Canadian Journal of Earth Sciences
Revue canadienne des sciences de la Terre

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Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2015-0110.R1
Manuscript Type:	Article
Date Submitted by the Author:	02-Sep-2015
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Dynamic Cause of Marginal Lithospheric Thinning and Implications for Craton Destruction: A Comparison of the North China, Superior and Yilgarn Cratons

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Abstract: We present a comparative tectonic analysis of the North China craton (NCC), which has lost parts of its root, with the Yilgarn and Superior cratons, which preserve their roots. We compare the geophysical structure and tectonic histories of these cratons to search for reasons why some cratons lose their roots, while others retain them. Based on the comparison and analysis of geological, geophysical and geochemical data, it is clear that the lithospheric thinning beneath craton margins is a common phenomenon, which may be caused by convergence between plates. However, craton destruction is not always accompanied by lithospheric thinning, except for cratons that suffered subduction and collision from multiple sides. The western block (also known as the Ordos Block) of the NCC, Yilgarn and Superior cratons have not experienced craton destruction; the common ground among them is that they are surrounded by weak zones (e.g., mobile belts or orogens) that sheltered the cratons from deformation, which contributes greatly to the long-term stability of the craton. Subduction polarity controlled the water released by the subducting plate, and if subducting plates dip underneath the craton, they release water which hydroweakens the overlying mantle, and makes it easy for delamination or sub-continental lithospheric mantle (SCLM) erosion to take place in the interior of the craton. Thus, subduction polarity during convergence events is an important element in determining whether a craton retains or loses its root.

Keywords: North China craton, Superior craton, Yilgarn craton, craton destruction

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1. Introduction

Cratons are tectonic units that account for about 50% of the Earth's land area. Main characteristics of cratons are that they can maintain long-term stability (>1 Ga), and deformation is extremely weak after their formation and stabilization. They formed in the Precambrian and are by definition more than 1.0 Ga old (Windley 1995), with most forming in the Archean (>2.5 Ga) and are composed of ancient continental crust and lower sub-continental lithospheric mantle (SCLM, or root). Due to the low heat flow values, low density, high seismic velocity, thick and refractory lithosphere (Morgan 1984; Shapiro et al. 1999; Sengör 1999; James et al. 2001), cratons have the ability to record a long geological history and avoid destruction in later geological events. However, studies about the Archean eastern North China craton (NCC; Fig.1B) indicate that it lost its root in the Mesozoic (Menzies 1993; Griffin et al. 1998; Zhang et al. 2002, 2003; Kusky et al. 2007a, b, 2014b; Zhu et al. 2012). Research on mantle xenoliths contained in diamond-bearing kimberlites combined with geophysical data show that the thickness of the lithosphere of the eastern NCC changed from 180-200 km in the Paleozoic to 80-100 km at present (Ma et al. 1984; Ma and Wu 1987; Yuan 1996; Menzies 1993; Griffin et al. 1998; Yang 2003; Kusky et al. 2007a; Li et al. 2009; Zhu and Zheng 2009). Furthermore, the nature of the lithosphere and the thermal state (Wu et al. 2008a) changed greatly at the same time, which reflects that the eastern NCC has been decratonized (Zhu et al. 2012), manifest as large-scale tectonic deformation and magmatic activity. Thus, stable cratons can become active under appropriate conditions, and the SCLM can become thinner, and its

1 composition can also be altered or the SCLM can be replaced with new fertile mantle ([Gao et al.](#)
2 [2002](#); [Zhai et al. 2007](#); [Zheng et al. 2007](#)).

3 In recent years much research has been done on searching for the dynamic cause of cratonic
4 lithospheric thinning (e.g., [Zhao et al. 2004](#); [Wilde et al. 2003](#); [Gao et al. 2002, 2009](#); [Xu 2001](#),
5 [2007](#); [Deng et al. 2004, 2007](#); [Zhu et al. 2012](#); [Miller et al. 2012](#); [Kusky et al. 2007a, 2014b](#)), with
6 much work focused on the destruction of the NCC. Continental-continental collision is considered
7 to be able to cause lithospheric thinning as this process can cause lithospheric or at least crustal
8 thickening ([Dewey and Burke 1973](#)), which leads to gravitational instability and therefore provides
9 advantages for the occurrence of delamination, and finally gives rise to the lithospheric thinning ([Li](#)
10 [1994, 1998](#); [Gao et al. 2002, 2009](#); [Zhang et al. 2002, 2003](#); [Wu et al. 2003, 2006](#)). Most scholars
11 also suggested that subduction of oceanic plates played a key role in the destruction of the NCC.
12 The subduction, flattening, and roll-back of the Izanagi (sometimes referred to as the Paleo-Pacific)
13 and Pacific plates is suggested by many to be the primary tectonic factor that led to the destruction
14 ([Griffin et al. 1998](#); [Wu et al. 2003](#); [Niu 2005](#); [Kusky et al. 2014b](#)). In addition, there are also some
15 scholars who regard mantle plumes as the cause of craton destruction ([Flower et al. 1998](#); [Zhao et al.](#)
16 [2004](#); [Wilde et al. 2003](#)). To clearly understand the problem, it is necessary to compare the unique
17 tectonic evolution of the NCC to the other representative Archean cratons, and conduct a
18 comprehensive comparison. Among the Archean cratons, we selected the Yilgarn and Superior
19 cratons which both preserve thick roots and are in a stable condition, to search for what in their
20 tectonic history is different that may have led to the destruction of one craton, and the preservation
21 of others.

22 The Archean Yilgarn and Superior cratons are located in southwest Australia and North

America, respectively. These two cratons still maintain a relatively stable state (Myers 1993; Kendall et al. 2002; Graham et al. 2004; Frederiksen et al. 2007; Huston et al. 2012) with no significant seismic activity, magmatic activity and little internal deformation. All three cratons formed in the Archean, but experienced different tectonic events since then, and now exhibit different states. Therefore, a comparison between these two different types of cratons can provide information about craton destruction and lithospheric preservation.

2. North China Craton

2.1 Geologic setting

The Archean North China craton (Fig.1) occupies about 1.5 million km² and is separated into the eastern and western blocks by the Central Orogenic Belt (Fig.1B; Griffin et al. 1998; Menzies and Xu 1998; Kusky and Li 2003; Zhao et al. 2005; Kusky et al. 2007a; Wu et al. 2008a; Santosh et al. 2010; Kusky 2011; Zhu et al. 2011, 2012). It is bounded by the Central China orogen (including the Qinling-Dabie belts) to the southwest, and the Central Asian Orogenic Belt to the north (Fig.1A; Davis et al. 2002; Windley et al. 2007; Kusky et al. 2007a), the eastern border is the Sulu orogen and its extensions in Korea along the Ogesheon belt (Oh and Kusky 2007). Most parts of the eastern NCC are dominated by lowlands with altitudes generally lower than 200 m. In contrast, the central and western NCC are characterized by high altitudes ranging from 500-3500 m (Jiang et al. 2013). Cenozoic rifts surround the Ordos Block (Fig.1B), to the east is Shanxi-Shaanxi Rift (SSR), and to the north is the Yinchuan-Hetao Rift (YHR, Fig.1B; Zhang et al. 1998).

Figure 1 around here

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1 The oldest rocks in the NCC were formed at about ~3.8 Ga, and crop out at Anshan (Liu et al.
2 1992; Wu et al. 2008b) and Xinyang (Fig.1B; Zheng et al. 2004). The dominant crustal age ranges
3 from ~2.5 Ga to ~3.0 Ga, with peaks in ages between ~2.7-2.9 Ga (Kusky et al. 2007c). The eastern
4 NCC mainly contains early to late Archean high and low grade tonalitic, trondhjemitic and
5 granodioritic (TTG) gneisses (Zhao et al. 2001). The western NCC is characterized by late Archean
6 to Paleoproterozoic metasedimentary belts which unconformably overlie the Archean basement
7 (Kröner et al. 2005). The Central Orogenic Belt includes a series of ~2.5-2.7 Ga amphibolite to
8 granulite facies terrains and ~2.5 Ga greenschist facies granite-greenstone terrains (Kusky 2011;
9 Kusky and Zhai 2012). This orogen contains a circa 2.5 Ga suture, where an arc is interpreted to
10 have collided with the Eastern block, a younger suture (~2.4-2.1 Ga) on the western side of the belt
11 where the Western block collided with the Central Orogenic Belt (Kusky et al. 2007a,c; Kusky and
12 Li 2003; J.P. Wang et al. 2013; Deng et al. 2013). The whole craton then experienced a high-grade
13 tectonothermal event at ~1.9-1.85 Ga, related to when the NCC joined the Columbia supercontinent
14 along its northern margin (Kusky et al. 2007c; Kusky and Santosh 2009; Peng et al. 2014).

15 Studies of mantle xenoliths contained in the diamond-bearing Ordovician kimberlites located
16 in the eastern NCC (Mengyin and Fuxian, Fig.1B) show that the lithospheric mantle beneath the
17 NCC had a thickness of about 200 km in the Paleozoic (~470 Ma) and maintained typical natures of
18 the Archean cratonic lithosphere mantle (Griffin et al. 1992, 1998; Menzies et al. 1993; Menzies and
19 Xu 1998; Wu et al. 2006). However, research on mantle-derived inclusions in Cenozoic basalts
20 indicates that the lithospheric thickness was reduced to about 80-120 km by then (Fan and Menzies
21 1992; Wu et al. 2003; Zheng et al. 2006), suggesting that more than 100 km of lithospheric thinning
22 has taken place in the eastern NCC since the early Paleozoic (Zhu et al. 2012). At the same time, the

nature of the SCLM changed essentially and is now characterized by high heat flow, abundant large earthquakes and significant deformation. The stable properties of the North China craton were lost in the Mesozoic (Liu et al. 2001; Zhang et al. 2002), which suggests that the eastern NCC had not only experienced lithospheric thinning but a wholesale destruction of the SCLM (Zhu et al. 2011, 2012). Compared to the eastern NCC, the western block has a lack of magmatic activity, has lower surface heat flow (Hu et al. 2000), a fairly thick crust (Li et al. 2006) and SCLM (Chen 2009). Recently, a series of geophysical data show that the SCLM thinning has also occurred in local parts of the western block (Chen 2009, 2010; Jiang et al. 2013), but the main body still preserves thick SCLM. So the western block has not been destroyed, only experienced local lithospheric thinning or modification (Zhu et al. 2012), and has remained relatively stable.

Since cratonization at ~1.8 Ga (Wilde et al. 2002; Zhao et al. 2001, 2005, 2006; Liu et al. 2009; Guo et al. 2005; Zhai et al. 2005; Zhai 2011; Zhao and Zhai 2013), the NCC remained relatively stable with thick Archean SCLM until the early Mesozoic. Because the destruction mainly took place after the Paleozoic, the following section is a review of its tectonic evolution since the Paleozoic (Table 1).

The Qaidam plate subducted beneath the NCC in the early Paleozoic and formed the active southern margin of the NCC (Yin and Nie 1996). The Yangtze craton subducted underneath the NCC in the Permo-Triassic (Li et al. 2004; Hacker et al. 2009), which resulted in exposure of ultra-high pressure rocks in the Dabie Shan and Sulu orogens, and westward extrusion of the southern Tarim-Qaidam block (e.g., Sengör 1985; Hacker et al. 2000; Ratschbacher et al. 2003). The subduction of Yangtze craton may have reached to depths up to 200 km (Ye et al. 2000; L. Wang et al. 2014), which definitely had an influence on the SCLM of the southern NCC (e.g., L.

1 [Wang et al. 2010, 2014](#)). The Tan-Lu fault may have initially formed during this convergent event
2 between the Yangtze craton and NCC ([Yin and Nie 1993](#)).

3
4 **Table 1 around here**
5

6 After the break-up of the Columbia (Nuna) supercontinent ([Rogers and Santosh 2002](#)), the
7 northern margin of the NCC was affected by the evolution of adjacent oceans, namely Panthalassa
8 ([Musashino 1990](#)), the predecessor of Paleo-Asian Ocean. The northern part of the NCC was
9 relatively stable from then until the Paleozoic ([Kusky and Li 2003](#); [Zhao et al. 2010](#); [Zhang et al.](#)
10 [2011](#)). During the early Paleozoic, the Bainaimiao arc developed along the southern part of
11 Xin-Meng orogen (part of the Central Asian Orogenic Belt, [Fig.1B](#)), but during this period, the
12 northern NCC was still stable. From the Paleozoic to early Permian, the Paleo-Asian Ocean to the
13 north of the NCC was open, so there were some active subduction zones which led to the growth of
14 the craton through accretion and arc magmatism along the northern margin ([Yin and Harrison 2001](#);
15 [Davis et al. 1996](#); [Xiao et al. 2003](#)). During the late Paleozoic, the northern margin of the NCC
16 changed into an active Andean style continental margin, and the Paleo-Asian Ocean subducted
17 beneath the NCC, which finally generated the collision between the NCC and the Mongolia block at
18 the end of the Permian ([Xiao et al. 2003](#); [Windley et al. 2007](#)). Studies of tectonism and magmatism
19 show that the closure of the Paleo-Asian Ocean affected the NCC ([Davis et al. 2001](#); [Xiao et al.](#)
20 [2003](#); [Zhang et al. 2003, 2010, 2011, 2012](#); [S.H. Zhang et al. 2014](#)). The gradual subduction of the
21 Paleo-Asian Ocean and collision of various terranes formed the Central Asian orogenic belt
22 ([Robinson et al. 1999](#); [Badarch et al. 2002](#); [Zhang et al. 2003](#); [Xiao et al. 2009a](#)). A series of studies
23 (e.g., [Xiao et al. 2003, 2009a, 2012](#); [Windley et al. 2007](#)) show that the evolutionary history of this

orogenic belt is very complicated, including the closure of the Paleo-Asian Ocean, the subduction of oceanic plates, the collisional event between the NCC and Siberia, and extensive magmatism (Li 2006). A triangular remnant embayment of the Palaeo-Pacific remained in eastern Asia in the mid-late Palaeozoic (Zorin et al. 1998; Zorin 1999) and is known as the Mongol-Okhotsk ocean which finally closed during the late Jurassic-Cretaceous (Tomurtogoo et al. 2005). About nine hundred plutons with late Paleozoic ages intruded in the region of Inner Mongolia located in the northern part of the NCC. Xiao et al. (2003) suggested that these plutons are related to the closure of Paleo-Asian Ocean, and that the ocean closure was associated with south-directed subduction beneath the northern margin of the NCC from 300-250 Ma. This closure combined the north China-Mongolia continent and the Siberia craton. The fact that the intrusions are younger from west to east along the line of the suture suggests that the ocean was closed in a scissor-like movement (Kravchinsky et al. 2002).

From the Paleoproterozoic to early Mesozoic, the NCC remained stable with deposition of thick clastic and carbonate rocks. It suffered different degrees of multi-phase deformation, but did not completely lose the basic attribute of a stable craton. In the late Palaeozoic, subduction and collisional events combined with volcanic activity changed the lithospheric thickness of the marginal areas and thermal character of the craton, which provided foundation for craton destruction in the Mesozoic and Cenozoic (Menzies and Xu 1998).

The evolution of the NCC was mainly controlled by the complicated interaction of Eurasian, Indian, Izanagi and Pacific plates during the Mesozoic and Cenozoic. The tectonic evolution during that period can be summarized as convergent events between the NCC and surrounding blocks. The subduction of the Izanagi plate under the Eurasian continent during the Mesozoic can be divided into

two phases: subduction and contraction during the Jurassic and subduction with extension in the overriding plate (Asia) during the Cretaceous (Wu et al. 2008a). This subduction caused the magmatism, deformation and mantle hydration in the eastern NCC during 200-100 Ma (Niu 2005; Kusky et al. 2007). Significant tectonic regime change of the eastern NCC happened in the late Jurassic to early Cretaceous, when the convergence dominated regime switched to an extensional regime (Menzies and Xu 1998; Kusky et al. 2014a). In the early Tertiary, the collision between India and Eurasia together with the rollback of the Izanagi plate may have caused the deformation of Shanxi-Shaanxi Rift and Yinchuan-Hetao Rift in the western NCC (Xu 2007).

2.2 Lithospheric structure

Geophysical methods are widely used in detecting the structure of the crust and mantle, and in recent years there has been a rapid increase of digital broadband seismic data in China. Results from these data show that the thickness of the SCLM of the eastern block of the NCC is about 60-70 km near the Tan-lu fault and increases to about 90-140 km to the northwest (An and Shi 2006; Huang and Zhao 2006; Huang et al. 2009; Li et al. 2009; Li and van der Hilst 2010; Zhao et al. 2012; Zhu et al. 2012). The eastern NCC can be subdivided into two parts represented and/or covered by the Southern North China Basin and Northern North China Basin (Fig.1B; Huang et al. 2009). The lithosphere of the Southern North China Basin is basically a high velocity lid down to 70-80 km (Fig.2B), and below that, the velocity decreases remarkably. On the other hand, the crust of the Northern North China Basin has a low velocity, and the thickness of the lithosphere is about 130 km (Huang et al. 2009). In general, the eastern NCC exhibits extremely thinned lithosphere that has already been acknowledged by numerous researchers (Menzies 1993; Kusky et al. 2007a; Zhu and

1 [Zheng 2009; Zhu et al. 2012](#)).

2 Along the northern margin of the NCC, the structure is different from that inside the NCC
3 ([Fig.2A](#)). The lithospheric structure beneath the northern margin is dominated by low velocity
4 anomalies, and thin lithosphere of no more than 100 km was imaged under a large part of the
5 northern margin of the NCC ([Tang and Chen 2008; Huang et al. 2009; Jiang et al. 2013](#)). Study of
6 Rayleigh waves in the northern NCC shows that the overall structure of this region is more similar
7 to tectonically active continental regions rather than a stable craton ([Tang and Chen 2008](#)). The
8 southern margin of the NCC also shows a relatively low velocity below 80 km ([Fig.2A; Zhao et al.](#)
9 [2012](#)). The eastern parts of the Qinling-Dabie belts are dominated by S-wave velocities of no more
10 than 4.4 km/s below 80 km ([Jiang et al. 2013](#)). Geophysical data above show that the lithosphere
11 beneath the marginal areas of the NCC has been thinned.

12 The strong contrast between the Ordos (Western) Block and the eastern NCC is an obvious
13 feature from the NCC tomographic images ([Fig.2](#)). [Fig.2](#) was derived from surface wave study
14 ([Jiang et al. 2013](#)) whose dominant resolution is confined to the depth interval of ca. 20–200 km.
15 Checkerboard tests ([Jiang et al. 2013](#)) performed to test the resolution were designed with a patch
16 size of $2^{\circ} \times 2^{\circ}$ (ca. 200 km) and velocity anomalies of $\pm 5\%$, results of the checkerboard inversions
17 verified that anomalies revealed by tomographic maps of S-wave velocities are reliable. We can
18 clearly infer from the seismic images that the western NCC has thick lithosphere represented by
19 high-velocity. The Central Orogenic Belt is generally characterized by low-velocity anomalies but
20 exhibits great heterogeneities, which may be the result of multiphase reworking of pre-existing
21 mechanically weak zones since the amalgamation of the craton. High V_p and V_s volumes, along
22 with low V_p/V_s volumes are apparent down to depths of ≥ 300 km beneath the Precambrian Ordos

Block (X. Wang et al. 2014), indicating that the SCLM is thick with lower temperatures, and has a relatively depleted geochemical composition. However, rifts around the Ordos Block are dominated by strong low-density anomalies (X. Wang et al. 2014; Kusky and Mooney 2015). High-resolution images derived from dense seismic-array data also indicate that the dramatically thinned SCLM is mainly present beneath the Yinchuan–Hetao and Shaanxi–Shanxi rifts, and thick SCLM is preserved beneath the Ordos Block (Chen 2009, 2010; Huang et al. 2009; Bao et al. 2011; Jiang et al. 2013; C.Y. Wang et al. 2014). Geophysical data suggests the coexistence of both thinned SCLM and a preserved cratonic root beneath the central and western NCC (Chen 2009, 2010; Jiang et al. 2013).

Figure 2 around here

Deep-seated xenolith data record the activities that the lithosphere suffered, which provides the most direct evidence for the structure of the SCLM. Therefore we present xenolith data related to the SCLM thinning beneath the marginal regions of the three cratons in this paper.

Xenoliths derived from Hannuoba which is located near the northern margin of the NCC records ages of $\sim 645 \pm 225$ Ma (Yu et al. 2007), ~ 470 Ma and ~ 315 Ma (Zhang et al. 2012), showing multi-episodic SCLM modification. A zircon age of ~ 470 Ma (Zhang et al. 2011) in the granulites has also been found in Fuxian near the northern margin of the NCC. In addition, a zircon age of ~ 462 Ma contained in xenoliths from the Mesozoic Liaodong basalt close to Fuxian (Zhang et al. 2011) also suggests that the Archean basement of the northern margin of the NCC suffered reworking in the Paleozoic. This is confirmed by seismic profiles across this margin (e.g., Huang et al. 2009). The situation of the southern margin is similar to that of the northern margin. The deep-seated xenoliths from the Mesozoic Jiagou (Fig.1B) dioritic porphyry in the southeastern

margin of the NCC comprise ages ranging from ~ 384 to ~ 410 Ma with a weighted mean age of $\sim 393 \pm 7$ Ma (Liu et al. 2013). Xenoliths derived from igneous precursors of various ages (~ 2.5 Ga, ~ 2.1 Ga and $\sim 393 \pm 7$ Ma) underwent variable degrees of metamorphic overprinting or reworking at ~ 2.1 Ga, ~ 1.8 Ga, $\sim 393 \pm 7$ Ma and $\sim 176 \pm 2$ Ma, respectively (Liu et al. 2013). Xenoliths from the Mesozoic igneous rocks of Xinyang also record age information of ~ 377 - 263 Ma (Zheng et al. 2008) and ~ 220 Ma (Li et al. 1993).

Heat flow investigations show a cold core is still preserved beneath the western NCC, with low heat flow value of < 50 mW/m², indicating a stable state. Heat flow in the Cenozoic rifts around the Ordos block is relatively high, which ranges from 63 to 98 mW/m² (Hu et al. 2000). The mean heat flow in the eastern NCC is ~ 64 mW/m², few boreholes in the ENCC even recorded high heat flow values of > 80 mW/m² (He 2015). Huang (2014) collected a large number of heat flow data in the NCC and concluded that the heat flow illustrates a generally eastward increasing trend. The lithosphere inferred from the result of geothermal study is in agreement with the geophysical and geochemistry research as discussed above.

Geological, geophysical and geochemical evidence show that the lithospheric thinning occurred beneath the eastern part, southern and northern margin of the NCC. The Ordos Block in the western part remained relatively stable with local lithospheric thinning of Shanxi-Shaanxi Rift and Yinchuan-Hetao Rift.

3. Yilgarn Craton

3.1 Geologic setting

The Yilgarn craton (Fig.5) formed in the Archean and is located in the southwestern part

(24°S-36°S, and 114°E-126°E) of the West Australian shield, covering an area of more than 750,000 km² (Mole et al. 2012). It is a typical Archean granite-greenstone terrane (Windley 1995; Myers 1990, 1995). To the north of the Yilgarn craton, lies another Archean craton named the Pilbara, separated by the Proterozoic Capricorn orogen (~1.7-2.2 Ga; Fig.3A, Occhipinti et al. 2004). The Proterozoic Pinjarra orogen is situated to the west of the Yilgarn craton, which is overlain by sedimentary rocks of the Perth basin. The southern border of the Yilgarn craton is the Proterozoic Albany-Fraser orogen (Fig.3A). The Yilgarn craton was formed between ~2670 and ~2650 Ma by accretion of micro continents (Myers 1990, 1993). Krapez et al. (2000) and Brown et al. (2001) interpreted the volcanic rocks in the craton to be associated with convergent margins, but other people such as Campbell and Hill (1988) and Hill et al. (1992) argued that the Yilgarn craton formed above a mantle plume. However, the geophysical data reported below shows no evidence to support a plume model, such as low velocity holes that would record the passage of a plume. In the past few decades, several specific subdivisions have been put forward for the Yilgarn craton (Gee 1979; Myers 1993, 1995; Tyler and Hocking 2001; Cassidy et al. 2004, 2006). Based on the plate tectonic environment, Cassidy et al. (2006) divided the Yilgarn craton into six terranes (Fig.3B): the Narryer terrane, Youanmi terrane, Kalgoorlie terrane, Kurnalpi terrane and Burtville terrane. Among them the Kalgoorlie terrane, Kurnalpi terrane and Burtville terrane are known as the Eastern Goldfields superterrane (Fig.3B).

Figure 3 around here

Much of the continental crust in the Yilgarn craton was generated after ~3.2 Ga (Griffin 2004a), and mainly between ~2720-2655 Ma which corresponds to the final assembly of different terranes (Huston et al. 2012). The oldest crustal components (~3.7 Ga) are identified in the middle of the craton (Griffin 2004a). Wilde et al. (2001) found samples in the region with the world's oldest zircon age of ~4.4 Ga, only 150 million years younger than the formation of the earth. The zircon

1 indicates the interaction of rocks with Earth's early hydrosphere, meaning that the ocean existed on
2 earth before ~4.4 Ga. [Harrison et al. \(2005\)](#) also supported that this zircon in the Yilgarn craton is
3 older than ~4.4 Ga based on the measurement of initial $^{176}\text{Hf}/^{177}\text{Hf}$ values. [Cassidy and](#)
4 [Champion \(2004\)](#) used Rb isotope and zircon data and combined the research on the upper crust by
5 [Swager \(1997\)](#) and [Krapez et al. \(2000\)](#), and then determined that the western Yilgarn craton is
6 older than the eastern parts.

7 Compared with the NCC, the Yilgarn craton has had a relatively simple tectonic evolution
8 history ([Table 2](#)) since cratonization.

9
10 [Table 2 around here](#)
11

12 The Yilgarn craton finished cratonization at around ~2655 Ma ([Czarnota et al. 2010](#); [Huston et](#)
13 [al. 2012](#)). During ~2215-1950 Ma, the Pilbara craton collided with the Yilgarn craton during a
14 multi-stage orogeny ([Cawood and Tyler 2004](#)). The Pilbara craton subducted under the Yilgarn craton,
15 and formed the Capricorn orogenic belt. Thereafter, these two cratons were combined as the West
16 Australian craton ([Cawood and Tyler 2004](#); [Occhipinti et al. 2004](#)). During the Yapungku-Capricorn
17 orogeny at ~1790-1780 Ma, the Western Australia craton subducted underneath the Northern
18 Australia terrane ([Bagas 2004](#); [Giles et al. 2004](#)) and formed a sinistral strike-slip fault along the SE
19 margin of the Yilgarn craton ([Huston et al. 2012](#)). During ~1345-1140 Ma, the amalgamation of the
20 West Australian craton and the Mawson craton (include the South Australian Element and large parts
21 of the east Antarctica continent; [Boger 2011](#)) formed the Albany–Fraser orogen ([Myers et al. 1996](#);
22 [Smithies et al. 2010](#); [Kirkland et al. 2011](#); [Boger 2011](#)). This orogeny mainly took place in two
23 stages: ~1345-1260 Ma (Stage I) and ~1215-1140 Ma (Stage II; [Clark et al. 2000](#); [Bodorkos and](#)

1 [Clark 2004](#)). The Pinjarra orogen which lies to the west of the Yilgarn craton comprises elements of
2 the Australo-Antarctic and Indo-Antarctic domains ([Cawood and Korsch 2008](#)). The orogen was
3 formed during the oblique collision of these continental fragments at about ~522Ma ([Collins 2003](#);
4 [Collins and Pisarevsky 2005](#)). During the late Jurassic, India, Antarctica and Australia began to
5 separate from each other ([Veevers 2006](#)) and completed the process at about ~120 Ma ([Norvick and](#)
6 [Smith 2001](#); [Veevers 2006](#)).

7 3.2 Lithospheric structure

8 The SCLM of the Yilgarn craton is characterized by high seismic velocity, it is depleted and
9 refractory, and has had little internal deformation since the Archean (e.g., [Fishwick et al. 2005](#);
10 [Fishwick and Reading 2008](#); [Jaques and Milligan 2004](#)). [Van Gerven et al. \(2004\)](#) used data from
11 broad-band seismometers to compute the possible compositional changes of the upper mantle of
12 Australia. They calculated the relative density-to-shear velocity scaling (ζ) of each main block. At
13 depths over 200 km under the Yilgarn craton, the ζ is negative, as a result, they speculated that the
14 thickness of the SCLM is about 200 km. [Goleby et al. \(2006\)](#) conducted a seismic survey line in the
15 Eastern Goldfields superterrane, and according to their results the lithospheric thickness is about 220
16 km. The tomographic results of [Fishwick and Reading \(2008\)](#) who used surface waves also support
17 that the Yilgarn craton has a thick SCLM. [Kennett and Blewett \(2012\)](#) integrated different seismic
18 data and then summarized the thickness of SCLM beneath the Yilgarn craton to be 200-250 km.
19 From north to south the lithospheric thickness has a tendency to become thinner, and from west to
20 east to gradually thicken. Thick SCLM can also be inferred from the bouguer gravity anomaly map,
21 and the Yilgarn craton has low gravity, because of its low density material ([Blewett et al. 2010](#)).

Figure 4 around here

Although there is not much data from kimberlites in the Yilgarn craton, studies of the existing diamondiferous kimberlites show the lithospheric thickness beneath the Yilgarn craton is around 175 km or thicker (Jaques and Milligan 2004; Graham et al. 2004). Geothermal data also show that the Yilgarn craton preserves thick lithosphere (Jaupart and Mareschal 2005), the whole Western Australian craton including the Yilgarn and Pilbara are characterized by low heat flow ranging from 40 to 50 mW/m² (Jaupart and Mareschal 2005). There is not any evidence demonstrating anomalous geothermal gradients in the Yilgarn craton (Weber et al. 2005). In addition, limited decoupling between the crust and SCLM is suggested by the coincidence of the Proterozoic mantle magma with the Archean mantle magma (Blewett et al. 2010).

Xenoliths derived from the eruption of kimberlites in the northern and eastern parts of the Yilgarn craton yield additional information about the SCLM (Jaques and Milligan 2004). Xenocryst-poor kimberlites from the Eastern Goldfields Province (EGP; Fig.3B) yielded ages of ~2025 ± 10 Ma (Graham et al. 2004). Additional age information of ~1300 Ma is recorded by xenoliths from the northern Yilgarn craton (Graham et al. 1999). Ultramafic rocks emplaced at about ~1900-1700 Ma (Shee et al. 1999) are recorded by diamond bearing kimberlites from the northern Yilgarn craton. Besides that, the Northampton and Mullingar complex (Fig.3B; located in the Pinjarra orogen) record deformation, magmatism and metamorphism that took place at around ~1080 Ma (Bruguier et al. 1999; Cobb et al. 2001).

More recent data have shown heterogeneities within the Yilgarn craton and confirm that the lithosphere is characterized by fast seismic velocity, with a depleted and refractory mantle, and fast

1 shear wave anomalies extending to more than 200 km (Simons et al. 1999, 2002; Reading and
2 Kennett 2003; Fichtner et al. 2009; Blewett et al. 2006, 2010; Fishwick et al. 2005, 2012; Fishwick
3 and Reading 2008; Ford et al. 2010; Kennett and Blewett 2012). Interestingly, Fishwick and
4 Reading (2008) point out that a distinct low velocity zone is observed at depths of 150-250 km in
5 the cross sections through the Archaean regions of the Yilgarn craton at 120°E (Fig.4B).
6 Checkerboard tests by Fishwick and Reading (2008) have verified the reliability of Fig.4. Both
7 perturbations and absolute velocities figures (Fig.4B) indicate that anomalies exist at depths of
8 150-250 km beneath the southern margin of the Yilgarn craton. Ford et al. (2010) carried out Sp
9 receiver functions to detect the lithosphere-asthenosphere boundary beneath Australia. The
10 calculated results of stations in the southern Yilgarn craton only show an anomaly at a depth of
11 about 90 km, which they interpreted as a mid-lithospheric discontinuity (MLD), but they did not
12 find any anomaly at greater depths that can represent the lithosphere-asthenosphere boundary
13 (LAB). Therefore, this anomaly interface could also be interpreted as the lithosphere-asthenosphere
14 boundary, which indicates a thinned lithosphere beneath the southern Yilgarn craton.

15 Seismic data above shows that the Yilgarn craton maintained stability and preserves a thick
16 lithosphere root of about 200 km, no abnormal velocity area was imaged beneath the interior of the
17 craton. However, in the marginal areas, the southern margin is represented by a zone of low wave
18 speed with perturbation of around -2% to -4% from 100 km to 150 km (Fig.4A), which is quite
19 different from the interior area, and no more anomaly at greater depths that represent the
20 lithosphere-asthenosphere boundary (LAB) can be found in this area from the seismic data.
21 Therefore the lithosphere of the southern margin is much thinner than other parts of the Yilgarn
22 craton. Overall, the Yilgarn craton has thick SCIM, while lithospheric thinning occurred beneath

local parts near the Albany-Fraser orogen. Geothermal and geochemistry data reflect that the Yilgarn craton is different from the eastern NCC and it is still a typical stable Archean craton.

4. Superior Craton

4.1 Geologic setting

The Superior craton (Fig.5) which is located in Canada is the largest craton in the world and the nuclei of the North American continent (e.g., [Thurston 2002](#)), covering an area of about 1.6 million km². The Archean Superior craton mainly contains Archean linear belts of volcanic-plutonic rocks, metasedimentary belts, and high-grade gneiss belts (Fig. 5). Most of these rocks formed between ~2.8 to ~2.7 Ga ([Hoffman 1989](#)). Since its formation around ~2.7 Ga through the accretion of arcs, accretionary prisms, and ribbon microcontinents ([Burke et al. 1976](#); [Card and Poulsen 1998](#); [Percival et al. 2006](#); [Evans and Halls 2010](#)), the Superior craton only experienced minor deformation. The Kapuskasing Structural Zone (KSZ; Fig.5B) divides the Superior craton into western and eastern parts ([Percival and McGrath 1986](#); [Evans and Halls 2010](#)). The western part mainly comprises granite-greenstone and metasedimentary belts or subprovinces ([Ferguson et al. 2004](#)) with east-west trends, which formed during the progressive accretion of arcs and micro-continents (e.g., [Thurston et al. 1991](#); [Kusky and Polat 1999](#); [Percival et al. 2006](#)). The Superior craton is also surrounded by orogenic belts (Fig.5A), including the Trans-Hudson orogen to the northwest (~1.8 Ga; [Hoffman 1988](#)), the Penokean orogen (~1.9-1.6 Ga; [Ludden and Hynes 2000](#); [Schultz and Cannon 2007](#)) to the south as well as the Grenville Province to the southeast (~1.3-1.0 Ga; [Hoffman 1988](#); [Thurston 2002](#); [Ludden and Hynes 2000](#); [Tohver et al. 2004, 2005](#)).

Figure 5 around here

The Superior craton formed during ~3.0-2.6 Ga through a series of magmatic arc complexes added by accretion events from the south (Card and Poulsen 1998; Percival et al. 2006). Since the late Archean, the craton has always kept a relatively stable state (Table 3; Thurston et al. 1991; Percival et al 2006; Frederiksen et al. 2007). Granite rocks intruded the craton during a short time after its cratonization. In the early Paleoproterozoic, numerous dyke swarms intruded the Superior craton and its southern margin developed sedimentary basins (Buchan and Ernst 2004; Bleeker and Ernst 2006). After ~2.0 Ga, the marginal regions of the Superior craton began to change to convergent tectonic regimes (Hoffman 1988; Ernst et al. 1995, 2014; Schulz and Cannon 2007).

The Archean Superior craton collided with the west Churchill Province between ~1.9-1.8 Ga, forming the Trans-Hudson orogen (Fig.5A; Eaton and Darbyshire 2010 and references therein; Schneider et al. 2007). The collision occurred after the closure of the Pacific-scale Manikewan Ocean (Eaton and Darbyshire 2010). Between ~1890 Ma and ~1830 Ma, an island arc collided with the Superior craton on the southern margin and ended the southward subduction, forming the Penokean orogen (Schultz and Cannon 2007). A notable intracratonic area in the Superior craton is the Kapuskasing Structural Zone which suffered significant deformation and brought deep-crustal rocks to the surface. The deformation is considered to be a response to collisional and orogenic events at about ~1.9 Ga (Percival and Peterman 1994). Later the region was affected by the Keweenawan Mid-continent rift between ~1.11-1.09 Ga (Thurston et al. 1991). The Mid-continent rift is a major structure that formed in an event that affected the southern Superior craton. It is a major continental rift in the world, extending about 2300 km in the central North America. The Grenville orogeny (Fig.5A) occurred in the late Proterozoic, which was the last major

accretionary/collisional event forming the Canadian Shield ([Hoffman 1988](#); [Thurston et al. 1992](#)). The orogeny includes two stages, the older one is the collision between Laurentia and Amazonia during ~1.2-1.12 Ga, the younger one is interpreted to represent intracratonic strike-slip at ~1.1 Ga ([Corrigan and van Breemen 1997](#); [Tohver et al. 2005](#)).

Table 3 around here

4.2 Lithospheric structure

Studies of surface waves from the western Superior craton ([Darbyshire et al. 2007](#); [Darbyshire and Lebedev 2009](#)) suggest that the thickness of lithospheric root extends more than 200 km ([Fig.6](#)), including the 40 km thick crust ([Kendall et al. 2002](#)). [Sol et al. \(2002\)](#) even estimated the thickness of the cratonic root of the western Superior craton to be around 300 km based on both P- and S-wave anomalies. Beneath the central regions of the Superior craton, the inferred thickness of lithosphere changes from 140 km to 200 km, and some areas in the center are estimated to be between 200 km to 240 km thick ([Darbyshire et al. 2007](#)). The depth of the lithosphere beneath Hudson Bay is generally greater than 180 km ([Darbyshire et al. 2011](#)). The continental-scale shear wave velocity models of North America show that most areas of the Superior craton are dominated by significant high-velocity anomalies to a depth of 200 km (e.g., [Godey et al. 2004](#); [van der Lee and Frederiksen 2005](#)).

Figure 6 around here

The tomographic velocity model derived by [Frederiksen et al. \(2007\)](#) shows the variation of lithospheric thickness across the Superior craton. The western part represents high velocity while the

eastern part is characterized by low velocity (Fig.6). The checkerboard resolution tests done by Frederiksen et al. (2007) used three-dimensional checkerboard patterns of alternating positive and negative spike anomalies, and excellent recovery of the input model indicates that the given seismic image (Fig.6) in this paper is reliable within at least 200 km in the vertical direction. Surface wave studies (Darbyshire et al. 2007) also show that the Superior craton exhibits large variations in lithospheric thickness ranging from 100 km to almost 220 km. The regions showing this difference are separated by a NNW-SSE oriented boundary (Fig.6) which also divides the regions with weak and strong shear wave splits (Darbyshire et al. 2007). In addition, they also discovered a 200-km-thick low-velocity anomaly corresponding to the failed-branch (Fig.6) of the ~1.1 Ga Keweenawan Mid-continent rift, and a linear low-velocity anomaly which may be caused by the Great Meteor hot spot track (Heaman et al. 2004).

The thick lithosphere extended to the diamond stability field beneath the Wawa area (Fig.5B; the southern Superior craton) prior to ~2.7 Ga, as inferred from studies of diamonds in Neoproterozoic lamprophyres and related volcanoclastic breccias (Stachel et al. 2006; De Stefano et al. 2006). Furthermore, diamonds contained in the ~2.697-2.700 Ga Wawa metaconglomerate were derived from the depleted mantle (Smith et al. 2012). Geothermal data of the Neoproterozoic Superior craton show that the maximum heat flow is about 41 mW/m² (Drury and Taylor 1987), and the inferred lithospheric thickness from the geothermal data for the southern Superior craton was at least 190 km at that time. Mantle xenoliths of coarse and deformed garnet peridotites (Meyer et al. 1994; Vicker 1997) from the Jurassic Kirkland Lake (also located in the southern margin of Superior craton; Fig.5B) kimberlites (Heaman et al. 2004) show a different situation. These garnet peridotites were plucked from depths between 60 and 150 km, suggesting that the maximum lithospheric thickness

1 was 150 km in the Jurassic. [Griffin et al. \(2004b\)](#) carried out a study on the composition of the
2 mantle below Kirkland Lake based on the geochemistry of garnet macrocrysts. The results indicate
3 that the upper mantle above 135 km is a typical Archean cratonic root. Below that depth, the mantle
4 is marked as metasomatized and fertile lherzolites, indicating a modified cratonic root ([Griffin et al.](#)
5 [2004b](#)). In addition, current geophysical data from the southern Superior craton have shown that the
6 lithospheric thickness is no more than 150 km ([Fig.6; Faure et al. 2011; Frederiksen et al. 2007](#)).
7 [Fig.6](#) clearly shows that the Wawa area is dominated by low velocity with perturbation of around
8 -2% below 100 km. That means the SCLM thinning also occurred along the southern margin of the
9 Superior craton. The SCLM thinning in areas adjacent to the Mid-continent rift happened before
10 ~1.1 Ga ([Miller et al. 2012](#)), because the maximum thickness of the SCLM of the area inferred from
11 the peridotitic garnet xenocryst's formation was already less than 150 km at ~1.1 Ga ([Kaminsky et al.](#)
12 [2002](#)).

13 [Queen et al. \(1996\)](#) reported $^{40}\text{Ar}/^{39}\text{Ar}$ phlogopite ages for a number of ~1140 Ma
14 lamprophyres from kimberlites in the Wawa area ([Fig.5B](#)). Also the Pele perovskite (in the Wawa
15 area) yields an age of $\sim 1172.3 \pm 4.6$ Ma ([Heaman et al. 2004](#)). The kimberlites from Kyle Lake have
16 ages of about $\sim 1076.2 \pm 3.8$ Ma ([Sage 2000; Kaminsky et al. 2002](#)). A Jurassic kimberlite pipe
17 ([Meyer et al. 1994; Vicker 1997](#)) near Kirkland Lake ([Fig.5B](#)) contains xenoliths with two main
18 stages of zircon growth. The first main stage of growth is around $\sim 2584 \pm 7$ Ma, $\sim 2629 \pm 8$ Ma,
19 $\sim 2633 \pm 3$ Ma, the younger one is between ~2.52 Ga and ~2.40 Ga ([Moser and Heaman 1997](#)). In
20 addition, the xenoliths there also show ages between ~157-152 Ma. The Attawapiskat kimberlites
21 record age information at around ~180 Ma and the Timiskaming kimberlites have ages between
22 ~155-127 Ma ([Heaman and Kjarvgaard 2000](#)).

Scully et al. (2004) did a geochemical analysis on xenocrysts contained in kimberlites in the Superior craton. The results show that a number of xenocrysts are similar with others found in other Archean cratons. However, significant variations in the garnet composition were also observed. Changes in mantle depletion with depth were inferred from the Rare-Earth Element features of different Mesozoic kimberlite pipes. Lithosphere with depths less than 120 km exhibits compositions which are closer to the primitive upper mantle (Scully et al. 2004). The Proterozoic pipes show greater depths of origin. Therefore, they proposed that the upper mantle of the Superior craton is highly heterogeneous.

Present heat flow models of the Superior craton indicate that the average geothermal gradient is about 42 mW/m^2 (Drury and Taylor 1987), which is almost the same as the worldwide average for Archean crust (41 mW/m^2 ; Nyblade and Pollack 1993). Research on the heat flow from 9 deep boreholes in the western Superior show that flow value is between 31 and 56 mW/m^2 , and the mean heat flow is $42 \pm 8 \text{ mW/m}^2$ (Rolandone et al. 2003). Perry et al. (2006b) also pointed out that the mean surface heat flow for the Superior craton is $40.9 \pm 0.9 \text{ mW/m}^2$ based on 70 land determinations. The Abitibi subprovince (Fig.5B) is distinguished from the rest of the craton, where large volumes of greenstones are responsible for the relatively high heat flow value in this area (Mareschal and Jaupart 2004). Geophysical and geochemical evidence presented above indicate that the Superior craton is underlain by lithospheric mantle that exhibits strong regional variations in anisotropy, velocity structure and composition (Darbyshire et al. 2007; Scully et al. 2004). The lithosphere has been thinned beneath the southern margin and xenolith data record a possible related event at about $\sim 1.1 \text{ Ga}$. On a larger scale, the Superior craton still has a thick SCLM (Carlson et al. 2005), and is characterized as a stable and typical Archean craton.

5. Discussion

5.1 Subduction polarity

When oceanic lithosphere subducts beneath cratons, water is carried into the mantle via hydrous mineral phases in the slab and overlying sediments, and the most important dehydration reactions occur at the facies transition from blueschist to eclogite, which releases abundant water to the overlying mantle (Peacock 1993). This hydrates and thereby weakens the upper mantle and therefore lowers the melting temperature (solidus), and decreases the mantle viscosity (Komiya and Maruyama 2006; Sandu et al. 2011; Kusky et al. 2014b). Water would preferentially be removed with the melt in the partial melting process of mantle peridotite and pyroxenite, leading to the formation of refractory lithospheric mantle (Karato 1986; Hirth and Kohlstedt 1996). The higher degree of mantle melting would eliminate more water and result in greater stability of the lithospheric mantle. Therefore, water plays a key role in the formation, modification and destruction process of the craton (Niu 2005; Maruyama and Okamoto 2007; Kusky et al. 2007a, 2014b). Thus, the long-lived subduction beneath the craton is important to its destruction. On the contrary, when subduction is dipping away from a craton, the hydration of the mantle is very limited which greatly contributes to the preservation of the SCLM.

The NCC experienced multiple episodes of sub-craton subduction, mantle hydration, and collisional events (Table 1; Maruyama et al. 2007; Windley et al. 2010; Xiao et al. 2003, 2009a, b; Kusky et al. 2007a, 2014b), and most polarities of the subduction were toward and underneath the NCC. For example, the Yangtze craton subducted underneath the NCC in the Permo-Triassic, the Solonker ocean floor was subducted underneath the NCC during 500-250 Ma and the Izanagi plate

1 also dipped underneath the NCC during 200-100 Ma. These subduction events (especially the
2 long-term subduction of the Izanagi plate during the Cenozoic) are suggested to have led to
3 remarkable hydroweakening of the lithosphere, which promoted the generation of melts and the
4 interaction with the overlying SCLM ([Maruyama and Okamoto 2007](#); [Windley et al. 2010](#); [Kusky et](#)
5 [al. 2007a, 2014b](#)). The oceanic lithosphere that subducted underneath the NCC is estimated to be
6 more than any other craton on the Earth ([Kusky et al. 2007b](#); [Windley et al. 2010](#)), and the loss of
7 the SCLM of the eastern NCC is related to the ingress of large amount of water released by the
8 subducted oceanic lithosphere. The dehydration of oceanic lithosphere triggered partial melting in
9 the overlying mantle wedge and played a key role in the generation of hydrous magmas which
10 reduced the viscosity of the lithosphere and made it behave similar to the asthenosphere, therefore
11 setting the stage for the loss of the SCLM ([Kusky et al. 2014b](#)). Consequently this gave rise to the
12 unsteadiness of mantle convection beneath the NCC, which softened the lithosphere and made it
13 easy for delamination or SCLM erosion to take place in the interior of the craton (e.g., [Sleep 2005](#)).

14 Compared with the NCC, most subduction events dipped away from the Yilgarn and Superior
15 cratons ([Table 2, 3](#)), or the craton experienced relatively little subduction, but had accretion of arcs
16 and other blocks along the margins of the craton. During the amalgamation of the West Australian
17 craton and the Mawson craton, the subduction was away from the Yilgarn craton. The subduction
18 also dipped away from the Yilgarn craton during the oblique collision ([Table 2](#)) of the
19 Australo-Antarctic and Indo-Antarctic elements. Regarding to the Superior craton ([Table 3](#)), when
20 an island arc collided with the Superior craton on the southern margin and forming the Penokean
21 orogen, the subduction was away from the Superior craton. The subduction still dipped away from
22 the Superior craton in the collision between Laurentia and Amazonia during ~1.2-1.12 Ga (Grenville

orogeny). So in these cases, the influence to the SCLM is limited. With almost no subduction and mantle hydration beneath the craton, which provides advantages for preserving the SCLM, this is the main reason that the Yilgarn and Superior cratons still maintained stability and currently still have cratonic characteristics.

As the subduction polarity controls the water released to the SCLM and water plays a key role in the formation and destruction of cratons, subduction polarity has a great effect on craton destruction. Based on the tectonic events that the NCC, Yilgarn and Superior cratons experienced, it is reasonable to conclude that the SCLM is more likely to be preserved when subduction has a long-term history of dipping away from the craton.

5.2 Marginal Lithospheric thinning

During the comparison of geological, geophysical and geochemical data of the three cratons, we found that the Yilgarn craton and Superior craton remained stable as a whole and also show properties of typical cratons, but lithospheric thinning still occurred in local regions. Geophysical data show that these regions are limited to the craton margins where plate subduction or collision has occurred.

Yilgarn craton

Seismic data combined with geothermal data show that most areas of the Yilgarn craton have lithosphere more than 200 km thick, while the situation beneath the southern margin is a little different. Fishwick and Reading (2008) have shown that the southern margin of the Yilgarn craton (Fig.4A; south of 32°S) is an abnormal area with low seismic velocities between 150 and 250 km (Fig.4B), but they presented no speculations about the reason. Considering the area is inside the

1 craton and far away from the continent-ocean transition region, and the depth of the lithosphere is
2 much shallower than other marginal areas of the craton inferred from the seismic image (Fig.4B),
3 this area most affected by subduction events. Therefore we suggest that it is cratonic lithosphere
4 which has been thinned.

5 The area is close to the Albany-Fraser orogen (Fig.3B) which formed in the collision between
6 the west Australian craton and the Antarctic continent (Myers et al. 1996) during ~1345-1140 Ma.
7 From the perspective of space, the area with lithospheric thinning corresponds to the Albany-Fraser
8 orogen. In addition, in the aspect of time, scarcely any tectonic events could have had effects on the
9 SCLM around the region except the Albany-Fraser orogeny during the evolution of the Yilgarn
10 craton. Based on U-Pb, Lu-Hf and Sm-Nd data, Kirkland et al. (2011) also suggested that the
11 southern margin of the Yilgarn craton has been re-worked during the formation of the Albany-Fraser
12 orogen. Therefore, the convergence between plates is most likely to be the dynamic factor of the
13 thinning of the SCLM at this region. So far, there are scantily few xenolith data from the Yilgarn
14 craton, while xenocryst-poor kimberlites in the Jewill field (Fig.3B) and lamprophyre contained in
15 the Buljah field record age information of ~1300 Ma (Graham et al. 1999). Besides that, the
16 Northampton and Mullingarra complex (basement of the Pinjarra orogen; Fig.3B) record that the
17 deformation, magmatism and metamorphism took place at around ~1080 Ma (Bruguier et al. 1999;
18 Cobb et al. 2001). The age information contained in the xenoliths and deformation at about 1080 Ma
19 may have connections with the subduction and collisional events during the Albany-Fraser orogeny
20 for the time is coincident but there is not any clear geological evidence. If this hypothesis is proven
21 true, then this would provide solid evidence that the abnormal area with low velocity was caused by
22 SCLM thinning related to the convergence between the Yilgarn craton and the Antarctic continent.

1 Superior craton

2 Evidence from geochemistry and geothermal data show that the lithospheric thickness of the
3 southern margin of the Superior craton thinned from at least 190 km in the Neoarchean to the
4 maximum thickness of 150 km in the Jurassic and geophysical data also shows that the lithosphere
5 beneath the southern margin of the Superior craton is quite different from the other parts. [Miller et al.](#)
6 [\(2012\)](#) proposed that the diamondiferous root below the southern Superior craton has disappeared,
7 and the lithospheric thinning beneath the southern Superior may be caused by tectonic erosion,
8 craton amalgamation, and ingress of asthenospheric fluids. However, they did not have a clear
9 explanation for the specific dynamic reason for the lithospheric thinning. The thick lithosphere
10 extending to the diamond stability field existed beneath this area (the Wawa area; [Fig.5B](#)) prior to
11 ~2.7 Ga, which was inferred from studies of diamonds in the Neoarchean lamprophyres and related
12 volcanoclastic breccias ([Stachel et al. 2006](#); [De Stefano et al. 2006](#)). However, detailed geophysical
13 data presented in section 4.2 shows that the southern Superior craton does not have high seismic
14 velocities indicative of lithosphere thicker than 150 km ([Frederiksen et al. 2007](#); [Faure et al. 2011](#)).
15 Geochemical studies of garnet macrocrysts from the mantle below Kirkland Lake also suggest that
16 the cratonic root has been modified since the lithosphere below 135 km is characterized by
17 metasomatized and fertile lherzolites ([Griffin et al. 2004b](#)).

18 After cratonization of the Superior craton, its southern margin experienced the Penokean
19 orogeny and Keweenawan Mid-continent rifting events. Both events may have caused lithospheric
20 thinning, but combined with the age data contained by the xenoliths for the timing of tectonic events,
21 we can have a better estimation about which dynamic factor caused lithospheric thinning.

22 Age information contained in xenoliths from the southern margin including ~1140 Ma
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1 lamprophyres (Queen et al. 1996), $\sim 1172.3 \pm 4.6$ Ma perovskite (Heaman et al. 2004), showing that
 2 the SCLM has been affected during $\sim 1150 \pm 20$ Ma. The Keweenaw Mid-continent rift formed
 3 before ~ 1.1 Ga (Thurston et al. 1991) and the Penokean orogeny took place between ~ 1.9 - 1.6 Ga
 4 (Ludden and Hynes 2000). According to the evidence provided by xenoliths, the SCLM thinning
 5 beneath the southern margin was more likely caused by the Keweenaw Mid-continent rifting event.
 6 Based on the major and trace element chemistry of garnets from kimberlites and alkaline intrusions,
 7 Scully et al. (2004) proposed that the mantle metasomatism in the southern Superior craton is
 8 contemporaneous with the ~ 1.1 Ga Mid-continent rift system. This major continental rift certainly
 9 had great significance in the modification of the mantle lithosphere. The peridotitic xenoliths from
 10 the Jurassic kimberlite near Attawapisk (Fig.5B) also provide evidence that the lithosphere of the
 11 Wawa-Abitibi terrane in the southern Superior craton was affected by the Mid-continent rift. The
 12 T_{RD} ages (Re depletion ages) of these xenoliths is between ~ 1.6 and $\sim 1.4 \pm 0.2$ Ga (Smit et al. 2014),
 13 which are consistent with ~ 1.5 and ~ 1.1 Ga whole-rock T_{RD} ages for xenoliths from Kirkland Lake
 14 in the southern Superior craton (Pearson et al. 1997). These whole-rock T_{RD} ages are interpreted as
 15 mixing ages derived during the ingress of the Mid-continent rift related melt into the Archaean
 16 Superior lithospheric mantle (Smit et al. 2014).

17 Active rifting is thought to be caused by the interaction between plates (Sengör and Burke
 18 1978), which can become one of the candidate models for craton destruction when combined with
 19 thermo-chemical erosion (Zheng and Wu 2009). Gordon et al. (1986) explained the origin of the
 20 Mid-continent rift as a result of the Grenville orogenic event based on evidence from isotope ages.
 21 Furthermore the Mid-continent rift is located in the Penokean orogen which underwent multi-stage
 22 deformation, and hence it is easy to be affected by younger deformation. Therefore, the SCLM

thinning along the southern margin of the Superior craton is essentially affected by plate convergence.

Comparison with the NCC

Studies show that the convergent events along the southern NCC, western subduction of the Izanagi plate and southern subduction and collisional events occurred at the northern margin of the NCC are responsible for the lithospheric thinning beneath the southern and northern margin and also had great effect on craton destruction (Deng et al. 2007).

Northern margin of the NCC

The reactivation and subsequent orogenesis along the northern margin of the NCC were initiated in the late Paleozoic (Y. Wang et al. 2013; Z. Zhang et al. 2014). Geochemical and isotopic compositions of the Triassic Shuiquangou (Fig.1A) volcanic rocks and basic-ultrabasic complex and granitic intrusive rocks from the northern margin of the NCC shows that the asthenosphere was involved in the formation of these rocks (Zhang et al. 2010). However, no mid-ocean ridge or hotspot existed on the northern margin of the NCC during the Mesozoic. These depleted asthenospheric mantle compositions may come from decompressional melting of upwelling asthenospheric mantle caused by the post-collisional orogenic extension and collapse. Therefore, the convergent events would be expected to modify the lithosphere along the northern margin of the NCC (Zheng et al. 2012). As the upwelling of asthenospheric mantle and the evidence for an asthenospheric source composition in magmatic rocks are regarded as the beginning of destruction, lithospheric thinning of the northern margin of the NCC may have already started in the late Triassic. The upwelling of asthenosphere modified the SCLM by providing heat and reactions between melts and peridotite. Basaltic magma produced by partial melting of asthenosphere and lithospheric

mantle even ascends to the crust-mantle boundary. So from the Paleozoic to early Mesozoic, the closure of the Paleo-Asian Ocean and subsequent collision modified the lithosphere of the northern NCC.

Xenolith data described in section 2.2 including age information of $\sim 645 \pm 225$ Ma (Yu et al. 2007), ~ 470 Ma (Zhang et al. 2012) and ~ 462 Ma (Zhang et al. 2011) from the northern margin of the NCC, suggesting that the northern NCC suffered reworking during the Paleozoic. From the evolutionary history of the NCC, we conclude that the age information contained in the xenoliths corresponds to the time of subduction of the Paleo-Asian Ocean and collisional events after that, which means that the convergence between plates affected the SCLM (Zheng et al. 2012). All of this temporal evidence from xenoliths demonstrates that the lithospheric thinning of the northern NCC was related to the convergence between plates.

Southern margin of the NCC

Paleozoic (~ 377 -260 Ma) and early Mesozoic (~ 228 -219 Ma) zircons are also found in eclogites and peridotites derived from Qinling-Dabie-Sulu Orogenic Belt (Fig.1B), showing the modification of the lower crust and the uppermost mantle (Zheng et al. 2012). Liu et al. (2013) reported the Mesozoic Jiagou (Fig.1B) dioritic porphyry has an average age of $\sim 393 \pm 7$ Ma. In addition, 80% of zircons in the eclogite xenoliths from the Xinyang area give Paleozoic (~ 377 -263 Ma) U-Pb ages (Zheng et al. 2012), indicating that the southern margin of the NCC suffered intense modification during that time. This age information contained in xenoliths point out that the convergence between the Yangtze craton and the NCC was responsible for the reworking of the southern NCC. Zheng et al. (2005) studied peridotite xenoliths in the early Mesozoic volcanic rocks from Xinyang (Fig.1B) located in the southern margin of the NCC. The results also show that the

SCLM beneath the southern NCC has been modified by the convergence between the Yangtze and North China cratons. [Zheng et al. \(2012\)](#) compared the mantle inclusions in volcanic rocks derived from different locations (such as the centre and margin of plate, Tan-lu fault) and different time (e.g., the Paleozoic, Mesozoic and Cenozoic), and the peridotite from the southern NCC and Dabie-Sulu ultrahigh pressure metamorphic zone, and pointed out the inhomogeneity among the structure and composition of the SCLM. They suggested that the integrity of the NCC was destroyed by the subduction and collision of Yangtze craton during Triassic times.

The descriptions above suggest that subduction and collisional events at the northern and southern margin of the NCC in the Paleozoic were responsible for the SCLM thinning beneath the northern and southern margin of the NCC and also played a key role in the early destruction of the eastern NCC. [L. Wang et al. \(2014\)](#) also showed that melts derived from deeply subducted eclogite in the Sulu orogen infiltrated the base of the NCC SCLM at circa 228-219 Ma, further enhancing destruction of the base of the southern NCC.

All three cratons discussed here have experienced different degrees of lithospheric thinning beneath their marginal regions. The convergence between plates can affect the SCLM, which is the dynamic cause of marginal lithospheric thinning, at the same time it can only cause the SCLM thinning of the marginal areas but have almost no effect on the internal regions of a craton. Here we emphasize that craton destruction includes lithospheric thinning and the essential change of the nature of the lithosphere and the thermal state, which is different from lithospheric thinning ([Zhu et al. 2012](#)). Thus, craton destruction is not always accompanied by lithospheric thinning, except for cratons which suffered subduction and collision from multiple sides. Therefore, subduction and collision from multiple sides is an important factor for the destruction of cratons.

5.3 Protection by weak zones

An interestingly similar feature among the western NCC, Yilgarn and Superior cratons which all preserve thick SCLM, is that they are all surrounded by younger orogens (Fig.1A, Fig.3A, Fig.5A). The multiple periods of orogenesis of the surrounding belts indicates that stress induced by collision, without significant deformation of the cratonic interiors, is concentrated in these weak zones around the craton. Thus the orogens may shelter the craton from deformation, which contributes to the long-term stability of the craton (e.g., Lenardic 2000, 2003).

Windley (1995) observed that Archean cratons are typically surrounded by younger mobile belts, but the mechanical relationship between these two types of tectonic units has not been totally understood. A series of numerical studies suggest that as soon as a craton is surrounded by weak zones, the stress associated with mantle convection is buffered and the craton is prevented from being further deformed (Lenardic et al. 2000, 2003; Yoshida 2012). Results of these numerical studies suggest that the weak (low-viscosity, e.g., mobile belts) continental margin contributes most to the longevity of the SCLM (Yoshida 2010, 2012). The long-lived and inherited weakness of mobile belts can increase the longevity of cratonic roots by buffering the craton from mantle derived stresses. Because the stresses induced by collision between plates are transmitted to the SCLM by the weaker lithosphere of the mobile belt, the stress level that the SCLM experiences when suffering collisional events is below its yield stress (Lenardic et al. 2000). This prevents lithospheric failure and allows the high viscosity of cratonic lithosphere to resist mantle recycling (Lenardic et al. 2000). The following section shows how this general phenomenon is appropriate for all three studied cratons.

The viscosity of the lithosphere beneath grabens around the Ordos Block is lower than that of the Ordos Block by 1-2 orders of magnitude (Zang et al. 2007). The two grabens around the Ordos Block correspond to the tectonic belt that existed in the NCC since the Archean. The Yinchuan-Hetao rift is located in the Paleoproterozoic North Hebei orogen (Fig.1B; Kusky et al. 2007c) and the Shanxi-Shaanxi rift roughly coincides with the Central Orogenic Belt (Fig.1B) that formed when the western block collided with the eastern block at about ~2.4-2.1 Ga (Kusky 2011; Kusky and Mooney 2015). Compared to the core part of the Archean craton, these tectonic belts may have remained relatively weak since their formation, and they have been affected by multiphase tectonic events during the long-term evolutionary process. Therefore, these areas have been modified or even thinned which is indicated by the geophysical data (see details in section 2.2). As differences of properties (even the thickness) exist between the craton and mobile belts, the tectonically weak belts around the Ordos Block are more likely to be affected by tectonic events. Studies show that the latest Cenozoic lithospheric modification may be attributed to the far-field effects of both Pacific plate subduction and the India-Eurasia collision (Deng et al. 2004; Xu 2007; Jiang et al. 2013). It is the modification and thinning of lithosphere beneath these weak zones that reduced the total stress that the craton suffered, which protected the western NCC from destruction.

Comparison of rheology of the SCLM beneath the eastern and western parts of China by Wang et al. (2001) indicates that the overall strength of lithosphere beneath extensional basins of the eastern part is relatively weak, with numerous ductile layers in the crust, while the western contractional basins have relatively high lithospheric strength. Numerical studies (Lu et al. 2011) indicate that higher strength of the lithosphere would increase the probability of the formation of rifts (e.g., the Yinchuan-Hetao and Shanxi-Shaanxi rifts; Fig.1B). Narrow rifts behave as weak

1 zones that reduce the total stress within craton and help to stabilize the craton (Lu et al. 2011).
2 Lithosphere with relatively low strength is more likely to deform uniformly without localized
3 deformation. Therefore, wide rifts are usually observed in destructed cratonic areas, e.g. the eastern
4 NCC. In addition, Lu et al. (2011) used numerical modeling to show that a craton in a cold thermal
5 state is more inclined to experience localized deformation. On the contrary, a craton with a hot
6 thermal state tends to experience widespread or uniform deformation. On the other hand, relatively
7 dry lithosphere would develop localized deformation easier than wet lithosphere. Thus, cratons can
8 develop localized deformation (e.g., narrow rifts) to keep their stability and protect the remaining
9 parts of the craton from subsequent deformation, which provides a reasonable explanation for the
10 destruction of the eastern NCC and the preservation of the western NCC. The western NCC has
11 higher lithospheric strength, and a relatively dry lithosphere with a cold thermal state. It experienced
12 localized deformation (e.g., the Yinchuan-Hetao and Shanxi–Shaanxi rifts), and these weak zones
13 kept the remaining parts of the western NCC stable. In contrast, the eastern NCC is more likely to
14 have uniform and widespread deformation due to its low lithospheric strength, hot thermal state and
15 water-rich lithosphere.

16 **Yilgarn craton**

17 The Yilgarn craton represents a similar situation to the western NCC, as it is surrounded by
18 three younger orogenic belts (Fig.3A). Surrounding orogens also have sheltered the craton from
19 deformation and therefore helped the Yilgarn craton to preserve the SCLM. The suturing events that
20 brought together the Yilgarn and Pilbara cratons to form the West Australian craton occurred during
21 ~2200-1620 Ma through multiple stages of orogenesis (Cawood and Tyler 2004) including the
22 ~2200 Ma Ophthalmian orogeny, the ~2000-1960 Ma Glenburgh orogeny, the ~1830-1780 Ma

Capricorn orogeny, and a unnamed event at the end of the Palaeoproterozoic (~1670-1620 Ma), reflecting that the weak mobile belts are the focus of stress. Hence, these orogens sheltered the Yilgarn craton from deformation. The Albany-Fraser orogeny also occurred in two stages (Pisarevsky et al. 2003; Clark et al. 2000; Bodorkos and Clark 2004) during ~1345-1140 Ma, which suffered most of the stress related to the convergent events. The Pinjarra orogen records (Fig.3A) a complex history involving multiple stages of reactivation from the Mesoproterozoic to early Cambrian (e.g., Fitzsimons 2003; Boger 2011). However, the Yilgarn craton has remained stable since its cratonization in the Archean. The different tectonic evolution between the Yilgarn craton and Pinjarra orogen show that the Pinjarra orogen accommodated the orogenic stresses through deformation owing to its relatively low lithospheric strength. Consequently, the deformation induced by intra-plate stresses of East Gondwana during the Permian and early Cretaceous would preferentially be concentrated in the Pinjarra rather than the adjacent craton (Hall et al. 2013). Therefore, these weak zones are significant to the stability of the Yilgarn craton.

Superior craton

The Superior craton has remained stable since the late Archean except for the formation of the Kapuskasing Structural Zone (KSZ; Fig.5B). Crustal shortening (Percival and Peterman 1994; Evans and Halls 2010) took place in this region, while the surrounding areas remained relatively stable. The integrated strength of the lithosphere beneath the KSZ is lower than in the surrounding regions (Perry et al. 2006), so the lower crust and upper mantle could be easily affected by the collisional events. This is the same as the western NCC and Yilgarn craton, the deformation associated with plate tectonics would have been focused at these weak regions, which is proven by the multi-stage tectonic activities in the surrounding orogens (Hoffman 1988; Thurston et al. 1992; Schulz and

Cannon 2007; Tohver et al. 2004, 2005). As a consequence, this contributes to the long-term stability of the Superior craton.

Lenardic et al. (2003) concluded that the degree of yield stress variation between mobile belts and cratons is found to be an effective and robust means for providing the longevity of the cratonic lithosphere. In fact, similar to the western North China craton, local modification or thinning of the SCLM and strong differences of crust-mantle structure between the nuclei of cratons and the surrounding mobile belts have also been observed in South Africa (Fouch et al. 2004), Europe (Gegersen et al. 2010), Siberia (Lesne et al. 2000), East Africa (Keranen and Klemperer 2008), and North America (Barruol et al. 1997; Yuan and Romanowicz 2010). Numerical simulation research as described above have already confirmed that the tectonically weak regions can protect cratons from deformation for long periods of geologic time (Lenardic et al. 2000, 2003; Lu et al. 2011; Yoshida 2010, 2012), which is an important factor for the long-term stability of cratons. In this paper we expounded this general phenomenon and pattern in the evolution of cratons using specific examples.

6. Conclusion

This overview of three cratons provides implications for craton destruction:

As most polarities of the subductions were toward the NCC, the subducted oceanic lithosphere would add water and hydroweaken the underlying mantle, which softened the lithosphere and made it easy for delamination or SCLM erosion to take place in the interior of the craton. Finally, the combination of interior lithospheric delamination or SCLM erosion and marginal lithospheric thinning caused by convergence between plates from multiple sides may have led to decratonization of the NCC. In the cases of the Yilgarn and Superior cratons, most subduction events were dipping

away from cratonic interiors, and the marginal lithospheric thinning area is limited. Furthermore, the weak zones around these two cratons provide advantages for their long-term stability which is the same as the western NCC. Collectively these factors lead to the phenomenon that some cratons lose their roots (e.g., the eastern NCC), while others retain them (e.g., the western NCC, Yilgarn and Superior cratons).

Based on the comprehensive comparison of geological, geophysical and geochemical data between cratons which have lost their roots and cratons that preserve them, we obtain the following preliminary conclusions:

(1) Subduction polarity has significant impacts on craton destruction;

The subduction polarity of subducting plates, specifically whether they dip under or away from the craton, controls the water released to the SCLM and underlying mantle wedge, which hydroweakens the SCLM. Thus, the lithosphere of the interior of the craton may become pre-weakened for delamination or SCLM erosion. Therefore subduction polarity has great effects on craton destruction.

(2) Convergence between plates can lead to marginal lithospheric thinning;

All three cratons analyzed here experienced SCLM thinning beneath their margins; evidence indicates that the thinning is most likely caused by convergence between plates. But craton destruction is not always accompanied by lithospheric thinning, except for cratons that suffered subduction and collision from multiple sides (e.g., the eastern NCC).

(3) Protection by weak zones around the craton.

As soon as a craton is surrounded by weak zones (e.g., mobile belts or orogens), the stresses associated with mantle convection are buffered and the craton is prevented from being further

1 deformed, which reflects a general phenomenon and pattern in the evolution of craton.

2 **Acknowledgements**

3 We express gratitude to Kevin Burke and John F. Dewey for their landmark papers in tectonics,
4 and for mentoring of the communicating author during his studies during the “golden years of
5 tectonics” at the State University of New York at Albany. Kevin and John not only made
6 fundamental contributions to understanding plate tectonics and the history of the Earth, but also
7 mentored a new generation of tectonicists, many of whom are leaders in the field today. This study
8 was financially supported by the National Natural Science Foundation of China (Nos. 91014002).

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Fig.1. (A) Schematic geological map of NCC and adjacent areas. (B) Simplified tectonic map of NCC (modified from Kusky et al. (2014a)). Numbers in rectangles are age information recorded by xenoliths (only presented data related with marginal lithospheric thinning), R means reworking. Xenoliths Data source: Zheng et al. 2004, 2005, 2012; Zheng and Wu 2009; Yu et al. 2007; Zhang et al. 2012; Liu et al. 2013. WNCC: Western part of the North China Craton (also known as the Ordos Block, e.g., OB); ENCC: Eastern part of the North China Craton; COB: Central Orogenic Belt; SSR, Shanxi–Shaanxi Rift; YHR, Yinchuan–Hetao Rift; NNCB, Northern North China Basin; SNCB, Southern North China Basin.

Fig.2. (A) Tomographic maps of S-wave velocities of the NCC (modified from Jiang et al. (2013)). The dashed lines mark the primary tectonic boundaries and the dotted lines mark two Cenozoic rift zones. (B) Velocity gradient along profile A-A', B-B', C-C' as showing in (A).

Fig.3. (A) Map showing location of the Yilgarn craton and adjacent orogens. NA, Northern Australia; SA, Southern Australia. (B) Regional map of tectonic terranes in the Yilgarn craton (modified from Nemchin and Pidgeon (1997)). LTR (Lithospheric thinning region) represents the area that might have experienced SCLM thinning as indicated by geophysical data (Fig.4B; Fishwick and Reading (2008)). EGP, Eastern Goldfields superterrane; BJ, Buljah lamprophyres; JK, Jewill kimberlites; NC, Northampton complex; MC Mullingar complex; LC, leeuwin Complex.

Fig.4. (A) Tomographic maps of S-wave velocities of the Yilgarn craton (modified from Fishwick and Reading (2008)). (B) Cross sections (blue line in (A)) showing the SCLM thinning area, dotted lines show the area that may experience SCLM thinning. Abbreviations used: Y, Yilgarn craton; C, Capricorn orogeny; P, Pilbara craton.

Fig.5. (A) Map showing location of the Superior and adjacent orogens. (B) Regional map of tectonic terranes in the Superior craton modified after Williams (1990), with kimberlite localities indicated. Abbreviations used: AC, Ashuanipi complex; AT, Abitibi terrane; BRT, Berens River terrane; BS, Bienville subprovince; DH, Douglas Harbour domain; ERT, English River terrane; EwT, East Wabigoon terrane; GD, Goudalie domain; ID, Inukjuak domain; KSZ, Kapuskasing Structural Zone; LG, La Grande subprovince; LM, Lake Minto domain; MRT, Minnesota River Valley terrane; ML, Munro Lake subprovince; NCT, North Caribou terrane; NSS, Northern Superior superterrane; PT, Pikwitonei terrane; QT, Quetico terrane; UD, Utsalik domain; WRT, Winnipeg River terrane; WT, Wawa terrane; WwT, West Wabigoon terrane. Numbers in rectangles are age information recorded by xenoliths, data source: Queen et al. (1996); Moser and Heaman (1997); Heaman et al. (2004); Sage (2000); Kaminsky et al. (2002).

Fig.6. Tomographic maps of S-wave velocities for depth from 100-250 km (modified from Frederiksen et al. (2007)). Green lines are boundaries of the Superior craton. The blue dotted line represents the interpreted boundary of lithosphere between western and eastern Superior craton. The red dotted lines are the Mid-continent rift. The yellow dotted line is a linear low-velocity anomaly related to the Great Meteor hot spot. FBR, failed branch of rift; MCR, Mid-continent rift.

Table 1. Summary of major post-cratonization geologic events in the North China craton, as described in detail in the text.

Time	Event	Subduction Polarity	Signature in SCLM	References
~1.9-1.85 Ga	Final amalgamation of NCC		Replacement of part of root?	Kusky and Li 2003 Wilde et al. 2002
~1.8 Ga	Post-collision extension		Mantle melting	Li et al. 2004 Kusky et al. 2007a
~700-250 Ma	Subduction from Dabie Shan	Toward the NCC	Hydration due to ingress of slab fluids?	Hacker et al. 2009 Kusky et al. 2007a
~500-250 Ma	Subduction from Solonker	Toward the NCC	Hydration due to ingress of slab fluids?	Davis et al. 1996 Xiao et al. 2003, 2012
~270-208 Ma	Indosinian orogeny		Shortening of N edge of SCLM	Kusky et al. 2007a
~200-100 Ma	Subduction under Pacific margin	Toward the NCC	Hydration due to ingress of slab fluids?	Zheng et al. 2009 Kusky et al. 2014
~50-0 Ma	Himalayan orogeny		Extrusion	Yin and Harrison 1996

Table 2. Summary of major post-cratonization geologic events in the Yilgarn craton, as described in detail in the text.

Time	Event	Subduction Polarity	References
~2.7-2.5 Ga	Final assembly of the Yilgarn craton		Huston et al. 2012 ; Myers 1993
~2.2-1.8 Ga	Capricorn orogeny	Toward the Yilgarn craton	Cawood and Tyler 2004
~1.79-1.78 Ga	Yapungku orogeny	Away from the Yilgarn craton	Bagas 2004 ; Giles et al. 2004
~1.3-1.0 Ga	Albany-Fraser orogeny	Away from the Yilgarn craton	Myers et al. 1996 ; Smithies et al. 2010
~1.21 Ga	Marnda Moorn igneous event		Wang et al. 2013
~522 Ma	Pinarra orogeny	Away from the Yilgarn craton	Collins 2003 ; Collins and Pisarevsky 2005
~120 Ma	Separated From Great Indian		Norvick and Smith 2001 ; Veevers 2006

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Table 3. Summary of major post-cratonization geologic events in the Superior craton, as described in detail in the text.

Time	Event	Subduction polarity	Reference
~3.0-2.6 Ga	Assembly of the Superior craton		Card and Poulsen 1998; Percival et al. 2006
~2.8-2.6 Ga	Kenoran orogeny	Away from the Superior craton	Percival et al. 2006; Thurston 1991
~2.5-2.0 Ga	Sedimentary basins spanning		Buchan and Ernst 2004
~2.0 Ga	Tectonic regime change		Hoffman 1988; Ernst et al. 1995, 2014
~1.9-1.6 Ga	Penokean orogeny	Away from the Superior craton	Schultz and Cannon 2007
~1.8 Ga	Trans-Hudson orogeny		Eaton and Darbyshire 2010
~1.1 Ga	Keweenawan Mid-continent rift		Thurston et al. 1991
~1.3-1.0 Ga	Grenville orogeny	Away from the Superior craton	Hoffman 1989; Tohver et al. 2005
~180-134 Ma	Emplacement of kimberlite along track of Great Meteor hotspot		Heaman and Kjarsgaard 2000; Heaman et al. 2004

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Fig.1. (A) Schematic geological map of NCC and adjacent areas. (B) Simplified tectonic map of NCC (modified from Kusky et al. (2014a)). Numbers in rectangles are age information recorded by xenoliths (only presented data related with marginal lithospheric thinning), R means reworking. Xenoliths Data source: Zheng et al. 2004, 2005, 2012; Zheng and Wu 2009; Yu et al. 2007; Zhang et al. 2012; Liu et al. 2013. WNCC: Western part of the North China Craton (also known as the Ordos Block, e.g., OB); ENCC: Eastern part of the North China Craton; COB: Central Orogenic Belt; SSR, Shanxi–Shaanxi Rift; YHR, Yinchuan–Hetao Rift; NNCB, Northern North China Basin; SNCB, Southern North China Basin.

Fig.2. (A) Tomographic maps of S-wave velocities of the NCC (modified from Jiang et al. (2013)). The dashed lines mark the primary tectonic boundaries and the dotted lines mark two Cenozoic rift zones. (B) Velocity gradient along profile A-A', B-B', C-C' as showing in (A).

Fig.3. (A) Map showing location of the Yilgarn craton and adjacent orogens. NA, Northern Australia; SA, Southern Australia. (B) Regional map of tectonic terranes in the Yilgarn craton (modified from Nemchin and Pidgeon (1997)). LTR (Lithospheric thinning region) represents the area that might have experienced SCLM thinning as indicated by geophysical data (Fig.4B; Fishwick and Reading (2008)). EGP, Eastern Goldfields superterrane; BJ, Buljah lamprophyres; JK, Jewell kimberlites; NC, Northampton complex; MC Mullingarra complex; LC, Leeuwin Complex.

Fig.4. (A) Tomographic maps of S-wave velocities of the Yilgarn craton (modified from Fishwick and Reading (2008)). (B) Cross sections (blue line in (A)) showing the SCLM thinning area, dotted lines show the area that may experience SCLM thinning. Abbreviations used: Y, Yilgarn craton; C, Capricorn orogeny; P, Pilbara craton.

Fig.5. (A) Map showing location of the Superior and adjacent orogens. (B) Regional map of tectonic terranes in the Superior craton modified after Williams (1990), with kimberlite localities indicated. Abbreviations used: AC, Ashuanipi complex; AT, Abitibi terrane; BRT, Berens River terrane; BS, Bienville subprovince; DH, Douglas Harbour domain; ERT, English River terrane; EwT, East Wabigoon terrane; GD, Goudalie domain; ID, Inukjuak domain; KST, Kapuskasing Structural Zone; LG, La Grande subprovince; LM, Lake Minto domain; MRT, Minnesota River Valley terrane; ML, Munro Lake subprovince; NCT, North Caribou terrane; NSS, Northern Superior superterrane; PT, Pikwitonei terrane; QT, Quetico terrane; UD, Utsalik domain; WRT, Winnipeg River terrane; WT, Wawa terrane; WwT, West Wabigoon terrane. Numbers in rectangles are age information recorded by xenoliths, data source: Queen et al. (1996); Moser and Heaman (1997); Heaman et al. (2004); Sage (2000); Kaminsky et al. (2002).

Fig.6. Tomographic maps of S-wave velocities for depth from 100-250 km (modified from Frederiksen et al. (2007)). Green lines are boundaries of the Superior craton. The blue dotted line represents the interpreted boundary of lithosphere between western and eastern Superior craton. The red dotted lines are the Mid-continent rift. The yellow dotted line is a linear low-velocity anomaly related to the Great Meteor hot spot. FBR, failed branch of rift; MCR, Mid-continent rift.