Reinitiation of modern-style plate tectonics in the Early Neoproterozoic: Evidence from a ~930 Ma blueschist-facies terrane in South China

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Abstract

When and how plate tectonics operated throughout Earth's history remains a matter of debate. The absence of low-temperature/high-pressure metamorphic rocks during Earth's middle age (~1.8–0.8 Ga) leads to some arguing for a temporary shutdown or change of modern-style plate tectonics throughout the Mesoproterozoic and Early Neoproterozoic. Here we report the first discovery of ~930 Ma blueschist-facies metamorphic rocks (Tianli Schist) from the Jiangnan Orogen, South China. Peak pressure–temperature conditions of ~425–475 ℃ and ~0.8–1.1 GPa recorded by these units represent an average geothermal gradient of ~14 ℃/km, characteristic of cold subduction zones. Available geologic and geochronologic constraints indicate that the Tianli Schist protolith was deposited in the southeast margin of the Yangtze Block and subsequently experienced blueschist-facies metamorphism during ocean closure at ~930 Ma. Combined with the global metamorphic record, our results suggest that modern-style plate tectonics did not operate continuously since initiation in the Early Paleoproterozoic, but was inhibited in the Mesoproterozoic and restarted again at the end of Earth's middle age.

Keywords:

blueschist; Tianli Schist; plate tectonics; Neoproterozoic; cold subduction

1. Introduction

Plate tectonics is unique to Earth, although when it initiated and how it has changed throughout geological time are subject to considerable debate (e.g., Condie et al., 2008; Dhuime et al., 2015; Stern, 2005). The paucity of the Precambrian rock record has led to many researchers relying on geochemical proxies or modeling to infer the operation (or not) of subduction and plate motion during the Proterozoic and Archean (e.g., Dhuime et al., 2015; Tang et al., 2016). While plate tectonics is widely believed to have emerged by the end of Archean, whether its style has remained the same since initiation is controversial (e.g., François et al., 2018; Stern, 2005). A diagnostic feature of modern-style plate tectonics is cold subduction that typically develops low-temperature (low-T) blueschist in the descending slab. Blueschist is traditionally defined as a glaucophane-bearing mafic metamorphic rock formed at a low geothermal gradient of 5–15 °C/km but some metasedimentary rocks can also suffer blueschist-facies metamorphism in subduction zones (Brown & Johnson, 2019; Ernst, 1988; Li, 2020). Thus, blueschist-facies metamorphism can be viewed as diagnostic evidence to track the continuity of plate tectonics since the Proterozoic.

The oldest blueschist-facies rocks occur in the ~2.1 Ga greenstone belts in Burkina Faso, West Africa Craton (Ganne et al., 2012), and have similar ages to four global eclogite-facies units (~2.1–1.8 Ga) that share low thermobaric ratios (low T/P) (Fig. 1a; Brown & Johnson, 2019). Together, these occurrences have been regarded as

documenting the initiation of cold subduction in the Paleoproterozoic (e.g., François et al., 2018) following a period of warm and shallow subduction during the Late Archean and Early Paleoproterozoic (Palin & Santosh, 2021). Curiously, blueschist-facies rocks are absent from Earth's subsequent middle age (~1.8–0.8 Ga; Cawood & Hawkesworth, 2014), which has been variably ascribed to a lack of exhumation and preservation bias (e.g., Pereira et al., 2021), a hotter mantle (e.g., Bang et al., 2021) or a more mafic ocean crust (e.g., Palin and White, 2016). The re-appearance of blueschists at ~0.8 Ga (Xia et al., 2019), followed by a near-continuous record of associated low T/P metamorphic rocks (Brown & Johnson, 2019), has been argued for the emergence of modern-style plate tectonics in the Neoproterozoic (Stern, 2005).

Scarce low T/P metamorphism during ~1.8–0.8 Ga correlates to an extended period of subdued orogenic vigor (Cawood & Hawkesworth, 2014; Spencer et al., 2021; Tang et al., 2021), which may suggest that modern-style plate tectonics experienced a hiatus or change during this period. Here we report the first discovery of a ~930 Ma blueschist-facies metasedimentary rock in South China that marks the oldest known Neoproterozoic example on Earth and supports the hypothesis that modern-style plate tectonics reinitiated in the Early Neoproterozoic after being inhibited temporarily in the Mesoproterozoic.

2. Geology and samples

The studied rock samples are collected from the Tianli Schist in the Jiangnan

Orogen, South China. The South China Block is surrounded by the North China Block, Songpan-Ganzi terrane, Indochina Block, and Pacific Ocean (Fig. 1b). It is composed of the Yangtze Block in the northwest and Cathaysia Block in the southeast, which amalgamated to form the Jiangnan Orogen (also called Sibao Orogen) during the Early Neoproterozoic (e.g., Li et al., 2009). The Tianli Schist is chiefly exposed in the eastern segment of the Jiangnan Orogen (Fig. 1b). It consists predominantly of quartz mica schists and calcareous mica schists, with minor quartzite, sandstone, and marble (Li, Wartho, et al., 2007; Wang et al., 2018). The protolith of the Tianli Schist was likely deposited in a shallow marine environment and subsequently underwent greenschist-facies metamorphism and intense deformation during the Jiangnan Orogeny (Li, Wartho, et al., 2007; Wang et al., 2018). Previous Ar–Ar dating of muscovite in this terrane yielded ages of ~1.04–0.94 Ga (Li, Wartho, et al., 2007). It is in fault contact with or unconformably overlain by the Late Neoproterozoic rift successions (Fig. 1c–d), which experienced less intense deformation (Wang et al., 2018). The unconformably overlying Taoyuan Formation mainly consists of volcanic and pyroclastic rocks with a rhyolite on the top dated at ~827 Ma (Li, Li, & Li, 2008). The Wenjialing Formation consists of basal conglomerate, volcanic clasts, and reddish sandstone (Li, Li, & Li, 2008; Wang et al., 2018). The Tingmeng Formation consists of feldspathic lithic sandstone, siltstone, and silty mudstone. The Liuyuan Formation consists of quartz-sandstone with well-developed cross-beddings and is overlaid by the Nantuo tillite and Ediacaran strata

(Li, Li, & Li, 2008; Wang et al., 2018).

Mica schist samples were collected within the Tianli Schist unit and away from the unconformity or fault with the upper Taoyuan and Wenjialing Formation (118°8'23.30" E, 28°28'12.05" N; Fig. 1c and Fig. 2a–c). They are deformed and dominantly composed of phengite (~40%), dolomite (~35%), and quartz (~16%), with minor biotite (~7%), albite (~3%), K-feldspar, tourmaline, and chlorite (Fig. 2d and Fig. 3). Phengite is textually equilibrated with quartz, dolomite, albite, and biotite (Fig. 3a–b), while K-feldspar and chlorite occur as retrograde phases replacing albite and biotite (Fig. 3c–d), respectively. Accessory rutile is commonly included within phengite or occurs as matrix phases (Fig. 3e–f). Monazite occurs both within phengite and in the grain boundaries intergrown with quartz and apatite (Fig. 3f). Petrographic observations define a peak mineral assemblage of phengite+dolomite+quartz+biotite+albite+rutile+monazite.

3. Analytical methods

Monazite and phengite grains were separated from the schist samples using conventional magnetic and heavy liquid techniques and picked under a binocular microscope. Monazite samples were mounted in epoxy resin. Transmitted and reflected light micrographs and cathodoluminescence (CL) images were used to document the internal structures and mineral inclusions of monazite.

3.1 Image investigations

CL images of monazite, backscattered electron (BSE) images of thin sections, and identification of mineral species were obtained using a Nova NanoSEM 450 field-emission scanning electron microscopy equipped with a Gatan MonoCL4 detector and Aztec electron backscatter diffraction system at the Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing (IGG-CAS).

The mineral proportions of the schist samples were obtained by qualitative X-ray elemental mapping of bulk thin sections. The analysis was performed using a Bruker M4 Tornado micro-X-ray fluorescence spectrometer equipped with poly-capillary lens and silicon drift X-ray detectors at the Guangzhou Tuoyan Analytical Technology Company, Guangzhou. The thin sections were scanned with a 50 kV accelerating voltage, 300 μA probe current, 30 μm pixel resolution, and 5 ms dwell time per pixel. The XMapTools software (Lanari et al. 2014) was used to process the elemental maps and to quantitatively obtain mineral proportions.

3.2 Mineral composition analysis

Elemental compositions of monazite, carbonate, and tourmaline were determined using a CAMECA SXFiveFE electron probe microanalyzer (EPMA) equipped with large LiF, polyethylene terephthalate (PET), and thallium acid phthalate (TAP) diffraction crystals at IGG-CAS. Monazite analysis was performed with a 10 kV accelerating voltage, 200 nA beam current, 10 μm defocused beam diameter, 20–30 s

144 counting time per element, and a single-spot acquisition time of ~6 min. Tourmaline
145 was analyzed with a 15 kV accelerating voltage, 20 nA beam current, focused beam,
146 10–20 s counting time per element, and a single-spot acquisition time of ~3 min.
147 Carbonate was analyzed with a 15 kV accelerating voltage, 10 nA beam current, 10
148 μm defocused beam diameter, 10–20 s counting time per element, and a single-spot
149 acquisition time of ~3 min. Other instrumental conditions are provided in Table S1.

150 Elemental compositions of mica, feldspar, and chlorite in two thin sections were
151 determined using a JXA-8100 EPMA equipped with TAP, LiF, PET, and LDE (a
152 synthetic multilayer crystal, W/Si) crystals at IGG-CAS, with a 15 kV accelerating
153 voltage, 20 nA beam current, 5 μm electron beam diameter, 10–20 s counting time per
154 element, and a single-spot acquisition time of ~3 min. Other instrumental conditions
155 are provided in Table S1.

156 Zr contents of rutile in thin sections were determined using a CAMECA
157 NanoSIMS 50L at IGG-CAS, with a 150 pA primary beam intensity, 200–300 nm
158 spot size, and a raster area of 2 μm. An in-house rutile standard, JDX (Zr = 1049 ± 91
159 ppm, 2 SD), was used to correct the Zr/Ti fractionation during analysis. To ensure the
160 correct location of small rutile grains, an area of 40 × 40 μm^2 was mapped around
161 each rutile. Then multiple spot analyses were performed on different domains for each
162 rutile. The precision of Zr/Ti ratios was ~10% (2 SE) during NanoSIMS analysis and
163 ~20% (2 SE) for a single sample spot after error propagation of standard correction
164 and TiO_2 variation of 5 wt.%.

3.3 Monazite U–Th–Pb and O isotopic analysis

Monazite U–Th–Pb isotopes were analyzed by a CAMECA IMS 1280HR secondary ion mass spectrometry (SIMS) at IGG-CAS following a similar analytical procedure to Li et al. (2013). The intensity of the primary beam was ca. 1.0 nA and the beam size was about 10 × 15 μm^2. Each spot measurement was acquired within ca. 12 min. The RW-1 monazite standard (^{206}Pb/^{238}U age = 904.15 Ma; Ling et al., 2017) was used to calibrate the Pb/U and Pb/Th ratios. Due to isobaric interference of ^{232}Th^{144}Nd^{16}O$_2^{++}$ on ^{204}Pb, the ^{207}Pb-based common Pb correction was used (Li et al., 2013). Decay constants are adopted from Steiger and Jäger (1977). Another monazite standard USGS-44069 was analyzed as an unknown sample to monitor the accuracy. The USGS-44069 monazite was measured nine times, yielding a ^{206}Pb/^{238}U age of 423.6 ± 8.1 Ma (2 SD) and ^{208}Pb/^{232}Th age of 426.5 ± 5.3 Ma (2 SD), which are consistent within errors with the recommended value of 424.9 ± 0.4 Ma (2 SD; Aleinikoff et al., 2006).

Monazite oxygen isotope was analyzed by a CAMECA IMS 1280 SIMS at IGG-CAS following a similar analytical procedure to Wu et al. (2020). The primary Cs^+ ion beam was accelerated at 10 kV, and the beam current was ~2.0 nA. The beam size was ca. 10 × 15 μm^2, and each spot measurement was acquired within ~3 min. The oxygen isotopic data were expressed as δ notation (‰) by normalizing raw ^{18}O/^{16}O ratios to Vienna Standard Mean Ocean Water compositions (^{18}O/^{16}O = 0.0020052). The internal precision of a single analysis was generally 0.3‰–0.4‰ (2 SE). During the analysis,

four monazite standards (Manangoutry, Itambe, USGS-44069, and UNIL1-Mnz1) were used to correct the instrumental mass fractionation and matrix effects (Wu et al., 2020). The monazite standard RW-1 was analyzed as an unknown sample to monitor the accuracy. Five measurements of RW-1 monazite yielded an average $\delta^{18}O$ value of 6.00 ± 0.26 ‰ (2 SD), which is identical within error to the recommended value of 6.30 ± 0.16 ‰ (2 SD; Wu et al., 2019).

3.4 Sm–Nd isotopic analysis of monazite, phengite, and bulk rock

Monazite Sm–Nd isotopes were analyzed by a Thermo Neptune Plus laser ablation multi-collector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS) equipped with a Geolas-193 laser-ablation system at IGG-CAS. The analytical procedures were similar to Liu et al. (2012), using a 16 μm spot size, 4 Hz repetition rate, and 10 J/cm^2 laser beam energy density. The $^{147}Sm/^{149}Sm$ (Dubois et al., 1992) and $^{144}Sm/^{149}Sm$ (Isnard et al., 2005) ratios were used for correcting the isobaric interference of ^{144}Sm on ^{144}Nd. The $^{146}Nd/^{144}Nd$ ratio (0.7219) was used to correct the measured $^{143}Nd/^{144}Nd$ ratios using the exponential law. The Namaqualand monazite standard (Liu et al., 2012) was used to correct the instrumental mass fractionation. An in-house monazite standard Iveland was analyzed as an unknown sample. Thirty measurements of Iveland monazite yielded an average $^{143}Nd/^{144}Nd$ ratio of 0.512944 ± 0.000024 (2 SD), which is consistent within error with the recommended value of 0.512963 ± 0.000093 (2 SD; Liu et al., 2012).

Phengite and whole rock Sm–Nd isotopes were analyzed by isotope dilution

thermal ionization mass spectrometry (ID-TIMS) at IGG-CAS. Distilled HF and HNO_3 were used to dissolve the mineral or whole-rock powers in Teflon capsules. A conventional two-stage ion exchange purification technique (Li, Chen, & Li, 2007) was used to extract Sm and Nd from matrix elements. Sm–Nd isotopic ratios were analyzed by a Thermo Scientific Triton Plus TIMS (Li, Chen, & Li, 2007). The Sm and Nd procedural blanks were generally <50 pg, requiring no correction of isotopic ratios. The $^{146}Nd/^{144}Nd$ ratio (0.7219) was used to correct the measured $^{143}Nd/^{144}Nd$ ratios using the exponential law. During the analysis, the standard solution JNdi-1 yielded a $^{143}Nd/^{144}Nd$ ratio of 0.512101 ± 0.000008 (2 SE) which is consistent within error with the reference value of 0.512115 ± 0.000007 (2 SD; Tanaka et al., 2000). Sample results were corrected to the JNdi-1 standard.

3.5 XRF major element analysis of bulk rock

Whole-rock samples were crushed and powdered to 200 meshes in an agate mortar. The dried rock powder was heated in a preignition silica crucible to 1000 ℃ for 1 h to calculate the loss-on-ignition content. Major element oxides were analyzed by a standard X-ray fluorescence spectrometer (XRF) on fused glass beads with analytical uncertainties <2% (2 RSD). One sample fraction was analyzed at IGG-CAS without determining ferrous oxide content. Three sample fractions were analyzed at the Wuhan SampleSolution Analytical Technology Company, Wuhan, with ferrous oxide determined by wet chemical analysis.

4. Results

4.1 Mineral and whole-rock chemistry

Representative mineral and whole-rock compositions are listed in Table S1. Phengite has high Si contents of 3.55–3.64 (3.58 on average) per formula unit (p.f.u.) and X_{Mg} (= molar Mg/[Mg + Fe^{2+}]) of 0.66–0.74 (0.70 on average; Fig. 4a), similar to the compositions of phengite formed under blueschist facies conditions (Fig. 4b). Biotite has 0.06–0.08 p.f.u. Ti and high MgO contents (17.88–19.07 wt.%), with X_{Mg} values of 0.71–0.73. K-feldspar has low contents of Na_2O (<0.2 wt.%) and CaO (<0.2 wt.%), while plagioclase is almost albite (CaO<0.1 wt.%). FeO is considerable in dolomite (3.14–6.12 wt.%) but is negligible in calcite (below the detection limit). Both chlorite and tourmaline have high MgO (20.77–23.87 wt.% and 7.80–11.02 wt. %, respectively) and FeO contents (15.07–16.16 wt.% and 3.39–10.20 wt.%, respectively). Monazite grains have low UO_2 (<0.05 wt.%) and high ThO_2 (1.32–5.61 wt.%) contents, and show enrichment in light rare earth elements, with 28.56–31.49 wt.% Ce_2O_3, 14.21–18.48 wt.% La_2O_3, and 10.32–12.18 wt.% Nd_2O_3. Eleven rutile grains, associated with phengite with Si contents of >3.60 p.f.u., have Zr contents ranging from 18 ppm to 38 ppm. No clear chemistry difference has been observed between the rutile inclusions and matrix rutile.

The Tianli Schist has high contents of Al_2O_3 (10.41 wt.%), CaO (9.69 wt.%), MgO (8.19 wt.%), and K_2O (4.96 wt.%), and low contents of TiO_2 (0.50 wt.%), MnO (0.07 wt.%), Na_2O (0.41 wt.%), and P_2O_5 (0.14 wt.%). The loss-on-ignition content is

considerable (17.63 wt.%) due to high amounts of dolomite and mica. Three sample fractions give an average FeO content of 2.79 wt.% determined by wet chemical analysis; thus, the whole rock shows a high X_{Mg} value (0.89).

4.2 Monazite U–Th–Pb–O isotopes

Monazite crystals from the Tianli Schist are mostly euhedral to subhedral in shape and exhibit patchy CL images (Fig. 5a). SIMS U–Th–Pb geochronology of monazite reveals a concordia age of 927.6 ± 3.2 Ma (2 SE, n = 29; Fig. 5a; Table S2). All monazite grains have $\delta^{18}O$ values ranging from 15.03 ‰ to 16.42 ‰ with a mean of 15.78 ± 0.65 ‰ (2 SD, n = 29; Fig. S1; Table S2).

4.3 Mineral and whole-rock Sm-Nd isotopes

LA-MC-ICP-MS monazite Sm–Nd isotopic analysis gives $^{143}Nd/^{144}Nd$ values ranging from 0.511038 to 0.511142 with an average of 0.511093 ± 0.000046 (2 SD, n = 25; Table S3). These data reveal an average initial εNd(930 Ma) value of –18.0 ± 0.3 (2 SD; Fig. S1) and two-stage Nd-model age (T_{DM2}) of 3.0 Ga. The monazite Sm-Nd isotope itself defines an errorchron age of 916 ± 160 Ma (2 SE; MSWD = 0.85; Fig. 5b).

TIMS phengite Sm–Nd isotopic analysis gives a $^{143}Nd/^{144}Nd$ value of 0.511131, with a calculated initial εNd(930 Ma) value of –17.6 (Table S3). TIMS Sm–Nd isotopic analysis of the whole rock yields a $^{143}Nd/^{144}Nd$ value of 0.511163, with a calculated initial εNd(930 Ma) value of –17.6 (Table S3). Monazite–phengite–whole rock Sm-Nd isotopes define an errorchron age of 1115 ± 170 Ma (2 SE; MSWD =

1.7; Fig. 5b).

5 Peak P–T estimate

The peak metamorphic conditions of the Tianli Schist were initially calculated using conventional thermobarometry. Peak temperature was estimated using the Zr-in-rutile thermometer. As quartz and zircon are commonly adjacent to rutile, the activities of tetragonal-ZrO_2 and α-quartz were treated as 1. The Zr-in-rutile thermometer (Kohn, 2020) was applied to 11 rutile crystals, yielding temperatures between ~425 °C and ~475 °C at 8–11 kbar (Fig. 6; see below) for the minimum and maximum Zr concentrations, respectively.

Independent constraints on peak P–T conditions were obtained via phase equilibria modeling in the system Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–TiO_2–H_2O–CO_2 (NCKFMASTHC) using Perple_X 6.8.6 program (Connolly, 1990, 2005) and the thermodynamic dataset of Holland and Powell (1998, updated 2002) and the Compensated Redlich Kwong (CORK) equation of state for H_2O–CO_2 fluids (Holland & Powell, 1991, 1998). MnO and Fe_2O_3 were neglected due to their low abundances. The following activity–composition models were used: garnet (Gt), phengite (Ph), glaucophane (Gln), omphacite (Omp), chlorite (Chl), dolomite (Dol), magnesite (Mgs) (Holland & Powell, 1998), biotite (Bt) (Tajčmanová et al., 2009), feldspar (Kfs/Pl/Ab) (Fuhrman & Lindsley, 1988), and F for H_2O–CO_2 fluid solution (Connolly & Trommsdorff, 1991). Quartz (Qz), rutile (Rt), calcite (Cal), titanite (Ttn),

291 and lawsonite (Law) were considered as pure phases. Mineral abbreviations are after
292 Whitney and Evans (2010). The B-rich phase, tourmaline, is not considered in the
293 modeling due to the lack of suitable activity–composition model. Therefore, we
294 estimate an effective bulk composition by subtracting minor tourmaline and
295 retrograde K-feldspar and chlorite from the XRF data. The effective bulk composition
296 used for the modeling is as follows (in mol.%): H_2O = 5.57%, CO_2 = 20.03%, SiO_2 =
297 39.97%, TiO_2 = 0.14%, Al_2O_3 = 4.48%, CaO = 11.17%, MgO = 12.21%, FeO =
298 3.42%, K_2O = 2.62%, and Na_2O = 0.38%. The modeled peak assemblages of
299 Ph+Bt+Dol+Ab+Rt+Qz+Cal are stable in a wide P–T range (Fig. 6). Although calcite
300 has not been observed in our samples, its modeled proportion in the peak-assemblage
301 stability field is <2 vol.%, probably due to the uncertainties of mineral activity-
302 composition models. We further use the isopleths of biotite Ti content and phengite Si
303 content (two components in mica having diffusion rates slower than Mg and Fe) to
304 determine the peak P–T conditions of the Tianli Schist (Fig. 6). The measured
305 phengite data show a positive correlation between X_{Mg} and Si contents (Fig. 4a),
306 indicating that some phengite domains might record prograde or retrograde
307 compositions, although no clear zoning was observed in individual grains. If we treat
308 the phengite with Si contents above 3.60 p.f.u. as the peak composition, the modeled
309 isopleths of Si in phengite (3.60–3.63 p.f.u.) and Ti in biotite (0.06–0.08 p.f.u.) match
310 well with the measured results. The overall consistency between the Zr-in-rutile
311 thermometer, modeled and observed assemblages, and mineral compositions indicate

peak P–T conditions of 425–475 °C and 8–11 kbar (Fig. 6), corresponding to blueschist-facies metamorphism (Fig. S2) with an apparent geothermal gradient of ~14 °C/km. The estimated blueschist-facies conditions are also consistent with the presence of albite at the peak stage (non-eclogite facies) and with the typical compositions of blueschist-facies phengite (Fig. 4b).

To further test the reliability of our results, we investigated the effects of bulk rock compositions (X_{Mg} [MgO/(MgO+FeO_t)], X_{Al} [Al_2O_3/(Al_2O_3+MgO+FeO_t)], X_{Ca} [CaO/(CaO+MgO+FeO_t+Na_2O)] and X_{H2O} [H_2O/(H_2O+CO_2)]) on the estimated peak pressure conditions, as the temperature can be well yielded by the Zr-in-rutile thermometer (450 °C, mean value, Fig. 6). Our results clearly show that the varying bulk X_{Mg}, X_{Al}, X_{Ca} and X_{H2O} can indeed affect the stability field of peak mineral assemblages (Bt+Ph+Ab+Dol+Rt+Cal+Qz) (Fig. S3). However, these compositional variables hardly change Ti-in-biotite values (0.06–0.08 p.f.u.) for the peak mineral assemblages. More importantly, the observed Si-in-phengite values above 3.60 p.f.u. point to a high pressure (>9 kbar), which has similar features to the blueschist phengite (Fig. 4b). As a result, metamorphic pressures are constrained within 9–10 kbar in order to match the modeled peak assemblages and the measured phengite Si and biotite Ti contents. Therefore, we conclude that bulk rock compositions do not significantly affect the estimated peak P–T conditions.

6. Discussion

6.1 Blueschist-facies metamorphism at ~930 Ma

Monazite grains from the Tianli Schist show equilibrium texture with phengite (Fig. 3f). They have patchy cathodoluminescence textures and high Th contents, which distinguishes them from typical hydrothermal monazite (Th < 1 wt.%; Wu et al., 2019). Therefore, we interpret that these monazite grains formed during the same metamorphic stage with phengite. This interpretation is further supported by their overall identical U–Th–Pb age (~930 Ma; Fig. 5a) and their consistency within error with the youngest mica $^{40}Ar/^{39}Ar$ age population reported in previous studies (Li, Wartho, et al., 2007). In addition, Sm–Nd isotopes of different monazite grains from Tianli Schist define an errorchron age of 916 ± 160 Ma (Fig. 5b), which is also consistent within error with their U–Th–Pb age. However, Sm–Nd isotopes of phengite and whole rock distinctly deviate from the monazite errorchron line, suggesting that Sm–Nd isotopic equilibrium was not achieved between phengite and monazite due to slow elemental diffusion during low-temperature metamorphism.

The deposition age of the Tianli Schist protolith is constrained between the youngest detrital zircon age group (~1530 Ma; Li, Wartho, et al., 2007; Wang et al., 2018) and the ~930 Ma blueschist-facies metamorphic event. The enriched O–Nd isotopic compositions of monazite and T_{DM2} model age of ~3.0 Ga suggest that the protolith of the Tianli Schist was derived from eroded supracrustal rocks from the Yangtze Block, as Archean rocks outcrop there, but not in the Cathaysia Block (e.g.,

353 Gao et al., 2011). This interpretation is further confirmed by provenance studies using
354 detrital zircon (Li, Wartho, et al., 2007; Wang et al., 2018). Between ~1.0 Ga and 0.9
355 Ga, the oceanic plate between the Yangtze Block and Cathaysia Block was
356 subducting beneath the southeast margin of the Yangtze Block, as documented by the
357 presence of 970–890 Ma Shuangxiwu continental arc (Li et al., 2009), ~1.0 Ga
358 Zhangshudun ophiolite (Chen et al., 1991), and ~0.97 Ga Xiwan adakitic granite (Li
359 & Li, 2003) in the southeast Yangtze. Glaucophane-bearing blueschist was also
360 previously reported in Xiwan, southeast Jiangnan Orogen, and dated at ~0.9 Ga by
361 $^{40}Ar/^{39}Ar$ dating of amphibole (Xu et al., 1992). Unfortunately, this blueschist is
362 poorly outcropped and lacks further precise dating and P–T constraints. Therefore, we
363 suggest that the Tianli Schist underwent blueschist-facies metamorphism due to
364 subduction of shelf/passive margin sediments at ~930 Ma.

365 **6.2 Reinitiation of modern-style plate tectonics at the end of Earth's middle age**

366 The Tianli Schist represents the second oldest blueschist-facies rock and the
367 oldest known example from the Neoproterozoic, pushing back the oldest confirmed
368 date for reinitiation of modern-style plate tectonics to ~930 Ma. Although several low
369 T/P eclogites are known from various terranes worldwide with ages of ~2.1–1.8 Ga
370 and ~1.1–1.0 Ga but not in between (e.g., Brown & Johnson, 2019; Fig. 7a), these
371 rocks were formed in relatively high temperatures (>600 ℃). The new ages and P–T
372 data presented here, support the hypothesis that the blueschist-facies metamorphism

and modern-style plate tectonics were inhibited between the Late Paleoproterozoic and the Early Neoproterozoic. The blueschist-facies metamorphic rocks that bracket Earth's middle age at ~2.1 Ga and ~0.9 Ga also preserve relatively low pressures of equilibration (~0.8–1.1 GPa) compared to those of Phanerozoic orogens (>1.5 GPa), which inhibits the formation of glaucophane (Ganne et al., 2012 and this study). This may be due to a relatively hotter upper mantle in the Proterozoic that produced elevated geothermal gradients in subduction zones (e.g., Chen et al., 2022; Roberts et al., 2022; Zou et al., 2023).

It has been proposed that the mantle has not monotonically cooled since the end of Archean, and that instead, it may have increased in temperature at ~2.1–1.8 Ga due to increased continental insulation associated with the supercontinent assembly of Nuna (Lenardic et al., 2011; Tamblyn et al., 2022). Thus, blueschist-facies and eclogite-facies metamorphism during this time would have recorded atypically hot geotherms compared to Phanerozoic examples (Fig. 7a). The 2.1–1.8 Ga amalgamation of Nuna developed worldwide ophiolites, low T/P eclogites, orogenic gold and volcanic-hosted massive sulfide deposits (Brown & Johnson, 2019; Cawood & Hawkesworth, 2014; Condie, 2018; Fig. 7a). It also induced intense continental collisions (e.g., Zhao et al., 2002), which generated a thick continental crust (Tang et al., 2021) and eroded more radiogenic continent materials into oceans (e.g., high $^{87}Sr/^{86}Sr$ ratios, Shields, 2007; Fig. 7b).

393 Noteworthily, there was a reduced level of breakup and reassembly of continental
394 fragments during the transition from Nuna to Rodinia supercontinent (~1.8–0.9 Ga)
395 due to the opening and closure of internal oceans rather than external oceans (Li et al.,
396 2019). This resulted in a paucity of passive margins, orogenic gold and volcanic-
397 hosted massive sulfide deposits (Cawood & Hawkesworth, 2014). The limited
398 continental collisions also resulted in relatively high εHf(t) values of detrital zircon,
399 an increased zircon Eu anomaly, and a decreased granite Sr/Y ratio (Cawood &
400 Hawkesworth, 2014; Tamblyn et al., 2022; Tang et al., 2021; Triantafyllou et al.,
401 2022). Although orogenesis occurred during Earth's middle age (e.g., the ~1.3–1.0 Ga
402 Grenville orogen), it is characterized by a large volume of massive anorthosites and
403 A-type granites and a warmer average geothermal gradient, which clearly
404 distinguishes from Paleoproterozoic and Neoproterozoic counterparts. These features
405 point to hot, thin-crust, and low-elevation orogens (Brown et al., 2020; Brudner et al.,
406 2022; Cawood & Hawkesworth, 2014; Spencer et al., 2021; Wu et al., 2023; Zhu et
407 al., 2022), which would have reduced continental exhumation and weathering (Zou et
408 al., 2023) and finally promoted a subdued seawater Sr isotope (Fig. 7b; Shields,
409 2007). Therefore, cold subduction would not have been possible in the
410 Mesoproterozoic (Tamblyn et al., 2022; Stern, 2023), explaining the absence of
411 blueschists and low-T eclogites at this time (Fig. 7a).
412 In contrast, at the end of the Earth's middle age (~1.0–0.9 Ga), blueschist-facies
413 metamorphism (i.e., Tianli Schist) re-emerged together with a cold average

geothermal gradient (Brown et al., 2020), suggesting reinitiation of modern-style plate tectonics. This process could be explained by sustained mantle cooling below a critical threshold for blueschist preservation (Tamblyn et al., 2022). Meanwhile, this enhanced orogenic vigor developed coeval arc magmatism worldwide (e.g., Valhalla accretionary orogen in Scotland, east Greenland, Svalbard, and Norway; Central Asian Orogenic Belt in China and Russia; arc magmatism in South China, East Antarctica, Saharan Metacraton, Arctic, Alaska, India, and Sri Lanka; Cawood et al., 2010; Elburg et al., 2015; Evuk et al., 2014; Kröner et al., 2013; Li et al., 2009; Malone et al., 2017; Yang, Santosh, et al., 2020; Yang, Xu, et al., 2020; Zong et al., 2017), generated high-elevation mountain ranges, and increased weathering and erosion of radiogenic continent material as reflected in an elevated thickness of magmatically active continental crust and normalized seawater Sr isotope (Fig. 7b; Shields, 2007; Tang et al., 2021; Zou et al., 2023). The supercontinent thermal insulation eventually diminished during Rodinia breakup, generating widespread low-T/P rocks (Brown and Johnson, 2019; Chen et al., 2022; Fig. 7a).

Our new discovery also suggests that metasedimentary rocks should be investigated in detail as a potential record of ancient blueschist-facies metamorphism that has been formerly unrecognized. Focused studies of Proterozoic metasediments in other terranes worldwide may lead to further new discoveries of low-T/high-P lithologies, which will place additional constraints on the evolution of geodynamic regimes on Earth through time.

7. Conclusion

This study presents the first discovery of the oldest blueschist-facies metamorphism in the Neoproterozoic. The Tianli Schist samples from the Jiangnan Orogen, South China, have phengite and rutile formed at blueschist-facies conditions (425–475 °C and 8–11 kbar) and under a geothermal gradient of ~14 °C/km. Precise monazite U–Th–Pb geochronological data show that the Tianli Schist underwent blueschist-facies metamorphism at ~930 Ma, indicating cold subduction existing in the Early Neoproterozoic. Based on the global metamorphic records, we propose that modern-style plate tectonics may not have operated continuously since its onset during the Early Paleoproterozoic; instead, it was inhibited in the Mesoproterozoic and restarted again at the end of Earth's middle age.

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Data Availability

Datasets related to this article can be found at https://doi.org/10.6084/m9.figshare.21724682.v2, an open-source online data

repository hosted at figshare (Wu et al., 2022).

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Figure captions

Figure 1. (**a**) A global map showing >0.8 Ga blueschist- and low-T/P eclogite-facies rock records (data extended from Brown & Johnson, 2019). (**b**) A simplified geological map showing the tectonic position of South China. (**c**) A geological sketch map of the Tianli Schist (modified from Wang et al., 2018). (**d**) A simplified

745 Neoproterozoic stratigraphic column of the Tianli region. JSF, Jiangshan–Shaoxing
746 Fault; QDSO, Qinling–Dabie–Sulu Orogen; LMSF, Longmenshan Fault; ASS,
747 Ailaoshan-Song Ma suture. Fm., formation.
748
749 **Figure 2.** (**a–c**) Field photographs of the Tianli Schist showing the deformed quartz
750 vein and mica schists. (**d**) Mineral maps of the whole thin section determined by
751 micro-XRF. Mineral abbreviations: Dol, dolomite; Ph, phengite; Qz, quartz; Ab,
752 albite; Bt, biotite.
753
754 **Figure 3.** Micrograph and backscattered electron (BSE) images of the Tianli Schist.
755 **(a–b)** Peak mineral assemblages of biotite, dolomite, phengite, and quartz. **(c–d)**
756 Retrograde chlorite and K-feldspar surrounding biotite and albite, respectively. **(e-f)**
757 Accessory minerals including apatite, monazite, magnetite, pyrite, rutile, tourmaline,
758 and zircon. Mineral abbreviations: Ab, albite; Ap, apatite; Bt, biotite; Chl, chlorite;
759 Dol, dolomite; Kfs, K-feldspar; Mag, magnetite; Mnz, monazite; Ph, phengite; Py,
760 pyrite; Qz, quartz; Rt, rutile; Tur, tourmaline; Zrn, zircon.
761
762 **Figure 4**. (**a**) Si content (per formula unit) and X_{Mg} (= molar Mg/[Mg + Fe^{2+}]) of
763 phengite from the Tianli Schist. (**b**) Muscovite–Celadonite–Pyrophyllite ternary plot
764 of phengite compositions from the Tianli Schist. Integrated muscovite data of
765 greenschist and blueschist from the literature (Table S1) are marked as green and
766 yellow.

Figure 5. (**a**) U–Th–Pb concordia diagram of monazite from the Tianli Schist using IsoplotR after common Pb correction (Vermeesch, 2018). Insets are representative cathodoluminescence images of monazite with yellow circles showing the analytical spot of U–Th–Pb dating. (**b**) Sm–Nd errorchron of monazite, muscovite, and whole rock from the Tianli Schist. Error bars and uncertainties are ±2σ.

Figure 6. P–T pseudosection modeling for the Tianli Schist in the NCKFMASTHC system. Isopleths of Ti-in-biotite (green), Si-in-phengite (red), and Zr-in-rutile (blue) compositions for constraining the peak P–T conditions (yellow area) are marked. Mineral abbreviations: Ab, albite; Arg, aragonite; Bt, biotite; Cal, calcite; Dol, dolomite; F, H_2O–CO_2 fluid; Gln, glaucophane; Kfs, K-feldspar; Omp, omphacite; Ph, phengite; Pl, plagioclase; Qz, quartz; Rt, rutile; Ttn, titanite.

Figure 7. (a) Blueschist- and eclogite-facies metamorphic rock records (data extended from Brown & Johnson, 2019). The lavender bar highlights the time interval of Earth's middle age (Cawood & Hawkesworth, 2014). **(b)** Normalized seawater Sr isotope curve (Shields, 2007) and reconstructed thickness of magmatically active continental crust (Tang et al., 2021). The dark grey bar and shadow show the time intervals of Nuna and Rodinia supercontinent assembly (Li, Bogdanova, et al., 2008; Pourteau et al., 2018). The blue and light grey bar represent the time window of Shuangxiwu arc magmatism and the Grenville orogen during Rodinia assembly (Li,

789 Bogdanova, et al., 2008; Li et al., 2009).