.1016/j.gca.2005.11.008.

Bédard, J.H., Brouillette, P., Madore, L., and Berclaz, A., 2003, Archaean cratonization and deformation in the northern Superior Province, Canada: An evaluation of plate tectonic versus vertical tectonic models: Precambrian Research, v. 127, p. 61–87, doi:10.1016/S0301-9268(03)00181-5.

Bédard, J.H., Harris, L.B., and Thurston, P.C., 2013, The hunting of the snArc: Precambrian Research, v. 229, p. 20–48, doi:10.1016/j.precamres.2012.04.001.

Begg, G.C., Hronsky, J.A.M., Arndt, N.T., Griffin, W.L., O'Reilly, S.Y., and Hayward, N., 2010, Lithospheric, cratonic, and geodynamic setting of Ni-Cu-PGE sulfide deposits: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 105, p. 1057–1070, doi:10.2113/econgeo.105.6.1057.

Benn, K., and Moyen, J.F., 2008, The late Archean Abitibi-Opactica terrane, Superior Province: A modified oceanic plateau, *in* Condie, K.C., and Pease, V., eds., When Did Plate Tectonics Begin on Planet Earth?: Geological Society of America Special Paper 440, p. 173–197, doi:10.1130/2008.2440(09).

Bercovici, D., and Mahoney, J., 1994, Double flood basalts and plume head separation at the 660-kilometer discontinuity: Science, v. 266, p. 1367–1369, doi:10.1126/science.266.5189.1367.

Bickle, M.J., 1986, Implications of melting for stabilization of the lithosphere and heat-loss in the Archean: Earth and Planetary Science Letters, v. 80, p. 314–324, doi:10.1016/0012-821X(86)90113-5.

Bickle, M.J., Nisbet, G.G., and Martin, A., 1994, Archean greenstone belts are not oceanic crust: The Journal of Geology, v. 102, p. 121–137, doi:10.1086/629658.

Bindschadler, D.L., 1995, *Magellan*—A new view of Venus geology and geophysics: Reviews of Geophysics, v. 33, p. 459–467, doi:10.1029/95RG00281.

Bindschadler, D.L., deCharon, A., Beratan, K.K., and Head, J.W., 1992a, *Magellan* observations of Alpha Regio: Implications for formation of complex ridged terrains on Venus: Journal of Geophysical Research, v. 97, p. 13,563–13,577, doi:10.1029/92JE01332.

Bindschadler, D.L., Schubert, G., and Kaula, W.M., 1992b, Coldspots and hotspots: Global tectonics and mantle dynamic of Venus: Journal of Geophysical Research, v. 97, p. 13,495–13,532, doi:10.1029/92JE01165.

Bjonnes, E.E., Hansen, V.L., James, B., and Swenson, J.B., 2012, Equilibrium resurfacing of Venus: Results from new Monte Carlo modeling and implications for Venus

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

surface histories: Icarus, v. 217, p. 451–461, doi:10.1016/j.icarus.2011.03.033.
Bleeker, W., 2002, Archaean tectonics: A review, with illustrations from the Slave craton, *in* Fowler, C.M.R., Ebinger, C.J., and Hawkesworth, C.J., eds., The Early Earth: Physical, Chemical and Biological Development: Geological Society of London Special Publication 199, p. 151–181.
Bottke, W.F., Vokrouhlicky, D., Minton, D., Nesvorny, D., Morbidelli, A., Brasser, R., Simonson, B., and Levison, H.F., 2012, An Archaean heavy bombardment from a destabilized extension of the asteroid belt: Nature, v. 485, p. 78–81, doi:10.1038/nature10967.
Boyd, F.R., 1989, Compositional distinction between oceanic and cratonic lithosphere: Earth and Planetary Science Letters, v. 96, p. 15–26, doi:10.1016/0012-821X(89)90120-9.
Brown, C.D., and Grimm, R.E., 1997, Tessera deformation and the contemporaneous thermal state of the plateau highlands, Venus: Earth and Planetary Science Letters, v. 147, p. 1–10, doi:10.1016/S0012-821X(97)00007-1.
Brown, C.D., and Grimm, R.E., 1999, Recent tectonic and lithospheric thermal evolution of Venus: Icarus, v. 139, p. 40–48, doi:10.1006/icar.1999.6083.
Byerly, G.R., Lowe, D.R., Wooden, J.L., and Xie, X., 2002, An Archean impact layer from the Pilbara and Kaapvaal cratons: Science, v. 297, p. 1325–1327, doi:10.1126/science.1073934.
Calvert, A.J., Sawyer, E.W., Davis, W.J., and Ludden, J.N., 1995, Archean subduction inferred from seismic images of a mantle suture in the Superior Province: Nature, v. 375, p. 670–674, doi:10.1038/375670a0.
Campbell, I.H., and Griffiths, R.W., 1990, Implications of mantle plume structure for the evolution of flood basalts: Earth and Planetary Science Letters, v. 99, p. 79–93, doi:10.1016/0012-821X(90)90072-6.
Campbell, I.H., and Griffiths, R.W., 2014, Did the formation of D″ cause the Archaean-Proterozoic transition?: Earth and Planetary Science Letters, v. 388, p. 1–8, doi:10.1016/j.epsl.2013.11.048.
Card, K.D., 1990, A review of the Superior Province of the Canadian Shield, a product of Archean accretion: Precambrian Research, v. 48, p. 99–156, doi:10.1016/0301-9268(90)90059-Y.
Carlson, R.W., Pearson, D.G., and James, D.E., 2005, Physical, chemical, and chronological characteristics of continental mantle: Reviews of Geophysics, v. 43, p. RG1001, doi:10.1029/2004RG000156.
Cavosie, A.J., Valley, J.W., and Wilde, S.A., 2007, The oldest terrestrial mineral record: A review of 4400 to 3900 Ma detrital zircons from Jack Hills, Western Australia, *in* van Kranendonk, M.J., Smithies, R.H., and Bennett, V.C., eds., World's Oldest Rocks: Amsterdam, Netherlands, Elsevier, p. 91–111.
Cavosie, A.J., Quintero, R.R., Radovan, H.A., and Moser, D.E., 2010, A record of ancient cataclysm in modern sand: Shock microstructures in detrital minerals from the Vaal River, Vredefort Dome, South Africa: Geological Society of America Bulletin, v. 122, p. 1968–1980, doi:10.1130/B30187.1.
Cawood, P.A., Kroner, A., and Pisarevsky, S., 2006, Precambrian plate tectonics: Criteria and evidence: GSA Today, v. 16, no. 10, p. 4–11, doi:10.1130/GSAT01607.1.
Cawood, P.A., Kroner, A., Collins, W.J., Kusky, T., Mooney, W.D., and Windley, B., 2009, Accretionary orogens through Earth history, *in* Cawood, P.A., and Kroner, A., eds., Earth Accretionary Systems in Space and Time: Geological Society of London Special Publication 318, p. 1–36.
Chardon, D., Choukroune, P., and Jayananda, M., 1996, Strain patterns, decollement and incipient sagducted greenstone terrains in the Archaean Dharwar craton (south India): Journal of Structural Geology, v. 18, p. 991–1004, doi:10.1016/0191-8141(96)00031-4.
Chardon, D., Choukroune, P., and Jayananda, M., 1998, Sinking of the Dharwar Basin (South India): Implications for Archaean tectonics: Precambrian Research, v. 91, p. 15–39, doi:10.1016/S0301-9268(98)00037-0.
Chardon, D., Peucat, J.-J., Jayananda, M., Choukroune, P., and Fanning, M., 2002, Archean granite-greenstone tectonics at Kolar (South India): Interplay of diapirism and bulk inhomogenous contraction during juvenile magmatic accretion: Tectonics, v. 21, doi:10.1029/2001TC901032.
Choukroune, P., Bouhallier, H., and Arndt, N.T., 1995, Soft lithosphere during periods of Archaean crustal growth or crustal reworking, *in* Coward, M., and Ries, A.C., eds., Early Precambrian Processes: Geological Society of London Special Publication 95, p. 67–86.
Choukroune, P., Ludden, J.N., Chardon, D., Calvert, A.J., and Bouhallier, H., 1997, Archaean crustal growth and tectonic processes: A comparison of the Superior Province, Canada, and the Dharwar craton, India, *in* Burg, J.-P., and Ford, M., eds., Orogeny Through Time: Geological Society of London Special Publication 121, p. 63–98.
Chown, E.H., Harrap, R., and Moukhsil, A., 2002, The role of granitic intrusions in the evolution of the Abitibi belt, Canada: Precambrian Research, v. 115, p. 291–310, doi:10.1016/S0301-9268(02)00013-X.
Coffin, M.F., and Eldholm, O., 1994, Large igneous provinces: Crustal structure, dimensions, and external consequences: Reviews of Geophysics, v. 32, p. 1–36, doi:10.1029/93RG02508.
Cohen, B.A., Swindle, T.D., and Kring, D.A., 2000, Support for the lunar cataclysm hypothesis from lunar meteorite impact melt ages: Science, v. 290, no. 5497, p. 1754–1756, doi:10.1126/science.290.5497.1754.
Collins, W.J., Van Kranendonk, M.J., and Teyssier, C., 1998, Partial convective overturn of Archaean crust in the east Pilbara craton, Western Australia: Driving mechanisms and tectonic implications: Journal of Structural Geology, v. 20, p. 1405–1424, doi:10.1016/S0191-8141(98)00073-X.
Condie, K., 2007, The distribution of Paleoarchean crust, *in* Van Kranendonk, M.J., Smithies, R.H., and Bennett, V.C., eds., Earth's Oldest Rocks (>3.2 Ga): Amsterdam, Netherlands, Elsevier, p. 971–983, doi:10.1016/S0166-2635(07)15012-X.
Covey, C., Thompson, S.L., Weissman, P.R., and MacCracken, M.C., 1994, Global climatic effects of atmospheric dust from an asteroid or comet impact on Earth: Global and Planetary Change, v. 9, p. 263–273, doi:10.1016/0921-8181(94)90020-5.
Daigneault, R., Mueller, W.U., and Chown, E.H., 2002, Oblique Archean subduction: Accretion and exhumation of an oceanic arc during dextral transpression, Southern volcanic zone, Abitibi Subprovince Canada: Precambrian Research, v. 115, p. 261–290, doi:10.1016/S0301-9268(02)00012-8.
Davies, G.F., 1979, Thickness and thermal history of continental crust and root zones: Earth and Planetary Science Letters, v. 44, p. 231–238, doi:10.1016/0012-821X(79)90171-7.
Davies, G.F., 1999, Dynamic Earth: Plates, Plumes and Mantle Convection: Cambridge, UK, Cambridge University Press, 458 p.
Davies, G.F., 2006, Gravitational depletion of the early Earth's upper mantle and the viability of early plate tectonics: Earth and Planetary Science Letters, v. 243, p. 376–382, doi:10.1016/j.epsl.2006.01.053.
Davies, G.F., 2007a, Dynamics of the Hadean and Archaean mantle, *in* Van Kranendonk, M.J., Smithies, R.H., and Bennett, V.C., eds., Earth's Oldest Rocks (>3.2 Ga): Amsterdam, Netherlands, Elsevier, p. 61–73, doi:10.1016/S0166-2635(07)15023-4.
Davies, G.F., 2007b, Controls on density stratification in the early mantle: Geochemistry, Geophysics, Geosystems, v. 8, p. Q04006, doi:10.1029/2006GC001414.
Davies, G.F., 2007c, Mantle regulation of core cooling: A geodynamo without core radioactivity?: Physics of the Earth and Planetary Interiors, v. 160, p. 215–229, doi:10.1016/j.pepi.2006.11.001.
Davis, D.W., Sutcliffe, R.H., and Trowell, N.F., 1988, Geochronological constraints on the tectonic evolution of a Late Archaean greenstone belt, Wabigoon Subprovince, northwest Ontario, Canada: Precambrian Research, v. 39, p. 171–191, doi:10.1016/0301-9268(88)90041-1.
de Wit, M.J., 1998, On Archean granites, greenstones, cratons and tectonics: Precambrian Research, v. 91, p. 181–226, doi:10.1016/S0301-9268(98)00043-6.
Elkins-Tanton, L., and Hager, B., 2005, Giant meteoroid impacts can cause volcanism: Earth and Planetary Science Letters, v. 239, p. 219–232, doi:10.1016/j.epsl.2005.07.029.
Farnetani, C.G., and Richards, M.A., 1994, Numerical investigations of the mantle plume initiation model for flood basalt events: Journal of Geophysical Research, v. 99, p. 13,813–13,833, doi:10.1029/94JB00649.
Farnetani, C.G., Richards, M.A., and Ghiorso, M.S., 1996, Petrological models of magma evolution and deep crustal structure beneath hotspots and flood basalt provinces: Earth and Planetary Science Letters, v. 143, p. 81–94, doi:10.1016/0012-821X(96)00138-0.
French, B.M., and Koeberl, C., 2010, The convincing identification of terrestrial meteorite impact structures: What works, what doesn't, and why: Earth-Science Reviews, v. 98, p. 123–170, doi:10.1016/j.earscirev.2009.10.009.
Ghent, R.R., and Hansen, V.L., 1999, Structural and kinematic analysis of eastern Ovda Regio, Venus: Implications for crustal plateau formation: Icarus, v. 139, p. 116–136, doi:10.1006/icar.1999.6085.
Ghent, R.R., and Tibuleac, I.M., 2002, Ribbon spacing in Venusian tessera: Implications for layer thickness and thermal state: Geophysical Research Letters, v. 29, p. 2000–2003, doi:10.1029/2002GL015994.
Gilmore, M.S., Collins, G.C., Ivanov, M.A., Marinangeli, L., and Head, J.W., 1998, Style and sequence of extensional structures in tessera terrain, Venus: Journal of Geophysical Research, v. 103, p. 16,813–16,840, doi:10.1029/98JE01322.
Glass, B.P., and Simonson, B.M., 2012, Distal impact ejecta layers: Spherules and more: Elements, v. 8, p. 43–48, doi:10.2113/gselements.8.1.43.
Golabek, G.J., Keller, T., Gerya, T.V., Zhu, G., Tackley, P.J., and Connolly, J.A.D., 2011, Origin of the Martian dichotomy and Tharsis from a giant impact causing massive magmatism: Icarus, v. 215, p. 346–357, doi:10.1016/j.icarus.2011.06.012.
Goodwin, A.M., 1982, Archean volcanos in southwestern Abitibi belt, Ontario and Quebec—Form, composition, and development: Canadian Journal of Earth Sciences, v. 19, p. 1140–1155, doi:10.1139/e82-098.
Goodwin, A.M., 1996, Principles of Precambrian Geology: London, Academic Press, 327 p.
Graupner, M., 2013, Geologic and Structural Mapping of Southern Tellus Regio, Venus [M.S. thesis]: Duluth, Minnesota, University of Minnesota, 80 p.
Green, D.H., 1972, Archaean greenstone belts may include terrestrial equivalents of lunar maria?: Earth and Planetary Science Letters, v. 15, p. 263–270, doi:10.1016/0012-821X(72)90172-0.
Griffin, W.L., and O'Reilly, S.Y., 2007, The earliest subcontinental lithospheric mantle, *in* Van Kranendonk, M.J., Smithies, R.H., and Bennett, V.C., eds., Earth's Oldest Rocks (>3.2 Ga): Amsterdam, Netherlands, Elsevier, p. 1013–1035, doi:10.1016/S0166-2635(07)15082-9.
Griffin, W.L., O'Reilly, S.Y., Abe, N., Aulbach, S., Davies, R.M., Pearson, N.J., Doyle, B.J., and Kivi, K., 2003, The origin and evolution of Archean lithospheric mantle: Precambrian Research, v. 127, p. 19–41, doi:10.1016/S0301-9268(03)00180-3.
Griffin, W.L., O'Reilly, S.Y., Doyle, B.J., Pearson, N.J., Coopersmith, H., Kivi, K., Malkovets, V., and Pokhilenko, N., 2004, Lithosphere mapping beneath the North American plate: Lithos, v. 77, p. 873–922, doi:10.1016/j.lithos.2004.03.034.
Griffin, W.L., Begg, G., O'Reilly, S.Y., and Afonso, J.C., 2009a, Paleo-Archean generation of the continental lithosphere: Geochimica et Cosmochimica Acta, v. 73, p. A466.
Griffin, W.L., O'Reilly, S.Y., Afonso, J.C., and Begg, G.C., 2009b, The composition and evolution of lithospheric mantle: A re-evaluation and its tectonic implications: Journal of Petrology, v. 50, p. 1185–1204, doi:10.1093/petrology/egn033.
Griffin, W.L., Begg, G.C., and O'Reilly, S.Y., 2013, Continental-root control on the genesis of magmatic ore deposits: Nature Geoscience, v. 6, p. 905–910, doi:10.1038/ngeo1954.
Griffiths, R.W., 1986, Dynamics of mantle thermals with constant buoyancy or anomalous internal heating: Earth and Planetary Science Letters, v. 78, p. 435–446, doi:10.1016/0012-821X(86)90010-5.
Grimm, R.E., 1994, Recent deformation rates on Venus: Journal of Geophysical Research, v. 99, p. 23,163–23,171, doi:10.1029/94JE02196.
Groves, D.I., Ho, S.E., Rock, N.M.S., Barley, M.E., and Muggeridge, M.T., 1987, Archean cratons, diamond and platinum—Evidence for coupled long-lived crust-mantle systems: Geology, v. 15, p. 801–805, doi:10.1130/0091-7613(1987)15<801:ACDAPE>2.0.CO;2.
Hamilton, W.B., 1998, Archean magmatism and deformation were not products of plate tectonics: Precambrian

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

Research, v. 91, p. 143–179, doi:10.1016/S0301-9268(98)00042-4.

Hansen, V.L., 1992, Regional non-coaxial deformation on Venus: Evidence from western Itzpapalotl Tessera: Houston, Texas, Lunar and Planetary Institute, Lunar and Planetary Science Conference XXIII, p. 478–479.

Hansen, V.L., 2006, Geologic constraints on crustal plateau surface histories, Venus: The lava pond and bolide impact hypotheses: Journal of Geophysical Research, v. 111, no. E11010, doi:10.1029/2006JE002714.

Hansen, V.L., 2007, Venus: A thin-lithosphere analog for early Earth?, *in* Van Kranendonk, M.J., Smithies, R.H., and Bennett, V.C., eds., Earth's Oldest Rocks (>3.2 Ga): Amsterdam, Netherlands, Elsevier, p. 971–983, doi:10.1016/S0166-2635(07)15081-7.

Hansen, V.L., and López, I., 2010, Venus records a rich early history: Geology, v. 38, p. 311–314, doi:10.1130/G30587.1.

Hansen, V.L., and Phillips, R.J., 1995, Formation of Ishtar Terra, Venus: Surface and gravity constraints: Geology, v. 23, p. 292–296, doi:10.1130/0091-7613(1995)023<0292:FOITVS>2.3.CO;2.

Hansen, V.L., and Tharalson, E.R., 2014, Geologic Map of the Agnesi Quadrangle (V–45), Venus: U.S. Geological Survey Scientific Investigations Map SIM-3250, scale 1:5,000,000, 1 sheet and pamphlet.

Hansen, V.L., and Willis, J.J., 1996, Structural analysis of a sampling of tesserae: Implications for Venus geodynamics: Icarus, v. 123, p. 296–312, doi:10.1006/icar.1996.0159.

Hansen, V.L., and Willis, J.J., 1998, Ribbon terrain formation, southwestern Fortuna Tessera, Venus: Implications for lithosphere evolution: Icarus, v. 132, p. 321–343, doi:10.1006/icar.1998.5897.

Hansen, V.L., Banks, B.K., and Ghent, R.R., 1999, Tessera terrain and crustal plateaus, Venus: Geology, v. 27, p. 1071–1074, doi:10.1130/0091-7613(1999)027<1071:TTACPV>2.3.CO;2.

Hansen, V.L., Phillips, R.J., Willis, J.J., and Ghent, R.R., 2000, Structures in tessera terrain, Venus: Issues and answers: Journal of Geophysical Research, v. 105, p. 4135–4152, doi:10.1029/1999JE001137.

Head, J.W., and Basilevsky, A.T., 1998, Sequence of tectonic deformation in the history of Venus: Evidence from global stratigraphic relations: Geology, v. 26, p. 35–38, doi:10.1130/0091-7613(1998)026<0035:SOTDIT>2.3.CO;2.

Herrick, R.R., Sharpton, V.L., Malin, M.C., Lyons, S.N., and Feely, K., 1997, Morphology and morphometry of impact craters, *in* Bouger, S.W., Hunten, D.M., and Phillips, R.J., eds., Venus II: Tucson, Arizona, University of Arizona Press, p. 1015–1046.

Herzberg, C., and Rudnick, R., 2012, Formation of cratonic lithosphere: An integrated thermal and petrological model: Lithos, v. 149, p. 4–15, doi:10.1016/j.lithos.2012.01.010.

Herzberg, C., Condie, K., and Korenaga, J., 2010, Thermal history of the Earth and its petrological expression: Earth and Planetary Science Letters, v. 292, p. 79–88, doi:10.1016/j.epsl.2010.01.022.

Ignacio, R., Ramo'n, C., and Francisco, A., 2005, Tectonic and kinematic study of a strike-slip zone along the southern margin of central Ovda Regio, Venus—Geodynamical implications for crustal plateaux formation and evolution: Icarus, v. 175, p. 320–334, doi:10.1016/j.icarus.2004.11.007.

Ingle, S., and Coffin, M.F., 2004, Impact origin for the greater Ontong Java Plateau?: Earth and Planetary Science Letters, v. 218, p. 123–134, doi:10.1016/S0012-821X(03)00629-0.

Ito, G., and Clift, P.D., 1998, Subsidence and growth of Pacific Cretaceous plateaus: Earth and Planetary Science Letters, v. 161, p. 85–100, doi:10.1016/S0012-821X(98)00139-3.

Ito, G., and Taira, A., 2000, Compensation of the Ontong Java Plateau by surface and subsurface loading: Journal of Geophysical Research–Solid Earth, v. 105, no. B11, p. 11,171–11,183, doi:10.1029/2000JB900036.

Ivanov, B.A., Melosh, H.J., and Pierazzo, E., 2010, Basin-forming impacts: Reconnaissance modeling, *in* Gibson, R.L., and Reimold, W.U., eds., Large Meteorite Impacts and Planetary Evolution IV: Geological Society of America Special Paper 465, p. 29–49, doi:10.1130/2010.2465(03).

Ivanov, M.A., and Head, J.W., 1996, Tessera terrain on Venus: A survey of the global distribution, characteristics, and relation to surrounding units from *Magellan* data: Journal of Geophysical Research, v. 101, p. 14,861–14,908, doi:10.1029/96JE01245.

Johnson, B.C., and Melosh, H.J., 2012, Impact spherules as a record of an ancient heavy bombardment of Earth: Nature, v. 485, p. 75–77, doi:10.1038/nature10982.

Jones, A.P., Price, G.D., Price, N.J., DeCarli, P.S., and Clegg, R.A., 2002, Impact induced melting and the development of large igneous provinces: Earth and Planetary Science Letters, v. 202, p. 551–561, doi:10.1016/S0012-821X(02)00824-5.

Jones, A.P., Wunemann, K., and Price, D., 2005, Impact volcanism as a possible origin for the Ontong Java Plateau (OJP), *in* Foulger, G.R., Natland, J.H., Presnall, D.C., and Anderson, D.L., eds., Plates, Plumes, and Paradigms: Geological Society of America Special Paper 388, p. 711–720, doi:10.1130/0-8137-2388-4.711.

Jordan, T.H., 1975, The continental tectosphere: Reviews of Geophysics and Space Physics, v. 13, p. 1–12, doi:10.1029/RG013i003p00001.

Jordan, T.H., 1981, Continents as a chemical boundary layer: Philosophical Transactions of the Royal Society of London, ser. A, Mathematical and Physical Sciences, v. 301, p. 359–373, doi:10.1098/rsta.1981.0117.

Jordan, T.H., 1989, Chemical-boundary layers of the mantle and core: Philosophical Transactions of the Royal Society of London, v. 328, p. 441–442, doi:10.1098/rsta.1989.0047.

Kamber, B.S., 2015, The evolving nature of terrestrial crust from the Hadean, through the Archaean, into the Proterozoic: Precambrian Research, v. 258, p. 48–82, doi:10.1016/j.precamres.2014.12.007.

Kemp, A.I.S., Hickman, A.H., Kirkland, C.L., and Vervoort, J.D., 2015, Hf isotopes in detrital and inherited zircons of the Pilbara craton provide no evidence for Hadean continents: Precambrian Research, v. 261, p. 112–126, doi:10.1016/j.precamres.2015.02.011.

Koeberl, C., 2006, Impact processes on the early Earth: Elements, v. 2, p. 211–216, doi:10.2113/gselements.2.4.211.

Koeberl, C., Claeys, P., Hecht, L., and McDonald, I., 2012, Geochemistry if impactites: Elements, v. 8, p. 37–42, doi:10.2113/gselements.8.1.37.

Korenaga, J., 2006, Archean geodynamics and the thermal evolution of Earth, *in* Benn, K., Mareschal, J.-C., and Condie, K.C., eds., Archean Geodynamics and Environments: American Geophysical Union Geophysical Monograph 164, p. 7–32.

Korenaga, J., 2008a, Urey ratio and the structure and evolution of Earth's mantle: Reviews of Geophysics, v. 46, p. RG2007, doi:10.1029/2007RG000241.

Korenaga, J., 2008b, Plate tectonics, flood basalts and the evolution of Earth's oceans: Terra Nova, v. 20, p. 419–439, doi:10.1111/j.1365-3121.2008.00843.x.

Kumar, P.S., 2005, An alternative kinematic interpretation of Thetis Boundary shear zone, Venus: Journal of Geophysical Research, v. 110, p. E07001, doi:10.1029/2004JE002387.

Kusky, T.M., 1998, Tectonic setting and terrane accretion of the Archean Zimbabwe craton: Geology, v. 26, p. 163–166, doi:10.1130/0091-7613(1998)026<0163:TSATAO>2.3.CO;2.

Kusky, T.M., and Polat, A., 1999, Growth of granite-greenstone terranes at convergent margins, and stabilization of Archean cratons: Tectonophysics, v. 305, no. 1–3, p. 43–73, doi:10.1016/S0040-1951(99)00014-1.

Langenhorst, F., and Deutsch, A., 2012, Shock metamorphism of minerals: Elements, v. 8, p. 31–36, doi:10.2113/gselements.8.1.31.

Lee, C.T.A., 2006, Geochemical/petrologic constraints on the origin of cratonic mantle, *in* Benn, K., Mareschal, J.-C., and Condie, K.C., eds., Archean Geodynamics and Environments: American Geophysical Union Geophysical Monograph 164, p. 89–114.

Lee, C.T.A., Luffi, P., and Chin, E.J., 2011, Building and destroying continental mantle: Annual Review of Earth and Planetary Sciences, v. 39, p. 59–90, doi:10.1146/annurev-earth-040610-133505.

Lenardic, A., Moresi, L.N., and Muhlhaus, H., 2003, Longevity and stability of cratonic lithosphere: Insights from numerical simulations of coupled mantle convection and continental tectonics: Journal of Geophysical Research–Solid Earth, v. 108, no. B6, p. 2303–2317, doi:10.1029/2002JB001859.

Lin, S., 2005, Synchronous vertical and horizontal tectonism in the Neoarchean: Kinematic evidence from a synclinal keel in the northwestern Superior craton, Canada: Precambrian Research, v. 139, p. 181–194, doi:10.1016/j.precamres.2005.07.001.

Lin, S., Percival, J.A., and Skulski, T., 1996, Structural constraints on the tectonic evolution of a late Archean greenstone belt in the northeastern Superior Province, northern Quebec (Canada): Tectonophysics, v. 265, p. 151–167, doi:10.1016/S0040-1951(96)00150-3.

Lin, S., Parks, J., Heaman, L.M., Simonetti, A., and Corkery, M.T., 2013, Diapirism and sagduction as a mechanism for deposition and burial of "Timiskaming-type" sedimentary sequences, Superior Province: Evidence from detrital zircon geochronology and implications for the Borden Lake conglomerate in the exposed middle to lower crust in the Kapuskasing uplift: Precambrian Research, v. 238, p. 148–157, doi:10.1016/j.precamres.2013.09.012.

Lowe, D.R., Byerly, G.R., Kyte, F.T., Shukolyukov, A., Asaro, F., and Krull, A., 2003, Spherule beds 3.47–3.24 billion years old in the Barberton greenstone belt, South Africa: A record of large meteorite impacts and their influence on early crustal and biological evolution: Astrobiology, v. 3, p. 7–48, doi:10.1089/153110703321632408.

Mareschal, J.C., and West, G.F., 1980, A model for Archean tectonism: 2. Numerical-models of vertical tectonism in greenstone belts: Canadian Journal of Earth Sciences, v. 17, p. 60–71, doi:10.1139/e80-006.

McGregor, A.M., 1951, Some milestones in the Precambrian of South Rhodesia, Parts 1–3: Transactions of the Geological Society of South Africa, v. 54, p. 27–71.

Miljković, K., Wieczorek, M.A., Collins, G.S., et al., 2013, Asymmetric distribution of lunar impact basins caused by variations in target properties: Science, v. 342, p. 724–726, doi:10.1126/science.1243224.

Moore, W.B., and Webb, A.A.G., 2013, Heat-pipe Earth: Nature, v. 501, p. 501–505, doi:10.1038/nature12473.

Moser, D.E., Flowers, R.M., and Hart, R.J., 2001, Birth of the Kaapvaal tectosphere 3.08 billion years ago: Science, v. 291, p. 465–468, doi:10.1126/science.291.5503.465.

O'Neill, C.O., Jellinek, A.M., and Lenardic, A., 2007, Conditions for the onset of plate tectonics on terrestrial planets and moons: Earth and Planetary Science Letters, v. 261, p. 20–32, doi:10.1016/j.epsl.2007.05.038.

O'Reilly, T.C., and Davies, G.F., 1981, Magma transport of heat on Io: A mechanism allowing a thick lithosphere: Geophysical Research Letters, v. 8, p. 313–316, doi:10.1029/GL008i004p00313.

Parmenter, A.C., Lin, S., and Corkery, M.T., 2006, Structural evolution of the Cross Lake greenstone belt in the northwestern Superior Province, Manitoba: Implications for relationship between vertical and horizontal tectonism: Canadian Journal of Earth Sciences, v. 43, p. 767–787, doi:10.1139/e06-006.

Parsons, B., and McKenzie, D., 1978, Mantle convection and thermal structure of plates: Journal of Geophysical Research, v. 83, no. B9, p. 4485–4496, doi:10.1029/JB083iB09p04485.

Pearson, D.G., 1999, The age of continental roots: Lithos, v. 48, p. 171–194, doi:10.1016/S0024-4937(99)00026-2.

Pearson, D.G., and Wittig, N., 2008, Formation of Archaean continental lithosphere and its diamonds: The root of the problem: Journal of the Geological Society of London, v. 165, p. 895–914, doi:10.1144/0016-76492008-003.

Percival, J.A., and Pysklywec, R.N., 2007, Are Archean lithospheric keels inverted?: Earth and Planetary Science Letters, v. 254, p. 393–403, doi:10.1016/j.epsl.2006.11.047.

Percival, J.A., and Williams, H.R., 1989, Late Archean Quetico accretionary complex, Superior Province, Canada: Geology, v. 17, p. 23–27, doi:10.1130/0091-7613(1989)017<0023:LAQACS>2.3.CO;2.

Percival, J.A., Stern, R.A., Skulski, T., Card, K.D., Mortensen, J.K., and Begin, N.J., 1994, Minto block, Superior province: Missing link in deciphering assembly of the craton at 2.7 Ga: Geology, v. 22, p. 839–842, doi:10.1130/0091-7613(1994)022<0839:MBSPML>2.3.CO;2.

Percival, J.A., McNicoll, V., Brown, J.L., and Whalen, J.B., 2004, Convergent margin tectonics, central Wabigoon Subprovince, Superior Province, Canada: Precambrian Research, v. 132, p. 213–244, doi:10.1016/j.precamres.2003.12.016.

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest

Phillips, R.J., and Hansen, V.L., 1994, Tectonic and magmatic evolution of Venus: Annual Review of Earth and Planetary Sciences, v. 22, p. 597–656, doi:10.1146/annurev.ea.22.050194.003121.

Phillips, R.J., and Hansen, V.L., 1998, Geological evolution of Venus: Rises, plains, plumes and plateaus: Science, v. 279, p. 1492–1497, doi:10.1126/science.279.5356.1492.

Phillips, R.J., Raubertas, R.F., Arvidson, R.E., Sarkar, I.C., Herrick, R.R., Izenberrg, N., and Grimm, R.E., 1992, Impact craters and Venus resurfacing history: Journal of Geophysical Research, v. 97, p. 15,923–15,948, doi:10.1029/92JE01696.

Polat, A., Kerrich, R., and Windley, B., 2009, Archaean crustal growth processes in southern West Greenland and the southern Superior Province: Geodynamic and magmatic constraints, *in* Cawood, P.A., and Kroner, A., eds., Earth Accretionary Systems in Space and Time: The Geological Society of London Special Publication 318, p. 155–191.

Pollack, H.N., 1986, Cratonization and thermal evolution of the mantle: Earth and Planetary Science Letters, v. 80, no. 1–2, p. 175–182, doi:10.1016/0012-821X(86)90031-2.

Poudjom Djomani, Y.H.P., O'Reilly, S.Y., Griffin, W.L., and Morgan, P., 2001, The density structure of subcontinental lithosphere through time: Earth and Planetary Science Letters, v. 184, p. 605–621, doi:10.1016/S0012-821X(00)00362-9.

Price, N.J., 2001, Major Impacts and Plate Tectonics: A Model for the Phanerozoic Evolution of the Earth's Lithosphere: London, Routledge, 354 p.

Rasmussen, B., and Koeberl, C., 2004, Iridium anomalies and shocked quartz in a late Archean spherule layer from the Pilbara craton: New evidence for a major asteroid impact at 2.63 Ga: Geology, v. 32, no. 12, p. 1029–1032, doi:10.1130/G20825.1.

Reese, C.C., Solomatov, V.S., and Baumgardner, J.R., 2002, Survival of impact-induced thermal anomalies in the Martian mantle: Journal of Geophysical Research, v. 107, p. E10, p. 5082–5093, doi:10.1029/2000JE001474.

Reese, C.C., Solomatov, V.S., Baumgardner, J.R., Stegman, D.R., and Vezolainen, A.V., 2004, Magmatic evolution of impact-induced Martian mantle plumes and the origin of Tharsis: Journal of Geophysical Research, v. 109, no. E8, doi:10.1029/2003JE002222.

Rey, P.F., and Houseman, G., 2006, Lithospheric scale gravitational flow: The impact of body forces on orogenic processes from Archaean to Phanerozoic, *in* Buiter, S.J.H., and Schreurs, G., eds., Analogue and Numerical Modelling of Crustal-Scale Processes: Geological Society of London Special Publication 253, p. 153–167.

Rey, P.F., Philippot, P., and Thebaud, N., 2003, Contribution of mantle plumes, crustal thickening and greenstone blanketing to the 2.75–2.65 Ga global crisis: Precambrian Research, v. 127, p. 43–60, doi:10.1016/S0301-9268(03)00179-7.

Roberts, J.H., and Arkani-Hamed, J., 2014, Impact heating and coupled core cooling and mantle dynamics on Mars: Journal of Geophysical Research, v. 119, p. 729–744, doi:10.1002/2013JE004603.

Robin, C.M.I., and Bailey, R.C., 2009, Simultaneous generation of Archean crust and subcratonic roots by vertical tectonics: Geology, v. 37, p. 523–526, doi:10.1130/G25519A.1.

Rogers, G.C., 1982, Oceanic plateaus as meteorite impact signatures: Nature, v. 299, p. 341–342, doi:10.1038/299341a0.

Romeo, I., Capote, R., and Anguita, F., 2005, Tectonic and kinematic study of a strike-slip zone along the southern margin of Central Ovda Regio, Venus: Geodynamical implications for crustal plateaux formation and evolution: Icarus, v. 175, p. 320–334, doi:10.1016/j.icarus.2004.11.007.

Ruiz, J., 2007, The heat flow during the formation of ribbon terrains on Venus: Planetary and Space Science, v. 55, p. 2063–2070, doi:10.1016/j.pss.2007.05.003.

Sandiford, M., Van Kranendonk, M.J., and Bodorkos, S., 2004, Conductive incubation and the origin of dome-and-keel structure in Archean granite-greenstone terrains: A model based on the eastern Pilbara craton, Western Australia: Tectonics, v. 23, p. TC1009, doi:10.1029/2002TC001452.

Schaber, G.G., Strom, R.G., Moore, H.J., Soderblom, L.A., Kirk, R.L., Chadwick, D.J., Dawson, D.D., Gaddis, L.R., Boyce, J.M., and Russell, J., 1992, Geology and distribution of impact craters on Venus: What are they telling us?: Journal of Geophysical Research, v. 97, p. 13,257–13,302, doi:10.1029/92JE01246.

Schubert, G., Solomatov, V.S., Tackley, P.J., and Turcotte, D.L., 1997, Mantle convection and the thermal evolution of Venus, *in* Bougher, S.W., Hunten, D.M., and Phillips, R.J., eds., Venus II: Tucson, Arizona, University of Arizona, p. 1245–1287.

Simons, M., Solomon, S.C., and Hager, B.H., 1997, Localization of gravity and topography: Constraints on the tectonics and mantle dynamics of Venus: Geophysical Journal International, v. 131, p. 24–44, doi:10.1111/j.1365-246X.1997.tb00593.x.

Simonson, B.M., and Glass, B.P., 2004, Spherule layers—Records of ancient impacts: Annual Review of Earth and Planetary Sciences, v. 32, p. 329–361, doi:10.1146/annurev.earth.32.101802.120458.

Slonecker, M.A., 2013, Geologic and Structural Mapping of Northern Tellus Regio, Venus [M.S. thesis]: Duluth, Minnesota, University of Minnesota, 77 p.

Smrekar, S.E., and Phillips, R.J., 1991, Venusian highlands: Geoid to topography ratios and their implications: Earth and Planetary Science Letters, v. 107, p. 582–597, doi:10.1016/0012-821X(91)90103-O.

Solomatov, V.S., and Moresi, L.-N., 1996, Stagnant lid convection on Venus: Journal of Geophysical Research, v. 101, p. 4737–4753, doi:10.1029/95JE03361.

Solomon, S.C., 1993, The geophysics of Venus: Physics Today, v. 46, p. 48–55, doi:10.1063/1.881359.

Solomon, S.C., Head, J.W., Kaula, W.M., McKenzie, D., Parsons, B., Phillips, R.J., Schubert, G., and Talwani, M., 1991, Venus tectonics: Initial analysis from *Magellan*: Science, v. 252, p. 297–312, doi:10.1126/science.252.5003.297.

Stern, R.J., 2008, Modern-style plate tectonics began in Neoproterozoic time: An alternative interpretation of Earth's tectonic history, *in* Condie, K.C., and Pease, V., eds., When Did Plate Tectonics Begin on Planet Earth?: Geological Society of America Special Paper 440, p. 265–280.

Tarduno, J.A., Sliter, W.V., Kroenke, L., Leckie, M., Mayer, H., Mahoney, J., Musgrave, R., Storey, M., and Winterer, E.L., 1991, Rapid formation of Ontong Java Plateau by Aptian mantle plume volcanism: Science, v. 254, p. 399–403, doi:10.1126/science.254.5030.399.

Tejada, M.L.G., Ravizza, G., Suzuki, K., and Paquay, F.S., 2012, An extraterrestrial trigger for the Early Cretaceous massive volcanism? Evidence from the paleo–Tethys Ocean: Scientific Reports, v. 268, doi:10.1038/srep00268.

Thompson, H., 2012, Ancient asteroids kept on coming: Nature, v. 484, p. 429, doi:10.1038/484429a.

Thomson, O.A., Cavosie, A.J., Moser, D.E., Barker, I., Radovan, H.A., and French, B.M., 2014, Preservation of detrital shocked minerals derived from the 1.85 Ga Sudbury impact structure in modern alluvium and Holocene glacial deposits: Geological Society of America Bulletin, v. 126, p. 720–737, doi:10.1130/B30958.1.

Thurston, P.C., Ayer, J.A., Goutier, J., and Hamilton, M.A., 2008, Depositional gaps in Abitibi greenstone belt stratigraphy: A key to exploration for syngenetic mineralization: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 103, p. 1097–1134, doi:10.2113/gsecongeo.103.6.1097.

Trofimovs, J., Davis, B.K., and Cas, R.A.F., 2004, Contemporaneous ultramafic and felsic intrusive and extrusive magmatism in the Archaean Boorara Domain, Eastern Goldfields superterrane, Western Australia, and its implications: Precambrian Research, v. 131, p. 283–304, doi:10.1016/j.precamres.2003.12.012.

Tuckwell, G.W., and Ghail, R.C., 2003, A 400-km-scale strike-slip zone near the boundary of Thetis Regio, Venus: Earth and Planetary Science Letters, v. 211, p. 45–55, doi:10.1016/S0012-821X(03)00128-6.

Van Kranendonk, M.J., 2010, Two types of Archean continental crust: Plume and plate tectonics on early Earth: American Journal of Science, v. 310, p. 1187–1209, doi:10.2475/10.2010.01.

Van Kranendonk, M.J., Collins, W.J., Hickman, A., and Pawley, M.J., 2004, Critical tests of vertical vs. horizontal tectonic models for the Archaean East Pilbara granite-greenstone terrane, Pilbara craton, Western Australia: Precambrian Research, v. 131, p. 173–211, doi:10.1016/j.precamres.2003.12.015.

White, R.S., and McKenzie, D., 1995, Mantle plumes and flood basalts: Journal of Geophysical Research, v. 100, p. 17,543–17,585, doi:10.1029/95JB01585.

Windley, B.F., 1984, The Evolving Continents: New York, John Wiley and Sons, 416 p.

Wyman, D.A., Kerrich, R., and Polat, A., 2002, Assembly of Archean cratonic mantle lithosphere and crust: Plume-arc interaction in the Abitibi-Wawa subduction-accretion complex: Precambrian Research, v. 115, p. 37–62, doi:10.1016/S0301-9268(02)00005-0.

MANUSCRIPT RECEIVED 17 FEBRUARY 2014
REVISED MANUSCRIPT RECEIVED 23 FEBRUARY 2015
MANUSCRIPT ACCEPTED 3 AUGUST 2015

Printed in the USA

Downloaded from http://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/5/563/3045885/563.pdf
by guest