Contrasting mechanisms and timescales of subduction and exhumation as recorded by Paleoproterozoic and Late Paleozoic high-pressure granulites in the North China Craton

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ABSTRACT

The North China Craton records multiple metamorphic events related to supercontinent assembly during the Paleoproterozoic, forming Columbia, and again during the Late Paleozoic, forming Pangea. Here we show that the Paleoproterozoic high-pressure granulites (HPGs) formed from E-MORB protoliths and record a clockwise *P–T–t* path with prograde metamorphism at 7.8–10.0 kbar and 780–820 °C, peak granulite-facies metamorphism at 12–12.3 kbar and ~860–880 °C, and

retrograde metamorphism at 8.7–9.1 kbar and 850–855 °C. Subduction initiated prior to 1.90 Ga, with final collision and orogeny at 1.88 Ga, followed by post-collision/exhumation at 1.80 Ga, defining a prolonged exhumation period (~90 Myr) that occurred at a slow velocity of ~0.16 ± 0.08 mm/y. Late Paleozoic HPGs are N-MORB type and record a near clockwise *P–T–t* path, with peak/post-peak amphibolite-facies metamorphism at 11.0–12.5 kbar and 860–890 °C, isothermal decompression to 7.2–7.5 kbar and 810–820°C, and retrogression to 5.5–7.2 kbar and 805–850 °C. Subduction initiated earlier than c. 340 Ma, exhumation and uplift initiated at 335–309 Ma and continued to 297–287 Ma. The exhumation was short-lived (~50 Myr) and relatively fast (0.38 ± 0.14 mm/y). When compared to granulite-facies metamorphism documented in many Paleoproterozoic HPGs, Late Paleozoic HPGs appear to commonly form with an initial period of steep subduction leading to eclogite-facies metamorphism, with subsequent exhumation to middle/lower levels of the crust. Our results further reveal that the exhumation velocity for supercontinent collision was facilitated and duration shortened through time, and that the exhumation mechanism might have been controlled by subduction angle, compression pressure and temperature.

INTRODUCTION

High-pressure granulite (HPG) and high-pressure eclogite (HPE) form in distinct tectonic environments; the former being characteristic of lower continental crust that overthickens during accretionary or collisional orogeny, and the latter most commonly forming within deeply subducted oceanic crust (Guo et al., 2002; Weller and St-Onge,

2017; O’Brien, 2021). Both lithologies therefore act as tracers of convergent plate margin activity in the geological past, which itself can be used as a proxy for the operation of plate tectonics through time (e.g., O’Brien and Rötzler, 2003; Hernández-Uribe et al., 2020; Palin and Santosh, 2021; Palin, 2022). The geological record contains very few examples of HPE in Precambrian terranes, although HPG is notably more common throughout the Proterozoic and Archean, and thus provide a more robust proxy for tracing the evolution of geodynamic processes through time (Stosch and Lugmair, 1990; Zhao et al., 2001; Yu et al., 2021; Li, 2022). While the driving forces for subduction (and thus plate tectonics) are fairly well understood (e.g. Forsyth and Uyeda, 1975), those that facilitate exhumation are more ambiguous, and may involve combinations of factors, such as buoyancy of the subducted crust compared to surrounding mantle (Platt, 1993; Chapman et al., 2019), erosion of the overburden (Avellaneda-Jiménez et al., 2020), thrust-driven exhumation at shallower level (Sherlock et al., 1999; Guo et al., 2022), tectonic extension and exposure of lower crustal levels (Gao and Klemd, 2003; Froitzheim et al., 2003), decoupling of the low-density hydrous rocks that are retained at shallow depths and dense rocks that sink into the deep mantle (Chen et al., 2013), delamination of lithospheric mantle, and slab break off (Sanborn-Barrie et al., 2019). As HPGs are common in the geological record and necessarily document the movement of deep-seated crustal lithologies to the Earth’s surface, investigating the rates of exhumation and peak thermal conditions of HPGs of various ages can provide insight into which mechanisms have influenced the evolution of different terranes throughout geological time (Platt, 1993; Agard et

al., 2009).

The North China Craton (NCC) is a complex tectonic domain that has experienced a prolonged geological history. Popular tectonic models propose that it comprises at least three discrete microblocks divided by three major Paleoproterozoic metamorphic belts. The Western Block formed during amalgamation of the Yinshan and Ordos blocks at c. 1.95 Ga, forming the Khondalite belt along the Inner Mongolia Suture Zone (Zhao et al., 2005; Santosh, 2010; Zhai and Santosh, 2011; Wan et al., 2015); the Jiao-Liao-Ji belt formed in the Eastern Block at 1.85 Ga (Wan et al., 2006; Liu et al., 2017, 2019); and the Trans North China Orogen (TNCO) formed during welding of the Western and Eastern Blocks at 1.85 Ga (Zhao et al., 2000; Li et al., 2016; Tang and Santosh, 2018) (Fig. 1). Additionally, the Central Asian Orogenic Belt (CAOB) records accretion and collision of multiple arcs between the Siberian Craton and the NCC associated with the closure of the Paleo-Asian Ocean during the Late Paleozoic (Xiao et al., 2003, 2013; Safonova and Santosh, 2014; Yang et al., 2022).

The North Hebei region is located in the northern margin of the NCC, and documents cratonization during the Neoarchean, formation of the TNCO during the Paleoproterozoic, and construction of the CAOB during the Paleozoic (Liu et al., 2007). Retrogressed high-pressure metamorphic rocks with diverse ages reported from the region have been correlated with both Paleoproterozoic (c. 1.85-1.80 Ga) subduction and Late Paleozoic (>350 Ma) subduction (Ni et al., 2006; Zhang et al., 2020), although the pressure–temperature (*P–T*) history of Paleoproterozoic metamorphism is strongly debated, with some authors arguing that the units are not

retrogressed HPE, but instead HPG due to the absence of omphacite bearing mineral assemblage or ultra-high pressure UHP condition (e.g., Zhang et al., 2016, 2020; Liu and Zhang, 2021). In addition, Ni et al., (2006) suggest the retrograde eclogite formed during Paleozoic (~325 Ma). Thus, the metamorphic history of this region is complex, and the exhumation history remains uncertain.

Here, we present new petrological analyses, microstructural observations, and the results of zircon and monazite geochronology performed on HPGs from the North Hebei region, which reveals two episodes of HPG metamorphism: an older event that occurred during the Paleoproterozoic related to the assembly of the Columbia supercontinent, and a younger event that occurred during the Late Paleozoic related to the assembly of Pangea. Petrological and zircon trace element data reveal that the Paleoproterozoic HPGs only reached granulite facies, whereas the Late Paleozoic HPGs may represent pervasively recrystallized former eclogite-facies lithologies. Phase equilibria based modelling reveals both Paleoproterozoic and Late Paleozoic HPGs record slow (0.16 ± 0.08 mm/y) and prolonged (~90 Myr) uplift and exhumation during the Paleoproterozoic, but fast (0.38 ± 0.14 mm/y) and short-lived (~50 Myr) uplift and exhumation during the Late Paleozoic. Thus, we propose that exhumation rate was facilitated and duration shortened through time.

GEOLOGICAL BACKGROUND AND SAMPLES

The east-west trending Chongli–Chicheng–Gubeikou–Chengde–Pingquan (CCGCP) fault divides the North Hebei region into two terranes containing Jurassic volcano-sedimentary formations in the north, and Precambrian granulite-facies

metamorphic basement, Mesoproterozoic–Neoproterozoic amphibolite-facies rocks and Paleozoic unmetamorphosed cover in the south (Fig. 1) (Kong et al., 2011; Chu et al., 2013; Zhang et al., 2016; Liu and Zhang, 2021).

The Hongqiyingzi Complex is located north of the CCGCP fault, and exposes meta-supracrustal rocks, Late Paleozoic dioritic–quartzodioritic gneiss, Paleoproterozoic garnet–biotite–plagioclase gneiss, and minor Neoarchean metamorphic relics. The meta-supracrustal rocks are dominated by meta-leptynite, garnet meta-leptynite, garnet-bearing schist, garnet–biotite–plagioclase gneiss with or without graphite, garnet amphibolite, and diopside-bearing marbles (Wang et al., 2012). Retrogressed eclogite and serpentinized ultramafic rocks occur as lenses or blocks that are aligned with the layering in host garnet–biotite–plagioclase gneiss (Wang et al., 2012).

The mafic protolith of the Hongqiyingzi retrograded eclogite formed at 438 Ma, with peak eclogite-facies metamorphism dated at 355 Ma, granulite-facies overprinting at 341 Ma, and a final stage of amphibolite-facies metamorphism at c. 322–326 Ma (Kong et al., 2011). Chu et al. (2013) proposed that Paleo-Asian Ocean subduction had begun by at least at 355 Ma and interpreted ages of 320–300 Ma to record orogenic uplift, with the relatively younger ages of 290–280 Ma corresponding to a separate large-scale thermal event. Late Paleozoic c. 326–321 Ma amphibolite facies metamorphism is widely reported in this region (Ni et al., 2006). Detrital zircons analyzed from the retrograde eclogite yield ages between 389 Ma and 1989 Ma (Chu et al., 2013), consistent with the ages of 369–1855 Ma obtained from

associated garnet–biotite–plagioclase gneiss (Kong et al., 2011). Geochemical data reveal that the protolith of this retrograde eclogite was likely mid-ocean ridge basalt (MORB) or island arc tholeiite (IAT) type basalt (Ni et al., 2006). South of the CCGCP fault, the protolith of the granulite has an IAT-type basalt geochemistry, formed at c. 2550–2500 Ma, and HPG metamorphism occurred during the Paleoproterozoic at c. 1900 Ma (Zhang et al., 2016, 2020, 2021).

The Paleoproterozoic HPGs studied in this work occur in two exposures. One location (CH-3) along a large roadcut with rocky hillocks has thick bands and layers of garnet-bearing metagabbro with pegmatite veins. Some melt domains contain coarse garnet and pyroxene (Fig. 1). The second location (CH-6) exposes a sequence several hundred meters in thickness of garnet-bearing mafic granulite, and lenses and sheets of garnet-bearing trondhjemite. A rocky hillock track exposes a 300–500-m-thick sequence of mafic granulite with abundant, but variably sized garnet. Thick layers of amphibolite are also present.

The Late Paleozoic HPGs also occur in two main exposures in the Chicheng region. One location (CH-1) contains medium- to coarse-grained HPGs, which are foliated, and form dismembered blocks/boudins distributed within garnet-bearing metapelite (Fig. 1). Garnet porphyroblasts are mantled by plagioclase coronae and so have a “white-eye socket” texture. Medium-grained garnet bearing metapelite is foliated, layered and folded. Evidence for partial melting is also preserved by medium- to coarse-grained, and layer-parallel meta-trondhjemite veins. Ultramafic rocks, mainly including pyroxenite and dunite, occur to the west of the exposure.

Garnet-bearing metapelite is intercalated with mafic granulite and ultramafic rock, and so might represent an accretionary mélange. In a separate locality (CH-5), fragmented layers and blocks/boudins of mafic granulite with blocks of meta-trondhjemite occur within fine-grained, bedded and laminated sequences of metasediments. Thinly weathered and soily layers of volcanic tuff also occur. To the north of the exposure is migmatitic diorite gneiss that contains folded bands of quarzofeldspathic melts injected parallel to the metamorphic foliation. The dioritic gneiss is intruded by pink granite and contains boudins of amphibolite. Highly fragmented residual sheets of leucocratic meta-trondhjemite are also present. Thus, the amphibolite–diorite–granodiorite–meta-trondhjemite appears to be fractionated sequence of under-plated basaltic magma, and likely represents the lower portion of continental arc. Overall, the mafic granulite may represent deeply subducted oceanic crust, metamorphosed and exhumed along the subduction channel and brought up as exotic blocks in the accretionary sediment mélange.

Three HPGs and one garnet trondhjemite from south of the CCGCP fault (CH-3-1681, CH-3-2A, CH-6-2, CH-6-3B), and three HPGs and one meta-trondhjemite from north of CCGCP fault (CH-1-1B, CH-1-3, CH-1-5, CH-5-1A) were collected for petrological and geochemical analyses, and U–Pb dating of zircon and monazite. Detailed descriptions of each sample are given in APPENDIXES 1ta.

ANALYTICAL METHODS

Mineral compositions

Back-scatter electron (BSE) images were acquired using a TESCAN MIRA3 LMH Schottky field-emission scanning electron microscope (FE-SEM) housed at the Colorado School of Mines, USA. Major element compositional analyses of minerals in each sample were acquired using a JEOL JXA 8900 electron microprobe housed at the Denver Microbeam Laboratory at the United States Geological Survey (USGS), Colorado, USA. Operating conditions included an acceleration voltage of 15 kV, a beam current of 20 nA, a spot size of 1 μm. A ZAF matrix correction for atomic number, absorption, and fluorescence was automatically applied to all analyses. Multiple spot analyses and line profiles were taken across all major minerals in each sample to determine the extent of inter- and intra-crystalline compositional variation according to microstructural position. Mineral composition data are given in Table S1.

Bulk-rock geochemistry

Bulk-rock compositional data were obtained via X-ray fluorescence (XRF) and ICP-MS housed at the Testing Center of the First Geological Institute of the China Metallurgical Geology Bureau, with titration used to determine individual amounts of Fe_2O_3 and FeO. Whole-rock composition data are given in Table S2.

Petrological modelling

Petrological modeling was performed to constrain the *P–T* history of key units. All phase diagrams were constructed using Theriak-Domino (de Capitani and

193Petrakakis, 2010) and the internally consistent thermodynamic data set ds-62 (Holland
194and Powell, 2011) in the Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–O_2
195(NCKFMASHTO) system. The following activity–composition (a–x) relations for
196solid-solution phases were used: orthopyroxene garnet, biotite, muscovite, and
197chlorite (White et al., 2014); augitic clinopyroxene, clinoamphibole and melt (Green
198et al., 2016); epidote (Holland and Powell, 2011); plagioclase (Holland and Powell,
1992003); and ilmenite (Powell et al., 2014). Pure phases comprised kyanite, quartz,
200rutile, sphene, albite and H_2O. Uncertainties related to the absolute positions of
201assemblage field boundaries on calculated phase diagrams are typically around ±1
202kbar and ±50 °C at the 2σ level (Powell and Holland, 2008; Palin et al., 2016a; Lanari
203and Duesterhoeft, 2019), with this variation being largely a function of propagated
204uncertainty on end-member thermodynamic properties within the data set.

205 Mineral compositions were recalculated to standard numbers of oxygens per
206formula unit (p.f.u.), with H_2O assumed to be present in stoichiometric amounts. The
207proportion of ferric iron was estimated using the software AX (Holland, 2009).
208Pressure and temperature were modelled with the AX recalculated mineral
209composition data via THERMOCALC.

210**Zircon and monazite U–Pb geochronology**

211 Laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS)
212zircon U–Pb geochronology was performed at the State Key Laboratory of Geological
213Processes and Mineral Resources (GPMR), China University of Geosciences
214(Wuhan), following the procedures outlined by Liu et al. (2010). Concordia diagrams

and weighted mean calculations were made using Isoplot/Ex_ver3 (Ludwig, 2003). Zircon U–Pb age data are given in Table S4.

U–Pb dating of monazite was also conducted by LA-ICP-MS at the Wuhan Sample Solution Analytical Technology Co., Ltd., Wuhan, China. Analytical conditions and procedures followed Zong et al., (2015). Isotopic data for monazite in all samples are given in Table S5.

RESULTS

Petrography and mineral chemistry

Paleoproterozoic HPGs and garnet trondhjemite

Sample CH-3-1 contains euhedral to subhedral garnet, clinopyroxene, and minor plagioclase, with randomly distributed needles of amphibole and ilmenite. All minerals are fine to medium grained (<1 mm) and form an equigranular texture, indicative of textural equilibrium. Garnet exhibits no significant compositional zoning from core to rim, having a composition $Alm_{56-58}Sps_{1-2}Prp_{16-21}Grs_{21-24}$. Plagioclase that surrounds garnet has a lower XCa = Ca/(Ca + Na + K) = 0.03 (Table S1) than grains in the matrix XCa = 0.42–0.45. Amphibole has Si = 6.47-6.70 c.p.f.u., Na = 0.50-0.57 c.p.f.u., Ti = 0.12-0.15 c.p.f.u., and XMg [= atomic Mg/(Mg + Fe)] = 0.59-0.63 and is thus magnesiohornblende (Fig. S1) (Leake et al., 1997). Clinopyroxene has Si = 1.94-1.99 c.p.f.u., Al = 0.08-0.14 c.p.f.u., Ca = 0.88-0.90 c.p.f.u., XMg = 0.68-0.70, and is augite (Morimoto et al., 1988).

Sample CH-3-2A is an amphibole-rich HPG containing coarse-grained garnet, amphibole, clinopyroxene, and plagioclase-quartz segregations. Minor and accessory

minerals include zircon, ilmenite and rutile. Large (<5 mm) poikiloblastic garnet contains inclusions of plagioclase, quartz, amphibole, ilmenite, and rutile, and have distinct compositional zoning, with almandine- and pyrope-rich cores and grossular-rich rims (Fig. 2). Garnet poikiloblasts are also surrounded by symplectites of intergrown clinopyroxene, plagioclase, and hornblende. Plagioclase-quartz segregations that occur as interlayers between aligned amphibole aggregations are interpreted as textural evidence of partial melting (Menzies, 1973). Plagioclase inclusions in garnet have lower XCa = 0.30-0.32 than grains in the symplectite (XCa = 0.34-0.47). High-Ti (0.11-0.26 c.p.f.u.) greenish to brownish amphibole inclusions have similar compositions to matrix grains (Si = 6.35-6.64 c.p.f.u., Na = 0.40-0.49 c.p.f.u., and high XMg = 0.60-0.73) and are classified as magnesiohastingite and/or magnesiohornblende (Fig. S1). Clinopyroxene has Si =1.92-1.97 c.p.f.u., Al = 0.07-0.13 c.p.f.u., Ca = 0.86-0.92 c.p.f.u., and XMg = 0.74-0.84, and thus is diopside/augite.

Sample CH-6-3B is a medium to fine grained HPG that contains clinopyroxene, amphibole, garnet, minor plagioclase, and exhibits an equigranular texture (Fig. 2). Accessory zircon and ilmenite are also present. Deformed amphibole aggregates wrap around garnet. Porphyroblastic garnet contains clinopyroxene, quartz, and ilmenite inclusions, and retains weak compositional variation from core to rim ($Alm_{61-63}Sps_1Prp_{15-16}Grs_{20-22}$). This garnet rim is bordered either by hornblende + plagioclase + clinopyroxene symplectitic or hornblende coronae, which formed during retrograde metamorphism. Matrix plagioclase has low XCa = 0.07-0.15, XNa = 0.79-0.89, and

so is albitic, and may have formed due to garnet growth at the expense of high XCa plagioclase and biotite (Stowell and Stein, 2005), and high XCa = 0.34, XNa = 0.66 andesine may have formed during peak metamorphism. Amphiboles have compositions of Si = 6.49-6.56 c.p.f.u., Na = 0.62-0.67 c.p.f.u., low XMg = 0.50-0.53, and high Ti = 0.19-0.25 c.p.f.u. (Fig. S1). Clinopyroxene inclusions and matrix has similar composition Si = 1.96-1.99 c.p.f.u., Al = 0.06-0.09 c.p.f.u., Ca = 0.79-0.94 c.p.f.u., and XMg = 0.49-0.65.

Sample CH-6-2 is a garnet-trondhjemite leucosome containing medium to fine grained elongated feldspar and quartz, that wrap around relict garnet.

Late Paleozoic HPGs and meta-trondhjemite

Sample CH-1-1B is a foliated garnet amphibolite containing garnet, plagioclase, and amphiboles (Fig. 2). Accessory zircon, ilmenite, and rutile also occur. Foliated amphiboles define the rock fabric. Porphyroblasts of garnet (<5 mm) are highly fractured and contain amphibole, rutile, and ilmenite inclusions at its rims, and are surrounded by plagioclase + hornblende symplectite. Partial melting is indicated by trondhjemitic, layer-parallel bands that exhibit clear dihedral angles between plagioclase and quartz (Holness et al., 2011). Quartz veins (<3 mm) are randomly located among the amphibole aggregates. Garnet preserves compositional zoning, with grossular (Grs_{21-16}), spessartine (Sps_{4-2}) and almandine (Alm_{58-55}) decreasing and pyrope (Prp_{17-27}) increasing from core to rim; such compositional zoning is characteristic of prograde metamorphism (Woodsworth, 1977). Amphibole has Si = 5.97-7.08 c.p.f.u., Na = 0.25-0.53 c.p.f.u., and XMg = 0.56-0.73 and is classified as

281magnesiohornblende and tschermakite. Plagioclase grains in coronae have XCa = 2820.36-0.51 (Fig. S1). Amphibole surrounding ilmenite is brownish and has high Ti = 2830.08-0.14 c.p.f.u. than greenish matrix amphibole Ti = 0.03 c.p.f.u., indicates ilmenite 284decompression and titanium exsolution. Ilmenite exhibits bimodal grain sizes with 285larger grains are surrounded by myrmekite and small grains.

286 Sample CH-5-1A is a garnet amphibolite containing medium to fine grained 287amphibole and plagioclase that wraps euhedral to subhedral garnet (Fig. 2). Accessory 288minerals include zircon, ilmenite and rutile. Garnet contains quartz inclusions and 289exhibits weak compositional zoning (Alm_{58-63}, Sps_{4-9}, Prp_{6-10}, Grs_{24-29}). Plagioclase in 290coronae has XCa = 0.41-0.47. Amphibole inclusions in garnet and matrix grains have 291similar compositions (Si = 6.37-6.66 c.p.f.u., Na = 0.41-0.44 c.p.f.u., and XMg = 2920.41-0.48) and are classified as ferropargasite (Fig. S1).

293 Sample CH-1-3 is an amphibolite that contains medium to fine grained foliated 294amphiboles, plagioclase and quartz. Pseudomorphs after garnet are represented by 295fine-grained intergrowths of plagioclase and amphibole.

296 Sample CH-1-5 is a coarse grained meta-trondhjemite dominantly comprised of 297elongated and foliated plagioclase, K-feldspar, quartz, and minor muscovite.

298**Whole-rock geochemistry**

299 The studied Paleoproterozoic HPGs have bulk-rock compositions straddling the 300picrobasalt/basalt fields on a total-alkali–silica (TAS) diagram, indicating an igneous 301protolith with an ultramafic/mafic composition (Fig. 3A). On a Nb/Y vs. Zr/TiO_2 plot, 302all HPGs straddle the fields for andesite to subalkaline basalt (Fig. 3B). Transitional

trace elements vary in concentration, with Ni = 44–131 ppm and Cr = 15–313 ppm, which are lower than primitive mantle values (Ni > 400 ppm, Cr > 800 ppm). Large ion lithophile elements (LILE) have variable Sr (31–215 ppm), Ba (14–200 ppm), and Rb (1.3–19.1 ppm) contents, and high field strength elements (HFSE) are depleted (Nb = 4.2–21 ppm, Ta = 0.36–1.26 ppm, Zr = 49–116 ppm, and Hf = 1.36–1.83 ppm). These geochemical features show negative Sr, Zr anomalies on a primitive mantle-normalized trace element diagram, which suggests LILE enrichment and relative HFSE depletion (Fig. 3). Chondrite-normalized REE patterns are similar to those for E-MORB, and $(La/Sm)_N$ = 0.6–1.4, $(Ce/Yb)_N$ = 1.0–1.8, and $(Gd/Yb)_N$ = 1.2–1.6 values indicate relative LREE and MREE enrichment over HREE (Fig. 3).

Late Paleozoic HPGs have bulk-rock compositions equivalent to basalt/basaltic andesite on a TAS diagram (Fig. 3A, B). Transitional trace elements have variable concentrations (Ni = 56–92 ppm, Cr = 84–214 ppm); LILEs have variable concentrations (Sr = 32–138 ppm, Ba = 24–118 ppm, and Rb = 2.4–28.7 ppm); and HFSEs have relatively low concentrations (Nb = 0.78–5.9 ppm, Ta = 0.12–0.38 ppm, Zr = 37.3–167 ppm, and Hf = 1.08–1.98 ppm). Chondrite-normalized REE patterns are similar to N-MORB and E-MORB, and $(La/Sm)_N$ = 0.6–1.7, $(Ce/Yb)_N$ = 0.3–0.8, $(Gd/Yb)_N$ = 0.4–1.1 values indicate slight HREE enrichment over LREE and MREE (Fig. 3E). Primitive mantle-normalized trace element diagrams show that these HPGs are characterized by negative Nb and variable La, Ce anomalies, analogous to profiles exhibited by E-MORB/N-MORB (Fig. 3F).

Phase equilibrium modelling

Calculated *P–T* pseudosections for three Paleoproterozoic and three Late Paleozoic HPGs are shown in Figs. 4 and 5. A bold dark grey line marks the low-grade stability limit of garnet, a bold white line marking the position low-temperature stability limit of hornblende, a black line marking the high-grade stability limit of rutile and the solidus, and an light grey bold line represents the THERMOCALC calculate average *P–T* (avPT) mode for each sample (Powell and Holland, 1994). The dark grey colored assemblage fields represent observed equilibria mineral assemblage in each sample to constrain the *P–T* condition. Expected molar volume proportion changes versus temperature (750–1000 °C) and pressure (0.5–1.5 kbar) are shown in Figure 6.

Paleoproterozoic HPGs

The observed mineral assemblage of Grt–Hbl–Aug–Ilm–Qtz in sample CH-3-1 was calculated to be stable at *P–T* conditions of 6.9–11.2 kbar and <840 °C (Fig. 4A). This field overlaps with the modelled average *P–T* conditions of 9.2–11.0 kbar and <820 °C for the paragenesis Hbl–Cpx–Grt–Pl–Ilm (Fig. 4A). Calculated mineral isopleths and modal proportions indicate that the observed minerals are dominated by augite (~17–23 vol. %), hornblende (~57 vol. %), garnet (~10–14 vol. %), melt (~7 vol. %) at 7.8–10.0 kbar and 780–820 °C (Figs. 4A, 6A). This assemblage transforms from an augite + hornblende dominated assemblage to garnet + augite dominated assemblage at 7 kbar and 800 °C. Quartz is stable up to 8.4 kbar and 835 °C,

plagioclase is completely consumed at 6.4 kbar and 785 °C. Together, these modelling results indicate peak *P–T* conditions of around 7.8–10.0 kbar and 780–820 °C (Fig. 4A).

A pseudosection for sample CH-3-2A is shown in Figure 4B, and the observed assemblage Grt–Pl–Hbl–Aug–Qtz–Ru plus partial melt is stable at >10.5 kbar and 850–890 °C. These conditions overlap with avPT conditions for the paragenesis Hbl–Cpx–Grt–Pl–Ilm of 11.9–12.3 kbar and 850–880 °C. Calculated mineral isopleths and modal proportions of the observed minerals are likely to have constituted garnet (~20–25 vol. %), hornblende (~18–32 vol. %), augite (~19–23 vol. %), plagioclase (~15 vol. %), melt (~20 vol. %) at 12–12.3 kbar and ~860–935 °C (Figs. 4B, 6B). Rutile inclusions in garnet indicate peak metamorphism at >11 kbar and 900 °C. Consumption of garnet would increase the proportion of hornblende and plagioclase-quartz segregation in accordance with the petrological observation, which records the transition from peak/post-peak to retrograde metamorphism (Figs. 1D, 6B). Hence, peak granulite facies metamorphic *P–T* conditions are constrained at 12–12.3 kbar and ~860–880 °C (Fig. 4B).

A pseudosection for sample CH-6-3B is shown in Figure 4C and the observed assemblage Hbl–Grt–Aug–Ilm–Qtz is stable at *P–T* conditions of 6–12.7 kbar and 800–880 °C. This field overlaps with calculated avPT conditions of 9.8–10.5 kbar, 800–855 °C. Calculated mineral isopleths and modal proportions of the observed mineral assemblage contains hornblende (~10–12 vol. %), garnet (~38–40 vol. %), augite (~35 vol. %) and melt (~7 vol. %) and stabilized at 8.7–9.1 kbar and 850–855

367°C (Figs. 4C, 6C). Exhumation of the HPG would transform a garnet + augite (>85
368vol. %) dominated assemblage to a garnet + hornblende + augite (>90 vol. %)
369dominated assemblage (Fig. 6C). Garnet breakdown would produce hornblende and
370quartz between 9.8–5.4 kbar and 870–760 °C. Abundant plagioclase would form at
371<5.4 kbar and 760 °C during retrograde metamorphism (Figs. 2, 4C). Thus, the post-
372peak metamorphism can be constrained at 8.7–9.1 kbar and 850–855 °C (Fig. 4C).

373*Late Paleozoic HPGs*

374 A pseudosection for sample CH-1-1B is shown in Figure 5A, where the observed
375assemblage Hbl–Grt–Pl–Qtz–Ru is stable at 11.0–12.5 kbar and 860–890 °C.
376Calculated mineral isopleths and modal proportions of phases in this sample are
377garnet (~28–33 vol. %), hornblende (~25–27 vol. %), quartz and plagioclase (~22 vol.
378%), melt (~15 vol. %) at 10.6–12.5 kbar and 860–890 °C (Figs. 5A, 6D). Rutile
379inclusions in garnet porphyroblast rims indicate peak metamorphism at conditions of
380>10.6 kbar and 890 °C. Consumption of garnet during decompression would lead to
381additional hornblende and plagioclase growth at lower grade, which is consistent with
382the observed microstructures (Fig. 6D). Plagioclase is stable at <10.7 kbar and 895
383°C, and so plagioclase coronae around garnet document decompression from peak
384conditions. Ilmenite stabilizes below 10.0 kbar and 875 °C, and its occurrence in
385myrmekite microstructures again confirms decompression at high temperature during
386retrograde metamorphism. No clinopyroxene was observed in the Late Paleozoic
387HPGs; however, the calculated proportion of augite is ~12 vol. %, that can be
388accounted for the transformation of the eclogite to garnet amphibolite (Ni et al.,

3892006). Thus, the metamorphic condition can be constrained at 11.0–12.5 kbar and 390860–890 °C (Fig. 5A).

391 A calculated pseudosection for sample CH-5-1A is shown in Figure 5B and the 392observed assemblage Hbl–Grt–Pl–Ilm–Qtz is stable at 7.2–11.2 kbar and 820–850 °C. 393Calculated mineral isopleths and phase proportions of the observed minerals contain 394garnet (~5 vol. %), hornblende (~34 vol. %), plagioclase (~12 vol. %), quartz (~6 vol. 395%), and melt (~15 vol. %) at 7.5 kbar and 810 °C (Figs. 5B, 6E). Garnet is calculated 396to break down to produce hornblende at >7.4 kbar and 810 °C (Fig. 6E). Coeval with 397the absence of rutile inclusions and the abundance of plagioclase coronae suggests 398retrograde metamorphism at <7.5 kbar and 815 °C. Combining the minerals isopleth 399with modelled proportion, the metamorphic condition can be constraint at 7.2–7.5 400kbar and 810–820°C.

401 A calculated pseudosection for sample CH-1-3 is shown in Figure 5C, where the 402observed retrograde assemblage Hbl–Pl–Ilm–Qtz is stable above the solidus at 5.5–4039.0 kbar and 805–850 °C. Mineral isopleth and modal proportion calculation of the 404observed minerals contain quartz (~ 5 vol. %), plagioclase (~23 vol. %), hornblende 405(~55–60 vol. %), melt (~12 vol. %) at 7.2 kbar and 805 °C (Fig. 6F). The onset of 406exhumation of the HPG is recorded by a major increase in the proportion of 407hornblende and plagioclase (>80 vol. %) during decompression. Garnet is 408pseudomorphed by these minerals at conditions below 8.6 kbar and 840 °C, recording 409the transition from granulite to amphibolite facies metamorphism (Fig. 6F). Thus, 410combining the minerals isopleth with modelled proportion, the metamorphic

conditions of this stage can be constrained at 5.5–7.2 kbar and 805–850 °C.

Zircon U–Pb geochronology

Six HPGs and two meta-trondhjemite samples were dated via zircon U–Pb geochronology, with all data given in Table S4 and shown in Figure 7. Monazite from two meta-trondhjemite samples were also analyzed, with U–Pb age data given in Table S5 and shown in Figure 7. Zircon and monazite trace element data are given in Table S6 and plotted in Figures S2-S4.

Paleoproterozoic HPGs and garnet-trondhjemite

Zircon grains in sample CH-3-1 are subhedral to anhedral in shape, up to 200 μm long and can be divided into two groups (Fig. 7A). Group 1 represents magmatic zircons with patchy zoning in cores surrounded by unzoned metamorphic rims, and Group 2 grains are metamorphic with patchy zoning or homogeneous unzoned. Thirty-two zircon spots were obtained from thirty grains, although three were excluded due to high degrees of discordance. The remaining 29 spots fall along a discordia line with an upper intercept age of 2501 ± 93 Ma and lower intercept age of 1797 ± 68 Ma (MSWD = 1.7, n = 29). Among the metamorphic grains, a single population produced a weighted mean $^{207}Pb/^{206}Pb$ age of 1904 ± 24 Ma (MSWD = 0.42, n = 20). Both magmatic and metamorphic zircons are characterized by LREE depletion, positive Ce and Sm anomalies, and negative Pr-Nd and Eu anomalies (Fig. S2A). However, the metamorphic zircons have relative LREE enrichment, and variable HREE depletion.

Zircon grains from sample CH-3-2A have metamorphic characteristics, and are subhedral to anhedral in shape, up to 200 μm long, and display patchy zoning or sector zoning. Thirty-four zircon spots were analyzed. After excluding one discordant spot, the remaining 33 spots yielded a weighted mean $^{207}Pb/^{206}Pb$ age of 1887 ± 15 Ma (MSWD = 0.52, n = 33) (Fig. 7B). These zircons are characterized by LREE depletion, prominent negative Pr-Nd, Eu anomalies, and positive Ce and Sm anomalies (Fig. S2B).

Zircon grains from garnet-trondhjemite sample CH-6-2 are prismatic or stumpy in shape and up to 200 μm in length. Most of the grains show core–rim structures and can be divided into oscillatory zoned magmatic zircon (Group 1), metamorphic zircon overprints on the magmatic grains (Group 2), and metamorphic overgrowths (Group 3) (Fig. 7C). Thirty-four spots were analyzed, of which two discordant spots were excluded. The remaining 32 spots define one magmatic and two metamorphic age populations. Group 1 produced a weighted mean $^{207}Pb/^{206}Pb$ age of 2554 ± 41 Ma (MSWD = 2.5, n = 12) magmatism. Group 2 produced a weighted mean $^{207}Pb/^{206}Pb$ age of 2501 ± 24 Ma (MSWD = 1.4, n = 15), and Group 3 yielded a weighted mean $^{207}Pb/^{206}Pb$ age of 2437 ± 33 Ma (MSWD = 0.23, n = 5) for metamorphism. Both magmatic and metamorphic zircons are characterized by LREE depletion, positive Ce anomalies, and negative Pr-Nd anomalies. However, the metamorphic zircons display strong HREE enrichment, variable LREE enrichment, and positive Sm and negative Eu anomalies (Fig. S2C).

Late Paleozoic HPGs and meta-trondhjemite

Metamorphic zircon grains from sample CH-1-1B are euhedral to subhedral in shape, and up to 250 μm in length. Core–rim structures are common, and the grains can be divided into three groups: Group 1 zircons are characterized by weakly luminescent cores, Group 2 zircons are characterized by relatively highly luminescent cores, and Group 3 zircons show clear metamorphic resetting in the core, whereby grains show weak patchy or striped zoning, or are unzoned entirely (Fig. 7D). Thirty-three zircon spots were analyzed and excluding eight highly discordant spots, the remaining 23 spots can be divided into three groups. Group 1 yielded a weighted mean $^{206}Pb/^{238}U$ age of 335 ± 10 Ma (MSWD = 0.22, n = 5), Group 2 yield weighted mean $^{206}Pb/^{238}U$ age of 309 ± 4 Ma (MSWD = 0.79, n = 11), Group 3 yield lower intercept age of 287 ± 13 Ma (MSWD = 0.56, n = 9). All grains are characterized by LREE depletion and MREE enrichment, whereas the metamorphic zircons in Group 3 are characterized by relatively low MREE and HREE contents (Figs. S3A-C).

The metamorphic zircon grains in sample CH-5-1A have a euhedral to subhedral morphology and up to 150 μm in length. Most of the grains display core–rim textures and can be divided into two groups. Metamorphic cores in Group 1 grains show weak oscillatory zoning, whereas cores in Group 2 are homogenous (Fig. 7E). Thirty-three zircon spots can be divided into two groups based on their textures. Group 1 yielded a weighted mean $^{206}Pb/^{238}U$ age of 328 ± 2.3 Ma (MSWD = 1.1, n = 19), Group 2 yield lower intercept $^{206}Pb/^{238}U$ age of 314 ± 4 Ma (MSWD = 1.8, n = 14). The zircons are characterized by LREE depletion and slight MREE enrichment, and negative Pr-Nd

and positive Ce anomalies. Nine of the Group 2 zircons – defined here as Group 3 – display relatively low MREE and HREE contents (Figs. S3D-F).

Metamorphic zircon grains in sample CH-1-5 have a prismatic to stumpy morphology and are up to 240 μm in length. These grains can be divided into three groups, Group 1 is characterized by core-rim textures and are strongly luminescent, Group 2 and 3 are weakly luminescent and have patchy or sector zoning (Fig. 7F). Thirty zircon spots from sample CH-1-5 were analyzed, and excluding one discordant spot, the remaining 29 spots clustered into three populations. Group 1 yield a weighted mean $^{206}Pb/^{238}U$ age of 341 ± 3 Ma (MSWD = 0.51, n = 4), Group 2 yields a $^{206}Pb/^{238}U$ age of 330 ± 3 Ma (MSWD = 0.18, n = 4). Group 3 yielded a weighted mean $^{206}Pb/^{238}U$ age of 297 ± 1.2 Ma (MSWD = 0.76, n = 17/21). Zircons from Group 1 and 2 are characterized by low REE and HREE contents, and positive Ce and negative Pr-Nd anomalies. However, Group 3 grains have high REE contents, with strong LREE and HREE enrichment, and prominent positive Ce, Sm and negative Pr-Nd, Eu anomalies (Fig. S2D).

Monazite U–Pb geochronology

Monazite grains in sample CH-6-2 are angular or round, and up to 170 μm in length. Most of the grains are homogenous in CL, although some show core–rim internal structures. Thirty monazite grains were analyzed; excluding three discordant spots, the remaining 27 spots define a discordia upper intercept age of 2500 ± 78 Ma and lower intercept age of 1890 ± 35 Ma (MSWD = 2.5, n = 27) (Fig. 8A). The monazite grains in sample CH-6-2 are characterized by strong HREE depletion and

prominent negative Eu anomalies (Fig. S4A).

Monazite grains in sample CH-1-5 are anhedral or round, with diameters of 100–220 μm. Most grains show core-rim structures or are unzoned. Thirty monazite grains were selected for U–Pb analysis, and all data plot along concordia and have a weighted mean $^{206}Pb/^{238}U$ age of 292 ± 1.2 Ma (MSWD = 1.7, n = 30) (Fig. 8B). Those monazite grains are mostly characterized by a strong HREE depletion and prominent negative Eu anomalies (Fig. S4B).

DISCUSSION

Petrogenesis and mantle source characters

The lithological assemblages of the Paleoproterozoic garnet–biotite–plagioclase gneiss, garnet-bearing mafic granulite, garnet-bearing trondhjemite, amphibolite and Paleozoic diorite–granodiorite–amphibolite– meta-trondhjemite suggest the protolith of both HPGs are tholeiitic basalt initial from the mid ocean ridge tranverse to the island arc, and metamorphosed during the subsequent subduction process. Geochemically, the Paleoproterozoic and Paleozoic HPGs mostly exhibit low silica and subalkaline, slightly enriched in LILE (Rb, Ba), and depletion in (Nb, Ta), suggest the depleted mantle was modified by slab dehydration fluids in a subduction environment. The REE contents in the studied Chicheng samples indicate that the precursor magmas of the Paleoproterozoic HPGs originated from an enriched mantle source, whereas the Late Paleozoic HPGs protoliths more closely resemble MOR-derived basalts. Immobile trace element (Dy, Yb, Ti, Y, Nb) ratios have also been used

to constrain the depth of partial melting (Furnes and Safonova, 2019; Pearce, 2008). Most HREE strongly fractionate into garnet during metamorphism and anatexis, and Yb has a stronger affinity with garnet over spinel (Jung et al., 2006). In a La/Yb vs Dy/Yb diagram, all HPG samples cluster between the garnet-peridotite and spinel-peridotite fields, indicating that their precursor mafic magmas were either mixed or formed at a transitional depth in the mantle where both garnet and spinel were stable. For a Proterozoic mantle potential temperature of ~1400–1500 °C (e.g., Herzberg et al., 2010), this transitional domain occurs at around 40–75 km depth at subsolidus conditions (Jennings and Holland, 2015; Weller et al., 2019). The Paleoproterozoic HPGs exhibit relatively high Dy/Yb ratios compared to Late Paleozoic HPGs, which implies partial melting from a relatively deep mantle source (garnet peridotite) for the former and a relatively shallow mantle source (spinel peridotite) for the latter (Fig. 9A) (Thirlwall et al., 1994; Jung et al., 2006). Elevated Nb/Y and Rb/Y ratios ~~and the low garnet content~~ of the Paleoproterozoic HPGs ~~also~~ suggests the direct involvement of hydrous fluids and melt in the mantle source region and deeper depth of partial melting ~~rather than anhydrous slab melting~~, as ~~garnet would otherwise remain in the slab as a residual phase~~ melts dehydration at greater depth than fluids (Fig. 9B) (Kepezhinskas et al., 1997). High ~~Nb/Y~~ Rb/Y ratios of the Paleozoic HPGs reflect fluid-mediated amphibolitization during retrograde metamorphism and shallow depth of partial melting (Kepezhinskas et al., 1997).

We thus interpret that both Paleoproterozoic and Late Paleozoic HPGs are sourced from low-Sr/Y calc-alkaline island arc basalt, with the decreasing Y content

and retention of garnet and hornblende in the mantle source corresponding to an increasing metamorphic grade (Fig. 9C) (Drummond and Defant, 1990). However, partial melting and crystallization of plagioclase enriches the source in Y and depletes it in Sr from peak to retrograde metamorphism (Defant and Drummond, 1990). Most of the studied samples straddle a mantle array and arc-type field in a Th/Yb vs. Nb/Yb diagram, indicating the Paleoproterozoic HPGs formed from subduction related-basalt derived from a relatively enriched mantle source (Fig. 9D) (Pearce et al., 2008). Ti, Nb and Yb are immobile elements in subduction zone environments, such that a TiO_2/Yb vs. Nb/Yb plot is useful approach to model the depth of mantle melting (Pearce et al., 2008). In this plot, the studied HPGs entirely cluster in the MORB array, and so suggest a shallow source depth (spinel peridotite facies). Figure 8E also shows an enriched mantle source for the Paleoproterozoic HPGs, alongside Th_N vs. Nb_N data in Figure 8F, whereas the Late Paleozoic HPGs span the N-MORB and E-MORB fields. Few samples plot away from the N-MORB and E-MORB field account for partial melting and extraction of plagioclase during retrograde metamorphism (Fig. 9F) (Saccani, 2015).

Studied Paleoproterozoic HPGs have a relatively higher MgO content (average = 7.8 wt. %) than Late Paleozoic HPGs (average = 6.91 wt. %) (White and Klein, 2014). Compatible trace elements, such as Cr, Co, and Ni, are more enriched in the Paleoproterozoic HPGs (Cr: 290 ppm, Co: 63 ppm, Ni: 131 ppm) than the Late Paleozoic HPGs (Cr: 113–214 ppm, Co: 50–58 ppm, Ni: 74–92 ppm). These signatures are analogous to comparisons between Archean and Phanerozoic

(meta)basalt, given that MORB is known to have been more mafic in the geological past, due to higher mantle potential temperatures (Condie et al., 2016). As the protoliths of the Paleoproterozoic and Late Paleozoic granulites are MORB-type basalt, this supports the interpretation given here that they experienced subduction, incorporation into the deep continental crust, and were later exhumed into lower–middle crust.

***P–T–t* constraints on exhumation**

The Chicheng Paleoproterozoic HPGs preserve a well-defined clockwise *P–T–t* path, although constraints on the conditions of prograde or peak metamorphism in the Late Paleozoic HPGs are lacking (Fig. 10). Many studies have reported that the Late Paleozoic Paleo-Asian Ocean subduction-related mafic granulites reached eclogite facies (Ni et al., 2006; Chu et al., 2013; Kong et al., 2011). However, a recent study by Zhang et al. (2020) proposed that eclogite-facies metamorphism relates to the Paleoproterozoic orogenic event in the TNCO. Since eclogite-facies metamorphism is only interpreted from REE profiles of zircon in the Late Paleozoic HPGs of this study (Fig. S3), we favor the former hypothesis, and suggest that more work is required to constrain the timing of peak metamorphism in units from the north margin of the NCC. Nonetheless, both Paleoproterozoic and Late Paleozoic HPGs record common granulite-amphibolite facies conditions that document geotherm changes during burial and exhumation.

Petrological, microstructural, and mineral compositional information shows that the Paleoproterozoic E-MORB-type HPGs experienced prograde metamorphism at

5847.8–10.0 kbar and 780–820 °C, and inclusions of rutile suggest peak metamorphic 585conditions exceeding 12–12.3 kbar and ~860–880 °C (Fig. 10A). This was followed 586by isothermal decompression at 8.7–9.1 kbar and 850–855 °C, consuming garnet to 587form plagioclase–hornblende–clinopyroxene coronae (Fig. 10A). The peak 588metamorphism equates to a geothermal gradient of ~21 °C/km, which is typical for 589HPG metamorphism (Brown, 2007), and the pressure changes from peak to retrograde 590metamorphism suggest the HPGs exhumed at least 14 km from a 40 km to 26 km 591depth (Fig. 10A). The clockwise *P–T–t* path for the Paleoproterozoic mafic HPGs is 592broadly consistent with previous studies in the Xuanhua Complex, Fuping Complex, 593Lüliang Complex, Wutai Complex, and Hengshan Complex of TNCO and suggests a 594subduction-collision-exhumation history, which interprets it as an orogenic belt that is 595correlated to the Columbia supercontinent amalgamation (Fig. 1A) (Table S7) (Zhao 596et al., 2000; Guo et al., 2002; O'Brien et al., 2005).

597 Late Paleozoic N-MORB/E-MORB type HPGs experienced peak/post-peak 598~~amphibolite facies~~ granulite facies metamorphism at 11.0–12.5 kbar and 860–890 °C, 599and later experienced isothermal decompression and cooling to 7.2–7.5 kbar and 810–600820°C, followed by decompression at 5.5–7.2 kbar and 805–850 °C, at which time 601garnet was consumed to make hornblende and plagioclase coronae (Fig. 10B). The 602peak geothermal gradient is ~20 °C/km, and the pressure changes from peak to 603retrograde metamorphism suggest the Late Paleozoic HPGs exhumed at least 18 km 604from 42 km to 24 km depth (Fig. 10B). Near clockwise *P-T-t* path is analogous to the 605widespread high-pressure retrograde eclogite in the Dunhuang Block, Western

Tianshan of CAOB and recorded an exhumation history, and large-scale emergence of eclogite in the Late Paleozoic suggest a steep subduction angle (Table S7) (e.g., Ni et al., 2006; Xiao et al., 2013; Wang et al., 2016; Zhang et al., 2018). When compared with the granulite-facies metamorphism of the Paleoproterozoic HPGs, we interpret that a relatively deeply buried (>42 km) crust might have been metamorphosed to eclogite-facies conditions and subsequently exhumed in the Late Paleozoic (Fig. 11).

Exhumation evolution history of the HPGs

The studied samples were collected from the northern margin of the North China Craton and document two metamorphic events related to amalgamation of the Columbia and Pangea supercontinents. Three E-MORB type HPGs located south of the CCGCP fault record Paleoproterozoic metamorphic ages and three N-MORB/ E-MORB type HPGs from north of the fault record Late Paleozoic metamorphic ages (Fig. 1B).

Paleoproterozoic HPGs and garnet-trondhjemite represent portions of Paleoproterozoic oceanic crust that crystallized at c. 2.50-2.55 Ga, and synchronously metamorphosed in the Trans North China Orogen that welded together the Eastern and Western Blocks of the NCC (e.g., Guo et al., 2005; Qian and Wei, 2016; Gao et al., 2021; Zhang et al., 2021). The c. 2.5–2.44 Ga metamorphic age recorded the cratonization of the NCC and overprinted by subsequent prograde metamorphism at c. 1904 Ma, driven by collision between the Western and Eastern Blocks (Figs. 7, 8). This was followed by peak metamorphism at c. 1887 Ma, and retrograde exhumation from the lower to middle crust at c. 1797 Ma (Fig. 11A) (Zhao et al., 2005; Zhai and

628Santosh, 2011; Tang and Santosh, 2018; Zhang et al., 2020). Variable HREE and Eu
629depletion in Paleoproterozoic HPGs and garnet-trondhjemite indicate they were
630experienced granulite-facies metamorphism at c. 1.9–1.8 Ga (Fig. S2D). Zhang et al.
631(2016) correlates the 1850 Ma granulite-facies metamorphism with continental
632subduction and amalgamation of the Eastern and Western Blocks during formation of
633the NCC. However, Zhang et al. (2020) proposed that the Late Paleoproterozoic 1850
634Ma metamorphism reached eclogite-facies conditions, and thus document subduction
635and exhumation of oceanic crust. Our study reveals the HPGs from Chicheng record
636granulite-facies prograde metamorphism at 1900 Ma and peak metamorphism soon
637afterwards (c. 1890–1880 Ma) (Fig. 11A), which is consistent with the two
638metamorphic events documented in the TNCO: a 1960–1900 Ma collisional event
639between the Eastern and Western Blocks, and retrograde cooling and exhumation in
6401880–1800 Ma (Tang and Santosh, 2018). Thus, subduction in the Chicheng region
641must have initiated prior to 1900 Ma, with HPGs initially buried to lower crustal
642depths. Later exhumation to the upper lower or middle crustal level was accompanied
643by metamorphism due to Eastern and Western Block collision (Fig. 11A). Our results
644together with the (1960–1800 Ma) granulite-facies tectonothermal event in TNCO,
645such as Xuanhua Complex, Fuping Complex, Huai’an Complex, Lüliang Complex,
646Hengshan Complex, Wutai Complex further reveal the Paleoproterozoic Columbia
647tectonothermal cycle is a long-lived subduction–collision–exhumation process that
648lasted at least ~160 Myr, and exhumation lasted ~80 Myr (Fig. 12) (Zhang et al.,
6492016; Guo et al., 2002; Zhao et al., 2000, 2001; O’Brien et al., 2005; Qian and Wei,

6502016; Gao and Santosh, 2019).

651 Late Paleozoic HPGs record a peak/post-peak amphibolite-facies metamorphism
652at 335–309 Ma, which we attribute to orogenic uplift of the Central Asian Orogenic
653Belt (Fig. 10B). The metamorphic age of 297–287 Ma in both HPGs and meta-
654trondhjemite is interpreted to record retrograde metamorphism and mineral growth
655during exhumation (Figs. 7F, 10B). The metamorphic age of 341 Ma from the meta-
656trondhjemite dates subduction of Paleo-Asian Oceanic crust (Fig. 7F). The lack of an
657Eu anomaly, low concentration and flat HREE patterns together with low Th/U values
658(0.002-0.118) from both Late Paleozoic HPGs and meta-trondhjemite, suggest they
659experienced eclogite-facies metamorphism before c. 340 Ma (Fig. S3, Table S6).
660Thus, the HPGs and meta-trondhjemite was contemporaneously formed by the
661metamorphism of the oceanic crust during the Late Paleozoic (Fig. 11B). Previous
662studies reported two periods of metamorphism in this region: firstly, the 355 Ma
663represent retrograde eclogite-facies metamorphism, following by amphibolite-facies
664metamorphism occurring at c. 320–300 Ma, finally overprinted by a c. 290–280 Ma
665large-scale thermal event (Kong et al., 2011; Chu et al., 2013). The history of these
666Late Paleozoic Chicheng HPGs is also coherent with amphibolite- to granulite-facies
667metamorphism (430–320 Ma) in the Western Tianshan, Dunhuang Block of western
668CAOB, thus Paleo-Asian Ocean subduction initiated to the west of Tianshan and
669propagated to the eastern North China Craton (Zhang et al., 2018; He et al., 2014).
670Thus, Paleo-Asian Ocean subduction initiation must have occurred earlier than c. 350
671Ma, the tectonic cycle and duration of the Late Paleozoic Pangea supercontinent

amalgamation is a rapid and short-lived tectonothermal event and completed in ~60 Myr, and exhumation lasted ~24 Myr (Fig. 12).

Tectonic implications for exhumation mechanism

The mechanisms controlling exhumation of buried or subducted crust depend on several parameters, including the depth of burial, velocity of exhumation, extent of serpentinization, crustal hydration state/viscosity and density, and timescale (Guillot et al., 2001; Pilchin, 2005; Agard et al., 2009). Among these, the rate of exhumation though time are investigated, as constrained by crustal burial and exhumation depths, and coupled with timescales from peak to retrograde metamorphic stages. Data from the Paleoproterozoic TNCO of Columbia supercontinent, Late Neoproterozoic–Cambrian Sri Lanka and Antarctica of Gondwana supercontinent, Middle–Late Paleozoic CAOB of Pangea supercontinent, and the Cenozoic Himalaya are used for estimating the exhumation velocity and results are shown in Fig. 12.

We show a long-lived exhumation (1888–1797 Ma) timescale for the Chicheng Paleoproterozoic HPGs, with ~14 km of overburden removed at a slow average rate of 0.16 ± 0.08 mm/y (Fig. 11A). Combined with the buried oceanic crust of the Xuanhua, Fuping, Lüliang, Wutai, Hengshan Complex from north to south in TNCO, a long-lived (70–140 ~~Ma~~Myr) exhumation history in the TNCO is assumed, with the crust uplifted 11–32 km (average ~15 km) at a slow exhumation rate (0.14 ± 0.09 – 0.29 ± 0.10 mm/y) (Fig. 12). However, Late Paleozoic HPGs from Chicheng were exhumed 18 km from a 42 km depth over a short timescale (335–287 Ma), relatively rapidly at 0.38 ± 0.14 mm/y (Fig. 11B). This Late Paleozoic oceanic crust which

likely experienced eclogite-facies metamorphism, was initially subducted to great depth, and exhumed by ~21 km depth into the lower or middle crust at 340–320 Ma at a rapid rate of 1.04 ± 0.34 mm/y (Ni et al., 2006; Kong et al., 2011). This is similar to deeply buried oceanic crust exposed in south Tianshan and the Dunhuang Block, western CAOB, meaning that the exhumation was likely sustained for c. 15-65 ~~Ma~~Myr, the Paleozoic oceanic crust was uplifted 18-51 km (average 30 km) by a fast exhumation velocity of 0.38 ± 0.14 – 3.4 ± 0.45 mm/y (Fig. 12). Thus, prolonged exhumation processes documented by Paleoproterozoic HPGs in the TNCO differ from much faster durations for Paleozoic examples in the CAOB (Fig. 12). Further comparison reveals relatively fast (0.19 ± 0.14 – 0.54 ± 0.10 mm/y, 15 km depth) exhumation processes for mafic granulites from Antarctic and Sri Lanka (40–65 ~~Ma~~Myr) fragments of the Gondwana supercontinent, and short-lived (3–15 ~~Ma~~Myr) rapid exhumation (1.25 ± 0.68 – 21.67 ± 2.27 mm/y, 65 km depth) process in Himalayan-Tibet, therefore, we propose that exhumation velocity and depth might also vary though time during supercontinent collision (Fig. 12) (Table S7).

Notably, although our results reveal the Paleoproterozoic HPGs (Grt–Hbl–Pl–Aug–Qtz–Ru, average MgO: 7.8 wt. %) are slightly denser than the Late Paleozoic rock (Grt–Hbl–Pl–Qtz–Ru, average MgO: 6.91 wt.%), no density diversity with the surrounding crust and relative shallow buried (<45 km) depth for the Chicheng HPGs, suggest buoyancy is not the driving force for exhumation (Fig. 11). However, the associated serpentinized ultramafic rocks with Late Paleozoic HPGs might have ‘lubricated’ the ascending units and accelerated the exhumation rate, and/or the low-

density serpentinite might have served as carrier for the ascending HPGs, together uplifted from the relative dense deep crust (Guillot et al., 2001; Pilchin, 2005). In addition, the relative steep subduction and high compression would increase the pressure which might facilitate an increase in exhumation velocity during uplifting (Fig. 11) (Stern et al., 2005; Palin et al., 2016b; O'Brien, 2021). Thus, the exhumation rate of the deep buried Late Paleozoic HPGs might be readily extruded though the parallel W-E trending CCGCP fault during exhumation (Fig. 1) (Chu et al., 2013; Zhang et al., 2016; Liu and Zhang, 2021). In addition, a secular decrease in crustal geothermal gradient since the Archean means that the relatively higher temperature may have played a role in decelerating the exhumation velocity of the Paleoproterozoic HPGs, and Late Paleozoic HPGs, which are easier to be uplifted in a cold environment (Fig. 11) (Agard et al., 2009; Palin et al., 2020; Maiti and Mandal, 2021).

CONCLUSIONS

Integrated zircon and monazite dating reveals Paleoproterozoic oceanic subduction initiated prior to 1904 Ma, and final collision of Eastern and Western Block at 1887 Ma, following by post-collision/exhumation at 1797 Ma. Thus, the tectonothermal cycle was long-lived process (>110 Myr) and exhumation lasted ~90 Myr. By contrast, Late Proterozoic ocean subduction initiated earlier than c. 340 Ma, peak amphibolite-facies metamorphism at 335–309 Ma, and 297-287 Ma is interpreted as retrograde metamorphism during exhumation. Therefore, the tectonic cycle and duration of Late Paleozoic subduction was short lived (>50 ~~Ma~~Myr) and the

738burial–exhumation cycle was completed within ~50 Myr. Petrological modelling
739suggests the Paleoproterozoic HPGs are E-MORB-type record a clockwise *P–T–t* path
740with granulite facies metamorphism. Late Paleozoic HPGs are N-MORB-type define
741an amphibolite-facies metamorphism. Comparison with granulite facies
742metamorphism of the Paleoproterozoic HPGs, Late Paleozoic HPGs might
743metamorphose to eclogite facies though a steep subduction process and subsequently
744exhumed. The slowly exhumed (0.16 mm/y) Paleoproterozoic HPGs and the fast
745exhumed (0.38 mm/y) Late Paleozoic HPGs suggest the exhumation velocity was
746facilitated and duration reduced through time, and exhumation mechanism might have
747been controlled by subduction angle, compression pressure, and temperature.

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FIGURE CAPTIONS

Figure 1 Geological map of North China Craton and North Hebei, and field photographs. (A) Geological map of North Heibei modified from Zhang et al., (2016). (B) Tectonic framework of North China Craton, with the study area marked by a red yellow box after Zhao et al. (2005) and Santosh, (2010). (C) – (F) Representative field photographs showing outcrops of samples investigated in this study. Abbreviations for the major geological features: CD, Chengde; NH, Northern Heibei; XH, Xuanhua; HA, Huai'an; HS, Hengshan; WT, Wutai; FP, Fuping; LL, Lüliang; ZH, Zanhuang; ZT, Zhongtiao; DF, Dengfeng; TH, Taihua. Other abbreviations: CCGCP, Chongli–Chicheng–Gubeikou–Chengde–Pingquan fault.

Figure 2 (A) – **(D)** Backscattered photomicrographs of representative mineral assemblages and microstructures for HPGs. **(E)** – **(F)** Microprobe elemental maps for Mn, Mg, Fe, and Ca for garnet grains from samples CH-1-1 and CH-5-1. Abbreviations: Pl – plagioclase; Hbl – hornblende; Grt – garnet; Cpx – clinopyroxene; Ilm – ilmenite; Qtz – quartz; Ru – rutile. Microprobe elemental maps for Mn, Mg, Fe, and Ca for garnet grains, with chemical zoning shown. Alm – almandine; Pyr – pyrope; Grs – grossular; Sps – spessartine. Dashed arrows are garnet compositional analysis location from rim to core to rim.

1117

1118**Figure 3** Geochemical plots of Paleoproterozoic and Late Paleozoic HPGs. (A) total
1119alkali vs. silica diagram plot (Wilson, 1989); (B) Nb/Y vs. Zr/TiO2 geochemical
1120classification diagram (after Winchester and Floyd, 1977); Chondrite-normalized
1121trace-element patterns for Paleoproterozoic (C) and Late Paleozoic (E) HPGs at
1122different metamorphic stages; Primitive mantle normalized trace element plot for
1123Paleoproterozoic (D) and Late Paleozoic (F) HPGs. Normalization values are from
1124Sun and McDonough (1989).

1125

1126**Figure 4** Pressure–temperature (*P–T*) pseudosections for the Paleoproterozoic mafic
1127granulite calculated by Theriak Donimo. Deep grey fields are constraint *P-T* range.
1128Bold black lines are THERMOCALC calculate average *P-T* condition. Black isopleth
1129lines are modelled proportion for clinopyroxene. Grey isopleth lines are modelled
1130proportion for garnet. White isopleth lines are modelled proportion for hornblende.
1131Abbreviations: pl – plagioclase; hbl – hornblende; grt – garnet; aug – augite; ilm –
1132ilmenite; qtz – quartz; ru – rutile; liq – silicate melt; opx – orthopyroxene.

1133

1134**Figure 5** Pressure–temperature (*P–T*) pseudosections for the Late Paleozoic mafic
1135granulite. Deep grey fields are constraint *P-T* range. Grey isopleth lines are modelled
1136proportion for garnet. White isopleth lines are modelled proportion for hornblende.
1137Abbreviations are same as Figure 4.

1138

1139**Figure 6** Modeled changes in mineral proportions of Paleoproterozoic and Late

Paleozoic HPGs. Pressure at 5–15 kbar and temperature at 750–1000 ºC. Dashed lines are modelled proportion analogous to the petrological observation results, arrows show the metamorphic *P-T* change direction under different metamorphic stage. Abbreviations are same as Figure 4.

Figure 7 Zircon U–Pb concordia plots, and CL images with analytical spots and ages. (A), (B) and (C) zircon U-Pb concordia plots for Paleoproterozoic HPGs and garnet-trondhjemite. (A) Light blue circles represent magmatic zircons, grey circles are metamorphic zircons for sample CH-3-1. (B) Grey circles are metamorphic zircons for sample CH-3-2A. (C) Green and red circles represent two magmatic zircon groups, blue circles are metamorphic zircons. (D), (E) and (F) zircon U-Pb concordia plots for Late Paleozoic HPGs and meta-trondhjemite. (D) Green, pink and yellow circles represent three metamorphic zircon groups for sample CH-1-1B. (E) Green and pink circles represent two metamorphic zircon groups for sample CH-5-1A. (F) Grey, pink and red circles represent three metamorphic zircon groups for sample CH-51-5.

Figure 8 Monazite U–Pb concordia plots, and BSE images with analytical spots and ages. (A) and (B) monazite concordia diagrams for Paleoproterozoic garnet-trondhjemite and Late Paleozoic meta-trondhjemite.

Figure 9 Discrimination diagrams used for the geochemical classification and

tectonic settings. (A) Dy/Yb vs La/Yb diagram (Jung et al., 2006) suggesting low degree partial melting of mantle in spinel peridotite facies. (B) Nb/Y vs. Rb/Y plot showing trends of increasing hydrous metasomatism, slab melt interaction, and increasing degree of partial melting (Kepezhinskas et al., 1997), (C) Sr/Y vs. Y diagram (Defant and Drummond, 1990) to discriminate adakites from other arc lavas. (D) – (E) are Th/Y vs. Nb/Yb and TiO2/Y vs. Nb/Yb diagrams after Pearce (2008), (F) ThN vs. NbN discrimination diagram after Saccani (2015).

Figure 10 Summary plots showing the interpreted *P–T–t* evolution. (A) Dashed line is predicted prograde to peak evolution progress, solid black lines represent peak to retrograde metamorphic progress. Green field represents prograde metamorphic *P-T* range and corresponding to the top-left microstructure. Purple field represents peak *P-T* range, and corresponding to the top-right microstructural. Pink field represents retrograde metamorphism and corresponding to the two right middle and bottom microstructure. (B) Solid black line represents predicted peak to retrograde metamorphic progress. The purple field represents peak *P-T* range, and correlate to the top-right microstructure showing garnet breaks down to form hornblende and plagioclase symplectite. Green and blue field represent retrograde *P-T* range, and corresponds to the two bottom microstructure and show garnet gradually breaks down until totally consumed.

Figure 11 Tectonic schematism showing exhumation process, include exhumation

depth, time, and rate. (A) Paleoproterozoic HPGs exhumed via shallow angle in the TNCO with small accretionary wedge, thick oceanic crust, and thin lithospheric mantle, (B) Late Paleozoic HPGs exhumed via steep angle in the CAOB with large accretionary wedge, thin oceanic crust and thick lithospheric mantle.

Figure 12 Exhumation parameters for the Paleoproterozoic HPGs from TNCO, Late Neoproterozoic to Cambrian HPGs from Antarctic, and Cenozoic HPGs from Himalayan. (A) Calculated exhumation velocity vs. age, (B) exhumation period, (C) exhumation depth of this and previous studies.

APPENDIXES

Fig. S1. Compositional diagrams in the HPGs. Classification diagram (A) and (B) for amphibole, (C) for clinopyroxene.

Fig. S2. Zircon chondrite-normalized REE diagrams. (A) and (B) for the Paleoproterozoic HPGs, (C) and (D) for the meta-trondhjemite samples.

Fig. S3. Zircon chondrite-normalized REE diagrams for the Late Paleozoic HPGs.

Fig. S4. Monazite chondrite-normalized REE diagrams for the meta-trondhjemite samples.

APPENDIXES 2

Table S1. Representative mineral compositions for all studied samples.

Table S2. Bulk-rock compositions derived from X-ray fluorescence analysis.

Table S3. Bulk-rock compositions utilised in petrological modelling (molar percentages of oxides).

Table S4. LA-ICP-MS zircon U–Pb data.

Table S5. LA-ICP-MS monazite U–Pb data.

Table S6. LA-ICP-MS zircon and monazite trace element data.

Table S7. Compilation exhumation velocity data of oceanic crustal rock from Chicheng, TNCO, CAOB, East Antarctica, and Himalayan.