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1	Precambrian Tectonic Evolution of Earth: an Outline
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11	Abstract
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12	Space probes in our solar system have examined all bodies larger than about 400 km in diameter and shown that
13	Earth is the only silicate planet with extant plate tectonics. Plate tectonics is unusual in our solar system and may

14 be unusual in time. Venus and Earth are about the same size at 12,000 km diameter, and close in density at 5.2 15 and 5.5 Kg.m-3 respectively. Venus and Mars are stagnant lid planets; Mars may have had plate tectonics and 16 Venus may have had alternating 0.5 Ga periods of stagnant lid punctuated by short periods of plate turnover. Plate 17 tectonics has clearly operated on Earth since the beginning of break-up of Rodinia at about 0.7 Ga, witnessed by 18 rock associations such as obducted supra-subduction zone ophiolites, blueschists, jadeite, ruby, continental thin 19 sediment sheets, continental shelf, edge, and rise assemblages, collisional sutures, and long strike-slip faults with 20 large displacements. Equally, from rock associations and structures, nothing resembling plate tectonics operated 21 prior to about 2.5 Ga. Contentious questions are: "when did plate tectonics start", "did plate tectonic style and 22 rock assemblages evolve with time?", and "what tectonic mechanism(s) was responsible for shaping pre-plate 23 tectonic Earth?" The many opinions on these issues have been summarised by Korenaga (2013). We conclude, 24 following Burke and Dewey (1973), that there is no evidence for subduction before about 2.5 Ga, and that plate 25 tectonics or, at least, some form of large lateral relative displacement mobilism evolved during the period from 26 2.5 to 2.1 Ga after which "modern assemblages", and long linear/curvilinear deformation belts are developed, and 27 palaeomagnetism indicates that large lateral relative motions among continents had begun since at least 1.88 Ga. 28 Prior to 2.5 Ga there was a stagnant lid. The "boring billion", from about 1.8 to 0.8 Ga, was a period of two super-29 continents, Columbia and Rodinia with substantial intra-plate magmatism and marginal accretionary tectonics.

Modern plate tectonics from about 0.8 Ga is correlated with major glaciations, including the Snowball Earth and
 the appearance of metazoan life. Our conclusions are based, almost wholly, upon geological data sets, including
 geochemistry, with minor input from modelling and theory.

33 Introduction.

34 Plate tectonics sensu stricto (Isacks et al., 1968; Le Pichon, 1968; Mckenzie and Parker, 1967; Morgan, 1968; 35 Wilson, 1965) is the relative motion among torsionally rigid plates with narrow boundary deformation zones in 36 the oceans but, generally, wider in the continents. Significant intra-plate deformation in the oceans is rare, except 37 very close to ridge axes but minor deformation is common in the continents, except for the stiff Archaean cratons 38 with a thick lithosphere, which are almost earthquake-free and commonly wrapped around by younger orogenic 39 terrains. Plate tectonics is a highly-efficient mechanism for global heat-loss by magmatism and hydrothermal 40 convection at the oceanic ridges, with minor conductive heat-loss, supplemented, episodically, by intra-plate 41 magmatism in large igneous provinces and hotspots above mantle plumes, and cooling by subduction. If there 42 was a time on Earth before plate tectonics, neither conduction nor radiation can account for the necessary heat-43 loss so that the only credible mechanism would have been pervasive plume upwelling and related massive mafic 44 magmatism with, possibly, some form of crustal-lithospheric fragmentation and foundering, drip, sagduction, and 45 localised, non-connected small subduction zones to allow surface materials into the mantle. Mantle temperature 46 and heat-loss have been diminishing since the origin of Earth (Bickle, 1978; Bickle, 1986; Korenaga, 2013); 47 therefore one might expect changes in tectonic style, involving a stagnant non-segmented lithospheric lid to plate 48 tectonics.

49 Plate tectonics, at a planetary scale, in our solar system, appears to be restricted, now, to Earth but may have 50 occurred in other planets, especially Venus, and may also be occurring in the moons of the giant gas planets. There 51 has been a plethora of suggestions and substantial disagreement on when plate tectonics began on Earth, and what 52 tectonic regime may have applied before plate tectonics, based upon a large range of criteria (e.g. Cawood et al., 53 2018; Johnson et al., 2017; Moyen and van Hunen, 2012; Smithies et al., 2007). Also, there has been a natural 54 tendency to lean on the doctrine of "the present is the key to the past" perhaps with the notion that plate tectonics 55 sensu stricto is too good a mechanism to waste. Indeed, since the advent of plate tectonics (Wilson, 1965), there 56 has been a naturally enthusiastic but over-zealous tendency to interpret the whole recorded history of Earth in 57 these terms. Uniformitarianism has meant a variety of things to geologists. The episodicity, periodicity, and 58 catastrophic geologically-instantaneous nature of many if not most processes indicates that uniformitarianism is

not strictly true, though this statement depends upon the time over which processes are differentiated and integrated. It is clear that tectonic processes were not uniform through time and that various forms of secular evolution have occurred, as shown in the compilations by Bradley (2011) of "everything through time". Heat production, the temperature of the asthenosphere, and heatflow have evolved with time, from three times the heat production in the early Archaean compared to today, with a corresponding progressive thickening and stiffening of the lithosphere, whatever form of tectonics is operating.

65 There is a plethora of opinions, summarised by Korenaga (2013), on the existence, origin and nature of "plate 66 tectonics" before 0.8 Ga. Stern (2005) argued, from plate subductability, blueschists, ruby, jadeite, and ophiolites, 67 that modern plate tectonics began at about 0.7 Ga, with a stagnant lid prior to that, interrupted by a period of plate 68 tectonics from about 2.1 to 1.7 Ga. Hamilton (2007) argued for a 2.0 Ga start based upon the first large deformation 69 belts, Grieve (1980) for 2.7 Ga, Brown (2006) for 2.8 Ga, Polat and Kerrich (1999) before 3.8 Ga, Nutman et al. 70 (2002) at 3.6 Ga, Komiya et al. (1999) and Kusky et al. (2018) at 3.8 Ga, Shirey et al. (2008) at 3.9 Ga, Hopkins 71 et al. (2008) at 4.2, and Harrison et al. (2005) before 4.4 Ga. There is no doubt, from palaeomagnetism (Evans 72 and Pisarevsky, 2008), that substantial relative horizontal motions took place during the Proterozoic from about 73 1.9 Ga. Smithies et al. (2005a), Cawood et al. (2006), Dewey (2007), Pease et al. (2008), Condie and Kroner 74 (2008), Brown (2008). Van Kranendonk et al. (2007) and Dhuime et al. (2011; 2012; 2015) argued from geology, 75 and numerical modelling, that some form of plate tectonics or, at least, lateral relative motion driven by 76 subduction, started at the beginning of the Neo-Archaean at about 3.1 Ga, marked also by an increase in the growth 77 rate of the continental crust. The uniformitarian plate tectonic view was challenged, initially, by Burke and Dewey 78 (1973), who saw the Archaean as a "permobile" regime characterized by pervasive magmatism and deformation 79 during density inversions when no part of Earth, on any scale, consisted of torsionally rigid lithospheric plates 80 sensu stricto on any scale. They argued that plate tectonics began during the early Palaeoproterozoic.

If plate tectonics is the operator, one might expect to see geological evidence of plates and plate boundaries such as the following: passive continental margin and shelf associations; supra-subduction zone ophiolite complexes *sensu stricto* indicating both fore-arc plate accretion and the relative horizontal motion of obduction; long linear and curvilinear deformation belts between widespread, flat-lying, little or undeformed epicontinental platform sequences: transcurrent faults with large displacements; paired metamorphic belts with adjacent high pressurelow temperature and high temperature-low pressure zones, adakites; subduction-accretion prisms; collision zones with ruby; foreland fold thrust belts especially thin-skinned; and palaeomagnetic evidence of the relative motion 88 of continents. These plate tectonic rock suite indicators might be expected to be arranged in belts and zones rather89 than in blobby patches.

90 It is interesting to speculate on how we might approach the problem of Archaean tectonics were we to know 91 nothing of plate tectonics. We know that Earth has evolved thermally and so perhaps it should not be surprising 92 that the principal heat loss mechanisms of plumes and plate tectonics have also evolved. Many papers depend on 93 one or a small number of distinctive rock types, such as boninite and andesite, but these are not definitive 94 indicators of subduction because the can also form, even in modern Earth, in intraplate settings. For example, 95 boninites have been found in The Manihiki Plateau (Golowin et al., 2017) and abundant calc-alkaline andesites 96 in the Basin and Range (Hawkesworth et al., 1995). Thus, reliance upon certain rock types and chemistries to give 97 definitive solutions may yield incorrect answers.

98 This paper focuses on rock assemblages, patterns, and relationships, with accompanying geochemistry and 99 petrology, that define tectonic regimes and involve mostly field observations and geological maps. In simple 100 terms, the question is asked "do geological patterns, at all scales, resemble those that we know were generated by 101 plate tectonics?" or do they differ. Changing heat production suggests the likelihood of progressive rather than 102 instantaneous change, perhaps with tipping points. If we see a geological pattern like, for example, that of the 103 Lower Palaeozoic of western Newfoundland, we may suppose that it was generated by an oceanic arc with an 104 ophiolitic fore-arc colliding with a rifted continental margin. Colloquially, avoiding ducks, "if it looks and behaves 105 like a cat, it probably is a cat". If it has scales and two legs and swims, it probably is not. Insufficient attention 106 has been paid to rock suites and their structures and arrangements that are seen in outcrop and across regions. For 107 example, andesite, tonalite, and diorite are not a definitive indicators of plate tectonics unless they have arc 108 chemistries and are in linear-arcuate belts that, in turn, indicate a related subduction zone. Also, too much reliance 109 has been placed on numerical modelling at the expense of empirical field data; modelling never trumps 110 observation. The key issue is tectonic style and observed geology. Obviously this is not possible for the Hadean 111 for which we have only a handful of primitive zircons, and where modelling and inference are necessary.

Plate tectonics can be dealt with in fully quantitative terms from 160 Ma by sequential magnetic anomaly and continental margin fitting from which poles and rates of relative plate motion can be deduced. Semi-quantitative plate tectonics is allowed by palaeomagnetism, faunas, facies, and paleoclimatology from the beginning of the Cambrian to the early Jurassic because oceanic crust with its magnetic anomalies has been subducted and vanished along suture zones. In the Pre-Cambrian, there are no faunas of use so that palaeomagnetism and facies allow only sketchy continental reconstructions and analyses of relative motion. All those who have written on the subject
and, probably most geologists, would agree that plate tectonics *sensu stricto* can be traced back, at least, to the
break-up of the supercontinent Rodinia from about 0.8 - 0.6 Ga.

120 The existence of Rodinia requires that it was assembled from older continental masses and fragments involving 121 subduction and collision. It is commonly considered that this was completed at about 1.0 Ga when the Grenville 122 and Namaqua Orogens were formed (Fig 1). There is clear evidence from palaeomagnetism (Evans and 123 Pisarevsky, 2008) that relative motion of continental masses (continental drift) was occurring back to at least 1.8 124 Ga and is permissible back to about 3.0 Ga. Extensive linear to arcuate zones of continental deformation, the 125 oldest of which are the Limpopo and Ubendides of southern Africa at about 2.0 Ga (Fig. 2), are characteristic of 126 continental collision zones. Perhaps the finest and clearest example of a Proterozoic collision zone is the Wopmay 127 Orogen (Hoffman and Bowring, 1984), which is part of an extensive network of late Palaeoproterozoic orogens. 128 These include the Trans-Hudson and Labrador Trough, the latter with the characteristic paired positive-negative 129 gravity anomaly, indicating that that the lithosphere was sufficiently strong to support it without relaxing non-130 elastically to perfect Airy isostasy. Moreover, there is no obvious horizontal strain in the cratons at the time that 131 the bounding Palaeoproterozoic orogens were developed, evidence that they were torsionally-rigid platforms. 132 Care should be exercised with this argument because even modern continental plate boundary zones may be very 133 wide and complex with large torsionally rigid blocks; this is seen in the Himalaya-Tarim and the Cordilleran-134 Rocky Mountain plate boundary zones. A further problem is that, if oceans existed during the Precambrian, they 135 have been subducted, except possibly in small protected patches like the South Caspian. In Phanerozoic suture 136 zones, ophiolites sensu stricto evolved in fore-arcs; therefore, although not ocean floor, an ophiolite fore-arc 137 indicates the subduction of oceanic lithosphere. It is fairly certain, from these general arguments, albeit with 138 caveats, that relative motion between and collision of continents was occurring back to at least 2.0 Ga, with a 139 lithosphere of some torsional strength suggestive of plate tectonics (Fig.1). It can be said, with some certainty, 140 that plate tectonics was operating in its present form since about 0.8 Ga, and something at least resembling plate 141 tectonics since 2.3 Ga. Therefore, in this paper, we explore, mainly, the tectonic history of Earth before 2.0 Ga.

Related critical questions are the volume, composition, and areal distribution of the continental crust. Were there many early small continents or was there a time when continental crust covered the globe with a subsequent loss (Armstrong, 1968; Fyfe, 1978) by massive tectonic erosion leaving stable deformation-resistant cratons? If plate tectonics was not operating in a world of discrete continents, what was between the continents and what was its 146 composition? How can the rate of crustal addition and growth be judged, and how much crustal subtraction and 147 loss took place (Dewey and Windley, 1981)? If there were discrete continents in a stagnant lid, what was 148 happening in the "oceans" between them? What are the lithospheric conditions necessary for subduction? Without 149 subduction, what were the mechanisms for getting water into the mantle to make continents.

150 Plate tectonic indicators in the rock record

151 We now look at the temporal distribution of a number of features that are characteristic of Phanerozoic plate 152 tectonics and providing confidence that plate tectonics was operating if many or all were present at a point in 153 space and time where a range of structural, petrological, facies, and stratigraphical features in the geological record 154 were helpful in distinguishing tectonic regimes. Also, there are geochemical trends and characteristics that help 155 to define the evolution of the mantle and continental crust. A particular problem is that there are time gaps in the 156 occurrence of some features, which may be genuine gaps because of episodicity (e.g. four main times of ophiolites 157 during the Phanerozoic) and periodicity, or false gaps caused by non-preservation at the surface by subduction, 158 subduction-erosion or temperature overprint. A good example is ultra-high pressure metamorphism, which is 159 almost entirely younger than 600 million years with outliers at about 2.3 Ga. Also a particular key feature of plate 160 tectonics, such as blueschists, may not have developed in a hotter regime. Times of supercontinents will have an 161 abundance of marginal arcs, much intra-plate rifting and potassic magmatism, and widespread continental 162 sediment sheets. Times of distributed continents will have abundant rifted margins, oceanic arcs with ophiolite 163 fore-arcs colliding with rifted margins, and Andean-style orogens. We should not expect to be able, always, to 164 tick off a convincing list of features that demand a plate tectonic solution. Figure 1, shows some key events and 165 sequences in the tectonic history of Earth.

166 Hadean (4.55 - 4.0 Ga).

Following the Gaia-Theia collision at 4.51 Ga, the 4.4 Ga Jack Hills zircons in Australia and the Palaeoarchaean 4.03 Ga Acasta gneiss in Canada are the only physical remnants of an event on Earth during the Hadean. The nature of the Jack Hills zircons and the lithology from which they were derived, whether silicic or, mafic, is disputed. Harrison et al. (2005) have argued that the trace chemistry of the zircons points to plate tectonic processes and the growth of continental crust for their origin. We suggest that an opposite conclusion may be drawn from simple geological reasoning. If there had been a substantial amount of Hadean granitic, island arc, or TTG crust, why did none of it survive? The survivability of buoyant, weak continental crust is the reason for mountains and for the preservation of TTG rocks from 4.03 Ga. The lack of extant Hadean continental crust
implies that there never was any and that the Hadean crust was largely mafic and later subducted. Furthermore,
the absence of granitoids suggests the absence of water until the first TTG's appear in the Eo-Archaean and no
plate tectonics.

178 We envisage the Hadean world to have been, like the Eo-Archaean Moon, one of hot, anhydrous conditions in a 179 stagnant mafic lid, bombarded by bolides. Our reconstruction of the Hadean is similar, in its essentials, to that of 180 Grieve (1980), summarized briefly as follows. Hadean Earth was a dry planet with heat production two to four 181 times that of today (Bickle, 1978; Bickle, 1986; England and Bickle, 1984). The stagnant lid lithosphere must 182 have been thin with a fifteen kilometre basaltic-komatiitic plateau-like crust, shallower than today, generated by 183 a high degree of partial melting of hot mantle (Debaille et al., 2013; Fischer and Gerya, 2016). Continuous bolide 184 bombardment generated multi-ring impact basins with massive excavation, lithospheric fracturing, and mafic 185 volcanism induced below impact sites with both shallow and deep plumes as the mechanisms of global heat-loss 186 and the origin of a probably komatiitic crust. Small felsic pools may have developed, from which the Jack Hills 187 zircons were derived, and there was likely reworking of intra-basin volcanics, and impact lithologies. Impact 188 craters, such as those on the Moon, probably reflect the geometry and topography of Earth's surface in Hadean 189 times, whereas the Vredefort crater in South Africa, although younger at 2025 Ma, preserves the geological 190 responses to cataclysmic impacts upon both basement granitic crust and younger sediments. The only Hadean 191 rocks to have been found to date remain the Acasta Gneiss (Bowring et al., 1989b).

192 Eo-Archaean (4.0-3.6 Ga)

193 The main key to understanding the early tectonic behaviour of Earth is the plagioclase-rich tonalite-trondjemite-194 granodiorite (TTG) of the earliest continental crust, the Ancient Grey Gneiss (AGG), exposed in the Kaapvaal, 195 Greenland, Slave (Acasta Gneiss; Bowring et al., 1989a, b; Bowring and Williams, 1999), Zimbabwe, and Pilbara 196 Cratons. The AGG is, in places, overlain unconformably by Archaean komatiite sequences but, more commonly, 197 has tectonic contacts at dome margins with mafic and ultramafic volcanic and sedimentary sequences (greenstone 198 belts). TTG's vary substantially from low-pressure derivatives, generated by the hydrous partial melting of garnet-199 free, plagioclase-rich, amphibolites and low-magnesium basalts at up to 900°C at about 35 km, to high-pressure 200 from garnet – amphibolites or eclogites at greater than 60 km depth (Adam et al., 2012). Eoarchaean TTG's are 201 low pressure. Therefore, the crust and mantle must have acquired water by about 4.0 Ga and there must have been 202 either a very thick (>35 km) plateau-like basaltic crust and/or a mechanism to transport mafic rocks into the

shallow mantle by subduction or sagduction. We suggest that Earth's water was acquired from icy comets during the late heavy bombardment, which also fractured the lithosphere to allow the sinking of giant mafic slabs that could have been the source from which the TTG's partially melted. We suggest that a thick mafic crust on a thin lithosphere with a geothermal gradient three to four times that of today, perhaps aided by impact-induced foundering, was the likely TTG source. Johnson et al. (2017) have argued, from petrogenetic modelling, that Eoarchaean TTG's were derived from the bases of early basalt sequences with continuous and multiple cycles of volcanism, burial, partial melting and remobilization of TTG's in which there is no a priori need for subduction.

210 Palaeo- and Mesoarchaean (3.6-2.8 Ga).

211 Palaeoarchaean rocks occur in all continents and have strikingly similar lithological assemblages and structures 212 that are wholly dissimilar from younger terrains. They form the earliest cratonic cores, such as the Kaapvaal and 213 Pilbara. Today, they are almost earthquake-free, except for mining-induced seismic activity, have low heat-flow, 214 and thick lithosphere, up to 350 km. They have resisted subsequent regional deformation and commonly form 215 hard knots around which younger orogens are moulded. They have, mostly and regionally, low-grade 216 metamorphism and are little eroded; when formed, they were roughly the same crustal thickness as today. 217 Typically, they have the classic TTG dome and greenstone keel structure (Collins et al., 1998; Choukroune et al., 218 1997). In the Kaapvaal (Anhaeusser et al., 1969) the thick Onverwacht komatiite /chert/greywacke sequence is 219 overlain by the Fig Tree argillite/chert and the Moodies conglomerate/sandstone. The TTG's are in domal 220 structures, typically about 30 km in diameter. The structural fabric generally varies from almost isotropic in the 221 cores of the domes, with increasing flattening fabrics outwards, to plane strain near and at the margins suggesting 222 ballooning. In the greenstones keels, strain is strong from more plane strain fabrics at their margins to vertical 223 stretching fabrics in their cores. Pilbara (Collins et al., 1998), Yellowknife (Drury, 1977) and Dharwar 224 (Choukroune et al., 1997) structures and fabrics are very similar. Bickle et al. (1980) have argued, for the Pilbara, 225 that a pre-doming deformation event was responsible for shortening and early flat-lying structures along the 226 greenstone-granite contacts. We interpret these structures as having formed along the contacts during diapiric 227 inversion. Jackson and Talbot (1989) and Van Kranendonk et al., (2007) have demonstrated the complex 228 polyphase structural complexities that can develop in TTG-greenstone terrains along the margins of mushroom-229 shaped diapirs – mechanisms include dome margin shear, compression by ballooning, horizontal shearing at the 230 base of overhangs, and compressional constriction in keel cores. Glazner (1994) has shown that mafic intrusions 231 may also enjoy solid-state vertical motions in the continental crust. His model involves the ascent of mafic magma

to a position of neutral buoyancy. Upon cooling, the now denser mafic body sinks at rates of several kilometres
 per million years to a new, deeper, level of neutral buoyancy. These silicic-mafic buoyancy-driven inversions may
 be important in driving crustal stratification

235 We interpret these relationships as follows. The Onverwacht (3.56–3.33 Ga) and Fig Tree were laid down as a 236 thick sequence on a hot Eoarchaean TTG basement, followed by crustal inversion, the hot mobilized TTG 237 basement rising as spreading/ballooning/inflating diapiric domes and the greenstones sinking and compressed 238 between the domes (e.g. Schwerdtner et al., 1983). The Moodies was probably mainly restricted to early 239 depressions between the domes and represents the stripping of the Onverwacht, Fig Tree, and some TTG from 240 the dome heads. from the rising domes. These events were terminated by little-deformed, late, high-level, sill-like 241 sheets of potassic granite and small syenite plutons. Although commonly strongly-deformed, the Onverwacht, Fig 242 Tree, and Moodies have a clear stratigraphy in the main Barberton Synclinoriun (Anhaeusser, 1969) that precludes 243 their arrangement in assembled terranes as argued by Lowe (1982). Also, the Onverwacht is clearly a natural 244 sequence of komatiite flows, is not a series of thrust stacks, is not oceanic (see arguments for the Neo-Archaean 245 Belingwe sequence of Zimbabwe below), and cannot be considered to be part of an ophiolite complex as has been 246 suggested (De Wit, 1982; De Wit, 1991). Grosch and Slama (2017) argued for the presence of an ophiolite-type 247 sequence preserved in the Barberton greenstone belt (BGB). They combined new field observations with detrital 248 U-Pb zircon geochronology and geochemistry on fresh drill-core material from the Kromberg type-section 249 sequence of mafic-ultramafic rocks in the Onverwacht Group of the BGB. Trace element geochemistry indicates 250 that the Kromberg metabasalts were derived from the primitive mantle. The ENd values and Nd model ages of the 251 metabasalts record a depleted Archaean mantle source similar to CHUR (chondritic uniform reservoir) with no 252 continental [TTG]) crustal contamination. U-Pb geochronology by laser ablation–ICPMS on detrital zircons from 253 an uppermost chert unit indicate a homogeneous age distribution and a gabbroic source in the greenstone belt, in 254 direct contrast to zircons from felsic conglomerates that structurally underlie the Kromberg sequence. Grosch and 255 Slama (2017) suggested that, collectively, the new data and field observations indicate that the 3.33 Ga Kromberg 256 mafic-ultramafic sequence formed in a juvenile oceanic setting and represents a remnant of tectonically accreted 257 oceanic crust, and that horizontal plate tectonic processes were operating on the Archaean Earth as early as 3.6 258 Ga. The principal problem in regarding the Kromberg sequence as an ophiolite, generated at an oceanic spreading 259 ridge, is that although it is thrusted (obducted) onto the Noisy Formation diamictites, turbidites and tuffs in the 260 section described, it appears, elsewhere, to be part of the regular Onverwacht volcanic stratigraphy. Also, although 261 there are some lithologies (dunites, gabbros, and pillow basalts) that appear ophiolitic, they are not arranged in

anything resembling an ophiolite sequence and there is, critically, no sheeted dyke complex or tectonisedperidotite.

The dome and keel structure was generated by vertical inversion of a lighter, hotter Eoarchaean TTG crust overlain by a heavier cooler komatiitic Palaeoarchaean volcanic sequence in a geothermal gradient three times that of today, which would have decreased viscosity and facilitated inversion. The Palaeoarchaean Isua Dome in west Greenland with its flattened and stretched envelope of low-grade meta-volcanics and sediments is very similar to the Barberton TTG domes and was probably also formed during crustal inversion The geology and evolution of Eo- and Palaeoarchaean terrains, in our opinion, is wholly inconsistent with developing in a plate tectonic Earth.

270 Neoarchaean (2.8 Ga to 2,5 Ga)

271 By the end of the Mesoarchaean, the Kaapvaal Craton was established upon which the Pongola, Witwatersrand, 272 Ventersdorp, and Transvaal volcanosedimentary sequences were deposited (Fig.1). The Pongola was probably 273 deposited in a rift indicating that the Kaapvaal lithosphere was sufficiently strong to crack and, in Ventersdorp 274 times, to be penetrated by komatiites, basalts and intermediate volcanics. Little-deformed platform carbonates and 275 banded iron formation of the Transvaal Group were deposited across the Kaapvaal Craton as three unconformable 276 sequences in two basins separated by the Vryburg Arch indicating a relatively stable platform and cratonization 277 by this time. Similarly, in Northwestern Australia, the stabilized Pilbara Craton is overlain by the sediments and 278 volcanics of the Fortescue Formation and the banded ironstones of the Hamersley Basin.

279 However, in the Zimbabwe Craton, the Superior Province and the Yilgarn Craton, the Neoarchaean record is quite 280 different where, within the period 3.1-2.5 Ga, a sub-linear TTG/greenstone crustal fabric was developed. In 281 Zimbabwe, the Belingwe-Bulawayan sequence cycles (2.9 -2.55 Ga) of komatiites, and mafic to silicic volcanics, 282 and sediments arranged in synclinal tracts ranging from broad and open and shallow (Belingwe) to tight and deep 283 (Bulawayo) The distinctive Zimbabwe cover sequence, including thick komatilites, shows minor facies variations 284 but is otherwise uniform and rests with a mapped unconformity upon a Palaeoarchaean TTG-komatiite-basalt 285 (Tokwe/Shibane/Sebakwian) basement (Orpen and Wilson, 1981)) and are intruded by a 2.55 Ga potassic granite 286 suite.. This arrangement is similar to the Palaeoarchaean TTG/greenstone inversion in the Kaapvaal Craton; 287 Zimbabwe well illustrates the various stages of basement- cover inversion from mild and gentle basins to deep 288 and steep keels.

289 The Superior Province has been interpreted, commonly, as having originated in a plate tectonic regime with 290 volcanic assemblages interpreted as volcanic arcs, sea mounts, oceanic plateaux fore-arcs, and back-arc basins 291 (Capdevila et al., 1982; Corcoran and Mueller, 2007; Davis et al., 1989; Davis et al., 1988; Desrochers et al., 292 1993; Dimroth et al., 1986; Hoffman, 1990; Jackson and Cruden, 1995; Ludden and Hubert, 1986; Mueller et al., 293 1996; Percival and Williams, 1989; Polat et al., 1998; Wyman, 1999). The Abitibi, and other, belts in the Superior 294 Province have 2.8-2.7 Ga (Percival and Williams, 1989) tholeiite/komatiite (Pyke et al., 1973), calc-alkaline, and 295 bimodal volcanics (Glikson, 1979; Glikson and Derrick, 1978; Goodwin, 1982; Schwerdtner et al., 1979), with 296 an underlying 3.8-2.8 Ga TTG basement and intruded by 2.7-2.65 "post-orogenic" potassic granites. Commonly, 297 the volcanic sequences are concentric around TTG domes (Goodwin, 1982) giving a regional blobby pattern 298 reminiscent of Pilbara, Dharwar, and Zimbabwe patterns. Volcanic sequences are commonly multi cyclic from 299 mafic to silicic (Goodwin, 1968), unlike stratigraphical arrangements in modern arcs. There are clear belts and 300 some linearity, which Bedard (2018) and Bedard el al. (2013)explain by compression resulting from the 301 convective drag of craton keels causing the collapse of weak zones, sagduction, and the collision of more rigid 302 zones. Calvert et al. (1995) have interpreted shallow-dipping reflection images as evidence of a subduction zone 303 but Ji and Long (2006) argue that other reflectors, such as folded structures can be responsible. Alternatively, the 304 structures could be thrusts developed under the compression inferred by Bedard.

305 We suggest that both Palaeoarchaean and Neoarchaean terranes were built by the same or very similar TTG 306 basement and basalt-komatiite cover inversion sealed by a terminal craton-sealing high-level granite-syenite 307 event, commonly in thick sills. The Yilgarn Craton has a north-north-west TTG/greenstone belt linearity with 308 TTG domes and greenstone keels and probably originated in the same way. All these Neoarchaean terranes have 309 an older basement upon which komatiite/basalt/calc-alkaline sequences rest, commonly unconformably, and are 310 sealed by a late potassic phase of granite/syenite magmatism. There are no obvious sutures within them. They are 311 arranged in more linear patterns than Palaeoarchaean-style terranes, the first of which was the 3.1 Ga 312 Pietersburg/Murchison Greenstone Belt along the Northern side of the Kapvaal Craton. We regard their volcanic 313 components as pseudo-arcs, containing a significant proportion of andesites but not of subduction-derivation. 314 Lastly, their deep structure is shown in the Kapuskasing Belt (Percival and Card, 1983) along which the Superior 315 Belt was turned up along its north-west margin. Here, almost the whole Superior crust is shown in cross-section 316 as follows from top to base: up to 10 km greenstone belt metavolcanics underlain by 10-20 km of tabular and 317 xenolithic gneissic tonalite and granodiorite, with a basal 20-25 km section of older gneissic granitoid

assemblages. This does not favour an interpretation of multiple colliding arcs but rather a volcanic assemblagebuilt upon an older granitoid crust in a stagnant lid.

320 Many Archaean mafic-ultramafic associations have been claimed to be ophiolites (Friend and Nutman, 2010;

- **321** Furnes et al., 2014; Furnes et al., 2015; Komiya et al., 1999; Polat et al., 2002) none of which resemble ophiolites.
- 322 They are scraps and slivers of, variably, pillow lava, gabbro, and serpentinite that could be of any origin; none
- 323 have sheeted dyke complexes, tectonized harzburgites, or an ordered ophiolite sequence.

324 Palaeoproterozoic (2.5-1.6 Ga)

325 During the early Proterozoic, the first arcs appeared in the Birimian (Combs, 2018) from about 2.3 Ga. Narrow 326 linear to curvi-linear deformation zones developed, bounded by cratonic platforms, including the 1.8-1.6 Ga 327 Capricorn Belt, the 2.0 Ga Limpopo and Ubendides of Africa, the 1.9-1.75 Ga Trans-Hudson, 328 Coronation/Wopmay, Cape Smith and Labrador Trough in Canada, and the 1.8-1.7 Ga Mazatzal of the US. The 329 North American orogens weld together seven Archaean provinces and have been interpreted as collision zones 330 (Hoffman et al., 1989). These belts and their margins have many of the hallmarks of Phanerozoic orogenic belts, 331 and were clearly developed in a plate tectonic milieu. They include the characteristic paired positive/negative 332 gravity anomalies, forelands with rifted margin and fore-deep sedimentary sequences, continental shelf prisms, 333 aulacogens, foreland fold and thrust belts, pre- and syn-collisional magmatic zones including adakites and 334 boninites (Wyman, 1999), transcurrent structures, UHT metamorphism and deformation, and substantial 335 basement re-activation. Structures indicative of subduction zones are seen in several places in these belts. This 336 suggests that some form of plate tectonics was operating on Earth from about 2.3 to 1.7 Ga although it is difficult 337 to gauge the degree of plate torsional rigidity. Also there are no blueschists, although thermal conditions may 338 have prevented the formation of low temperature/high pressure metamorphism. The Jormua ophiolite appears not 339 to be a typical supra-subduction zone-ophiolite but rather a slice of an ocean-continent trasition zone obducted 340 about 100 km onto an adjacent platform. Holder et al., (2019) have shown that paired metamorphic belts exist 341 back to the early Proterozoic, but the distinction between the hot, low-pressure and cold high-pressure belts 342 diminishes back in time to zero at 2.5 Ga. The oldest eclogites appear to be outliers to Neo-Proterozoic-343 Phanerozoic occurrences. The period 2.3-1.7 also includes ophiolites indicative of sea-floor spreading. Although 344 no relics of normal oceanic crust are preserved, except perhaps as accreted fragments, the Jormua ophiolite appears 345 to represent a slice of an ocean-continent transition OCT) zone obducted about 100 km onto an adjacent platform 346 and has characteristics of modern OCT terranes such as Galicia.

347 In the Palaeoproterozoic, a global mafic outburst is witnessed by a number of LIPS with dykes, sills, and plutons, 348 for example the 2.055 Ga Bushveld Complex, the 2.45 Ga Great Dyke, the 2.45 Ga Matachewan dyke swarm 349 (Hoffman, 1990) and the 2.2 Nipissing mafic dyke and sill complex, all evidence for vigorous plume activity 350 accompanied by extensional deformation in a brittle lithosphere. This vigorous mafic magmatism and a 351 widespread coherent oceanic lithosphere may have been the harbinger and cause of the development of plate 352 tectonics. At 2.5 Ga, there was a massive high volatile flux from the mantle at at least ten times the present rate 353 (Marty et al., 2019) based upon an Archaean ¹²⁹xenon deficiency relative to modern compositions. Marty et al. 354 argue that this massive outgassing could not have occurred in a plate tectonic regime but represents a 300 million 355 year burst of mantle activity. Also there was a global outbreak in sanukitoid (high Mg granitoids) intrusion at 2.5 356 Ga. We suggest that, from 2.5 Ga, there began a phase of vigorous convection pluming, mafic magmatism, and 357 lithospheric fragmentation that initiated true subduction and the establishment of plate tectonics, which has 358 continued unabated until today.

359 The (so-called) Boring "Billion" (1.8-0.8 Ga)

360 A stumbling block has been argued against plate tectonics during the period 1.8-0.8 Ga, the so-called boring 361 billion. This was the period of the Columbia super-continent leading to the Rodinia supercontinent, which began 362 to break up and disperse during the Neoproterozoic. Stern (2005) suggested that the boring billion was an interval 363 of stagnant lid with widespread potassic magmatism and minor rifting in US granite/rhyolite terrains expressed 364 in Pikes Peak 1.08, Tishomingo 1.375, Wolf River 1.468, and the Wausau Syenite at 1.45-1.45 Ga and 1.34-1.08 365 Ga (Bickford and Van Schmus, 1985; Bickford et al., 1981). A stagnant lid is difficult to reconcile with the 366 horizontal extension in the Keneewenan Rift, the Grenville, Namaqualand-Natal, and Sveco-Norwegian 367 Orogenies at about 1.0 Ga, and the Gothian (1.5-1.4 Ga) and Hallandian (1.5-1.4 Ga) Orogenies (Stephens, 2020). 368 There are also reasonably well-preserved, likely subduction initiation, ophiolites of about 1.0 Ga age in the East 369 Sayan terrane in southern Siberia (e.g. Belyaev et al., 2017) and in the Yangtse Craton in central China (e.g. Peng 370 et al., 2012). Thus, we see no reasons to deny plate tectonics during the boring billion.

371 Igneous rock types

372 Igneous rocks potentially provide the 'smoking gun' needed to establish the existence of plate tectonic processes
373 in the Archean. However, mid-ocean ridge basalts (MORB), as indicators of divergent plate margins, generally
374 owe their compositions to shallow melting of slightly depleted mantle with moderate potential temperatures

375 (Mckenzie and Bickle, 1988). As these conditions would likely not apply to an ocean ridge in an older, and hence 376 hotter, Earth, the absence of clear examples of Archean MORB (e.g. Pearce, 2008) is insufficient reason to rule 377 out plate tectonics. Much more robust indicators, therefore, are rock types related to convergent plate margins, 378 where subduction of cooler crustal materials beneath hotter mantle should give distinctive compositions that are 379 significantly independent of the age of the Earth at the time. These volcanic arc rock types group as: 1) the products 380 of crystallization of water-rich magmas, such as the calc-alkaline basalt-andesite-dacite-rhyolite (BADR) series, 381 2) the products of melting of hydrated, depleted mantle such as the boninite-high-Mg andesite (HMA) series, and 382 3) the products of subducted slab melting such as adakites. Their respective plutonic equivalents include 383 granodiorites, gabbro-norites and TTGs. We examine these three groups in turn below.

384 Arc-like BADR sequences are rare but do exist in the Archean. Perhaps the best examples are the c. 2.7 Ga Blake 385 River and Confederation Assemblages of the Superior craton (Wyman and Hollings, 2006), part of the c. 2.8-2.7 386 Ga Youanmi terrane of the western Yilgarn craton (Wyman and Kerrich, 2012) and the 3.12 Ga Whundo Group 387 on the western edge of the Pilbara Craton (Smithies et al., 2005b). They differ in detail from Phanerozic arc 388 assemblages, in particular by being less porphyritic and having a complex range of rock types in addition to the 389 BADR series which may include boninites, high-Mg andesites (HMA), and Nb-enriched basalts (NEB), picrites 390 and adakites. The principal proponents of Archean plate tectonics have attributed this difference to the prevalence 391 of hot and relatively flat subduction in the Archean (Polat and Kerrich, 2006), noting that these rock types are 392 found on present day Earth in areas with hot slab-mantle interfaces such as ridge subduction and slab windows. 393 They also highlight the fact that these rock types do typically have the chemical signatures of subduction such as 394 negative Nb anomalies.

395 Arguments against a subduction origin for Archean BADR series were summarized by Bédard et al. (2013). A 396 particular question raised by Bédard and others is why such BADR series are so rare in the Archean and why they 397 are typically bimodal (basic-acid), with andesites relatively rare. To paraphrase, and extend their argument, surely 398 1.5 Ga of hot subduction, with slab fusion adding silica to mantle wedges and abundant water (from subducted 399 serpentinite in particular) available to promote calc-alkaline fractionation, should have produced many more 400 BADR sequences and, in particular, much more andesite. In response to geochemical arguments, they point to 401 evidence that the negative Nb anomalies and related characteristics may be generated by magma-crust interactions 402 as well as subduction (e.g. Pearce, 2008). They propose that Archean BADR sequences are the product of 403 interaction between hot, plume-derived basic magma and Archean crust rather than plate subduction. While

404 accepting the present lack of consensus, we favour the conclusion of Smithies et al. (2018), based on detailed 405 interpretation of the Nb anomalies, that most Archean BADR series are indeed the products of magma-crust 406 interaction but that a small subset do involve some form of transient subduction or sagduction process not 407 necessarily, linked to global plate tectonics.

408 Of the high-Si, high-Mg rock types, boninites are one of the principal rock types to have been linked to subduction. 409 Boninites were once regarded as diagnostic of subduction, but the discovery of boninites in the Cretaceous 410 Manihiki oceanic plateau (Golowin et al., 2017; Timm et al., 2011), in particular, demonstrated that boninites can 411 also form in plume terranes. It is therefore useful to subdivide past boninite lavas into two groups: 'high-Si 412 boninite, HSB' (with SiO₂>57 at MgO =8 wt. %); and 'low-Si boninites, LSB' (with SiO₂=52-57 at MgO =8 wt. 413 (Pearce and Reagan, 2019). The HSB group is characteristic of the boninite type area (the Bonin forearc south of 414 Japan), has been found in many ophiolites and basal arc sequences in Phanerozoic and Proterozoic orogenic belts, 415 is characteristic of subduction initiation terranes, and is subduction-specific. The oldest example found so far on 416 Earth marks the start of arc volcanism in the Trans-Hudson Belt at 1.9 Ga (Wyman, 1999). The LSB group co-417 exists with the HSB group in subduction initiation terranes but is found also at slab edges and in a few oceanic 418 arcs, forearc basins and back-arc basins, and in the Manihiki intraplate setting. LSB have been reported from 419 several Archean settings.

In the detailed evaluations of proposed Archean boninites by Smithies et al. (2004) and Pearce and Reagan (2019), three sets of boninite localities emerged as most likely to be linked to subduction, all classifying as LSB and all having subduction-like geochemical signatures. The first set (the 'Whundo type' of Smithies et al, 2004) comprises the boninites from within the BADR sequences of the Yilgarn (Lowrey et al., 2019), Superior (e.g. Boily and Dion, 2002) and Pilbara (Smithies, 2002) cratons. These are the most likely to have had an origin in an active volcanic arc, and the same arguments for and against plate subduction made in the above discussion of BADR series apply to these boninites.

The second set of boninites (the 'Mallina type') are associated with silicic high-Mg basalts (SHMB) and their plutonic equivalents (high-Mg diorites, or sanukitoids) and are found within intracratonic rifted basins, notably the c. 3.0 Ga Mallina Basin in the Pilbara terrane) (Smithies, 2002; Sun et al., 1988) and large igneous provinces (such as that hosting the 2.7 Ga Stillwater complex, Helz, 1985). These boninite-SHMB associations are restricted to the period 3.0 to 2.0 Ga (Pearce and Reagan, 2019). As the authors cited above have all shown, high-Si, low-Mg compositions are indicative of depleted mantle lithosphere refertilized by a subduction-like component. Thus, 433 although their geological settings show that they are not directly subduction-related, they do carry an inherited 434 record of older subduction events. Their depleted sources and restricted time period are most consistent with a 435 subduction-like refertilization event related to craton accretion. The final set of boninites is that reported for some 436 of the oldest volcanic series, within the small, highly deformed and metamorphosed 3.8-3.7 Ga volcanic sequences 437 at Isua, SW Greenland (Polat et al., 2002) and Nuvvuagittuq, Northern Quebec (O'Neil et al., 2011; Turner et al., 438 2014). Pearce and Reagan (2019) found that the Isua 'boninite like-rocks' were most similar to the low-Ti basalts 439 - boninite rocks from the Manihiki plateau, but that the Middle Unit of the Nuvvuagittuq sequence did appear to 440 have robust characteristics of low-Si boninites. More work is needed to confirm these inferences and link to plate 441 tectonic processes.

442 Adakite is another rock type that was once viewed as a subduction indicator, but now has alternative 443 interpretations. Adakites, and their supposed plutonic equivalents, TTGs, have long been explained in terms of 444 subduction, specifically melting of subducted oceanic crust (e.g. Martin et al., 2005). There is no shortage of 445 TTGs in the Archean with the less-differentiated members becoming deeper in origin (residual amphibole to 446 residual garnet), more magnesian and less silicic from the earliest to latest Archean. This is explained by Martin 447 et al. (2005) in the context of a plate tectonic Earth in terms of increasing subduction dip, in which the high-Si 448 adakites (HSA: >3Ga) are silicic slab melts that have ascended directly to the surface, while low-Si adakites 449 (LSA: 3.0-2.5Ga) are slab melts that have reacted with, or instigated melting of, a peridotite mantle wedge before 450 reaching the crust. It is now, however, clear that these rocks could be the product of crustal melting (of plume-451 derived basalts), in which case the changing characteristics could simply be the result of increasing depth of 452 melting in a steepening crustal geotherm with time. There are many arguments for why crustal melting is the more 453 viable interpretation, the ability of slab melting to generate the observed volume of TTGs being one (Bédard, 454 2013). Nonetheless, the much less voluminous adakite volcanic rocks within BADR sequences could derive 455 directly from either slab melting, crustal melting or fractional crystallization involving amphibole and/or garnet, 456 as is the case with recent arc-related adakites (e.g. Castillo, 2006).

457 Overall, therefore, the direct evidence from igneous rock types for plate tectonics is small, with just a handful of 458 localities having rocks indicative of active subduction or sagduction and no evidence for plate tectonics on a 459 global scale. These localities comprise a small number of BADR series that can better be explained by subduction 460 than crustal assimilation together with a comparable number of intraplate boninite-SHMB series that require 461 refertilization of depleted cratonic lithosphere by subduction-like components prior to a second melting event. In

- 462 our view, as with (Bedard, 2006; Bedard et al., 2013) and Smithies et al. (2018) in particular, and as discussed
 463 further in this paper, this can be achieved without global-scale plate tectonics. If geological arguments for stagnant
 464 lid Archean tectonics are strong enough, as we believe they are, then processes such as localized craton accretion
- 465 could explain any subduction signatures so far.

466 Broad patterns over time

467 Bradley's (2011) compilation shows that there are a number of compelling changes and trends illustrating gradual 468 and sudden changes in earth history that are likely the results of changes from plume to plate tectonics in a cooling 469 earth. The time distribution of almost everything shows rapid change from about 3.0 to 2.0 Ga, through the late 470 Archaean and early Palaeoproterozoic, with the biggest switch in observed geology at the end of the Archaean 471 Eon. Dome and keel greenstone tectonics dominates the tectonic style of the Archaean, whereas mantled gneiss 472 domes are a minor component of the Proterozoic and younger Earth. The Archaean was a more mafic earth with 473 TTGs and komatiites; komatiite is known after 2.5 Ga only in the Eocene of Gorgona, Columbia. From 3.0-2.0 474 Ga, there was a steepening rise in the K₂O/Na₂O ratio and decline in MgO and Cr of igneous rocks and a 475 steepening rise of Rb/Sr in sediments. At 2.5 Ga, there was a sharp increase in TiO2, La, Zr, and Sr in igneous 476 rocks.

477 Mineral deposits with time

Broad trends emerge when relating ore deposit types to geologic age, surficial environment and tectonic setting.
The mechanisms by which the crust has evolved, and the amalgamation and dispersal of continents over time,
have played critical roles in the formation and preservation of all deposit types.

481 Figure 2 illustrates the distribution of major ore deposit types as a function of time, and also with respect to the 482 existence of supercontinents. The distribution of ore deposit types over time reveals a broad association with the periodicity of supercontinent assembly and break-up, with fewer deposits forming in periods of supercontinent 483 484 stasis. Porphyry and epithermal ore forming systems, orogenic Au and MVT Pb-Zn deposits are typically 485 associated with convergence and orogeny (Figure 2) and mineralized districts tend to form close to craton margins 486 and paleo-sutures. Other deposit types may also be linked to dispersal of the supercontinents and divergent 487 margins. It is evident that only diamondiferous kimberlite and PGM-rich layered mafic intrusions have an affinity 488 with intracratonic settings and tend in many cases to form preferentially during periods of supercontinent stasis. 489 The long-lived Mesoproterozoic supercontinent Nuna (or Columbia) is noted for an absence of orogenic ore

- 490 deposit types, but does contain significant accumulations of continental sediment-hosted metal deposits (SedEx)
- 491 as well as kimberlites and intrusion-related iron oxide-copper-gold (IOCG) ores.

492 Evidence exists that the early atmosphere and the precursors to the present oceans formed only at the end of the 493 Hadean era, once the main period of accretion and meteorite bombardment had terminated at circa 3900 Ma 494 (Kasting, 1993). The implications for metallogenesis are that sedimentary and hydrothermal process are likely to 495 have been inconsequential in the Hadean, and any ore deposits that did form at that time were, therefore, largely 496 igneous in character. It is conceivable, for example, that oxide and sulfide mineral segregations accumulated from 497 anorthositic and basaltic magmas at this time. The Eoarchaean, 3800 Ma old, Isua supracrustal belt and associated 498 Itsaq gneisses of western Greenland, for example, comprise mafic and felsic metavolcanics, as well as 499 metasediments, and resembles younger greenstone belts from elsewhere in the world. The Isua belt contains a 500 major chert-magnetite banded iron-formation component as well as minor occurrences of copper-iron sulfides in 501 banded amphibolites and in iron-formations (Appel, 1990). The largest iron-formation contains an estimated 2 502 billion tons of ore at a grade of 32% Fe. Scheelite mineralization has also been found in both amphibolite and 503 calc-silicate rocks of the Isua belt, an association which suggests a submarine-exhalative origin. The coexistence 504 of banded iron-formations and incipient volcanogenic or sedimentary exhalative, massive sulfide deposits points 505 to circulation of seawater through oceanic crust. Although the zones of known mineralization in the Isua belt are 506 sub-economic, at 3800 years old they represent the oldest known ore deposits on Earth.

507 Evidence exists through the Meso- and Neoarchaean Eras for substantial crustal amalgamations, such an early 508 Vaalbara continent and later Superia and Sclavia continents (Bleeker, 2003). The existence of Vaalbara (a 509 combination of parts of the Kaapvaal Craton in southern Africa and the Pilbara Craton in Western Australia) 510 receives support from the similarities that exist in the nature and ages of Archaean greenstone belts and 511 supracrustal sequences on the Pilbara and Kaapvaal cratons (Cheney, 1996; Martin et al., 1998), a feature that is 512 especially striking when comparing the Superior-type banded iron-formations of the two regions. It was 513 previously thought that the late Archaean Witwatersrand basin on the Kaapvaal Craton is unique but exploration 514 in Western Australia, has recently revealed the existence of sedimentary sequences of similar age that also appear 515 to be well-endowed with gold mineralization.

516 The Neoarchaean era represents a period of significant crustal growth and the development of abundant 517 mineralization, formed by processes not unlike those taking place later in Earth history, involving plate 518 subduction, arc magmatism, continent collision and rifting, and cratonic sedimentation. A wide variety of ore519 forming processes therefore characterizes this period of Earth history. Well mineralized examples of continental 520 crust that formed in the period 2800–2500 Ma are represented by the granite-greenstone terranes of the Superior 521 Province of Canada, as well as the Yilgarn and Zimbabwe cratons. Greenstone belts formed from arc-related 522 volcanism host important volcanogenic massive sulfide (VMS) Cu-Zn ore bodies, such as those at Kidd Creek 523 and Noranda in the Abitibi greenstone belt of the Superior Province. Off-shore, in more distal environments, 524 chemical sedimentation gave rise to Algoma type banded iron-formations, examples of which include the Adams 525 and Sherman deposits, also in the Abitibi greenstone belt. Greenstone belts formed at this time also comprise 526 komatiitic basalts that, under conditions favorable for magma mixing and contamination, exsolved immiscible 527 Ni-Cu-Fe sulfide fractions to form deposits such as Kambalda in Western Australia and Trojan in Zimbabwe. 528 During and soon after periods of compressive deformation, major suture zones became the focus of hydrothermal 529 fluid flow derived either from metamorphic devolatilization or late-orogenic magmatism. This resulted in the 530 formation of the varied but common styles of orogenic gold mineralization that are typical of most Meso- and 531 Neoarchaean granite-greenstone terranes worldwide. Examples include important deposits such as the Golden 532 Mile in the Kalgoorlie district of Western Australia and the Hollinger-McIntyre deposits of the Abitibi greenstone 533 belt.

534 Early intracratonic styles of sedimentation, often in foreland basinal settings, gave rise to concentrations of gold 535 and uraninite represented by the ca. 3.0 to 2.7 Ga Witwatersrand basin in South Africa. At least some of this 536 mineralization is placer in origin and was derived by eroding a fertile Archaean hinterland that appears to have 537 been elevated compared to the adjacent basin. The passive margins to these early continents would have developed 538 stable platformal settings onto which laterally extensive Superior type banded iron-formations were deposited. A 539 very significant period for deposition of iron ores such as those of the Hamersley and Transvaal basins of Western 540 Australia and South Africa respectively, as well as the Mesabi range of Minnesota, seems to have been around 541 the Archaean–Proterozoic boundary at 2500 Ma, by which time continental crust was both thick and increasingly 542 rigid.

From a metallogenic perspective, the Palaeoproterozoic is significant because of the major changes that occurred to the atmosphere, especially the rise in atmospheric oxygen levels at around 2300 Ma (GOE). Prior to this, the major oxygen sink was the reduced deep ocean where any photosynthetically produced free oxygen was consumed by the oxidation of volcanic gases, carbon, and ferrous iron. In this environment banded iron-formations, as well as bedded manganese ores, developed, as indicated by the widespread preservation of both Algoma and Superior 548 type iron deposits. The increase in ferric/ferrous iron ratio in the surface environment that accompanied 549 oxyatmoinversion at 2300 Ma, and the accompanying depletion in the soluble iron content of the oceans, resulted 550 in fewer BIFs forming after this time (Bekker et al., 2014; Bekker et al., 2010). The stability of easily oxidizable 551 minerals such as uraninite and pyrite is also to a certain extent dependent on atmospheric oxygen levels and it is, 552 therefore, relevant that major Witwatersrand-type placer deposits did not form after about 2000 Ma.

Metallogenic patterns during the Paleoproterozoic Era were dominated by wide-ranging orogenic processes accompanying plate movements associated with the break-up of Superia and Sclavia and the assembly of Nuna between circa 1800 Ma and 1400 Ma. The break-up of Superia was accompanied, between 2000 and 1700 Ma, by the creation of new oceanic crust and the formation of volcanogenic massive sulfide Cu–Zn deposits such as Flin Flon in Canada, Jerome in Arizona, and the Skellefte (Sweden) – Lokken (Norway) ores of Scandanavia, which are subduction-related in boninite-arc tholeiite sequences.

At 2055 Ma on the Kaapvaal craton, the enormous Bushveld complex, with its world-class PGE, Cr, and Fe–Ti– V reserves, was emplaced, as was the Phalaborwa alkaline complex with its contained Cu–P–Fe–REE mineralization - both generated in intraplate settings (a plume-related Large Igneous Province (LIP) in the case of the Bushveld). In West Africa the period between 2100 Ma and 1900 Ma saw the development of substantial juvenile crust along the margins of the Man Craton during the Eburnean orogeny, accompanied by the formation of extensive orogenic gold mineralization.

565 The amalgamation of Nuna was followed by a long period of cratonic stability that resulted in the deposition, 566 between 1800 Ma and 1500 Ma, of marginal marine sedimentary basins that host the important SEDEX Pb-Zn 567 ores of eastern Australia (Mount Isa, Broken Hill, and McArthur River) and South Africa (Aggeneys and 568 Gamsberg). Anorogenic magmatism was widespread in Nuna times - in South Australia, for example, the 1590 569 Ma Roxby Downs granite-rhyolite complex (Johnson and Cross, 1995), host to the enormous magmatic-570 hydrothermal Olympic Dam iron oxide-Cu-Au-U deposit, was emplaced. Anorogenic granite magmatism at 571 1880 Ma may also have given rise to the later stages of IOCG style mineralization (such as Estrela, Volp, 2005) 572 in the giant deposits of Carajas, Brazil.

In summary, as more evolved continents formed from late Archaean times and conventional plate tectonic
processes appear to have developed, the range of ore deposit types widened, although many have not been
preserved because of tectonic reworking and erosion. The Neoarchaean Era, especially from around 2700 Ma,

576 was a period of significant global orogenesis and many ore deposits formed at this time tend to be arc-related and 577 magmatic-hydrothermal to hydrothermal in nature, not unlike those typifying the Phanerozoic Eon when plate 578 tectonic processes were fully extant. The early stages of the Proterozoic Eon were characterized by a number of 579 major crust-forming orogenies, but the period from about 1800 Ma appears to have been marked by longer periods 580 of tectonic quiescence and continental stability. Consequently, although mineral deposits are not unequivocally 581 diagnostic of evolving crust forming processes, it is nevertheless apparent (Figure 2) that a cyclic pattern linked 582 to the development of supercontinents becomes more readily apparent during the Archaean-Proterozoic transition, 583 and it is from this interval of time, therefore, that a conventional style of plate tectonics is likely to have developed.

584 Diamonds

585 A key question, is whether any biogenic surface carbon has been taken into the mantle to depths over 150 km to 586 make diamond and if so how this relates to tectonic processes. Diamonds and their inclusions are the only natural 587 samples brought to the surface from depths of >150 km. Most common lithospheric inclusions can be subdivided 588 into the eclogitic (e-type) and peridotitic (p-type) with garnet, clinopyroxene and sulphides being common for 589 both sets. Less commonly, olivine and orthopyroxene are reported in p-type inclusions. Because of the very small 590 amount of impurities, it is impossible to date diamonds; however, diamond inclusions make plausible material for 591 dating. Over the last few decades a number of studies have been done on dating diamond inclusions. Shirey and 592 Richardson (2011) compiled the silicate- and sulphide dates from garnet (bulk), clinopyroxene (bulk) and sulphide 593 (single grain) inclusions worldwide and showed that the oldest inclusions are peridotitic (up to 3.5 Ga), while 594 eclogitic inclusions are younger than 3 Ga. Based on this finding, they concluded that the onset of plate tectonics 595 should have occurred after 3 Ga.

596 This was contested by two recent studies on the other sets of diamonds and their inclusions. Smart et al.. (2016) 597 analysed carbon and nitrogen isotopic compositions of Archaean placer diamonds from the Kaapvaal Craton 598 which formed between 3.1 and 3.5 Ga, and concluded that diamonds must have crystallised from a source 599 previously exposed to the surface, arguing the onset of subduction and plate tectonics between 3.1 and 3.5 Ga. 600 Smit et al., (2019) investigated mass-independent fractionation (MIF) of sulphur isotopes, recorded prior to 2.4 601 Ga (Farquhar et al., 2000). Because of the change in atmospheric oxygen between 2.4 and 2.09 Ga, the style of 602 sulphur isotope fractionation changed and only mass-dependent fractionation was recorded after 2.09 Ga. Smit et 603 al., (2019) analysed sulphur isotopes in sulphide inclusions in diamonds from the West Africa, Zimbabwe, 604 Kaapvaal and Slave cratons and reported that MIF sulphur was not observed in the oldest, 3.5 Ga sulphide

inclusions from the Slave craton. Younger (<3 Ga) diamond inclusions from the other studied cratons, however,
contained MIF sulphur, consistent with the hypothesis that subduction operated from around 3 Ga. Despite very
different approaches, all studies on diamonds and diamond inclusions agree at approximately 3-3.1 Ga age for the
onset of subduction.

609 We suggest that diamonds and their inclusions cannot provide a definitive answer to the time of onset of plate 610 tectonics. Sample bias is a serious hurdle to overcome. Dating diamond inclusions (usually very small in size) is 611 not a trivial task, while the lack of samples from a wide range of locations and compositions may constrain 612 obtaining representative ages. Thus, it is possible that eclogitic inclusions older than 3 Ga exist, but have not been 613 sampled yet. The carbon and nitrogen isotopic compositions of diamonds themselves also cannot provide a 614 definitive answer about their source. Carbon isotopic compositions of mantle diamonds have been extensively 615 used to discriminate their origin. Among eclogitic and peridotitic diamond inclusions, the peridotites are 616 significantly more uniform isotopically (with the mean in $\delta^{13}C = -5\%$) and are considered to have formed from 617 mantle carbon, while the eclogites show a wide range of carbon isotopic values ($\delta^{13}C = -41$ to +5%) attributed to 618 a possible subduction origin. The organic carbon formed at the surface is isotopically much lighter ($\delta^{13}C = -40$ to 619 0‰) (Cartigny et al., 2014).

620 The wide range of carbon isotopic compositions in eclogitic diamonds, does not necessarily reflect a subduction 621 origin (e.g. Cartigny et al., 1998a; Cartigny et al., 1998b). First, to yield such extremely light isotopic signatures 622 $(\delta^{13}C = < -25\%)$ the subducted material would need to consist of only organic carbon, with no input from surface 623 carbonates that dominate surface sediments (~80%) and are isotopically heavier ($\delta^{13}C = -15$ to -5%). Second, it 624 is likely that eclogitic and peridotitic diamonds are produced by different mechanisms and involve different fluid 625 speciation. For instance, it has been shown that, even at high temperatures, significant isotopic fractionation could 626 occur between the oxidised (CO₂) and the reduced (CH₄) forms of carbon by means of Rayleigh distillation (Javoy, 627 1972). This could lead to the light δ^{13} C in diamonds being precipitated from the methane-rich fluids. Methane-628 related diamond crystallisation in the Earth's mantle was also detected by combined carbon-nitrogen isotopic 629 studies that do not support a recycled origin of carbon (Cartigny et al., 1998b; Javoy, 1972). Mikhail et al. (2014) 630 have also shown that diamonds in equilibrium with iron carbide are isotopically lighter because of the high carbon 631 isotopic fractionation during the interaction between mantle carbon and native iron at very reduced conditions. 632 Thus, although the surface origin of carbon cannot be ruled out, there are a number of mantle processes that cause 633 carbon isotopic fractionation, triggering the formation of isotopically light diamonds.

634 The lithosphere

635 Lithospheric density, thickness and strength, are likely to have increased in response to diminishing 636 asthenospheric temperature diminishes during planetary history. Cooling increases in a plate tectonic planet as 637 slabs are injected deep into the hot mantle, ultimately causing the asthenosphere to cool, thin, stiffen and, 638 eventually to exterminate plate tectonics. Intra-plate alkalic basalts will increase in proportion. The early mantle 639 was hotter by 100-200°C; the lithosphere was, therefore, thinner, weaker and more buoyant, impeding onset of 640 plate tectonics. Lithospheric strength is also essential for plate tectonics to take place; oceanic lithosphere must 641 be strong enough to remain torsionally coherent during plate motion and subduction. Lithospheric strength 642 increases with thickness, so lithospheric density and strength have increased together as the Earth has aged and 643 cooled. There are three aspects of lithospheric weakness that are required for subduction and plate tectonics to 644 happen. First, the lithosphere-asthenosphere boundary must be weak enough for the lithospheric plates to move 645 over it, accomplished by the hotter and weaker asthenosphere and aided by the concentration of volatiles at the 646 interface. Such a weak zone is likely to have existed since lithosphere first formed, very early in Earth history. 647 The second essential weakness required for establishing self-sustaining and asymmetric (one-sided) subduction 648 is evident in the two-dimensional (2-D) numerical experiments of Gerya et al. (2008) which show that the stability, 649 intensity, and mode of subduction require a zone of weak hydrated rocks above the subducted slab. The weak 650 interface is maintained by the release of fluids from the subducted sediments, oceanic crust, and serpentinized 651 upper mantle as the slab sinks and is pressurized and heated. The third weakness required for subduction and plate 652 tectonics is for the development of large-scale, laterally extensive (~1000 km long) weakness through the 653 lithosphere where a subduction zone can nucleate. Without such weakness, subduction and plate tectonics cannot 654 occur. Such trans-lithospheric weaknesses are produced continuously today on Earth by plate boundaries but it 655 is less obvious how the first trans-lithospheric weakness was produced on a stagnant-lid Earth. It seems likely 656 that the first one may have been produced by interaction of a large mantle plume head with old "oceanic" 657 lithosphere (Gerya et al., 2015), during the global outbreak of mafic magmatism between 2.5 and 2.0 Ga.

658 Subduction and subductability

Plate tectonics requires lithosphere that is dense enough to sink under its own weight. It must be strong enough to remain coherent during self-sustaining subduction, and weak enough to form localized plate boundaries. Such conditions are only likely to happen in a mature silicate planet, as Earth is today. Special circumstances are required for plate tectonics; the planet must be dominantly silicate and conditions must be appropriate for self663 sustaining subduction to occur. Because global plate motions are mostly powered by the sinking of the negatively 664 buoyant oceanic lithosphere in subduction zones (although significant roles for ridge push and mantle convection 665 drag for modifying these motions has been repeatedly proposed) the lithosphere must be denser than the 666 underlying asthenosphere.

667 Self-sustaining subduction of the oceanic lithosphere, as the engine of plate tectonics, cannot occur on a constant-668 size Earth without spreading ridges to balance the loss of area by subduction. Therefore, in a stagnant lid planet 669 (>2,5 Ga), MORB cannot be erupted and the non-continental crust must be formed in some other way, probably 670 by the vertical accumulation of basalts, komatiites, and minor differentiated silicics. Because of the higher 671 temperature of the asthenosphere, the lithosphere would have been thinner and the mafic-ultramafic crust thicker. 672 It seems unlikely that the pre-2.5 Ga crust was wholly continental because subduction would have been unable to 673 start. Also, the present crustal thickness of Archaean cratons was achieved immediately at the end of crustal 674 inversion and the intrusion of late potassic granites and sanukitoids. Most of Earth's continental crust was 675 generated by 2.5 Ga. possibly by the end of the Eoarchaean. Had this crust been globally enveloping, some two 676 thirds of this continental crust would have to have been lost by some process such as tectonic decretion by 677 subduction erosion after 2.5 Ga. Subduction, and hence plate tectonics in a silicate planet, needs lithosphere that 678 is sufficiently dense, stiff, and coherent to sink at weak zones rather than float. The integrated density of present-679 day oceanic lithosphere younger than about 20 million years will float unless attached to older lithosphere, 680 whereas lithospere older than 20 million years is unstable and can sink. Stern calls this the crossover time, when 681 the integrated density of the lithosphere equals the density of the asthenosphere (3.25). A young stagnant lid of 682 non-continental lithosphere will be more buoyant than young MORB lithosphere because of its thick komatiitic 683 crust and its thinner lithosphere resulting from a hotter asthenosphere. A stagnant lid of oceanic lithosphere above 684 a hotter asthenosphere that constantly thermally eroded the lid would not progressively thicken as does lithosphere 685 moving away from a ridge. Therefore, the crossover time of a stagnant lid lithosphere would have been 686 correspondingly longer and preventing the onset of subduction and plate tectonics. We see no evidence of plate 687 tectonics until after 2.5 Ga; therefore, the Archaean crossover time was probably very long, subduction was not 688 possible, and plate tectonics was delayed until after 2.5 Ga.

A stagnant lid is the default mode of planetary tectonics (Stern et al., 2018; Stern et al., 2016). It is the potential
energy resulting from denser lithosphere on above` weak asthenosphere that provides most of the energy for plate
motion (Forsyth and Uyeda, 1975; Lithgow-Bertelloni, 2014). Such a density inversion does not exist for

692 continental lithosphere or for oceanic plateaus, where low-density crust reduces the overall lithospheric density.
693 Such a density inversion was less likely early in Earth history, when hotter mantle resulted in thinner mantle
694 lithosphere and thicker oceanic crust.

695 Continental crust formation and destruction

696 One of the unique features of Earth is its two types of crust: 40 km thick dioritic continental crust and 6 km thick 697 basaltic oceanic crust. Human beings only exist because there are continents for us to evolve on and exploit. If 698 plate tectonics has been extant since 4.6 Ga, Earth's continental crust must be growing with time and must be a 699 product of plate tectonics. On the other hand, if plate tectonics began between 2.5 and 2.3 Ga, there must have 690 been another mechanism for forming continental crust because there are many crustal tracts older than this.

701 The present-day flux from mantle to crust is basaltic and yet the average composition of the continental crust is 702 andesitic. This is the crust composition paradox. A new solution to this paradox is proposed whereby the secular 703 evolution in the composition of the continental crust reflects a changing flux from mantle to crust over time. Thus 704 it is proposed that the present-day composition of the continental crust is a time-integrated average. Crustal growth 705 curves show that at least 50% of the continental crust had formed by the end of the Archaean (Dhuime et al., 2012, 706 2015). A mass balance model based upon a tonalite-trondhjemite-granodiorite (TTG) composition compositional 707 model for the Archaean continental crust shows that the post-Archaean mantle to crust flux was predominantly 708 basaltic and likely a mix of arc-plume basalts. Trace element modeling, however, reveals that additional processes 709 also contributed to the average crust composition. Balancing Y, Ho, and Yb concentrations requires a 710 garnetiferous mafic granulite composition for the lower Archaean crust, which in turn drives the post-Archaean 711 flux toward a high-magnesium andesite. This suggests that there was a slab melt contribution to the continents, in 712 addition to basalt. An excess of fluid mobile elements in the continental crust can be explained either by the 713 addition of a slab melt or small fraction melts. A deficiency in Sr requires that the post-Archaean crustal 714 composition has been modified by erosion. Both Archaean and post-Archaean continental crust contain 715 contributions from basalt and a slab melt. In the Archaean crust the slab melt contribution is dominant. In the 716 post-Archaean crust the basaltic contribution is dominant.

717 There remains, however, the problem of the global distribution of Archaean crust. Some, possibly a substantial718 amount (difficult to ascertain), was recycled as sediment, melted or partially melted to form younger granitoid

plutons and silicic/intermediate volcanics, and structurally reworked into younger orogens. Possible geometric
solutions are;

721 1. There was a global continental coverage of Archaean crust that was segmented and shortened into the extant
722 Archaean cratons, which cannot be correct because there is insufficient post-Archaean shortening in the Archaean
723 cratons.

2. During the later plate tectonic regime, large amounts of a global Archaean continental crust were lost by tectonic decretion/subduction erosion. Tectonic decretion is sufficient, today, that crustal subtraction is slightly larger than addition (Dewey and Windley, 1981). Armstrong (1968) argued that the volume of the continental crust has remained roughly constant since about 3.8 Ga, either by a subtraction addition balance or that all crustal growth was during the Archaean. Fyfe (1978) went further, arguing for continental growth until about 2.3 Ga then continental diminution to the present day.

3. The present volume of continental crust in the present cratonic nuclei is all that ever existed. This means that about 86% of Earth, during the Archaean, was covered with remnant Hadean oceanic lithosphere, none of which is preserved directly or its existence suggested by rocks generated by plate tectonics. We think it inconceivable that a non-plate tectonic inversion process could have been occurring to generate the structure of the continental crust, while, in a surrounding ocean, plate tectonics was generating ridges, arcs and collisions from which none of the rock results are preserved. Therefore, we incline towards massive tectonic decretion from about 2.3 Ga and subduction of any remaining Hadean lithosphere.

737 Evolution

738 Biological evolution is driven by isolation and competition. Isolation allows new species to evolve and 739 competition selects the organism that has best adapted to the environment. Plate tectonics is an mechanism for 740 creating and destroying biological environments such as continents, continental shelves, deep ocean basins, island 741 arcs, and mountains, and therefore is also an unparalleled engine for isolating and recombining ecosystems, 742 favouring speciation and competition. In contrast, stagnant lid tectonics has far less ability to create and destroy 743 biological environments. The implications for evolution are obvious: plate tectonics favours rapid evolution, 744 stagnant lid tectonics does not. Yet we see almost three and a half billion years of stromatolites and other primitive 745 organisms with a rapid metazoan evolution in the Ediacaran. Thus, almost two billion years of plate tectonics did

- not produce a metazoan evolution. Perhaps, in spite of the Great Oxidation event at 2.3 Ga, oxygen levels only
- became sufficient, when they climbed rapidly from about 3% to 13% at about 0.7 Ga.

748 Conclusions

749 We consider 2.5 Ga, the end of the Archaean as the beginning of a time of fundamental change in Earth's tectonic 750 behaviour that led to the establishment of plate tectonics by about 2.3 Ga. Consensus is not a necessary prerequisite 751 for truth but there is general agreement, for all the reasons given by Stern (2005) that plate tectonic has operated 752 since at least 0.7 Ga. These include ophiolites, blueschists, UHP metamorphism, sutures and collision zones, 753 paired metamorphic belts, zones of substantial crustal shortening, major strike-slip faulting, and linear adakite 754 zones. Collectively and from their geological arrangements, these have a clear plate tectonic fingerprint. 755 Palaeomagnetism shows that there was relative motion among continental cratons back to at least 1880 (Condie 756 and Kroener, 2008), The Wopmay and other Palaeoproterozoic orogens such as the Trans-Hudson, the Capricorn, 757 and Limpopo, are belts of, variably, adakites, miogeoclinal platforms, the classic paired positive-negative gravity 758 anomalies, suturing, and horizontal shortening, all of which suggest a plate tectonic origin back to at least about 759 2.0 Ga. These orogenic belts weld together Archaean cratons and are best-described as collisional orogens, 760 themselves an indication of the former subduction of some form of oceanic tract., either a stagnant lid lithosphere 761 or lithosphere generated by sea-floor spreading; if the latter, this pushes plate tectonics back before 2.0 Ga. The 762 2.04-2.24 Ga Birimian in the Reguibat of Mauritania and the Man-Leo Shield of Cote d'Ivoire (Combs, 2018) is 763 very similar to the Neoproterozoic Pan-African and models for the Phanerozoic of central Asia (Sengor et al., 764 1993), comprising huge tracts of accreted volcanic arcs, also takes plate tectonics back to at least 2.3 Ga. Holder 765 et al (2019) made the key analysis and observation that paired metamorphic belts, distinctive of convergent plate 766 boundary zones, become weaker in their bimodality back in time until the bimodality vanishes between 2.2 and 767 2.4 Ga, supporting the advent of plate tectonics at this time.

Many workers in Archaean (>2.5 Ga) terrains have developed plate tectonic models for these rocks (see Korenaga, 2013)), although there is a small group of dissenters (Davies, 1992; Dewey, 2018a; Dewey, 2018b; Dewey, 2019; Hamilton, 1998; Hamilton, 2003; Hammond and Nisbet, 1992; Padgham, 1992). The blobby to sub-linear TTG dome and greenstone keel pattern that dominates the upper crust from about 3.6-2.6 Ga and the stratigraphical sequences of the Archaean show no patterns that can be related to plate boundary zones, and no lithological assemblages like those of the plate tectonic world. The dome and keel patterns were developed by crustal inversion of a light, hot, TTG crust beneath a colder, heavier, volcanic crust above with TTG domes, containing radionuclides and gold, rising and ballooning and greenstones sinking by sagduction and drip. The greenstone
keels are characterised by steep to vertical prolate fabrics while the TTG fabrics change from almost isotropic in
dome centres through increasing flattening to plane strain at the margins. Strain patterns are, locally, more
complex and polyphase but do not imply development prior to or independent of diapiric ballooning (Snowden
and Bickle, 1976). The common TTG gneisses may have been formed by flow and flattening beneath the plutons.
Archaean TTGs are distinct from later TTGs in being silica-alumina-soda rich, potash-iron magnesium poor with
low heavy REE and fractionated rare earths.

782 Komatiites, are mainly Archaean; they cannot be oceanic crust because they occur with sediments, basalts and 783 silicic rocks in thick sequences on continental crust; they were likely developed by peridotite partial melting in 784 plumes at mantle potential temperatures at least 200°C higher than present. The sediments, usually water-lain are 785 commonly volcaniclastic but do not show the structural patterns of accretionary prisms. Claims have been made 786 that ultramafic-basalt associations are parts of ophiolite complexes and represent slices of obducted oceanic 787 lithosphere (such as at Isua or in the Kromberg Formation at Barberton; Grosch and Slama, 2017), but although 788 they contain pillow lavas and are therefore sub-aqueous, do not resemble, even remotely, the classic ophiolite 789 sequence of the Phanerozoic, lacking key components such as sheeted dyke complexes and residual tectonized 790 harzburgites. Archaean lithologies, structures, and patterns bear only a passing resemblance to younger ones but 791 never in the same arrangements, and relationships. The Archaean has lithologies, patterns and structures that are 792 unknown or rare in later terrains and contains very few that are characteristic of plate boundary zones.

793 We emphasise that one cannot use a simple shopping list of features and characteristics either all of or any one 794 which must be observed to establish a plate tectonic origin. Also, no broad rock type name such as andesite, 795 tonalite, or boninite can be used as definitive; these are merely names that conceal wide variations in petrology, 796 geochemistry and significance. It is necessary to specify the precise petrology and chemistry in rock suites that 797 show great variation among a variety of tectonic situations. Particularly egregious is the use of the general name 798 adakite as pejorative proof of a volcanic arc; this is tautology. Scraps of serpentinite, gabbro, and pillow basalt do 799 not necessarily constitute an ophiolite - lithologies must be arranged in sequence with sheeted dykes to have any 800 validity as oceanic crust and mantle generated at a ridge axis. We suggest that there are three types of evidence 801 used to argue for plate tectonics in the Precambrian, especially during the Archaean:

802 1. Based upon incorrect data.

- **803** 2. Based upon incorrect interpretations of correct data.
- **804** 3. Based upon correct data and interpretation.

805 It is the latter which best describes plate tectonics back to 2.0 Ga and, perhaps to 2.5 Ga. It is the pattern and 806 arrangement of rock types in linear and curvilinear belts that separate older platforms and cratons and are 807 indicative of extinct plate boundary zones, just as for recent and present ones. The difference is that whereas today 808 we can see oceanic lithosphere entering modern subduction zones, old subduction zones can be inferred only from 809 sutures, "adakite" belts, orogens and palaeomagnetism.

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- 1159 Figure 2: Distribution of major ore deposit types as a function of time and the supercontinent cycle (after Robb,
- 1160 2020); IOCG Iron Oxide-Copper-Gold/SEDEX Sedimentary Exhalative/VMS Volconogenic Massive
- 1161 Sulphide/MVT Mississippi Valley Type

