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The Largest Mountain Belt of the Last Billion Years: The East African Orogen, its tectono-topographic evolution and global significance.

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ABSTRACT

Hutton's Principle of Uniformitarianism suggests that the present is the key to the past, and in the present, Earth's topography is dominated by the Himalaya. The geological record clearly preserves past mountain ranges, but few have purported to have the effect on Earth's surface systems (such as the atmosphere, hydrosphere, biosphere and ocean chemistry) that the Himalaya does today. The Neoproterozoic East African Orogen is an exception. Not only is it the scale of the modern Himalaya, but it affected ocean chemistry in a similar manner to the Himalaya. Here we posit that the East African Orogen is the only other Himalayan-scale mountain range, and the only orogen that contends with the Himalaya as the largest mountain range of the last billion years.

In this manuscript we review and reconstruct the orogen. We then use existing metamorphic data triaged to exclude subduction-zone metamorphism, to estimate the elevation-time evolution of the orogen. We demonstrate that the available metamorphic record supports our assertion that the East African Orogen is amongst the largest orogens in the geological record. We also integrate spatially-located metamorphic data with a full-plate tectonic reconstruction of the evolving orogen to show how different plate interactions created the evolving topography.

INTRODUCTION

Mountain belts transport material from the deep Earth to the reactive crucible of the planet's surface, where it interacts with the atmosphere and hydrosphere and biosphere. This element-transfer role of mountains controls such fundamental uniformitarian Earth processes as the supply and extraction of CO_2 to and from the atmosphere, for example chemical weathering of large mountain ranges can push the planet towards icehouse conditions (Godderis et al., 2017). Weathering of mountain ranges also supplies nutrients that facilitate biological productivity and help drive evolution (Cox et al., 2016b; Yang et al., 2020). A major step in understanding how our planet evolved and developed as a system is to understand the history, location and evolution of past mountain ranges (e.g. Merdith et al., Submitted).

We can track the fingerprints of past mountain ranges by variations in the products of extremely long half-life radioactive decay systems into the oceans. The most useful of these is ^{87}Sr, which forms by radioactive decay of ^{87}Rb that is relatively common in the lithosphere and has a suitably long half-life of 4.92×10^{10} years. Strontium is soluble in water and has a residence time in the oceans on the order of millions of years (Veizer et al., 1999; Vollstaedt et al., 2014), meaning that it is well mixed in the oceans. It also readily precipitates with carbonates. These features mean that strontium isotopes form an effective proxy for the erosion of old rocks—rocks that get bought to the surface and eroded in mountain ranges. Therefore, the strontium isotope curve for carbonates through the last billion years contains within it the globally homogenized orogenic record through that time. For example, the growth of the Himalaya over the last 50 Ma is reflected in a sharp rise in radiogenic strontium (Derry and France-Lanord, 1996). The only other instance in the last billion years with a similar rise was between ca. 850–500 Ma, particularly focused between ca. 660–500 Ma (Shields, 2007; Cox et al., 2016a; Chen et al., 2022). This period coincides with the formation of the Gondwana-forming orogens (Collins and Pisarevsky, 2005; Cawood and Buchan, 2007; Schmitt et al., 2018)(Fig. 1), of which the East African Orogen is the largest in both spatial footprint and elevation (Fritz et al., 2013; Clark et al., 2015)(Figs. 1, 2).

Geographic reconstructions that divide the earth into land and sea have been a research focus since the development of the continental drift hypothesis (Wegener, 1912). Palaeomagnetic-based continental drift-like models that focus on continent distributions (Dalziel, 1997; Li et al., 2008) have evolved into full-plate tectonic reconstructions that use the wealth of palaeotectonic information held in global rock systems to attempt to reconstruct plate tectonics in deep time (Domeier and Torsvik, 2014; Domeier, 2016; Merdith et al., 2017a; Merdith et al., 2021; Li et al., 2023; Cao et al., 2024). Restoring the horizontal motion of Earth's lithosphere is not enough, though, to accurately model the importance of the geosphere on Earth surface systems. We also need to understand the geographic expression and distribution of Earth's surface topography (and bathymetry). This has been attempted in the Phanerozoic by looking at preserved rocks and attributing an elevation to that part of the world at the time of rock formation. For example, the presence of preserved deep marine, shallow marine and terrestrial sedimentary rocks indicates past flooded continents and coastal elevations. The presence of mountains is usually invoked from the preservation of metamorphic rocks on the surface, exhumed by erosion of the orogenic topography (e.g. Scotese, 2021). More quantitative approaches using the geochemistry of igneous rocks that formed in specific tectonic settings (like volcanic arcs, or mountain belts) have been developed that provide estimates of crustal thickness at the time of crystallisation, and therefore information about past topography (Tang et al., 2021; Zhu et al., 2022; Zhou et al., 2025). These approaches require magmatism, and many of the largest mountain ranges in the world have limited magmatism during the times at which they have maximum elevation (e.g. the Himalaya today). Here, we take a complementary approach to look at past-orogenic elevations, by reviewing the pressure-time history preserved in metamorphic rocks and use these to estimate past-orogenic elevation (Collins et al., 2022b).

In this manuscript, we first review the extent and nature of the preserved rocks within the East African Orogen (Fig. 1), as one of the largest orogens known in the geological record, place them within a modern full-plate tectonic framework, and qualitatively discuss the spatial and temporal evolution of major topography. We aim that our method forms a focus for future investigations

into quantitative estimates of elevation to understand how orogenic evolution modulates Earth system.

THE CONCEPT OF THE EAST AFRICAN OROGEN

The East African Orogen (EAO; Figs. 1, 2) was defined by Stern (1994). Here we restrict it to the regions where rocks crop out or are found in the subsurface that were formed and/or deformed in a mountain range that ran through the centre of Palaeozoic Gondwana (Fig. 1). The orogen forms a broad band from northernmost Gondwana—Egypt and the Middle East, and now dispersed terranes that reconstruct to this region—through eastern Africa and Madagascar (Figs. 1, 2). South of Madagascar, in reconstructed Gondwana, the orogen bifurcated into two branches (Fig. 2). A western branch joins with the Zambezi Belt of Zambia and Zimbabwe and then to the Lufilian Arc and the central Damara Belt (Goscombe et al., 2020)(Figs. 1, 2). An eastern branch passes through southernmost India, Sri Lanka and into the Lützow-Holm Bay area of eastern Antarctica (Plavsa et al., 2015) ('Lut' in Fig. 2). Here we will focus on the region north of this bifurcation, where the EAO is clearly the result of ocean closure and subsequent continent-continent collision between Neoproterozoic India and the Neoproterozoic African continents—the southern Congo-Tanzania-Bangweulu continent and the Saharan metacraton (Figs. 1, 2). The boundary between these two continents is the Central African Fold Belt (otherwise known as the Oubanguides belt)(Figs. 1,2), which consists of a series of thrust sheets, known as the Yaoundé-Yangana nappes (Toteu et al., 2022)("Y-Y" in Fig. 2) transported south over the Congo-Tanzania-Bangweulu Block. The belt marks a site of ocean closure that separates a Neoproterozoic arc terrane (Poli- Mayo Kebbi arc) from the Congo-Tanzania-Bangweulu continent and can be traced west into the Sergipano belt of NE Brazil (Neves et al., 2006)(Fig. 1).

The EAO divided Gondwana into an eastern part, whose major components now form South America and Africa, and a western part, whose now dispersed successors form Australia, East Antarctica, India and much of East Asia (Fig. 1). In our discussion here we refer to directions in a reconstructed Gondwana reference frame with Africa oriented as present day. The significance of the EAO was increased when McWilliams (1981) supported early speculation that plate

tectonics operated in the Precambrian (Burke et al., 1976) and interpreted the palaeomagnetic data of the times to reflect the existence of two large, separately travelling, continents that collided to form Gondwana in the late Proterozoic. Stern (1994), delineated and named the EAO and suggested that this was the site of collision between these east and west Neoproterozoic Gondwana continents. This led to a hunt for the suture of the newly named Mozambique Ocean that was hypothesized to have separated these continents (Shackleton, 1996).

In 1997, Ian Dalziel began the push to map the deep time geography of the planet by publishing a major continental-drift style reconstruction for the Neoproterozoic (Dalziel, 1997) that attempted to combine known geology with the palaeomagnetic record. This model coincided with Joe Meert and Rob Van der Voo (1997) pointing out that the developing geochronological record for the formation of Gondwana appeared to show three main periods of orogenesis. The earliest orogeny, between 800–650 Ma, was correlated with the East African Orogeny and was located mainly in East Africa. A second orogeny was interpreted as representing the amalgamation of the South American part of Gondwana between 600–530 Ma. A third, eastern Gondwanan orogeny, was suggested to have occurred at ca. 550 Ma. They named this the Kuunga Orogeny and suggested that it involved the final amalgamation of Gondwana (Meert and Van der Voo, 1997).

At about this time, several researchers realized that the vast region of the EAO preserved evidence of considerable orogenic complexity, like that of Phanerozoic orogens. Kröner et al. (2000) and Collins and Windley (2002) suggested that a major piece of pre-Neoproterozoic continental crust lay within the East African Orogen. This palaeocontinent was subsequently named ‘Azania’ (Collins and Pisarevsky, 2005) (Figs. 1, 2). It was proposed that Azania collided with the Congo-Tanzania-Bangweulu continent at ca. 650 Ma and that this collision was the event responsible for the East African Orogeny. Neoproterozoic India was hypothesized to have later collided with the then combined Azania/Congo-Tanzania-Bangweulu continent at ca. 550 Ma forming a Malagasy Orogeny (Collins and Pisarevsky, 2005). Meanwhile, Johnson and Woldehaimanot (2003) published a key review of the northern EAO in the Arabian Nubian Shield

(Fig. 2). They demonstrated that this vast area, which incorporated all of Arabia and much of NE Africa, consisted of a series of different-aged terranes that amalgamated at different times in a similar, but better resolved, manner that Collins and coworkers were proposing in the southern EAO. Johnson et al. (2011) and Fritz et al. (2013) published major reviews of the northern and southern parts of the EAO.

The last 20 years has seen the identification of further pre-Neoproterozoic crust within the EAO that has extended the number and complexity of continental terranes within the orogen (Be'eri-Shlevin et al., 2009; Be'eri-Shlevin et al., 2012; Boger et al., 2014a; Eyal et al., 2014; Boger et al., 2015; Clark et al., 2020). Both margins of the EAO have been significantly revised and extended. Here we suggest that the eastern margin be extended to include the basement of Oman, Seychelles, Pakistan and NW India (Alessio et al., 2018; Blades et al., 2020; Collins et al., 2021b; de Wall et al., 2022) (Figs. 1, 2). In these regions, juvenile Tonian (1000 to ca. 720 Ma, this and all stratigraphic ages taken from Cohen et al., 2025) crust forms the basement, but in contrast to the western EAO, orogenesis ended in the late Tonian or earliest Cryogenian (ca. 720–635 Ma).

The original definition of the western margin of the EAO as the boundary with a “Saharan Metacraton” (Abdelsalam et al., 2002) is also far from clear and needs significant revision, especially with the identification of Ediacaran (ca. 635–539 Ma) juvenile terranes in northern Chad in the centre of the erstwhile “Saharan Metacraton” (Şengör et al., 2020; Shellnutt et al., 2026). In Chad, a pre-Neoproterozoic terrane has been identified and named the Poli-Mayo Kebbi-Lake Fitri Terrane (PMKF)(Fig. 1). To date there are no pre-Neoproterozoic rocks known from this region and it appears to be a juvenile Tonian terrane (Shellnutt et al., 2026). Southeast of the PMKF lies the Adamawa-Yadé-Guéra-Ouaddaï (AYGO) Terrane (Fig. 1)(Shellnutt et al., 2026). This contains gneisses with protoliths that date back to the Stenian (1000–1200 Ma)(Blades et al., 2021a), with detrital and inherited zircons that reach back to the Archaean (Djerossem et al., 2020; Blades et al., 2021a). These contrast with other parts of the Saharan basement, for example Gebel Uweinat in SW Egypt, that preserve Archaean rocks that have largely missed later orogenesis (Zhang et al., 2019)(Fig. 1). A boundary between pre-

196 Neoproterozoic and Tonian crust marks the edge of the EAO as shown in figures 1 and 2. Juvenile
197 Neoproterozoic terranes then characterize the western parts of the EAO (west of Afif-Abas and
198 Azania, Fig. 2). The 800–650 Ma East African Orogeny of Meert and Van der Voo (1997) affected
199 these regions, but the later Malagasy Orogeny (correlated with the Angudan Orogeny in Arabia,
200 Collins and Pisarevsky, 2005) largely, did not (with only local shear zones being active at this
201 time).
202
203 The origin and tectono-geographic location of the EAO have been revised considerably. The
204 original suggestion, that the orogen preserved a Tonian continental rift sequence that
205 represented the break-up of Rodinia (Stern, 1994), is no longer tenable, as extensive Stenian crust
206 has been identified along the west margin of the EAO (de Wit and Linol, 2015; Blades et al.,
207 2021b) and within the EAO in Madagascar, Sinai and East Antarctica (Eyal et al., 2014; Tucker et
208 al., 2014; Jacobs et al., 2015; Elburg et al., 2016; Archibald et al., 2018). In addition, the age of
209 800 Ma for the beginning of the East African Orogeny (Meert and Van der Voo, 1997) is
210 challenged by extensive 850–750 Ma magmatism in central Madagascar (Handke et al., 1997;
211 Archibald et al., 2016; Archibald et al., 2017), pre-800 Ma magmatism in Ethiopia (Blades et al.,
212 2015; Bowden et al., 2020), Arabia (Whitehouse et al., 2016; Alessio et al., 2018), NW India and
213 SE Pakistan (Pandit et al., 2003; Shakoor et al., 2019), South India (Plavsa et al., 2015) and Sri
214 Lanka (Kröner and Jaeckel, 1994). A window into the basement in NE Oman demonstrates the
215 presence of high-grade metamorphism and deformation at ca. 820 Ma (Alessio et al., 2018) that
216 links with pre-760 Ma tectonothermal activity recorded in the Pakistan and India basement west
217 of the Western Margin Fault of the Aravalli-Delhi Fold Belt India (Deb et al., 2001; de Wall et al.,
218 2022), showing that pre-800 Ma orogenesis is not limited to magmatism. We suggest that the
219 orogen should now be thought to represent over 500 million years of punctuated subduction-
220 accretion. This new paradigm means that the orogen does not represent a Wilson-like, internal
221 ocean (Nance et al., 1988), opening and closing in approximately the same position. Instead, it
222 seems clear that the orogen lay in a tectono-geographic position like that of the modern Pacific-
223 rim, where subduction of an external ocean over hundreds of millions of years resulted in a
224 punctuated record of accretion, extension, arc magmatism and collision (Merdith et al., 2017b;

Şengör et al., 2020; Collins et al., 2021b; Merdith et al., 2021). This process is similar to the Proterozoic to Mesozoic history of the Central Asian Orogenic Belt (Kröner et al., 2014; Xiao et al., 2020) and the Phanerozoic history of the Tasmanides of Australia and the broader Terra Australis Orogen (Cawood, 2005; Glen and Meffre, 2009; Rosenbaum, 2018).

In the next section, we summarize the Stenian to Cambrian (ca. 1100–520 Ma) history of the now dispersed parts of the EAO before we discuss the tectonic geography of their amalgamation and evolution of the resulting mountain range(s).

THE FAR NORTHERN EAST AFRICAN OROGEN

Displaced terranes in Greece, Türkiye and Iran

Tectonic windows into Neoproterozoic to early Palaeozoic crust occur as discontinuous exposures throughout the central crystalline basement of Europe and from the Alps to Türkiye, Syria and Iran in the Alpine–Mediterranean mountain belt (von Raumer et al., 2003). The orogenesis preserved in these locations are variously attributed to the Cadomian or East African orogen events (often just termed 'Pan-African'). The precise original location of these now exotic terranes are hard to define. However, the limited Mesozoic ocean crust development in the western Mediterranean and (near?)continuity of the Apulia promontory with both northern Gondwana and northern Europe means that the Aegean and eastern Mediterranean terranes are restricted to originate close to the Egyptian margin of north Gondwana (Stampfli, 2005; Torsvik and Cocks, 2013)(Fig. 2), close to where the East African Orogen met the circum-Gondwana Ocean (which at this part of geological time and at this location is named the Tornquist Sea).

The terranes most pertinent here are the Greek Pelagonian Zone, the Turkish Tauride Terranes (the Menderes Massif and the Bitlis Massif), as well as the dismembered terranes of the Sanandaj–Sirjan Zone of Iran as these formed the far northern extensions of the EAO (Fig. 2). The Pelagonian Zone of Greece contains the relatively small Florina Terrane, which preserves orthogneisses dated as crystallizing between ca. 730–690 Ma (Anders et al., 2006). Crystallization

ages of this antiquity have not been recorded from the Menderes Massif that is found along tectonic strike in west Türkiye. Here, an old kernel of Neoproterozoic clastic metasedimentary rocks are intruded by 570–520 Ma granites (Gessner et al., 2004; Koralay et al., 2004) with rare metabasites that preserve Cambrian eclogite metamorphic assemblages (Candan et al., 2016). A similar Proterozoic clastic sequence is preserved in the Bitlis Massif of SE Türkiye, here also, purported subduction-related magmatic rocks intrude the sequence at ca. 550–530 Ma (Ustaomer et al., 2009). The Iranian Sanandaj-Sirjan Zone, contains a series of inliers of Neoproterozoic–Cambrian metasedimentary rocks and granitoids decorating the Cenozoic suture (Jamshidi-Badr et al., 2013; Fergusson et al., 2016; Honarmand et al., 2018). Further east in Iran, the 'Central and East Iranian Microcontinent' (CEIM) consists of three blocks (the Lut, Tabas and Yazd) separated by tectonic belts. In one of these belts, 575–525 Ma granitoids that have been attributed to an Andean-like volcanic arc, suggest that this region shares a similar Ediacaran–Cambrian tectonic setting to the other basement terranes in Türkiye and Iran (Ramezani and Tucker, 2003; Rossetti et al., 2015). The continental volcanic-arc tectonic setting of these has been questioned by Zoleikhaei et al. (2025), who argued that the detrital zircon and apatite record in Ediacaran to Palaeozoic sandstones, is better explained by a rift setting at this time with no arc-like edifice.

Both Ediacaran–Cambrian intrusions in the Sanandaj–Sirjan Zone and in the Menderes Massif preserve inherited zircon that stretches back to ca. 3.6 Ga and hafnium isotopes on younger grains that suggest ancient continent was a component of the source of these rocks (Nutman et al., 2014). Whether this ancient source was crystalline basement beneath those regions, or detritus then preserved in Proterozoic metasedimentary rocks is unclear.

THE NORTHWESTERN EAST AFRICAN OROGEN: THE ARABIAN NUBIAN SHIELD (ANS)

Egypt, Sudan, Saudi Arabia, Eritrea, central and north Ethiopia, Yemen, northern Somalia

The ANS (Fig. 2) is dominated by low grade volcano-sedimentary sequences and associated plutonic and ophiolitic remnants. These are reconstructed into a complex mix of terranes

representing accreted arcs that record subduction polarity reversals that are reviewed and summarised in a number of papers (Johnson et al., 2011; Robinson et al., 2014; Kashghari et al., 2025).

Pre-Stenian (> 1200 Ma) crust in the ANS forms two broad belts. One extends from the Palaeoproterozoic Khida region of Saudi Arabia (which also underlies much of the Siham-Khida Terrane—previously part of the Afif Terrane—to the north; Stoeser and Frost, 2006)("SK" in Fig. 2) to the Abas Terrane of Yemen (Yeshanew et al., 2015; Whitehouse et al., 2023)("Ab" in Fig. 2). The Neoproterozoic Al-Bayda Terrane separates the Abas Terrane from the Palaeoproterozoic–Archaean Al-Mafid Terrane of southern Yemen ("Ab" and "AMa" respectively in Fig. 2). The Al-Mafid Terrane is traced south across to the Somaliland escarpment (Sassi et al., 1993), and further south through eastern Ethiopia (Kröner and Sassi, 1996) to southern Somalia (Küster et al., 1990), central Madagascar (Kröner et al., 2000) and into southern India (Collins and Windley, 2002; Collins et al., 2007; Plavsa et al., 2014)(Fig. 2). Collins and Pisarevsky (2005) named this pre-Stenian microcontinent 'Azania' after the ancient Greek name for the east African Coast (Schoff, 1912).

The next oldest terrane in the ANS is the late Mesoproterozoic Sa'al Metamorphic Complex (1.03–1.02 Ga) in Sinai, marking the initiation of magmatism in the northern-most Arabian Nubian Shield (Be'eri-Shlevin et al., 2012; Eyal et al., 2014)(Fig. 2). The original location of this Stenian terrane, is uncertain, due to it being surrounded by younger terranes, but coeval subduction-magmatism occurred within the Saharan basement (Blades et al., 2021b) and the Sinai terrane may form a rifted part of this continent.

The Tonian to Cryogenian history of the ANS is marked by formation of oceanic volcanic arcs and arcs built on Azanian (or Afif-Abas) crust that amalgamated to form a larger intra-Mozambique Ocean terrane separate from both Neoproterozoic India and African Gondwanan continents. A number of terranes in the ANS are correlated as equivalents, separated by the opening of the Red Sea, from south to north, these are the Asir (a composite of the Bidah, An Nimas, Al Qarah-

Malahah and Tathlith terranes of Kashghari et al., 2025) and Tokar/Barka terranes, the Haya and Jiddah terranes, the Hijaz and Gabgaba/Gebeit terranes, and the Eastern Desert and Midyan terranes (Johnson et al., 2011)(Fig. 2). It is unclear whether the combined Asir-Tokar/Barka terrane and Haya-Jiddah terranes were ever on separate plates as, in Saudi Arabia, no clear suture is seen between them (Kashghari et al., 2025). In SE Sudan and Eritrea, the Tokar-Barka terrane (Fig. 2) has been divided into a number of terranes (Drury and De Souza, 1998), in a similar manner to its Saudi Arabian equivalent (Johnson et al., 2011). We suggest that these terranes formed a complex middle Tonian continental amalgam.

The older, Tonian to earliest Cryogenian, amalgamation history of the ANS is marked by approximately ENE-WSW oriented sutures (present orientation) between juvenile Neoproterozoic ocean-arc terranes (Collins et al., 2021a; Collins et al., 2021b)(Fig. 2). The oldest of these sutures is between the Jiddah-Haya and Gabgaba/Gebeit-Hijaz terranes (the Bi'r Umq–Nakasib suture)("bu" in Fig. 2), which is dated at ca. 780–750 Ma (Johnson et al., 2011). This suture created the kernel of a late Tonian microcontinent. The Midyan-Eastern Desert terrane collided with this kernel ca. 715 Ma along the Yanbu-Sol Hamed suture (Robinson et al., 2015a) ("y" in Fig. 2). Both sutures evolved from SE-dipping subduction zones (Robinson et al., 2015a).

The older ENE–WSW sutures are bound by younger NNW–SSE oriented Cryogenian to Ediacaran sutures and terranes that represent a fundamental kinematic change in Mozambique Ocean subduction (Collins et al., 2021a; Collins et al., 2021b). The oldest of these is the 680–640 Ma Nabitah suture (Flowerdew et al., 2013)("ns" in Fig. 2), which forms the eastern margin of the intra-Mozambique Ocean island-arc terrane microcontinent (discussed above), against Tonian-Cryogenian continental arcs (Siham and Suwaj arcs of Kashghari et al., 2025)("SK" and "Sj" in Fig. 2) built on the Afif-Abas microcontinent. This now Cryogenian engorged Afif-Abas microcontinent collided with the active margin of the Sahara basement along the Sudanese Keraf Suture ("ks" in Fig. 2). This collision occurred in late Cryogenian to early Ediacaran times (ca. 650–580 Ma)(Abdelsalam et al., 1998). Further to the east, in the most easterly exposed terrane—the Saudi Ar Rayn Terrane—juvenile calc-alkaline magmatism stretches from ca. 690 Ma to 615 Ma

(Doebrich et al., 2007). Turbiditic sediment deposition in the Ad Dawadmi basin ("AD" in Fig. 2) that separates the Ar Rayn Terrane ("AR" in Fig. 2) from the Afif-Abas microcontinent continued until at least 620 Ma (Cox et al., 2012), but was locally intruded by ca. 630 Ma adakitic magmas (Cox et al., 2019). This sequence was metamorphosed to greenschist-facies grades at ca. 620 Ma (Cox et al., 2012). Further east still, broad N–S magnetic highs, beneath the Arabian Phanerozoic sedimentary sequence (Johnson and Stewart, 1995), suggest younger arc terranes now buried beneath the Rub al-Khali Basin (Fig. 2). Geochronological data from basement-intercepting boreholes throughout Saudi Arabia exclusively yield Neoproterozoic ages from these buried terranes (Collins et al., 2024). Deformation that we interpret as related to the collision and amalgamation of these buried terranes is prominent in the subsurface of western Oman where it is known as the Western Deformation Front ("WDF" in Fig. 2) and delineates the Angudan Orogeny (Loosveld et al., 1996; Al-Barwani and McClay, 2008; Gómez-Pérez and Morton, 2025). Al Rawahi et al. (2026) demonstrated from provenance changes in the late Ediacaran Fara Formation of Oman that closure of this suture occurred by ca. 551 Ma. We correlate this latest Ediacaran–Cambrian Angudan orogeny with the Malagasy orogeny of Collins and Pisarevsky (2005). This event is interpreted as the final suture of the Mozambique Ocean (Gómez-Pérez et al., 2024; Gómez-Pérez and Morton, 2025). A zone of lithospheric mantle shear wave anisotropy in far southwestern Oman may locate this suture in the upper mantle (Al-Lazki et al., 2012), this broadly correlates with highly strained Ediacaran volcano-clastics found in the subsurface and at outcrop (Gómez-Pérez and Morton, 2025). The transition to post-tectonic magmatism within the eastern terranes of the Arabian Nubian Shield begins from ca. 605 Ma with post-tectonic magmatism (Doebrich et al., 2007; Cox et al., 2012; Robinson et al., 2014; Robinson et al., 2015b; a; Cox et al., 2019) and pull-apart basins developed along the large strike-slip faults (known as the Najd Faults) that cut the region (Moore, 1979; Meyer et al., 2014; Nettle et al., 2014; Shen et al., 2022; Hamimi et al., 2023).

The basement of Yemen is marked by considerable pre-Neoproterozoic crust found in the Al-Mahfid and Abas terranes (Windley et al., 1996; Whitehouse et al., 1998). The Abas terrane is linked with the Siham-Khida terrane of Saudi Arabia (Whitehouse et al., 1998; Yeshanew et al.,

2015) to form the Afif-Abas continent of Collins and Pisarevsky (2005)(Fig. 2). Whereas the Al-Mahfid terrane has been linked with pre-Stenian crust found through Somaliland, eastern Ethiopia, southern Somalia, central Madagascar and the Madurai Block of southern India that formed the continent Azania (Whitehouse et al., 2001; Collins and Pisarevsky, 2005). The Abas Terrane lies east of an interpreted late Tonian to Cryogenian arc, called the Amlah Terrane (Whitehouse et al., 2025)("Aml" in Fig. 2) and an oceanic suture between this and the Saudi Arabia Tathlith terrane (Kashghari et al., 2025) is inferred by the change in younging direction of preserved volcanic arc intrusions (Flowerdew et al., 2013; Whitehouse et al., 2025). The Abas terrane is largely composed of ca. 790–725 Ma granitic gneisses with variable $\varepsilon(Nd)_t$ values as evolved as -11, suggesting significant ancient crust in the source magmas (Whitehouse et al., 1998; Yeshanew et al., 2015). Some rocks may date back to the early Tonian and a single Neoarchaean core has been analysed hinting at the age of older basement (Whitehouse et al., 1998). The juvenile Al Bayda terrane ("ABa" in Fig. 2) separates the Abas terrane from the Al Mahfid terrane. The latter contains significant Neoarchaean crust and was interpreted by Collins and Windley (2002) to form the northern part of a Neoproterozoic continent that lies within the EAO (Azania of Collins and Pisarevsky, 2005). The Al Mukalla terrane ("Amk" in Fig. 2) lies east of the Al Mahfid in Yemen and was correlated by Whitehouse et al. (2001) with the Maydh (or Mait) terrane of northern Somalia (Somaliland)("M" in Fig. 2), which contains meta-tholeiitic basalts with pillow-lavas textures and mid-ocean ridge like chemistries (Dal Piaz et al., 1993).

Pre-Neoproterozoic rocks are found along the southern escarpment margin of the Gulf of Aden in northern Somalia (Somaliland) and eastern Ethiopia form a > 600 km long inlier between the latitudes of 41–47° E, from Hirna in east Ethiopia to the town of Maydh in northern Somalia. These were included within Azania by Collins and Pisarevsky (2005). In Somaliland, the oldest components of this inlier are the Qabri Bahar and Mora complexes (Sassi et al., 1993)("QM" in Fig. 2), which consists of paragneisses and orthogneisses with local migmatites and granulite-facies relics (Sassi and Visonà, 1993). The east Ethiopian continuation of this belt makes-up the unimaginatively named 'Lower Unit' and Bora Group of the Harar-Hirna area (Kazmin, 1975; Berhe, 1981)("HH" in Fig. 2) that contain ca. 2000 Ma tonalitic gneiss xenoliths in Tonian

399 granitoids (Teklay et al., 1998; Yeshanew et al., 2017). In addition, Archaean and earliest
400 Palaeoproterozoic zircon grains have been reported from granitoids interleaved with
401 metasedimentary gneisses (Teklay et al., 1998). All analysed rocks in eastern Ethiopia have
402 evolved neodymium isotope values, suggesting that either there are older rocks at depth, or the
403 a sedimentary source was derived from more ancient crust, or both (Teklay et al., 1998;
404 Yeshanew et al., 2017).
405
406 In Eritrea, the Tokar/Barka terrane (Johnson and Woldehaimanot, 2003) consists of juvenile 850–
407 650 Ma old supracrustal and intrusive rocks that range from low greenschist facies to granulite
408 facies (Andersson et al., 2006). Peak metamorphism of these rocks is dated at ca. 590 Ma
409 (Andersson et al., 2006). Rocks of the Tokar/Barka terrane in Eritrea pass south into the Tsaliet
410 Group of Tigray in northern Ethiopia where they conformably pass up into Tonian strata of the
411 Tambien Group that are well dated to between ca. 820 Ma and < 719 Ma (Swanson-Hysell et al.,
412 2015; Park et al., 2019). This region has missed the high-grade Ediacaran metamorphism seen
413 both to the east and west.
414
415 To the southwest, the Western Ethiopian Shield (WES) is one of the larger regions of ANS outcrop
416 (Fig. 2). It marks the transition from the relatively low-grade ANS to the higher metamorphic
417 grades of the Mozambique Belt (Abdelsalam and Stern, 1996). The WES preserves no evidence
418 of pre-Neoproterozoic crust, with the oldest rocks in the area dated at ca. 855 Ma (Blades et al.,
419 2015) with geochemical evidence suggesting that they formed above an intra-oceanic subduction
420 zones (Blades et al., 2019). Magmatism (re-melting of juvenile crust) and deformation is seen at
421 810–770 Ma and at 660–635 Ma in the Western Ethiopian Shield, with post tectonic magmatism
422 at ca. 584 Ma (Blades et al., 2015; Blades et al., 2019; Bowden et al., 2020). All the rocks in the
423 Western Ethiopian Shield have juvenile ƐHf(t) values of +4.5 to +10, suggesting that the influence
424 of continental crust was minimal.
425
426 **THE WESTERN MARGIN OF THE EAST AFRICAN OROGEN**
427 ***Sudan, Chad, Central African Republic, South Sudan, Uganda***

The western margin of the East African Orogen can be reasonably well located in northern Sudan where > 900 Ma schists and gneisses are overthrust by 800–900 Ma gneisses and low-grade metavolcanic rocks of the Rahaba–Absol terrane (Harms et al., 1994; Schandelmeier et al., 1994; Johnson and Woldehaimanot, 2003)("RA" in Fig. 2). Thrusting was over here before 700 Ma (Abdelsalam et al., 1995). As you get further south, though, the western margin becomes hard to locate as much of the south-east of the region often denoted as the Sahara Metacraton (Abdelsalam et al., 2002) has a basement made up of late Mesoproterozoic rocks that formed in volcanic arc terranes (Shellnutt et al., 2017; Blades et al., 2020; Djerossem et al., 2020; Plunder et al., 2026). This suggests that this part of the basement to Chad and Sudan forms an earlier accretionary orogen that simply gets younger eastward into the EAO—part of a large and long-lived accretionary system, possibly stretching to the eastern edge of the West African Craton (Şengör et al., 2020; Lom et al., 2024; Shellnutt et al., 2026).

The Nuba Mountains lie at a critical point to the west of the traditional western boundary of the EAO, yet north of the Ediacaran tectonism that marks the Central African Fold Belt (Fig. 2). The eastern part of the exposed basement rocks consist of low-grade volcano-sedimentary rocks that are separated from a higher-grade gneiss terrane in the central Nuba Mountains by the Kabus ophiolitic mélange (Abdelsalam and Dawoud, 1991). This boundary has long been used to mark the western boundary of the ANS and therefore the East African Orogen (Abdelsalam and Dawoud, 1991; Johnson and Woldehaimanot, 2003). Mohammed (2017), however, showed that these gneisses are Neoproterozoic and to the west of them lies a ca. 780 Ma ophiolitic succession that he named the Arid Unit. This suggests that juvenile Neoproterozoic crust continues west of the Nuba Mountains and that these mountains lie close to the triple junction where the EAO meets the Central African Fold Belt (Fig. 2).

THE NORTHEASTERN EAST AFRICAN OROGEN

NW India, Pakistan, Oman, UAE, Afghanistan, SE China, Seychelles, N Madagascar

There are three main windows into the Omani basement exposed east of the 'Western Deformation Front'. These all lie east of the possible lithospheric suture marked by shear wave splitting in far southwestern Oman (Al-Lazki et al., 2012). The tectonic windows are Al Jobah (northern Huqf area), Jebel Ja'Alan (NE Oman) and Mirbat region of Dhofar in the south-east ("Hu", "JJ" and "Mir" in Fig. 2). These have been interpreted to represent arc accretion during the Tonian (890–720 Ma) (Stern and Johnson, 2010; Rantakokko et al., 2014; Whitehouse et al., 2016; Alessio et al., 2018), with magmatism and deformation/metamorphism broadly younging from the NE to the SW, towards the putative final Mozambique Ocean suture (Blades et al., 2020). In addition, clasts bought up from > 8km depth in salt diapirs in Oman and the United Arab Emirates preserve volcanic rocks with ages between 560–545 Ma that contain inherited zircons back to the Neoarchaean (Thomas et al., 2015). Detrital zircons in sedimentary enclaves show a similar age range (Thomas et al., 2015; Arboit et al., 2024) as do in-situ metasedimentary basement rocks of the Jebel Ja'Alan region (Whitehouse et al., 2016; Blades et al., 2020). These suggest a depositional connection to crust of considerable antiquity. A number of workers have suggested that Afghanistan may be possible source for these (Thomas et al., 2015; Alessio et al., 2018; Blades et al., 2020), as this Archaean/Palaeoproterozoic terrane likely lay with the NW Neoproterozoic Indian accretionary system and was interpreted in Merdith et al. (2021) as correlating with the SW of the Yangtse craton of China as a microcontinent accreted to the Neoproterozoic Indian margin in the Tonian.

In Gondwana, Oman lay adjacent to NW India in Gondwana (Figs. 1, 2), and by Ediacaran times, at least, a similar sedimentary package was deposited across Rajasthan, Pakistan and through Oman that contains excellent petroleum source rocks and evaporites (Cozzi et al., 2012). These were being deposited in a stable, passive margin environment (Allen, 2007) at the same time that rocks in the eastern Saudi Arabian Shield were forming in an active subduction margin (Cox et al., 2012). The pre-Ediacaran basement rocks in Rajasthan and Pakistan, also, share similarities with those of Oman. Granitoids have been dated from Rajasthan ("Raj" in Fig. 2) and from Nagar Parkar in eastern Sind (Pakistan, "NP" in Fig. 2) at ca. 1.1 Ga (Raza et al., 2012; Meert et al., 2013). But no evidence of pre-existing crust occurs west of the Western Margin Fault of the Aravalli-

Delhi Orogen (“wmf” in Fig. 2). Tonian granitoids and rhyolites occur in inliers through northwest India and Pakistan, where they cluster into crystallization ages of ca. 990–970 Ma (Pandit et al., 2003), ca. 860–820 Ma (Davies and Crawford, 1971; Deb et al., 2001; Van Lente et al., 2009; Just et al., 2011) and ca. 775–760 Ma (Gregory et al., 2009; Van Lente et al., 2009; Ashwal et al., 2013; Meert et al., 2013). The latter magmatic and extrusive phase forms one of the largest felsic igneous provinces on the planet, which is also traced to the Seychelles (Torsvik et al., 2001; Tucker et al., 2001)(“Sey” in Fig. 2), and may correlate with the northern Bemarivo Belt of Madagascar where the magmatism dates from ca. 750-705 Ma (Collins, 2006; Thomas et al., 2009; Armistead et al., 2019)(“Bem” in Fig. 2).

Armistead et al. (2019), showed that ca. 750–705 Ma rocks from the northern Bemarivo Belt of far north Madagascar preserve similar juvenile εHf(t) signatures to rocks in south China, Rajasthan, Oman and the Seychelles, that had previously been suggested to have formed a volcanic arc system on the margin of Neoproterozoic India (Wang et al., 2017; Cawood et al., 2018). The period of magmatism in this proposed arc was long-lived, beginning at around ca. 850 Ma and ending at around ca. 700 Ma. When reconstructed in this fashion (Armistead et al., 2019), there is a general southward to southeastward younging trend (reconstructed orientation), with the oldest record coming from China, and progressing to younger rocks through Oman, Malani, Seychelles and the northern Bemarivo Belt.

Tonian palaeomagnetic data from Oman (Swanson-Hysell et al., 2025; Antonio et al., 2026) support the correlation of Oman with Neoproterozoic India at the time (Alessio et al., 2018). Sedimentological data from the Cryogenian and Ediacaran deposits suggest that this link lasts through until Ediacaran formation of Gondwana (Cozzi et al., 2012; Gómez-Pérez et al., 2024; Gómez-Pérez and Morton, 2025). Only colliding with the rest of Arabia in the late Ediacaran (Collins and Pisarevsky, 2005; Blades et al., 2020; Merdith et al., 2021; Cao et al., 2024; Gómez-Pérez and Morton, 2025; Al Rawahi et al., 2026).

THE CENTRAL EAST AFRICAN OROGEN

Southern Ethiopia, Uganda, Kenya, NE Tanzania, southern Somalia

The grade of metamorphism of rocks exposed on the surface increases through Ethiopia from the broadly greenschist-facies rocks of the ANS in the north to upper amphibolite and granulite-grade rocks that dominate the Mozambique Belt south of central Ethiopia. We interpret this transition to reflect greater Cryogenian to Cambrian orogenesis in the south with higher and more extensive orogenic topography of the overlying mountains (see discussion below for more on palaeoelevation).

South of the Western Ethiopian Shield (WES, Fig. 2), the Karamoja Belt of NE Uganda ("Kmja" in Fig. 2) is dominated by the Karasuk Supergroup (Vail, 1985; Westerhof et al., 2014), which are metaigneous rocks that crystallized between ca. 740 Ma and ca. 680 Ma (Mänttäri et al., 2011; Westerhof et al., 2014) and are intruded by post-tectonic granitoids with U–Pb crystallization ages of ca. 660 Ma (Mänttäri et al., 2011). Similar timing of magmatism is seen within the WES (Blades et al., 2015), however, there is a notable difference between the Southern Ethiopian Shield ("SES" in Fig. 2) and the Karamoja Belt. In the Southern Ethiopian Shield, a ca. 660 Ma event that is coeval with the late magmatism in the eastern WES, is termed the 'Moyale Event' (Teklay et al., 1998) this is associated with multiply deformed dismembered ophiolites (Abebe et al., 2025). Further east still, Yeshanew et al. (2017) identified Ediacaran granitic magmatism (600 – 560 Ma) in the Harar region of eastern Ethiopia ("HH" in Fig. 2). The Ediacaran ages extend known magmatic events in the region to much younger ages than previously reported, and considerably younger than any orogenesis in the Western Ethiopian Shield. Two of the granitoid samples dated at ca. 600 and ca. 585 Ma are not deformed, whereas the younger gneisses at ca. 568 and ca. 560 Ma are strongly deformed. These were interpreted to represent a low-strain domain during orogenesis (Yeshanew et al., 2017) and suturing between two lithospheric blocks (WES and Eastern Ethiopia—Azania?). These mid-late Ediacaran ages are much younger than any evidence for tectonism found in the WES, and any suggested ocean closure from the ANS west of the Saudi Arabian Afif Terrane (Johnson et al., 2011), so rather than representing suturing between the WES and eastern Ethiopia, we suggest that the post-ca. 560 Ma deformation may

reflect late intra-continental deformation, as seen elsewhere in the ANS (e.g., Kusky and Matsah, 2003).

To the south, in Kenya, southern Somalia and Tanzania, the East African Orogen is dominated by high-grade gneisses of the Mozambique Belt (see Fritz et al., 2013 for a review)("Moz Belt" in Fig. 2). In the region of the town of Lodwar in NW Kenya ("Ld" in Fig. 2), amphibolite-grade orthogneisses yield protolith ages of ca. 800 Ma that were metamorphosed at ca. 630 Ma (Blades et al., 2026). These are separated from the NE edge of the Congo-Tanzania-Bangweulu Block by the Cryogenian to early Ediacaran West Pokat ophiolite (Vearncombe, 1983; Ries et al., 1992)("Pok" in Fig. 2) and are themselves bound to the east by metasedimentary gneisses that record peak metamorphic conditions of 8.4–10.6 kbar and temperatures of 750–830°C at ca. 625 Ma (McIntyre, 2023). The Lodwar orthogneisses likely correlate north with the Didesa Terrane (West Ethiopian Shield; Blades et al., 2015). This region passes south into the Western Granulites of SW Kenya and Tanzania ("WG" in Fig. 2) that involve highly metamorphosed pre-Neoproterozoic crust that was metamorphosed to granulite facies at ca. 630–615 Ma (Möller et al., 2000; Sommer et al., 2003; Cutten et al., 2006; Apen et al., 2020) and locally extruded westward over the Congo-Tanzania-Bangweulu Block (Eguíluz et al., 2024).

The eastern part of the Mozambique Belt (Burr Massif, Somalia, Fig. 2) consists of a tract of largely pre-Neoproterozoic continental crust, strongly deformed, partially melted and metamorphosed in the Ediacaran/Cambrian (ca. 600 and ca. 530 Ma, Küster et al., 1990; Lenoir et al., 1994). Extensive tectonothermally reworked pre-Neoproterozoic crust lies further west in the Neoarchaean Tanzanian craton (Muhongo et al., 2001; Johnson et al., 2003; Reddy et al., 2003; Tenczer et al., 2013). Near peak, granulite-grade metamorphism along the western part of this belt (ca. 640 Ma), has been dated by metamorphic zircons (Coolen et al., 1982; Muhongo, 1994; Appel et al., 1998; Muhongo et al., 2001). Hauzenberger et al. (2007) identified that the age of metamorphism further east, in the Galana River area ("GR" in Fig. 2), is considerably younger (580–530 Ma) overlapping the age of the 'Malagasy Orogeny' of Collins and Pisarevsky (2005).

THE SOUTHERN EAST AFRICAN OROGEN: THE MOZAMBIQUE BELT AND THE EAST AFRICAN OROGEN BIFURCATION

Tanzania, northern Mozambique, Madagascar

In Gondwana, Madagascar lay directly off the east coast of Tanzania and SE Kenya (Reeves et al., 2004)(Figs. 1, 2). Terranes in this area have been tentatively extended into western Madagascar (Fritz et al., 2013). For example, in the South Kenyan Galana River region, the Sobo Unit has been correlated with the pre-Neoproterozoic terranes of western Madagascar, while the Galana and Sagala Units are linked with the Vohibory Terrane of far SW Madagascar (Fritz and Hauzenberger, 2021)("V" in Fig. 2). The Rantosara shear zone of southern Madagascar also has been correlated with deformation along the Galana-Mutito Shear Zone (Hauzenberger et al., 2007). Unlike the Ranotsara shear zone ("rsz" in Fig. 2), though, the Galana-Mutito Shear Zone separates distinct terranes with different protolith origins and histories (Schreurs et al., 2010; Fritz et al., 2013; Fritz and Hauzenberger, 2021). To the east of the Galana-Mutito Shear Zone lies the microcontinent Azania (Fig. 2), whereas to the west are a series of Neoproterozoic terranes known as the Kasigau and Kurase Units in Kenya (Frisch and Pohl, 1986), the Eastern Granulites in Tanzania (Maboko, 2000)("EG" in Fig. 2) and the Cabo Delgado Nappe Complex in northern Mozambique (Viola et al., 2008)("CD" in Fig. 2). These originated as Neoproterozoic volcanic arcs that consumed the western branch of the Mozambique Ocean (Kröner et al., 2003; Cutten et al., 2006; Mole et al., 2018; Fritz and Hauzenberger, 2021).

Subduction of a strand of the Mozambique Ocean that separated the Congo/Tanzania/Bangweulu continent from Azania began as early as 1080 Ma with the formation of an intra-oceanic volcanic arc that is preserved today as the ca. 1080–980 Ma Dabolava Arc of western Madagascar (Tucker et al., 2014; Archibald et al., 2018)("Db" in Fig. 2). This arc system can be traced south in reconstructed Gondwana into the 'Tonian Oceanic Arc Superterrane' (TOAST) region of East Antarctica (Jacobs et al., 2015; Merdith et al., 2017b; Merdith et al., 2021)(Fig. 2). Later formed, early Tonian, supra-subduction arc systems included magmatism on the surrounding pieces of continental crust as well as within the ocean itself. These include the

Galana and Sagala units with emplacement ages between 970 Ma and 850 Ma and the ca. 930–900 Ma Vohibory arc terrane of SW Madagascar (Jöns et al., 2005; Jöns and Schenk, 2008; Collins et al., 2012)(Fig. 2). Protoliths to granulites in the Pare–Usambara and Uluguru Mountains ("Ulu" in Fig. 2) of Eastern Tanzania preserve subduction-related magmatism dated to ca. 700–800 Ma, which is broadly coeval with the extensive 850–750 Ma Itsindro-Imerona arc that intruded throughout central Madagascar (Handke et al., 1997; Archibald et al., 2016; Archibald et al., 2017). Finally, the youngest known subduction-arc terrane in this area forms the ca. 700–650 Ma Ntaka terrane of southern Tanzania (Mole et al., 2018)("Nt" in Fig. 2).

In Madagascar, accreted Late Mesoproterozoic to Tonian arc systems make up much of the western part of the exposed basement of the island (Collins, 2006; Tucker et al., 2014; Archibald et al., 2018; Boger et al., 2019; Collins et al., 2022a). The central part of Madagascar (the Antananarivo Domain, "Tana" in Fig. 2) forms a pre-Neoproterozoic continent made up of Archaean orthogneisses (Tucker et al., 1999; Tucker et al., 2011; Armistead et al., 2018; Archibald et al., 2023), overlain and interleaved with Palaeoproterozoic and Mesoproterozoic sequences of meta-sedimentary rocks (Cox et al., 1998; Cox et al., 2004; Boger et al., 2014b; Archibald et al., 2015; Armistead et al., 2021; Costa et al., 2021). To the east of this pre-Neoproterozoic unit lies a series of Neoproterozoic metasedimentary rocks that include pods of ultramafic rocks (the Anaboriana-Manampotsy belt). These rocks separate central Madagascar from the Antongil and Masora domains ("At" and "Ms" in Fig. 2, respectively); the former, at least, is a displaced part of the Western Dharwar Craton (Tucker et al., 1999; Tucker et al., 2014; Armistead et al., 2018). Collins et al. (2003) suggested that the Anaboriana-Manampotsy belt forms a cryptic suture (the Betsimisaraka Suture, "bet" in Fig. 2) between a Neoproterozoic continent (named 'Azania' by Collins and Pisarevsky, 2005) and Neoproterozoic India (Collins and Windley, 2002). This interpretation was countered by Tucker et al. (2011; 2014) who argued that central Madagascar was a continuation of the Dharwar Craton and that the Anaboriana-Manampotsy belt originated as an intra-continental basin. The Anaboriana-Manampotsy belt passes north to the Bemarivo Domain of far north Madagascar (Collins and Windley, 2002)("Bem" in Fig. 2). Here, Armistead et al. (2019) demonstrated that the northern part of this region formed a juvenile ca. 750–700 Ma

terrane that they named the Bobakindro Terrane. These authors argued that it formed a continuation of the volcanic arc terranes that accreted to Neoproterozoic India during the Tonian and Cryogenian (Blades et al., 2020; Collins et al., 2021b) and correlated south with the Betsimisaraka Suture to separate Neoproterozoic India from Azania and African-derived terranes.

The foreland to the Lurio Belt of northern Mozambique consists of the Stenian Nampula Complex ("Na" in Fig. 2), which is relatively weakly deformed and metamorphosed in the Neoproterozoic (Kroner et al., 1997; Sacchi et al., 2000; Bingen et al., 2009). The Lurio foreland disrupts the orogenic trend of the EAO and splits it into southwestern and southeastern branches with a triple junction that lay directly south of Madagascar and Southern India (Schmitt et al., 2018; Clark et al., 2020)(Figs. 1, 2). The southwestern branch passes through Malawi and as it tracks further west, it is progressively known as the Zambezi Belt, the Lufilian Arc (Fig. 2) and eventually the Central Damara Orogen in Namibia ("Da" in Fig. 1). The southeastern branch forms a broad orocline, curving through Sri Lanka where it encompasses the Wanni and Highland Complexes ("W" and "HC" in Fig. 2, respectively). The orocline wraps around the Stenian Vijayan Complex (correlated with the Nampula Complex by Collins and Pisarevsky, 2005)("V" in Fig. 2) and passes into East Antarctica in the region of Lutzöw Holm Bay ("Lut" in Fig. 2).

THE SOUTHWEST BRANCH OF THE EAST AFRICAN OROGEN

Northern Mozambique, Malawi, Zambia

Northern Mozambique and southeastern Zambia experienced extension and deposition in the early Neoproterozoic. Metasedimentary units in the Marrupa and Unango complexes of Mozambique (Bingen et al., 2009)("M-U" in Fig. 2) and the Nyimba-Sinda Terrane of Zambia (Alessio et al., 2019a)("N-S" in Fig. 2) likely reflect basin development contemporaneous with late Tonian/early Cryogenian rift-related magmatism in NE Mozambique (Bjerkgard et al., 2009), Malawi (Ashwal et al., 2007) and east Zambia (Johnson et al., 2006). This rifting is interpreted as associated with the breakup of central Rodinia and connected with the initial opening of the Pacific Ocean basin (Merdith et al., 2017a). Subsequent collision between the Kalahari continent and the Congo-Tanzania-Bangweulu Block formed some of the younger orogenesis that made-

up Gondwana (Schmitt et al., 2018; Gomes et al., 2025). Peak metamorphism within the central Damara Orogen ("Da" in Fig. 1) culminated at 530–525 Ma (Goscombe et al., 2017), which matches similar Cambrian peak metamorphism in the Zambezi Belt (John et al., 2004). Post-peak cooling was well underway through much of the belt by the end of the Cambrian (Alessio et al., 2019b; Vincent et al., 2026).

THE SOUTHEAST BRANCH OF THE EAST AFRICAN OROGEN

South India, Sri Lanka, East Antarctica

Southern India has been divided into a series of tectonic units with distinct protolith origins and tectonothermal histories. From north to south these are: the Salem Block (sometimes called the Northern Granulite Terrane); the Palghat-Cauvery shear system ("pcss" in Fig. 2); the Madurai Block ("Mad" in Fig. 2); the Achancovil Zone ("asz" in Fig. 2); the Trivandrum Block (also known as the Kerala Khondalite Belt); and the Nagercoil Block (both amalgamated as "TN" in Fig. 2) (e.g. Santosh et al., 2009; Collins et al., 2014). The Palghat-Cauvery shear system has long been noted to separate autochthonous parts of Neoproterozoic India from more exotic terranes to the south (Bartlett et al., 1998; Cenki et al., 2004; Collins et al., 2014; Plavsa et al., 2015).

The northwest of the Madurai Block is composed of charnockite massifs with Sm–Nd and Rb–Sr whole rock model ages in the range of 3170–2200 Ma (Bartlett et al., 1998; Bhaskar Rao et al., 2003). U–Pb zircon ages from charnockites in this northwest Madurai Block have yielded a consistent range of crystallisation ages between ca. 2.55 and 2.50 Ga (Ghosh et al., 2004; Brandt et al., 2011; Plavsa et al., 2012). This northern region is separated from the rest of the predominantly metasedimentary Madurai Block first by a lithological boundary that broadly follows the Karur-Kamban-Painavu-Trichur lineament (known as the KKPT) (Ghosh et al., 2004) that can also be seen as a zone of S-dipping reflectors in deep seismic surveys (Rajendra Prasad et al., 2007). South of the KKPT ("kkpt" in Fig. 2) is a broad zone of highly deformed and metamorphosed Archaean/Palaeoproterozoic metasedimentary rocks that continue until a line approximately from Rajapalayam to Madurai (Plavsa et al., 2014; Kumar et al., 2017). Neoproterozoic metasedimentary rocks with Mesoproterozoic to Neoproterozoic metaigneous

689 plutons that have continental arc-like chemistries dominate the region south of this
690 Rajapalayam–Madurai line and pass into the Achankovil Zone where Neoproterozoic rocks are
691 deformed into a prominent SE–NW striking high strain zone (Guru Rajesh and Chetty, 2006).
692
693 Some workers have subdivided the southern part of Kerala and Tamil Nadu into the Trivandrum
694 and Nagercoil Blocks (Santosh et al., 2003; Santosh et al., 2006). U–Pb zircon data from the
695 charnockites that are common in the region demonstrate that both blocks preserve similar ages
696 and really seem to be differentiated just on the dominance of charnockite in the Nagercoil Block
697 and metasedimentary rocks in the Trivandrum Block (Cenki et al., 2004; Kröner et al., 2012;
698 Johnson et al., 2015; Kumar et al., 2017; Clark et al., 2020). The metaigneous protoliths formed
699 at ca. 2040 Ma in an oceanic magmatic arc setting and are distinct from elsewhere in southern
700 India, but likely correlate with the basement of the Androyen and Anosyen terranes of southern
701 Madagascar (Tucker et al., 2014), possibly the Highland Complex of Sri Lanka (Kitano et al., 2018)
702 and even with parts of East Africa (Clark et al., 2020). The area including all these regions, with
703 the exception of East Africa, was named “SMIWH” (South Madagascar, India, Wanni-Highland
704 Complex) by Tucker et al. (2014), who argued that this terrane accreted to the Dharwar Craton
705 by the late Palaeoproterozoic. We note that this terrane is separated from the pre-1000 Ma crust
706 of India by the Neoproterozoic southern Madurai and Achancovil terranes (Plavsa et al., 2012;
707 Plavsa et al., 2015), which leads us to disagree with the proposed Palaeoproterozoic accretion
708 and suggest that this terrane combined with cratonic India during the Ediacaran high-grade
709 metamorphism seen throughout the region (Collins et al., 2007; Clark et al., 2009; Collins et al.,
710 2014; Taylor et al., 2014; Plavsa et al., 2015). We also point out that no reliable evidence exists
711 for pre-Stenian basement in the Sri Lankan Wanni Complex, and that it shouldn’t be considered
712 part of this terrane; the acronym SMIWH then is inaccurate.
713
714 We suggest that the contiguous Gondwanan terrane encompassing the Androyen/Anoysen
715 domains of Madagascar, the Trivandrum Block of southmost India and the Highland Complex of
716 Sri Lanka be called ‘Pandya’ after i) the named used for the metamorphic belt of southern India

(Pandyan Mobile Belt, Ramakrishnan, 1993), ii) the ancient Tamil dynasty of southern India, whose lands encompassed parts of Sri Lanka at their greatest extent.

DISCUSSION

The Plate Tectonic Evolution of the East African Orogen

The tectonic geography represented by the various terranes of the East African Orogen can be difficult to deduce where extreme metamorphism and deformation has transposed all physical evidence and often rendered geochemical evidence enigmatic. Such obfuscation is a particular challenge in the southern East African Orogen—a region known as the Mozambique Belt (Fritz et al., 2013). Along orogenic strike, though, outcrop metamorphic grade decreases towards the north—from granulite and upper amphibolite grade rocks in the southern EAO to greenschist facies and sub-greenschist facies rocks preserved in the northern parts of the EAO . By comparison, in the northern EAO most surface rocks experienced metamorphism of no more than greenschist-facies and original sedimentary/volcanic features are commonly preserved as are near-original chemistries of the rocks. It is here in the Arabian-Nubian Shield where many pre-continent/continent collision tectonic events that affected the region can be best defined (Johnson et al., 2011; Collins et al., 2021a; Collins et al., 2021b).

The East African Orogen—the Largest Mountain Belt of the Last Billion Years?

There is a good case to suggest that the EAO is the largest collisional orogen of the last billion years simply based on its footprint on a reconstructed Gondwana. Extending across approximately 100 degrees of longitude at low to moderate southern latitudes (Collins et al., 2021b) the orogen was over 10,000 kms long and extend over 1000 km at its widest point (in the region now preserved in Mozambique, Madagascar and Southern India)(Figs. 1, 2). The scale is comparable to the modern Alpine-Himalaya chain, and like this modern orogen, the EAO was large enough to strongly change the oceanic radiogenic strontium isotopic curve (Chen et al., 2022). In fact, global ocean $^{87}Sr/^{86}Sr$ values only rise dramatically twice in the last billion years; once from 50 Ma to present, due to erosion of the growing Himalaya, and in the late Neoproterozoic (Chen et al., 2022). Zhu et al. (2022) looked at the scale of orogens through time

from a different perspective. By examining the rare earth element chemistry of dated detrital zircons, they showed that there are two main times of 'supermountain' formation, one in the Palaeoproterozoic and a second contemporaneous with the EAO. The youth of the Himalayan orogen is the likely reason for no significant record of it in these detrital zircons. Therefore, as well as the EAO (along with contemporaneous Gondwana-forming orogens in Africa and South America) being of a scale to affect ocean water composition, it also thickened the crust enough to significantly affect the trace-element chemistry of zircon-producing magmas. This suggests that the orogen was not only spatially extensive, but also formed mountains of an elevation at least as high as those seen in Asia today. As proposed by Zhu et al. (2022), erosion of this 'supermountain' would likely stimulate biological productivity and widely effect the wider planetary system (e.g., Squire et al., 2006; Halverson et al., 2010). The orogen's broad E–W orientation and location in relatively low to moderate latitudes likely accentuated its erosion and multiplied its effect on other Earth systems (in a similar manner to the Late Palaeozoic orogens demonstrated by Godderis et al., 2017). Merdith et al. (Submitted) demonstrated the importance of this within a computational whole-Earth, where they traced the evolution of tectonic plate boundaries, to automatically isolate specific tectonic environments that are associated with mountain building, such as continental arcs, continental collisions, rifts and large igneous provinces. From this they simulated orogenic growth, and erosion, over the last 1 Ga, using parameters and limits derived from the present-day expression of topography on Earth to provide a quantitative and tractable basis for palaeoclimate or landscape evolution modelling over the last 1 Ga. This semi-automated topography model uses 'rules' relating orogenic topography to the nature and location of plate boundaries. The next step is to use geological data to constrain topographic evolution directly from nature.

To obtain a better understanding of the role the EAO plays in the tumultuous changes to the climate, biosphere and atmosphere in the Neoproterozoic an understanding of the spatial and temporal evolution of elevation through the orogen is needed. Unravelling topography from the now denuded and dispersed remains of the East African Orogen mountains is a complex task. We have taken the approach here of mapping pressure and time determinations from the literature

on a reconstruction of Gondwana (Merdith et al., 2021). We limit our analyses to exclude low geothermal gradient metamorphism (< 450 K/GPa) as this is usually interpreted as subduction-zone related (Brown and Johnson, 2019). In this situation, pressure (depth) will not necessarily relate to surface elevation due to the non-isostatic nature of these environments. We have also ignored the possible effects of differential stress (Schmalholz and Podladchikov, 2014; Tajčmanová et al., 2021; Zuza et al., 2022) due to the difficulty of quantifying its effect on mean stress, metamorphic determinations (pressure).

A simple isostatic correction to present day elevation can be produced from metamorphic pressure estimates to estimate palaeo-elevation (Fig. 3). Given a peak metamorphic pressure, P_s, at the time of metamorphism, the height of eroded material, *h*, can be estimated from an assumed density (2850 kg m^{-3}) for the eroded material, ρ, and gravity, g, by

$$h = \frac{P_s}{\rho g}.$$

Assuming the lower crust and mantle lithosphere at the time of metamorphism is the same today, their contributions cancel out of the isostatic balance; thus mantle density, ρ_m, and present day elevation, ε_{pd} are the only additional parameters needed to compute the elevation at the time of peak metamorphism, ε_{EAO}. Using the local isostatic balance equations: (1) $\rho h = \rho_m h_m$ and (2) $h = h_m + \varepsilon_{EAO} - \varepsilon_{pd}$, where h_m is the thickness of mantle needed to compensate the change in load, we estimate the palaeo-elevation by,

$$\varepsilon_{EAO} = \varepsilon_{pd} + \frac{P_s}{\rho g}\left(1 - \frac{\rho}{\rho_m}\right).$$

A total of 197 metamorphic conditions have been compiled (Table 1), supplementing the P–T database by Brown and Johnson (2019). Present-day elevation was determined by averaging pixels in a 25 km radius about the sample location, though in many cases locations are approximated from a map or description. While it is beyond the scope of this study to produce a rigorous elevation model, which ideally would incorporate thermal effects and compositional uncertainties (Hasterok and Chapman, 2007), where pressure ranges have been provided an elevation range was computed (Table 1).

804 Predicted elevations range from near sea-level to a maximum of ca. 10 km, with an average of
805 5.1 ± 1.6 km, similar in elevation to the modern Tibetan Plateau, and over a similar area (Figs. 1,
806 5). The modelled EAO palaeo-elevations are greater than the global average palaeo-elevations
807 (Fig. 4) and are normally distributed. One may expect a more exponential distribution for
808 elevation, though it is unsurprising that elevation predicted from peak metamorphic pressures
809 does not. The primary reason for this difference is the lack of precise recording of pressures at
810 low metamorphic grades, which are difficult due to kinetic rates at low temperatures. Hence our
811 estimates likely yield elevation estimates in the orogenic plateau and range crest (Fig. 5). Since
812 there is no obvious decrease in elevation at the eastern and western edges of the plateau, the
813 area of elevated terrain is likely much greater than the distribution of model estimates.
814
815 Based on these simple elevation estimates, we suggest that the high topography varied spatially
816 in time as the orogen evolved. High elevations were created in the first phase of the orogen
817 (680–640 Ma; i.e. the East African Orogeny of Collins and Pisarevsky, 2005), recorded as high
818 elevations > 6 km in the Tanzanian Eastern Granulites associated with shortening during
819 subduction and collision of Azania with the Congo-Tanzania-Bangweulu Block (Fig. 5). In the
820 second phase (640–580 Ma), the elevation remained high in the Mozambique Belt and southern
821 Azania. At this time, many regions in the Arabian-Nubian Shield also reached peak metamorphic
822 pressure conditions, though at lower maxima that result in a lower topography (~ 2.5–4 km, Fig.
823 5 and Table 1). The final stage of the EAO (The Malagasy Orogeny of Collins and Pisarevsky, 2005)
824 resulted in the development of an extensive orogenic plateau, with the highest modelled
825 elevations (8–10 km) in the south, in present-day coastal East Antarctica, and throughout a broad
826 plateau that extended from the margin of the Congo–Tanzania–Bangweulu continent, across
827 Azania in present-day Madagascar, Sri Lanka and the Southern Granulite Terrane of India (Figs.
828 1, 2, 5). It is unlikely that the highest elevations are accurate as regional compensation would
829 likely have resulted in lower elevations, however, we assert that the aerial extent of the high
830 elevations in the plateau regions is reasonable, i.e., within the bounds of elevation observed in
831 Tibet and the Himalaya today (Fig. 4). On the basis of these estimates, we suggest that the EAO
832 was amongst the largest, highest mountain belts known on Earth, rivalling the Alpine-Himalayan

Orogen of today and lending support to hypotheses that have the orogen fundamental to earth surface systems tipping points, such as facilitating carbon burial and planetary oxygenation that are broadly coeval with the orogen's apogee (e.g., Zhu et al., 2022).

Suggestions for further work

A grand challenge of earth science is to be able to map the evolution of Earth's surface systems through its history. This is a herculean task, but a worthy task, as it is only by developing a deeper understanding of the geological record that can drive more naturalistic models of the planet that we can understand how our planet works, and how the geosphere interacts with the atmosphere to develop the climate, hydrosphere and biosphere (e.g., Mills et al., 2025). Developing a model of the Earth system over the ca. 500 million years of the Neoproterozoic involves reconstructing both the plate tectonic evolution of the planet, and the evolution of the planet's surface elevation.

The reconstruction presented here (Figs. 1, 2), and the palaeo-geographically controlled palaeo-elevation estimates (Fig. 5), are only the first attempt at integrating the full plate tectonic record with the metamorphic record to track the evolution of this one (albeit large) mountain range. Much of the geology of the East African Orogen is poorly known, especially when compared with that of Phanerozoic mountain ranges such as the European Alps, North American Appalachians or Chinese Qinling mountains. Much work is needed to unravel the paleo-plate tectonic evolution of the orogen, especially in the countries of east and northeast Africa, Arabia, Madagascar, South Asia and East Antarctica. This is in addition to the other Neoproterozoic orogens that lace South America, Australia and parts of Asia.

We hope to have demonstrated that when triaged carefully, the metamorphic record preserves palaeo-topographic information that is difficult to recover by other methods. We note that much of the available metamorphic record through Proterozoic orogens is governed by point analyses on restricted outcrops where high variance lithologies occur that are used to constrain the

862 pressure-temperature and time evolution of that one location. We encourage future pressure-
863 time campaign-style studies that may be able to determine the geometry of orogenic topography
864 and along-strike variations that will considerably improve the resulting models.
865
866 As an example of what is possible. We undertook a reconnaissance pressure-time study of
867 southern India that was reported in Collins et al. (2022b). Here we dated garnets using the in-situ
868 Lu-Hf technique over an orogenic transect (Tamblyn et al., 2022) and examined the trapped
869 pressures of quartz inclusions using RAMAN spectroscopy to obtain peak pressures. Many
870 exhumed orogens are garnet-rich and a similar approach could be done through many of the
871 Proterozoic orogens to better understand their topographic evolution.
872
873 The answer in the end involves a community-wide approach to supporting geological research
874 globally, and especially in the global south. Then integrating plate-tectonic reconstructions
875 developed from these works into a global reconstruction. Topography can then be added by
876 combining estimates of past elevation from the proximity of a region to a plate boundary (in the
877 style of Merdith et al., Submitted) with estimates from geochemistry of igneous rocks (e.g. Zhu
878 et al., 2022; Zhou et al., 2025) and estimates from the metamorphic record as discussed here.
879
880 **CONCLUSIONS**
881 The East African Orogen is one of the largest known orogens on the planet, on a similar, or greater
882 scale, than the present-day Alpine-Himalaya orogen. Reconstructions of the orogen show that it
883 not only bisected Gondwana, but it also developed a thick orogenic plateau that rivals Tibet for
884 elevation and area (Fig. 1). The East African Orogen evolved from a Mesoproterozoic accretionary
885 orogen that developed to the east of the West African Craton and north of the Congo-Tanzania-
886 Bangweulu Block. Throughout the Tonian accretionary orogenesis characterized the orogen with
887 considerable volcanic-arc development on the margins of the growing nuclei of Africa (soon to
888 be central Gondwana) and within the Mozambique Ocean. The collision of Azania (and other
889 intra-Mozambique Ocean continents) with the active margin of Africa caused an early continent-

890 continent collisional orogeny—the East African Orogeny. This was later followed by the Ediacaran
891 collision of Neoproterozoic India with the growing kernel of central Gondwana.
892
893 The available metamorphic record supports the designation of the East African Orogen as, if not
894 the largest, then one of the largest mountain ranges in Earth history (Fig. 4). Past ground
895 elevation estimates constructed from the triaged metamorphic record are located on the full-
896 plate tectonic reconstruction of Cao et al. (2024), which is fundamentally based on Merdith et al.
897 (2021), and reveals the elevation consequences of the series of plate-tectonic events (Fig. 4). This
898 also demonstrates the along-strike variability along the orogen, with contemporary palaeo-
899 elevations in the Arabian-Nubian Sheild region being considerably lower than in the Mozambique
900 Belt, reflecting the geometry and kinematics of the plate-movements at the time. Our plate-
901 tectonic/palaeo-elevation approach is suggested as a way forward in reconstructing the shape of
902 the Earth in deep time—a key parameter in future earth systems models of the planet.
903
904 **ACKNOWLEDGMENTS**
905 ASC and MLB are funded by Australian Research Council grants FL240100114 and DP250102570.
906 ASM is funded by DE230101642. ASC thanks the numerous colleagues and students who have
907 assisted on the projects that led to his fascination with the East African Orogen and the formation
908 of Gondwana.
909

910

911 **Table 1 - Peak metamorphic conditions and model paleo elevations within the East African Orogen.**

912 **(see spreadsheet).**

FIGURE CAPTIONS

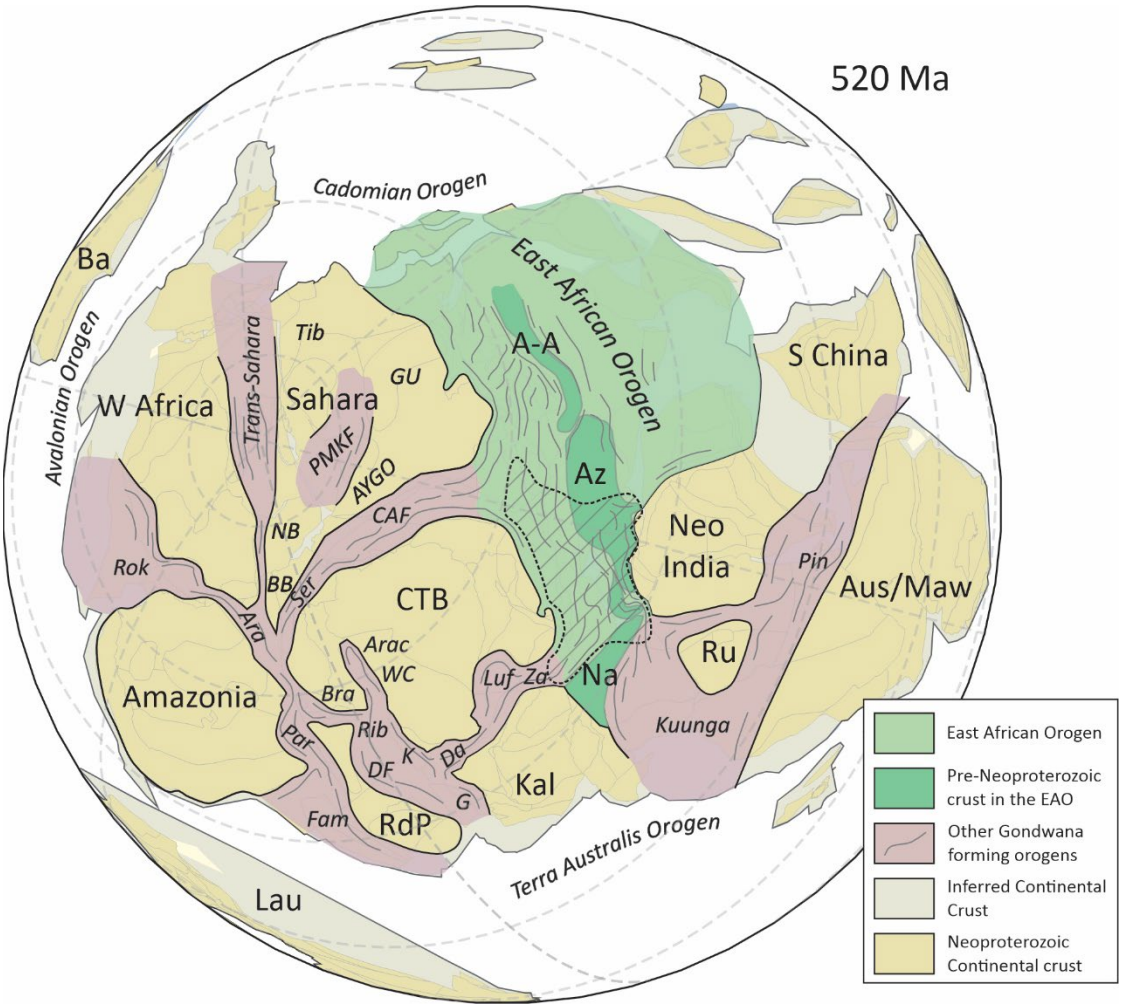

Figure 1: Reconstruction of Gondwana in the early Cambrian (520 Ma), after Merdith et al. (2021) with the East African Orogen (EAO) outlined along with major regions of Pre-Neoproterozoic crust reworked within the EAO; A-A = Afif-Abas Terrane, Az = Azania, Na = Nampula Terrane. Other major Gondwana-forming orogens are outlined, which are; Ara = Araguaia Belt, Arac = Aracuaí Orogen, Bra = Brasília Belt, CAF = Central African Fold Belt, Da = central Damara Belt, DF = Dom Feliciano Belt, Fam = Famatinian Orogen, G = Gariep Belt, K = Kaoko Belt, Kuunga = Kuunga Orogen, Luf = Lufilian Arc, Par = Paraguay Belt, Pin = Pinjarra Orogen, Rib = Ribeira Belt, Rok = Rokelide Orogen, Ser = Sergipano Orogen, Trans-Sahara = Trans-Saharan Orogen, WC = West Congo Orogen, Za = Zambezi Belt. The major Neoproterozoic continental blocks are; Aus/Maw =

Australia/Mawson Continent, Ba = Baltica, CTB = Congo-Tanzania-Bangweulu (also includes the São Francisco craton of Brazil), GU = Gebel Uweinat, Kal = Kalahari, Lau = Laurentia, Neo India = Neoproterozoic India, PMKF = Poli-Mayo Kebbi-Lake Fitri Terrane, W Africa = West Africa (including the São Luis Block of Brazil), RdP = Río de la Plata (and including the putative Paranapanema craton), Ru = Ruker Terrane (likely a continuation of Neoproterozoic India), Sahara = Saharan basement (including the Nigeria-Benin block [NB], the Brazilian Borborema Province [BB], Tibesti [Tib], Gebel Uweinat [GU] and the Adamawa-Yadé-Guéra-Ouaddaï Terrane [AYGO]), S China = South China. The broad sites where three of the Gondwana-peripheral orogens will form are also marked, these are the Avalonian Orogen, the Cadomian Orogen, and the Terra Australia Orogen. The cross-hatched zone in the EAO is where the highest-grade metamorphic rocks from the East African orogen and Malagasy orogens are found, which approximates to the position of the Mozambique Belt and forms the past position of the large orogenic plateau discussed in this paper. Broad orogenic strike lines are added within the Gondwana-forming orogens.

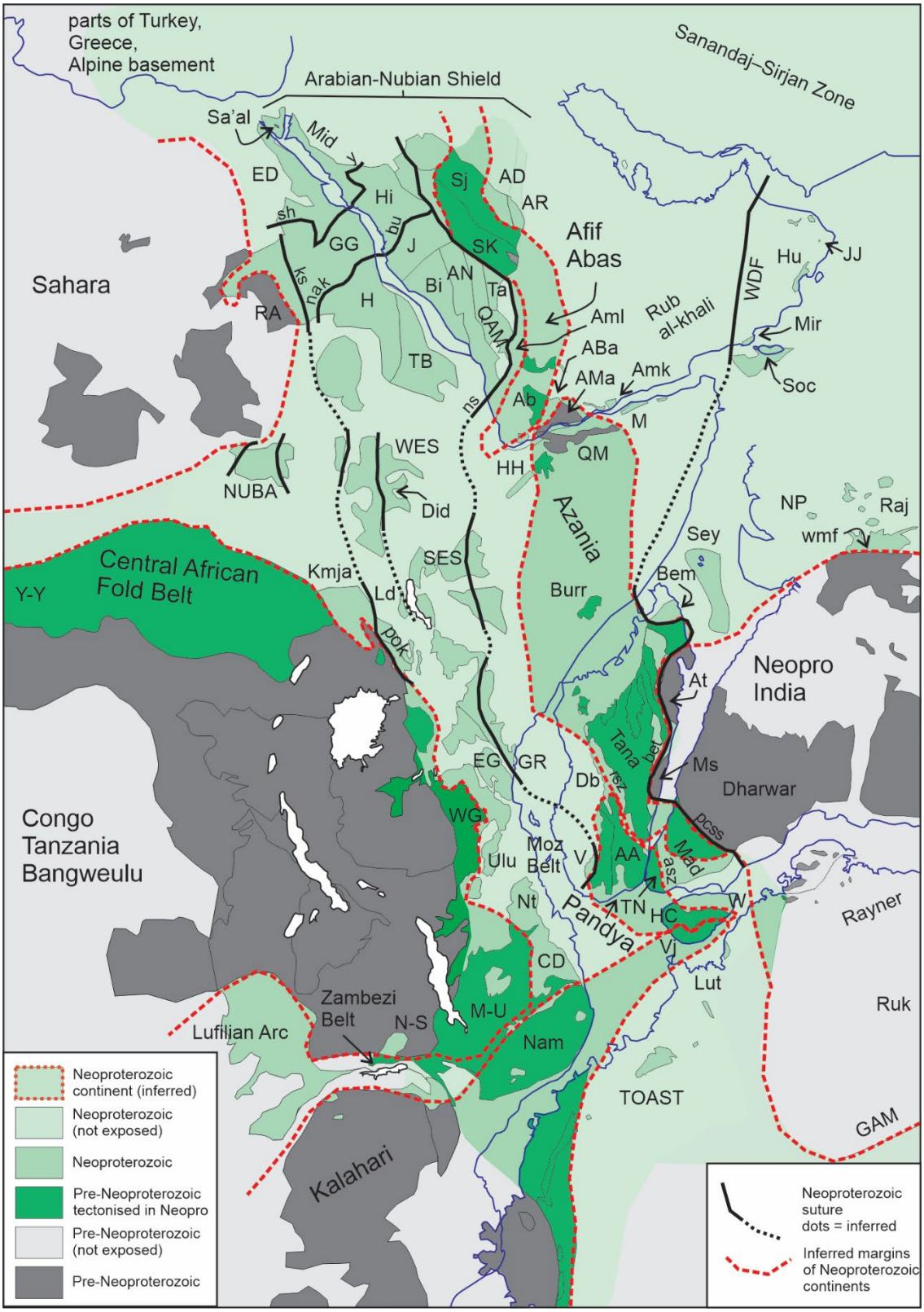

Figure 2: Reconstruction of Central Gondwana at 550 Ma (using the GPlates reconstruction of Merdith et al. 2021 as the basemap). Crust that was significantly deformed and metamorphosed

(tectonised) in the Neoproterozoic, or formed in the Neoproterozoic, are coloured shades of green. Abbreviations used are: AA = Anosyen-Andasibe domains, Ab = Abas terrane, ABa = Al Bayda terrane, AD = Ad Dawadmi terrane, AMa = Al Mafid terrane, Amk = Al Mukalla terrane, Aml = Amlah terrane, AN = An Nimas terrane, AR = Ar Rayn terrane, At = Antongil Domain, Bem = Bemarivo Belt (the Bobakindro terrane is the northernmost part of this belt), Bi = Bidah terrane, CD = Cabo Delgado Complex, Db = Dabolava Suite (arc), Did = Didesa terrane, ED = Eastern Desert of Egypt, EG = Eastern Granulites, GG = Gabgaba/Gebeit terrane, GR = Galana River, H = Haya terrane, HC = Highland Complex, HH = Harar-Hirna, Hi = Hijaz terrane, Hu = Huqf, J = Jiddah terrane, JJ = Jebel Ja'alan, Kmja = Karamoja Belt, Ld = Lodwar, Lut = Lutzöw-Holm Bay, M = Maydh (or Mait) complex, Mad = Madurai Block, Mid = Midyan terrane, Mir = Mirbat, Ms = Masora Domain, M-U = Marrupa-Unango complexes, Nam = Nampula Complex, NP = Nanga Parkar, N-S = Nyimba–Sinda Terrane, Nt = Ntaka terrane, NUBA = Nuba Mountains, QAM = Al Qarah-Malahah terrane, QM = Qabri Bahar and Mora complexes, RA = Rahaba-Absol terrane, Raj = Rajasthan, Ray = Rayner Complex, Ruk = Ruker domain, Sa'al = Sa'al Metamorphic Complex, Sahara = Sarahan basement (including pre-Neoproterozoic crustal terranes and Neoproterozoic juvenile terranes), SES = Southern Ethiopian Shield, Sey = Seychelles, Sj = Suwaj, SK = Siham-Khida terrane, Soc = Socotra, Ta = Tathlith terrane, Tana = Antananarivo Domain, TB = Tokar/Barka terrane, TN = Trivandrum-Nagercoil blocks, Ulu = Uluguru Mountains, V = Vohibory Domain, Vj = Vijayan Complex, W = Wanni Complex, WES = Western Ethiopian Shield, WG = Western Granulites, Y-Y Yaoundé-Yangana nappes. Shear zones and sutures are marked with the following acronyms; asz = Achancovil shear zone, bet = Betsimisaraka Suture (in the Anaboriana-Manampotsy belt), bu = Bi'r Umq suture, kkpt = Karur-Kamban-Painavu-Trichur lineament, ks = Keraf Suture, nak = Nakasib suture, ns = Nabitah suture, pok = West Pokat suture, pcss = Palghat-Cauvery Shear System, rsz = Ranotsara shear zone, sh = Sol-Hamed Suture, wmf = Western Margin Fault, y = Yanbo suture.

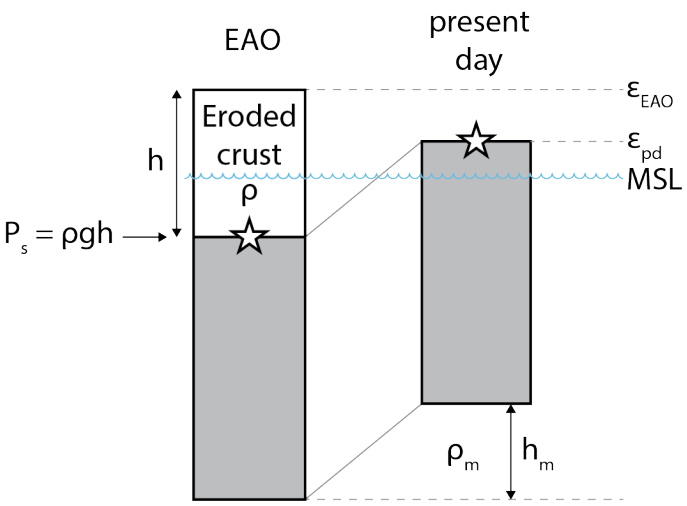

Figure 3: Isostatic columns used to estimate elevation during the East African Orogen from metamorphic pressure estimates. Isostatic column during the East African Orogen (left) included overburden that eroded (average density, ρ, and thickness, h), exposing the metamorphic pressure (P_s, star) at present day (right).

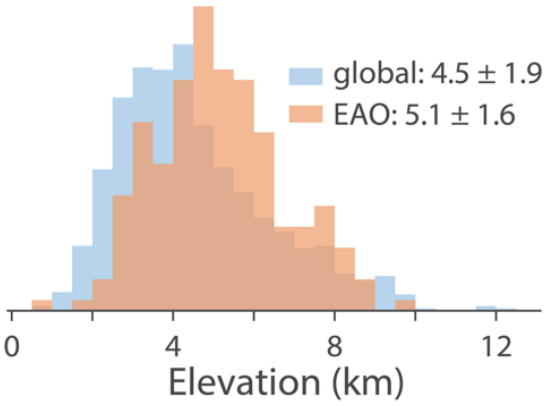

Figure 4: Palaeo-elevation estimates for the East African Orogen (EAO) compared to the global dataset excluding EAO data. Individual palaeo-elevation estimates are provided in Table 1. The global estimates are derived from pressures included in an expanded metamorphic database originally by Brown and Johnson (2019)). Data were prefiltered for metamorphic pressures below 2 GPa and metamorphic thermal gradients above 450 K/GPa to remove most subduction-related events.

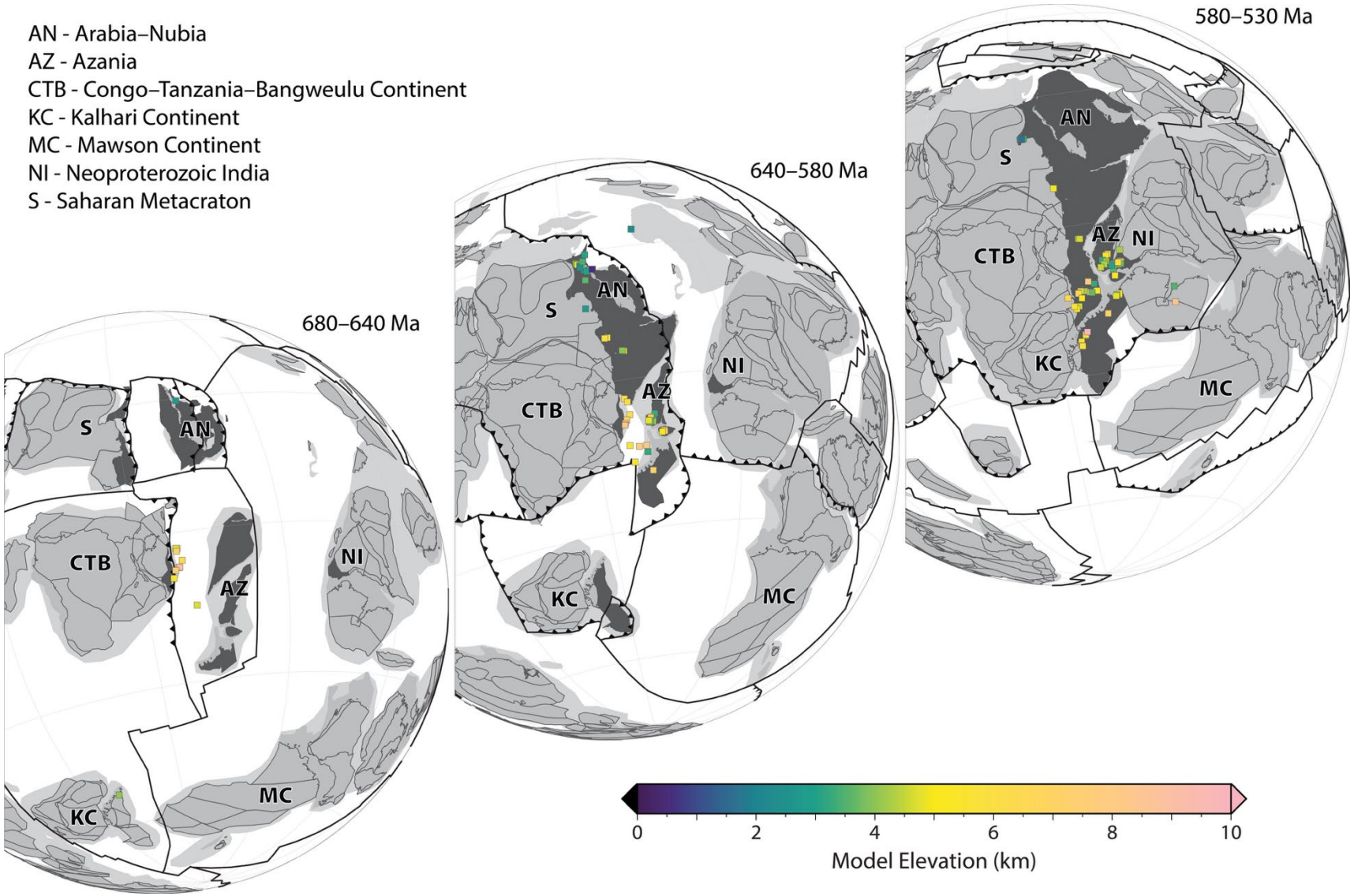
AN - Arabia–Nubia
AZ - Azania
CTB - Congo–Tanzania–Bangweulu Continent
KC - Kalhari Continent
MC - Mawson Continent
NI - Neoproterozoic India
S - Saharan Metacraton
680–640 Ma
640–580 Ma
580–530 Ma
S
AN
CTB
AZ
NI
KC
MC
0
2
4
6
8
10
Model Elevation (km)

Figure 5: Palaeo-elevation estimates for the East African Orogen. The P–T locations have been reconstructed from the tectonic model by Cao et al. (2024) at time steps 660, 600, and 550 Ma, corresponding to the palaeo-elevation estimates at 680–640, 640–580, and 580–520 Ma, respectively. The plate models are fixed relative to the Congo-Tanzania-Bangweulu Block.

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