

Middle Neoproterozoic (Tonian) polar wander of South China: Paleomagnetism and ID-TIMS U-Pb geochronology of the Laoshanya Formation

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16 **Key Points:**

- 17 • High-precision ID-TIMS U-Pb dating yields deposition ages of the Laoshanya
18 Formation (Yangjiaping, South China) between 809 and 804 Ma.
19 • Discordant paleomagnetic directions in the red beds are a conflation of three
20 temporally distinct field records residing in hematite.
21 • Our results do not support rapid true polar wander or abnormal magnetic field
22 geometry around 805 Ma.

Abstract

Paleomagnetic records of middle Neoproterozoic (820-780 Ma) rocks display high amplitude directional variations that lead to large discrepancies in paleogeographic reconstructions. Hypotheses to explain these data include rapid true polar wander, a geomagnetic field geometry that deviates from a predominantly axial dipole field, a hyper-reversing field (>10 reversals/Ma), and/or undiagnosed remagnetization. To test these hypotheses, we collected 1057 oriented cores over a 85 m stratigraphic succession in the Laoshanya Formation (Yangjiaping, Hunan, China). High precision U-Pb dating of two intercalated tuff layers constrain the age of the sediments between 809 and 804 Ma. Thermal demagnetization isolates three magnetization components residing in hematite which are not time-progressive but conflated throughout the section. All samples possess a north and downward directed component in geographic coordinates at temperatures up to 660°C that is ascribed to a Cretaceous overprint. Two components isolated above 660°C reveal distinct directional clusters: one is interpreted as a depositional remanence, while the other appears to be the result of a mid-Paleozoic (460-420 Ma) remagnetization, which is likely widespread throughout South China. The high-temperature directions are subtly dependent on lithology; microscopic and rock magnetic analyses identify multiple generations of hematite that vary in concentration and distinguish the magnetization components. A comparison with other middle Neoproterozoic paleomagnetic studies in the region indicates that the sudden changes in paleomagnetic directions, used elsewhere to support the rapid true polar wander hypothesis (ca. 805 Ma), are better explained by mixtures of primary and remagnetized components, and/or vertical axis rotations.

Plain Language Summary

Paleomagnetic directions recorded in 820-780 million year old rocks from South China exhibit large amplitude changes that vary rapidly, which have been interpreted to indicate extraordinarily fast motion of Earth's crust and mantle, up to 90° within a five million year span, with respect to the spin axis of the core. This hypothetical phenomenon, called rapid true polar wander, could be responsible for dramatic global environmental change at that time. To test this theory, we collected over 1000 samples from a well exposed section where the incongruous directions are found. Our measurements suggest that some of the rocks acquired a new magnetic signal during a pervasive remagnetization event in South China around 440 million years ago, long after original deposition of the rocks. New hematite growth has a demagnetization spectrum that partially overlaps or completely obscures the original magnetic signal, which was previously unrecognized. This implies that rapid true polar wander is likely an artefact of magnetic overprinting in ancient rocks from South China. Our results suggest that South China was in a relatively stable position at high latitudes 809-804 million years ago. We find no evidence to support exceptionally fast continental drift or an abnormal geomagnetic field geometry during that time.

1 Introduction

Magnetic directions preserved in rocks and sediments can be used to estimate past positions of the geomagnetic pole, known as virtual geomagnetic poles (VGPs). A sequence of rocks spanning several million years in time may record progressive changes in VGP position, which can be used to construct an apparent polar wander path (APWP). APWPs represent the combined effect of tectonic plate motion and the net rotation of the mantle and crust with respect to Earth's rotational pole, called true polar wander (Evans, 2003). A key assumption is that Earth's magnetic pole aligns with the geographic (spin) axis, which is known as the geocentric axial dipole (GAD) hypothesis. If the GAD hypothesis is valid, then APWPs can be used to determine past locations of tectonic

72 plates and reconstruct cycles of continental assembly and breakup; globally correlated
73 APWPs can be used to estimate rates of true polar wander.

74 A key challenge in paleomagnetism is to build accurate APWPs back into the
75 Precambrian (Kodama, 2013). Of particular interest is the Neoproterozoic Era, which saw
76 the end of prolonged environmental stability between 1.8 – 0.8 Ga, a period referred to as
77 the Boring Billion (N. M. Roberts, 2013), as well as intermittent episodes of Snowball
78 Earth and extreme greenhouse conditions (Hoffman et al., 1998), culminating in the
79 Cambrian explosion and diversification of complex life after ca. 539 Ma (Cohen et al.,
80 2013, version 2023/04). An explanation of Earth’s paleogeography and unique climatic
81 behaviour in the Neoproterozoic is crucial for discerning drivers of long-term climate
82 change and biological evolution. Central to this challenge is unravelling the assembly of
83 the Rodinia supercontinent prior to ca. 900 Ma and its break-up after ca. 750 Ma
84 (Z. X. Li et al., 2008), but Rodinia’s configuration is still debated. Torsvik (2003)
85 described Rodinia as ‘a jigsaw puzzle where we must contend with missing and faulty
86 pieces and have misplaced the picture on the box’. The South China Block (SCB)
87 constitutes one of the most enigmatic puzzle pieces, whose Neoproterozoic history has
88 plagued efforts to build a cohesive model of Rodinia (Evans et al., 2016; Cawood et al.,
89 2020; Jing et al., 2021; Park et al., 2021).

90 Resolving the position and orientation of the SCB in the Neoproterozoic, particularly
91 around 820 – 800 Ma, is complicated by the hypothesized occurrence of one or two rapid
92 true polar wander (TPW) events which may have rotated Earth’s mantle and lithosphere
93 55 – 90° with respect to the spin axis (e.g., Z. X. Li et al., 2004; Niu et al., 2016; Jing et
94 al., 2020). Large-scale oscillating TPW events (e.g., Fig. 1a), known as inertial
95 interchange true polar wander (IITPW), have been postulated to resolve discordant
96 paleomagnetic directions from other continents and are linked to dramatic global
97 environmental change observed around the same time (Kirschvink et al., 1997). Sudden
98 directional changes observed in South China have been compared with paleomagnetic
99 observations from Svalbard (Maloof et al., 2006), and speculatively correlated with the
100 onset of a global negative carbon isotope anomaly, the Bitter Springs Stage, known as the
101 Bitter Springs TPW hypothesis (Swanson-Hysell et al., 2012).

102 The debate over South China’s position in Rodinia is convolved with the Bitter Springs
103 TPW hypothesis, as paleomagnetic data from the SCB undergo rapid paleolatitude
104 variations between 820 and 780 Ma. Various reconstructions of Rodinia place the SCB in
105 a central, marginal or completely isolated position (Fig. 1b). Classic ‘missing link’ models
106 place the SCB at a low latitude between Laurentia and Australia (e.g., Z. X. Li et al.,
107 2008), while other models such as SWEAT (Southwest US – East Antarctica) place the
108 SCB adjacent to India and/or NW Australia, allowing for a direct fit between Australia
109 and Laurentia (Yao et al., 2019). It is also possible that South China was entirely
110 disconnected from Rodinia and occupied a position near the north pole (e.g., Park et al.,
111 2021). Increasingly, models appear to challenge a SWEAT configuration in the
112 Neoproterozoic, although more and more studies suggest that Australia, India and South
113 China were geographically close through the late Cryogenian to Ediacaran (S. Zhang et
114 al., 2021, and references therein).

115 Maloof et al. (2006) and Swanson-Hysell et al. (2012) suggested that testing of the Bitter
116 Springs TPW hypothesis necessitates robust paleomagnetic and geochronologic studies of
117 continuous sedimentary sequences from several cratons. If the rapid TPW hypothesis is
118 valid, then all strata that span the Bitter Springs Stage [ca. 810 – 800 Ma Halverson et
119 al. (2022)], should show similar excursions in their isotopic and paleomagnetic signals.
120 Resolving these discordant paleomagnetic directions, currently observed only in Svalbard
121 and South China, calls for integrative studies and is an issue of ‘major geodynamic
122 significance’ (Evans et al., 2016). Raub et al. (2015) notes that the TPW events
123 hypothesized by Maloof et al. (2006) are not directly dated so cannot be precisely
124 correlated with data from South China. This allows for considerable flexibility in

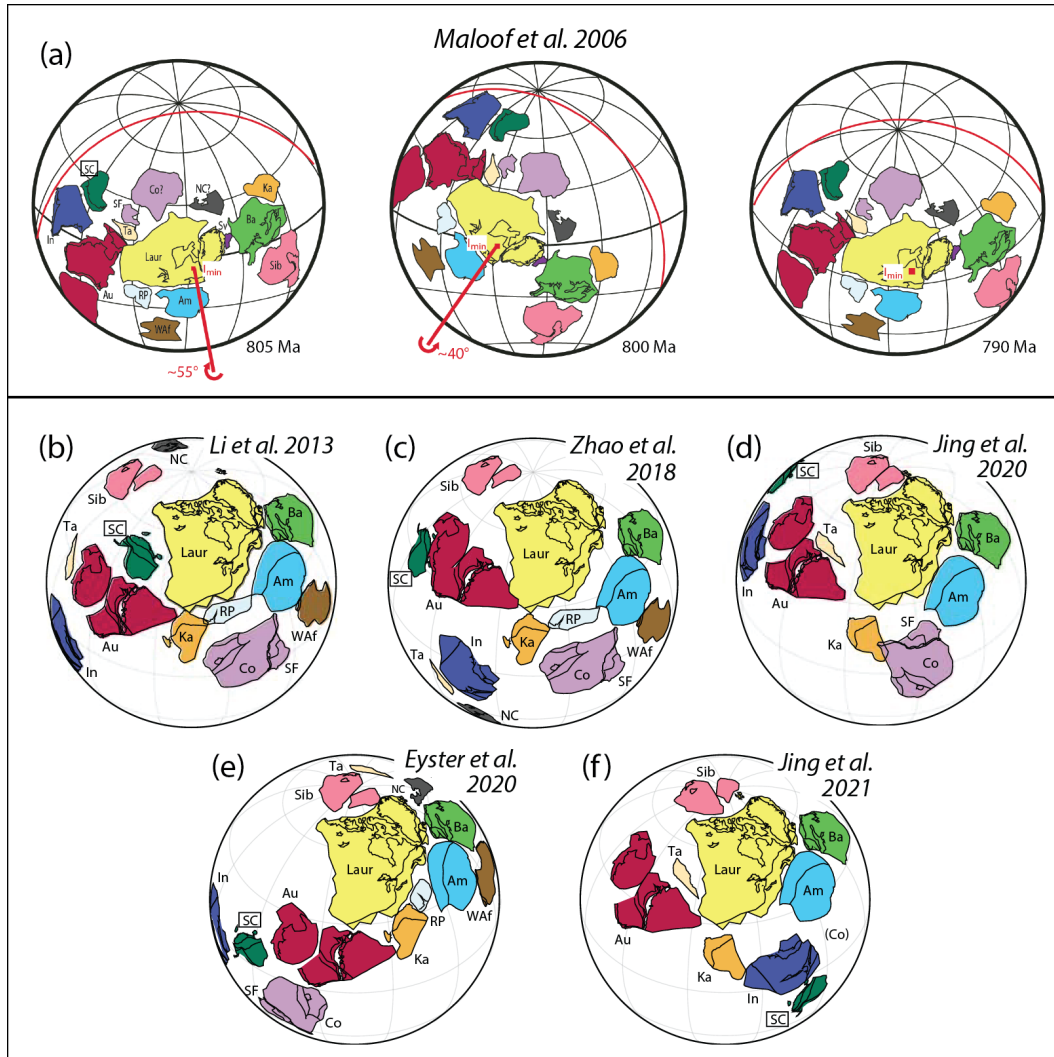


Figure 1. (a) Paleogeographic reconstructions of Rodinia during a hypothesized rapid true polar wander (TPW) event between 805 and 790 Ma; the TPW rotation axis (I_{min}) is marked in red (after Maloof et al., 2006). (b-f) Proposed reconstructions of Rodinia with Laurentia fixed in present North American coordinates, modified from Evans (2021). South China occupies completely different positions in each reconstruction (see Z. X. Li et al., 2013; Zhao et al., 2018; Jing et al., 2020; Eyster et al., 2020; Jing et al., 2021). Am = Amazonia, Au = Australian cratons including Mawsonland, Ba = Baltica, Co = Congo, In = India, Ka = Kalahari, Laur = Laurentia, NC = North China, RP = Rio Plata, SC = South China, SF = Sao Francisco, Sib = Siberia, Sv = Svalbard, Ta = Tarim, WAF = West Africa.

125 paleogeographic models around 800 Ma, so the TPW hypothesis and the precise
 126 configuration of Rodinia ‘must await better constraints’ (Raub et al., 2015).

127 Other plausible explanations for South China’s discordant paleomagnetic data include
 128 rapid tectonic plate motion or non-dipolar to hyper-reversing geomagnetic fields driven by
 129 nucleation of Earth’s inner core (e.g., Driscoll, 2016). Peculiar APWPs could also be an
 130 artefact of undiagnosed remagnetization. For example, Q. R. Zhang and Piper (1997)
 131 warned that South China experienced several episodes of tectonic reworking and suffered
 132 extensive sub-tropical weathering, thereby highlighting challenges to paleomagnetism,
 133 especially going far back in time. Indeed, numerous studies from South China are based
 134 on red sedimentary rocks that have complicated magnetization histories; hence, caution is
 135 required when interpreting paleomagnetic signals in red beds.

136 The Yangjiaping (Hunan Province, China) area has a fairly complete geologic record from
 137 the Neoproterozoic through the Paleozoic, including Marinoan tillites and cap carbonates
 138 (Macouin et al., 2004). Below the tillites lie the Laoshanya Formation red beds, which
 139 were previously dated at Yangjiaping to 809 ± 16 Ma (Yin et al., 2003). There, we
 140 collected 1057 paleomagnetic cores in a 85 metre-thick section, freshly exposed along a
 141 river, as well as 14 cores in the underlying Lengjiayi Group. We present a paleomagnetic
 142 study of those samples, combined with rock magnetic, mineralogical and microscopic
 143 analyses, as well as ID-TIMS U-Pb zircon geochronology of intercalated tuffs, in order to
 144 shed light on Neoproterozoic geomagnetic field behaviour and the paleogeography of
 145 South China around 820 – 800 Ma.

146 **2 Geological setting**

147 **2.1 Nanhua rift basin**

148 The South China Block (SCB) consists of two major tectonic elements, the Yangtze
 149 craton to the northwest and the Cathaysia block to the southeast (Fig. 2a). The SCB
 150 assembled between 1000 – 820 Ma in a series of suprasubduction zone arc-backarc systems
 151 and accretionary orogenesis (Cawood et al., 2013), with final assembly/suturing completed
 152 ca. 830 - 816 Ma (Park et al., 2021), an event referred to as the Jiangnan or Sibao
 153 orogeny. There is still controversy over the timing of their amalgamation, but it is
 154 generally accepted that the two blocks formed the SCB when they collided along the
 155 Jiangshan-Shaoxing suture zone (J. Wang et al., 2015, and references therein). After ~ 820
 156 Ma, during the hypothesized breakup of Rodinia, the Nanhua rift basin developed along
 157 the suture zone (S. Zhang et al., 2008), which accumulated sediments through the
 158 Neoproterozoic to early Paleozoic (Charvet, 2013). Several rifting phases likely occurred
 159 in the basin as indicated by bimodal magmatism ca. 830 – 820 Ma; the lower sequences
 160 are dominated by volcanoclastic sedimentary rocks that are interpreted to be continental
 161 rift deposits (J. Wang & Li, 2003).

162 Rifting terminated in the Nanhua basin in the late Ordovician to early Silurian (460 – 400
 163 Ma), perhaps due to inversion of the basin in response to continental convergence (Xu et
 164 al., 2016), an enigmatic episode known by several names including the Caledonian
 165 (Y. Wang et al., 2007), Wuyi-Yunkai (Z. X. Li et al., 2010) and Kwangsi (Zheng et al.,
 166 2020) orogenic event. We refer to this episode as the Wuyi-Yunkai Orogeny. An
 167 explanation for the geodynamic processes that triggered the Wuyi-Yunkai Orogeny is a
 168 persistent problem, although an angular unconformity between South China’s
 169 pre-Devonian and Devonian strata is widespread (Charvet, 2013). Mid-Paleozoic
 170 metamorphic and magmatic rocks, as well as evidence of folding and thrusting, are found
 171 throughout the Jiangnan orogen and Cathaysia block, but the tectonic response to the
 172 Wuyi-Yunkai Orogeny is not well understood within the Yangtze craton (Zheng et al.,
 173 2020).

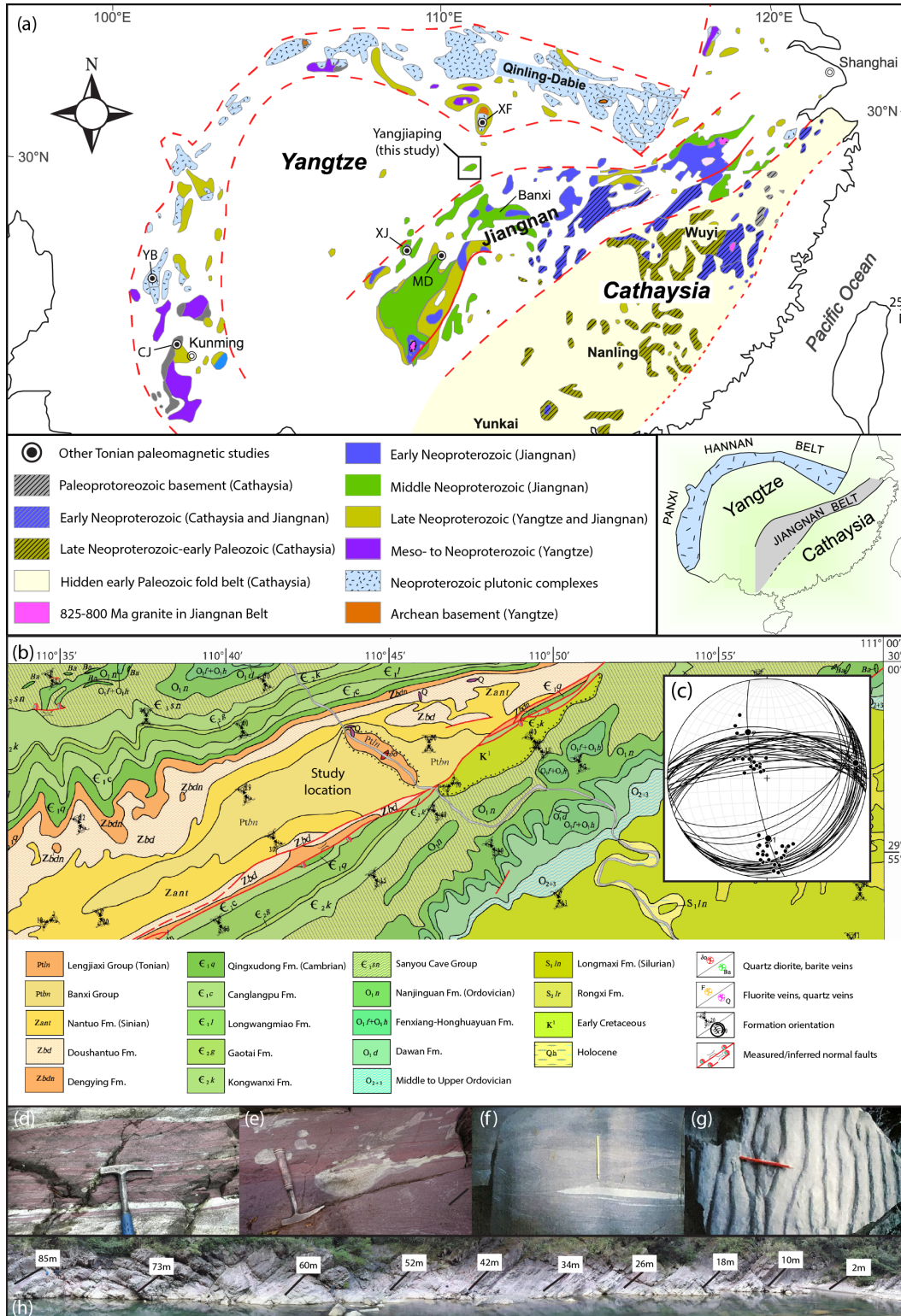


Figure 2. (a) Simplified regional map highlighting the Neoproterozoic in South China (after Cawood et al., 2018). Table S1 lists the Tonian studies (black circles with initials). (b) Geological map of the Yangjiaping area, northwest Hunan (after C. Li et al., 2017). (c) Measured bedding strikes and dips in the area indicates near-cylindrical folding, with a slightly plunging (11°) axis trending 80°. (d-g) Field photos of the Laoshanya Formation showing (d) alternating green-grey and purple-red sandstones, (e) leaching of the red beds (whitish color) that cuts across bedding, (f) cross-bedding, and (g) ripple marks. (h) Composite photo of sampled section with stratigraphic heights.

174 2.2 Banxi group

175 One of the key rifting successions in the Nanhua basin is the Banxi Group siliclastic
 176 sequence (ca. 820 – 730 Ma) that crops out towards the south-eastern edge of the Yangtze
 177 craton and is widely exposed in Hunan Province (S. Zhang et al., 2008). The Banxi
 178 Group is thought to have formed in an intra-continental rift environment and is
 179 characterized by thick sedimentary successions and bi-modal magmatism (Cawood et al.,
 180 2018). The Banxi Group in Hunan Province has been considered equivalent to the
 181 Danzhou Group in northern Guangxi Province (H. Zhang, 1998) and the Xiajiang Group
 182 in eastern Guizhou Province (Lan et al., 2015; Park et al., 2021); however, the age of the
 183 base of the Danzhou Group is poorly constrained (Song et al., 2017).

184 Around the study area, the Banxi Group overlies the Lengjiayi Group with an angular
 185 unconformity (the Wuling unconformity), which crops out throughout the orogen and is
 186 broadly constrained as 830 – 813 Ma (D. Li et al., 2022). In north-central Hunan, the
 187 Cangshuipu Group volcanics were emplaced between the Banxi and Lengjiayi groups.
 188 These volcanics were dated at 824 – 814 Ma (J. Wang et al., 2003; Y. Zhang et al., 2015),
 189 which provide an additional constraint on the lower age of the Banxi Group. The Banxi
 190 Group consists of the Madiyi Formation in the lower part and the Wuqiangxi in the upper
 191 (S. Zhang et al., 2008). The Xihuangshan disconformity (800 – 779 Ma) separates the
 192 Madiyi and Wuqiangxi Formations (D. Li et al., 2022), referred to elsewhere as the
 193 Zhangjiawan and Xieshuihe Formations, respectively (J. Wang et al., 2003).

194 The Madiyi Formation crops out across Hunan and is thought to have formed in a shallow
 195 marine, oxidizing environment. In northwest Hunan, the formation comprises alternating
 196 sequences of distinct dark purple-red mudstones and green sandstones (Fig. 2h). A
 197 regional study of detrital zircons in the Madiyi Formation yielded age peaks (lower limits)
 198 at ~ 808 Ma in the lowermost part, and 803 – 800 Ma in the uppermost part (D. Li et al.,
 199 2022). The Madiyi Formation is characterized as relatively Fe_2O_3 -rich and may act as a
 200 geochemical barrier in the region: $\sim 80\%$ of the gold deposits in northwest Hunan are
 201 hosted in the Madiyi Formation (S. X. Yang & Blum, 1999). The Laoshanya Formation is
 202 synonymous with the Madiyi Formation in Yangjiaping (Yin et al., 2003), although recent
 203 work suggested the Madiyi Formation may be slightly younger (Y. Zhang et al., 2015;
 204 Xian et al., 2020; Park et al., 2021). We use Laoshanya to refer specifically to the outcrop
 205 in Yangjiaping, and Madiyi to refer to the wider formation, as our dating shows them to
 206 be the same age (Section 4.1).

207 2.3 Yangjiaping section

208 The Yangjiaping section lies in the northwestern Nanhua rift basin (Fig. 2b) (X. Zhang et
 209 al., 2000; Song et al., 2017). Yangjiaping is an important landmark for the Precambrian
 210 geology of South China and was selected as a ‘candidate stratotype section’ for the
 211 Nanhua system (Yin et al., 2003). At Yangjiaping, the Laoshanya Formation is 148
 212 metres thick (Yin et al., 2004) and presents a continuous sequence of alternating
 213 purple-red and green-grey sandstones and siltstones (Fig. 2d). For comparison, Madiyi in
 214 the southeast part of the basin may be 3500 metres thick (S. Zhang et al., 2008). The
 215 Laoshanya Formation is well-preserved and shows evidence of a shallow marine,
 216 near-shore or tidal environment, with cross-bedding (Fig. 2f) and ripple marks (Fig. 2g).
 217 Some beds contain sub-vertical quartz and calcite veinlets up to several centimetres wide.
 218 White to pale green ‘mottling’ is ubiquitous (Fig. 2e), often cross-cutting the purple beds,
 219 which is evidence of localized leaching of iron by secondary fluids (Parry et al., 2004).

220 The Laoshanya Formation is conformably overlain by the Xiushuihe (Wuqiangxi)
 221 Formation (Fig. 3), dated at 758 ± 23 Ma by the Sensitive High Resolution Ion
 222 Microprobe (SHRIMP) method (Yin et al., 2003). Twelve metres below the
 223 Xiushuihe/Laoshanya contact lies a tuff whose SHRIMP zircon U-Pb date constrained the
 224 upper age of the Laoshanya Formation to 809 ± 16 Ma (Yin et al., 2003). The Laoshanya

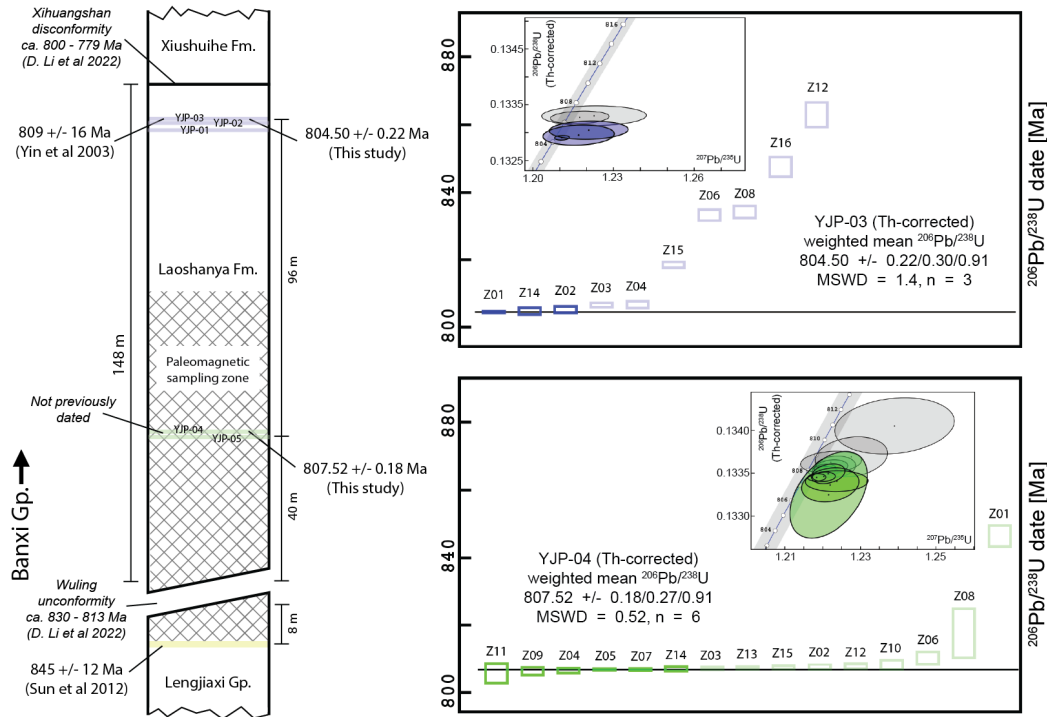


Figure 3. Geochronology of the Laoshanya Formation in Yangjiaping with locations of interbedded tuffs (blue, green and yellow stripes). A five metre thick conglomerate marks the base of our section, below which 14 cores were sampled in the Lengjiaxi Group. Previously published dates are shown on the left side of the column; our new dates (YJP-03 and YJP-04) are shown on the right. Coloured rectangles indicate Th-corrected $^{206}\text{Pb}/^{238}\text{U}$ dates for individual zircons with 2σ analytical uncertainty; transparent rectangles indicate results excluded from weighted means. Corresponding $^{206}\text{Pb}/^{238}\text{U} - ^{207}\text{Pb}/^{235}\text{U}$ concordia diagrams are inset.

225 Formation (strike/dip = $255^\circ/58^\circ$) unconformably overlies the Lengjiaxi Group
 226 ($258^\circ/87^\circ$), dated elsewhere at 860 – 835 Ma (Y. Zhang et al., 2015). A five metre thick
 227 conglomerate distinguishes the unconformity, which demarcates the base of our section.
 228 SHRIMP U-Pb dating of laminated tuff from the underlying Lengjiaxi Group in
 229 Yangjiaping yielded 845 ± 12 Ma (Sun et al., 2012).

230 The Doushantuo Formation (599 ± 4 Ma) in Yangjiaping yielded a single-polarity
 231 magnetization that passed a fold test and was interpreted to be primary, indicating a
 232 low-latitude position for South China in the Ediacaran (Macouin et al., 2004). However,
 233 S. Zhang et al. (2015) remarked that the formation does not contain the pattern of
 234 reversals common for the Ediacaran elsewhere; the paleopole lies close to other early
 235 Cambrian and Silurian poles for South China; and the positive fold test implies only that
 236 the magnetization is pre-folding and not necessarily primary. Macouin et al. (2004)
 237 recognized the similarity of their pole to early Silurian poles, although they considered the
 238 loop in South China's APWP between the Ediacaran and Silurian to be a robust feature
 239 and not an artefact of remagnetization. More detailed rock magnetic data suggested an
 240 episode of magnetite dissolution during early diagenesis (Macouin et al., 2012).

241 South China was shaped by large-scale Mesozoic deformation in response to the collision
 242 of the North and South China blocks that ended by the late Jurassic (Gilder & Courtillot,
 243 1997; Yan et al., 2003). This collision likely produced the folding at Yangjiaping seen in

244 Fig. 2(b) (Gilder et al., 1999). J. Yang et al. (2021) identified angular unconformities
 245 between Middle Triassic and Jurassic strata, and between Cretaceous and pre-Cretaceous
 246 strata 20 – 40 km south of Yangjiaping. At Yangjiaping, the Laoshanya Formation
 247 through the late Ordovician Wufeng Formation [442 ± 8 Ma (Xie et al., 2012)] have
 248 consistent strikes $250 - 280^\circ$ and dips $55 - 70^\circ$, suggesting a lack of significant
 249 deformation until at least the end of the Ordovician. Similarly, no angular unconformity
 250 exists between the Neoproterozoic Danzhou Group and Ordovician sedimentary rocks in
 251 northern Guangxi, ~ 500 km to the south (S. Li et al., 2022).

252 **3 Methods and equipment**

253 **3.1 Sampling campaigns**

254 We collected paleomagnetic samples in September 2019 along a river at Yangjiaping (Fig.
 255 2b) (Shimen County, Hunan Province) opposite the J01 County Road (29.970°E ,
 256 110.732°N). 1057 oriented cores, 2.5 cm in diameter, were drilled over a stratigraphic
 257 thickness of 85 m starting at the base of the Laoshanya Formation (Fig. 3), with a
 258 nominal spacing of 5-10 cm; occasionally we drilled 4 or 5 cores in the same horizon with
 259 each sample separated along strike by 10's to 100's of cm. An additional 14 cores were
 260 drilled over a stratigraphic thickness of 5 m at the top of the underlying Lengjiayi Group
 261 (below the conglomerate). All cores were oriented with sun (when possible) and magnetic
 262 compasses. The median magnetic anomaly from the sun compass was $-3.4 \pm 0.6^\circ$ ($n =$
 263 269), indistinguishable from that expected (-3.7°) from the IGRF2020 model (Alken et al.,
 264 2021) at Yangjiaping; all compass azimuths were corrected accordingly. In August 2021,
 265 five ~ 2 kg samples were collected from suspected volcanic tuff horizons within the
 266 Yangjiaping section for U-Pb dating (Fig. 3).

267 **3.2 U-Pb geochronology**

268 High precision U-Pb zircon dating was performed using chemical abrasion - isotope
 269 dilution - thermal ionization mass spectrometry (CA-ID-TIMS) at the University of
 270 Geneva. Mineral separation involved crushing and milling bulk samples to $<300 \mu\text{m}$,
 271 followed by concentration on a Wilfley table, Frantz magnetic separator and heavy liquids
 272 (methylene iodide, density 3.32 g/ml). Crack and inclusion-free zircon crystals were
 273 hand-picked under a binocular microscope, thermally annealed at 900°C for 48 hours, and
 274 then partially dissolved in concentrated hydrofluoric acid (HF) at 210°C for 12 hours in a
 275 ParrTM bomb vessel (Widmann et al., 2019). The partially dissolved grains were then
 276 extracted and washed in 6N HCl in 3 ml Savillex beakers overnight (min. 12 h) at 80°C .
 277 Individual zircons were washed 4x with 7N HNO₃ in 3 ml Savillex beakers placed in an
 278 ultrasonic bath. Samples were then transferred into 200 μl Savillex microcapsules, mixed
 279 with 1-2 drops of EARTHTIME ²⁰²Pb-²⁰⁵Pb-²³³U-²³⁵U (ET2535) tracer solution (Condon
 280 et al., 2015; McLean et al., 2015) and dissolved with 2-3 drops of concentrated HF in
 281 ParrTM digestion vessels at 210°C for 48 hours. After dissolution, samples were dried at
 282 120°C , re-dissolved in 3N HCl, and then U and Pb were separated using anion exchange
 283 column chemistry. Chemical preparation was carried out in a class 100 clean-air facility.
 284 U and Pb were loaded together on outgassed, zone-refined, Re filaments with a
 285 silica-gel/phosphoric acid emitter solution (Gerstenberger & Haase, 1997), and measured
 286 on either a Thermo Scientific Triton TIMS or an IsotopX Phoenix TIMS.

287 The measured isotopic ratios were corrected for interferences of ²³⁸U¹⁸O¹⁶O on ²³⁵U¹⁶O₂
 288 using a ¹⁸O/¹⁶O composition of 0.00205 based on repeat measurements of the U500
 289 standard. Mass fractionation of U was corrected using a double isotope tracer with a
 290 ²³⁵U/²³³U of 0.99506 ± 0.00005 (1σ). The Pb blank isotopic composition is ²⁰⁶Pb/²⁰⁴Pb
 291 $= 17.84 \pm 0.37$ (1σ), ²⁰⁷Pb/²⁰⁴Pb $= 15.25 \pm 0.33$ (1σ) and ²⁰⁸Pb/²⁰⁴Pb $= 37.05 \pm 0.90$
 292 (1σ), based on total procedural blank measurements. Repeat analyses of the ET100
 293 solution (²⁰⁶Pb/²³⁸U date: 100.173 ± 0.003 Ma; Schaltegger et al., 2021) yielded a value

294 of 100.1678 ± 0.0046 Ma (MSWD = 3.2, n = 32/40). All zircon $^{206}\text{Pb}/^{238}\text{U}$ dates were
 295 corrected for ^{230}Th - ^{238}U disequilibrium using a $\text{Th}/\text{U}_{\text{magma}}$ of 3.5 ± 1.0 .

296 To determine the U-Pb age of each tuff, we take the youngest cluster of ≥ 3 analyses,
 297 assuming the zircons crystallised until the time of eruption and all Pb-loss is mitigated by
 298 the chemical abrasion procedure, in line with previous U-Pb ID-TIMS studies which show
 299 that youngest date clusters are generally a reliable measure for eruption age (e.g.,
 300 Schaltegger et al., 2015; Edward et al., 2023). Older dates are assumed to be from
 301 recycled zircons or associated with inherited cores, wall rock contamination, and/or
 302 prolonged growth in magma chambers. We report weighted mean U-Pb age uncertainties
 303 at the 2σ level in the format $A \pm X/Y/Z$, where A = weighted mean age, X = analytical
 304 uncertainty, Y = combined analytical and tracer uncertainty, and Z = combined
 305 analytical, tracer and decay constant uncertainty (Schoene et al., 2006).

306 3.3 Paleomagnetism

307 Paleomagnetic samples were measured at the Paleomagnetic Laboratory,
 308 Ludwig-Maximilians-Universität (LMU-Munich). Cores were cut into 2.2 cm high
 309 cylindrical specimens using a non-magnetic saw. Specimens were thermally demagnetized
 310 in a magnetically-shielded, paleomagnetic oven with an accuracy of $\pm 1^\circ\text{C}$ (Volk, 2016).
 311 Remanent magnetizations were measured using a 2G Enterprises superconducting rock
 312 magnetometer via the automated SushiBar system (Wack & Gilder, 2012). Before
 313 treatment and after each heating step, bulk susceptibilities were measured at room
 314 temperature with a Bartington MS2 susceptibility meter. Repeat measurements on a
 315 subset of specimens were measured with an Agico JR6 spinner magnetometer to check for
 316 consistency. 22 specimens underwent stepwise alternating field (AF) demagnetization up
 317 to 90 mT prior to thermal demagnetization, which generally removed $<10\%$ of the natural
 318 remanent magnetization (NRM) so was discontinued. All paleomagnetic measurements
 319 were made inside a 90 m^3 magnetically shielded room with an average residual field <500
 320 nT.

321 3.4 Rock magnetism

322 Based on the paleomagnetic experiments, select samples were subjected to a suite of rock
 323 magnetic investigations. Samples were crushed into powder using a porcelain mortar
 324 (unsieved). Hysteresis parameters and isothermal remanent magnetization (IRM)
 325 acquisition curves in applied fields up to 1.8 T were measured with a Princeton
 326 Measurements Corporation MicroMag 3900 vibrating sample magnetometer (VSM) at
 327 LMU-Munich. Low temperature experiments were performed on the same instrument
 328 down to 77 K with a liquid nitrogen cryostat. Thermal susceptibility curves were
 329 measured up to 700°C in air using a Agico MFK1-FA Kappabridge with a CS-4 furnace
 330 at the University of Oslo. Anisotropy of magnetic susceptibility (AMS) measurements on
 331 99 non-demagnetized core specimens were carried out with an Agico MFK1-FA
 332 Kappabridge at the University of Tübingen, with a peak field intensity of 200 A/m and
 333 an operating frequency of 976 Hz. Anisotropy of anhysteretic magnetic remanence (AMR)
 334 measurements were made on the same 99 specimens using the SushiBar at LMU-Munich
 335 with a peak AF field of 90 mT and a DC bias field of 0.1 mT (Wack & Gilder, 2012).

336 3.5 Mineralogical and microscopic analyses

337 25 samples were selected for mineralogical analyses. Polished thin sections were first
 338 observed using transmitted and reflected light microscopy at LMU-Munich. Scanning
 339 electron microscopy (SEM) analyses were performed with a Hitachi SU5000 Schottky
 340 Field-Emission Gun (FEG) SEM (LMU-Munich), and on core slices polished to $0.5\ \mu\text{m}$
 341 with a Zeiss Ultra55 FEG SEM at the Institut de Minéralogie, Physique des Matériaux et
 342 de Cosmochimie (Paris). Backscattered secondary electron (BSE) images were acquired

343 using an angle selective backscattered detector with accelerating voltages of 15 - 20 kV,
 344 and working distances of 7.5 - 10 mm. Elemental compositions were determined by energy
 345 dispersive X-ray spectrometry (EDXS) using an QUANTAX EDS detector after copper
 346 calibration. Semi-quantification of the spectra was achieved using the ESPRIT software
 347 package (Bruker) and the phi-rho-z method.

348 Results from EDXS elemental composition data obtained by SEM were cross-checked with
 349 XRD measurements to more confidently identify the mineralogy. Four \sim 10 g core samples
 350 were finely ground and homogenized in an agate mortar for X-ray diffraction (XRD) at
 351 the Institut de physique du globe de Paris (IPGP, France). XRD analyses on the powders
 352 were performed using a Panalytical Xpert Pro transmission diffractometer with a copper
 353 anode operated at 45 kV and 40 mA and a slit of 0.5° at a 240 mm radius. The 2θ scan
 354 was performed in continuous mode from 4° to 90° (2θ) with steps of 0.0001° . Peak
 355 identification was performed using pattern search on the X-ray diffraction database of
 356 reference spectra Crystallography Open Database (<http://www.crystallography.net>).

357 4 Results and observations

358 4.1 U-Pb geochronology

359 Samples YJP-01, YJP-02 and YJP-03 lie 12 m below the top of the Laoshanya Formation
 360 (Fig. 3); YJP-01 and YJP-03 are 1.8 m apart. Samples YJP-04 and YJP-05 (0.6 m apart)
 361 lie 40 m above the base of the formation, near the middle of the paleomagnetic sampling
 362 zone, 96 m below YJP-01. Song et al. (2017) measured a stratigraphic thickness of 184 m
 363 for the Laoshanya Formation in Yangjiaping (perhaps along the road); our measurements
 364 on the riverside outcrop match the stratigraphic thickness of 148 m obtained by Yin et al.
 365 (2004).

366 Analyses of 44 individual zircons from the upper three samples (YJP-01 to YJP-03)
 367 yielded dates predominantly between 863 and 804 Ma based on Th-corrected $^{206}\text{Pb}/^{238}\text{U}$
 368 ratios, with minor ($n < 4$) clusters at 1.0, 1.7, 2.0 and 2.4 Ga. Sample YJP-03 yielded the
 369 youngest age plateau at $804.50 \pm 0.22/0.30/0.91$ Ma (MSWD = 1.4, $n = 3$), which we
 370 interpret as the depositional age for this part of the section. YJP-01 and YJP-02
 371 contained many subhedral or rounded zircons that yielded dates between 861 and 807 Ma,
 372 with no distinct plateau, which we interpret as inherited or recycled.

373 We analyzed 29 zircons from the lower sample group. Sample YJP-04 contained hundreds
 374 of pristine euhedral zircons and yielded the youngest age plateau at $807.52 \pm$
 375 $0.18/0.27/0.91$ Ma (MSWD = 0.52, $n = 6$), which is interpreted as the depositional age.
 376 YJP-05 contained large subhedral zircons that yielded scattered dates between 1015 and
 377 818 Ma, with an age plateau at $818.91 \pm 0.18/0.27/0.92$ Ma (MSWD = 0.45, $n = 3$).
 378 Given that YJP-05 is situated only 0.6 m below YJP-04, we consider it unlikely that this
 379 is a true depositional age as it implies a hiatus of 11.4 Myr which is not supported by
 380 field observations. $\text{Th}/\text{U}_{\text{zircon}}$ ratios for YJP-05 are also lower (< 0.75) than those of
 381 YJP-03 and YJP-04 (generally 0.9 to 1.2), suggesting a different provenance or
 382 metamorphic overgrowth, and increased likelihood of inheritance. Moreover, a
 383 depositional age > 814 Ma would be superpositionally inconsistent with the inferred ages
 384 of the underlying Wuling unconformity and the Cangshuipu volcanics.

385 Assuming no significant depositional hiatuses exist, these results yield an accumulation
 386 rate of 32 ± 4 m/Myr (or 3.2 ± 0.4 cm/kyr) for the Laoshanya Formation (with
 387 analytical uncertainty only). Using linear extrapolation (Blaauw & Christeny, 2011), our
 388 findings suggest that the Laoshanya Formation was deposited between 808.7 [median, 95%
 389 range = 809.6-808.0] and 804.2 [median, 95% range = 804.7-803.5] Ma (Fig. S4), and can
 390 be broadly correlated with the lower Madiyi Formation and upper Xiajiang Group. Our
 391 results agree well with the regional compilation of D. Li et al. (2022) who reported age
 392 peaks in detrital zircons at ~ 808 Ma for lower Madiyi and 803 to 800 Ma for upper

Table 1. Summary of paleomagnetic results from Yangjiaping (29.970°N, 110.732°E). n = number of specimens. D and I are declination and inclination in (g) geographic (in situ) and (s) stratigraphic (tilt-corrected) coordinates (°). κ and α_{95} are the precision parameter and 95% confidence ellipse (Fisher, 1953). Pos./neg. rev. test = positive or negative reversal test after McFadden and McElhinny (1990). A, B and C refer to the magnetization component.

Name	n	D _g	I _g	D _s	I _s	κ	α_{95}	Comments
Lengjiaxi Gp.								Strike/dip = 258/87
Lengjiaxi A	14	24.6	58.3	7.8	-22.2	269.3	2.4	Neg. fold test
Laoshanya Fm.								Strike/dip = 255/58
Laoshanya A	1002	24.5	55.9	5.9	4.1	29.3	0.8	Neg. fold test
Laoshanya B	100	33.8	-61.9	133.8	-46.9	15.8	3.7	
Laoshanya C	81	149.4	42.5	30.3	73.8	33.6	2.8	Neg. rev. test
Laoshanya C*	44	150.9	43.4	25.4	74.1	48.3	3.1	Pos. 'B' rev. test

*Filtered for specimens that decay linearly to the origin and have MAD < 10° (Fig. S3).

393 Madiyi. The age of $807.52 \pm 0.18/0.27/0.91$ Ma in the middle (40 m height) of the
394 paleomagnetic sampling zone is assigned to the paleopole in the following sections.

395 4.2 Paleomagnetic results

396 1188 specimens were thermally demagnetized up to 680°C with progressive decreases in
397 temperature steps from 100 to 2°C (Fig. 4). Many specimens demonstrated erratic
398 demagnetization behaviour above 660°C which was accompanied by a large (tenfold)
399 increase in bulk susceptibility, indicating alteration. Three coherent components (A, B
400 and C) were identified using principal component analysis (Kirschvink, 1980), as described
401 below and shown in Fig. 4 with summaries in Fig. 5 and Table 1. Specimens often
402 appeared to contain variable concentrations of all three components, which could be
403 resolved and quantified with vector unmixing analyses (Tonti-Filippini & Gilder, 2023).
404 Some samples also contained a spurious component which demagnetized below 100-200°C
405 and generally aligned with the present day field; this is interpreted to be a viscous
406 remanent magnetization (see Butler, 1992, Chapter 3) and not considered further.

407 4.2.1 Components A, B and C

408 1016 specimens (86%) contained a coherent north and down (in situ coordinates) directed
409 component that unblocked mostly between 100 and 660°C (Figs. 4a and d). This
410 component is referred to as component A and is ubiquitously present in both the
411 Laoshanya Formation (n = 1002) and the Lengjiaxi Group (n = 14). Component A often
412 occurs in combination with a higher temperature component (B or C), although in ca. 6%
413 of the specimens, it is the sole magnetization component. A fold test is negative at 99%
414 confidence limits (McElhinny, 1964), indicating component A was acquired after folding.

415 181 specimens (15%) displayed coherent demagnetization behaviour from 660 to 680°C
416 that decays to the origin on Zijderveld diagrams (Fig. 4b and c). The high-temperature
417 directions form two clusters on a stereonet (Figs. 5b and c), which are referred to as
418 components B (n = 100) and C (n = 81). Component B consists of single-polarity,
419 southeast and up directions in tilt-corrected (stratigraphic, s) coordinates (Fig. 5b).
420 Despite spanning 85 metres of section, component B directions are present only in one
421 polarity.

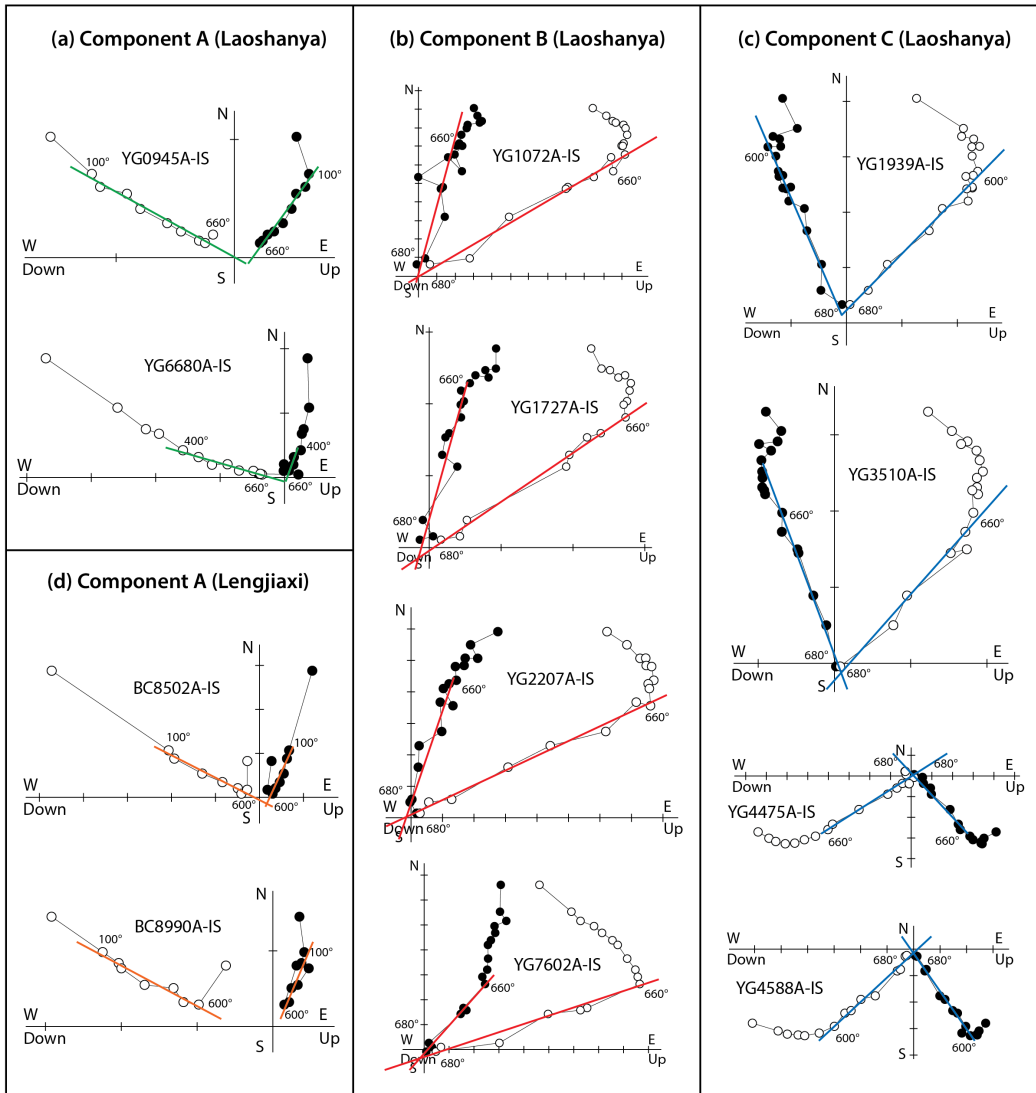


Figure 4. Zijderfeld (1967) diagrams of representative samples (in geographic coordinates) from the three component groups (A-C) from the Yangjiaping section displayed with best-fitting ChRM components not forced to the origin. All tick marks are 10^{-3} A/m. In (a), only component A is present, removed below 660°C . Specimens from component B (b) group generally unblock to the origin between 660 and 680°C with single-polarity directions, after removal of component A. Component C (c) unblocks to the origin between 660 and 680°C and has dual-polarity ChRM directions after removal of component A. (d) Specimens from the Lengjiaxi Group (below the conglomerate) contain component A, but not B or C.

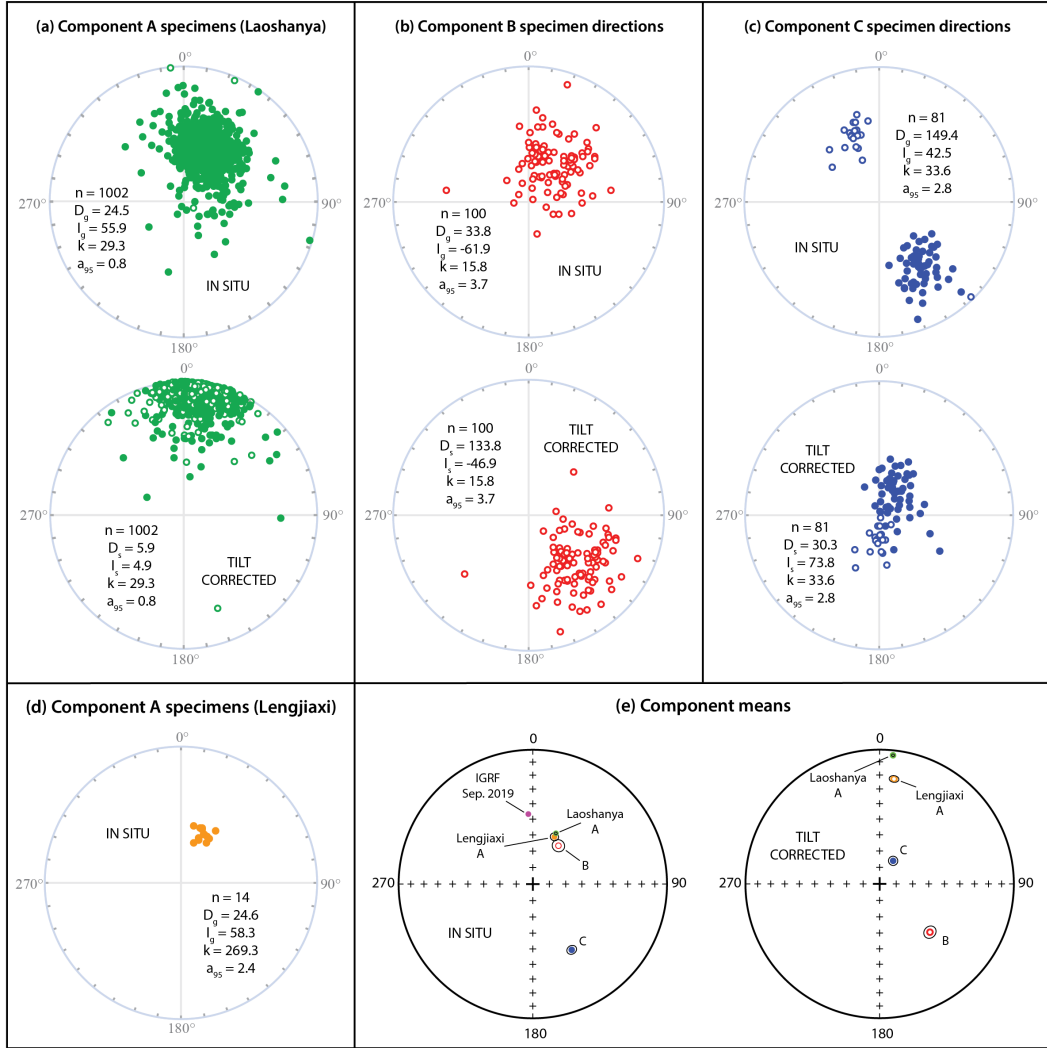


Figure 5. Stereonet plots of component A-C paleomagnetic directions in geographic (in situ) and stratigraphic (tilt-corrected) coordinates. Filled circles are lower hemisphere, open circles upper hemisphere. (a-c) Components A and B have a single-polarity while component C has dual-polarities. (d) Specimens from the Lengjiayi Group only contain component A, indistinguishable to that from the Laoshanya Formation in geographic coordinates. (e) Component mean directions for components A, B and C are distinguishable from the expected IGRF direction (Alken et al., 2021). α_{95} ellipses are shown in black; α_{95} for Laoshanya A component is smaller than the point indicating the direction.

422 Component C ($n = 81$) comprises dual-polarity directions with steep inclinations in tilt
 423 corrected coordinates; 62 samples have positive inclinations and 19 negative (Fig. 5c). A
 424 reversal test (McFadden & McElhinny, 1990) is negative at the 95% confidence level.
 425 However, the reversal test is positive (B classification) if the specimens are filtered for the
 426 directions that decay linearly to the origin with maximum angular deviations (MAD) \leq
 427 10° ($n = 44$, Fig. S3). Large MAD values likely arise from unresolved overlapping
 428 components (B+C). Components B and C were only found in the Laoshanya Formation
 429 so no fold test could be performed at the local level. Below, we show the results of a
 430 regional fold test, which is negative for component B and positive for component C.

431 **4.2.2 Sedimentology**

432 Median bed thickness defines three distinct parts (Fig. 6a). In the lowermost 20 m,
 433 median bed thickness ranges from 0.10 to 0.25 m for sandstones and 0.05 to 0.10 m for
 434 siltstones. This part preserves abundant ripple marks, suggesting a shallow, tidal or
 435 near-shore environment. Between 20 and 54 m, the median thickness of the sandstone
 436 beds increases up to 0.35 m. This part contains graded bedding and some cross-bedding,
 437 indicating more energetic deposition. Median thickness increases towards the top of the
 438 section (54-85 m) for both siltstones (up to 0.35 m) and sandstones (up to 0.4 m) which
 439 could indicate a more stable, perhaps deeper, depositional environment; cross-bedding and
 440 ripple marks are absent. These three parts compare well with stratigraphic units 1-5
 441 described by Yin et al. (2004).

442 We compared the paleomagnetic results against the stratigraphic log (Fig. 6a) to test the
 443 correlation between lithology and magnetization components. Specimens containing
 444 component B arise mainly in the siltstone and fine sandstone beds (59%), with 41% taken
 445 from coarse sandstone beds (Figs. 6b and c). 77% of the specimens having component C
 446 come from coarse sandstones, while 23% come from finer sediments. Component C
 447 appears more frequently in the middle part (20-54 m), while component B is more
 448 abundant in the bottom (0-20 m) and top (54-85 m) parts. The highest density of
 449 specimens containing component C occurs in the thick sandstone beds between 45 and 50
 450 m. This is not an artefact of sampling density as the spacing between specimens was kept
 451 constant at 5-10 cm throughout the section.

452 **4.2.3 Vector unmixing**

453 To quantify component contributions to the NRM, we performed vector unmixing
 454 analyses (Tonti-Filippini & Gilder, 2023) on 1152 specimens from the Laoshanya
 455 Formation (Fig. 7). A goodness of fit (GOF) $\geq 70\%$ was achieved for 640 specimens using
 456 the mean component directions from Table 1 and the median demagnetization curves
 457 derived by Tonti-Filippini and Gilder (2023), i.e., demagnetization data for more than half
 458 the specimens could be accurately modelled by superimposed contributions from
 459 components A, B and/or C. In the following sections, we grouped the specimens by their
 460 dominant component, i.e., the estimated component contribution must be at least $\sim 40\%$
 461 of the total NRM to qualify for that component group. However, 98% of the specimens
 462 contain some combination of two or three of the components.

463 **4.3 Rock magnetic results**

464 Thermal remanence curves for group A samples gradually decay between 100 and 660°C
 465 (Figs. 4a and d), consistent with fine-grained or pigmentary hematite (Collinson, 1974).
 466 Those in groups B and C unblock in a narrow range between 660 and 680°C (Figs. 4b
 467 and c), consistent with crystalline hematite (Özdemir & Dunlop, 2002, 2005). To further
 468 characterize remanence carriers, we carried out rock magnetic experiments on fresh
 469 specimens from cores chosen to best represent component groups A, B, and C.

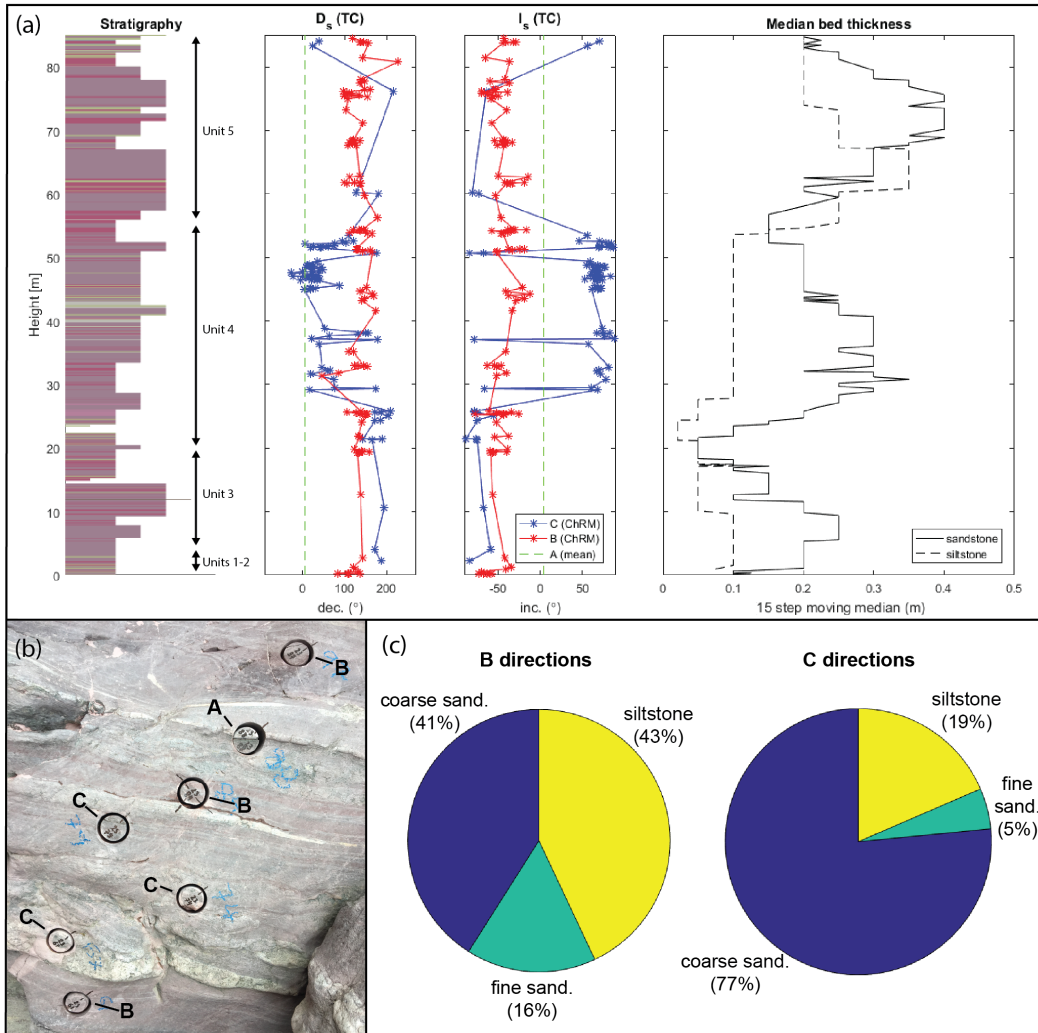


Figure 6. Lithological characteristics of paleomagnetic directions. (a) Graphical representation of the paleomagnetic section (Laoshanya Formation, Yangjiaping), with stratigraphic units identified by Yin et al. (2004). The section is shown alongside the paleomagnetic results (D_s and I_s = declination/inclination in stratigraphic [tilt corrected] coordinates) and median bed thickness by lithology. (b) Field photo of interbedded siltstone/sandstone that contain the components A-C as indicated. (c) Component C was more frequently identified in the coarse sandstone beds (nominal grain size $> 250 \mu\text{m}$), while component B was found more in the fine-grained sandstones (63 to $250 \mu\text{m}$) and siltstones ($< 63 \mu\text{m}$).

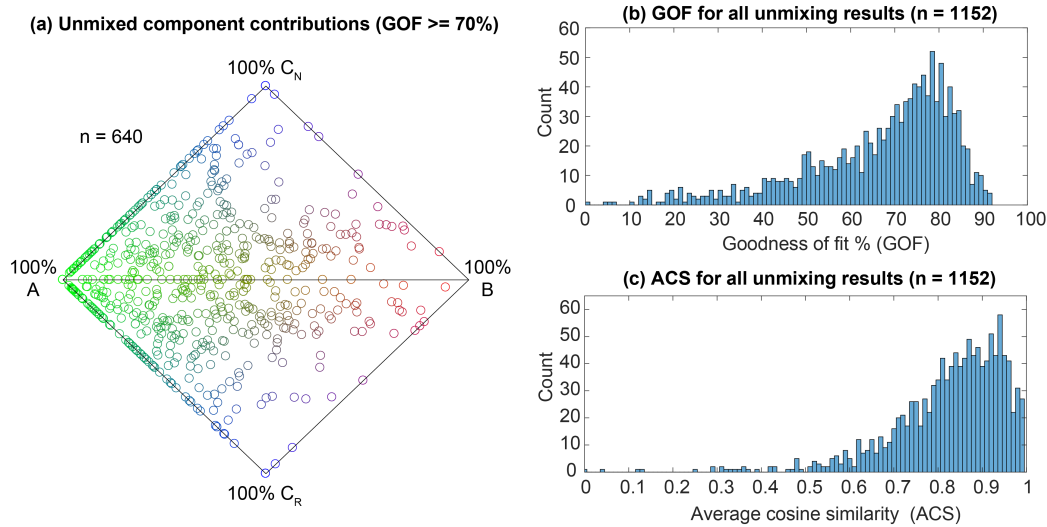


Figure 7. Results of vector unmixing analyses. (a) Specimen unmixing results with goodness of fit (GOF) $\geq 70\%$ ($n = 640$), shaded by component contribution where green = A, red = B and blue = C. Results are shown on a double ternary plot as a proportion of total NRM, unmixed using median component curves derived in Tonti-Filippini and Gilder (2023), representing estimated combinations of components A, B and C_N (positive inclinations) or C_R (negative inclinations). Distributions of (b) GOF and (c) average cosine similarity (ACS) for all Laoshanya unmixing results ($n = 1152$).

470 **4.3.1 High temperature experiments**

471 Fig. 8 shows representative magnetic susceptibility vs. temperature curves (9 out of 25
 472 measured). All samples exhibit a rapid decline in susceptibility between 660 and 700°C,
 473 indicative of the Néel temperature in hematite. Another commonality, although variable
 474 in magnitude, is the existence of a phase with a Curie temperature around 100–200°C
 475 that is non-reversible upon cooling. This could be indicative of titanium-rich
 476 titanohematite, titanomagnetite or goethite.

477 The thermal susceptibility curves can be divided into three categories:

- 478 1. Samples in the first category (Figs. 8a-b, d-e and g) have reversible behavior with a
 479 Hopkinson peak around 660°C, characteristic of single-domain (SD) hematite grains
 480 (Dunlop, 1974).
- 481 2. Samples in the second category (Figs. 8c and h-i) have irreversible behavior upon
 482 cooling below 600°C, with a new or enhanced Hopkinson peak around 560°C, near
 483 the Curie temperature of magnetite. This is indicative of the creation of SD
 484 magnetite at high temperatures, which also explains the increase in bulk
 485 susceptibility above 600°C observed in the thermal demagnetization experiments;
 486 the susceptibility of magnetite is several orders of magnitude higher than hematite
 487 (O'Reilly, 1984).
- 488 3. The third category (Fig. 8f) has steep slopes in the cooling curves around
 489 560-600°C, indicative of magnetite. However, the susceptibility of the cooling
 490 curves may be lower than the heating curves, suggesting some pre-existing
 491 magnetite was oxidized to hematite. At higher temperature, new magnetite was
 492 created, sometimes in lower concentration when the cooling curve is below the
 493 heating curve.

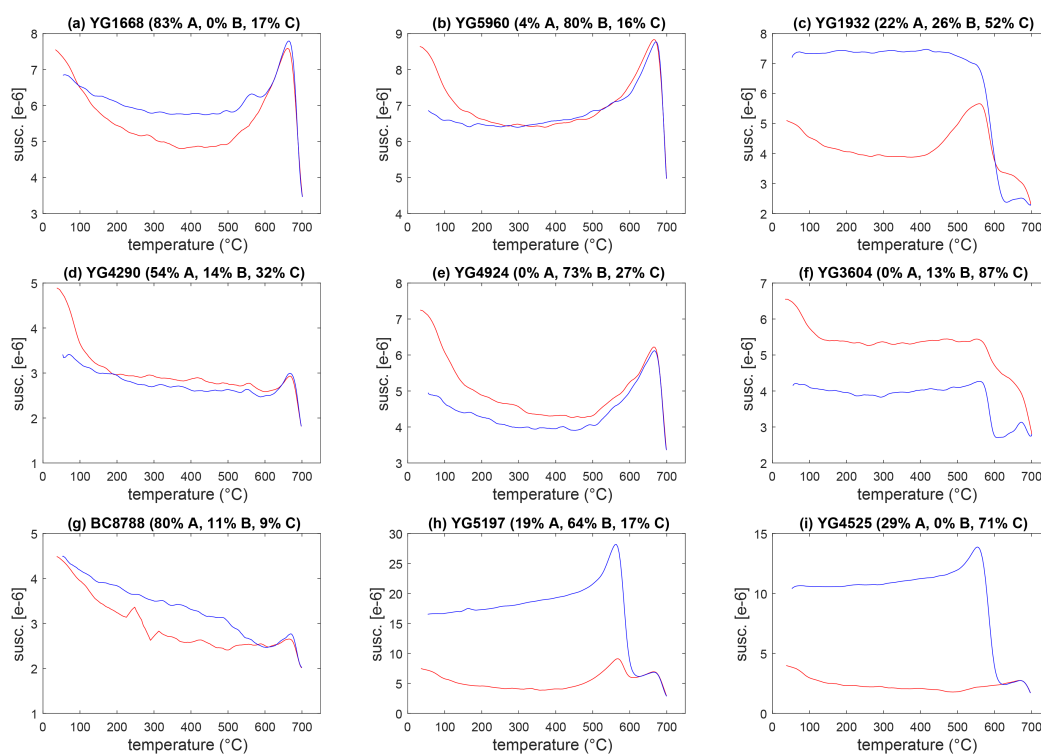


Figure 8. Susceptibility (not mass-normalized) vs. temperature curves for each component group between 20 and 700°C (in air) (red lines indicate heating, blue lines cooling). Component contributions estimated by vector unmixing analysis are shown in the subtitles. Most specimens show a Hopkinson peak around 660°C, with rapid decay up to 700°C, consistent with SD hematite. The Hopkinson peaks appear more pronounced in specimens with a dominant (> 50%) component A (a, d, and g) or B (b, e, and h), compared to those with a dominant component C (c, f, and i). Some specimens, e.g., (c), (e) and (i), exhibit a Hopkinson peak upon cooling around 560°C, likely due to the creation of magnetite during heating. All samples have a phase with a Curie temperature below 150°C that is destroyed by heating, which may indicate the presence of high-Ti titanohematite, titanomagnetite or goethite.

4.3.2 Low temperature experiments

Low temperature experiments were performed on the VSM using a liquid nitrogen cryostat (Fig. S1). A magnetizing field of 1.8 T was applied at room temperature (293 K) then switched off. Magnetic remanence was then measured in a null field down to 100 K and then upon warming to room temperature at a rate of 0.25°/s. Pure hematite typically shows a drop in remanence through the Morin transition (Özdemir et al., 2008) around ~250 K; however, none of the Yangjiaping samples displayed either a measurable Morin transition or a Verwey transition (characteristic of magnetite), at least until 100 K (note that temperature stability of the cryostat decreases approaching 100 K). Dekkers and Linssen (1989) observed similar low temperature behaviour in natural hematite and suggested that impurities (e.g., adsorbed silica) might suppress the Morin transition. Small amounts of Ti (e.g., 1 wt.%) can also suppress the Morin transition (Özdemir et al., 2008), so the absence of a measurable Morin transition is not uncommon. We found no evidence for a Verwey transition in the ca. 100-123 K range indicative of magnetite.

4.3.3 Hysteresis (VSM)

Hysteresis parameters were determined for 22 samples up to 1.5 T (Figs. 9a-b), although hematite can require >10 T to fully saturate (Rochette et al., 2005). Linear corrections (75-90% to 1.5 T) were subtracted to account for paramagnetic contributions. The corrected hysteresis loops for samples with dominant A or B components (Figs. 9c-d) are broad and do not saturate, consistent with single domain (SD) hematite (Özdemir & Dunlop, 2014). Samples in group C (Fig. 9e) show narrow wasp-waisted curves, indicating populations of grains with contrasting coercivity. These wasp-waisted loops are similar to multidomain (MD) hematite behaviour observed by Özdemir and Dunlop (2014). Compositional variations in titanohematite grains could also explain wasp-waisted hysteresis behaviour (A. P. Roberts et al., 1995), as can mixtures of magnetite and hematite.

Samples in group C have bulk coercivities (H_c) an order of magnitude lower and saturation magnetizations (M_s) 2 – 3 times higher than those in groups A and B (Fig. 9f). Values of H_c between 10 and 100 mT in group C imply hematite grain sizes of 10-300 μm (Özdemir & Dunlop, 2014), while values between 300 and 600 mT (for groups A and B) imply smaller hematite grain sizes, on the order of $d = 0.1\text{-}3.0 \mu\text{m}$. However, we note that these numbers pertain to very crystalline hematite, and natural hematite can have a maximum in coercivity at a much smaller grain size. Kletetschka and Wasilewski (2002) estimated the SD to MD grain size transition at 100 μm for hematite, suggesting that the group C hematite is more MD rich than groups A and B. As would also be expected for MD hematite, most samples from group C have only minor or non-existent Hopkinson peaks (Fig. 8).

4.3.4 Anisotropy (AMS and AMR)

We measured the anisotropy of magnetic susceptibility (AMS) on 99 non-demagnetized core specimens (Figs. 10a-e and S2a-c). Anisotropy degrees (P) were <1.1, with maximum axes (K1) trending 235° on average in the horizontal plane (tilt-corrected coordinates). Intermediate (K2) and minimum (K3) axes spread over a NW-SE great circle, with some clustering in the vertical and horizontal planes. AMS directions are not markedly different among the three groups, although group C has a more pronounced tectonic fabric (Fig. 10c). The intermediate tectonic fabrics are typical of sedimentary rocks in thrust-and-fold belts (Saint-Bezar et al., 2002). However, the K2-K3 plane implies a 325° shortening direction (235°-trending fold axis), which differs from the measured strike of our section (255°) by 20° (Fig. 10e).

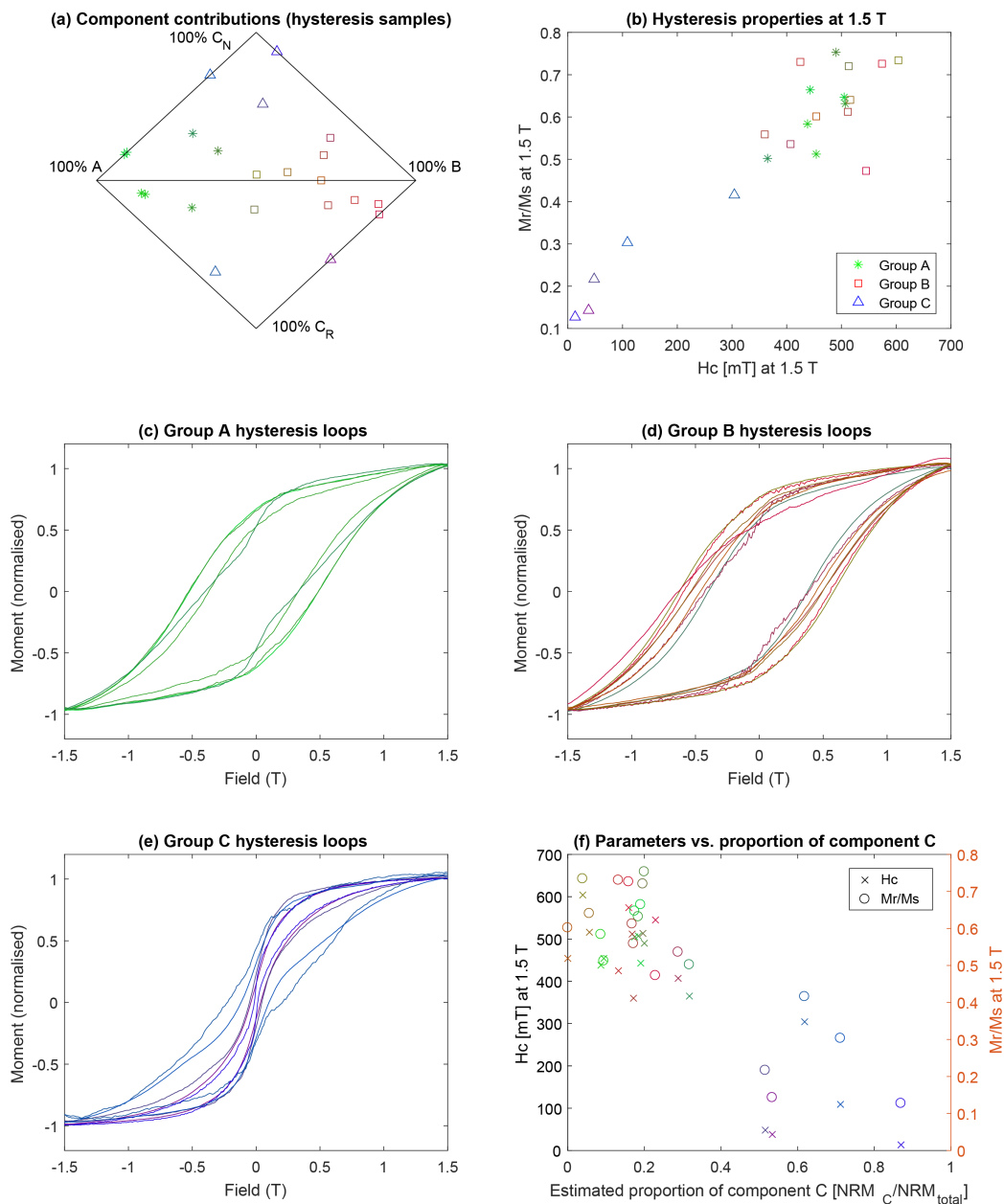


Figure 9. High field experiments. (a) Vector unmixing results for samples used in the hysteresis experiments, with dominant components indicated by blue triangles (C), red squares (B) and green stars (A). The same colour shading is preserved in b-f. (b) Samples with higher Mr/Ms ratios have higher coercive forces (Hc). Lower Hc values can indicate larger hematite grain sizes (Özdemir & Dunlop, 2014), and/or mixtures of magnetite with hematite. (c-e) Room temperature hysteresis curves for specimens from component groups A, B, and C. Groups A and B have wide curves consistent with single domain hematite. Group C has wasp-waisted loops, indicating populations of grains with distinctly different coercivity spectra. (f) Mr/Ms and Hc values (at 1.5 T) correlate negatively with the proportion of component C.

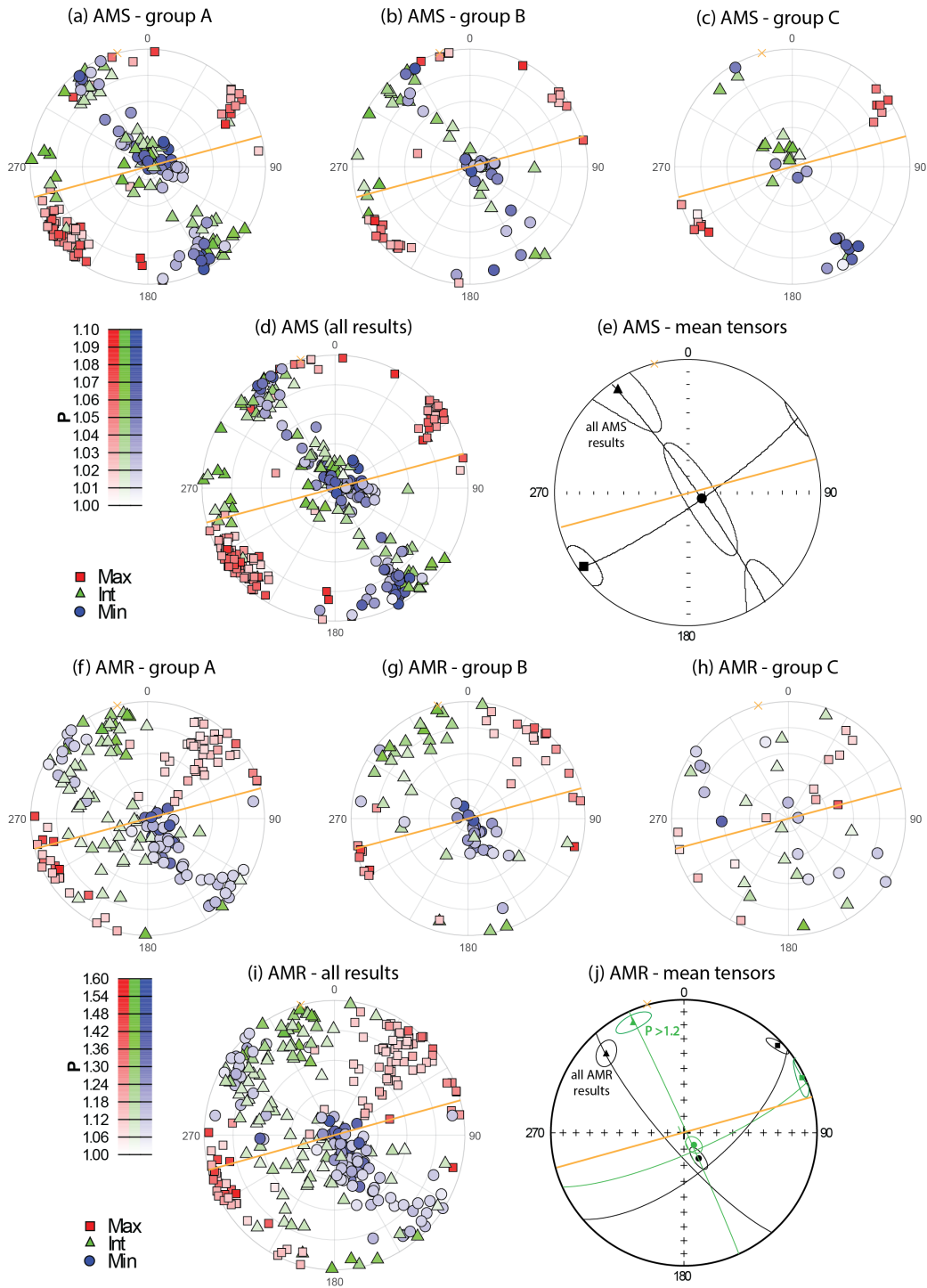


Figure 10. (Caption next page.)

Figure 10. (Previous page.) Stereonet plots of principal axes directions for anisotropy of magnetic susceptibility (AMS) and anisotropy of magnetic remanence (AMR) measured in Laoshanya samples. The strike in Yangjiaping is shown in orange. Symbols are shaded by anisotropy degree (P). Note the different scales between AMS and AMR. (a-c) AMS principal axes directions of the three component groups in tilt-corrected coordinates. Groups A and B have sedimentary fabrics (K3, the minor axis, lies perpendicular to the bedding plane) with varying degrees of a tectonic overprint. Maximum anisotropy axes (K1) lie approximately parallel to strike, while some K3 axes are rotated toward the horizontal along a great circle trending perpendicular to strike. Group C records a strong tectonic fabric, with horizontal K3 directions and K2 (intermediate axis) directions perpendicular to bedding. (d) AMS principal axes directions for all samples. (e) The mean K2-K3 plane defines a shortening direction oriented 325° . (f-h) AMR principal axes directions for groups A and B are generally compatible with those from AMS. Group C has mostly incoherent fabrics. (i) Taken together, the most anisotropic samples have AMR K1 directions that parallel more closely to the fold axis direction than AMS. (j) Mean AMR tensors divided by shape anisotropy. Samples with $P > 1.2$ are more consistent with the regionally-defined compression axis (345°) – all results (black) = 316° , for $P > 1.2$ (green) = 335° .

542 Anisotropy of magnetic remanence (AMR) was measured on the same 99 specimens (Figs.
 543 10f-j and S2d-f) using the refinement method of Wack (2023). Groups A-C have similar
 544 fabrics as AMS, although more scattered (Figs. 10f-h). P values are generally < 1.2 with
 545 mean principal axis directions are similar to those for AMS, with maximum axes (M1)
 546 declinations trending 227° , compared to 235° for AMS (K1). Considering all results
 547 together, 29 specimens have $P > 1.2$ (mean $P = 1.12$ for $n = 99$). Mean M1 axes for those
 548 ($P > 1.2$) specimens trend 245° , 20° clockwise relative to the total population (Figs. 10i-j).
 549 The more anisotropic specimens occur in groups A and B, suggesting they absorbed more
 550 of the tectonic stress that produced the deformation in Yangjiaping. This is not surprising
 551 since finer-grained sediments (Section 4.2.2) with higher clay contents typically absorb
 552 more tectonic strain (Kodama, 2012).

553 4.4 Mineralogy and microscopy

554 X-ray diffraction peaks are dominated by quartz (25-34 %), albite (12-20 %), muscovite
 555 (42-54 %) and chlorite (2-5 %) (e.g., Fig. S5). Hematite is the main iron-bearing mineral
 556 whose concentration varies from 2 to 6 wt.%. Chlorites also contain up to 4 % Fe (EDXS
 557 data). SEM observations and EDXS analyses accord well with the XRD data. SEM
 558 images show that fine-grained micas and silicates (chemically consistent with muscovite
 559 and chlorite) cluster around coarse grains of quartz and alkaline feldspar. Accessory
 560 minerals, such as vermiculite, calcite, apatite, zircon, monazite and rutile, were chemically
 561 characterized by EDXS.

562 SEM images (Fig. 11) show abundant hematite in all samples. Fine, needle-like hematite
 563 flakes, generally $< 1 \mu\text{m}$, are ubiquitous but more dominant in group A samples (Figs.
 564 11a-f). Samples from groups B and C contain large (30–100 μm), Ti-rich hematite and
 565 martite (hematite pseudomorphic after magnetite) grains showing trellis textures of
 566 exsolution typical of Ti-magnetite and/or (hemo-)ilmenites (Figs. 11g-r). Samples from
 567 group B also show abundant 1–2 μm hematite platelets scattered throughout the matrix,
 568 or in dense clusters, which presumably replaced and/or oxidized Fe-rich grains (Figs.
 569 11g-k). Larger grains in group B appear to be more leached, with empty Ti-rich (rutile)
 570 lattices left behind, surrounded by hematite platelets (Fig. 11l). Group C samples

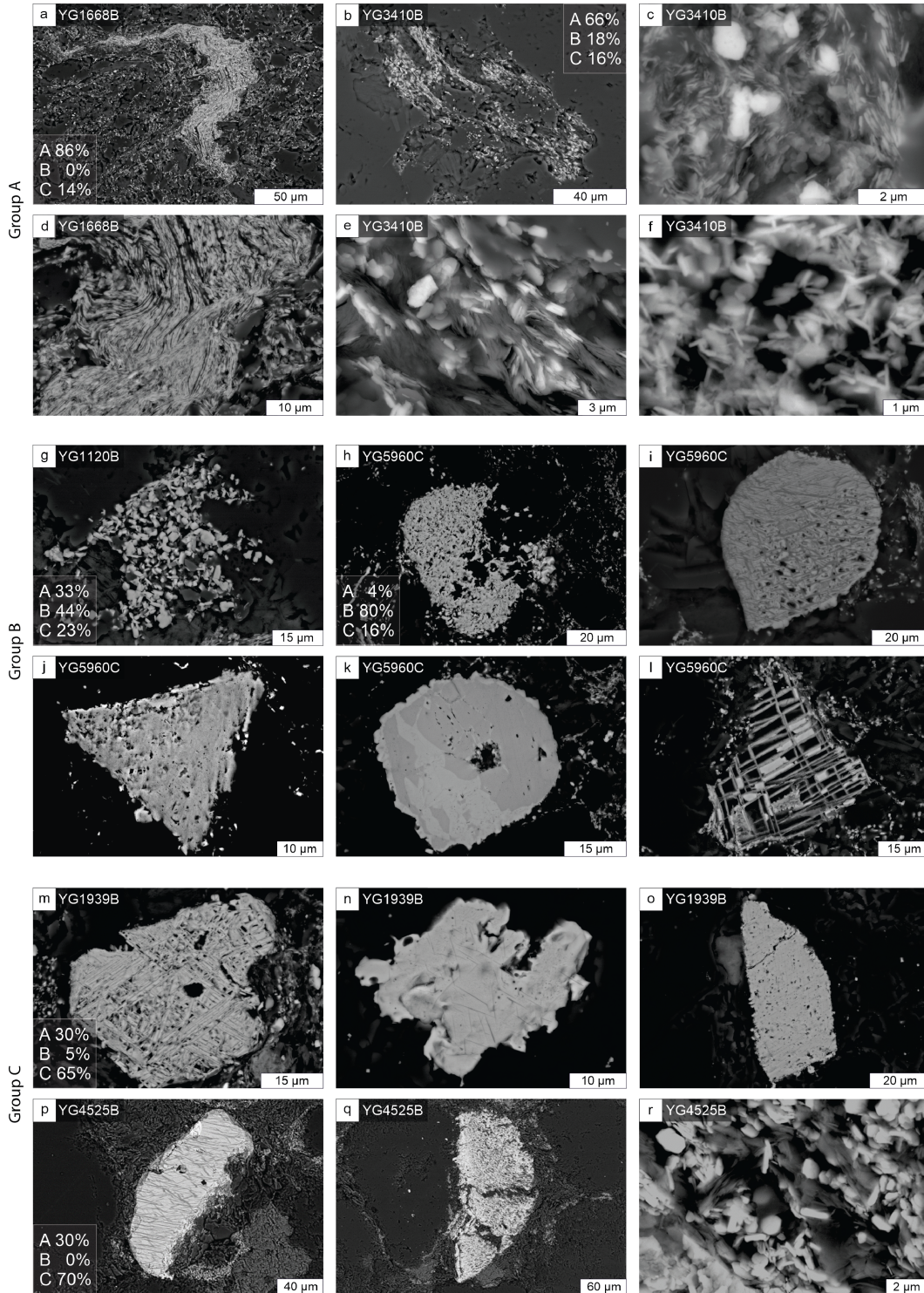


Figure 11. SEM images indicating component proportions estimated by vector unmixing analyses. (a-f) Specimens dominated by component A show pervasive Fe remobilization and fine, needle-like hematite flakes, generally $<1 \mu\text{m}$. (g-l) Component B-rich specimens contain dense clusters of hematite platelets ca. $1\text{--}2 \mu\text{m}$ which appear to have formed in situ, perhaps from the leaching of Ti-rich trellis structures. (m-r) Specimens with a high proportion of component C generally have fewer fine-grained hematite flakes and platelets, and are dominated by large ($30\text{--}100 \mu\text{m}$) Ti-rich hematite and martite grains with exsolution features typical of converted Ti-magnetite and/or (hemo-)ilmenite.

571 generally contain fewer hematite flakes in the matrix than the other sample groups (Figs.
 572 11m-p), consistent with minimal Fe remobilization. Overall, these observations suggest
 573 there was greater Fe remobilization in groups A and B than in group C.

574 **5 Interpretation and discussion**

575 Our experimental results offer several insights into the nature of the remanent
 576 magnetizations in the Laoshanya Formation:

- 577 1. SEM analyses show that group C samples contain abundant detrital (ca. 30 – 100
 578 μm) Ti-rich hematite grains indicative of MD grain sizes, consistent with an igneous
 579 or metamorphic source (Basu & Molinaroli, 1989), and therefore a primary/detrital
 580 remanence (DRM). These samples also contain a reduced presence of $< 1\text{-}2\ \mu\text{m}$
 581 (SD-sized) hematite flakes relative to the samples from groups A and B. Hysteresis
 582 curves for group C samples appear wasp-waisted, indicating distinct magnetic
 583 populations or dominantly MD grains. Thermal-susceptibility curves for group C
 584 also showed minor to non-existent Hopkinson peaks, consistent with MD behaviour.
- 585 2. Hysteresis parameters suggest that samples from groups A and B are dominated by
 586 SD hematite. Thermal-susceptibility curves for groups A and B also show distinct
 587 Hopkinson peaks. The inference of SD hematite is further supported by SEM
 588 observations of abundant sub-micron hematite particles in these samples, consistent
 589 with Fe remobilization and secondary fluid circulation, potentially from the
 590 dissolution of Fe-rich minerals (Walker et al., 1981). The fact that the SEM images
 591 show the SD hematites to be secondary products supports a chemical formation,
 592 and hence, a CRM.
- 593 3. AMR fabrics indicate group A and B specimens may have absorbed more tectonic
 594 stress, with higher anisotropy and an inferred strain axis rotated 20° relative to the
 595 other component groups, consistent with higher clay content/more fine-grained
 596 material.

597 **5.1 Components in the Laoshanya Formation**

598 Of the three magnetization components isolated in this study, component A fails the fold
 599 test between the Laoshanya and Lengjiaxi formations. The corresponding direction at 0%
 600 unfolding is $D = 24.5^\circ$, $I = 55.9^\circ$, $\alpha_{95} = 0.8^\circ$, composed solely of normal polarity, similar
 601 to a widespread overprint component found throughout South China, originally described
 602 by Kent et al. (1987). The corresponding pole at $\text{lat} = 68.6^\circ\text{N}$, $\text{lon} = 176.7^\circ\text{E}$, $dp/dm =$
 603 $0.8^\circ/1.1^\circ$ is near-sided with respect to Cretaceous reference poles for South China. The
 604 unfolding path of component A intersects the Cretaceous poles of Huang et al. (2018) at
 605 10-20% unfolding; however, the reference pole is based mostly on sedimentary rocks,
 606 which can be affected by inclination shallowing. Given that the overprint is a CRM and
 607 the reference poles are based mostly on continental sediments with DRM, one would
 608 expect a CRM to be immune from inclination shallowing, as found by Meng et al. (2022).
 609 Hence, we interpret component A to be a Cretaceous overprint acquired during the
 610 normal superchron [120-83 Ma (Ogg, 2012)].

611 The interpretation of components B and C is more complicated. Taken at face value, both
 612 components could be primary, as they display near-identical thermal remanence
 613 behaviour. Fig. 12 plots the corresponding poles in both geographic and stratigraphic
 614 coordinates, together with a 450 to 120 Ma APWP for South China (Huang et al., 2018).
 615 The component B pole in tilt-corrected (TC) coordinates [LS-B (TC)] lies in proximity to
 616 the 450 Ma segment of the reference APWP. Component B in geographic/in situ (IS)
 617 coordinates [LS-B (IS)] and the TC and IS poles for component C lie far from the
 618 reference curve.

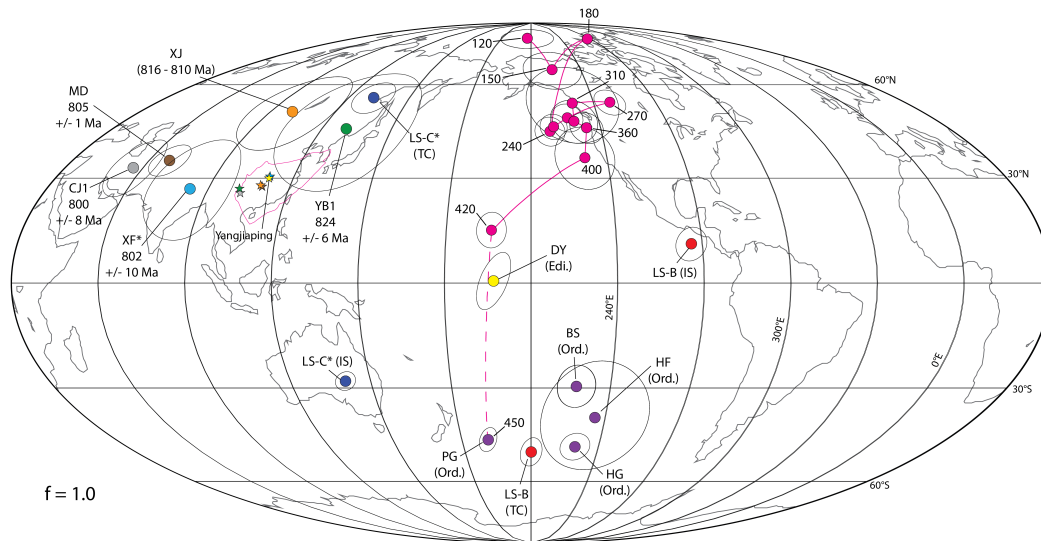


Figure 12. Laoshanya (LS) Formation poles from components B and C in situ (IS) and tilt-corrected (TC) coordinates plotted with the APWP for South China from Huang et al. (2018) shown in pink. LS-C* and Tonian (820-800 Ma) poles are shown in normal (northern hemisphere) polarity for the sake of comparison, although an inverted (southern hemisphere) configuration is also possible (Jing et al., 2021). LS-B is shown in reversed (southern hemisphere) polarity as the LS-B (TC) pole lies close to the 450 Ma segment of the APWP (the line is dashed as there is only one study pole, PG, older than 420 Ma). LS-B (IS) and LS-C* (IS and TC) lie far from the reference path. Table S1 provides details of relevant Tonian poles (820-800 Ma). Ordovician (Ord.) poles are shown in purple (Table S2). The Ediacaran (Edi.) pole from the Doushantuo Formation in Yangjiaping (DY) is shown in yellow. Stars indicate site locations. No shallowing correction was applied for this initial comparison ($f = 1.0$).

619 Fig. 12 also plots our data alongside select Tonian to pre-Devonian poles from the SCB
 620 (Tables S1 and S2, assessed in Supporting Information [SI] Section B). LS-B (TC) lies
 621 among several Ordovician poles for the SCB. Together with the rock magnetic and
 622 microscopic observations (Section 4), we interpret component B to be an Ordovician-aged
 623 chemical remagnetization. This interpretation is consistent with a mineralogical study in
 624 Yangjiaping that indicates the Banxi Group may have experienced very low-grade
 625 metamorphism (ca. 260°C) in the mid-Paleozoic (H. Wang et al., 2014). Estimated
 626 temperatures increase to 360°C towards eastern Hunan and the Jiangnan belt, whose heat
 627 source was linked to the Wuyi-Yunkai (Caledonian) Orogeny (H. Wang et al., 2016).
 628 Moreover, Cawood et al. (2018) suggested that the Banxi Group and its equivalents in the
 629 Jiangnan belt were reworked between 460 – 420 Ma; this timing is supported by
 630 metamorphic and petrogenetic analyses together with $^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb dating across
 631 the Wuyi-Yunkai orogen (Z. X. Li et al., 2010).

632 The component C pole, LS-C* (TC), lies close to Tonian poles from the 824 ± 6 Ma
 633 Yanbian Dykes (Niu et al., 2016) and 816 – 810 Ma Xiajiang Group (Park et al., 2021).
 634 The similarity with other Tonian-aged poles, as well as its dual-polarity, lead us to
 635 interpret component C as primary. Reversal frequency in the Tonian (ca. 807.5 Ma) based
 636 on seven polarity intervals between 20 – 54 m (34 m), where the highest density of
 637 component C directions are found (Fig. 6a), yields 0.21 reversals per metre or 6.7
 638 reversals/Myr (rev/Myr) given an accumulation rate of 32 m/Myr. This is a minimum
 639 value, as the true number could be higher if the signal was obscured by component A or B
 640 overprints. Our estimate for the Tonian is higher than the present rate of 4 – 5 rev/Myr
 641 for the past 5–10 Myr (Ogg, 2012), but is within estimated rates for the rest of the
 642 Phanerozoic, which often exceed 5 – 10 rev/Myr (Torsvik et al., 2021), and the Ediacaran
 643 at 6 – 24 rev/Myr (Meert et al., 2016; Hounslow et al., 2018). For comparison with other
 644 studies, we have grouped the component C directions into ‘sites’ of $n=8-12$ (Table S3),
 645 although we use the filtered specimen mean ($n=44$) to define our pole (LS-C*).

646 5.2 Paleomagnetic poles from South China

647 We assess relevant Tonian (820-800 Ma) poles in Section B (SI). We excluded poles CJ3
 648 (Jing et al., 2020) and YB2 (Niu et al., 2016) because they lack reversals and fail the
 649 statistical requirements of Meert et al. (2020). Poles XJ, CJ1 and MD were derived from
 650 sedimentary rocks; Park et al. (2021) corrected these poles for inclination shallowing with
 651 a blanket flattening factor (f) of 0.6. However, none of these studies applied any E/I or
 652 anisotropy-based methods to justify this factor. We applied the E/I correction method
 653 (Tauxe & Kent, 2004) to Laoshanya component C directions ($n = 81$) but could not
 654 obtain a meaningful result (i.e. our data were labelled ‘pathological’), which may be due
 655 to contamination by the component A and B overprints. Our E/I assessment of all 169
 656 specimen data from Park et al. (2021) corrected the mean inclination from -72.5° to
 657 -75.7° , equivalent to $f = 0.81$, although we recognise that the E/I method may
 658 overestimate shallowing effects if applied to data spanning several sites (e.g., Meng et al.,
 659 2017). Therefore, we find it difficult to justify using $f = 0.6$. To estimate a flattening
 660 factor, we applied a range of f values on inclinations from sedimentary poles (CJ1, MD,
 661 XJ and LS-C*) from 1.0 to 0.6 and compared them to those from igneous poles (XF* and
 662 YB1), keeping the igneous inclinations fixed (Table S4). A value of $f=0.8$ produced the
 663 lowest standard deviation across all poles, so we used this value to correct the
 664 sedimentary poles. This assessment relies on an assumption that all the 820-800 Ma poles
 665 are coeval, which is discussed further below.

666 Fig. 12 shows that the Tonian poles appear to undergo high amplitude and rapid motion
 667 between 820 and 800 Ma, which some workers have previously interpreted as reflective of
 668 a rapid TPW oscillation (e.g., Z. X. Li et al., 2004; Niu et al., 2016; Jing et al., 2020).
 669 However, Park et al. (2021) showed that the arc distances between the ~ 815 and ~ 805
 670 Ma (Xiajiang and Madiyi) poles are much smaller than what would be predicted by the

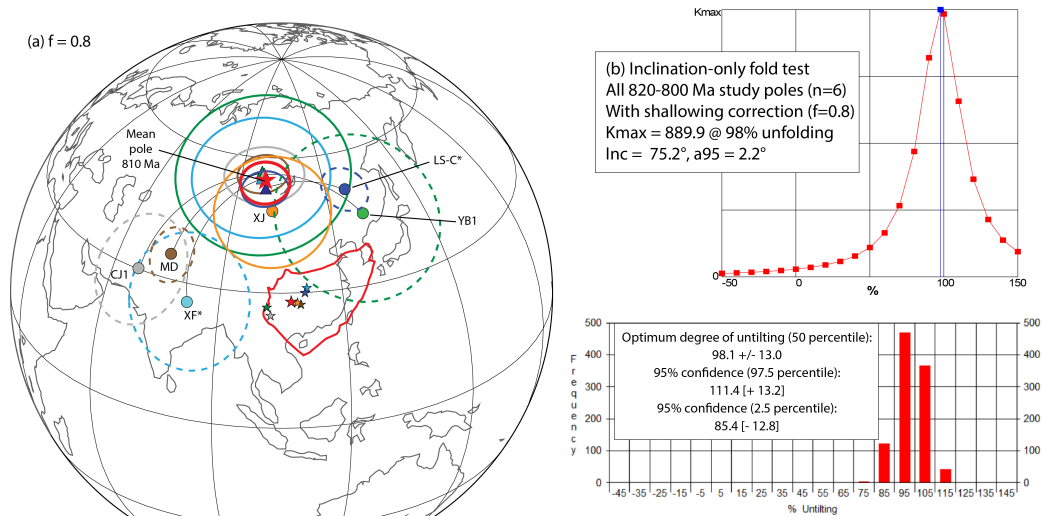


Figure 13. Scenario 1: Can vertical axis rotations explain discordant paleomagnetic data in the Tonian? (a) If allowed to rotate, all Tonian poles (820-800 Ma) collapse to define a group pole ca. 810 Ma. XJ was held fixed as its site location (and pole) lies between all the other sites (and poles). The outline of the SCB and the group mean pole are in red. Dashed circles indicate pole locations before vertical axis rotations. Poles are summarized in Table S1 and Section B (SI). (b) An inclination-only fold test with parametric sampling (Watson & Enkin, 1993; Enkin & Watson, 1996); k maximizes at $98 \pm 13\%$ unfolding.

671 Bitter Springs TPW hypothesis, although they considered that this could be explained by
672 differential plate motion. Park et al. (2021) also described how the data could be
673 interpreted to represent a stable high-latitude position for South China at ca. 825-805
674 Ma, inconsistent with rapid TPW, if the Svalbard data are removed as constraints. Our
675 new data from Laoshanya supports this interpretation, as our ~ 808 Ma pole lies between
676 the Xiajiang and Madiyi poles chronologically, but does not fall on the inferred TPW
677 path. In fact, the pre-Bitter Springs Stage XJ pole and syn-Bitter Springs Stage LS-C*
678 pole overlap within uncertainty after shallowing corrections are applied (Fig. 13).
679 Furthermore, the Laoshanya (808 Ma) and Madiyi (805 Ma) formations are both precisely
680 dated and very similar in age, but the poles are 50° apart, implying extremely rapid
681 motion of ca. $17^\circ/\text{Myr}$, which is an order of magnitude greater than TPW rates estimated
682 for the Phanerozoic (Torsvik et al., 2012). Therefore, in the following sections, we
683 consider alternative hypotheses to explain the data.

684 5.2.1 Scenario 1: Vertical axis rotations dispel TPW

685 All the 820–800 Ma poles for South China show similarly steep inclinations, implying a
686 high paleolatitude for the SCB (Fig. 12). We can be particularly confident in inclinations
687 from sedimentary rocks, whose paleohorizontal can be robustly measured by the bedding
688 attitudes. However, all paleomagnetic poles can be susceptible to uncertainty associated
689 with vertical axis rotations, especially when going far back in time, like in the
690 Precambrian. Vertical axis rotations can lead to poles distributed along small circles
691 centered on the sampling sites, as shown in Fig. 13(a). Therefore, the large changes in
692 declination can be explained without the need for IITPW, if rotations are permitted.
693 Although the interior of South China experienced only minor rotations ($<20^\circ$) since the
694 Cretaceous (Meng et al., 2022), Permo-Triassic rocks exhibit a large variability in

695 declination that matches the rotational amplitude of the Tonian data (Tan et al., 2007;
 696 Gilder et al., 2008). We therefore entertain the possibility in Scenario 1 that the SCB
 697 experienced internal vertical axis block rotations that streaked the Tonian poles out about
 698 a small circle centered on the sampling sites.

699 Under this scenario, all poles are considered to be based on primary, approximately coeval
 700 magnetizations, obtained at similar paleolatitudes. We arbitrarily kept XJ fixed as it
 701 passed a regional fold test and its site and pole lie in the middle of the other sites and
 702 poles. After allowing the other 820-800 Ma poles to freely rotate about their respective
 703 sampling sites, the poles come into close coincidence (Fig. 13a). If the data were primary,
 704 a regional fold test on the directions would be meaningless given the dispersion in
 705 declination, yet an inclination-only fold test would be independent of the differential
 706 rotations. An inclination-only fold test (Enkin & Watson, 1996) yielded a maximum
 707 precision parameter κ_{max} at $98 \pm 13\%$ unfolding with an inclination of 75.2° (Fig. 13b).
 708 Therefore, by accounting for vertical axis rotations, the 820-800 Ma South China poles
 709 can be explained without the need to invoke rapid TPW, extreme plate motion or
 710 anomalous field behaviour. Assuming that Xiajiang is representative of the overall
 711 continent (without vertical axis rotation), we thus calculate a mean pole of $P_{lat} = 54.1^\circ$,
 712 $P_{lon} = 99.0^\circ$ ($A_{95} = 5.0^\circ$, $N = 6$ studies) which we consider representative of the Tonian
 713 for the SCB at ca. 810 Ma.

714 The rotations required by Scenario 1 are quite large (over 90° between CJ1 and
 715 LS-C/YB1), so it is worth considering whether these are realistic with respect to the
 716 tectonics of South China. The structural framework of the Yangtze craton is dominated
 717 by a large curved orocline, readily observed in regional maps and satellite imagery. Tan et
 718 al. (2007) and Gilder et al. (2008) showed that this trend is reflected in paleomagnetic
 719 directions from late Permian to Middle Triassic rocks in South China. Differences in
 720 rotations can exceed 180° among sites when integrating over the past ca. 300 Myr.
 721 Rotation magnitudes could be even more dispersed when integrated over an additional
 722 500 Myr.

723 **5.2.2 Scenario 2: Pervasive Ordovician remagnetization**

724 In Scenario 2, we consider the possibility that remagnetization is pervasive in the SCB.
 725 We fit a small circle running through three, presumably primary Ordovician poles (HF,
 726 HG and BS), centered on their mean site location (Fig. 14a). When plotted in the
 727 southern hemisphere, the ‘Tonian’ CJ1, MD and XF* poles lie near the small circle swath,
 728 suggesting that Ordovician remagnetization may be a common feature that better
 729 explains the origin of their remanences. Indeed, the Madiyi (MD) pole was derived from
 730 silty mudstones, like component B in the Laoshanya Formation. This is curious, as silty
 731 mudstones can carry primary detrital remanences, as revealed through intraclast
 732 conglomerate tests (Tauxe et al., 1980; Opdyke & DiVenere, 2004; Swanson-Hysell et al.,
 733 2019). The Chengjiang (CJ1) study also reports a mid-inclination secondary component
 734 (CJ3) which is offset from the primary component (à la Laoshanya components B and C).
 735 Therefore, an Ordovician remagnetization advocated in Scenario 2 provides an alternative
 736 explanation for the differences in pole positions ca. 805 Ma, whereby some of the
 737 magnetization components are wholly or partially overprinted. Instead of large amplitude
 738 and rapid TPW, there could be mixing of components acquired at distinctly different
 739 times.

740 Remagnetization could have occurred during a mid-Paleozoic deformation event (i.e. the
 741 Wuyi-Yunkai Orogeny). Consistent with this idea, Chang et al. (2022) mapped the Banxi
 742 Group as unconformably overlain by Devonian strata, thereby supporting an
 743 Ordovician-Silurian deformation event. 90 km to the south-east of Yangjiaping, an
 744 angular unconformity occurs at the top of the Middle Ordovician Guniutan Formation
 745 (Schmitz et al., 2010), however, near Yangjiaping, the formations above and below the

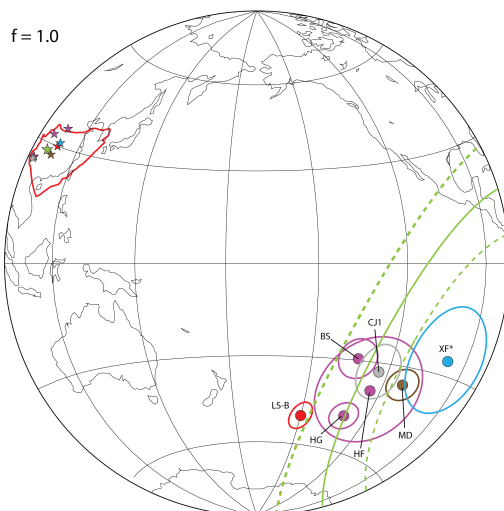


Figure 14. Scenario 2: Can an Ordovician remagnetization account for the discordant directions in South China? When plotted in the southern hemisphere, the CJ1, MD and XF* poles (Table S1) overlap with several Ordovician poles (purple, see Table S2), suggesting they may be affected by Ordovician remagnetization. No corrections for inclination shallowing were applied in this comparison ($f=1.0$). A small circle (green) is plotted through HF, HG and BS poles, centered on the mean site location (PG was excluded from the small circle calculation as it has a significantly different inclination). Dashed lines show 95% confidence limits.

746 Guniutan Formation have similar bedding orientations. Chen et al. (2014) described a
 747 significant hiatus between the Ordovician Wufeng and Silurian Longmaxi (Lungmachi)
 748 formations in Zhangjiajie, 100 km south of Yangajiaping. Zheng et al. (2020) dated the
 749 onset of this hiatus to 447 ± 1.4 Ma using rhyolitic tuffs in the top of the Wufeng
 750 Formation, 30 km to the north of Yangajiaping, and linked the hiatus to
 751 deformation/uplift caused by the Wuyi-Yunkai Orogeny. An Ordovician age for the
 752 deformation and remagnetization is consistent with the apparent age of component B in
 753 the Laoshanya Formation, as its pole overlaps several Ordovician poles.

754 5.3 Summary of discussion

755 We show that rapid pole variations postulated for ca. 820-800 Ma can be explained by
 756 vertical axis rotations and/or regional remagnetization in South China, without invoking
 757 extreme plate motion, rapid TPW or abnormal field geometry. South China's poles for
 758 this time are distributed along a small circle centered on the mean site location and can
 759 be brought into close alignment if vertical axis rotations are permitted. Furthermore,
 760 some poles derived from Tonian-aged rocks lie within a swath of Ordovician poles (when
 761 plotted in the same hemisphere), leading us to conclude that paleomagnetic data from
 762 other Madiyi-equivalent formations may reflect a regional overprint, not primary
 763 remanences. A complicated mixture of primary and secondary magnetizations may have
 764 led some studies to support a rapid TPW (or IITPW) event. Similarity between South
 765 China's Tonian and Ordovician paleolatitudes could be contributing to the confusion, if
 766 an Ordovician remagnetization is widespread.

767 Our results call for careful consideration of paleomagnetic signals in sedimentary
 768 sequences containing diverse lithologies, particularly those with hematite-bearing (red)
 769 mudstones and sandstones. Further work is needed to unravel the response of the Yangtze

770 craton to the Wuyi-Yunkai Orogeny and understand mechanisms for (re)magnetization of
 771 pre-Silurian strata. If the discordant poles are indeed artefacts of a tectono-thermal event
 772 in the Ordovician, then any proposed remagnetization mechanism must be capable of
 773 resetting or overprinting paleomagnetic signals across a range of lithologies and magnetic
 774 mineralogies, as is also true for the Cretaceous overprint component.

775 **6 Conclusions**

776 The Laoshanya Formation in Yangjiaping contains a complicated paleomagnetic record
 777 which offers a unique opportunity to investigate remanence acquisition processes in
 778 hematite-bearing sedimentary rocks. Remarkably, the red beds seem to have preserved a
 779 primary (depositional) Tonian signal (component C), as well as secondary magnetizations
 780 residing in hematite created in the Ordovician (B) and/or Cretaceous (A). From our
 781 study we conclude the following:

- 782 1. Stepwise thermal demagnetization of 1152 samples isolated three end-member
 783 magnetization components (A-C) that combine in a complex and out-of-sequence
 784 manner throughout the 85 m of sampled section. Vector unmixing identifies the
 785 proportions of the three components, which are distinguished by their rock
 786 magnetic characteristics and microscopy, and may respectively represent the most
 787 common hematite phases found in red beds: pigmentary coatings, intergranular
 788 authigenic grains, and detrital specularite grains.
- 789 2. Component A is an overprint acquired during the Cretaceous normal superchron, a
 790 common direction found throughout South China. Thermal remanence curves for
 791 this component are consistent with fine-grained or pigmentary hematite, assumed
 792 to have formed by the oxidation of Fe-rich minerals. This interpretation is
 793 consistent with rock magnetic experiments and microscopic observations of
 794 abundant ($< 1 \mu\text{m}$) hematite flakes and needles in all specimens.
- 795 3. Component B is found mostly in fine-grained beds and likely formed authigenically
 796 at temperatures up to 260°C in Yangjiaping, and up to 360°C in Tonian-aged red
 797 beds farther south. In the Laoshanya Formation, the component resides in fine
 798 idiomorphic hematite crystals ($1\text{--}2 \mu\text{m}$) and/or in dense clusters/pseudomorphs of
 799 microplaty hematite. The corresponding pole lies near several Ordovician poles
 800 from the SCB, suggesting that the age of magnetic acquisition is Ordovician. We
 801 link component B to a regional low-grade tectono-thermal ($< 260^\circ\text{C}$) event in the
 802 late Ordovician, which may have partially or fully remagnetized pre-Silurian
 803 formations on the Yangtze craton.
- 804 4. Component C is carried mostly by coarse-grained hematite and is interpreted to be
 805 a primary remanence based on a positive reversal test. High-precision U-Pb dating
 806 of zircons extracted from two tuff horizons in the section yielded dates of $807.52 \pm$
 807 $0.18/0.27/0.91 \text{ Ma}$ and $804.50 \pm 0.22/0.30/0.91 \text{ Ma}$, which constrain the
 808 depositional age of the section between 809 and 804 Ma (using linear
 809 extrapolation). The primary nature of Component C is further supported by its
 810 agreement with some coeval poles from South China. Our new pole supports a
 811 stable, high-latitude position for the SCB between 809 and 804 Ma, and therefore a
 812 peripheral location with respect to the Rodinia supercontinent.
- 813 5. A stable position for the SCB likely extends further back and forth in time, as
 814 differences in paleomagnetic poles between 820 and 800 Ma can be explained by
 815 vertical axis rotations and/or regional remagnetization. This inferred stability is
 816 therefore inconsistent with the hypothesized Bitter Springs TPW event.

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 829 their constructive suggestions which improved our work.

830 **Author contributions**

831 BR and SG jointly conceived of the idea for this project and wrote the funding proposal
 832 together. JT, BR, EM, MW, SG and XZ carried out initial fieldwork in 2019. JM carried
 833 out supplementary sampling in 2021. JT was responsible for paleomagnetic and rock
 834 magnetic measurements with significant input from SG. EM, FD and JT were jointly
 835 responsible for the microscopic observations and mineralogical analyses. JT and AP
 836 carried out sample preparation and U-Pb geochronology with supervision by US. JT
 837 wrote the manuscript with significant input from BR and SG, and in consultation with
 838 the other authors. All authors have read and approved the final manuscript.

839 **Open Research**

840 Paleomagnetic data (Tonti-Filippini et al., 2024) are available in the MagIC database:
 841 <https://earthref.org/MagIC/19728>. DOI: 10.7288/V4/MAGIC/19728.

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