New pressure-temperature-time constraints on initial India-Asia collision from Eastern Himalayan Syntaxis eclogite-facies metamorphic rocks

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ABSTRACT

Eclogite-facies rocks exposed within the Himalayan Range can be used to constrain the timing and characteristics of the India–Asia collision, which remain strongly debated. Many collision-related high- and ultrahigh-pressure lithologies are exposed around the western Himalaya; however, the rarity of similar units in the eastern Himalaya limits our understanding of the along-strike geodynamic evolution of India–Asia collision and the characteristics of the subducted India plate. Here, we document ~~newly identified~~ collision-related, low-temperature eclogite-facies metapelites from the easternmost Indus-Yarlung suture zone, which retain vital records of the timing of

the initial India–Asia collision in this region. Petrology and phase diagram modeling indicate that the metapelite experienced cold subduction along a low thermal gradient (~12 ℃/km), reaching peak low-temperature eclogite-facies conditions (ca. 1.5 GPa and 590 ℃), and was retrogressed to amphibolite-facies conditions during exhumation. Zircon and monazite U–Th–Pb chronology reveals new eclogite-facies ages of ca. 47 Ma, constraining collision of the Indian and Asian continents in the eastern Himalayan syntaxis at ca. 50 Ma, which is consistent with the timing in the western Himalaya and east-central Himalaya. This demonstrates that the initial India–Asia collision occurred quasi-synchronously along orogenic strike, although the maximum depth of subduction of continental material differed. Moreover, the subduction dynamics of continental crust within the collisional orogen exhibit pronounced lateral heterogeneity. These findings further demonstrate substantial along-strike variations in slab dip angles during the closure of the Neo-Tethys Ocean, and such spatiotemporal complexity must be incorporated into future geodynamic models to better resolve orogenic evolution mechanisms.

Keywords: P–T path, eclogite facies rock, continental collision, Eastern Himalayan Syntaxis, Indus-Yarlung suture zone

1. Introduction

The India–Asia collision is one of the most widely and intensively studied tectonic events of the Phanerozoic (Yin and Harrison, 2000). The timing of the initial collision provides important data for geodynamic models of Himalayan–Tibetan orogenic evolution, mechanisms of plateau uplift, intracontinental deformation, the impact of Himalayan orogenesis on global cooling, changes in marine geochemistry, and faunal exchange between the continents (Ding et al., 2022). A significant debate persists regarding whether the initial India–Asia collision was synchronous (Najman et al., 2010; Hu et al., 2015; 2016) or diachronous (Wu et al., 2014) along its ~2500 km length. As such, discovery of key rock types within the Himalayan Range or Tibetan Plateau that formed during terminal closure of the Neo-Tethyan Ocean would provide critical new insight into this issue and allow hypothesis testing.

Significant advancements have been made in the past two decades in constraining the timing of key stages of India–Asia continental collision and subsequent orogenic growth. The timing of initial collision has been independently defined by many different workers via a range of techniques (see reviews by Yin and Harrison, 2000; Guillot et al., 2003; Najman, 2010; van Hinsbergen et al., 2011). These studies have yielded interpreted collision ages that span the period 70 Ma to 35 Ma, although most workers show a preference for ages around 55–50 Ma (Najman et al., 2010; van Hinsbergen et al., 2011; Wu et al., 2014; DeCelles et al., 2014; Ding et al., 2016; Hu et al., 2015, 2016).

During collisional orogenesis, deep subduction of the continental crust can only occur immediately following the onset of collision between both continental terranes, when the buoyant continental crust is attached to descending fragments of the oceanic lithosphere. The age of high- or ultrahigh-pressure (HP/UHP) rocks preserved in

collisional orogens is thus ideal to constrain the minimum age of collision, with a lag time calculated by combining these ages of peak metamorphism with the subduction velocity of the underthrust continent (e.g. Ding et al., 2016). However, (U)HP rocks are rarer in the Himalaya than in other continental orogenic belts, including the contemporaneous Alpine orogen. The two most-studied UHP eclogites within the orogen are both exposed in the western Himalaya, and have peak metamorphic ages recording collision between India and Asia in the west at ca. 51 Ma (Kaghan; Kaneko et al., 2003; Parrish et al., 2006; Donaldson et al., 2013) and 50.8 ± 1.4 Ma (Tso Morari; St-Onge et al. 2013). While some retrograde eclogites have been reported in the central and eastern Himalaya, most have been subjected to (ultra) high-temperature overprinting during exhumation, driven by crustal thickening of the Himalaya during the process of continental convergence in the Middle Oligocene, or to the southern Tibet thrust fault (Wang et al., 2021; Zhang et al., 2024). This (ultra) high-temperature overprinting occurred at c. 30 Ma (Wang et al., 2021) and c. 17–14 Ma (Li et al., 2019; Wu et al., 2022; Zhang et al., 2024), essentially removing all evidence for the timing of eclogite-facies metamorphism in these rocks. Recent studies have shown that metamorphic rocks in the east-central Himalaya underwent medium-pressure (MP) metamorphism, and the initial timing of the India-Asia collision is constrained at ca. 50 Ma in the east-central Himalaya (Ding et al., 2016). However, no reports of collision-related UHP or HP rocks have been made in the eastern Himalaya (O'Brien et al., 2019; Fig. 1A), directly impeding our understanding of the along-strike geological evolution of the Tibetan Plateau and Himalayan range.

In the eastern Himalayan syntaxis (Fig. 1B), detailed investigations into early subduction-associated high-pressure low-temperature (HP-LT) metamorphism are lacking. In this study, we document the petrological characteristics and U–Pb zircon

geochronology from two samples of low-temperature eclogite-facies schists in the Lulang area, eastern Himalayan syntaxis, southern Tibet, situated within the eastern Indus-Yarlung suture zone (IYSZ: Fig. 1B and C). These rocks represent relics of sedimentary material that was subducted alongside oceanic crust during closure of the Neo-Tethys Ocean, and provide key constraints on the age and geodynamics of initial India–Asia collision in the eastern IYSZ. This allows us to define the age of initial collision in the western Himalayan Range, resolving the controversy regarding synchronous vs. diachronous initial collision and shedding new light on along-strike variations in the geodynamic character of continental subduction.

2. Geological background

The 2500 km long IYSZ marks the boundary between the India and Asia plates, and preserves remnants of the Neo-Tethys oceanic lithosphere in the form of ophiolite complexes and associated metamorphosed oceanic sediments (O'Brien et al., 2019). In southeastern Tibet, ITSZ units separate the Gangdese Arc (Asian plate) to the north from the Namche Barwa Complex (Indian plate) to the south (Zhang et al., 2022). This paleo-accretionary prism consists of a wedge-shaped complex comprising metavolcanic rocks, phyllite, slate, marble, greenschist-facies to amphibolite eclogite-facies metasedimentary-matrix mélange, and serpentinite-matrix mélange (e.g., Palin et al., 2015; Li et al., 2023; 2024). Cenozoic UHP and HP rocks are rare throughout the Himalayan orogen (Fig. 1A); most occur in the Western Himalaya, within or just south of the IYSZ at Kaghan, Tso Morari, Stak, Sapi-Shergol, and Zildat (O'Brien et al., 2019 with references therein), although one example has been recently documented from Milin (Li et al., 2023; 2024), close to Lulang in the eastern Himalaya. Previous studies show that deformed mylonite units from the eastern IYSZ

experienced mylonitization and peak amphibolite-facies metamorphism at ~610 ± 30 ℃ at 23.4 ± 0.7 Ma, and exhumation-driven cooling of these units to 420–500 ℃ occurred at ca. 11–8 Ma (Palin et al., 2015). Additionally, the presence of HP-LT (eclogite-facies) rocks is further confirmed by the preservation of phengite + chloritoid assemblage in schists in the Milin area of the eastern IYSZ (Li et al., 2023; 2024). These HP-LT rocks underwent multistage metamorphism, with peak metamorphism recording pressures up to 16–18 kbar and temperatures of 520 ℃, and later amphibolite-facies retrograde metamorphism overprinting at 25–21 Ma (Li et al., 2023). Yet, despite their significance, Li et al. (2023) did not define the age of peak metamorphism, which is critical for comparison with similar units along strike.

Our study focuses on the Lulang area of the easternmost IYSZ at the Eastern Himalayan Syntaxis (Fig. 1B) where the aforementioned metasediments are situated. The Lulang metasedimentary unit is part of a narrow ophiolitic mélange (Fig. 1C; Yin et al., 2003; Palin et al., 2015), and is a relic of a paleo-accretionary prism formed by the northward subduction of Neo-Tethyan oceanic crust. Rock types in this area are mainly dismembered ophiolites and mélange and include coherent metabasite, metapelite and ultramafic and mafic pods. Of those, we focused on metapelitic rocks as they commonly preserve textural evidence of their tectonothermal history within garnet porphyroblasts and reaction microstructures indicating multistage metamorphism. Hence, their mineral assemblages are well suited to the estimation of the *P–T* conditions of multistage metamorphism during the terminal stages of subduction of the Neo-Tethys Ocean associated with India–Asia collision.

3. Petrology and mineral chemical compositions

Three samples of garnet–mica schist (23LL06-2, 23LL06-3, and 23LL08-2) were examined in this work, and all analytical methods and data for mineral analyses are

presented in the Supplemental Material[1]. The mineral abbreviations in this paper follow Whitney and Evans (2010). In hand specimen, all units show a clear schistosity and have a porphyroblastic texture (Figs. 2 and S1). They contain garnet porphyroblasts (25–30 vol. %) and widespread white mica (25–30 vol. %) and quartz in the matrix, with biotite (10–15 vol. %), chlorite (~5 vol. %), and staurolite (~5 vol. %). Accessory minerals include ilmenite, rutile, zircon, apatite, and allanite (Figs. 2 and S1). In addition, samples 23LL06-2 and 23LL06-3 contain minor graphite in their matrix (Fig. 2). Samples 23LL06-2 and 23LL08-2 were used for detailed petrographic study and thermobarometry (phase equilibrium modeling) as they are considered representative of the studied units.

Garnet porphyroblasts in samples 23LL06-2 and 23LL08-2 are up to 12 mm in diameter (Figs. 2 and S1), and are euhedral to semi-euhedral. For samples 23LL06-2 and 23LL06-3, garnet core-mantle domains contain inclusions of rutile (Rt1), plagioclase (Pl1), quartz, muscovite (Ms1), chlorite and biotite (Bt1). (Fig. 2A-C). Rim domains include plagioclase (Pl2), quartz and biotite (Bt2), paragonite and phengite (Ms2), but lack chlorite (Fig. 2D and E). In sample 23LL08-2, garnet core-mantle domains contain ilmenite, plagioclase (Pl1), paragonite (Pa1), quartz, muscovite (Ms1), chlorite (Chl1) and biotite (Bt1). (Fig. 2G). Rim domains lack plagioclase (Pl2), but contain staurolite, phengite (Ms2), quartz and chlorite (Chl2) (Fig. 2G and H). Inclusion suites in garnet therefore place constraints on the metamorphic history of the samples. Early high-pressure and initial decompression periods are primarily evidenced by (1) garnet rims and phengite, and (2) their xenoblastic rims, which penetrate the white mica (Ms3) matrix and/or appear replaced by retrograde biotite (Bt3) + chlorite (Chl3) + quartz + plagioclase (Pl3). These retrograde minerals are fine-grained, forming reaction rims around peak metamorphic

minerals and locally filling fractures (Figs. 2E, F and 2I). Staurolite is scarce and occurs as euhedral patchy crystals 0.1–0.5 mm long (Fig. 2G and H), and white mica is abundant and appears as a 0.2–2 mm flakes (Fig. 2G).

Compositional zoning profiles in garnet for samples (a) 23LL06-2 and (b) 23LL08-2 are similar (Table S1), and are characterized by increasing X_{Alm} and X_{Prp}, with X_{Sps} and X_{Grs} decreasing from the core to the rim (Figs. 3A, B and Fig. S2). In particular, the outermost garnet rim in sample 23LL08-2 shows a slight decrease of X_{Prp} from 0.02 to 0.03 and increase of X_{Grs} from 0.23 to 0.21 (Fig. 3B and Fig. S2). White mica inclusions are both phengite and paragonite, and their X_{Na} and X_{Mg} contents both correlate with Si apfu (Fig. 3C-F). Usually, paragonite in sample 23LL06-2 (a) and 23LL08-2 (b) has high X_{Na} (0.39–1.21[a]; 0.60–0.90[b]) and low Si apfu (2.78–3.05[a]; 2.97–3.07[b]) and X_{Mg} (0.05–0.25[a]; 0.09–0.40[b]), whereas phengite has low X_{Na} (0.05–0.28[a]; 0.10–0.30[b]) and generally higher Si apfu (3.01–3.30[a]; 3.08–3.18[b]) and X_{Mg} (up to 0.68) values (Fig. 3C-F). Chlorite shows a limited range in composition, with octahedral Al between 1.50 and 1.70 apfu, and X_{Mg} = 0.36–0.42 (Table S2).

Based on a detailed petrographic and mineralogical study, three metamorphic assemblages were recognized: a pre-peak stage (M1), garnet core + Ms1 + Pa1 ± Bt1 + Qz + Rt + Chl + Pl1; a peak stage (M2), garnet rim + Ph (Ms2) + Pa2 ± Bt2 + Qz + Rt + Pl2 ± St; and a retrograde stage (M3), Ms3 ± Pa3 + Bt3 + Qz + Ilm + Pl3 ± St.

4. Metamorphic P–T-t determination

4.1 Phase equilibrium modeling

We performed phase equilibrium modeling to constrain the P–T conditions of stages M1 to M3, taking advantage of prograde zoning profiles preserved in garnet

porphyroblasts alongside their inclusion suites. Phase equilibrium modeling was performed for samples 23LL06-2 and 23LL08-2 in the system NCKFMASHT (Na_2O–CaO–K_2O–FeO–MgO–MnO_2–Al_2O_3–SiO_2–H_2O–TiO_2), which is a good representation of most natural metasedimentary rocks (White et al., 2000). Calculations were performed using the GeoPS v3.32 program (Xiang and Connolly, 2022) with the internally consistent thermodynamic dataset ds62 (Holland and Powell, 2011) and solid-solution models for biotite, chloritoid, chlorite, cordierite, garnet, ilmenite, white mica (muscovite–paragonite), silicate melt (White et al., 2014), plagioclase and K-feldspar (Holland et al., 2022). Kyanite, sillimanite, rutile, quartz and water are considered pure phases. Analysed XRF-derived bulk rock compositions are presented in Table S1. The absence of Fe^{3+}-bearing minerals, such as epidote and magnetite, the presence of rutile, measured compositions of Fe−Mg-bearing minerals (garnet + biotite) in the metapelites, and the presence of minor graphite indicate that both samples formed in a reduced setting. Consequently, Fe_2O_3 was omitted as a component in pseudosection construction. For each model, the rock was considered to be water-saturated due to the abundance of hydrous minerals that show equilibrium relationships. We also considered the effects of cation sequestration into growing garnet porphyroblasts by calculating effective bulk compositions for the prograde-to-peak and retrograde stages of the metamorphic cycle. All analytical and computational methods, and a discussion of the reliability of the modeling results and sources of uncertainty and data for whole-rock geochemistry are presented in Supplemental Material[1]. The normalized XRF-based bulk rock compositions of samples 23LL06-2 and 23LL08-2 are presented in Table S2.

Pre-peak stage M1

For sample 23LL06-2, the ***M1*** assemblage Grt + Chl + Ms + Bt + Rt + Qz + H_2O is stable over *P–T* conditions of 470 – 520 ℃ and 0.5 – 0.8 GPa (Fig. 4A). The measured compositions of garnet cores (X_{Prp} = 0.03, X_{Spe} = 0.17, X_{Alm} = 0.54, X_{Grs} = 0.26; see Fig. 4A) together with modeled isopleths for garnet in the calculated pseudosection further constrain the *P–T* conditions of formation during ***M1*** to ~500 ℃ and ~0.7 GPa (Fig. 4A).

For sample 23LL08-2, the inferred ***M1*** mineral assemblage is stable on the calculated pseudosection at *P–T* conditions of 490–550 ℃ and 0.6–0.7 GPa (Fig. 4C). The measured composition of garnet cores (X_{Prp} = 0.08, X_{Alm} = 0.69, X_{Grs} = 0.15 and X_{Spe} = 0.08, Fig. 4C) intersect modelled isopleths of garnet at ~520 ℃ and ~0.65 GPa in the Grt + Chl + Ms + Pa +Pl + Ilm + Qz + H_2O stability field (Fig. 4C), which is consistent with the inclusion assemblage (chlorite, paragonite, ilmenite, muscovite, plagioclase, quartz) in the Grt core (Fig. 2G).

Peak stage M2

For sample 23LL06-2, ***M2*** occupies a field of Grt + Bt + Ms (Ph)+ Pa + Pl + Rt + Qz + H_2O, that extends over a wide *P–T* range of 0.8–1.7 GPa and 550–660 ℃ (Fig. 4A). Isopleths for the modelled composition of X_{Prp} and X_{Grs} are plotted in Fig. S3. Calculated Si isopleths for white mica exhibit slightly positive dP/dT slopes (Fig. 4A), which makes this variable particularly useful for constraining pressure during ***M2***. The observed inclusion assemblage within Grt rims includes paragonite and phengite, plagioclase, quartz and biotite (Fig. 2); isopleths for the modelled composition of phengite and biotite are also plotted in Fig. 4A. The highest Si content (3.23 apfu) of phengite in the matrix, and the Ti, X_{Mg} of biotite (0.095, 0.48), combined with X_{Prp} values of Grt rims (X_{Prp} = 0.09, Fig. 4A) restrict the *P–T* conditions of ***M2*** to 1.4 GPa

and ~590 ℃ (Fig. 4A), representing low-temperature eclogite-facies metamorphism.

For sample 23LL08-2, the ***M2*** mineral assemblage of Grt + Ms (Ph) + Pa + Chl + Rt + St + Qz + H_2O is stable over a narrow *P–T* field of 1.08–1.5 GPa and 570–610 ℃ (Fig. 4C). Isopleths for the modelled composition of X_{Prp} and X_{Grs} are plotted in Fig. S3. The assemblage observed within garnet rims includes chlorite, paragonite, staurolite, phengite and rutile (Fig. 2F). The X_{Grs}, X_{Alm} and X_{Prp} values of garnet rims (X_{Grs} = 0.03, X_{Alm} = 0.78, X_{Prp} = 0.17, Fig. 4C) constrain the *P–T* conditions of ***M2*** to ~1.4–1.5 GPa and 580–595 ℃ (Fig. 4C); these conditions are taken to represent peak pressure and are consistent with low-temperature eclogite-facies metamorphism.

Amphibolite-facies retrograde M3

In both samples 23LL06-2 and 23LL08-2, the key features of the ***M3*** assemblage are low-Si muscovite, garnet partially replaced by retrograde biotite + chlorite + quartz + plagioclase (Fig. 2) and the replacement of rutile by ilmenite. For sample 23LL06-2, the X_{Mg} (0.53) isopleths of chlorite and X_{Mg} (0. 18) isopleths of staurolite constrain the *P–T* conditions of ***M3*** to 0.45 GPa and 560 ℃ (Fig. 4B) at the St + Chl + Bt +Ilm stability field, suggesting later decompression and/or cooling. For sample 23LL08-2, the retrograde assemblage of Bt + Ms + Chl + Ilm + Qz + Pl (Ab) + Pa + H_2O in a narrow field represents ***M3*** with *P–T* conditions of 0.45–0.55 GPa at 540–555 ℃ (Fig. 4D). The ***M3*** assemblage does not include garnet, and garnet is absent from the modelled phase assemblage at these *P–T* conditions (Fig. 4D). The consumption of garnet is supported by the presence of resorbed garnet margins adjacent to pressure shadows (Fig. 2).

In summary, phase equilibria modeling shows that both analyzed Lulang schists reached roughly similar peak-metamorphic conditions of ca. 1.4–1.5 GPa and 590 ℃.

They both record clockwise *P–T* paths, with a prograde increases in *P* and *T* and a late cooling and decompression stage. This HP eclogite-facies peak metamorphism was overprinted by relatively low pressure (LP) retrogression of the lower amphibolite-facies. It is evident, that although these rocks witnessed similar metamorphic conditions and *P–T* paths, they have different mineral assemblages and varying modes during the prograde, peak and retrograde processes because of their differences in the bulk rock chemical compositions and degrees of retrogression.

4.2 Zircon geochronology and trace element compositions

Analytical methods and zircon dating results are presented in the Supplemental Material[1]. Most zircons from garnet–mica schist 23LL08-2 show core–rim structures (Fig. 5A). The inherited detrital cores have variable size, shape, and distinct zoning patterns, whereas the overgrowth rims show weak patchy zoning or two-layer zoning (inner rim zone: Zrn-I; outer rim zone: Zrn-II) (Fig. 5A). Zrn-I and Zrn-II have no clear difference in age and rare earth element (REE) characteristics, with low and variable Th/U ratios (0.0004–0.18, Table S4) and weakly fractionated to flat heavy rare earth element (HREE) patterns (Fig. 5C). A total of 108 spots yielded a $^{206}Pb/^{238}U$ age range from 51.1 ± 4.27 Ma to 36.5 ± 4.97 Ma, with a peak at ca. 47–43 Ma (Fig. 5E). The mean square of weighted deviates (MSWD) values of the weighted mean dates of 23LL08-2 (45.6 ± 0.13 Ma, MSWD = 8.5, n =108) (Fig. 5G) indicate that the analyses do not represent single populations. Except for a few anomalies, the REE patterns for the sample display lack Eu anomalies and have flat heavy REE slopes (Fig. 5C). The REE data indicates that zircons with $^{206}Pb/^{238}U$ ages > 48 Ma are relatively enriched in HREE contents (Fig. 5C) compared to those with $^{206}Pb/^{238}U$ ages < 48 Ma. Furthermore, no significant correlation is observed between ΣHREE or $(Gd/Lu)_N$ ratios and $^{206}Pb/^{238}U$ ages for zircons within the 47–43 Ma age range (Figs. 6A

and C). However, zircons older than 45 Ma show higher and more variable $(Eu/Eu^*)_N$ ratios compared to zircons younger than 40 Ma (Fig. 6E).

Most zircons from garnet–mica schists (23LL06-2 and 23LL06-3) consist of a large core domain and one or several rims, the cores exhibit well-defined sector zoning or concentric oscillatory zoning made up of irregular zones (Fig. S4A and B). Overgrowth rims show patchy zoning and low Th/U ratios (0.003–0.148, Table S3) and REE contents (<565 ppm, Table S3) and weakly fractionated to flat or even depleted HREE patterns (Fig. S5). All the analytical spots yield $^{206}Pb/^{238}U$ ages ranging from 62.65 ± 13.59 Ma to 20.38 ± 0.16 Ma (Fig. S5, Table S4), two clusters 47–44 Ma and ca. 23 Ma are identified in age spectra. Excluding spots in each sample with relatively high uncertainty, high discordance, and/or ages that differ significantly from the main age cluster ages, calculated weighted mean ages are 47.44 ± 0.43 Ma (MSWD = 4.5, n = 15) and 46.7 ± 1.8 Ma (MSWD = 3.0, n = 9) for 23LL06-2 and 23LL06-3, respectively (Fig. S5). Zircon domains with ages of ca. 60–50 Ma, 40–37 Ma and ca. 23 Ma have relatively high concentrations of HREE (2110–4774 ppm) and $(Lu/Gd)_N$ ratios (65.8–185) (Table S3); there was a general trend to lower concentrations of HREEs and $(Lu/Gd)_N$ ratios with decreasing dates (Fig. S5), whereas domains with ages of 40–37 Ma and ca. 23 Ma show the opposite trend: concentrations of HREEs (42–9265 ppm) and $(Lu/Gd)_N$ ratios (1.39–275) markedly increase with decreasing age (Fig. S5).

4.3 Monazite geochronology and trace element compositions

In situ monazite geochronology was performed from several thin sections to link ages to textures, and thus stages along each rock's metamorphic *P*-*T* path. Analytical procedures and detailed results are presented in Supporting Information S1 and Table

S3. Analyzed monazite grains (25-100 μm) occur as prismatic to subrounded inclusions within garnet rims (Fig. 5B and Fig. S6) and display no significant chemical zonation in backscattered electron (BSE) imaging (Fig. 5B and Fig. S6). Forty $^{208}Pb/^{232}Th$ spots yielded consistent $^{208}Pb/^{232}Th$ ages in a narrow range of 44.1–50.3 Ma (Table S3), and produced a weighted mean age of 47.15 ± 0.36 Ma (n = 40; MSWD = 4.4, Fig. 5H). In a Tera-Wasserburg diagram, all spot analyses lie on a discordia line and define a lower intercept age at 48.20 ± 0.43 Ma (MSWD = 7.5; Fig. 5F). The REE data reveal that monazite grains with $^{208}Pb/^{232}Th$ ages > 48 Ma have higher HREE concentrations than younger (<48 Ma) grains; however, no significant systematic correlation was observed between ΣHREE, $(Gd/Lu)_N$ or $(Eu/Eu^*)_N$ and $^{208}Pb/^{232}Th$ ages over the 47–44 Ma interval.

5. Bulk-Rock Nd Isotope Results

Analytical methods and data for Sr-Nd isotope analyses are presented in the Supplemental Material[1]. Measured $(^{87}Sr/^{86}Sr)_i$ ratios ranged from 0.732814 to 0.734423, $^{143}Nd/^{144}Nd$ ratios ranged from 0.511687 to 0.511742, εNd(0) values are from −18.24 to −17.20, and εNd(t) values are from −16.98 to −18.05 (Fig. 8 and Table S3). There are no distinct differences in $^{143}Nd/^{144}Nd$ ratios, $(^{87}Sr/^{86}Sr)_i$ ratios, εNd(0), and εNd(t) values for different samples (Fig. 8 and Table S3), indicating that they share a common tectonic and metamorphic history. The T_{DM} values are 2,424–2,026 Ma (Fig. 6 and Table S3).

6. Discussion

6.1 Geochronology interpretation and P–T–t evolution

In most metamorphic rocks, garnet is an important repository of Lu/Gd and HREEs, while plagioclase is enriched in Eu. Thus, at equilibrium, the Lu/Gd, HREE concentrations and Eu/Eu* of zircon/monazite should be negatively correlated with the modal proportion of garnet and plagioclase in a metamorphic rock in which they grow (Rubatto et al., 2006; Yakymchuk et al., 2017). On chondrite-normalized rare earth element (REE) diagrams (Fig. 6), relatively shallow slopes for HREEs and lower $(Lu/Gd)_N$ ratios reflect a larger proportion of garnet in the system, whereas steeper slopes and higher ratios reflect lower proportions. Conversely, lower $(Eu/Eu^*)_N$ ratios reflect a larger proportion of plagioclase in the system, and vice versa for higher ratios. Our petrological modeling and microstructural analysis indicates that garnet porphyroblasts grow during prograde metamorphism and are partially replaced during retrograde metamorphism, therefore reaching their maximum volume proportion at the metamorphic peak (Fig. S3). By contrast, plagioclase is another highly pressure-sensitive metamorphic mineral that breaks down during burial, but crystallizes again during exhumation and retrograde metamorphism. With this framework in mind, changes in HREEs, $(Lu/Gd)_N$ and $(Eu/Eu^*)_N$ values in accessory minerals in both studied schists can be combined with our new age data (Fig. 6A–D) to constrain the timing of peak eclogite-facies metamorphism in this part of the Himalayan orogen.

The zircon and monazite ages collected from each studied samples show two distinct peaks at ca. 47 Ma and 23 Ma (Fig. 8A and B). Spot analyses for both zircon and monazite that are older than 48 Ma have relatively high HREE contents (Fig. 5C, D and S5), indicating that garnet nucleated and began to grow along the prograde burial path (M1) at this time (Fig. 7C). Zircon and monazite that formed during the ca.

47 Ma peak (or slightly later in the 47–43 Ma range) shows no significant difference in REE contents (Fig. 7C), likely indicating protracted accessory mineral growth or recrystallization at high-pressure eclogite-facies conditions (M2 stage). We interpret this due to (1) the shallower HREE profiles compared to >48 Ma grains, and lower $(Lu/Gd)_N$ ratio and higher $(Eu/Eu^*)_N$ consistent with (re)crystallization in the presence of garnet and absence of plagioclase (Fig. 5, 6 and S5), (2) monazite inclusions within garnet rims that are spatially associated with M2 (peak-pressure) assemblages, recording coeval growth with the host garnet (Fig. S6), and (3) Raman analyses of zircon rims that showed garnet + rutile + phengite inclusions (Fig. S7), providing a maximum age constraint on peak eclogite-facies conditions. We therefore interpret the upper limit of the date peak (i.e., ca. 47 Ma) is the oldest age of eclogite-facies metamorphism (ca. 1.5 GPa at 590 ℃) in these rocks at Lulang. Scattering of monazite/zircon U-Th-Pb ages is a common feature of Himalayan rocks (e.g., Warren et al., 2011; Tobgay et al., 2012), which often reflects the timeframe over which monazite/zircon grew (e.g., Simpson et al., 2000; Corrie and Kohn, 2008; Janots et al., 2008). Thus, as the chemical composition of 47–43 Ma of monazite and zircon in samples was uniform, this suggests that that scatter in their ages is most likely due to protracted accessory mineral growth over several million years rather than post-crystallization isotopic disturbance.

A small number of age data within the 37–40 Ma range are interpreted to record zircon recrystallization during retrograde metamorphism. This is indicated by systematically elevated $(Lu/Gd)_N$ and $(Eu/Eu^*)_N$ ratios (Fig. 6 and Fig. S5), which are diagnostic of recrystallization processes occurring under plagioclase-stable and garnet-absent conditions, and consistent with petrographic evidence documenting garnet breakdown and feldspar crystallization during the M3 retrograde metamorphic

stage (Fig. 7C). In addition, zircon age spectra from both mica schists exhibit a distinct 23 Ma peak in both kernel density estimation and probability density plots (Fig. 7A and B), potentially recording thermal overprinting associated with late-stage exhumation. This thermal pulse correlates with a 25–21 Ma regional tectonothermal event documented across the eastern Himalayan syntaxis, including amphibolite-facies overprinting of granulite terranes at Namche Barwa (ca. 24 Ma; Zhang et al., 2010, 2012), initiation of IYSZ ductile shearing (ca. 23–25.5 Ma; Xu et al., 2012; Palin et al., 2015), and amphibolite-facies retrogression of HP-LT eclogites in the Milin complex (Li et al., 2023, 2024). These coeval thermal events record a regional pulse of crustal extrusion and thermal resetting during the late Oligocene, but which fortunately did not erase geochronological or petrographic evidence of earlier eclogite-facies metamorphism.

To summarize, our new zircon data provide the first robust age for high-pressure metamorphism in the Lulang area related to the earliest stages of India-Asia collision and Himalayan orogenesis. When coupled with the results of phase equilibrium modeling, we provide a high-resolution *P–T–t* path for schists in the Lulang area that preserve the prograde, peak, and retrograde portions of a single metamorphic cycle (Fig. 7C). Prograde metamorphism (M_1) at >48 Ma records burial, whereas peak metamorphism (M_2) at 47–43 Ma includes both the pressure peak and the thermal peak, identifying the timing of maximum burial during subduction. Subsequent retrograde metamorphism is chronologically constrained by two distinct zircon age populations: 40-37 Ma dates correspond to M_2-M_3 stage decompression recording initial exhumation, while ca. 23 Ma ages document final-stage exhumation following M_3 retrogression.

6.2. An Indian provenance

The Lulang schists display uniform Nd–Sr isotopic compositions (εNd(0) = −17.20 to −18.24; εNd(t) = −17.04 to −18.05; $^{87}Sr/^{86}Sr(i)$ = 0.7328–0.7344; Fig. 8A and B) that overlap with εNd(t) (−10 to −25) and $^{87}Sr/^{86}Sr$ (0.71–0.90) ranges for Tethyan and Greater Himalayan rocks (Fig. 8A). These values contrast sharply with similar data for rocks belonging to the the South Lhasa terrane (εNd(t) = +7 to −7; Fig. 8A and B) and accretionary prism metasediments (εNd(0) = −10.8 to −5.9; Wei et al., 2020; Fig. 8B). Furthermore, all the samples exhibit REE patterns characterized by significant LREE enrichment and flat HREEs distributions, coupled with negative Eu anomalies and absence of Ce anomalies, which are comparable to those of upper continental crust rocks (Fig. 8C). Such features sharply contrast with signatures of subduction zone sediments, which commonly exhibit Ce depletion and MORB-like REE patterns (Rudnick and Gao, 2003). These geochemical features demonstrate that the studied schists from Lulang were derived from continental crust-like sources rather than subduction zone-related sediments. This isotopic evidence, combined with the REE signatures, precludes derivation from subduction-related or accretionary systems, instead favoring a cratonic origin from the Indian continental margin.

6.3 Age of India-Asia collision

Constraining the timing of the initial collision between India and Asia is a key parameter for understanding its geodynamic and tectonic evolution (e.g., DeCelles et al., 2014; Wu et al., 2014; Hu et al., 2016). Proposed ages in the literature range from 60 to 35 Ma (e.g. Yin and Harrison, 2000; Aitchison et al., 2007; Gibbons et al., 2015; Wei et al., 2020) with most workers favoring the period 55–50 Ma (e.g. Najman et al., 2010; Donaldson et al., 2013; Ding et al., 2016; Zhao et al., 2024). It has further been

suggested that initial collision may have been diachronous along the length of the Himalayan Range, having occurred as early as 60–65 Ma in the central Himalaya, but at ~50 Ma in western and east-central Himalaya (e.g. DeCelles et al., 2014; Wu et al., 2014; Ding et al., 2016, 2017; Hu et al., 2016).

Plate collision should immediately precede deep subduction of telescoped portions of continental crust that remain attached to the remnants of descending oceanic lithosphere. As such, (U)HP metamorphism of Indian continental margin rocks serves as a robust proxy for initial collision. Early studies of UHP eclogite in the western Himalaya demonstrated crustal subduction to sub-arc depths of 80–160 km at ca. 53–51 Ma in the Tso Morari dome (Fig. 8A and B; de Sigoyer et al., 2000; Wilke et al., 2010) and at ca. 46 Ma in the Kaghan Valley (Fig. 8A and B; Kaneko et al., 2003). The apparent ca. 6 Myr age difference in the time of UHP metamorphism was initially taken to indicate diachronous subduction of a jagged Indian margin (Guillot et al., 2008). However, Donaldson et al. (2013) subsequently reported in-situ U/Th–Pb and REE analyses of zircons in two samples of Tso Morari eclogite, with ages ranging from 53 Ma to 37 Ma and peaking at c. 47–43 Ma. Flat heavy REE slopes and the absence of Eu anomalies were noted as being compatible with eclogite-facies zircon crystallization. Their results were consistent with those from Kaghan Valley, arguing instead that a single, protracted UHP metamorphic event may have occurred contemporaneously across the western Himalaya (Donaldson et al., 2013). Additionally, the Barrovian metamorphism of the Yardoi gneiss domes between 45 and 52 Ma, with a peak age of c. 47 Ma (Fig. 8B; Ding et al., 2016), may be explained by the initiation of northwest to east-central India–Asia collision synchronous along the margin at ca. 50 Ma (Ding et al., 2016). So far, no true eclogite has been found in the IYSZ rocks located near to the eastern Himalayan syntaxis, instead, early eclogite-

facies metamorphism of basic granulites may have occurred around 65~40 Ma and suggests that continental collision between India and Asia at the eastern Himalayan syntaxis initiated after 65 Ma (Fig. 8B; Ding et al., 2001).

The present geological boundary between India and Asia is marked by the IYSZ, which contains remnants of the ancient Neo-Tethys oceanic lithosphere and metasedimentary rocks derived from terrigenous sediments deposited on the Indian continental substrate (O'Brien et al., 2019). The bulk rock Sr–Nd isotope and REE data presented here show that the metasedimentary rock in the easternmost IYSZ formed from sediments derived from the Indian plate (Fig. 8). These terrigenous sediments were subducted into the mantle and subjected to HP eclogite-facies metamorphism during collisional orogeny in the Early Cenozoic, some which are presently exposed in the eastern IYSZ. Therefore, the HP metamorphic ages of ~47 Ma obtained here place a new, tight constraint on the timing of subduction of the Indian continent in the easternmost Himalaya. If (U)HP metamorphism occurs along the leading edge of a continental margin, it will postdate initial continent subduction by ~1–4 Myr, depending on convergence velocity and subducting slab dip (Kaneko et al., 2003; Leech et al., 2005). Therefore, the HP-LT metamorphic record in the easternmost Himalaya indicates that the initial Indian subduction occurred at 52–48 Ma (50 ± 2 Ma), consistent with the western and east-central Himalaya (Fig. 8A, Kaneko et al., 2003; Parrish et al., 2006; Donaldson et al., 2013; St-Onge et al. 2013; Ding et al., 2016). The 50 ± 2 Ma collisional time is also supported by a deformation pulse across the IYSZ dated at c. 51 Ma, which coincides with the first major episode of shortening in the upper plate after the initial India-Asia contact (Zhao et al., 2024).

Geochronological estimates for the age of initial India-Asia collision along the western, east-central, and easternmost collisional zones are thus 53–51, 50, and 50 ± 2

Ma, respectively (Fig. 9A and B). Assuming a similar velocity of convergence along the strike of the Himalayan belt, their consistent ages suggest the geometry of the collisional interface between the Indian and Asian continents in the west, east-central and east were either regularly linear or in line with the continental margin of Asia. Either way, this implies that the onset of the collision was synchronous along orogenic strike (Fig. 8B), at least as far as can be determined from the metamorphic rock record.

6.4 Differential subduction of the Indian continent

The contrasting metamorphic records along the Himalayan orogen reveal first-order variations in the subduction style and behaviour of the Indian plate. The presence of UHP metamorphic rocks in the western Himalaya indicates that the northwestern margin of the Indian continent was subducted to at least 100 km into the mantle (e.g. Kaneko et al., 2003; Guillot et al., 2008; O'Brien et al., 2019), placing minimum constraints on angle of dip and the rate of descent of the subducting Tethyan oceanic slab (Leech et al., 2005). In contrast, in the east-central Himalaya, the upper crust of the Indian continent experienced Early Eocene MP amphibolite-facies metamorphism, which is thought to be the result of low-angle or nearly flat subduction of the Indian continent along this sector (Ding et al., 2016). Tomographic and seismic data suggests that the Indian plate currently dips at a very shallow angle (~9°) beneath southern Tibet, reaching a maximum crustal depth of 80 km in the central Himalayan region (Guillot et al., 2008) before detaching and sinking into the mantle. As such, the dip angle of the Indian plate beneath southern Tibet has shallowed significantly since initial collision, prohibiting the formation of post-collisional UHP rocks.

Importantly, our work has revealed for the first time that the leading edge of the Indian continent in the easternmost Himalaya also experienced Early Eocene eclogite-facies metamorphism, as documented in the western Himalaya, albeit with subduction reaching notably shallower depths (~50 km). This discovery establishes a systematic along-strike transition in Paleogene subduction dynamics during closure of the Neo-Tethys Ocean. Robust *P–T–t* information obtained from rocks in both the western and eastern Himalayan syntaxes indicate that peak eclogite-facies metamorphism was reached at similar times in both localities, but at different depths within the mantle. This can be most simply explained by a strong shallowing of slab dip angle from west to east, although the geodynamic reasons for such along-strike variation in subduction angle remain unclear and offer scope for fruitful future investigation.

~~7. Concluding remarks~~

~~Metapelite exposed at the easternmost IYSZ, witnessed HP low-temperature eclogite-facies under P–T conditions of ca. 1.5 GPa at ca. 590 ℃. This indicates that subducted to a depth of about 50 km and then exhumation to lower amphibolite-facies conditions. Our new, U–Pb ages of zircons firstly provide dates for the timing of HP-LT metamorphism in the easternmost Himalaya (ca. 47 Ma) and hence for the initial India–Asia contact (50 ± 2 Ma) in the eastern Himalaya, and is consistent with the timing of eclogite-facies metamorphism in the western Himalaya, and east-central Himalaya. This demonstrates that the initial India–Asia collision occurred quasi-synchronously along orogenic strike, although the maximum depth of subduction of continental material differed. The observed depth discrepancies are most likely attributable to differences in subduction angles. These findings demonstrate marked~

~~spatial heterogeneity in continental crustal subduction behavior within collisional orogens.~~

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FIGURE CAPTIONS

Fig. 1. Geological sketch map of the Himalayan orogenic system.

(A) Generalized geologic map of the Himalayan orogen (modified after Guillot et al., 2008) showing the locations and metamorphic ages of UHP/HP/MP rocks associated with the collision in the Himalaya. Data sources: Shangla, Kaghan, Stak, Sapi-Shergol, Zildat, Tso Morari, Lopu, Yardoi (Ding et al., 2016; O'Brien et al., 2019 with references therein), Milin (Li et al., 2023; 2014), Paxiang (Ding et al., 2001), Lulang (This study). (B-C) Geological map and Cross-section of the Eastern Himalayan Syntaxis (after Palin et al., 2015).

Fig. 2. Representative microstructural features of samples (plane-polarized light, TIMA and BSE).

(A-B) The core of Garnet porphyroblasts includes abundant Qz, Ms1, Bt1, Chl1, Pl1 and Ilm inclusions. (C-D) Small Qz, Ms2, Bt2, Pa, Pl2 and Rt inclusions within Grt rim. (E-F) Later Bt3, Pl3, Ms3, St and Chl developed around the garnet rim and in the matrix. (G) Thin-section scans of sample 23LL08-2 with mineral identification performed by TIMA, TIMA image of core, rim and outer matrix of garnet porphyroblast showing mineral assemblages formed during the prograde, peak, and retrograde stages. (H-I) Inclusion mineral grains of Qz, Ph, Chl, Pa and St within the garnet rim, ilmenite locally replacing rutile in the matrix. The mineral abbreviations are after Whitney and Evans (2010).

Fig. 3. Garnet profiles and white mica chemistry from the studied samples.

(A) Compositional end member profile AA’ of garnet porphyroblast in Fig. S1A. (B) Compositional end member profile BB’ of garnet porphyroblast in Fig. S2A. (C, D) Si

apfu–X_{Na} plots and (E, F) X_{Mg}–Si apfu plots for K-Na-white mica in samples. Dashed lines highlight the best linear fit for the data points.

Fig. 4. **Pseudosection calculations of sample 23LL06-2 and 23LL08-2.**

(A) The pre-peak stage (M1) and peak stage (M2) modeling of measured bulk-rock composition and saturated H_2O content for sample 23LL06-2, M1 is defined by the intersection of X_{Alm}, X_{Grs}, X_{Spe} and X_{Prp} isopleths in the Grt core; M2 is defined by the intersection of X_{Alm}, X_{Grs} and X_{Prp} isopleths in Grt rim and the X_{Mg} isopleths of biotite and highest Si isopleths in phengite. (B) The amphibolite-facies stage (M_3) modeling of effective composition by excluding all the garnet (25 vol.%) and saturated H_2O content for sample 23LL06-2, M3 is defined by the intersection of X_{Mg} isopleths of chlorite and staurolite and assemblage of Ms + Chl + Rt + Bt + Ilm + St stability field, respectively. (C) The pre-peak stage (M1) and peak stage (M2) modeling of measured bulk-rock composition and saturated H_2O content for sample 23LL08-2, M1 is defined by the intersection of X_{Alm}, X_{Grs}, X_{Spe} and X_{Prp} isopleths in the Grt core; M2 is defined by the intersection of X_{Alm}, X_{Grs} and X_{Prp} isopleths in Grt rim and the X_{Mg} isopleths and Ti content of biotite. (D) The amphibolite-facies stage (M_3) modeling of effective composition by excluding all the garnet (25 vol.%) and saturated H_2O content for sample 23LL08-2, M3 is defined by the assemblage of Ms + Chl + Rt + Bt + Ilm + St stability field, respectively. BC: bulk rock composition; EBC: Effective bulk rock composition.

Fig. 5. **Age and composition of the metamorphic zircons and monazites.**

(A) CL images for zircons, (B) BSE images for monazites, (C-D) REE patterns for zircons and monazites, respectively, (E) Zircon U-Pb concordia diagrams, error

ellipses represent ±2σ, (F) Monazite U-Pb Tera-Wasserburg diagram, error ellipses represent ±1σ, (G) Weighted average U–Pb age for zircons, (H) Weighted average Th–Pb age for monazites.

Fig. 6. REE results for metamorphic zircons and monazites.

(A, C, E) ΣHREE, $(Lu/Gd))_N$ and $(Eu/Eu^*)_N$ ratios for zircons, all error bars represent the analytical error (2σ). (B, D, F) ΣHREE, $(Lu/Gd))_N$ and $(Eu/Eu^*)_N$ ratios for monazites, all error bars represent the analytical error (1σ).

Fig. 7. Zircon ages of the HP garnet–mica schists and reconstructed *P–T–t* path.

(A, B) Density and probability distribution and histogram of zircon and monazite ages; (C) *P–T–t* paths of the sample in this study and other representative schists from the easternmost Himalaya. Additional *P–T* paths for HP-MP rocks from Milin, eastern Himalaya, are also shown (Li et al., 2023; 2024, Palin et al., 2015).

Fig. 8. Diagram of whole-rock Sr-Nd isotopes and REE of samples

(A) Sr-Nd isotopic compositions for the metasedimentary rocks from Lesser, Higher, and Tethyan Himalaya, and the Jurassic rock in Southern Lhasa, the early Paleozoic ophiolite from IYSZ. The isotope data are cited from the compilation by Cao et al. (2022), Tian et al. (2017), Zeng et al. (2012), Richards et al. (2005; 2006), Ahmad et al. (2000), Imayama et al. (2008; 2014), Argles et al. (2003), Rashid et al. (2018), Hou et al. (2012), Ji et al. (2022) and Zeng et al. (2015). (B) εNd(0)-T_{DM} plot of samples from the Himalaya and southern Tibet Data composed sources from (Wei et al., 2020 with references therein). (C) Chondrite-normalized rare earth element patterns for study samples. The upper continental crust value is from Rudnick and Gao (2003).

Fig. 9. Tectonic map of the India–Asia collision zone.

(A) Cartoon showing the (quasi-) synchronous collision of India-Asia continents in the western and eastern Himalaya, we apply the common term Greater India to refer to the part of the Indian plate that has been subducted underneath Tibet since the onset of Cenozoic continental collision. (B) Time chart for Tso Morari, Kaghan, Yardoi and Namche Barwa, comprising data from multiple sources, including this study. (C) Approximately 47 Ma: a possible geodynamical models explain variations in ultrahigh-pressure and high-pressure metamorphism across the Himalayas (not to scale).