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The nature and origin of cratons constrained by their surface geology

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To the memory of Nicholas John (Nick) Archibald (1951–2014), master of cratonic geology.

ABSTRACT

Cratons, defined by their resistance to deformation, are guardians of crustal and lithospheric material over billion-year time scales. Archean and Proterozoic rocks can be found in many places on earth, but not all of them represent cratonic areas. Some of these old terrains, inappropriately termed "cratons" by some, have been parts of mobile belts and have experienced widespread deformations in response to mantle-plume-generated thermal weakening, uplift and consequent extension and/or various plate boundary deformations well into the Phanerozoic.

It is a common misconception that cratons consist only of metamorphosed crystalline rocks at their surface, as shown by the indiscriminate designation of them by many as "shields." Our compilation shows that this conviction is not completely true. Some recent models argue that craton formation results from crustal thickening caused by shortening and subsequent removal of the upper crust by erosion. This process would expose a high-grade metamorphic crust at the surface, but greenschist-grade metamorphic rocks and even unmetamorphosed supracrustal sedimentary rocks are widespread on some cratonic surfaces today, showing that craton formation does not require total removal of the upper crust. Instead, the granulitization of the roots of arcs may have been responsible for weighing down the collided and thickened pieces and keeping their top surfaces usually near sea level.

In this study, we review the nature and origin of cratons on four well-studied examples. The Superior Province (the Canadian Shield), the Barberton Mountain (Kaapvaal province, South Africa), and the Yilgarn province (Western Australia) show the diversity of rocks with different origin and metamorphic degree at their surface. These fairly extensive examples are chosen because they are typical. It would have been impractical to review the entire extant cratonic surfaces on earth today. We chose the inappropriately named North China "Craton" to discuss the requirements to be classified as a craton.

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INTRODUCTION

The purpose of this paper is to point out the kind of surface (or near surface) geological data that may be used to provide *binding constraints* for models proposed for the formation and persistence of cratons. As defined and elaborated on below, cratons are defined *only* by their resistance to deformation, and they may be of any age. Neither age, nor rock composition, nor morphology are parts of the definition of a craton, although cratons of different ages may have certain differences among them. They may preserve crustal and lithospheric material over billion-year time scales. Archean and Proterozoic rocks are present in many places on earth, but not all of them represent cratonic areas. Some of these old terrains, inappropriately termed "cratons," have been parts of mobile belts and have experienced extensive later deformation. What is and what is not a craton and how they form are still subjects of controversy largely because of the somewhat loose usage of the term and because their field geology has not everywhere been carefully assessed. The purpose of this paper is to review the concept of craton and critically discuss some recent ideas (e.g., Şengör, 1999; McKenzie and Priestley, 2008, 2016; Wang et al., 2014b; McKenzie and Tribaldos, 2018; Capitanio et al., 2020; Priestley et al., 2021) concerning their formation in the light of some relevant field geological observations.

DEFINITION, HISTORICAL BACKGROUND, AND ATTRIBUTES

Definition

It is a common misconception that cratons consist only of metamorphosed crystalline rocks at their surface, as shown by the indiscriminate designation of them by many as "shields," regardless of how extensive their non-metamorphic sedimentary cover may be (e.g., Gibbs and Barron, 1993; Hartmann and Delgado, 2001; Furnes et al., 2015). According to our present understanding, a craton consists of a piece of continental lithosphere isolated from the convecting mantle and resistant to alpinotype orogenies, i.e., those convergent plate boundary events creating intense penetrative deformation, metamorphism and being accompanied, to various degrees, by island arc-tholeiitic and calc-alkalic magmatism, within the Phanerozoic era. Cratons were also produced during the Archean and the earlier Proterozoic, but all of those are now incorporated into larger cratonic entities that formed in the late Proterozoic to earliest Cambrian interval, i.e., no longer forming independent cratons by themselves (Hoffman, 1988). *What defines a craton is thus only its strength as revealed by the absence of deformation within it while intense deformation may be going on along its margins*; neither its age, nor its composition are parts of the definition of a craton. As we explain below, areas that have commenced, yet not completed, their cratonization ("quasicratons") for various reasons (e.g., interference from mantle plumes, as in the Saharides: Şengör et al., 2020 and 2021; repeated deformation under the influence of the events around their periphery before cratonization was completed, as in Central and Western Europe: Küpper, 1965; Şengör, 1995; Kley, 2013; or in the U.S. Rocky Mountains: Burchfiel and Davis, 1975; Snyder et al., 1976 and Coney and Reynolds, 1977; also see Şengör et al., 2018a; or in Central and Eastern Asia: Şengör

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and Natal'in, 1996; Calais et al., 2006; Tunini et al., 2017; also see below etc.) usually resist alpinotype orogeny, but may be deformed by germanotype orogeny[1], i.e., non-penetrative block uplifts and subsidences of the type of the U.S. Rocky Mountains or the Boothia Uplift in Canada or the Harz Mountains in Germany or the Cainozoic Tien Shan in Central Asia. Such areas may be cut by major keirogens, i.e., belts of strike-slip deformation, such as the Central African Shear Zone or the Transbrasiliano lineament, but only if the craton is of late Proterozoic age. A fully developed craton cannot thus be penetrated by mechanical means; it may be completely destroyed only by heating from below by mantle plumes to remove its protective lithospheric armor (Şengör, 2001). Although some popular earth science dictionary definitions restrict the craton to the continental crust (Neuendorf et al., 2005; Allaby, 2008), today it is well known that formation and preservation of a craton is intimately linked to the composition, thermal state, and state of stress of its underlying lithosphere, although these factors are *not* a part of the *definition* of a craton, but its currently accepted *explanation*, as emphasized above (Şengör, 1999; Aulbach, 2012). Thus, in forming a craton, the histories of their crust and the underlying subcontinental lithospheric mantle are generally coupled in space and time.

[1]Orogeny is the term that covers the entirety of processes related to subduction and collision of buoyant lithospheric rafts including continents (Şengör, 1990, 2020). However, under orogeny, commonly the processes that create the great orogenic belts of our world are understood generally involving penetrative deformation, metamorphism and widespread, generally calc-alkalic magmatism. Understood as such, this term alone would leave out such lithospheric and crustal structures as the U.S. Rocky Mountains, the Cainozoic Tien Shan, the Cretaceous Harz Mountains, or the Jurassic Donbass structures. For that reason, it seems helpful to use the terms alpinotype for the orogenic structural ensembles that create generally penetrative deformation, widespread regional metamorphism and calc-alkalic magmatism and germanotype for those orogenic structures that create generally non-penetrative block uplifts and basins with little to no regional metamorphism and generally alkalic magmatism. These terms were defined by Hans Stille in 1920 to emphasize the difference between the structures in the Alps and the structures in Central Europe that formed in response to the same orogenic events (Stille, 1920). He later called the U.S. Rocky Mountains "germanotyp disloziertes Felsengbirge" (=germanotype-dislocated Rocky Mountains) as opposed to the "alpinotyp disloziertes Felsengebirge" (=alpinotype-dislocated Rocky Mountains, which are now generally referred to as the Sevier belt: Stille, 1940, fig. 59 on p. 242; notice here that Stille uses the designation "Rocky Mountains" for the entire North American Cordillera as recommended by Powell (1876, p. 4–5), but no longer followed). The distinction alpinotype versus germanotype has been found useful even in extraterrestrial geology (e.g., Byrne et al., 2014; also see Şengör, 2018).

A Terminological Problem

It should also be kept in mind that when a craton is identified it must be specified to what time period the definition applies. For example, the designation Slave Craton in Canada applies only to an interval between 2.69 and 1.88 Ma, because after that date it became only an integral part of a larger cratonic entity, the nuclear Canadian/Greenland, craton. That entity was then enlarged to include the Yavapai and Mazatzal orogens in the south and the Grenville orogen in the east, finally forming the entire Laurentian Craton (Suess, 1909, p. 284[2]), including Greenland, until the Paleocene-Eocene opening of the Davis Strait/Baffin Bay arm of the northern Atlantic Ocean: (e.g., Agranier et al., 2019). Similarly, the Dharwar Craton in India was an independent craton until it was framed by the Proterozoic orogenic belts that finally formed a part of the Gondwanian Craton by the Cambrian period. Speaking of a Slave Craton or a Dharwar Craton today would be like calling a twig a complete tree. It would be more informative and less confusing if such well-defined, smaller (and commonly older) areas in larger cartons areas are referred to as tectonic provinces—as is done by some geologists working in Canada (e.g., Heaman and Pearson, 2010)—within the present-day cratons or blocks with the necessary qualifiers, because they are not now independent cratonic structures, but parts of larger cratons. This terminological problem is not irrelevant to the discussions on the nature and origin of cratons, because Archean and Proterozoic parts of the present-day cratons have certain differences both in structure and evolutionary path. Bearing in mind the discussion presented in this section would also help to avoid such inapposite terms as the "Euxinic Craton" referring to the tiny Moesian Platform with a complex Phanerozoic deformational history (e.g., Balintoni and Ballica, 2016).

Brief History

The observation that large areas on continents have been free of any strong visible deformation since at least the beginning of the Cambrian seems to have been already made in the first half of the nineteenth century. It had been mostly attributed to the inability of the earth's internal heat to reach into the higher regions. First, Murchison in 1840 (Collie and Diemer, 2004) during his expedition in western Russia, then James Hall (Hall, 1842, 1858), after studying the geology of the Midwest of the USA, noticed considerable expanses of horizontal and undeformed Phanerozoic formations occupying very wide areas. Murchison thought, following a conversation with the German geologist Baron Leopold von Buch, that the absence of deformation was a consequence of the strength of the underlying "volcanic" rocks presumed to underlie those visible at the surface. It was Dana, who considered the interior of continents as "comparatively stable areas" (Dana, 1863, p. 734–735) in contrast to the oceans that subsided under the influence of the earth's contraction and compressed the continents between them. Within continents, Dana distinguished a stable center from less stable margins.

[2]Inspired by the "Laurentian surface" of Lawson (1890) and "Laurentian Peneplain" of Wilson (1903). For the usage of the adjective "Laurentian," see especially Wilson's footnote 2 on his p. 616. Suess took Wilson's "Laurentian" and used it to denote the entire North American craton.

At this early stage, before the term craton was coined, what was found surprising was the absence of surface deformation in large areas. That was the observation everyone agreed on, but its interpretations varied. While many, such as von Buch, Murchison, and Dana thought the absence of deformation was because the underlying rocks were strong and thus resistant to deformation, others, such as James Hall, thought, quite incomprehensibly, that the flat-lying sedimentary rocks were underlain by areas less resistant than mountainous areas (Hall, 1859, p. 86–87).

The more direct path leading to the invention of the term "craton" via the idea of "consolidation" supposedly creating strong areas, begins with Eduard Suess (1875). The first definition of consolidation ("Erstarrung" in Suess' publications, which his English translator Hertha Sollas rendered as rigefaction, i.e., stiffening: Suess, 1909, p. 625) was based on the character of the ground. The more contracted or stiffened parts of the earth's crust caused by cooling and shrinkage were thought to be more resilient against shortening deformation. Between them lie the weaker zones, which make up the geosynclines, i.e., the mother-throughs of future mountain chains, an idea borrowed from Élie de Beaumont (1828a, 1828b, 1852) and Dufrénoy and Élie de Beaumont (1848) via Dana (1847a, 1847b, 1863) and James Hall (1859). In 1883, Suess showed that on continents, mountain belts are found bordering vast areas of deformed and intruded Precambrian rocks covered by flat-lying Phanerozoic deposits that show no orogenic deformation since the Cambrian. He called such places "Tafel" (=Table), his definition of them was based strictly on the geology seen at the surface. This term became "plate-forme" in the French translation and then translated into English as "platform." Suess' original interpretation followed Dana's: areas of "tables" were regions where cooling had progressed the most (Suess, 1875).

Suess later changed his ideas of the nature of rigefaction and held responsible subsidence of the folded neighboring regions for the cessation of the orogenic stresses in the "table" areas. He emphasized that it was not demonstrated that the disappearance of deformation from such areas was a consequence of advanced cooling. Suess was thus inclined to ascribe the absence of deformation to absence of stresses, not to any increase in local strength of rocks. His student Leopold Kober (1921, p. 8), disagreed with this later interpretation by Suess and thought that the disappearance of deformation in certain areas was a result of the stiffening of the underlying rocks because of preceding orogenic deformation. He thus called such, what he considered *strong*, areas *Kratogen* (usually anglicized as cratogen) and divided the earth's lithosphere into three realms: cratogens, orogens and geosynclines. The term cratogen Kober derived from the two Ancient Greek words κράτος (kratos) meaning might, strength, also act of strength, and γένεσις (genesis) meaning origin, manner of birth, creation, so the term cratogen means "born strong."

The German geologist Hans Stille agreed with Kober and originally adopted his terminology, but in 1933 introduced the shorter version, *Kraton* (anglicized as craton), for what Kober had called *Kratogen* (Stille, 1936) and defined stiffened grounds as a consolidation state that no longer allows alpinotype deformation (i.e., penetrative and intense deformation, commonly accompanied by calc-alkalic magmatism and regional metamorphism) to occur but only germanotype deformation (i.e., non-penetrative deformation creating block uplifts or subsidences). According to Stille's tectonic theory, which is fully developed on the example of the structure of the Americas (Stille, 1940), the tectonic state of the ground is divided, like Kober's earlier classification, into three mobile areas occupied by orthogeosynclines, and they supposedly generated alpinotype orogens characterized by penetrative deformation. They form the large mountain belts of our planet. Partially or semi-consolidated areas ("quasicratons" in his terminology) can only localize germanotype orogeny involving non-penetrative block faulting, and fully consolidated areas allow no orogeny at all (Fig. 1).

Thus, the starting point of the creation of the concept of craton was the surface geology. In fact, the word craton is used today to designate large parts of the lithosphere that show no alpinotype orogenic deformation at least in the Phanerozoic. The terms "Tafel," "platform," and "craton" were used as their equivalents in the past, which caused a misunderstanding in the literature, such as East European Platform and East European Craton used as equivalents by different authors. It is important to use these terms correctly, because they denote distinct entities.

As used today, a *shield* (Suess, 1883) refers to large areas of ancient, partly crystalline but invariably strongly deformed rocks of Precambrian age that are exposed (because of subsequent erosion) in cratons, and a *platform* defines regions where the structural basement rocks are covered by horizontal or gently tilted sediments (Fig. 2). An old and rarely used Scandinavian term *glint* represents the boundary line between shields and platforms. Until Suess generalized its usage outside Scandinavia, the term glint (often in the form of *klint*) was confined to Scandinavian and circum-Baltic usage designating a mostly limestone questa resting on the Precambrian crystalline rocks of the shield. In this paper, we follow Suess' terminology while describing the cratonic areas. When we look at the historical development of the idea reviewed above, it is remarkable how they were formulated very early on in the history of geology, yet they could not always stick, because they commonly lacked a generally agreed upon theoretical foundation in which they could be adequately explained.

Over the past few decades, extensive research on cratons revealed common characteristics even though they experienced different overprinting by younger tectonic events. We here follow Stille's classification of cratons as follows:

- No craton: Any sort of deformation possible.
- Quasicraton: Alpinotype deformation no longer possible, but germanotype deformation still possible.
- Full craton: No copeogenic[3] deformation is possible, but falcogenic deformations are possible (e.g., the uparching of the Canadian and the Baltic shields after the end of glaciations).

But the full cratons also show two subtypes: the not so fully consolidated ones still show a weakened form of germanotype deformation (Marshak and Paulsen, 1996, 1997; Marshak et al., 1998), whereas full consolidation no longer allows any copeogenic deformation. Geophysical studies show that the lithospheric thicknesses not only of different cratons, but also various types of cratonization within a single cratonic area, vary (see Şengör, 1999). Archean lithosphere is usually 200–220 km thick, whereas the Proterozoic lithosphere can reach up to 300–350 km thickness (Jordan, 1975, 1988; Artemieva, 2006). Whereas the sub-cratonic keels are thinner in the southern hemisphere cratons (e.g., South Africa, Western Australia, South America, and India), thicker keels are found in the northern hemisphere (e.g., Baltic Shield, Siberian Platform, West Africa, and possibly the Canadian Shield) (Artemieva and Mooney, 2001, 2002; Artemieva, 2019) (Fig. 3). In cratons, Archean crust is generally thinner than the Proterozoic crust (27–40 km versus 40–55 km) (Baranov and Bobrov, 2018).

Cratons have a neutral or positive buoyancy that prevents them from sinking into the mantle. Although they are underlain by thick lithosphere, their top surfaces are usually near sea level. In the more stable areas, the long-term mean rate of denudation of the cratonic surfaces has been on the order of 1.5–4 m per million years assuming a static sea level; if the sea level is rising, the denudation takes a much longer time; if it is falling, a peneplain in the Davisian sense can never form on a craton (Pitman and Golovchenko, 1991; Flowers et al., 2006; Hall et al., 2021). The present-day elevations above sea level of the world's cratons do not exceed 1000 m, except in southern Africa (Zimbabwe, Kaapvaal and their continuation in Antarctica, Grunehogna), where the recent high topography has been attributed to the dynamic uplift of the proposed the Bouvet-Karoo mantle plume (Burke, 1996; Lithgow-Bertelloni and Silver, 1998; Burke and Gunnell, 2008; Jacobs, 2009; Artemieva and Vinnik, 2016) and in the Nilgiri Hills in southernmost India, where younger transpression has thickened the crust and raised the average elevation to above 2 km and crustal thickness to 52 km (Şengör and Natal'in, 1996; Das et al., 2015; Praseeda et al., 2015). The rest of the cratons reside between 100 and 600 m elevation where they commonly exhibit low grade to non-metamorphic rocks coeval with the time of cratonization. This indicates the cratons as a whole were never sites of high elevation and had little erosive removal after they stabilized.

RECIPES FOR A CRATON

In order to produce a craton, some conditions must be met. It requires: (*a*) depletion of volatiles from the sub-cratonic upper mantle, which helps the subcontinental mantle to be buoyant compared with the oceanic mantle (Pollack, 1986; Carlson et al., 2005; Capitanio et al., 2020; Priestley et al., 2021; but see Wang et al., 2014b); (*b*) geothermal gradient must be low (Ranalli and Murphy, 1987); (*c*) radioactive elements must be differentiated and depleted in

[3]*Copeogenic* deformations are those that cut the lithosphere; *falcogenic* deformations are those that only bend the lithosphere with accompanying minor faulting that may even give rise to some seismicity. These terms were introduced in Şengör (2003) and the reader should see there for an extensive discussion defending their introduction.

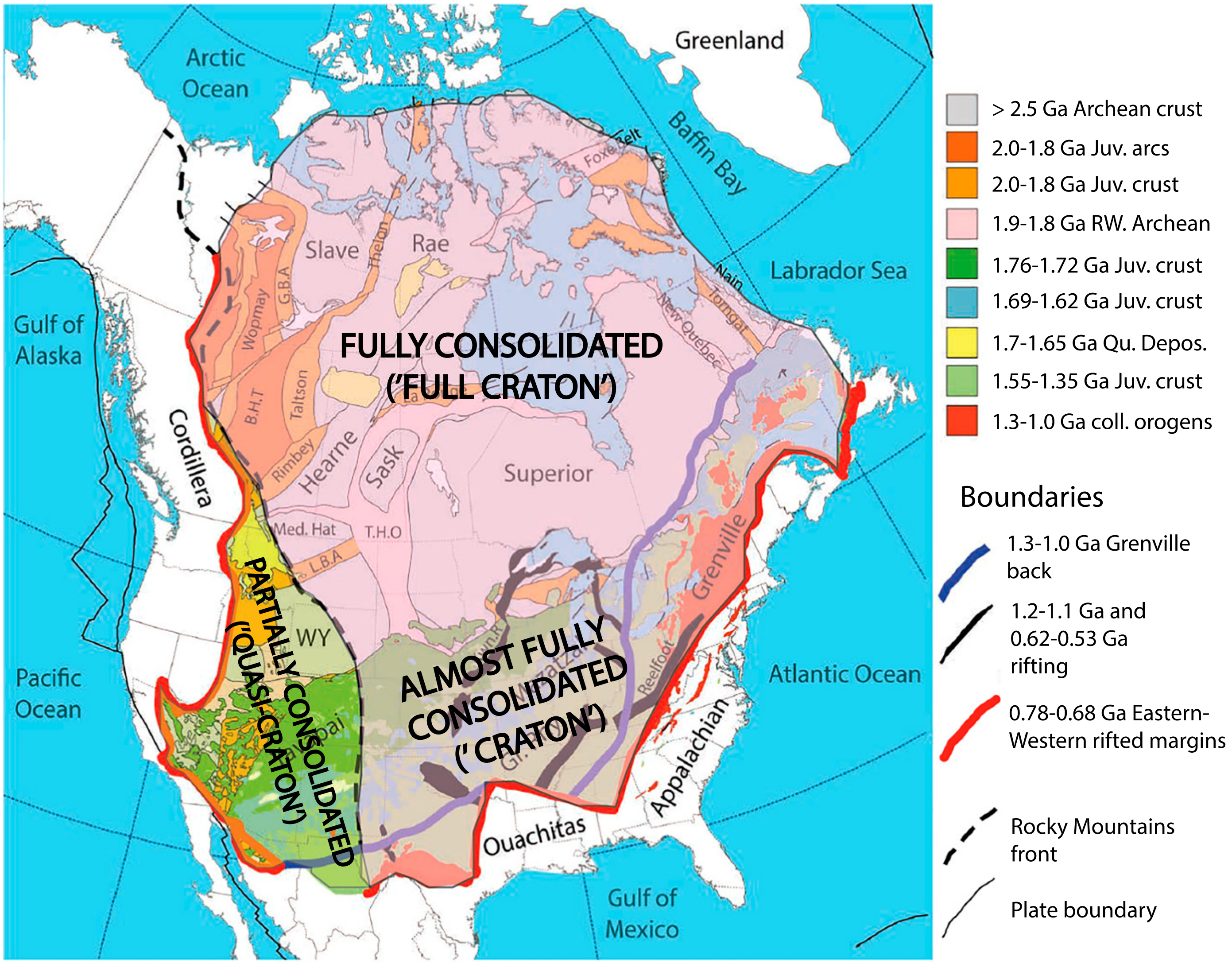

Figure 1. The North American Craton showing its constituent orogens and areas of different degrees of consolidation and their correspondence with the age of last orogeny (modified from Clouzet et al., 2018, figure 1). B.H.T.—Buffalo Head Terrane; G.B.A.—Great Bear Arc; Grt. Rhy—Granite-Rhyolite Province; Kwn.R—Kewanowan Rift; L.B.A—Little Belt Arc; Med. Hat—Medicine Hat Block; T.H.O.—Trans-Hudson Orogen; WY—Wyoming; RW—reworked; coll.—collisional; Juv.—juvenile; Qu. Depos.—Quartzite deposits.

order to decrease heat production at the base of the cratonic crust (Jaupart and Mareschal, 1999; Michaut et al., 2009), and (*d*) continental crust of intermediate or felsic composition must be generated in a substantial volume, probably through arc-like processes (Taylor and McLennan, 1985, 1995; Şengör et al., 1993; Şengör and Natal'in, 1996; James and Fouch, 2002).

According to Şengör (1999), a craton begins forming in a typical, evolving, subduction-accretion system with its underlying depleted mantle wedge (Fig. 4A). When continental collision occurs involving such a wedge (Fig. 4B), it thickens the lithosphere and the crust by shortening, but the predominantly mafic lower part pulls down and creates a lithospheric keel as recently shown by the preferred orientation of the fast axes of the olivine crystals in the subcontinental mantle (Priestley et al., 2021). Şengör (1999) pointed out that this does not give rise to much uplift of the surface of the crust, because of extensive granulitization of the lower crust increasing its density to ≥ 3 g/cm^3 (Fig. 4C). The formation of granulite-facies rocks increases the strength, and it becomes extremely difficult to destroy the cratons they underlie (Hawkesworth et al., 2010). This increase in weight prevents the rise of the surface elevations.

There is increasing evidence indicating that subduction-accretion processes were active during the earliest Archean, even perhaps in the latest Hadean (Harrison, 2009, 2020; Windley et al., 2021; Hyung and Jacobsen, 2020; Turner et al., 2020; Korenaga, 2021). Trace-element characteristics of Archean tonalite-trondhjemite-granodiorite gneisses resemble those predicted for Phanerozoic subduction-derived granitoids and thus are consistent with the growth of Archean continental crust at convergent plate margins (Arndt, 2013; Windley et al., 2021). In addition, the presence of ophiolites, but mainly ophirags, in many Eoarchean to Neoarchean Archean greenstone belts (Kusky, 1989; Armstrong et al., 1990; Polat et al., 2002, 2008; Hofmann and Kusky, 2004; Smithies et al., 2005; Mukhopadhyay et al., 2012; Turner et al., 2014; Furnes et al., 2015; Grosch and Slama, 2017) and the lithospheric structure derived from S-wave receiver functions suggest cratonic lithosphere was formed by arc accretion (Miller and Eaton, 2010). De Wit et al. (1992) favored the

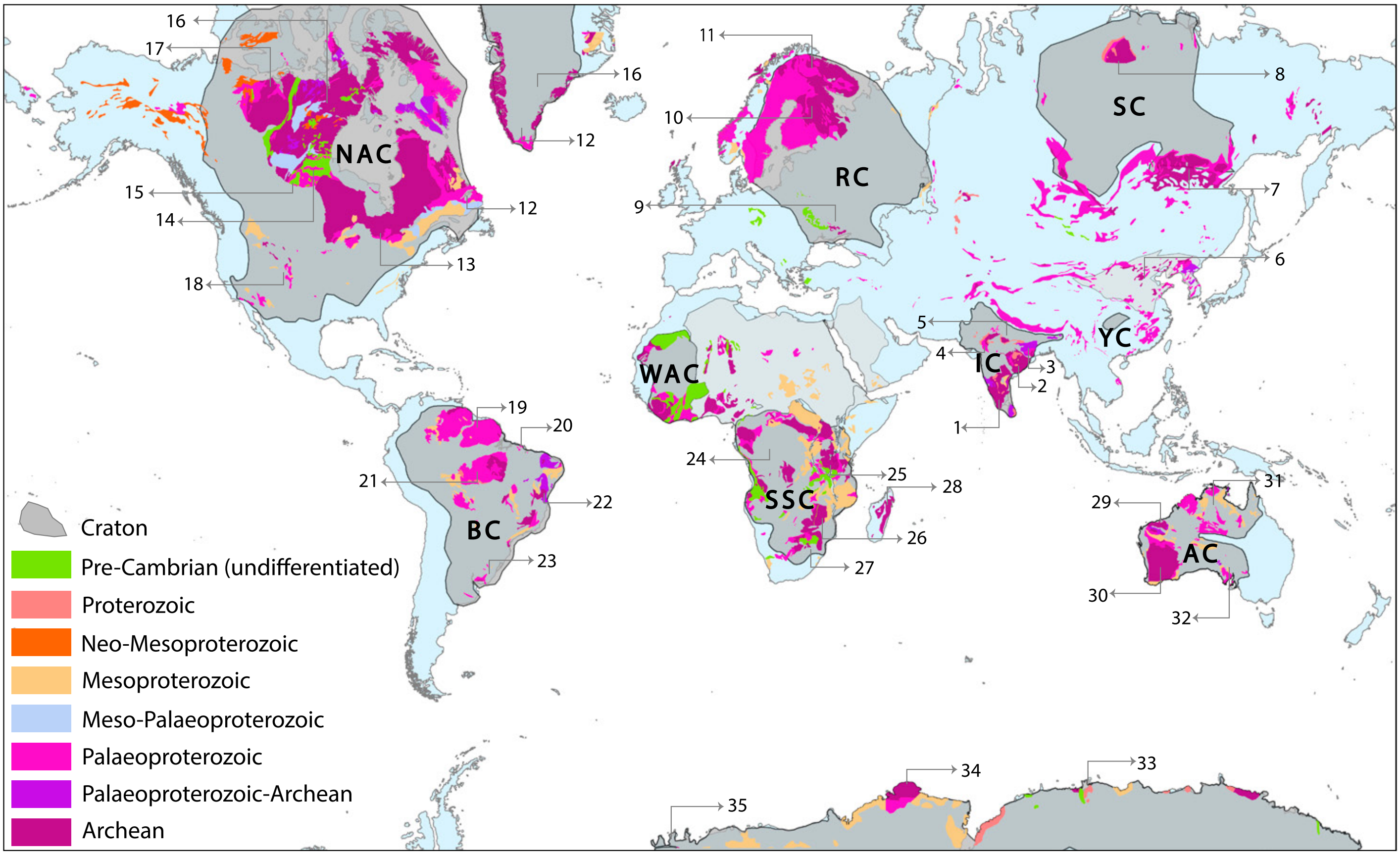

Figure 2. Cratons of the world. The dark gray color indicates the entire craton, i.e., the shield(s) plus the surrounding platforms. The light gray color indicates areas listed as cratons in the literature but do not meet the criteria. AU—Australian Craton; BC—Brazilian Craton; IC—Indian Craton; NAC—North American Craton; SC—Siberian Craton; SSC—Sub-Saharan Craton; Y—Yangtze Craton; WAC—West African Craton. The constituent older cratons commonly mentioned in the literature, that are now no longer independent cratons, but simply provinces within the larger cratonic entities: (1) Dharwar; (2) Bastar; (3) Singhbhum; (4) Aravalli; (5) Bundlkhand; (6) North China (this has never been a craton as we discuss herein); (7) Aldan; (8) Anabar; (9) Sarmatia; (10) Karelia; (11) Kola; (12) Nain; (13) Superior Province; (14) Hearne; (15) Sask; (16) Rae; (17) Slave; (18) Wyoming; (19) Guyana; (20) São Luis; (21) Amazonia; (22) São Francisco; (23) Rio de la Plata; (24) Congo; (25) Tanzania; (26) Zimbabwe; (27) Kaapval; (26 and 27) Kalahari; (28) Madagascar; (29) Pilbara; (30) Yilgarn; (31) North Australia; (32) Gawler; (33) Vestfold; (34) Napier; (35) Grunehogna.

subduction process over plume origin based on the Archean age metamorphosed peridotite and basalt (eclogite) xenoliths from Kaapvaal province (Shirey et al., 2004). Nothing else but subduction seems capable not only in placing oceanic crust on to the overriding plate or carrying material to great depths, but also in the ability to bring them back to the surface in an orderly fashion, which no diapiric process, as defended by Hamilton (2007, 2019)[4], for example, would be capable of doing. Hydrated mineral inclusions in the Hadean Jack Hills zircons, Western Australia, indicate their host protoliths were exposed at the earth's surface in the presence of water (shales, probably implying subduction-accretion material around sialic "rockbergs" perhaps not too dissimilar to "subduction zones" seen in present-day lava lakes: see Harrison, 2009) and then transported to a depth to form parenting orthogneiss (Mojzsis et al., 2001; Harrison 2009). A recent review of the geochemistry of zircons from Jack Hills reveals important petrogenetic information that implies the initiation of subduction zone processes as early as 3.6 Ga (Ackerson et al., 2021).

Kusky (1993) suggested that Archean lithospheric mantle formed beneath arcs by accretion of oceanic lithosphere and trapped subarc mantle wedge. Using major element and Re-Os isotope data from cratonic mantle xenoliths, Servali and Korenaga (2018) proposed that Archean cratonic mantle was derived from highly depleted oceanic lithosphere that originated beneath mid-ocean ridges or similar spreading settings. Higher ambient mantle temperatures in the Archean likely caused higher degrees of partial melting in the oceans (but not in the continents: Burke and Kidd, 1978), resulting in highly depleted oceanic residual lithospheric mantle that had olivines with high forsterite (92–94) contents (Herzberg, 2018). If Servali and Korenaga's geochemical interpretation is correct, then the accretion (underthrusting) of such depleted oceanic lithosphere beneath Archean oceanic island arcs could have formed the nuclei of the cratonic mantle, and the accretion of these island arcs eventually would have led to the formation of first larger continents and cratons around 3.2 Ga (Windley et al., 2021). The light rare earth element, large ion lithophile element, and Pb-enriched, but Nb-Ta–depleted trace-element systematics

[4]Warren Hamilton, one of the giants of geology in our times, passed away on 26 October 2018, leaving behind an almost complete manuscript of this paper, lacking only the abstract and the conclusions, which Gillian Foulger completed and prepared for publication, earning the eternal gratitude of the geological community. We disagree with Hamilton's view of the Precambrian tectonics, but his immense knowledge and keen insights have helped us in sharpening our ideas.

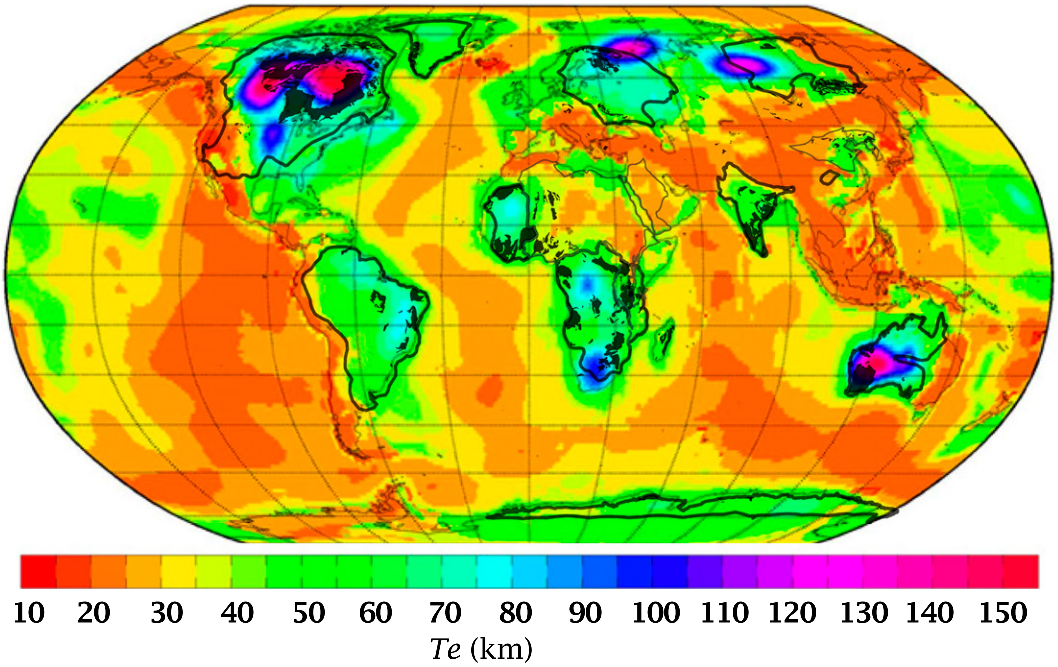

Figure 3. Distribution of the cratons and their Archean rocks (shown in black) compared with effective elastic thickness (Te) of the lithosphere (km) based on the yield strength envelope (Tesauro et al., 2012). Thick black line indicates borders of cratons as shown in Figure 2. Thin dark gray line indicates regions that are previously thought to be cratons.

of cratonic mantle suggest that accreted oceanic lithosphere beneath Archean island arcs was later overprinted by subduction-derived melts and/or fluids (Polat et al., 2018).

According to McKenzie and Priestley (2008) two possible ways of lithosphere thickening can lead to craton formation. Figure 5A demonstrates homogeneous crustal thickening with constant heat flow at the base of the lithosphere, Figure 5B represents shortening by pure shear, a uniform crustal thickening resulting from bulk shortening, consistent with the model by Şengör (1999; see Fig. 4 herein) and the observations in Priestley et al. (2021). In order to prevent delamination, the mantle lithosphere must be depleted during a previous melting. Figure 5C shows crustal thickening by emplacing a thick crustal layer on top of a craton as a result of thrusting or lower crustal flow. All these model cases are followed by a thermal relaxation phase (Le Pichon et al., 1997). McKenzie and Priestley's (2008) calculation shows that the temperature of the middle crust can rise over 900 °C in 60 Ma because of the crustal radioactivity unless extension occurs. This process ends up creating high pressure and high temperature rocks as already envisioned by Argand in 1922, when he postulated, within the framework of Wegener's theory of continental drift, a large-scale underthrusting of India under Asia (Argand, 1924; see Fig. 6). Note however, that Argand's Tibetan underthrusting does not account for the entire crustal thickening under Tibet; under the Kuen-Lun, he also postulated shortening and thickening of Asian crust.

It has been proposed recently that Tibet is an actualistic analogue of how cratons begin forming (McKenzie and Priestley, 2008, 2016; McKenzie and Tribaldos, 2018). The presence of a thick lithosphere beneath Tibet is inferred using surface wave tomography employing Rayleigh waves, and it indeed resembles that beneath cratons. Although attractive, this model implies that a substantial amount of the thickened crust, nearly 35–40 km, must somehow be removed to bring its surface to near sea level, where most major Archean cratons reside in the present day. Wholesale underthrusting of the Indian plate beneath the entire Tibetan plateau (e.g., Argand, 1924; Powell and Conaghan, 1973; Zuza et al., 2019), would create a persistent buoyancy that would cause erosion, presumably down to sea level, of the crust and when it stabilizes it would reveal only high-grade rocks, namely amphibolite to granulite grade. Based on the possible density distribution of the Indian plate beneath the Tibetan plateau (Hetényi et al., 2007), we have generated a scenario to estimate the metamorphic facies related to a given depth. Figures 7A and 7B show the distribution of metamorphic facies with depth and erosional (tectonic or attritional) surfaces with corresponding rock assemblages that will be exposed. A similar thing would happen if the shortening in Tibet is accommodated purely by bulk shortening of the plateau lithosphere as envisaged by Dewey and Burke (1973), which is also compatible with the observations by Priestley et al. (2021). We now know that a combination of these two processes seems to have happened, the Indian lithosphere having been claimed to reach as far north as the Banggong Co-Nu Jiang suture, i.e., to the boundary between the Lhasa and the East and West Qiang Tang arc fragments (Zuza et al., 2019). However, the growth of Tibetan topography from its center outwards (Wang et al.,2014a) makes a wholesale Indian underthrusting difficult as does the old observation of the antecedent rivers in the Himalaya (e.g., Medlicott and Blanford, 1879, p. 676–677; Valdiya, 1996), favoring more a manner of bulk crustal thickening as suggested by Dewey and Burke (1973). This is also supported by both widespread folding as clearly visible on Google Earth images of, and some thrusting on, the Lhasa Block (e.g., Şengör, 1981) contrary to some recent claims. The recent observation by Wang et al. (2021) that the entire 80 km thickness of the Tibetan crust in southern Tibet under the Lhasa block has a felsic composition indicates a crustal structure not unlike that present now under the Altiplano in the central Andes, where the entire 65-km-thick crust seems to consist of quartz-rich, felsic rocks (Zandt et al., 1996). In Tibet, a thick magmatic arc crust of the Lhasa Block that includes the Transhimalaya may be underlain by thinned continental margin rocks of India, where shortening results from the convergence that thickens both the crust and the underlying mantle parts of the lithosphere both by thrusting and bulk shortening. However, this mechanism can only operate in length scales less than 1000 km and cannot explain vast cratonic areas generated at the same time interval unless the unlikely scenario of having more than three contiguous Tibet-type plateaux

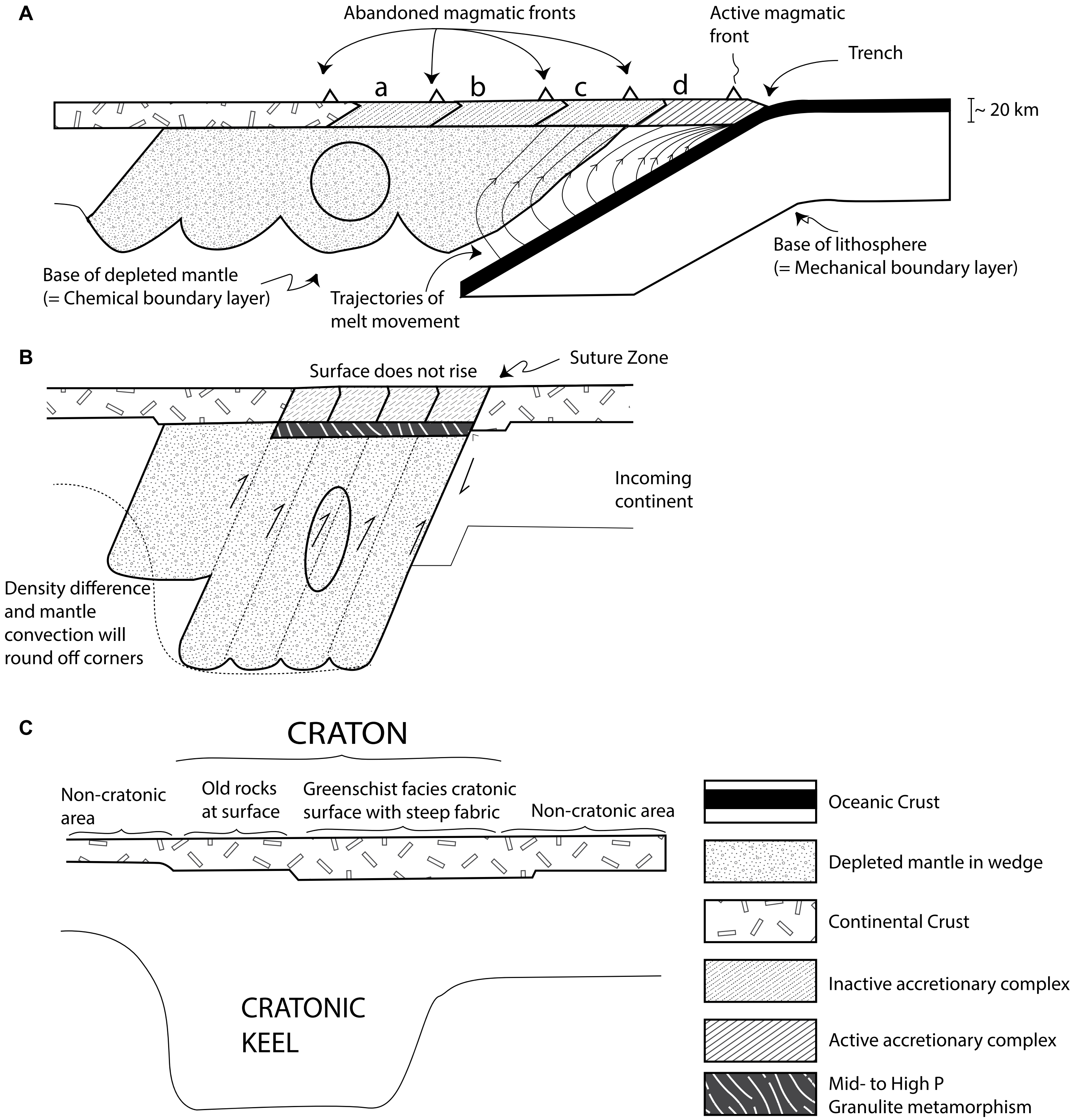

Figure 4. A cross-sectional model showing how to build a craton by subduction-accretion (Şengör, 1999). A case of homogeneous crustal thickening with constant heat flow at the base of the lithosphere. We added the strain ellipses to indicate the parallelism of Şengör's (1999) interpretation with that of Priestley et al. (2021, fig. 3). In A, the letters a, b, c, and d signify progressively younger subduction accretion prisms together with their now abandoned arcs added to the continent, thereby enlarging it.

that are assumed to form simultaneously with a unified 3000 km underthrusting! By contrast, the largest tectonic detachment underlying an Archean upper crustal assemblage we know of is that imaged in the Yilgarn Province in Western Australia (Ahmat et al., 1993), the width of which is barely 150 km, and in it such shear zones as that of Bardoc penetrate the sub-detachment basement for more than 20 km across strike (Ahmat et al., 1993). We show below, using surface geological observations, that some Archean cratons had little erosive removal after

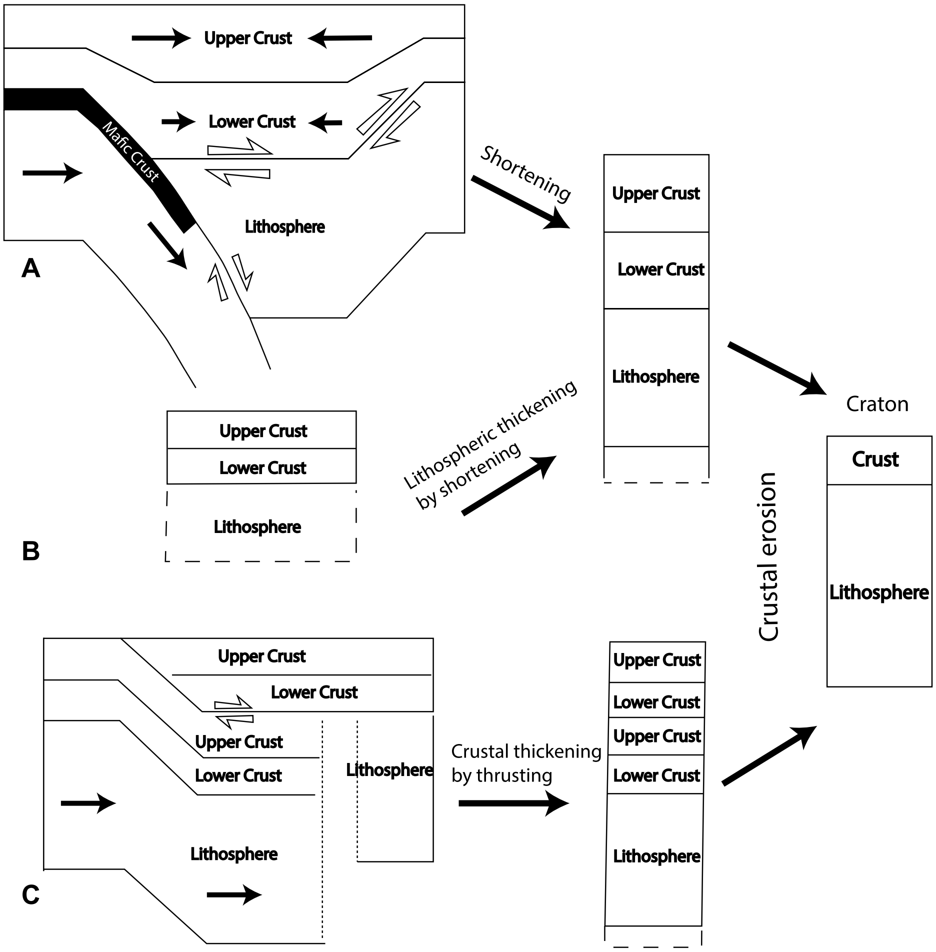

Figure 5. Schematic drawings of 2-D scenarios corresponding to (A and B) the uniform thickening of crust and lithosphere and, (C) crustal stacking model based on Le Pichon et al. (1997) and McKenzie and Priestley (2008).

they stabilized, and none ever was a Tibet-type high plateau as a whole.

MYTHS AND FACTS

Many consider that most cratons consist only of polydeformed and metamorphosed crystalline rocks on their surface, as shown by the commonly indiscriminate usage of the term "shield" as mentioned above (e.g., Gibbs and Barron, 1993; Hartmann and Delgado, 2001; Furnes et al., 2015). In North America, in the Superior Province and the Slave Province, in India, in the Dharwar, in the Baltic, in Karelia, in the Amazonian, Kaapvaal and Zimbabwe provinces in Africa, and the Pilbara and Yilgarn provinces in Australia, not only high-temperature–high pressure facies metamorphic rocks of Archean age, but also coeval low-grade or even completely unmetamorphosed and only gently deformed sedimentary rocks crop out at their surface (Table 1). When viewed from this angle, none of the cratons listed above show a uniform Tibet-like structure in their make-up.

The Superior Province, the Canadian Shield

The Superior Province represents the largest Archean exposure on earth, which is surrounded by Proterozoic orogenic belts together forming the Laurentian or the North American Craton (Hoffman, 1988). It consists of a large variety of plutonic, volcano-plutonic, and sedimentary rocks plus their metamorphic equivalents from subgreenschist to granulite facies (Card, 1990; Percival et al., 2012). These rocks are disposed in east-west–trending subprovinces ranging from 3.8 to 2.67 Ga (Polat and Kerrich, 2001; Percival et al., 2012). It is widely accepted that the dominant structural trend was created by subduction-accretion processes during the late Archean (Poulsen et al., 1992; Wyman et al., 2002; Percival et al., 2012) and cratonization took place between 2680 and 2600 Ma (Percival et al., 2012). Among these subprovinces, Wawa-Abitibi is the largest domain of juvenile rocks in the Superior Province and shows the lowest degree of metamorphism nowhere exceeding 4 kb (corresponding to ~12 km depth). The majority of the Superior Province is characterized by greenschist facies rocks. Figures 8 and 9 show examples of sedimentary and volcanic rocks, respectively, metamorphosed in greenschist facies from six different widely separated localities in the Superior Province. Notice that the rocks shown are deformed by medium to zero magnitude strain within the frame of our pictures. None could have possibly come from a depth of 30–40 km as required by the craton formation model of McKenzie and Priestley (2008, 2016) and McKenzie and Tribaldos, (2018).

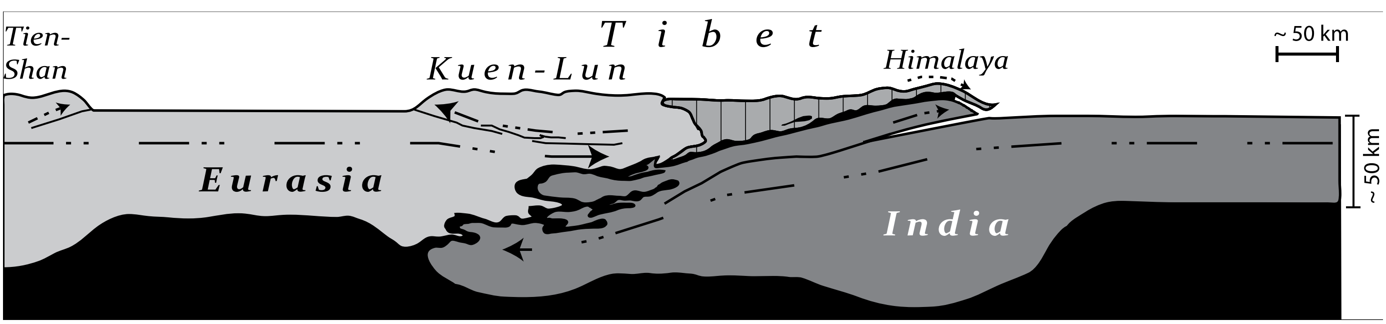

Figure 6. Cross-section of Tibet (Argand, 1924).

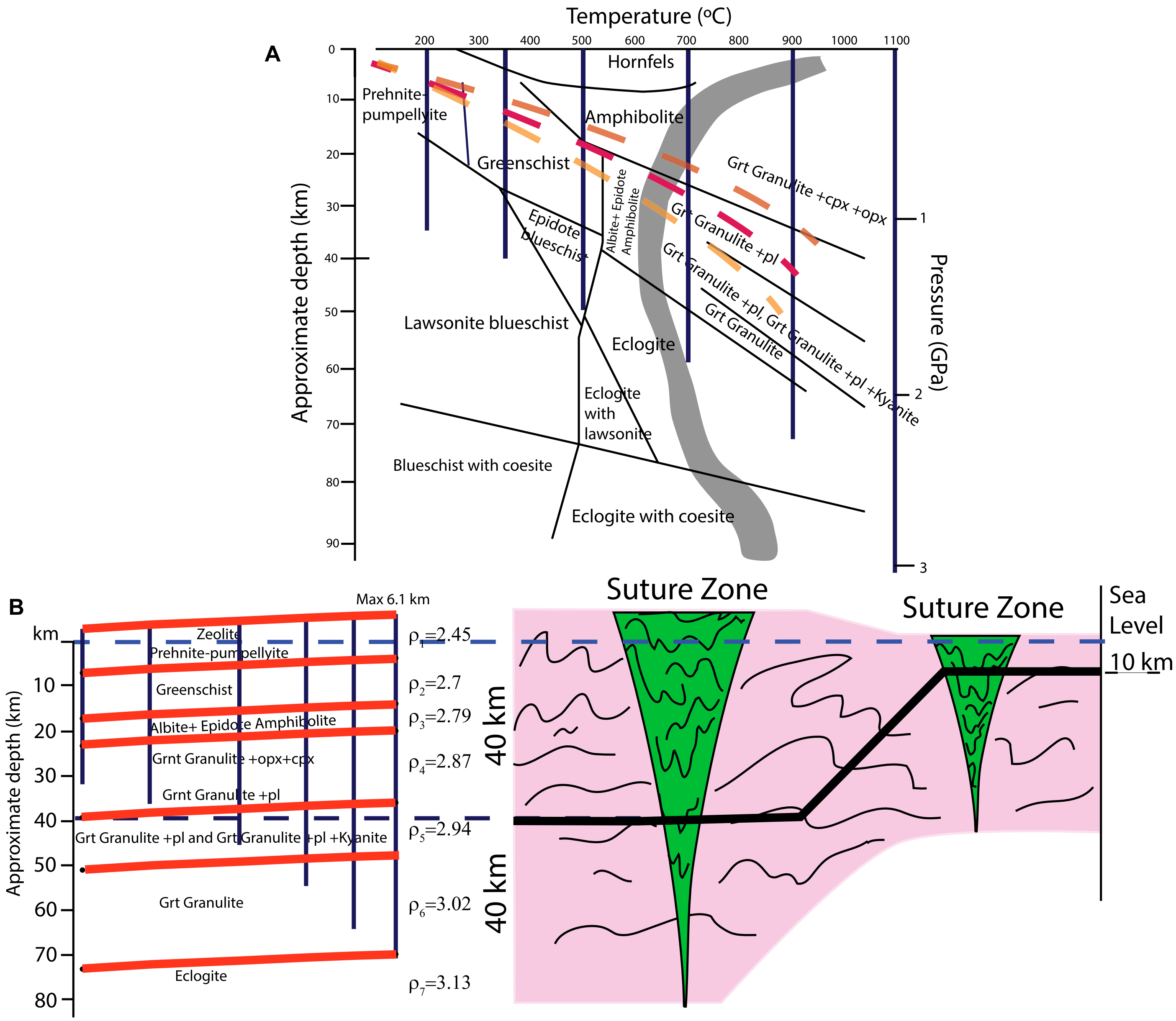

Figure 7. (A) Limits of the metamorphic facies as a function of temperature and pressure. Modified from Bousquet et al. (1997). Orange dashed lines indicate the range of adiabatic thermal gradient, gray band delimits the wet melting zone. Red dashed line indicates the average adiabatic thermal gradient that have been used in our calculations. (B) Metamorphic facies based on the average adiabatic thermal gradient shown with red dashed line in A. In the chart, horizontal red solid lines indicate facies boundaries, vertical black lines indicate adopted crustal thickness and related surface uplift. Blue dashed line corresponds to the erosion surface.

The Barberton Mountain, the Kaapvaal Province

The Kaapvaal province of South Africa is one of the oldest Archean segments in the cratons on the earth. The cratonization started around 3.1 Ga when major granitoid batholiths intruded and essentially has been completed by ca. 2.6 Ga (Nguuri et al., 2001). The Archean rocks of Kaapvaal were subsequently affected by magmatic and accretionary events. It is surrounded by Proterozoic mobile belts: on the south and east by the Namaqua-Natal belt (1.1–1.9 Ga), on the north by the Limpopo belt (2.3 Ga), and on the west by the Kheis overthrust belt (ca. 2.0 Ga). The north-central Kaapvaal craton was disrupted at ca. 2.05 Ga when the largest known layered mafic intrusion in the world, the Bushveld Complex, emplaced into a stable cratonic setting.

The Barberton Greenstone Belt (3.2–3.53 Ga) of eastern Kaapvaal consists of well-preserved Paleoarchean volcanic and sedimentary rocks generally at low metamorphic grade and in some places entirely devoid of penetrative deformation (Grosch et al., 2011; de Wit et al., 2011). One of the best examples is the Onverwacht Group which includes the Komati Formation in the lower levels and the Hooggenoeg, Kromberg and Mendon formations in the upper levels (de Wit et al., 2011; Grosch and Slama, 2017). Within the Onverwacht Group the Hooggenoeg Formation comprises a 2.5–3.0-km-thick section of predominantly tholeiitic and komatiitic basalts

TABLE 1. PROVINCES IN CRATONS SHOWING LOW-GRADE METAMORPHISM

Province	Age	Unit name	Lithology	Metamorphic grade	Source
Kaapvaal	3.07 Ga*	Dominion Group	Sandstone along with minor conglomerate and argillaceous horizon followed by thick succession of mafic-intermediate lavas interlayered with felsic lavas and pyroclastic rocks	Lower greenschist facies and minor deformation	Burke et al. (1986); Agangi et al. (2020)
Zimbabwe	2.7 Ga*	Upper Greenstone	Conglomerate, sandstone, siltstone, limestone, cherty limestone, stromatolitic limestone, and minor banded iron formation	Low-grade (sericite and chlorite)	Kusky and Kidd (1992)
Dharwar	2.7 Ga*	The Talya Conglomerate	Conglomerates interbedded with mudstone and sandstone	Greenschist facies	Ojakangas et al. (2014)
Superior Province	2.7 Ga*	Wawa-Abititi	Calc-alkalic to alkalic metavolcanic and metasedimentary rocks	Pumpellyite-amphibolite facies	Percival et al. (2012); Lodge et al. (2013)
Greenland	3.8 Ga†; 2.7 Ga§	Isua supracrustal belt	Metabasites	Greenschist to amphibolite facies	Hayashi et al. (2000)
Yilgarn	2.7–2.6 Ga*	Diemals Formation	Siliciclastic sedimentary rocks	Greenschist-facies	Morris et al. (2007)
Pilbara	3426–3350 Ma*	Strelley Pool Formation	Laminated carbonate rock, stromatolitic carbonate, evaporites, sandstone, conglomerate, volcanic ash, volcaniclastic sedimentary rocks	Lower-greenschist	Hickman (2008)
Brazilian Shield	2758 ± 39 Ma†	Grão Pará Group	Basaltic to rhyolitic metavolcanic units at the base, overlain by banded iron formation, and topped by an alluvial sedimentary sequence	Low to medium grade	Olszewski et al. (1989)
Karelia	2853 ± 11 Ma†	South Vygozersky Greenstone	Plagiogranites	Greenschist facies to epidoteamphibolite	Myskova et al. (2015)

*Age of deposition.
†Age of crystallization.
§Age of metamorphism.

at the base whereas the upper part is dominated by felsic volcaniclastic conglomerates, sandstones, and tuffs. Figure 10 shows various outcrop photographs all taken along its course, and the little to no strain they exhibit is shown particularly well by the well-preserved immiscibilty spherules seen in the pillow lavas (Fig. 10A) and the preserved sedimentary structures (Figs. 10B, 10C). Figure 10D shows pillow lavas from the Kromberg Formation which is described as a thick, continuous sequence of komatiite and basaltic volcanic rocks capped by carbonated shallow-water volcaniclastic sediments and cherts. The pillow lavas rest conformably on carbonated tuff. Along the Komati River, both the volcanic rocks and the sedimentary rocks show no metamorphism and very gentle deformation. The low-grade to unmetamorphosed rocks exposed at the surface are inconsistent with Barberton Mountain's having had a full crustal thickness on top of it as required by the model of craton formation by McKenzie and Priestley (2008, 2016) and McKenzie and Tribaldos (2018), and the common misconception that cratons consist typically of metamorphic rocks at their surface.

The Yilgarn Province, Western Australia

The Yilgarn province (Fig. 11) consists mainly of plutonic, volcanic, and sedimentary Archean rocks that formed between 3.73 and 2.60 Ga ago and, with the exception of southwestern Yilgarn high-grade province, was metamorphosed at low grade and in places not at all (Myers, 1993). In the Kalgoorlie region, around the western shores of the great salt pan of Lake Lefroy, there is a highly folded turbidite section sitting on a sequence of tholeiite basalts and komatiites with interlayered albite cherts (Kalgoorlie Group: Swager et al., 1992; Nick Archibald, personal commun. *in* Şengör and Natal'in, 1996). Above these rocks is another turbidite section, considered equivalent to what is called the Black Flags Group (Swager et al., 1992). Turbidites, together with their underlying komatiites are sliced by sled-runner–type thrust faults, which themselves were folded and steepened as a result of an east-west phase of shortening.

Although much of Yilgarn shows the structural and metamorphic pattern described above, the southwestern part of the craton, where extensive granulite facies metamorphism with migmatites and charnokites formed some 2.64 and 2.62 Ga ago, presents a much deeper level of erosion. It is this area that has very old detrital zircons predating any granitic magmatism in this part of the craton. The southwestern high-grade part of the Yilgarn seems to have been attached to the rest of the low-grade craton by 2.6 Ga ago (Cassidy et al., 2006). Within this area Smithies et al. (2018) noted that geochemical signatures indicate northeast-southwest–trending belts that are truncated by the province boundaries and do not correlate with the lower crustal density anomaly. Here geographical variations in proxies for source melting pressure define a broad northeasterly trend that truncates the north-northwesterly trending eastern boundary of the Southwestern Yilgarn Province (Smithies et al., 2018). However, the same northeasterly trends seem to be present also in the rest of the craton, albeit in a much less pronounced manner, although even this, according to Smithies et al. (2018) needs corroboration. If true, however, Smithies et al. (2018) believes that it would invalidate the west-to-east accretion of the various "terranes" in the Yilgarn. However, this suggestion ignores the very great amounts of strike-slip faulting in the Yilgarn as a whole during the Archean.

We have called the southwestern high-grade region the "McKenzie and Priestley-Land" because it does expose mid to lower crustal rocks as required by their model, in contrast to the "Şengör et alii-Land," where the grade of metamorphism reaches amphibolite grade in only very restricted areas, resembling very much the Altaids (Şengör et al., 2018b) and the Saharides in the Arabian Peninsula and Egypt (Şengör et al., 2020) having undergone all the vicissitudes described in detail by Goscombe et al. (2019) plus much strike-slip faulting in both senses. Yet the entire thing is now one single province. The schematic cross section in our Figure 11 illustrates our suggestion that the McKenzie and Priestley-Land may have been thrust over the Şengör et alii-Land in part and thus became uplifted just like Tibet. How it became denuded is another problem. Whether and how much that thrusting may have also had a strike-slip component is as yet unknown.

The North China Block

The purpose of this section is to provide an example of the inappropriate usage of the term craton for any old terrain exposing Archean and Proterozoic rocks without considering the entire geological history of the region under consideration. Hence the somewhat detailed discussion.

Main Geological Characteristics

What is called the "North China Craton" is one of the most intensively studied Precambrian terrains in the world. Its tectonic history, however, is one of the most controversial subjects in

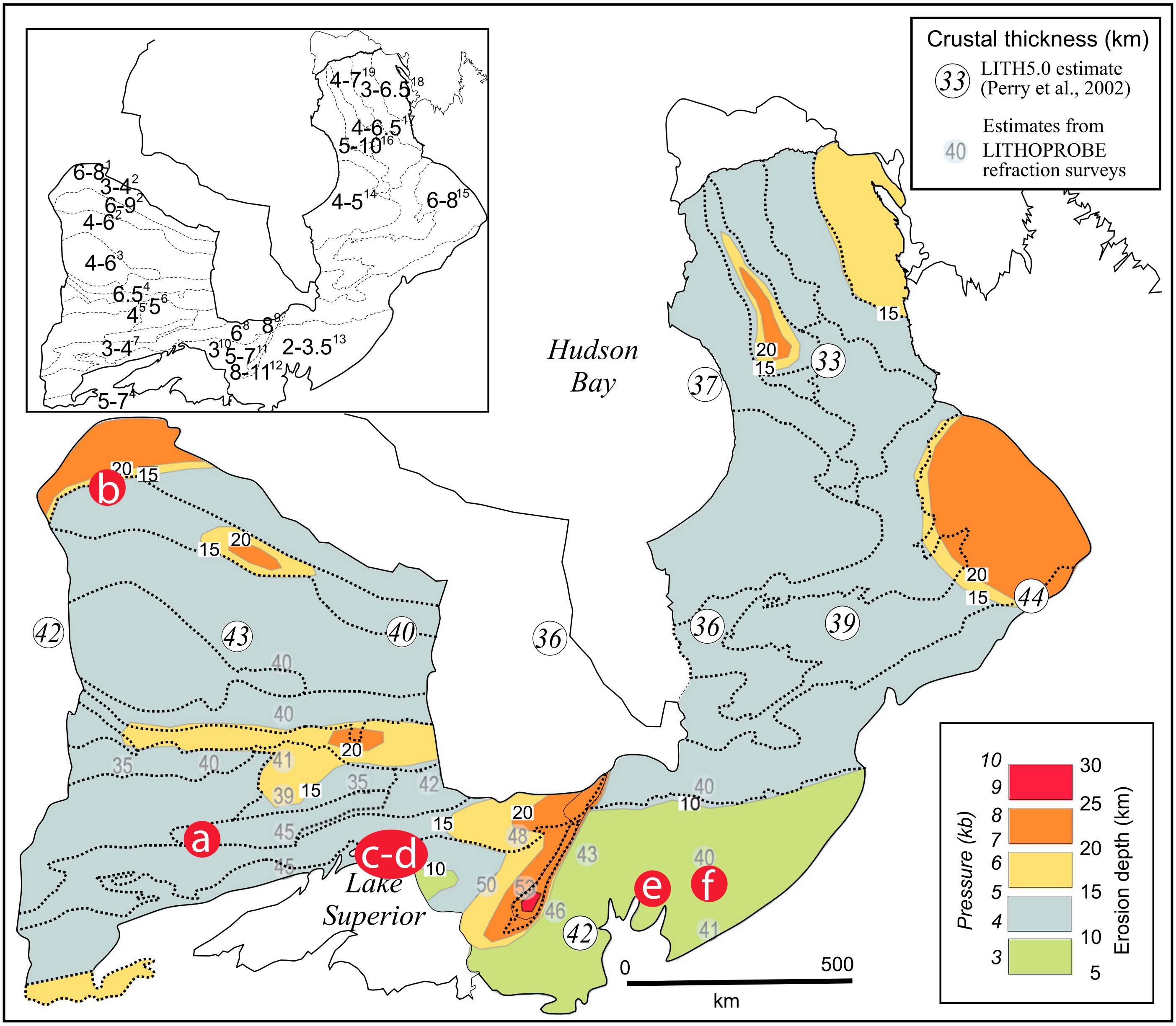

Figure 8. Contour map (colors) showing general level of erosion across the Superior Province based on igneous and metamorphic pressure estimates. Circled numbers show crustal thickness in kilometers (Percival et al., 2012). Letters represent the localities of photos presented in Figure 9.

Precambrian geology (Kusky et al., 2007a, 2016, 2020; Zhao et al., 2005, 2012; Zhai and Santosh, 2011; Zhai, 2011, 2014; Jiang et al., 2018; Wang et al., 2018; Zhao et al., 2021) in part because of the misleading implications of the term craton as applied to it. We here call it simply the North China Block for the purposes of the following discussion, following the usage of a number of authors before us. We note that the term "North China Block" was also used by Kusky et al. (2007a) on the basis of the assumption that the eastern part of the block lost its cratonic lithosphere in the Mesozoic.

The North China Block is bordered to the north by the Manchurides (Şengör and Natal'in, 1996; Xiao et al., 2003), to the south-southwest by the Qinling-Dabie orogenic belt, to the west by the Qilian part of the Kuen-Lun (Şengör and Okuroğulları, 1991), and to the east by the Su-Lu Belt (Fig. 12), although the Tan-Lu fault (Lu et al., 1983; Wang and Dou, 1997, Huang et al., 2015) and the associated Cainozoic Bo-Hai Basin (Allen et al., 1998; Yang and Xu, 2004) have very largely disrupted its older structure since the Mesozoic. Many models have been proposed to explain the tectonic evolution of what has long been called the North China Craton and its relationship with the neighboring Proterozoic and Phanerozoic orogenic belts (Kusky and Li, 2003; Zhai et al., 2010; Zhao and Zhai, 2013; Santosh et al., 2016; Kusky et al., 2016, 2020). Our reason for calling it North China Block is because we feel that it does not fulfill the criteria for being a craton for the reasons explained below.

Archean rocks in the North China Block range in age from 3.8 to 2.5 Ga, but most of them formed between 2.9 and 2.5 Ga. The block, however, did not form as a distinct entity until the closure of the ocean between the Eastern

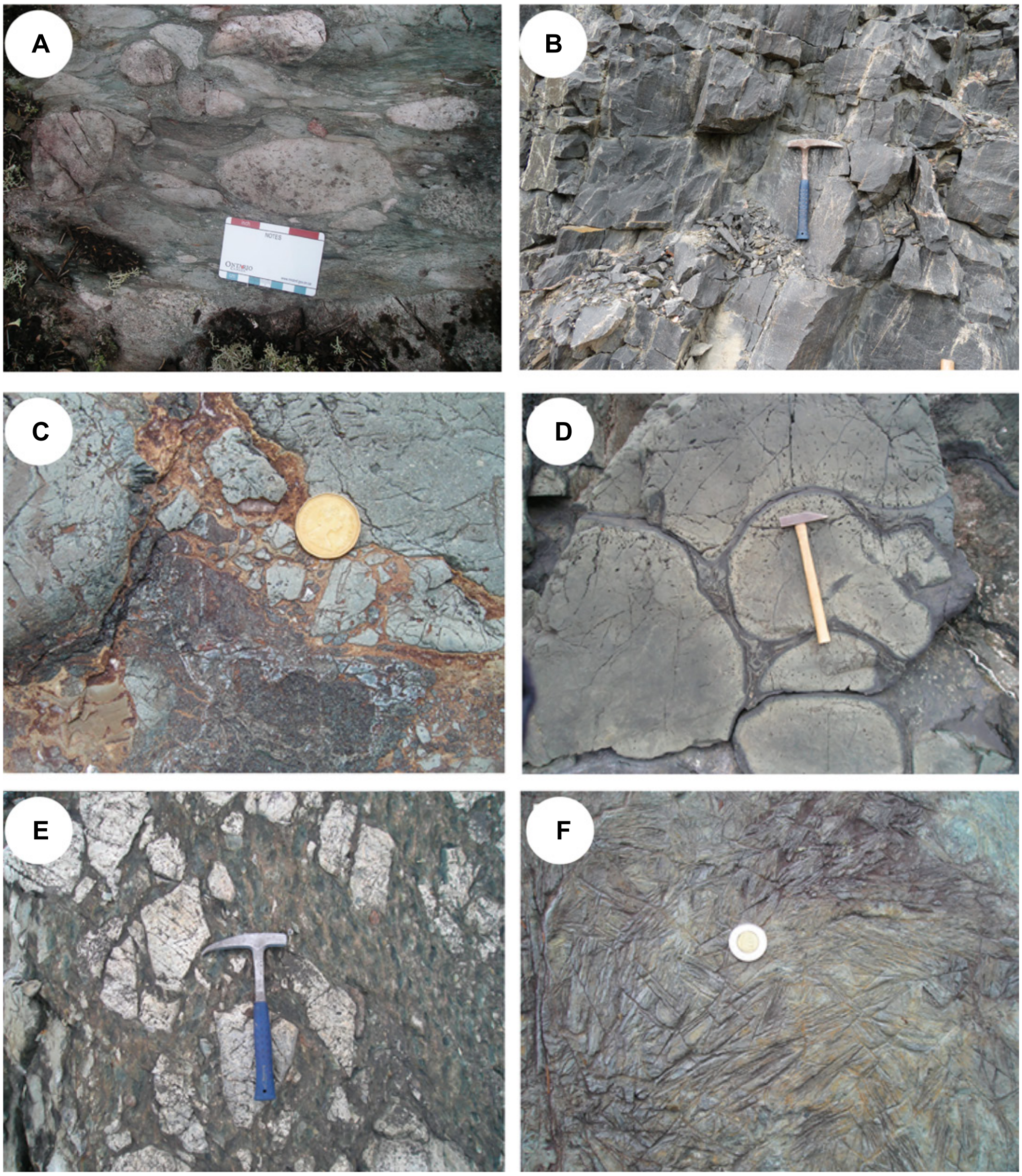

Figure 9. Photographs of outcrops of rocks deformed with middle to no strain from the Superior Province in Canada. (For localities see Figure 8; all photographs are by Ali Polat.) (A) Bad Vermillion Lake Belt, conglomerates. (B) Cross Lake Sandstone. (C, D) Ca. 2.7 Ga undeformed, greenschist facies pillow breccia and pillow basalts in the Schreiber-Hemlo greenstone belt, Wawa subprovince. (E, F) Autoclastite and komatiite, with spinifex texture, in the Abitibi subprovince, Superior Province. Photograph (C) is from Windley et al. (2021) and photograph (D) is from Polat et al. (2012).

and Western Blocks by 2.43 Ga. (Kusky et al., 2016; Fig. 12). But that date does not signify the end of significant internal deformation of the block and therefore one cannot speak of "cratonization" at 2.43 Ga. The North China Block consists in reality of three main tectonic units including the Eastern Block, Western Block, and the intervening Central Orogenic Belt (Fig. 12). The Eastern Block is predominated by Neoarchean granitoids with subordinate greenstone belts. It has been repeatedly deformed by strike-slip and extensional tectonism since the Mesozoic (Şengör and Natal'in, 1996; Yang and Xu, 2004 and Huang et al., 2015). The geology of the Western Block is poorly known because it is mostly covered by Proterozoic and Phanerozoic sedimentary rocks. The Ordos Basin constitutes a major part of the Western Block. The Central Orogenic Belt consists mainly of poly-deformed Neoarchean granitoids, mélanges and ophirags (Kusky et al., 2016, 2020).

The North China Block shares the major lithological, structural and geochemical characteristics of many Archean provinces including the Superior Province of Canada, suggesting that it originated through collision of different tectonic units such as island arcs, accretionary complexes, ophirags, oceanic plateaus, and continental blocks (Zhai, 2014; Kusky et al., 2016). The ages of accretion become progressively younger outward from the Central Orogenic Belt to the northwest.

In contrast to real cratons, the North China Block, however, has been affected extensively by Phanerozoic orogenic and taphrogenic events (Wang et al., 2018; Yang et al., 2019; Sun et al., 2020), which makes it very different from true Archean cratons. The northern margin of the North China Block is marked by the Paleoproterozoic Inner Mongolia–Northern Hebei Orogen, which consists of lithotectonic assemblages of the so-called "Khondalite Belt" to the south-southeast and the Alashan Block (also known as Alxa Block) to the north-northwest (Fig. 12).

According to the tectonic evolution model of Kusky et al. (2016) and more recent studies (Bi et al., 2018; Wang et al., 2018; Yang, et al., 2019; Jia et al., 2020), the major geological events in the North China Block and the neighboring regions started by the assembly of the Eastern Block between 2.7 and 2.6 Ga. This is followed

Figure 10. Photographs from the outcrops of the Barberton Greenstone Belt along the Komati River (all photographs are by A.M.C. Şengör). None of the rocks shown in these photographs show any metamorphism and barely any deformation. (A) Undeformed immiscibility spherules in Archean pillow lavas along the Komati River. (B) Well-preserved soft sediment deformation in turbiditic sandstones. Flame structures of finer-grained sediment (pale) injected into the overlying sandstone (purple-brown). (C) Poorly sorted, matrix supported conglomerates showing almost «Zero» deformation. (D) Archean pillow lavas from the Kromberg Complex.

by formation of island arcs between the Eastern and Western Blocks between 2.6 and 2.5 Ga which is resulted in the closure of the ocean around 2.43 Ga. Rifting of the newly formed continental block at 2.4–2.35 Ga along its center and northern margin resulted in the development of an Atlantic-type margin to the north. This margin was converted to an Andean-type margin through arc-polarity reversal by 2.3 Ga following a collision with an island arc. The Andean-type margin tectonic, magmatic, sedimentary and metamorphic processes affected much of the North China Block by 1.9 Ga.

The southern margin of the North China Block underwent rift-related magmatic, sedimentary, and metamorphic activities between 2.36 and 2.24 Ga that were followed by an orogenic event between 1.96 and 1.85 Ga and a rifting event between 1.84 and 1.74 Ga (Jia et al., 2020), indicating that the southern margin of the block was tectonically active throughout the early Proterozoic. Sun et al. (2020) proposed the development of an Andean-type margin in the southern part of the block at ca. 1.87 Ga. The eastern margin also underwent multiple rifting, subduction and collision events in the Paleoproterozoic between 2.47 and 1.86 Ga (Li et al., 2011; Yang et al., 2019), resulting in development of an Andean-type margin between 2184 and 2127 Ma (Bi et al., 2018). Continent-continent collision between the northern margin of the North China Block and the Columbia supercontinent resulted in granulite facies metamorphism and deformation across the block between 1.88 and 1.79 Ga. The North China Block was rifted away from the hypothesized Columbia supercontinent by 1.65 Ga, resulting in the development of a new Atlantic-type margin in the northern part of the block (but see Ding et al., 2020 for palaeomagnetic data suggesting a rifting 1.2 Ga ago[5]).

A Mesozoic orogenic event, nucleated on a Paleozoic extensional structure, and of medial Jurassic to Cainozoic age took place in the Helanshan between the Alashan and Ordos Blocks, which implies that these blocks could not have existed as a single cratonic entity before the Mesozoic (see especially Teilhard de Chardin and Licent, 1924; Yang and Zhang, 1987) or by the late Permian based on the palaeomagnetic results reported by Yuan and Yang (2015).

Deformation of the Luliangshan and Taihangshan in the Mesozoic (Wang et al., 2006) also indicates that an older collisional orogenic belt (Zhang et al., 2006; Zhao et al., 2006) between the Eastern and Western Blocks was reactivated and overprinted possibly during the medial Triassic (Kobayashi, 1942). Beneath the Mesozoic belt, the entire lithospheric mantle seems refertilized by younger magmatism (Tang et al., 2008). During the Eocene, this belt localized extensional deformation creating such structures as the Fen Wei Rift and the Jizhong Basin to its immediate east, which later became re-shortened (Zhao and Windley, 1990; Sun, 2005).

The entire periphery of the North China Block was affected by Mesozoic magmatism (Xiong'er Shan-Waifang Shan granites; Han et al., 2007) resulting from subduction under the North China Block (Wu et al., 2019) and subsequent collision. The Sulu ultrahigh pressure metamorphic belt of Shandong (Zheng et al., 2006) is nothing more than the easterly continuation of the Dabie Shan ultrahigh pressure metamorphic belt related to this northward subduction during the late Paleozoic–earliest Mesozoic (Okay et al., 1989, 1993; Okay and Şengör, 1992; Xu et al., 1992). In the north of the North China Block, the Carboniferous granites extend as far south as 41°N parallel as witnesses to southward subduction below the North China Block (Teilhard de Chardin 1932; Zhang et al., 2007) (Fig. 12).

Early Jurassic (190 Ma) to early Cretaceous (younger than 100 Ma in Liaoning, in the Songliao Basin: Davis, 2003) adakitic and

[5]Ding et al.'s data come from three localities: two within the Yin Shan orogenic belt and one in the so-called "Central Orogenic Belt" separating the Ordos Block from the eastern part of the North China Block (see their fig. 1). Data are not sufficient to tie the Ordos to the entire North China Block 1.2 Ga ago. Thus, we note Ding et al.'s observations, but reserve our judgment about the time of rifting they infer.

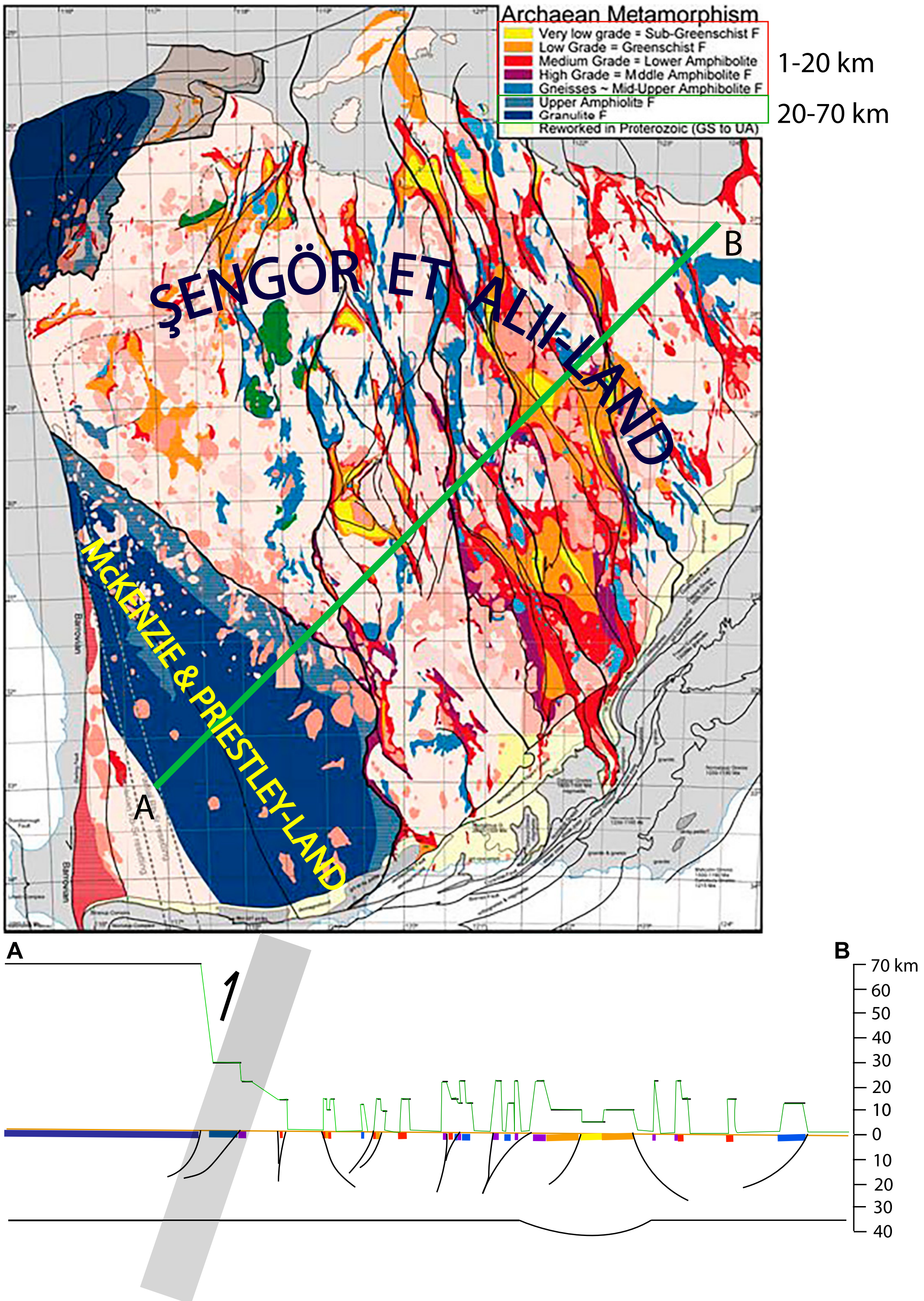

Figure 11. A simplified tectonic/metamorphic map of the Yilgarn Province (simplified from Goscombe et al., 2019). The schematic cross section shows the various heights inferred from metamorphic grade while the respective metamorphisms were going on. The areas designated by author's names simply indicate which areas of the Yilgarn are more compatible with which authors' hypotheses and are certainly not attempts to rename the provinces within the Yilgarn nucleus. They refer to discussions in this paper only. The reason we did not use the existing province and/or "terrane" names is that none of them covers what we wish to show. Notice the great difference between the McKenzie and Priestley-Land and the Şengör et alii-Land in terms of inferred elevations. Both now form, however, a single craton with similar crustal thickness. GS—greenschist; UA—upper amphibolite.

coeval non-adakitic magmatism spreads from the northwestern margin of the North China Block (Fig. 12). It seems clear from Davis' discussion of the available observations that these are not products of multiple episodes of lithospheric "delamination" or "decratonization" but products of the westward subduction, under thickened continental lithosphere (see also Wu et al., 2019), of a plate in the present space of the western Pacific Ocean much like the similar magmatism in the Andes (Davis, 2003). This episode of shortening in northeast China was followed by one of widespread extension around 130–120 Ma ago in an area exceeding 2 million km^2 (Davis, 2003; Davis et al., 2002; Liu et al., 2005), not dissimilar to the early Cainozoic extension that destroyed a North American "Altiplano" during the early Cainozoic and created the Cordilleran extensional metamorphic core complexes similar to those in northeastern China (Davis et al., 2002; Liu et al., 2005). The

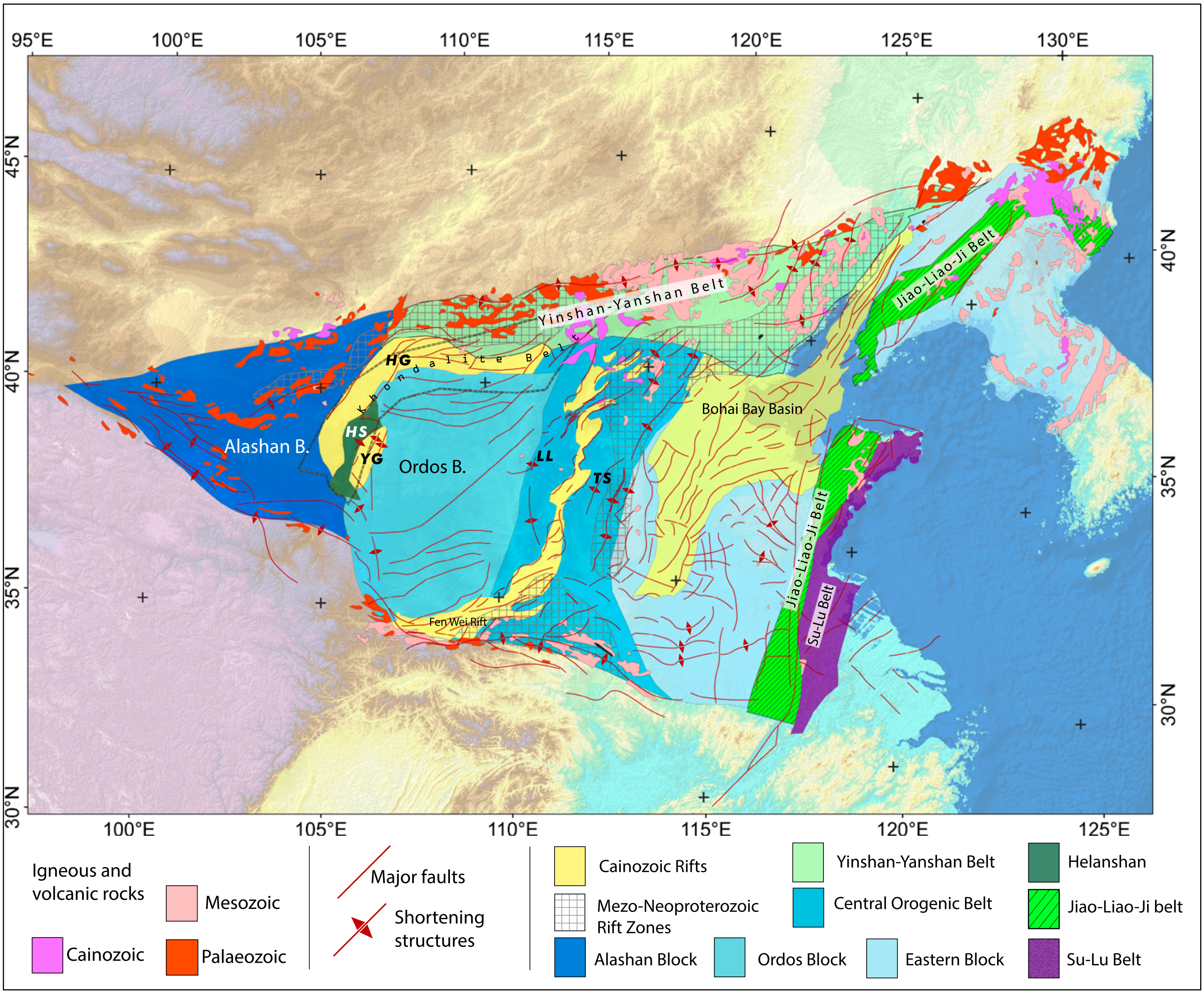

Figure 12. Summary map of the North China Block showing the major blocks and sutures. Note that all the tectonic elements shown here were repeatedly redeformed during the Phanerozoic and until about the Jurassic it is not possible with the available observations to be sure of their positions with respect to one another. Abbreviations as follows: HG—Hetao Graben; HS—Helanshan; LL—Luliangshan; TS—Taihangshan; YG—Yinchuan Graben. Redrawn from Deng et al. (2003), Kusky et al. (2016), Bi et al. (2018), Pang et al. (2020).

present-day thin lithosphere along the eastern edge of the North China Block is clearly a result of this late phase stretching (see Zhao and Zheng, 2007). As Şengör (2001) has argued, losing chunks of the lithospheric mantle (the preferred explanation of Zhao and Zheng, 2007) under zones of mantle upwelling is not incompatible with the stretching interpretation (see also Kusky et al., 2007b).

Did the North China Block Have a Thick, Continuous Precambrian Lithospheric Root?

The North China Block is distinct from many Archean cratons with its thinner (80–130 km) subcontinental lithospheric mantle beneath the eastern part (Kusky et al., 2007b; Wang et al., 2018; Wu et al., 2019; Xia et al., 2020). Based mainly on the presence Ordovician (ca. 480–450 Ma) diamondiferous kimberlite pipes from two locations and on the widespread occurrence of late Mesozoic and Cainozoic spinel lherzolites in the Eastern Block, many studies (Menzies et al., 1993; Zheng et al., 2007; Wu et al., 2019) argued that the cratonic mantle beneath the eastern part of the North China Block was thinned/removed and replaced by the asthenospheric mantle, leading to the decratonization of the eastern segment of the block. Using geophysical data, Xia et al. (2020) showed that most of the cratonic mantle beneath the North China Block was reworked in response to Phanerozoic tectonic and magmatic processes, which reduced the lithospheric thickness to 100–130 km beneath the northern, eastern, and western segments of block. The study by Xia et al. (2020) implies that thinner lithospheric mantle is not restricted to the eastern part but occurs throughout beneath the North China Block.

Although proposed ages for lithospheric thinning/removal range from 2.5 Ga to 120 Ma, most models favor decratonization and the loss of the lithospheric root in the Mesozoic (Menzies et al., 1993; Griffin et al., 1998; Zhai et al., 2007; Chen et al., 2008; Gao et al., 2008; Yang et al., 2008; Windley et al., 2010; Li et al., 2011; Zhu et al., 2012; Wang et al., 2018; Wu et al., 2019). A detailed review of these models is beyond the scope of this study. Accordingly, we present a short summary of these models. The majority of these models invoke hydration and weakening of the lithospheric mantle by subduction of oceanic lithosphere to the north and south, and flat subduction of the Paleo-Pacific oceanic crust beneath the North China Block in the Mesozoic (see Davis et al., 2001; Davis, 2003; Kusky et al., 2014; Zheng et al., 2018; Wu et al., 2019 for a review).

Windley et al. (2010) suggested that the Archean lithospheric mantle beneath the supposed craton was hydrated and weakened by the south-dipping slabs of Permian-Triassic Solonker and the Jurassic Mongol-Okhotsk subduction zones to the north, the north-dipping slabs of Permian-Triassic Qinling-Dabishan and Song Ma subductions to the south, and the west-dipping Jurassic-Cretaceous (200–100 Ma) Paleo-Pacific oceanic crust to the east. These processes eventually led to the thinning and delamination of the lithosphere beneath the Eastern Block. Similarly, Zhu et al. (2011, 2012) suggested that the westward subduction of the paleo-Pacific oceanic lithosphere beneath the North China Block played an important role in its lithospheric destruction. Wu et al. (2019) proposed that the lithospheric loss stemmed from Cretaceous orogenic collapse in response to Jurassic crustal thickening resulting from flat subduction of the paleo-Pacific oceanic lithosphere, carrying an oceanic plateau, between 210 and 140 Ma. None of the models above, however, take the timing of structural events into consideration and they are all incompatible with the tectonic calendar as documented by Davis (2003). Phanerozoic tectonics clearly shows that there never was a unified Precambrian entity worthy of the designation craton. This is much like the situation in Europe during the Phanerozoic where repeated tectonism did not allow the formation of a craton (Şengör et al., 2018a).

The question is: did the North China Block have a thick, continuous, rigid cratonic lithosphere before the Mesozoic? In other words, was it cratonized by the Paleoproterozoic (1.7 Ga) and remained intact until the Mesozoic? Although deformation in the early Cretaceous was mainly restricted to the Eastern Block, deformation in the Jurassic extended throughout the block (fig. 4 in Wu et al., 2019). Could a craton with a thick ($\sim$200 km) lithospheric mantle root undergo multiple phases of orogenic events and rifting? A record of widespread geological activity in the Proterozoic, Paleozoic, Mesozoic, and Cainozoic suggest that a thick cratonic lithospheric root beneath the North China Block could not have been fully developed by the end of Paleoproterozoic as commonly assumed. The development of a continuous, rigid lithospheric root might have been impeded by the incidence of the Paleoproterozoic orogenic and rifting events in the northern, southern, and eastern margins of the block, as outlined above, causing only a partial development of a lithospheric root. This partially developed lithospheric root was likely susceptible to destruction by tectonic and thermal erosion by the circum–North China subduction zones of various Paleozoic ages and would have provided the preconditions for its complete removal in the Mesozoic, but the internal taphrogenic and orogenic events along the Helanshan and the Taihangshan clearly show that there never was a united North China craton from under which a lithospheric armor could have been removed.

Although most studies assumed that the North China Block was underlain by a thick ($>$150 km) lithospheric mantle before the Jurassic (see Wu et al., 2019, for a review), its lateral distribution beneath the block before the Jurassic is not well known. Was such thick lithosphere restricted only to the eastern part of the block where the diamondiferous kimberlites occur, or did it exist beneath the whole block? Was the lithospheric mantle removed from beneath the block in the Mesozoic? If it was removed, how was it removed? These questions are the subjects of intense debate over the past 30 years (Kusky et al., 2007a, 2007b; Wu et al., 2019; Xia et al., 2020). The known diamondiferous kimberlite pipes are restricted to the eastern margin of the Eastern Block. If the whole North China Block was underlain by a thick lithospheric mantle such as that sampled by the Ordovician kimberlite pipes, one would expect to see the occurrence of such kimberlite pipes in other parts of the North China Block. Kimberlite pipes in other parts of the block either have not been discovered or they are not present due the absence of the thick lithospheric mantle beneath most of the block, which implies that a larger part of the block never developed a lithospheric layer thicker than 150 km to form diamonds.

Collectively, the tectonic evolution of the North China Block indicates that it underwent multiple phases of deformation, magmatism, metamorphism and sedimentation in its northern, southern, and eastern margins plus within it, around the Ordos Basin since the Neoarchean (Teilhard de Chardin and Licent, 1924; Kobayashi, 1941, 1942; He et al., 1983; Yang and Zhang, 1987; Darby et al., 2001; Davis et al., 2002), which raises the question whether the North China Block can be defined as a single, stable craton. No craton is known to us that exhibits this much tectonism and for such a long time within it. Consequently, we do not recognize the North China Block assembly as a craton: neither now nor any time in the past.

CONCLUSIONS

Ophiolitic remnants, igneous rocks with arc signature, structural thrust belts, juxtaposed rocks of different origin and ages, and the lithospheric structure derived from seismic reflectors in the Archean rock record show a high resemblance to the present-day subduction-accretion systems suggesting that subduction was operating throughout the Archean. This indicates that cratons formed by the amalgamation of magmatic arcs and associated subduction-accretion complexes that were eventually involved in a final continental collision that greatly thickened the underlying mantle lithosphere (Şengör, 1999; Priestley et al., 2021). The amalgamation may involve not only subduction, but also much strike-slip faulting. Both subduction and strike-slip faulting may intercalate episodes of local extension onto different parts of the system thus complicating its geological history. Even the final collision is probably only rarely head-on. This can be supported by the absence of high elevations in these cratonic areas as deduced from the surface geology and denudation rates. This leads us to the conclusion that craton formation does not require total removal of upper crust or exhumation of the entire lower crust. Instead, the granulitization of the roots of arcs may have been responsible for weighing the collided and thickened pieces down. The best candidate of a craton forming event is seen in the Altaids, a Paleozoic superorogenic system which has not been deformed orogenetically since the early Cretaceous (Şengör et al., 2014a, 2014b; 2018b).

We emphasize that when a craton is named, it must be indicated for what time period in the history of the earth that name and the area it implies are meant. Moreover, not every Archean or Proterozoic terrain is a craton. Our Figure 2 shows wide areas of Proterozoic and even Archean regions that are not cratons. Shields are multi-colored showing the ages of the underlying rocks identified in the legend. The platforms are shown in gray, except where sparse basement pieces crop out. The gray-colored platforms have two hues: darker gray indicates true cratonic platforms, about which there is no serious disagreement among geologists. Lighter gray indicates areas considered cratonic by most geologists, but we do not consider them as such because of a protracted deformation history that reaches into our own day, although they display a strength against convergent plate boundary deformations superior to that in orogenic zones, as defined in Şengör (1984), corresponding with Kober's (1921) orogens. Considering any ancient piece of crust, a craton has led to much confusion in the past as illustrated by the history of research on the Saharides, where an entire orogenic belt has been overlooked because of the abstruse concept of a "metacraton" (Şengör et al., 2020). Another example is illustrated by what is known as "North China Craton," It has undergone multiple episodes of deformation, magmatism, metamorphism, and sedimentation since the Archean, indicating that its geological characteristics are inconsistent with the definition of a craton.

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