Tectonic evolution and deep mantle structure of the eastern Tethys since the latest Jurassic

Authors

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Abstract

The breakup of Pangea in the Jurassic saw the opening of major ocean basins at the expense of older Tethyan and Pacific oceanic plates. Although the Tethyan seafloor spreading history has been lost to subduction, proxy indicators from multiple generations of Tethyan ribbon terranes and the active margin geological histories of volcanism and ophiolite obduction events can be used to reconstruct these ancient oceanic plates. The resulting plate reconstructions reconcile observations from ocean basins and the onshore geological record to provide a regional synthesis, embedded in a global plate motion model, of the India-Eurasia convergence history, the accretionary growth of Southeast Asia and the Tethyan-Pacific tectonic link through the New Guinea margin.

The global plate motion model captures the time-dependent evolution of plates and their tectonic boundaries since 160 Ma, which are assimilated as surface boundary conditions for numerical experiments of mantle convection. We evaluate subducted slab locations and geometries predicted by forward mantle flow models against P- and S-wave seismic tomography. This approach harnesses modern plate reconstruction techniques, mantle convection models with imposed one-sided subduction, and constraints from the surface geology to address a number of unresolved Tethyan geodynamic controversies.

Our synthesis reveals that north-dipping subduction beneath Eurasia in the latest Jurassic consumed the Meso-Tethys, and suggests that northward slab pull opened the younger Neo-Tethyan ocean basin from ~155 Ma. We model the rifting of 'Argoland', representing the East Java and West Sulawesi continental fragments, which were transferred northward in latest Jurassic times from the northwest Australian shelf – likely colliding first with parts of the Woyla intra-oceanic arc in the mid-Cretaceous, and accreting to the Borneo (Sundaland) core by ~80 Ma. The Neo-Tethyan ridge was likely consumed along an intra-oceanic subduction zone south of Eurasia from ~105 Ma, leading to a major change in the motion of the Indian Plate by ~100 Ma, as observed in the Wharton Basin fracture zone bends.

We investigate the geodynamic consequences of long-lived intra-oceanic subduction within the Neo-Tethys, requiring a two-stage India-Eurasia collision involving first contact between Greater India and the Kohistan-Ladakh Arc sometime between ~60 and 50 Ma, followed by continent-continent collision from ~47 Ma. Our models suggest the Sunda slab kink beneath northwest Sumatra in the mantle transition zone results from the rotation and extrusion of Indochina from ~30 Ma. Our results are also the first to reproduce the enigmatic Proto South China Sea slab beneath northern Borneo, as well as the Tethyan/Woyla slab that is predicted at mid-mantle depths south of Sumatra. Further east, our revised reconstructions of the New Guinea margin, notably the evolution of the Sepik composite terrane and the Maramuni subduction zone, produce a better match with seismic tomography than previous reconstructions, and account for a slab at ~30°S beneath Lake Eyre that has been overridden by the northward advancing Australian continent. Our plate reconstructions provide a framework to study changing patterns of oceanic circulation, long-term sea level driven by changes in ocean basin volume, as well as major biogeographic dispersal pathways that have resulted from Gondwana fragmentation and accretion of Tethyan terranes to south- and southeast-Eurasia.

Keywords:

Tethys, Pangea, tectonics, geodynamics, Sundaland, Southeast Asia

1 Introduction

Southern Eurasia, Southeast Asia and New Guinea represent a unique example of long-term tectonic convergence between multiple tectonic domains that has resulted in a complex assemblage of continental fragments, intra-oceanic arcs, ophiolite belts and marginal basins (Figs. 1 and 2). The Southeast Asian continental promontory, known as Sundaland, has grown through successive accretionary episodes resulting from the breakup of Pangea (Acharyya, 1998; Audley-Charles, 1988; Metcalfe, 1994), and subsequent northward transfer of Gondwana-derived continental ribbon terranes and microcontinents on the Tethyan oceanic "conveyors" towards Eurasia. Importantly, the region records a complex interaction between the Tethyan and (proto-) Pacific tectonic domains, which has opened and consumed successive oceanic basins and gateways (Metcalfe, 1999), and has had major consequences for biogeographic dispersal pathways such as the origin of the Wallace Line (Burrett et al., 1991; de Bruyn et al., 2014; Lohman et al., 2011), oceanic circulation (Gaina and Müller, 2007; Gourlan et al., 2008; Heine et al., 2004), global climate and sea level (Morley, 2012b; Wang, 2004; Xu et al., 2012), and the development of economic resources (Goldfarb et al., 2014; Zaw et al., 2014).

Plate tectonic reconstructions play a pivotal role in unravelling the complexity of this region and provide a platform to address long-standing geological questions in a geodynamic context. We apply a modern approach of modelling entire plates, their evolving plate boundaries and the terranes they carry. This study aims to synthesise previously published onshore and offshore geological constraints, as well as incorporate decades of developments in plate tectonic reconstructions, into a modern plate motion model to document the post-Pangea geodynamic evolution of southern Eurasia, Southeast Asia and New Guinea since the Late Jurassic in a regional and global context. Despite significant technological and methodological advancements in plate reconstruction approaches, very few reconstructions of the eastern Tethys exist in an open-access digital form that can be tested and expanded by the scientific community.

As part of this work, we release detailed plate reconstructions for the eastern Tethys (from the India-Eurasia collision zone eastward to Papua New Guinea, Figs. 1-3) that are embedded in a self-consistent global plate motion model, as a collection of digital geometry files and rotation parameters compatible with the open-source and cross-platform plate reconstruction tool, GPlates (www.gplates.org). We provide a brief background to previous regional tectonic reconstructions in Section 1.1, as well as tomographic and numerical modelling approaches in Sections 1.2-1.3 that have been used to gain insight into the tectonic and geodynamic processes controlling the regional evolution. In Sections 2 to 4, we outline our approach of building modern plate reconstructions for the three key regions that comprise the eastern Tethys, including i) the India-Eurasia convergence zone, ii) Southeast Asia, and iii) the New Guinea margin, and compare our approach and findings with previous work. In Section 5, we show how modern plate reconstructions that incorporate evolving plate boundaries can be used with numerical models of mantle flow to predict mantle structure, study the distribution of ancient slabs, and test alternative plate motion scenarios where geological constraints are vague or interpretation is ambiguous. In Sections 6 and 7 we highlight the implications of our work in a regional and global context, and provide some key findings from our modelling of the tectonic and geodynamic evolution of the entire eastern Tethyan domain.

The coupled global plate reconstructions and mantle flow models provide a context for better understanding the latest Jurassic rifting events from northern Gondwana (Metcalfe, 1994; Pigram and Panggabean, 1984), which opened the Neo-Tethys at the expense of the Meso-Tethys ocean basin (Fig. 4a). This rifting episode transferred the 'Argoland' ribbon continent, which included East Java, West Sulawesi and Mangkalihat (Hall, 2012; Zahirovic et al., 2014), north towards Eurasia, while also marking the onset of major intra-oceanic subduction systems along southern Eurasia and northern New Guinea. In the absence of preserved seafloor spreading histories for Neo-Tethyan evolution, we test alternative scenarios of subduction using geodynamic models of mantle flow that are compared with the present-day mantle structure interpreted from seismic tomography.

The improvement in the methods applied to plate reconstructions and increasing levels of detail in complex regions have implications for linking plate tectonic evolution with the deep mantle and other Earth systems, and have been extensively used to better understand biogeographic dispersal and evolutionary pathways (Monod and Prendini, 2015; Rolland et al., 2015), long-term climate and sea level change (Herold et al., 2014; Huber and Goldner, 2012; Lee et al., 2013; Müller et al., 2008; Scotese et al., 1999; Spasojevic and Gurnis, 2012; van der Meer et al., 2014), and paleo-bathymetry and oceanic gateway evolution (Gaina and Müller, 2007). Improved plate tectonic reconstruction techniques have enabled the quantification of time-dependent convergence rates (Lee and Lawver, 1995; Sdrolias and Müller, 2006; Whittaker et al., 2007), and inferences on regional and global plate re-organization events (Matthews et al., 2011; Matthews et al., 2012), as well as providing insight into the size distribution of tectonic plates (Morra et al., 2013) and factors controlling the speed of tectonic plates (Zahirovic et al., 2015). The plate reconstructions presented in this work have important implications for our understanding of the mid-Cretaceous seafloor spreading pulse (Seton et al., 2009) that may have led to higher eustatic sea levels (Müller et al., 2008), the proposed major regional and global plate reorganization at ~105-100 Ma (Matthews et al., 2012) that may be linked to the subduction of the Neo-Tethyan mid oceanic ridge. In addition, plate reconstructions of the Tethyan domain have consequences for understanding the atmospheric carbon budget resulting from the initiation and abandonment of major Andean-style and intra-oceanic Tethyan subduction zones (Jagoutz et al., 2016; van der Meer et al., 2014).

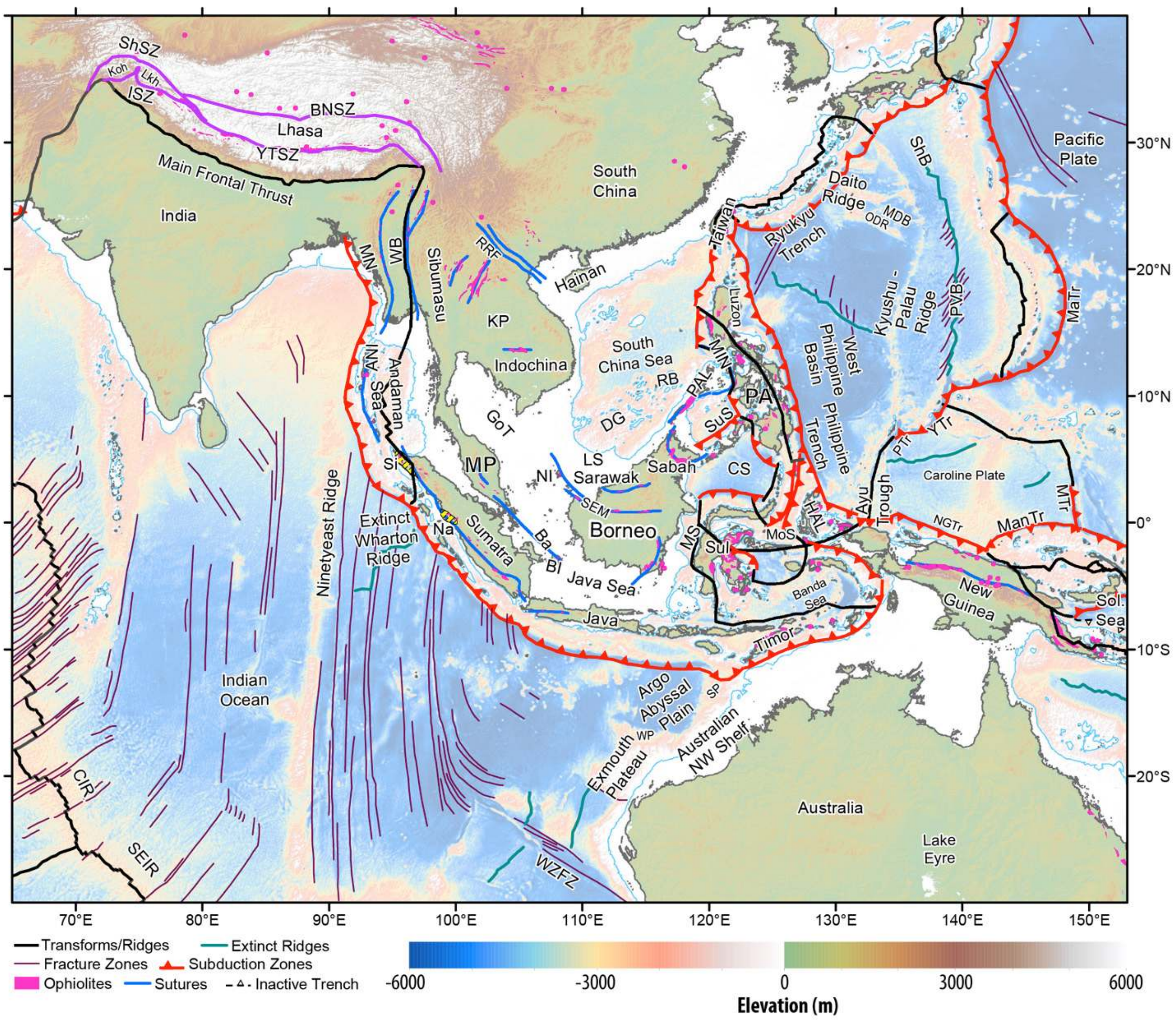

Fig. 1. Regional tectonic setting of southern Eurasia, Southeast Asia and New Guinea. The plate boundaries are modified from Bird (2003), the topography is from Amante et al. (2009), and the seafloor fabric is from Matthews et al. (2011). Southeast Asian sutures (blue) and ophiolites are modified from Hutchison (1975), with additional ophiolites for New Guinea from Baldwin et al. (2012), and for Southeast Asia from Pubellier et al. (2004). The Tethyan sutures in the Indian segment of the margin (violet lines) are from Yin and Harrison (2000). ANI – Andaman-Nicobar Islands, Ba – Bangka Island, BI – Billiton Island, BNSZ – Bangong-Nujiang Suture Zone, CIR – Central Indian Ridge, CS – Celebes Sea, DG – Dangerous Grounds, GoT – Gulf of Thailand, HAL – Halmahera, ISZ – Indus Suture Zone, Koh-Lkh – Kohistan-Ladakh, KP – Khorat Plateau, LS – Luconia Shoals, ManTr – Manus Trench, MaTr – Izu-Bonin-Mariana Trench, MDB – Minami Daito Basin, MIN – Mindoro, MN – Mawgyi Nappe, MoS – Molucca Sea, MP – Malay Peninsula,

MS – Makassar Straits, MTr – Mussau Trench, Na – Natal, NGTr – New Guinea Trench, NI – Natuna Island, ODR – Oki Daito Ridge, PA – Philippine Arc, PAL – Palawan, PTr – Palau Trench, PVB – Parece Vela Basin, RB – Reed Bank, RRF – Red River Fault, SEIR – Southeast Indian Ridge, ShB – Shikoku Basin, ShSZ – Shyok Suture Zone, Si – Sikuleh, SP – Scott Plateau, Sul – Sulawesi, SuS – Sulu Sea, WB – West Burma, WP – Wombat Plateau, WZFZ – Wallaby Zenith Fracture Zone, YTr – Yap Trench, YTSZ – Yarlung-Tsangpo Suture Zone.

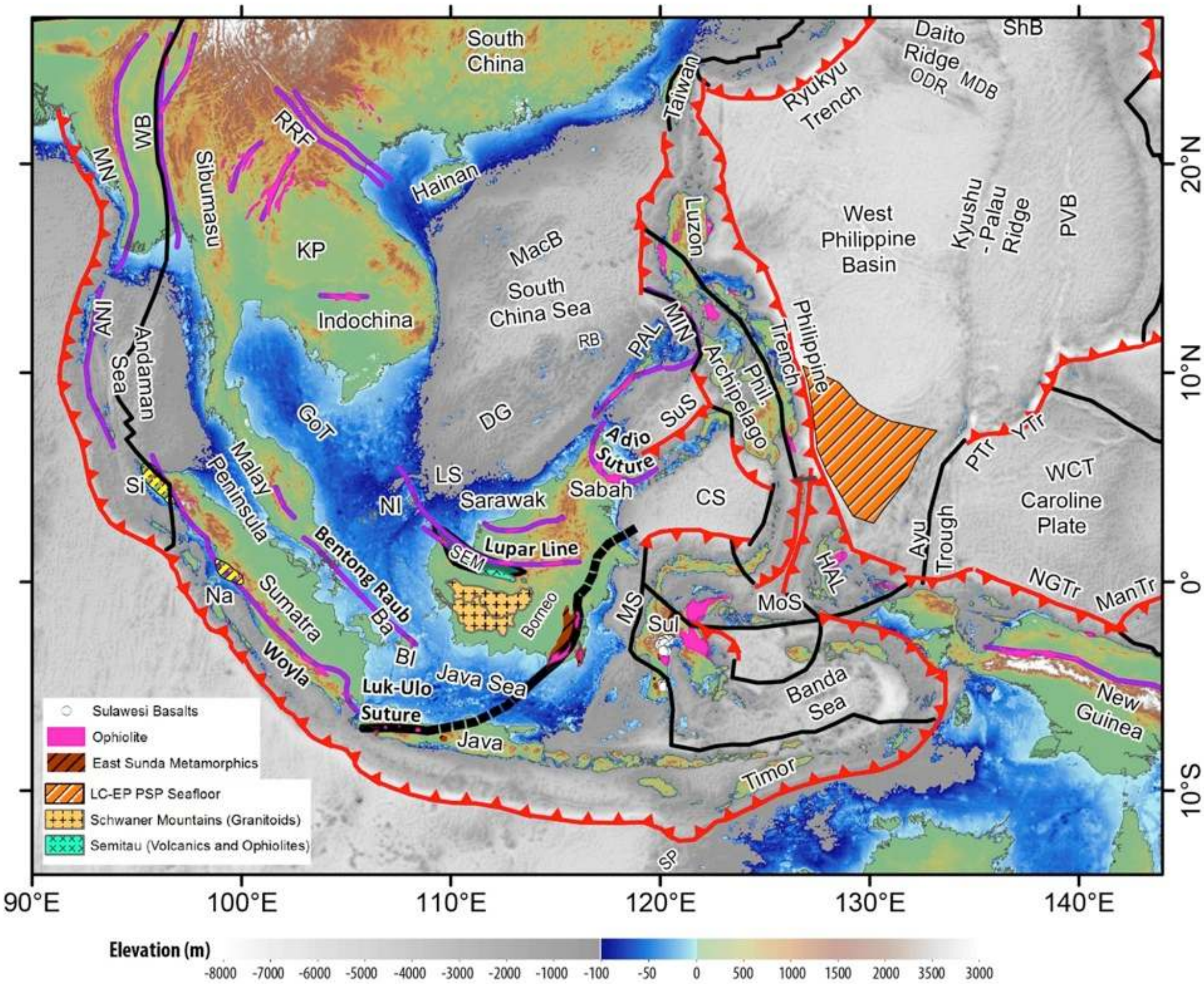

Fig. 2. Regional tectonic framework of Southeast Asia and New Guinea, with high-resolution Global Multi-Resolution Topography of depths shallower than 100 m from Ryan et al. (2009). The Cretaceous Luk Ulo-Meratus sutures are depicted as the thick black line through Java and Borneo. Abbreviations follow those used in Fig. 1. LC-EP PSP – Late Cretaceous(?)–early Paleogene Philippine Sea Plate seafloor crust, WCT – West Caroline Trough.

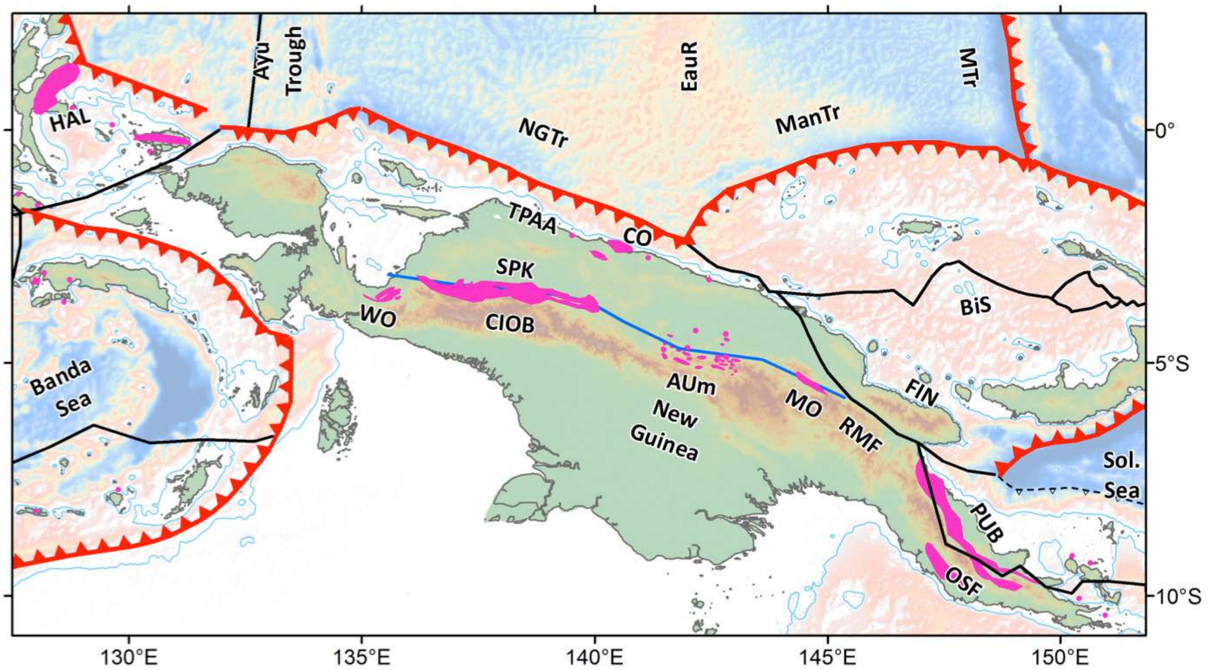

Fig. 3. Tectonic framework and topography of New Guinea. AUm – April Ultramafics, BiS – Bismarck Sea, CIOB – Central Irian Ophiolite Belt, CO – Cyclops Ophiolite, EauR – Eauripik Rise, FIN – Finisterre Terrane, MO – Marum Ophiolite, OSF – Owen Stanley Fault, PUB – Papuan Ultramafic Belt, SPK – Sepik Terrane, Sol. Sea – Solomon Sea, TPAA – Torricelli-Prince Alexander Arc, WO – Weyland Overthrust. Other abbreviations follow Fig. 1.

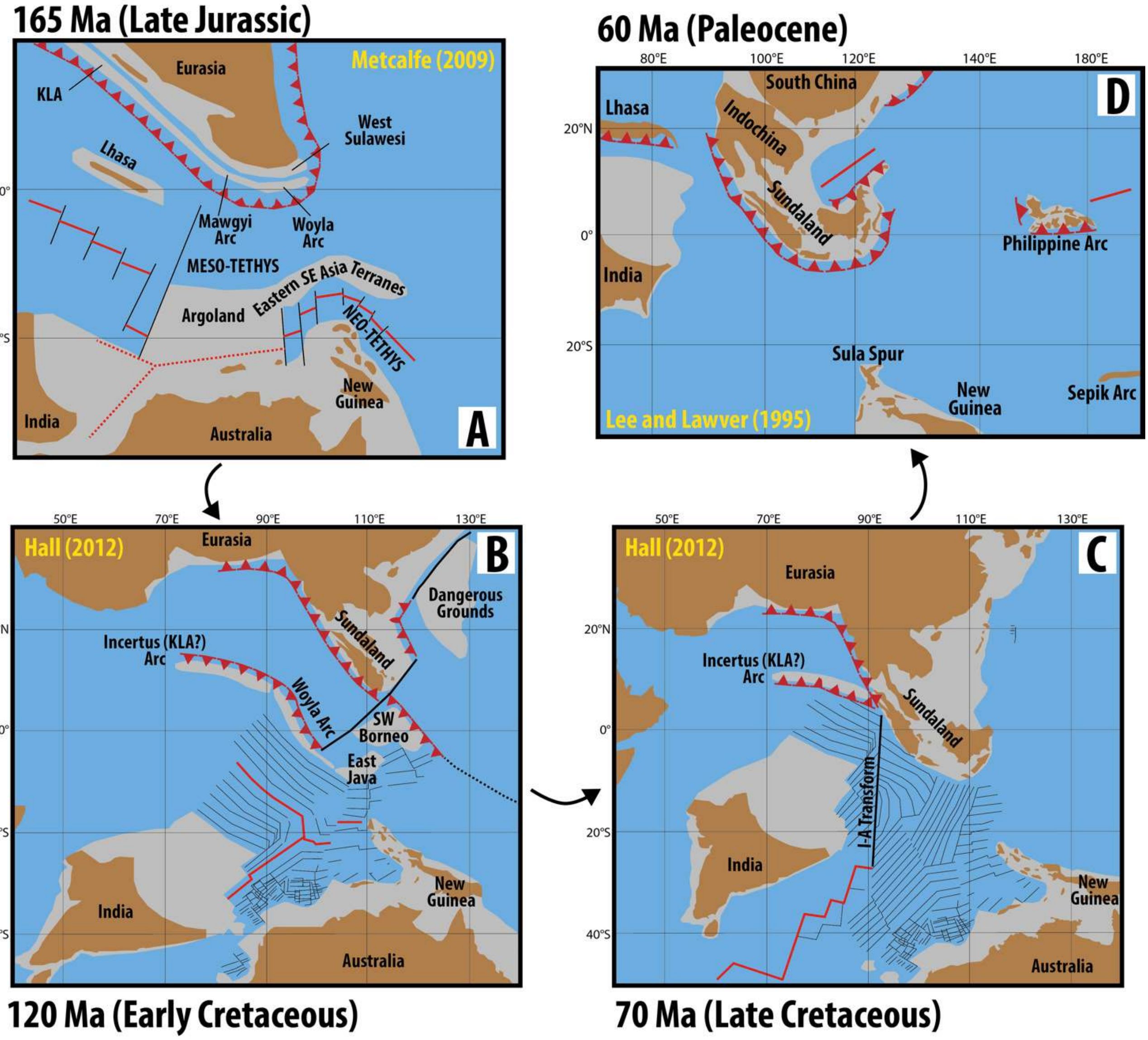

Fig. 4. A wide range of tectonic reconstructions have been proposed for the eastern Tethys between the India-Eurasia convergence zone and New Guinea. A) The Late Jurassic rifting event along northern Gondwana has been modelled as a westward propagating rift from New Guinea towards Argoland, and joining up as a triple junction in between India and Australia in the model of Metcalfe (2009). The rifting mechanism is implied as northward slab pull from Tethyan subduction along southern Eurasia. B) The model of Hall (2012) instead invokes a south-dipping subduction zone along northern Gondwana in the latest Jurassic, leading to the opening of the Neo-Tethys as a large back-arc basin. The related Incertus Arc likely represents the Kohistan-Ladakh (KLA) and Woyla arc systems in the Neo-Tethys. C) By the Late Cretaceous, subduction polarity reverses across the Incertus Arc to produce northward slab pull along a north-dipping intra-oceanic

subduction zone. The Hall (2012) model imposes a subduction hiatus along southern Sundaland between 90 and 45 Ma, which requires the segmentation of the Neo-Tethys across a transform that cross-cuts Tethyan seafloor fabric at ~90°E (I-A Transform). D) The model of Lee and Lawver (1995) presents eastern Tethyan plate reconstructions since 60 Ma in a South China fixed reference frame. The size of Greater India is similar as proposed in Hall (2012), but is about twice the northward extent presented in this study, largely to accommodate an India-Eurasia continent-continent collision at ~55 Ma. The Lee and Lawver (1995) model also presents all plate rotation parameters, which enables the reproducibility and testability of this model. A common feature between the models (A-D) is relatively less detail for the New Guinea region, which has been difficult to reconstruct due to the lack of data and the dominance of complex interactions between Asian and Pacific subduction systems.

1.1 Plate tectonic models of the eastern Tethys

As many generations of plate reconstructions have been proposed for the eastern Tethyan tectonic domain, it is useful to understand the historical context and help categorize successive generations of models that have been proposed. Even before the acceptance of plate tectonic principles, Southeast Asian geology was of great interest due to significant hydrocarbon (Halbouty et al., 1970; Wennekers, 1958) and metallogenic (Brown, 1951; Leith, 1926; Matthews, 1990; Penrose, 1903) discoveries. Early attempts to explain the geology of Southeast Asia led to a large number of competing hypotheses. Fairbridge (1963) explained the geological affinities between Southeast Asia and Gondwana by invoking a range of mechanisms from now-abandoned ideas of mantle contraction, mantle expansion, rising and sinking land bridges, galactic expansion, to then emerging ideas of continental drift. A few years on, Audley-Charles (1966) provided the first synthesis of stratigraphic evidence to describe the region’s Mesozoic paleogeographic evolution in the context of continental drift, with special reference to paleo-latitude indicators from paleo-climatic and paleo-magnetic data.

It was only in the 1970s that plate tectonic principles of subduction, seafloor spreading and transform tectonic boundaries (Forsyth and Uyeda, 1975; Le Pichon, 1968; McKenzie, 1969; McKenzie and Parker, 1967) were invoked in first-generation continental reconstructions (Table 1) to explain the present-day tectonic complexity of the eastern Tethys (Fitch, 1972; Hamilton, 1979; Katili, 1971). These models were subsequently used to create the first schematic "plate reconstructions" (Katili, 1975) that largely focused on the dominance of active volcanic arcs and associated subduction zones in controlling the tectonic complexity of the region. Importantly, the work of Katili (1975) identified a number of parallel and arcuate paleo-arc systems, which recorded post-Permian subduction of Tethyan oceanic crust. Pioneering work in the 1970s and 1980s applied paleomagnetic techniques to infer that parts of Southeast Asia originated from the northern Gondwana margin (McElhinny et al., 1981), and more specifically somewhere between north Africa and Greater India (the portion of India currently under-thrust beneath Eurasia) (Stauffer, 1983). Although the origin of Southeast Asian continental fragments from Arabia or Africa in the Paleozoic have since been discounted (see Metcalfe, 1988; Metcalfe, 1994; Metcalfe, 1999; Veevers, 2004), these early works established the wider notion of Southeast Asian crustal accretion via the northward transfer of continental fragments originating from the northern Gondwana margins (Fig. 4A).

Pioneering reconstructions of Gondwana breakup, and the northward transfer of crustal fragments towards Asia, were largely presented as schematic scenarios portraying the drift of continents with consideration of some major regional plate boundaries. Pigram and Panggabean (1984) and Audley-Charles (1988) combined regional stratigraphic composite wells to identify a major Late Jurassic breakup unconformity across the NW Australian margin, which suggested that a number of continental fragments had detached to form the north Gondwana passive margin and open the "Mesozoic Tethys" ocean basin. Based on the interpretation of rift-drift sedimentary sequences, including the timing of the post-breakup unconformity, Pigram and Panggabean (1984) provided schematic reconstructions of the Late Jurassic drifting episode and concluded that seafloor spreading initiated sometime in the Early Jurassic along New Guinea and Middle Jurassic along the NW Australian shelf. The generally northward transfer of Gondwana terranes opened successive Tethyan ocean basins including the rifting of a ribbon continent comprising Iran, North Tibet (Qiangtang) and Indochina to open "Tethys II" in the late Permian (Audley-Charles, 1988). A subsequent major rifting phase in the Late Jurassic opened the "Tethys III", detaching fragments including South Tibet (Lhasa), West Burma, Malaya, Borneo, Sulawesi, Sumatra and a number of Banda allochthons (Audley-Charles, 1988). Importantly, the work of Audley-Charles (1988) and Audley-Charles et al. (1988) introduced a paleogeographic reconstruction framework using the computerised University of Cambridge Atlas plotting workflow, which we classify as a second generation reconstruction methodology (Table 1). This early generation of reconstructions assessed the prior continental affinities and inferred major rifting phases using biostratigraphic constraints, as well as made use of paleomagnetic syntheses and structural interpretations from seismic sections to infer rift and drift histories.

The third generation of plate reconstructions, largely developed in the late 1980s and throughout the 1990s (e.g., Besse and Courtillot, 1988; Daly et al., 1991; Jolivet et al., 1989; Lee and Lawver, 1994; Lee and Lawver, 1995), made use of extensive identifications of marine magnetic anomalies from the Indian Ocean and West Pacific calibrated to a geological timescale (e.g., Taylor and Hayes, 1980; Taylor and Hayes, 1983). The seafloor spreading histories, supplemented with paleomagnetic data from the continental blocks (e.g., Haile et al., 1977), were applied to make plate reconstructions using rigid body motions on the surface of a sphere (i.e., Euler rotations). The Jurassic to recent plate reconstructions of Besse and Courtillot (1988) and Scotese et al. (1988) were an important benchmark for subsequent plate motion models, as the work synthesised marine magnetic anomalies and continental paleomagnetism, yet also took into account the plate boundary evolution. Pertinent to this study, the work of Besse and Courtillot (1988) and Scotese et al. (1988) enabled reproducibility by providing finite rotation parameters.

Although the plate reconstructions of Lee and Lawver (1994) and Lee and Lawver (1995) covered only the Cenozoic evolution of Southeast Asia (Fig. 4d), these were the first detailed regional reconstruction that published testable and reproducible finite rotation parameters that quantitatively described the motion of Southeast Asian crustal elements, building on the more regional approach presented in Jolivet et al. (1989). The relative plate motions, provided as finite rotations, were linked into a plate motion hierarchy that tied back to the South China block, and thus only provide a regional perspective (Lee and Lawver, 1994; Lee and Lawver, 1995). However, the provision of Euler rotations significantly increased their utility even over more recent models as they allow for reproducibility and refinement by subsequent researchers.

A major improvement in regional plate reconstructions was presented in Hall (1996), and subsequent works by the same author (Hall, 2002; Hall, 2012) (Fig. 4B-C), where the regional plate reconstructions were embedded in a global plate circuit – that links Australia and India back to Africa, and Asian fragments through Eurasia, North America and Africa. Using a global plate circuit combines relative plate motions with a frame of reference with respect to the mantle using a hotspot frame (e.g., Müller et al., 1993), which enables linkages between the plate-mantle system. In the absence of hotspot tracks (i.e., before ~120 Ma), plate reconstructions make use of paleomagnetic reference frames (e.g., Hall and Spakman, 2015), which enable the reconstruction of paleo-latitudes of climate-sensitive data, and can be corrected for True Polar Wander to create more explicit links between the plate-mantle system in deep time. The reconstructions presented in Hall (1996) and Hall (2002) provide a regional post-Jurassic evolution of the India-Eurasia convergence zone, Southeast Asia and New Guinea embedded in a detailed synthesis of relevant data, and are presented in 1 Myr interval snapshots. Such high temporal resolution is important for capturing major plate boundary reconfigurations and resulting changes in plate motion magnitudes and directions, such as the change in India's plate motions and northward advance from ~100 Ma

(Gibbons et al., 2015; Matthews et al., 2012; van Hinsbergen et al., 2011). Although the reconstructions are presented in 1 Myr intervals, no relative or absolute plate rotation parameters have been provided, which limits the testability of such models.

These first- to fourth-generation plate reconstructions provide considerable detail and insight into the tectonic evolution of the eastern Tethys, but cannot be easily linked to methods that take into account the geodynamic evolution of the plate-mantle system. Schematic reconstructions cannot be linked to numerical models of convection as they usually lack the continuous network of plate boundaries through time that enables the use of plate velocities as surface boundary conditions. As plate motions are inextricably linked to mantle convection (Hager and O'Connell, 1981; Turcotte and Oxburgh, 1972), and since much of the Tethyan seafloor spreading history has since been subducted (Hutchison, 1975; Şengör et al., 1988), some authors have inferred plate motion histories from high velocity seismic anomalies as given by mantle tomography models (Hall and Spakman, 2003; Hall and Spakman, 2015; Replumaz et al., 2004; van der Voo et al., 1999b; Wu et al., 2016). We expand on these approaches and make use of our most recent plate reconstructions coupled to numerical models of mantle convection that are validated using seismic tomographic images and a suite of onshore and offshore geological constraints.

Table 1. Generations of continental and plate reconstructions depicting the kinematic and geodynamic evolution of Southeast Asia.

Generation of reconstruction	**Description**	**Examples**
First	Schematic reconstructions of continental motions.	Pigram and Panggabean (1984) Metcalfe (1988)
Second	Continental reconstructions are made using digital approaches, with schematic paleo- plate boundaries.	Audley-Charles et al. (1988) Rangin et al. (1990)
Third	Additionally provide regional reconstructions using seafloor spreading histories, constraints from onshore geology	Besse and Courtillot (1988) Scotese et al. (1988)

	(paleomagnetism, stratigraphy, seismic, structural, biogeography, etc.) and an incomplete network of plate boundaries. Although these models are classified as 3rd generation reconstructions, they have a significant advantage over any other reconstructions that do not provide Euler rotation parameters that are provided in 5th generation models. These models are important examples of reproducible and testable plate reconstructions of Southeast Asia.	Jolivet et al. (1989) Lee and Lawver (1994) Lee and Lawver (1995)
Fourth	Regional reconstructions embedded in a global rotation hierarchy, constraining relative plate motions using seafloor spreading histories that are tied to an absolute hotspot or paleomagnetic reference frame. Synthetic seafloor spreading histories are generated in regions and times where seafloor has been subducted.	Hall (1996) Hall (2002) Hall (2012) Stampfli and Borel (2002)[#]
Fifth	A continuous global network of evolving plate boundaries is modelled, with complete model rotation parameters and digital geometry files released for testability and reproducibility. Such models can be linked to regional and global geodynamic numerical calculations that link plate tectonics with underlying mantle convection. Some of these models incorporate regional refinements that include retro-deformation of continental crust to provide better full-fit reconstructions of Pangea.	Gurnis et al. (2012) Seton et al. (2012) Zahirovic et al. (2012) Zahirovic et al. (2014) Domeier and Torsvik (2014)^ Gibbons et al. (2015) This study
Future	Build on previous approaches with stronger emphasis on quantifying uncertainties, and using ensemble computer modelling that incorporates all constraints (offshore and onshore) simultaneously and all relevant uncertainties to derive a quantitative "best-fit" plate reconstruction that is fully consistent with plate boundary forces and mantle convection models. Global plate reconstructions incorporate all major regions of deformation to provide better full-fit reconstructions and address the oversimplification of plate rigidity assumptions.	Such models are not yet available, and represent an aspirational goal to produce better plate tectonic reconstructions.

^ The Domeier and Torsvik (2014) reconstructions cover the Late Paleozoic global plate motion history, and include major blocks of Southeast Asia.

[#] The model of Stampfli and Borel (2002) has linked plate boundaries and synthetic seafloor spreading histories, which are important components of 5th-generation models, but only provides snapshots without rotation parameters or (evolving plate boundary) geometries.

1.2 Seismic tomography constraints

Seismic tomographic techniques have provided an important link between the present-day arrangement of plate boundaries and deep mantle structure around Southeast Asia resulting from long-term subduction of Tethyan and Pacific lithosphere. High-resolution P-wave seismic tomographic models have demonstrated that the Sunda slab from the subduction of the Indo-Australian oceanic plate penetrates to depths of ~1500 km (Li et al., 2008; Widiyantoro and van der Hilst, 1996), and that it is distinct from the older and deeper Tethyan slabs (Widiyantoro and van der Hilst, 1996). Due to the complexity of Tethyan convergence, and the lack of preserved seafloor spreading histories, interpretations of mantle structure provide an important additional insight into the past geometry and evolution of active margins in the region. For example, Hall and Spakman (2002) and Hall and Spakman (2003) used a P-wave seismic tomographic model to infer Cenozoic subduction histories in the vicinity of the northern Australian margin. Hall and Spakman (2003) interpreted the Bijwaard and Spakman (2000) P-wave seismic tomographic model to suggest north-dipping subduction north of Australia along the Philippine Archipelago occurred between 45 and 25 Ma, and inferred that little subduction occurred north of Australia since 25 Ma due to the likelihood of a margin dominated by strike-slip motion rather than convergence. Hall and Spakman (2015) recently attributed the 1600 km deep Sunda slab to subduction since 45 Ma, but discounted the possibility that the Proto South China Sea slab is in the upper mantle, and concluded that it is instead likely in the lower mantle at ~1200 km depth. Further south, a large east-west slab beneath Australia (including Lake Eyre) at ~800-1200 km depths has been interpreted to be the result of north-dipping subduction that ceased following accretion of the Sepik Terrane along New Guinea at ~50 Ma (Schellart and Spakman, 2015).

The India-Eurasia Tethyan mantle structure was interpreted in van der Voo et al. (1999b) where a global P-wave seismic tomographic model (Bijwaard et al., 1998) was used to infer the subduction history related to post-Jurassic subduction (Fig. 5). The large slabs, with a generally northwest-southeast trend and largely at mid-mantle depths, were interpreted to be the result of two simultaneous north-dipping subduction zones in the Neo-Tethys (van der Voo et al., 1999b), a scenario which requires a two-stage India-Eurasia collision. Hafkenscheid et al. (2006) elaborated on this approach by quantifying Tethyan slab volumes and inferring average slab sinking rates in the mantle. Hafkenscheid et al. (2006) tested end-member scenarios of convergence, including long-lived Andean-style subduction following Norton (1999) and Şengör and Natal'in (1996). Instead, the analysis by Hafkenscheid et al. (2006) suggested that an additional intra-oceanic subduction zone, following Stampfli and Borel (2002), could better reproduce the volume and distributions of slabs interpreted from 3D seismic tomography. The preferred scenario in Hafkenscheid et al. (2006) invoked an arc-continent collision between Greater India and the Spong Arc, likely contemporaneous with the Kohistan-Ladakh Arc (McDermid et al., 2002), at ~65-60 Ma, with continent-continent collision occurring at ~48 Ma, and infer a “free sinking rate” (i.e., when not attached to a subducting plate) of 3 and 2 cm/yr in the upper and lower mantle, respectively.

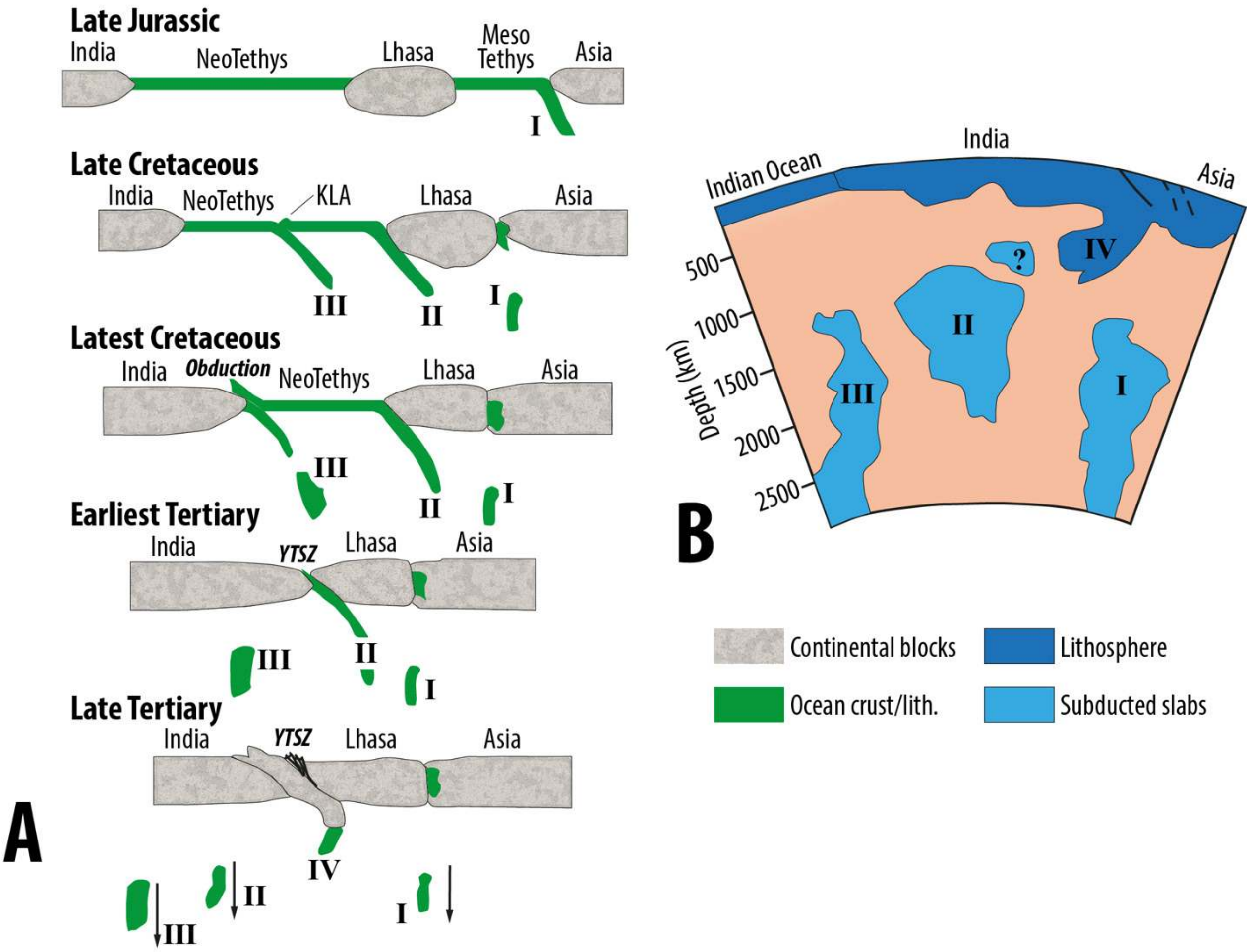

Fig. 5. A) Schematic synthesis of Tethyan subduction history accommodating India-Eurasia convergence, as interpreted from P-wave seismic tomography by van der Voo et al. (1999b). B) The three slab volumes in the lower mantle are interpreted as representing intra-oceanic subduction and a two-stage India-Eurasia collision. Figure adapted from van der Voo et al. (1999b). Note that the Tethyan ocean basin nomenclature in van der Voo et al. (1999b) differs slightly from the terminology used in this study. KLA – Kohistan-Ladakh Arc, YTSZ – Yarlung Tsangpo Suture Zone.

Incorporating 3D seismic tomographic interpretations, Replumaz et al. (2004) combined an assumption of vertical slab sinking with a tectonic reconstruction of Southeast Asia in a Siberia reference frame, and interpreted the pre-collision geometry of the southern Eurasian active margin using tomographic depth slices. The analysis of tomography linked to a “retro-deformation” model of block motions in Southeast Asia (using available fault offsets and slip rates) for Cenozoic times suggests that the India-Eurasia continental collision occurred sometime between 55 and 40 Ma, based on changes in slab morphology, and suggests a sinking rate of 5 cm/yr in the upper mantle and 2 cm/yr in the lower mantle. The resulting upper mantle sinking rates are slightly higher than the values suggested by Hafkenscheid et al. (2006). A similar approach of age-coding slabs in P- and S-wave seismic tomographic depth slices, assuming constant and vertical slab sinking was used in Zahirovic et al. (2012) and Zahirovic et al. (2014) as a general estimate of the location of Tethyan subduction zones that were then implemented into a global plate motion model. One important distinction was the use of multiple P- and S-wave seismic tomographic models, which is an important consideration in determining the distribution of Tethyan slabs. To supplement the assumption of vertical slab sinking, Zahirovic et al. (2012) used numerical mantle convection models kinematically driven by time-dependent plate reconstructions, and found that an intra-oceanic subduction scenario, as suggested by van der Voo et al. (1999b), Hafkenscheid et al. (2006) and Aitchison et al. (2007), better reproduced the Tethyan mantle structure than Andean-style subduction alone.

More generally, the approach of age-coding slabs in seismic tomographic depth slices has been applied globally to derive average slab sinking rates (Butterworth et al., 2014; van der Meer et al., 2010), and to propose a subduction reference frame using the assumption of vertical sinking and constant sinking rates (van der Meer et al., 2010). The cataloguing of global slab volumes by van der Meer et al. (2010) suggests that a ~15 to 20° longitudinal global shift of all continents is required to account for the observed distribution of post-Jurassic slabs in the mantle. Such an observation is an important first-order constraint of paleo-longitude in the absence of preserved hotspot tracks during the Late Jurassic and Early Cretaceous, and provides an estimated average global slab-sinking rate of 1.2 ± 0.3 cm/yr. A similar synthesis of slabs interpreted from seismic tomography in Butterworth et al. (2014) suggests a comparable average sinking rate of 1.3 ± 0.3 cm/yr for the whole mantle. However, such an approach does not take into account the contrasting viscosities of the upper and lower mantle, or the effects of slab stagnation and lateral slab advection from mantle flow, which may be an important factor contributing to Tethyan mantle structure (Becker and Faccenna, 2011; Zahirovic et al., 2012). To address this, Butterworth et al. (2014) made use of global numerical modelling of mantle flow to test competing absolute reference frames against present-day seismic tomographic constraints, and suggest that the longitudinal correction argued in van der Meer et al. (2010) is likely too large. The numerical modelling approach in Butterworth et al. (2014) highlighted the need to account for variable slab sinking rates resulting from factors such as oblique convergence, diachronous collisions and suturing, as well as two orders of magnitude increase in viscosity between the upper and lower mantle. The slab sinking rates from numerical mantle convection models in Butterworth et al. (2014) suggests a global mantle sinking rate of 1.5 to 2.0 cm/yr, which is also consistent with the 2.0 ± 0.8 cm/yr mantle sinking rate inferred from mantle flow modelling (Steinberger et al., 2012). However, other work applying mantle flow modelling highlights the time-varying nature of slab sinking rates, which is an important consideration when interpreting slabs from the present-day snapshot in seismic tomography (Bower et al., 2013).

1.3 Numerical modelling of Tethyan geodynamics

The evolution of the Tethyan realm has been the focus of decades of research, to better understand the India-Eurasia collision and the complex tectonics of Southeast Asia and New Guinea. A wide range of physical (analogue) and numerical experiments at crustal, lithosphere and mantle scales have revealed important aspects of the plate-mantle system that are responsible for the geodynamics of the Tethyan, Eurasian and Pacific tectonic domains. Since our approach requires modelling in a spherical domain with assimilation of plate reconstructions, only numerical methods are appropriate to study the long-term eastern Tethyan subduction history in a regional and global framework.

Wide ranges of numerical approaches exist to model mantle behaviour – including forward or backward advection models (including inverse and adjoint approaches), forward models with data assimilation, and fully geodynamic models that do not have imposed boundary conditions. Forward models assume that the plate motion histories are a reasonably good recorder of plate-mantle evolution, and use the plate motions as a surface kinematic boundary condition to predict mantle structure that can be compared to seismic tomography. Backward advection models use seismic tomography (as a present-day snapshot of the mantle) as an input where the seismic velocity anomalies are converted to density perturbations, assuming that the bulk of the anomaly has a thermal source, and the sign of gravity and time reversed to compute the past position of the mantle material (Glišović and Forte, 2014; Liu and Gurnis, 2008; Steinberger and O'Connell, 1998). The backward advection models take into account the complex present-day mantle structure, but can only be successfully used for times since ~70 Ma due to the inherent issues of irreversible thermal diffusion and the interaction of the boundary layers with internal flow (Bunge et al., 2003; Conrad and Gurnis, 2003; Steinberger and O'Connell, 1998). More advanced approaches using adjoint models overcome the limitations of irreversible backward advection (Liu and Gurnis, 2008; Spasojevic et al., 2009), but have yet to be applied to the Tethyan domain. Since our region of interest requires deeper time considerations, we use forward geodynamic flow experiments that are tested against mantle tomography.

1.3.1 Numerical models of India-Eurasia convergence

Following the interpretation of discrete Tethyan slab volumes at mid-mantle depths beneath India in P-wave seismic tomographic models by van der Voo et al. (1999b), a numerical approach using a 2D box (with 16.5 km mesh resolution) was used by Jarvis and Lowman (2005) to interpret the inferred Tethyan mantle structure (Fig. 6). A number of experiments were conducted, specifically varying the poorly-constrained viscosity contrast between the upper and lower mantle, with the results requiring a lower mantle that was at least 30 times more viscous than the upper mantle to maintain Tethyan slabs at mid-mantle depths (Jarvis and Lowman, 2005). The resulting upper and lower mantle viscosity contrast from Jarvis and Lowman (2005) was also consistent with earlier estimates of a 10 to 30 times more viscous lower mantle from global models fitting geoid anomalies over slabs (Hager, 1984).

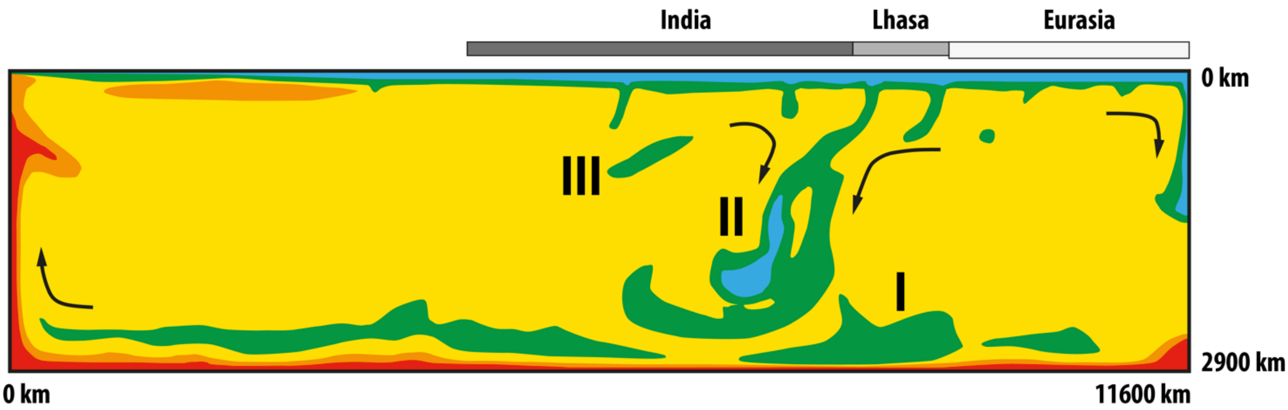

Fig. 6. Numerical 2D box model of India-Eurasia convergence, adapted from Jarvis and Lowman (2005), provided an important benchmark for quantifying and testing tectonic reconstruction scenarios of India-Eurasia convergence and Tethyan geodynamics.

Due to the limitation of a 2D box set-up, the India-Eurasia convergence modelled by Jarvis and Lowman (2005) required a simplified convergence history along a single transect despite a complex active margin with periods of oblique convergence. The applied velocity boundary condition led to symmetric downwellings rather than one-sided subduction. However, the work of Jarvis and Lowman (2005) highlighted the need for quantitative approaches to test plate reconstructions, while suggesting a lower limit on the viscosity contrast between the upper and lower mantle. The approach utilised simple kinematic boundary conditions for the convergence velocities, which were assumed to be ~17 and 6 cm/yr (Besse and Courtillot, 1988) before and after the India-Eurasia collision, respectively, at 42 Ma. Their subsequent numerical approach made use of a simple sinking slab in a 2D and 3D Cartesian box (Jarvis and Lowman, 2007), which suggested a viscosity contrast of a factor of 100 to 300 between the upper and lower mantle to maintain Tethyan slabs in the mid-mantle. These results suggested that Jurassic slabs likely retain a thermally anomalous signature with respect to the ambient mantle, enabling their detection by seismic tomographic techniques.

Becker and Faccenna (2011) used a 3D global approach to investigate the plate driving forces acting on the circum-Tethyan regions, and converted P- and S-wave seismic tomographic models to density anomalies driving instantaneous mantle flow models. They found that there was a dominant first-order mantle conveyor belt with northward velocities in the shallow mantle beneath India, sinking of mantle material near the suture zone, and accompanying southward flow that is interrupted by mantle upwelling in the region of the Carlsberg and Central Indian ridges. The work highlights the power of global models in capturing the complexity of slab interactions from circum-Tethyan subduction, with results suggesting that large-scale mantle flow and an associated Tethyan conveyor supports ongoing indentation by India. Similarly, a 3D global spherical approach using CitcomS (Zhong et al., 2000) was applied in Zahirovic et al. (2012), where plate kinematic boundary conditions were applied from 140 Ma to test end-member subduction scenarios accommodating India-Eurasia convergence. The forward numerical model predictions were compared to slabs interpreted from seismic tomography (Zahirovic et al., 2012), and showed that the mantle structure could be better reproduced when taking into account intra-oceanic subduction and a two-stage India-Eurasia collision (Aitchison et al., 2007; Hafkenscheid et al., 2006; van der Voo et al., 1999b). This earlier work highlights the need to test end-member plate reconstruction scenarios using mantle flow models, and comparisons to mantle structure from seismic tomography as an additional criterion for reconciling surface geology.

More recently, a 3D approach was also employed in Yoshida and Hamano (2015), who ran a forward convection model from Pangea times, but without applying a kinematic boundary condition or being able to incorporate one-sided subduction. Although many of the experiments failed to reproduce present-day arrangements of continents (such as predicting a problematic fit of

Antarctica with South America), one aspect of the models reproduced the approximate present position of India and highlights the requirement of long-lived subduction along southern Eurasia since Pangea breakup (Yoshida and Hamano, 2015). When considering the motion of India towards Eurasia, the anomalously high velocities (more than 14 cm/yr) of India between ~80 and 65 Ma can be modelled numerically through a viscous coupling mechanism between two simultaneous north-dipping subduction zones in the Neo-Tethys prior to India-Eurasia collision (Jagoutz et al., 2015). This modelling approach suggests that two subduction zones in the Neo-Tethys are required to account for the high convergence rates as long-lived (~20 Myr) accelerations cannot be explained by plume influences, which are likely to diminish over a shorter timeframe of several millions of years (van Hinsbergen et al., 2011).

When considering the mantle-surface interaction from Tethyan tectonics,, the work of Pusok and Kaus (2015) used a 3D numerical box model that captured both subduction processes and the resulting topographic response to the India-Eurasia collision, providing insight on the formation of oroclines in the eastern and western syntaxes of the convergence zone, as well as the uplift of the Tibetan Plateau and lateral expulsion of continental material. Major advances are also being made in reducing the uncertainties in the rheology of the mantle and lithosphere in such numerical models, with the recent work (Baumann and Kaus, 2015) highlighting a new, and currently computationally-intensive, approach of parallel inversion of observables including the gravity field, topography and GPS velocities to better model the lithospheric and crustal rheology. Other advances in inverse methods have the potential to fully incorporate the details of slabs and their coupling to lithospheric plates with fault-zones between plates with fully non-linear rheologies, which remains one of the largest uncertainties in mantle convection modelling (Ratnaswamy et al., 2015; Worthen et al., 2014). Such approaches provide a framework for geodynamic computations that capture realistic non-linear rheologies, including strain rate weakening and yielding, to better account for plate velocity observations, complex slab-trench interactions, and intra-plate deformation that goes beyond the simplifying assumption of plate rigidity (Alisic et al., 2012; Alisic et al., 2010).

1.3.2 Numerical modelling of Southeast Asia and New Guinea geodynamics

Few geodynamic models of mantle-, lithospheric- and crustal-scale evolution exist for the tectonically complex and less constrained Sundaland and New Guinea regions than for other parts of the Tethyan tectonic domain. For example, the synthesis by van Ufford and Cloos (2005) of at least six competing proposed scenarios for the Cenozoic evolution of New Guinea highlights the uncertain chronology of major tectonic events, as well as poorly-constrained subduction polarities. As a result, much of the numerical modelling has been restricted to understanding the present-day geodynamic character of the region. Ghose et al. (1990) used focal mechanism solutions to build a 3D finite element numerical experiment of subducted slabs and generalised mantle structure in the Sundaland region to compute the flow and stress field acting on the overriding continental promontory. The results indicate significantly higher plate coupling across the Sumatra segment of the Sunda Trench, resulting in a higher seismogenic potential than the Java region. This may be due to lower coupling assumed to be due to the lubricating effect of soft sediments in the trench (e.g., Clements and Hall, 2011), as well as the subduction of older Indian Ocean crust than along the Sumatra segment (Ghose et al., 1990).

North of New Guinea, the geodynamic significance of the Philippine Sea Plate has been the subject of a number of studies that employ numerical modelling to quantify the effects of Izu-Bonin-Mariana subduction initiation (Gurnis et al., 2004; Hall et al., 2003; Leng and Gurnis, 2015) on the Pacific Plate boundary forces, and its contribution to a change in Pacific Plate motion between ~50 and 40 Ma based on force balance calculations (Faccenna et al., 2012). Temporally linked to the inception of (proto-) Izu-Bonin-Mariana subduction, major changes in subduction along New Guinea in the Eocene have been invoked to explain the acceleration of Australia's northward motion (Schellart and Spakman, 2015; Zahirovic et al., 2014) from ~43 Ma (Williams et al., 2011). Schellart and Spakman (2015) identified a subducted slab at depths between ~800 and 1200 km beneath Lake Eyre in eastern Australia, and argued based on a simple Stokes flow model that the topographic depression is caused by dynamic subsidence induced by the sinking of a slab that detached along New Guinea at ~55-45 Ma. However, the complex interaction of slabs from regional subduction zones plays an important role that can only be tested in regional and global geodynamic numerical simulations that capture the time-dependent evolution of Southeast Asian plate boundaries.

2 Methods

2.1 Plate tectonic reconstructions

Reconstructions of the Tethyan domain have taken many forms over decades of research (see Section 1.1, Table 1, Fig. 4), with the post-Pangea plate reconstruction timeframe (since ~200 Ma) generally associated with lower uncertainties than earlier times due to greater preservation of oceanic crust (Zahirovic et al., 2015). Due to the ambiguity in reconstructing regions with no preserved seafloor spreading records and/or poor geological constraints, testing alternative scenarios becomes an avenue to evaluate the uncertainty inherent in plate reconstructions. In this study we present a new post-Jurassic plate motion model spanning the Tethyan region from the westernmost India-Eurasia convergence segment, in the vicinity of Kohistan-Ladakh, eastward to Southeast Asia (including Sundaland and the proto-South China Sea) and Papua New Guinea. The model is also compared to the previous synthesis of the region presented in Zahirovic et al. (2014) for Southeast Asia and New Guinea, and in Gibbons et al. (2015) for the India-Eurasia convergence zone, highlighting the alternative kinematic scenarios that can account for the constraints from marine and onshore data (Supplementary Animation 1). The refinements presented in this study focus on an alternative model for the Neo-Tethys, transferring the East Java-West Sulawesi blocks from the Argo Abyssal Plain on the NW Australian shelf towards Sundaland, as well as refinements to the evolution of the Kohistan-Ladakh, Woyla, and Philippine intra-oceanic arcs, in the context of the evolving Sundaland and New Guinea continental margins (see Section 3).

Relative plate motions are derived from preserved seafloor spreading histories (as in third and fourth generation reconstructions from Table 2), where seafloor magnetic anomalies are identified and combined with directional constraints from fracture zones to compute the Euler rotation that defines the relative rigid body motions on the surface of a sphere. We use the Global Seafloor and Magnetic Lineation Database of previously-published magnetic anomaly picks (Seton et al., 2014) and fracture zone geometries (Matthews et al., 2011), which were combined to compute relative plate motion parameters in previous studies (see discussion and references in Seton et al., 2012), preferably using the least-squares best-fit statistical method following Hellinger (1981) and Royer and Chang (1991). The rotation parameters are calibrated to the geomagnetic polarity reversal timescale of Gee and Kent (2007), which is an updated timescale compared to the one used in our previous plate reconstruction of the Tethyan region in Zahirovic et al. (2014) and Gibbons et al. (2015). In the absence of preserved seafloor spreading histories, we use onshore geological constraints to estimate the pre-rift position of continental fragments, the timing and trajectory of rifting, as well as the age and location of the accretion events (Tables 2-3). We construct synthetic oceanic plates that are consistent with plate tectonic driving mechanisms and reasonable relative plate motions across plate boundaries (e.g., convergence across subduction zones, divergence across mid-oceanic ridges, strike-slip motion along transform faults).

The relative plate motions form a chain that is hierarchical and typically ties all plate motions back to Africa, largely due to the central position of Africa within Pangea and relative stability because it is surrounded by mid-oceanic ridges (Torsvik et al., 2008). The motion of Africa is expressed with respect to the underlying mantle, using Indo-Atlantic or global hotspot tracks since

~100 Ma to derive a frame of reference for the global plate motion model. For earlier times when no hotspot tracks are available, the True Polar Wander- (TPW) corrected paleomagnetic reference frame of Steinberger and Torsvik (2008) is used. Due to the lack of paleo-longitudinal constraints in a paleomagnetic reference frame resulting from the radial symmetry of the Earth's magnetic dipole, we apply a 10° longitudinal shift gradually between 70 and 105 Ma to the TPW-corrected reference frame, following Butterworth et al. (2014) and van der Meer et al. (2010) to provide a better paleo-longitudinal link between subduction zones and Jurassic and Cretaceous subducted slabs interpreted from seismic tomography.

The combination of relative plate motions with an absolute reference frame enables the computation of absolute plate motions through time, modelled using the GPlates software (Boyden et al., 2011). Evolving plate boundaries are constructed using continuously-closing plate polygon algorithm (Gurnis et al., 2012), which provides global coverage of plates through time in 1 Myr intervals. The Tethyan plate motions are embedded in a global model, which is based on the synthesis in Seton et al. (2012) with regional refinements that are documented in Müller et al. (2016). The time-dependent plate boundaries, seafloor age-grids and plate velocities are assimilated into the numerical models of mantle convection, described in Section 2.3. By considering the evolution of the entire plate (Gurnis et al., 2012; Stampfli and Borel, 2002) rather than only focusing on continental blocks, plate reconstructions can be linked to geodynamic models (Conrad and Lithgow‐Bertelloni, 2004; Lithgow‐Bertelloni and Richards, 1998). The coupling of plate kinematics to geodynamic models provides the opportunity to reproduce the mantle structure interpreted from seismic tomography as well as reconstruct past mantle flow using the present-day surface geology and tectonics as constraints.

Table 3. Constraints used to construct plate motion model.

Region	**Event**	**Timing**	**Dating method/ interpretation**	**Interpretations based on data and**

				models
Australian NW Shelf	Onset of rifting	Sometime in Late Jurassic	Stratigraphic rift-drift sequences	Pigram and Panggabean (1984)
	Triple junction mid-oceanic ridge configuration	Latest Jurassic	Geometrical requirement, and evidence of possible plume influence	Audley-Charles (1988), Audley-Charles et al. (1988), Gibbons et al. (2012), Rohrman (2015)
	Onset of seafloor spreading	155 ± 3.4 Ma	K-Ar of basaltic basement	Gradstein and Ludden (1992)
West Sulawesi, East Java, Mangkalihat and easternmost Borneo	Onset of rifting	Late Jurassic	Biostratigraphic constraints in Paremba Sandstone and shallow marine sandstones in Bantimala Complex	Sukamoto and Westermann (1992), Wakita (2000)
	Onset of seafloor spreading	~158-155 Ma	K–Ar of diorite, microgabbro and basaltic dyke	Polvé et al. (1997)
	Oldest seafloor spreading magnetic anomalies	M25A M26 (~153-155 Ma)	Magnetic anomaly identifications from shiptracks	Heine and Müller (2005) Gibbons et al. (2012)
	Youngest preserved seafloor spreading magnetic anomaly in the Argo Abyssal Plain region	M10Ny, 128.9 Ma	Magnetic anomaly identifications from shiptracks	Gibbons et al. (2012)
	Suturing of ‘Argoland’ to	~80 Ma	Stratigraphy, K–Ar and U–Pb of	Wakita (2000) Clements and Hall

	southwest Borneo core		metamorphics, synthesis of previous studies	(2011)
New Guinea	Rifting on northern New Guinea (opening of Sepik ocean basin)	Late Jurassic 172 Ma ~157 ± 16 Ma	Jurassic granite in Bena Bena Terrane SSZ ophiolites in Central Ophiolite Belt	Davies (2012) Permana (1998)
	Subduction influence on eastern New Guinea	Early Cretaceous	Kondaku Tuffs	Dow (1977), Rickwood (1954)
	Onset of Sepik ocean basin subduction	Maastrichtian (~71 to 66 Ma) 68 Ma	Stratigraphic correlation and dating using foraminifera High-temperature metabasites on West Papuan Ophiolite	Worthing and Crawford (1996) Davies (2012)
	Sepik Terrane docking with New Guinea	35-31 Ma ~30 Ma	Ar-Ar age of Emo metamorphics Cooling histories from exhumation	Worthing and Crawford (1996) Crowhurst et al. (1996)
	South-dipping subduction	~18-8 Ma	Maramuni Arc volcanics	Hill and Hall (2003), Page (1976)
	Halmahera Arc collision	~14 Ma	Compression in PNG Mobile Belt, apatite fission track geochronology	Hill and Raza, (1999), Kendrick (2000)
Lhasa	Onset of Neo-Tethyan subduction	~170 Ma (to 137 Ma)	Calc-alkaline granites and granitoids	Zhang et al. (2012)

	Onset of intra-oceanic subduction along Kohistan-Ladakh Arc	~154 Ma	Matum Das tonalite	Schaltegger et al. (2003)
	Subduction along Zedong Terrane	161.0 ± 2.3 Ma, ~156 Ma, 152.2 ± 3.3 Ma	Dacite breccia, Andesite dyke/breccia, Quartz diorite, Andesitic dyke	McDermid et al. (2002)
	Magmatic hiatus on Lhasa	~137 to 109 Ma, ~75 to 60 Ma	Magmatic gap in Gangdese Batholith	Ji et al. (2009), Wen et al. (2008), Chung et al. (2005)
	Initiation of Kohistan-Ladakh back-arc basin subduction along Lhasa	~109 Ma	Resumption of arc volcanism in Gangdese Batholith	Ji et al. (2009), Wen et al. (2008)
	Maximum southward position of Kohistan-Ladakh Arc	~100 Ma	Equatorial paleo-latitudes from mid- to Late Cretaceous red beds	Zaman and Torii (1999)
	Kohistan-Ladakh collision with Greater India	~60 to 50 Ma	Cessation of calc-alkaline magmatism, stratigraphic constraints of collision, slowdown in Indian Ocean seafloor spreading at ~52 Ma, change in arc magma chemistry by ~50 Ma	Khan et al. (2009) Hu et al. (2015) Cande et al. (2010) Bouilhol et al. (2013)

	Kohistan-Ladakh collision with Eurasia	~47 to 40 Ma	Slowdown in India-Africa seafloor spreading, Indian Ocean microplate formation, completion of Andean-style subduction (Linzizong), change in arc magma chemistry by ~40 Ma	Cande and Patriat (2015) Matthews et al. (2016) Chung et al. (2005) Bouilhol et al. (2013)
West Burma	Onset of Neo-Tethyan subduction	~163-152 Ma	Jadeite geochronology	Shi et al. (2008, 2014)
	Onset of Neo-Tethyan intra-oceanic subduction	~156-150 Ma	Biostratigraphic ages of cherts constraining age of Naga Ophiolite formation	Baxter et al. (2011)
	Subduction of Woyla back-arc basin	~113-110 Ma (Albian) ~105-90 Ma 95 ± 2 Ma	Albian unconformity on West Burma Wuntho-Popa Arc SSZ formation of Andaman Ophiolite	Morley (2012a) Mitchell et al. (2012) Pedersen et al. (2010)
Sumatra	Onset of Neo-Tethyan subduction	~170 Ma ~165-140 Ma (?)	Onset of arc volcanism in Sumatra segment Minor UHP/VHP metamorphism	McCourt et al. (1996) Parkinson et al. (1998)
	Subduction of Woyla back-arc basin	From ~115 Ma	Peak in UHP/VHP metamorphism in	Parkinson et al. (1998)

			Meratus and Luk Ulo sutures	
		~105-75 Ma	Wuntho-Popa Arc volcanism to the west, and Woyla intrusions	Mitchell et al. (2012), McCourt et al. (1996), Wajzer et al. (1991)
	Woyla Arc accretion	~75-62 Ma	Magmatic gap of arc volcanics on Sumatra	McCourt et al. (1996)
	Onset of Sunda subduction	62 Ma	Arc volcanism on Sumatra	McCourt et al. (1996)
West Java/ East Borneo	Onset of NeoTethyan subduction	~180-165 Ma	Schist in Meratus Complex	Wakita et al. (1998)
		~170 Ma (Bajocian)	Radiolarians	Wakita et al. (1998)
		~160 Ma	Zircon age spectra	Clements and Hall (2007)
	Late stage of Woyla/Barito back-arc basin subduction along Sunda continental margin	~100 Ma	Peak in zircon age spectra	Clements and Hall (2007)
		~100-93 Ma	Cenomanian/Turonian Meratus Ophiolite obduction	Pubellier et al. (2004), Yuwono et al. (1988)
	Suturing of East Java	~80 Ma	Stratigraphy, K–Ar and U–Pb of metamorphics, synthesis of previous studies	Wakita (2000) Clements and Hall (2011)
	Onset of Sunda	65 Ma	Subduction-related	Guntoro (1999), van

	subduction		rocks on Sulawesi	Leeuwen (1981)
Philippine Arc	Onset of south-dipping subduction along New Guinea (Sepik)	156.3 ± 2.0 Ma and 150.9 ± 3.3 Ma	SSZ ophiolitic crust from the Lagonoy Ophiolite	Encarnación (2004)
		142 ± 4 Ma	Ophiolite crystallisation from Gag Island, Halmahera	
	Continued arc volcanism	126 ± 3 Ma and 119 ± 2 Ma	SSZ volcanics from Cebu Island	Deng et al. (2015)
		99.9 ± 7.0 Ma	Ar-Ar age of the Calaguas Ophiolite	Geary et al. (1988), Geary and Kay (1989)
		100 ± 4 Ma	Arc rocks reported from Obi Island on Halmahera	Hall et al. (1995b)

2.2 Insights from seismic tomography

The distribution of ophiolites, intra-oceanic arc fragments and a complex network of sutures within southern Eurasia (Figs. 1-3), Southeast Asia and New Guinea preserve the remnants of oceanic basins that have been lost to subduction. Although the consumption of oceanic basins leaves physical evidence in the form of arc volcanics, accreted seamounts and ocean floor sediments, and ophiolites, the present-day mantle structure illuminated using seismic tomographic methods holds additional clues to the geodynamic evolution of these regions (Hafkenscheid et al., 2006; Replumaz et al., 2004; van der Voo et al., 1999a; van der Voo et al., 1999b).

As an estimate of the location of subduction through time, depth slices of fast seismic velocity anomalies are age-coded according to an assumption of vertical slab sinking with an average sinking rate. In this study we compare our revised plate reconstructions with the publicly-available P-wave seismic tomography depth slices from Li et al. (2008), assuming a sinking rate of 3 and 2 cm/yr in the upper and lower mantle, respectively, following Hafkenscheid et al. (2006). Hafkenscheid et al. (2006) also noted that the upper mantle sinking rates are likely to be similar to the convergence rate at the trenches, which may suggest even higher sinking rates for the circum-Tethyan region in the context of Australia’s 6-8 cm/yr, the Pacific’s ~8 cm/yr, and India’s ~5 cm/yr root mean square velocities since ~40 Ma (Zahirovic et al., 2015). To investigate, we test a higher end-member sinking rate of 8 cm/yr in the upper mantle, which is likely only meaningful for the Cenozoic as constrained by seafloor spreading histories and detailed hotspot tracks for the Pacific. The sinking rates applied in this study are significantly higher than the ~1.2-1.3 cm/yr whole-mantle average global slab sinking rates (Butterworth et al., 2014; van der Meer et al., 2010), with a similar slower and constant sinking rate scenario applied to age-coding of slabs in P- and S-wave seismic tomography in Zahirovic et al. (2014). However, a faster sinking rate, with differential rates in the upper and lower mantle, was found to better reproduce the evolution of major Tethyan and Southeast Asian subduction zones (Zahirovic et al., 2014). Importantly, we note that the assumption of vertical and temporally constant sinking rates along a single subduction zone, not to mention across a range of subduction zones in a region, is likely an oversimplification and requires testing using numerical simulations of mantle flow, as was carried out in Butterworth et al. (2014).

We compare our numerical mantle flow predictions to a number of P- and S-wave seismic tomography models, because tomographic models are typically constructed using a variety of methods, which incorporate different seismic phases and parameterisations (Grand, 2002; Romanowicz, 2008). P-wave models tend to have higher resolutions than S-wave models, due to the limited number of S-wave phases that can be used in seismic tomographic inversions (Widiyantoro et al., 1998). Beyond the inherent higher resolution of P-wave models in well-sampled continental regions, the Li et al. (2008) global seismic tomography model has additional coverage by incorporating coverage using the Chinese Seismographic Network, leading to a better sampling of the Tethyan and Asian mantle.

The limitation of P-wave models is that they tend to bias their sampling of the mantle beneath continental crust, leading to lower seismic velocity anomaly amplitudes in oceanic regions. For example, a subducted slab that may straddle oceanic and continental regions (such as the Tethyan slabs) may appear "faded" beneath the oceanic regions. Although S-wave models tend to have lower resolution, they offer more equal sampling of the mantle beneath oceans and continents (Grand, 2002). Due to the lack of permanent seismic stations in the oceans (except for some stations located on islands) and over Antarctica, the coverage and sampling for both P- and S-wave seismic tomography models is poorer for the southern hemisphere and all oceanic regions (Romanowicz, 2008). As regional tomographic models can have edge artefacts (Foulger et al., 2013), and typically are not represented as seismic velocity anomalies with respect to the global mantle, we focus on using only global tomographic models in our comparisons.

2.3 Coupled plate reconstructions and mantle convection numerical models

To better understand the geodynamic implications of the plate reconstructions, and go beyond the assumption of constant and vertical slab sinking used in simple interpretations of mantle structure, we couple the plate kinematics to computations of mantle flow. We use the mantle convection modelling code CitcomS (Zhong et al., 2000) (https://geodynamics.org/cig/software/citcoms/), modified to progressively assimilate surface plate velocities, the thermal structure of the lithosphere and the shallow thermal structure of subducting slabs (Bower et al., 2015) from our plate motion model (Fig. 7a). The temperature and thickness of the lithosphere is derived using a half-space cooling model and the synthetic age of the ocean floor. Slabs are assimilated into the mantle to a depth of 350 km but convection is entirely dynamic away from slabs and below the lithosphere. We computed numerical models from 230 Ma to the present to capture the post-Pangea mantle evolution, with global plate reconstructions of the pre-Late Jurassic described in Müller et al. (2016). However, we analyse the mantle evolution since the latest Jurassic (~160 Ma) for which time the plate reconstructions are regionally refined, and the mantle flow models have reached a dynamic equilibrium from the synthetic initial condition (Flament et al., 2014). Initially at 230 Ma, slabs are inserted down to 1400 km depth, with a 45º dip down to 425 km and 90º below 425 km. In the initial conditions, slabs are twice as thick in the lower mantle than in the upper mantle to account for advective thickening observed in tests in which slabs are only initially inserted in the upper mantle. The initial condition includes a basal thermochemical layer 113 km thick just above the core–mantle boundary (CMB) that consists of material 3.6% denser than ambient mantle. This condition suppresses the formation of plumes, but does not impede the formation of large-scale mantle upwellings. The surface and CMB are isothermal at 273 K and 3100 K, respectively (Fig. 7b). Subduction zones that appear (initiate) during the model run are progressively inserted as slabs in the uppermost mantle (Bower et al., 2015). The kinematic boundary conditions, generated in GPlates, and the thermal volume conditions for the lithosphere and shallow subduction, are assimilated in 1 Myr intervals, as described in Bower et al. (2015). The average model resolution, obtained with ~13 × 10^6 nodes and radial mesh refinement, is ~50 × 50 × 15 km at the surface, ~28 × 28 × 27 km at the core–mantle boundary (CMB), and ~40 × 40 × 100 km in the mid-mantle.

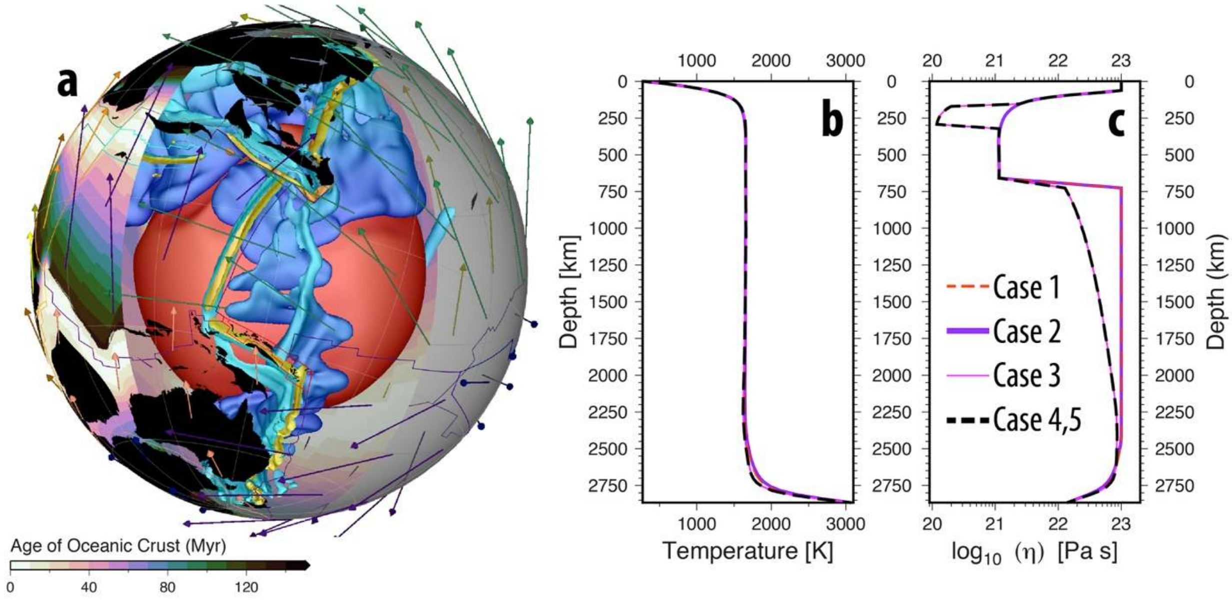

Fig. 7. a) Assimilation of plate velocities (arrows), subduction zones (yellow) and thermal lithospheric thickness based on the age of the seafloor from plate reconstructions in GPlates, with 3D volume contours of dynamic cold slabs (blue) and hotter upwellings (red) predicted by the mantle convection code CitcomS (here using Case 5, see Section 2.3 and Table 5). Reconstructed present-day coastlines (black) are provided as a reference. Slab colouring is a function of depth from light blue (shallow mantle) to darker blue (deep mantle). Seafloor age-grid is applied globally, but is cut out in this schematic to highlight the internal mantle structure. b) Horizontally-averaged present-day mantle temperature, and c) present-day average viscosity in the five numerical cases (see Section 2.3, and Table 5).

The vigour of mantle convection is defined by the Rayleigh number, Ra, where:

$$Ra = \frac{\alpha_0 \rho_0 g_0 \Delta T h_M^3}{\kappa_0 \eta_0},$$

in which α is the coefficient of thermal expansion, ρ the density, g the acceleration of gravity, ΔT the temperature difference between surface and CMB, h_M the thickness of the mantle, κ the thermal diffusivity, and η the viscosity; the subscript "0" indicates reference values (see Table 4). The viscosity of the slabs and mantle are stress- and temperature-dependent, following

$$\eta = \eta_0(r)\exp\left(\frac{E_\eta}{R(T+T_\eta)} - \frac{E_\eta}{R(T_b+T_\eta)}\right)$$

where $\eta_0(r)$ is a depth-dependent pre-factor defined with respect to the reference viscosity, η_0, E_η is the dimensional activation energy (E_{UM} in the upper mantle and E_{LM} in the lower mantle), R is the universal gas constant, T is the temperature, T_η is a temperature offset, and T_b is the ambient mantle temperature outside the thermal lithosphere, slabs or the basal thermal boundary layer (see Table 4). Although the viscosity of the upper mantle can be estimated in studies of post-glacial rebound (Fjeldskaar et al., 2000; Gasperini and Sabadini, 1989; Lambeck et al., 1998), the viscosity of the lower mantle is less well constrained, which has resulted in a wide range of proposed viscosity profiles. Previous approaches have argued for a factor of 10 increase in viscosity between the upper and lower mantle (Paulson et al., 2007), while others have argued for a factor of 30 (Hager, 1984) or 100 (Forte and Mitrovica, 1996; Steinberger and Calderwood, 2006). We vary the viscosity profile (Fig. 7c, Table 5) with cases 1 to 4 based on the plate reconstructions from Zahirovic et al. (2014) and Gibbons et al. (2015), and a fifth case based on the refined plate reconstructions presented in this study. The viscosity of the lower mantle in each cases is either 100 times more viscous than the upper mantle, or increases gradually with depth from a factor of 10 at the base of the transition zone (660 km) to a maximum of 100 in the lowermost mantle (Steinberger and Calderwood, 2006). Cases 3, 4 and 5 also incorporate a low-viscosity asthenosphere, which has been suggested to be an important decoupling layer that enables the elevated velocities of typically oceanic plates (Becker, 2006; Debayle and Ricard, 2013). Since paleo-longitudes are less well constrained earlier than ~100 Ma, we incorporate the van der Meer et al. (2010) subduction reference frame and their time-dependent longitudinal shift into Case 3. By using a variety of radial viscosity profiles, different absolute reference frames, and plate reconstructions between the five cases, allows us to capture some of the uncertainties involved in our approach of modelling deep-time plate reconstructions and mantle convection, and help test end-member plate reconstructions of the Tethyan region.

The time-dependent mantle structure is presented in 3D visualisations made with GPlates and as a series of vertical cross-sections that are reconstructed with the overriding plate to capture the evolution of subduction, plotted using Generic Mapping Tools (Wessel et al., 2013). The predicted present-day mantle structure is qualitatively compared to equivalent slices of P- and S-wave seismic tomography models, where fast seismic velocity anomalies are compared to slab contours (10% colder than ambient mantle, representing temperatures colder than ~1270°C) from the mantle convection models.

Table 4. Parameters common to all model cases. Subscript “0” denotes reference values.

Parameter	**Symbol**	**Value**	**Units**
Rayleigh Number	Ra	7.84×10^{7}	
Thermal expansion coefficient	α_0	3×10^{-5}	K^{-1}
Density	ρ_0	4000	$kg\ m^{-3}$
Gravity acceleration	g_0	9.81	$m\ s^{-2}$
Temperature change	ΔT	2825	K
Temperature offset	T_η	452	K
Background mantle temperature	T_b	1685	K
Mantle thickness	h_M	2867	km
Thermal diffusivity	κ_0	1×10^{-6}	$m^2\ s^{-1}$
Reference viscosity	η_0	1×10^{21}	Pa s
Activation energy (upper mantle)	E_{UM}	100	$kJ\ mol^{-1}$
Activation energy (lower mantle)	E_{LM}	33	$kJ\ mol^{-1}$
Activation temperature	T_η	452	K
Universal gas constant	R	8.31	$J\ mol^{-1}\ K^{-1}$

Radius of the Earth	R_0	6371	km

Table 5. Model set-up for Case 1 – 5. Also refer to Fig. 7b-c.

<table>
<tr><th></th><th>Case 1</th><th>Case 2</th><th>Case 3</th><th>Case 4</th><th>Case 5</th></tr>
<tr><td>Mesh nodes</td><td colspan="5">129 × 129 × 12 (nodes on the surface)
× 65 (depth levels)</td></tr>
<tr><td>Viscosity relative to Reference Viscosity (Lithosphere 0-160 km depth, Asthenosphere 160-310 km depth, Upper mantle 310-660 km depth, Lower Mantle > 660 km depth)</td><td>1,0.1,1,100</td><td>1,1,1,100</td><td colspan="3">1,0.1,1,10→100
linear increase of viscosity from 10 to 100 with depth in the lower mantle to approximate the viscosity profile of Steinberger and Calderwood (2006)</td></tr>
<tr><td>Plate reconstructions for the eastern Tethys</td><td colspan="2">Zahirovic et al. (2014)</td><td>Slab-calibrated longitudinal positions from Zahirovic et al. (2014)</td><td>Zahirovic et al. (2014)</td><td>This Study</td></tr>
</table>

3 Regional tectonic evolution

3.1 Late Jurassic plate boundary configuration and rifting mechanism along northern Gondwana

The Late Jurassic is marked by a major rifting event along northern Gondwana (Pigram and Panggabean, 1984) (Figs. 4a and 8), which transferred a number of continental blocks (including East Java, West Sulawesi, Mangkalihat and east Borneo) northward towards Eurasia (Hall, 2012;

Zahirovic et al., 2014), with the only portions of the seafloor spreading system preserved in the Argo Abyssal Plain on the NW Australian shelf (Gibbons et al., 2013). Beyond the oldest preserved oceanic crust, the plate configuration can only be inferred from proxy indicators found on continents. One pertinent argument is that Audley-Charles (1988) and Audley-Charles et al. (1988) required a triple junction plate boundary configuration in the Late Jurassic and Early Cretaceous in the vicinity of the NW Australian shelf where northward slab pull from subduction along southern Eurasia was the driving mechanism for detaching the Neo-Tethyan ribbon terrane (also in Fig. 4a). This northward slab pull detached the continental fragments forming passive margins along the northern and southern boundaries of the ribbon terrane, the preferred scenario presented here. An alternative scenario has south-dipping subduction along northern Gondwana in the Late Jurassic, leading to the opening of the Neo-Tethys and transfer of continental fragments northward through slab rollback (Hall, 2012) (Fig. 4b). Both mechanisms are thought to be capable of detaching continental fragments (see Stampfli and Borel, 2002), but, the south-dipping subduction end-member requires continuous arc volcanism, some of which ought to be preserved on the drifting ribbon terranes.

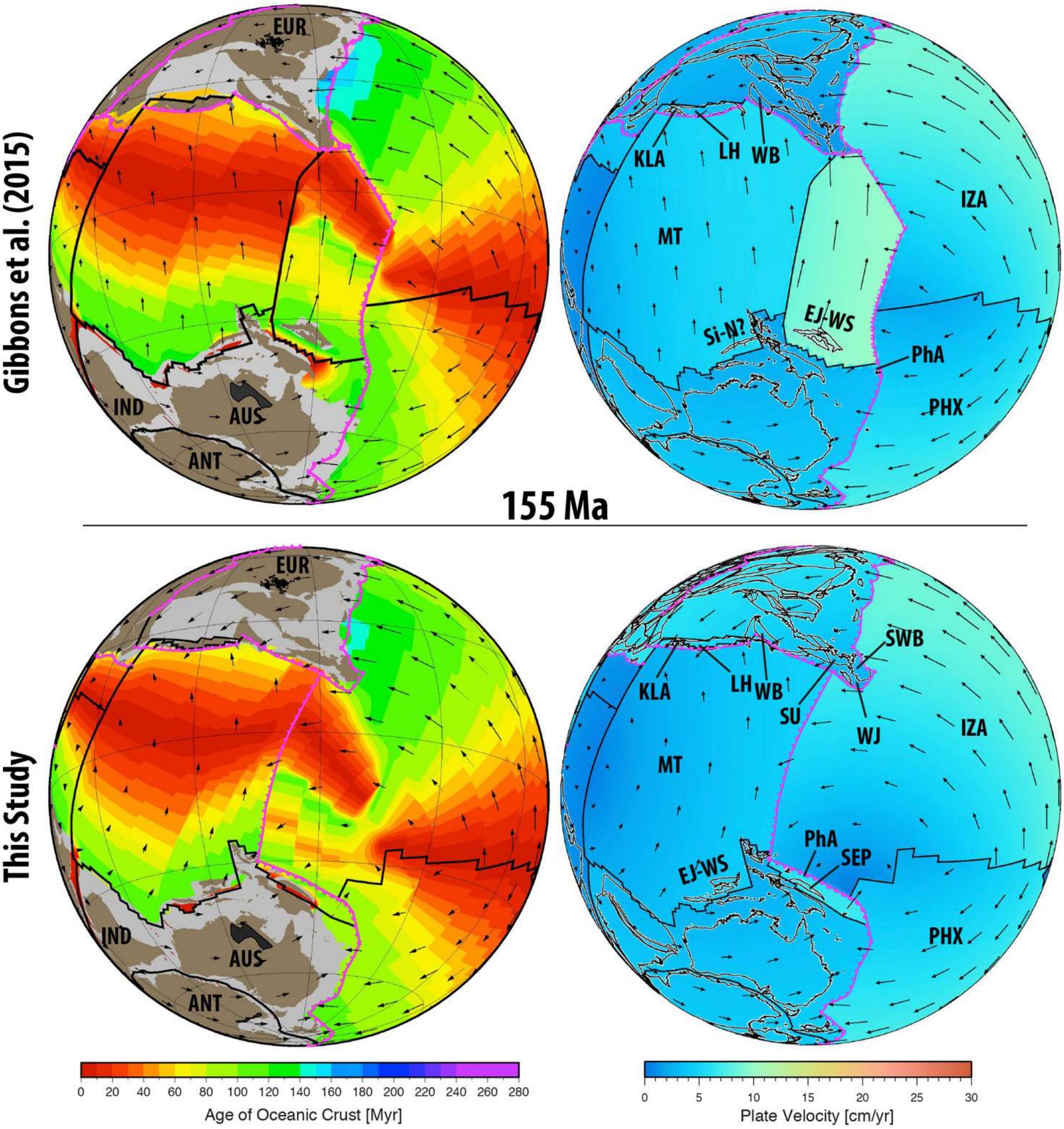

Fig. 8. Reconstruction of Neo-Tethyan ocean basin opening along northern Gondwana in the latest Jurassic. Both the Zahirovic et al. (2014) and Gibbons et al. (2015) models (top) invoke East Java and West Sulawesi rifting from New Guinea, as the simplest tectonic scenario to transfer the blocks northwards toward Southeast Asia, and a possible origin of the Sikuleh and Natal (Si-N) fragments from the Argo Abyssal Plain. Southwest Borneo (SWB), West Java (WJ), Sumatra (SU), West Burma (WB) and Lhasa (LH) form the active Eurasian continental margin. Seafloor spreading in the Neo-Tethys is driven by north-dipping subduction along southern Eurasia (EUR), consuming

the Meso-Tethyan (MT) oceanic crust and resulting in the incipient formation of the Kohistan-Ladakh Arc (KLA). In the revised reconstructions, East Java and West Sulawesi (EJ-WS) are the 'Argoland' continental fragment originating in the Argo Abyssal Plain on the NW Australian shelf. South-dipping subduction along New Guinea is modelled to detach the Sepik Terrane (SEP) from the New Guinea margin through slab rollback, generating the embryonic components of the Philippine Archipelago (PhA). The scale bars are relevant to all plate reconstruction figures. Grey regions represent the extent of continental crust, dark grey represents large igneous provinces and other plume products, and thin brown lines represent reconstructed fracture zones. ANT – Antarctica, AUS – Australia, IND – India, IZA – Izanagi Plate, PHX – Phoenix Plate. Orthographic reconstructions are centred on 115°E, 15°S. See Supplementary Animation 2, 3 and 4.

The NW Australian shelf, the putative source of the Argoland ribbon terrane records some Late Jurassic and Early Cretaceous volcanic plateaus (e.g., Scott and Wombat plateaux, Joey Rise, etc. – see Fig. 1), rift-related volcanics and seaward dipping reflectors (Heine and Müller, 2005; Rohrman, 2015; von Rad et al., 1992). An earlier phase of rhyolitic volcanism between ~213 and 190 Ma erupted on the Wombat Plateau (von Rad and Exon, 1983; von Rad et al., 1992), but cannot be temporally linked to the latest Jurassic (~155 Ma) rifting and seafloor spreading phase recorded on the NW Australian shelf. Although the latest Jurassic NW Australian margin was volcanic, little evidence exists that it was dominated by an Andean-style active margin (von Rad and Exon, 1983; von Rad et al., 1992). Although the seismic interpretations by Hopper et al. (1992) of the margin's volcanic history do not indicate widespread plume activity, the recent work of Rohrman (2015) suggests a plume origin for the large volume of underplated material and widespread sills interpreted from seismic sections in the Exmouth Plateau region (Fig. 1). One critical aspect of the latest Jurassic event is that the onset of seafloor spreading is well-constrained by a 155 ± 3.4 Ma K-Ar age of the oldest seafloor in the Argo Abyssal Plain (Gradstein and Ludden, 1992), consistent with rapid tectonic subsidence in the latest Jurassic on the NW Australian shelf (Heine and Müller, 2005; Rohrman, 2015; Tovaglieri and George, 2014), and the identification of M25A (Heine and Müller, 2005) or M26 (Gibbons et al., 2012) as the oldest magnetic anomalies (~153-155 Ma) in the seafloor spreading record.

Due to the lack of latest Jurassic arc volcanics on the NW Australian Shelf, together with strong indicators of north-dipping subduction initiation along southern Eurasia (see following sections), we prefer northward slab pull as the driving mechanism for rifting and seafloor spreading to open the Neo-Tethys from ~155 Ma. Although more work is required to test whether a plume model can explain the volcanism on the NW Australian shelf in the latest Jurassic (Rohrman, 2015), such a scenario would be consistent with the triple junction scenario invoked for this region (Audley-Charles, 1988; Audley-Charles et al., 1988; Gibbons et al., 2015; Gibbons et al., 2012; Zahirovic et al., 2014), and the similarity to the East African rift-plume interaction (Burke and Dewey, 1973; Montelli et al., 2006; Yirgu et al., 2006). Since the Neo-Tethyan seafloor spreading history is incomplete, it remains difficult to ascertain which continental blocks rifted from the Argo segment of the Australian margin (Table 2). Rifting of East Java and West Sulawesi from New Guinea was invoked as a preferred scenario in our base models in the Late Jurassic (Gibbons et al., 2015; Zahirovic et al., 2014), with the possibility that micro-continental fragments along Sumatra (such as the now-disputed Natal and Sikuleh fragments, Fig. 2) had an origin in the Argo Abyssal Plain, following Audley-Charles et al. (1988), Metcalfe (1994) and Heine and Müller (2005). However, recent zircon age spectra analyses from East Java suggesting strong affinities with the NW Australian Shelf (Sevastjanova et al., 2015; Smyth et al., 2007), led Hall (2012) to argue that East Java was the enigmatic “Argoland” fragment (Table 2). We present both a NW Australian shelf and a New Guinea origin for Argoland in our alternative plate reconstruction scenarios, and evaluate their plate kinematic and geodynamic consequences on the Neo-Tethyan tectonic evolution.

Table 2. Previously-interpreted continental fragments originating from northern Gondwana in the eastern Tethys in the Late Jurassic.

Model	***All continental fragments originating from northern Gondwana in the Late Jurassic eastern Tethys***	***Argoland fragment***
Audley-Charles et al. (1988)	South Tibet (Lhasa), West Burma, Malaya, Sumatra, East and West Borneo fragments, West Sulawesi	West Burma
Metcalfe (1994)	West Burma, Sikuleh, Natal, West Sulawesi, Mangkalihat, Banda Allochthons	West Burma
Heine and Müller (2005)	West Burma	West Burma
Hall (2012)	Southwest Borneo core, East Java, West Sulawesi	East Java and West Sulawesi
Zahirovic et al. (2014) Gibbons et al. (2015)	Sikuleh, Natal, East Java, Southeast Borneo, West Sulawesi	Sikuleh, Natal and other possible fragments that may be in the Mawgyi Nappe along West Burma
This Study	East Java, Eastern Borneo, Mangkalihat, West Sulawesi and Sepik (New Guinea)	East Java, Eastern Borneo, Mangkalihat and West Sulawesi

The ~155 Ma onset of seafloor spreading in the Argo segment of the north Gondwana margin is consistent with Jurassic sedimentary rift-drift sequences (Pigram and Panggabean, 1984), and mafic rocks that are as old as 158 Ma on West Sulawesi (Polvé et al., 1997), likely representing the drift of the East Java and West Sulawesi continental fragments (Zahirovic et al., 2014). The early seafloor spreading history is preserved in the Argo Abyssal Plain, with the youngest marine magnetic anomaly of M10Ny (Gibbons et al., 2013) representing an age of 128.9 Ma, after which the seafloor spreading history is unconstrained. As discussed extensively in Zahirovic et al. (2014), and summarised below in Section 3.4, the East Java and West Sulawesi fragments may have collided with an intra-oceanic arc in the mid-Cretaceous (Wakita, 2000), and sutured to Sundaland by 80 Ma. However, the Neo-Tethyan full seafloor spreading velocity required by the ~115 Ma arc-continent collision approaches ~25 cm/yr between ~128 and 115 Ma (see Supplementary Fig. 1), which is likely to be an upper limit for plate velocities in the post-Pangea timeframe (Stampfli and Borel, 2002; Vérard et al., 2012; Zahirovic et al., 2015). An alternative explanation for the ~115 Ma peak in very high pressure (VHP) metamorphic rocks in the Luk Ulo-Meratus Suture Zone (Parkinson et al., 1998) (Fig. 2), includes the initiation of subduction of the Woyla/Barito back-arc basins, which reduces the synthetic seafloor spreading velocities to ~11 cm/yr (Figs. 10-11). We adopt the latter option that does not introduce a geodynamically implausible velocity spike in Tethyan plate velocities.

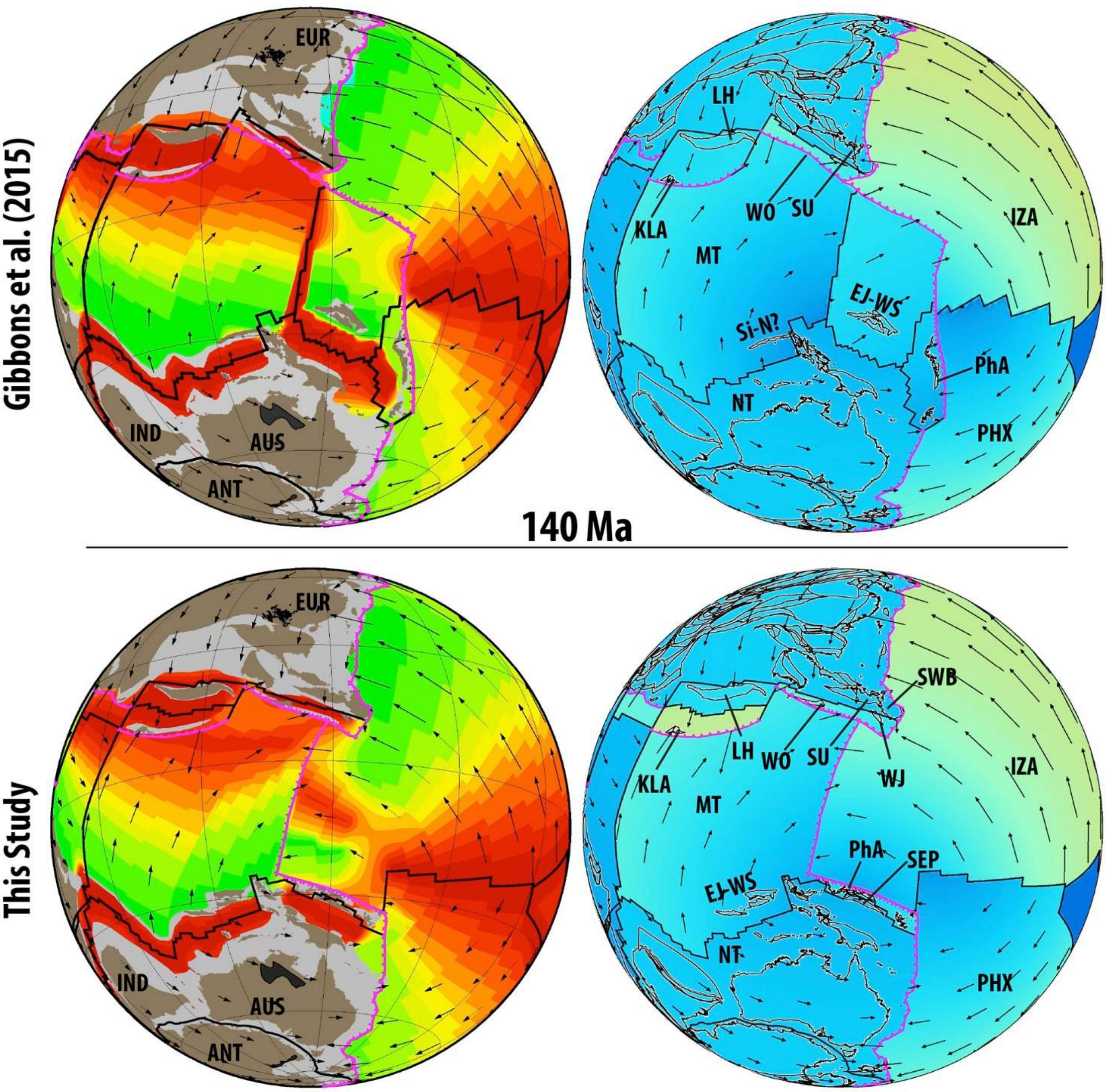

Fig. 9. Ongoing Meso-Tethyan subduction leads to the opening of the Kohistan-Ladakh (KLA) and Woyla (WO) back-arc basins, as well as the Neo-Tethys (NT) along northern Gondwana. The Philippine Archipelago (PhA) is detached from the Sepik Terrane (SEP) through a northward ridge jump and continued rollback of the Izanagi slab.

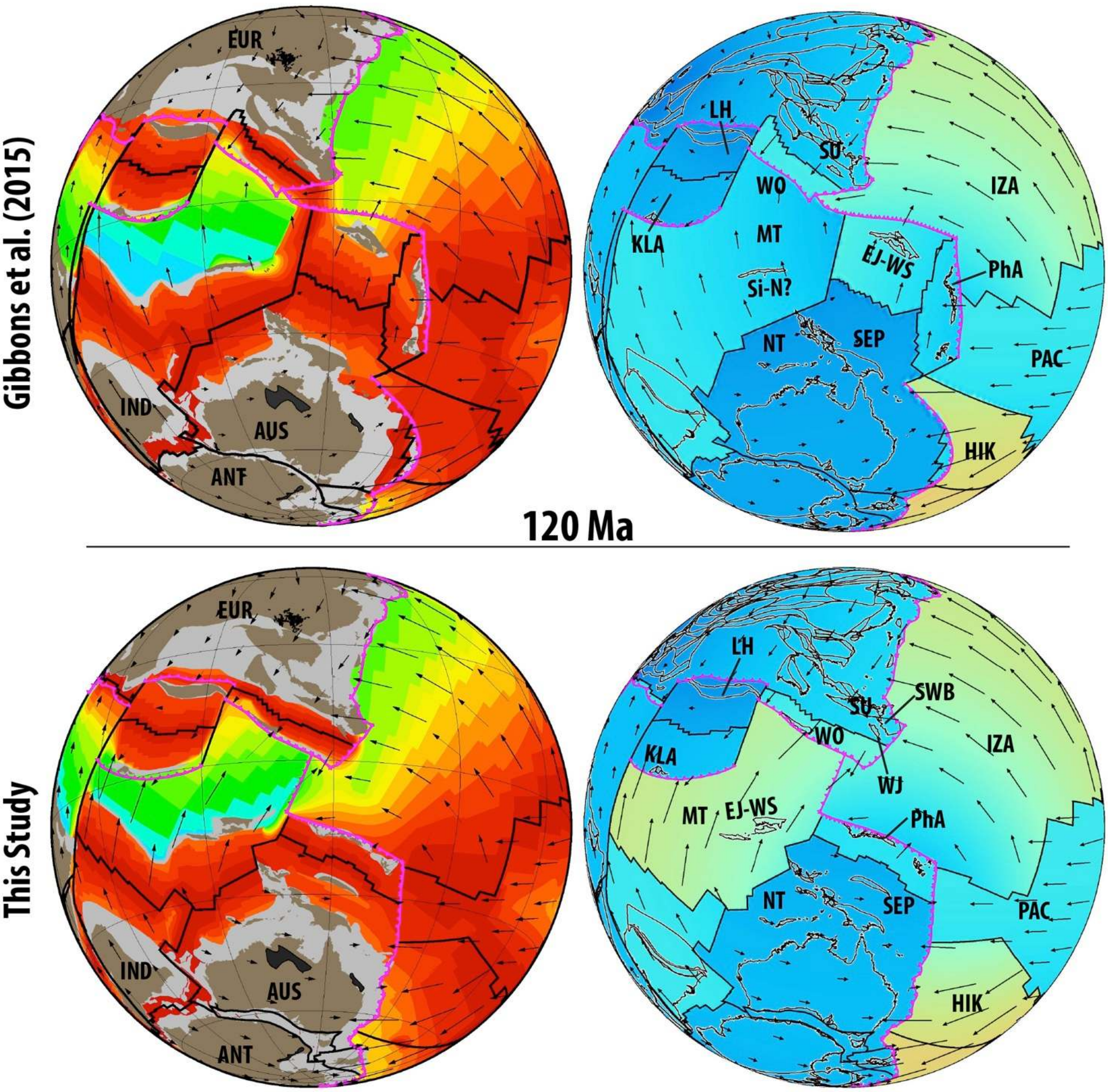

Fig. 10. In the models of Zahirovic et al. (2014) and Gibbons et al. (2015) the Neo-Tethyan ridge system is abandoned by 120 Ma. In this study, seafloor spreading continues because of ongoing subduction of the Meso-Tethyan Plate. Although the oldest preserved seafloor spreading constraints

for the Neo-Tethys are ~128 Ma in the Argo Abyssal Plain, we impose continued seafloor spreading in the Neo-Tethys that is driven by northward slab pull. HIK – Hikurangi Plate.

3.2 Active margin evolution in the Lhasa segment

The Late Jurassic is characterized by the asynchronous development of an active margin along southern Eurasia that was consuming the Meso-Tethyan ocean basin along a north-dipping subduction zone (Fig. 8). Along southern Lhasa, subduction-related calc-alkaline granites and granitoids ranging in age from ~170 to 137 Ma indicate the onset of north-dipping Meso-Tethyan Andean-style subduction (Zhang et al., 2012). The onset of subduction-related magmatism within the Kohistan Arc, which has no continental basement (Burg, 2011; Jagoutz and Schmidt, 2012), at ~154 Ma in the form of the Matum Das tonalite (Schaltegger et al., 2003) suggests the rollback of a Meso-Tethyan slab (e.g., Petterson and Windley, 1985; Pudsey, 1986) and the possible origin of Kohistan and Ladakh as fore-arc oceanic crust, following the generic model of forearc formation proposed by Flower and Dilek (2003) and Stern et al. (2012). Further east along the present-day suture zone, the interpreted intra-oceanic Zedong Terrane records latest Jurassic ages of subduction-related igneous suites including a dacite breccia dated as 161.0 ± 2.3 Ma, a number of samples with an age of ~156 Ma (andesite dyke, andesite breccia and quartz diorite), and an andesitic dyke with an age of 152.2 ± 3.3 Ma (McDermid et al., 2002). The Kohistan-Ladakh, Zedong and more-broadly Neo-Tethyan intra-oceanic subduction zone likely became established through continued southward slab rollback between ~150 and 120 Ma, which is marked by a magmatic hiatus in the Gangdese Batholith on Lhasa until ~110 Ma (Ji et al., 2009; Wen et al., 2008b). However, while the magmatic evolution of the Kohistan-Ladakh intra-oceanic arc is well-studied, its paleo-latitudinal position remains controversial and poorly constrained. Burg (2011) and Gibbons et al. (2015) place Kohistan on the equator at ~100 Ma (Fig. 11) based on the magnetisation of mid-Late Cretaceous red beds (Zaman and Torii, 1999), suggesting a maximum southward extent for the Neo-Tethyan intra-oceanic arc. In addition, although Kohistan and Ladakh form the only significant preserved remnants of the Early Cretaceous intra-oceanic arc within the Yarlung-Tsangpo Suture Zone, additional ophiolites with intra-oceanic affinity are embedded in the suture zone east of Kohistan and Ladakh (Aitchison et al., 2000; Hébert et al., 2012).

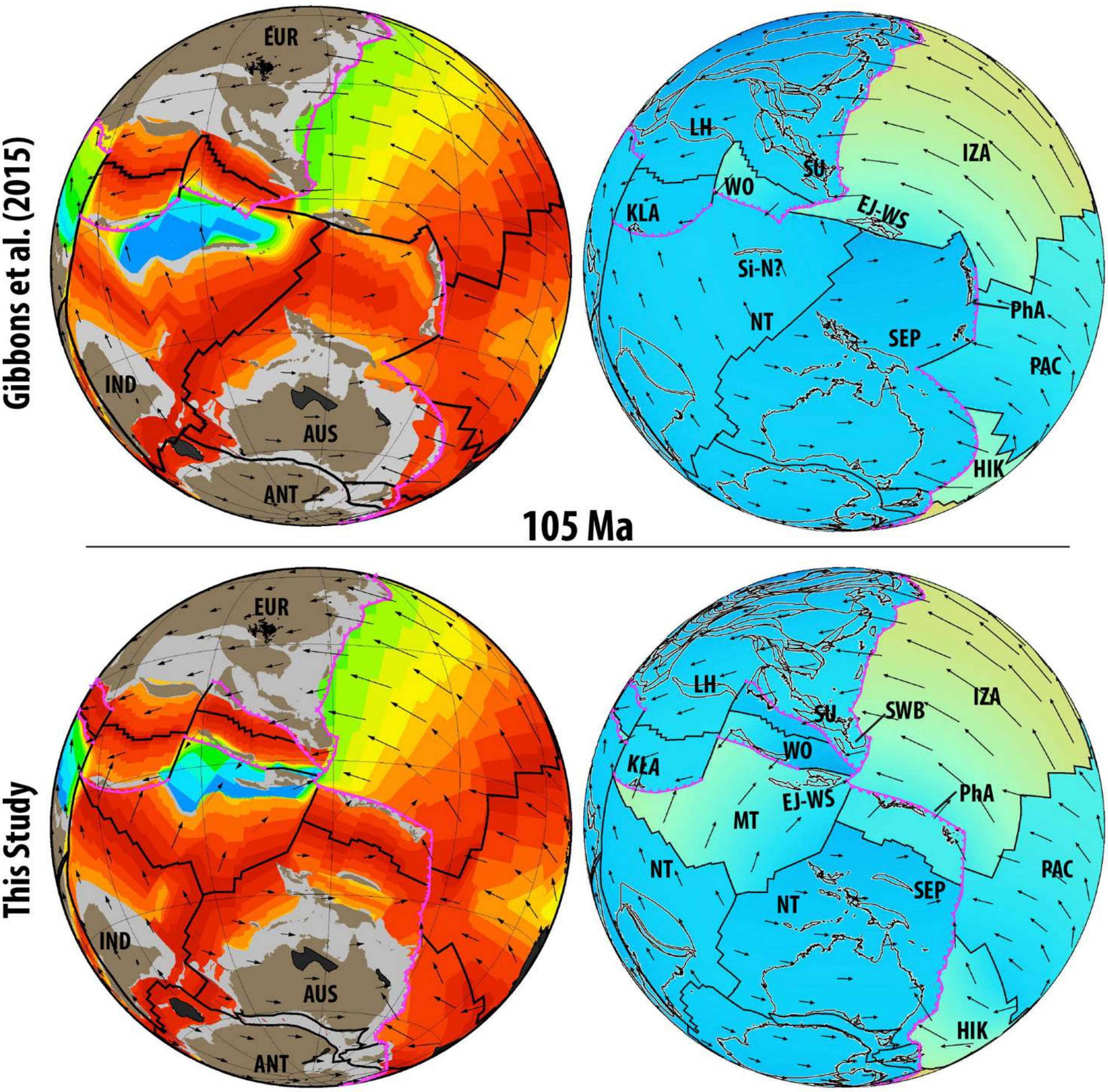

Fig. 11. The change in the motion of India from largely counterclockwise in the Early Cretaceous to largely northward, is recorded in fracture zone bends in the Wharton Basin at ~105-100 Ma (Matthews et al., 2012). In this study, the Neo-Tethyan ridge is consumed at the Kohistan-Ladakh

(KLA) intra-oceanic subduction zone from ~105 Ma, leading to a greater northward slab pull acting on the Indian Plate (IND), which we interpret as causing the change in India Plate motion. The collision of the East Java-West Sulawesi continental fragments possibly impeded subduction at the Woyla Arc at ~105 Ma, and led to obduction of the Meratus ophiolite in the Cenomanian-Turonian (~100-93 Ma) (Pubellier et al., 2004; Yuwono et al., 1988). The Kohistan-Ladakh and Woyla arcs likely occupied near-equatorial latitudes by ~100 Ma, with both back-arc systems subducted northward from ~115 Ma in the Sunda segment and from ~110 Ma along Lhasa, resulting in two coeval north-dipping subduction zones in the Neo-Tethys (see Section 3.2).

The Aptian to Albian (~126-100 Ma) Yasin Group sedimentary sequence on the Kohistan-Ladakh Arc are intercalated with syn-tectonic arc volcanics, and subsequent intrusions of diorites and granodiorites (Pudsey et al., 1985; Rehman et al., 2011). A large portion of the magmatic products were emplaced during a key timeframe between ~110 and 90 Ma (Petterson and Windley, 1985; Rehman et al., 2011; Schärer et al., 1984), with significant magmatic accretion forming the Sapat Complex on the Kohistan Arc between ~105 and 99 Ma (Bouilhol et al., 2011). This short-lived "trenchward migration of the hot mantle source" (Bouilhol et al., 2011) may indicate the arrival of the Neo-Tethyan mid-oceanic ridge with slab window formation, consistent with the ~95 Ma high-temperature granulite metamorphism in the Jijal Complex and the peak metamorphic event in the Kamila Amphibolite unit (Petterson, 2010). The demise of the Neo-Tethyan mid-oceanic ridge between ~110 and 90 Ma along the Kohistan-Ladakh intra-oceanic arc (Fig. 11) likely had substantial geodynamic implications for the region at the time, resulting in stronger northward slab pull acting on the Indian Plate. Although the slab window likely temporarily impeded subduction, the progressively increasing northward slab pull from ~105 Ma likely resulted in a major change in the direction and speed of the Indian Plate, observable in the significant bends observed in the Wharton Basin fracture zones (e.g., Wallaby-Zenith Fracture Zone) at ~105-100 Ma that required a ~50° clockwise reorientation of the Indo-Australian spreading system (Matthews et al., 2012), and possibly triggered a regional plate-reorganization event (Matthews et al., 2011). In addition, the onset of northward slab pull on the Indian Plate from ~100 Ma may explain the paleomagnetic observations of rifting within Greater India in the Late Cretaceous (van Hinsbergen et al., 2012).

The intersection of the Neo-Tethyan ridge with the Kohistan-Ladakh intra-oceanic subduction zone, as modelled in this study, may have temporarily interrupted subduction due to the decrease of negatively buoyant oceanic lithosphere entering the trench, with convergence accommodated along the active continental Eurasian margin from ~110 Ma rather than along the intra-oceanic subduction zone. The Gangdese Batholith recorded a major pulse of granitic magmatism from ~109 to 80 Ma (Ji et al., 2009), indicating the resumption of Andean-style subduction along Lhasa that is contemporaneous with intra-oceanic subduction along Kohistan-Ladakh, and hence signifies the onset of two simultaneous north-dipping subduction zones in the Neo-Tethys from ~110 Ma.

The Late Cretaceous evolution of the Kohistan-Ladakh Arc has conflicting interpretations. Conventional models suggest a Late Cretaceous collision and suturing of Kohistan-Ladakh to Eurasia (Clift et al., 2002; Debon et al., 1987; Treloar et al., 1996), whereas more recent works, which have incorporated detailed geochronology and structural interpretations of Kohistan, have concluded that instead of suturing to Eurasia, the Late Cretaceous is punctuated by an arc rifting and splitting episode by ~85 Ma (Bouilhol et al., 2011; Burg, 2011; Burg et al., 2006), which suggests Neo-Tethyan slab rollback rather than collisional processes. The ~75-60 Ma magmatic gap in the Gangdese Batholith (Chung et al., 2005; Wen et al., 2008b) may imply that the majority of India-Eurasia convergence was accommodated by subduction along Kohistan-Ladakh rather than by subduction along Lhasa. A scenario that precludes Kohistan-Ladakh collision with Eurasia in the Late Cretaceous requires that an intra-oceanic arc first accreted onto Greater India (Chatterjee et al., 2013). A similar model proposes that the Muslim Bagh Ophiolite represents the Kohistan-Ladakh forearc and was obducted onto the leading edge of Greater India at ~65 Ma in near-equatorial latitudes, which resulted in the cessation of calc-alkaline magmatism on Kohistan-Ladakh during ~65-61 Ma (Khan et al., 2009). This scenario requires the accompanying suturing between Kohistan-Ladakh and Greater India along the Indus Suture to occur earlier than the closure of Shyok Suture. Such a scenario is consistent with the plate reconstructions presented in this study, in which Greater India reaches equatorial latitudes at ~65 Ma. A recent stratigraphic analysis presented in Hu et al. (2015) suggests the India-Eurasia collision was underway by 59 ± 1 Ma, which we interpret as the initial arc-continent collision, consistent with the tectonic evolution of the Kohistan-Ladakh Arc. The major slowdown in spreading across the Central and Southeast Indian Ridges at ~52 Ma (Chron 23o, Cande et al., 2010) may indicate the complete abandonment of the intra-oceanic subduction zone, and the completion of the initial arc-continent collision between Greater India and the Neo-Tethyan intra-oceanic arc. Recent geochemical analyses of granitoids from the Kohistan-Ladakh Arc indicate a major change in magma chemistry (Nd and Hf isotopes) and the arrival of the Greater Indian continental margin into the subduction zone by 50.2 ± 1.5 Ma (Bouilhol et al., 2013), which is consistent with the cessation of intra-oceanic subduction by this time.

The continent-continent collision between Greater India and Eurasia likely occurred at ~47 Ma, recorded in the marked slowdown of seafloor spreading at Chron 21o along the Southeast Indian Ridge (Cande and Patriat, 2015) and the contemporaneous formation of an Indian Ocean microplate near the Ninetyeast Ridge (Matthews et al., 2016). Suturing along the Shyok Suture Zone between the two continents was likely complete by 40.4 ± 1.3 Ma (Bouilhol et al., 2013), which accounts for the ~60-40 Ma Andean-style emplacement of the Linzizong Volcanics in Lhasa (Chung et al., 2005). The ~47-40 Ma continent-continent collision timing is consistent with an additional slowdown and change in spreading direction along the Central and Southeast Indian Ridges, the inception of a short-lived Indian Ocean microplate (Matthews et al., 2016), and the abandonment of the Wharton Ridge sometime between ~43 and 36 Ma (see discussion in Gibbons et al., 2015).

3.3 Convergence along the West Burma and Sumatra margin segment

Eastern Sumatra, as part of the Sibumasu ribbon terrane, docked with the Eurasian margin sometime in the Late Triassic to Early Jurassic (Metcalfe, 2011), and has since recorded Meso- and Neo-Tethyan subduction and accretion histories. The Woyla Terrane, which accreted to Sumatra in the Late Cretaceous (Morley, 2012a), plays a key role in elucidating the geodynamic setting of the Sumatran active margin since the Late Jurassic. However, the nature of the Woyla Terrane crust and the subduction polarity and history has given rise to a number of competing models for the tectonic evolution of Sumatra. The Jurassic to Cretaceous Woyla Group of sedimentary and volcanic units has previously been interpreted as an arc built on re-worked continental basement (Barber and Crow, 2003; Cameron et al., 1980), largely due to the presence of a tin geochemical signature in the Sikuleh granitoids (Fig. 2) that may have been interpreted as analogous to the Southeast Asian tin belts that were built on continental crust (e.g., Bangka and Billiton Islands; Searle et al., 2012). Parts of the Woyla basement near Sikuleh, Natal and Bengkulu are composed of quartzite and phyllite (Acharyya, 1998), and overprinted by widespread granitoid intrusions largely Late Cretaceous in age (Barber and Crow, 2003), leading some authors to interpret these micro-blocks as Gondwana-derived continental fragments (Görür and Sengör, 1992; Haile, 1979; Metcalfe, 1994; Metcalfe, 2002; Metcalfe and Irving, 1990). The paleomagnetic study of a Jurassic limestone sample by Haile (1979) suggests that the crust in the vicinity of Sikuleh (Locality H in Haile, 1979) was at 26°S in the Jurassic. This result was used by Metcalfe (1994) to suggest a Gondwana origin for the proposed micro-continental fragment.

The continental nature of the Sikuleh part of the Woyla Terrane (Si, Fig. 2) is rejected by Barber and Crow (2003), who instead propose an intra-oceanic arc origin. The Woyla Group consists, at least in part, of accreted fragments that include seamounts, reef fragments, ophiolites and associated ocean floor sedimentary sequences (Barber and Crow, 2003; Wajzer et al., 1991), but no clear continental basement can be identified, much like the Kohistan-Ladakh Arc in the central Neo-Tethys. Paleontological constraints from a single foraminifera specimen within the Batang Natal Megabreccia provide a Late Triassic age (Wajzer et al., 1991), and suggest that the oceanic crust that was consumed in the Woyla intra-oceanic subduction system in the Cretaceous was at least Late Triassic in age, consistent with the age of Meso-Tethyan oceanic crust subducted along the Sumatra segment predicted by our reconstructions for the Cretaceous (Fig. 10). In addition, the accretion of highly disrupted lenses of oceanic crust and sedimentary sequences onto the Woyla Terrane is consistent with the observations of accreted oceanic plate stratigraphy further east in the Luk Ulo-Meratus Suture Zone between East Java-West Sulawesi and the core of Borneo (Wakita, 2000; Wakita and Metcalfe, 2005).

3.3.1 Development of the Woyla intra-oceanic arc

The Woyla Terrane, largely represented by the Woyla Group of sedimentary sequences and intrusions, likely developed on an active intra-oceanic margin (with possible continental basement) in the Early Cretaceous (Figs. 9-11), separated from mainland Sumatra by a marginal sea (Rock et al., 1983; Wajzer et al., 1991). However, the origin of the Woyla Arc has recently been debated, with a model proposing a Gondwana origin for both the Woyla and the Kohistan-Ladakh Arc (Hall, 2012) as the result of continued rollback of a south-dipping subduction zone (Fig. 4b-c). Alternatively, the model of Zahirovic et al. (2014), and the one presented here, invoke a southern Eurasia origin of the Kohistan-Ladakh and Woyla intra-oceanic island arcs. The scenario invoking south-dipping subduction along northern Gondwana in the Late Jurassic could be corroborated by the preservation of contemporaneous arc rocks on Greater India (Tethyan Himalayas) or the NW Australian Shelf, which are not yet documented. The scenario invoking north-dipping Meso-Tethyan subduction to detach the Argoland continental fragments and open the Neo-Tethys in the latest Jurassic can be corroborated by the subduction history along Lhasa, West Burma (Myanmar) and Sundaland. Subduction is suggested to have initiated along the West Burma block at ~163-152 Ma by jadeite geochronology (Shi et al., 2008; Shi et al., 2014). This age is similar to that of the 154 Ma Matum Das tonalite within the Kohistan-Ladakh arc (Schaltegger et al., 2003) to the west. The formation of the Naga Ophiolite during ~156-150 Ma, based on Kimmeridgian-lower Tithonian cherts (Baxter et al., 2011), suggests a close temporal and geodynamic association between the Kohistan-Ladakh (Lhasa segment), Mawgyi (West Burma segment) and Woyla (Sumatra segment) intra-oceanic arcs along which Meso-Tethyan oceanic crust began subducting in the latest Jurassic.

3.3.2 Subduction of the Woyla back-arc basin

The resumption of Andean-style subduction in the central Neo-Tethys along Lhasa is well-constrained to ~109 Ma, based on the onset of subduction-related magmatism in the Gangdese Batholith (Ji et al., 2009). However, the timings of subduction initiation along West Burma and Sumatra are less well constrained. An Albian (~113-100 Ma) unconformity on West Burma (Morley, 2012a) may indicate compression related to subduction initiation, which is contemporaneous with observations in Lhasa, and the supra-subduction formation of the Andaman Ophiolite at 95 ± 2 Ma (Pedersen et al., 2010) may suggest the onset of rollback and extension in the overriding plate. Here we interpret the ~115 Ma peak in Ultra- and Very-High Pressure metamorphics in the Luk-Ulo Suture Zone (Figs. 10-12) (Parkinson et al., 1998) as indicators of subduction initiation of the Woyla back-arc basin to account for the Albian unconformity on West Burma. The 105 to 90 Ma dioritic and granodioritic intrusions into the Wuntho-Popa Arc (Mitchell et al., 2012), west of the Sagaing Fault, suggest continuity of the contemporaneous Lhasa subduction zone into the West Burma segment of the margin. However, subduction to consume the Woyla back-arc basin may (also) have been south-dipping as argued in Morley (2012a).

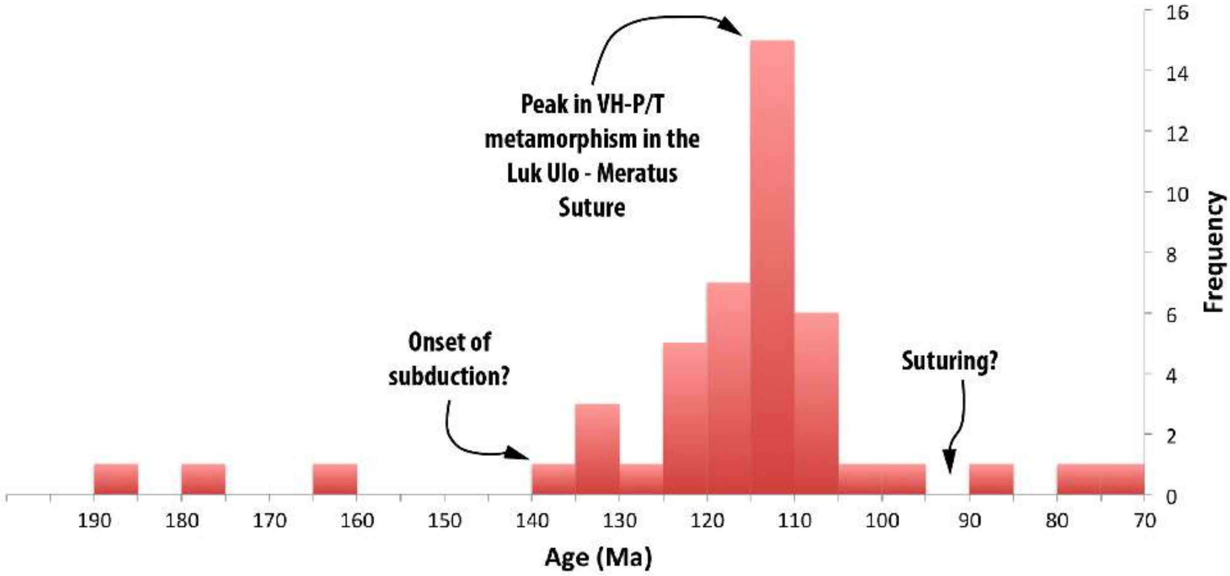

Fig. 12. Very High and Ultra High Pressure (VHP/UHP) metamorphic rocks in the Luk Ulo-Meratus suture on Java and Borneo (Fig. 2) record well-established subduction from at least ~140 Ma. A significant spike at ~115 Ma has been previously interpreted as a collision between East Java-West Sulawesi and the Woyla intra-oceanic arc (Gibbons et al., 2015; Zahirovic et al., 2014). However, in the refined plate reconstructions presented herein, the origin of East Java from the Argo Abyssal Plain would require excessive seafloor spreading rates, and instead we interpret the peak in VHP/UHP metamorphism to represent the onset of Woyla back-arc subduction along West Burma and Sundaland. Figure modified from Parkinson et al. (1998).

Although the Kohistan-Ladakh Arc is loosely constrained to near-equatorial paleo-latitudes during the mid-Cretaceous (Burg, 2011), no latitudinal constraints exist for the Woyla Arc. However, some constraints are available for the closure of the Woyla back-arc basin, and the collision of the intra-oceanic arc with Sumatra (Figs. 13-14). The Woyla Group is intruded by a number of Late Cretaceous igneous bodies, including the 84.7 ± 3.6 Ma (K-Ar) Batu Madingding diorite and the 78.4 ± 2.5 Ma (K-Ar) andesite in the southwest Batang Natal section (Wajzer et al., 1991), after which a significant magmatic gap is interpreted to represent collision of the Woyla Terrane with Sumatra. Hall (2012) argued that no subduction occurred on the Woyla/Sumatra segment of the Tethyan margin between 90 and 45 Ma (Fig. 4c), largely due to the presence of a regional unconformity that was interpreted to signify the absence of subduction-related dynamic subsidence on the overriding plate (Clements et al., 2011). However, only a ~10-15 Myr magmatic gap associated with a hiatus in subduction between ~75 and 62 Ma (Fig. 15) can be accounted for in the volcanic record on Sumatra (McCourt et al., 1996; Zahirovic et al., 2014). However, some of these (~10 Myr) magmatic gaps may be due to sampling issues, and future work may reveal more continuous subduction histories. The choice to impose a ~45 Myr (Hall, 2012) rather than a ~10-15 Myr Zahirovic et al. (2014) subduction hiatus has significant geodynamic implications for the region, where the continued northward motion of the Indian Plate needs to be accommodated by an oceanic transform that cuts across older oceanic lithosphere and pre-existing structural fabric in the reconstructions of Hall (2012) (Fig. 4c).

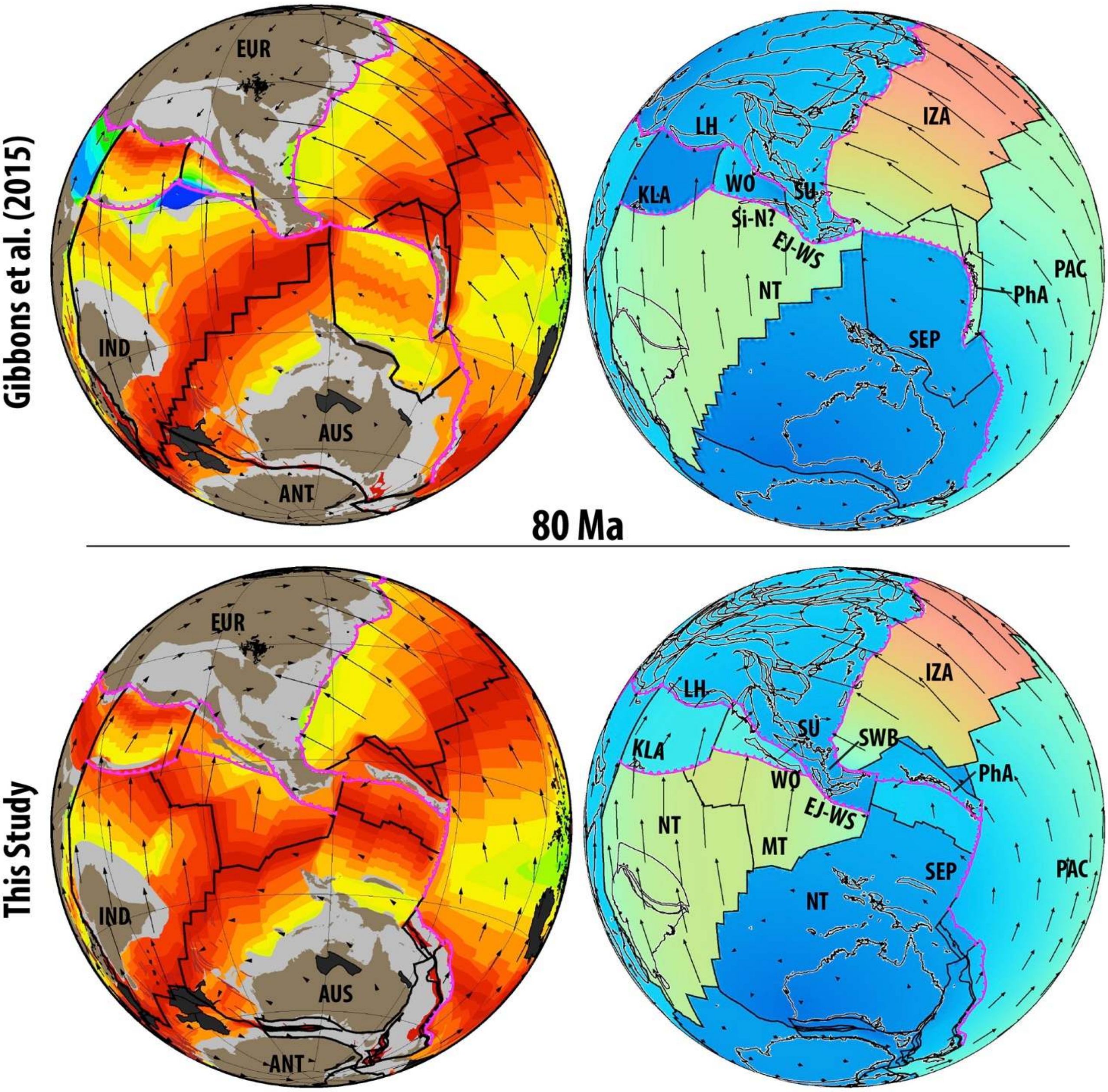

Fig. 13. India’s northward motion accelerated from ~80 Ma. The subduction polarity likely reversed along the Philippine Arc and suturing of the East Java and West Sulawesi continental fragments to the Southwest Borneo Core was complete by this time. The Woyla Terrane was approaching the Sumatran margin by this time. Additional polygons in the lower panels for Australian-Antarctic and Lord Howe-Tasman Sea regions represent areas of deforming continental crust.

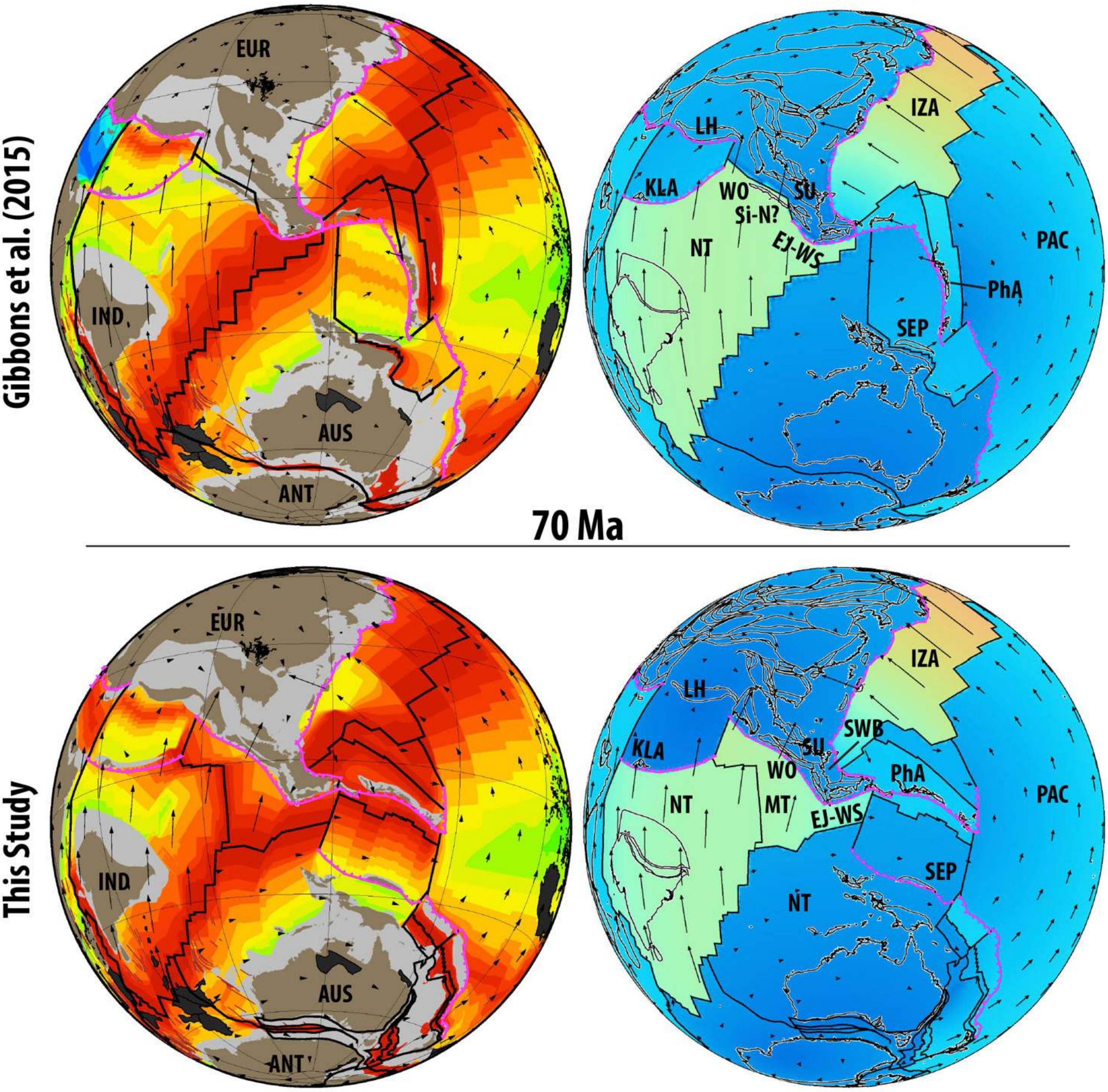

Fig. 14. The Woyla Arc collided with Sundaland (SU) by ~75 Ma, and impeded subduction in this segment of the margin for ~10 Myr. The Meso-Tethyan Plate was still likely being subducted along the Sunda Trench based on the refined plate reconstructions, with Wharton Ridge arrival near eastern Sundaland between ~70 and 60 Ma in both reconstructions. In this study, we reconstruct the subduction of the Sepik oceanic gateway from ~71 Ma based on the age of the Emo volcanics (Worthing and Crawford, 1996).

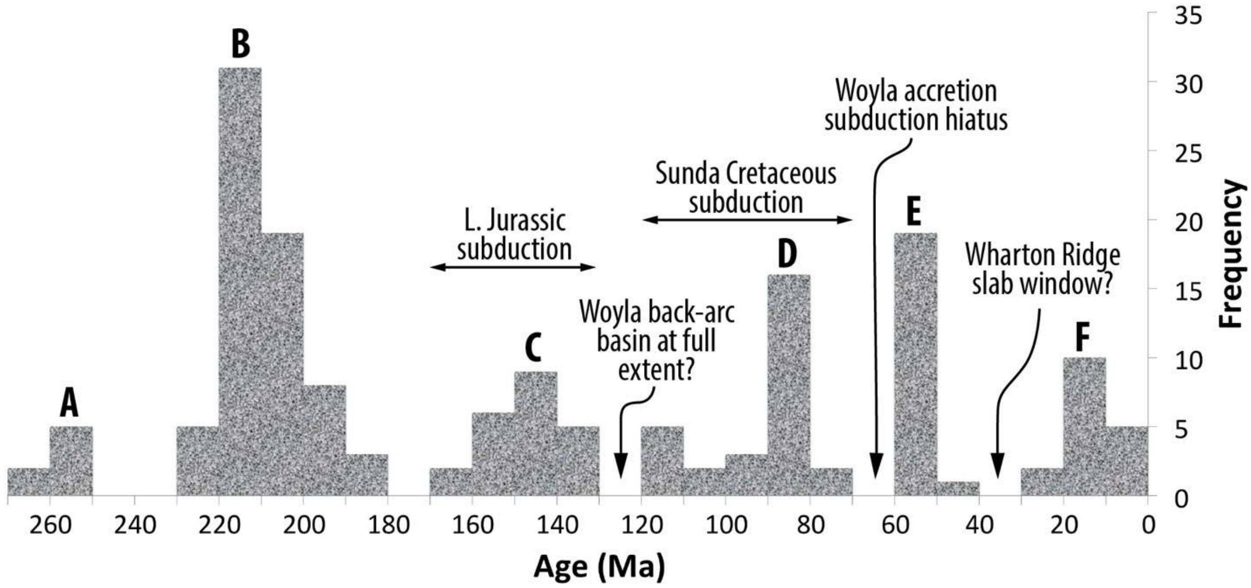

Fig. 15. A synthesis of arc volcanism on Sumatra, adapted from McCourt et al. (1996), highlights a number of short-lived magmatic hiatuses likely related to back-arc basin rifting and/or collisional processes that impeded subduction along the Sumatran continental margin.

The post-Cretaceous history of the Sumatran margin is less controversial, with magmatism related to the subduction of the Neo-Tethys and Indian oceans (McCourt et al., 1996), and widespread basin rifting and flooding occurring since Paleocene times on Sundaland (Doust and Sumner, 2007). One important geodynamic consideration for the Sumatra margin is the interaction of the (extinct) Wharton Ridge with the Java-Sunda trench, with the model of Whittaker et al. (2007) suggesting a long-lived slab window sweeping westward from eastern Sundaland (near Java) from ~75 Ma to the present. Such a scenario implies a time-dependent along-trench thermal anomaly affecting the Sundaland continent, and more importantly, the subduction of young oceanic crust (and hence thinner oceanic lithosphere) has important implications on the long-wavelength mantle-driven topography on the overriding plate (e.g., Flament et al., 2015, for Patagonian uplift associated with the Chile Triple Junction; Guillaume et al., 2009). The combination of a subduction hiatus in the Late Cretaceous, as well as the subduction of very young oceanic crust in the Eocene along the Java-Sunda trench would likely result in widespread regional dynamic uplift that has been proposed by Clements et al. (2011) to account for a widespread Late Cretaceous to Paleocene regional unconformity across Sundaland, as explored in Zahirovic et al. (In Review).

3.4 Accretionary history of the Java and Borneo margin segment

A key region recording the evolution of Southeast Asia in the context of Eurasian, Tethyan and (proto-) Pacific convergence is the Sundaland continental promontory. The core of Sundaland is composed of north-eastern Sumatra, West Java and the Southwest Borneo block (Hall, 2012; Metcalfe, 1988; Metcalfe, 2011; Zahirovic et al., 2014). The promontory is largely Phanerozoic continental crust (Hall, 2011), with accreted intra-oceanic and allochthonous continental fragments – some, like East Java, carrying Archean zircon signatures (Smyth et al., 2007). The continental fragments making up Sundaland have largely Tethyan-Gondwanan affinities based on paleontological, stratigraphic and paleo-magnetic constraints, as reviewed in Metcalfe (2006).

3.4.1 Subduction and accretion history of southern Sundaland

The onset of Late Jurassic subduction in the eastern Tethys is best represented by the ~180-165 Ma schists found within the Meratus Complex on the eastern periphery of the Southwest Borneo core, as well as the presence of Bajocian (~170 Ma) and younger radiolarians embedded in the Meratus Suture Zone (Wakita et al., 1998). Zircons shed into the Ciemas and Bayah Formations on West Java, have ages of 160 Ma and younger (Clements and Hall, 2007), and likely represent the onset of subduction along this margin. The Late Jurassic-Early Cretaceous age of subduction onset on this segment is consistent with the establishment of a major subduction system along southern Eurasia, spanning at least from western Lhasa to the easternmost Tethyan margin on eastern Sundaland (see previous sections for full chronology). A continuous record of very-high-pressure (VHP) metamorphic rocks (Fig. 12), including greenschists, blueschists, granulites, eclogites and jadeite-bearing metamorphics from ~140 Ma in the Luk-Ulo and Meratus region of Sundaland (Parkinson et al., 1998) suggests a well-established subduction zone in the Early Cretaceous. The VHP metamorphism peak at ~115 Ma has previously been interpreted as an arc-continent collision of Gondwana-derived continental fragments (including East Java, West Sulawesi, Mangkalihat, and eastern Borneo) with the eastward continuation of the Woyla Arc (Zahirovic et al., 2014). However, in this study we prefer an interpretation of Woyla back-arc basin subduction initiation at this time to account for the UHP/VHP metamorphism, as discussed in earlier sections.

A significant spike in the zircon age spectra at ~100 Ma (Clements and Hall, 2007) may indicate the arrival of the East Java-West Sulawesi in the vicinity of Sundaland. The obduction of the Meratus Ophiolite in the Cenomanian/Turonian between ~100 and 93 Ma (Pubellier et al., 2004; Yuwono et al., 1988) is consistent with the Cenomanian radiolarians found in the Meratus Complex (Wakita et al., 1998). The final closure of the Barito Sea back-arc basin along southern Sundaland occurred by ~80 Ma based on the lack of volcanic-derived zircons in fore-arc sandstones (Clements and Hall, 2011; Wakita, 2000). A Late Cretaceous to Paleocene (~72 to 65 Ma) unconformity on southwest Sulawesi (Milsom, 2000) may indicate collisional (uplift/denudation) processes, a subduction hiatus, or a combination of both. A resumption of subduction likely occurred in the Paleocene (Yuwono et al., 1988), with ~65-58 Ma (K-Ar) subduction related rocks (Guntoro, 1999), a 63 Ma tuff reported on South Sulawesi (van Leeuwen, 1981), and continuous calc-alkaline and tholeiitic volcanism occurring between 51 and 17 Ma on the Western and northern arm of Sulawesi (Elburg et al., 2003). Ongoing subduction and major deformation (Figs. 16-17), largely due to the arrival of the Australian continental margin, namely the Sula Spur (Figs. 18-19), started with the obduction of the East Sulawesi Ophiolite at ~20 Ma (Oligocene to Miocene) in a continent-continent collision setting (Bergman et al., 1996). The subsequent compressional deformation, and widespread oroclinal bending of Sundaland are discussed at length in Hutchison (2010) and Zahirovic et al. (2014), and are summarised in the following sections.

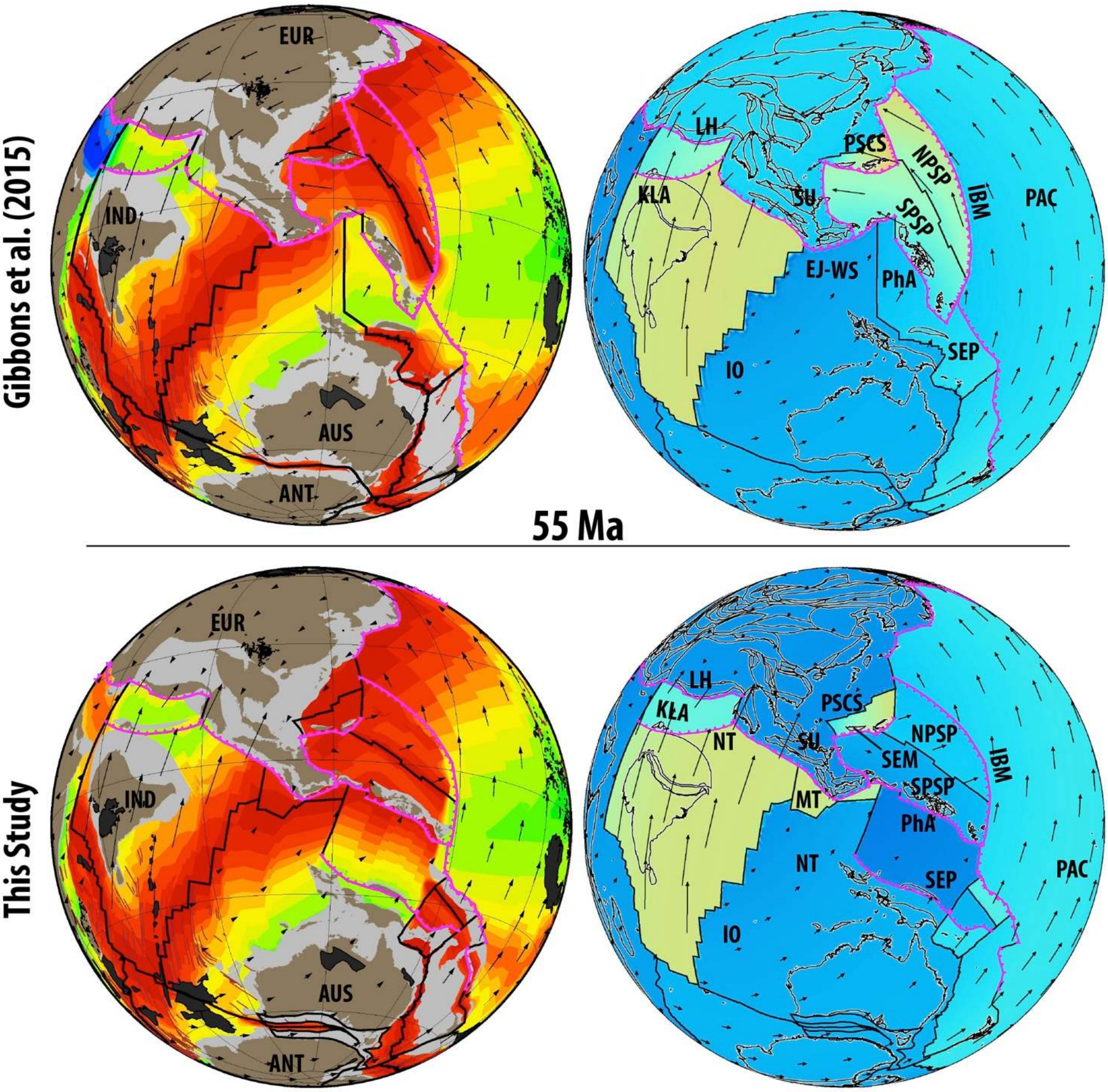

Fig. 16. The 55 Ma reconstruction marks the initial stages of contact between Greater India and the Kohistan-Ladakh Arc to close the Indus Suture Zone, leading to major changes in spreading rate and direction on the India-Antarctica ridge system. The rollback of the Izanagi slab opens the Proto South China Sea (PSCS) from ~60 Ma in a Tyrrhenian-style back-arc system. Subduction is initiated at ~55 Ma along the Izu-Bonin-Marina Arc (IBM) to consume Pacific (PAC) oceanic crust. IO – Indian Ocean, NPSP – North Philippine Sea Plate, SPSP – South Philippine Sea Plate, SEM – Semitau Block.

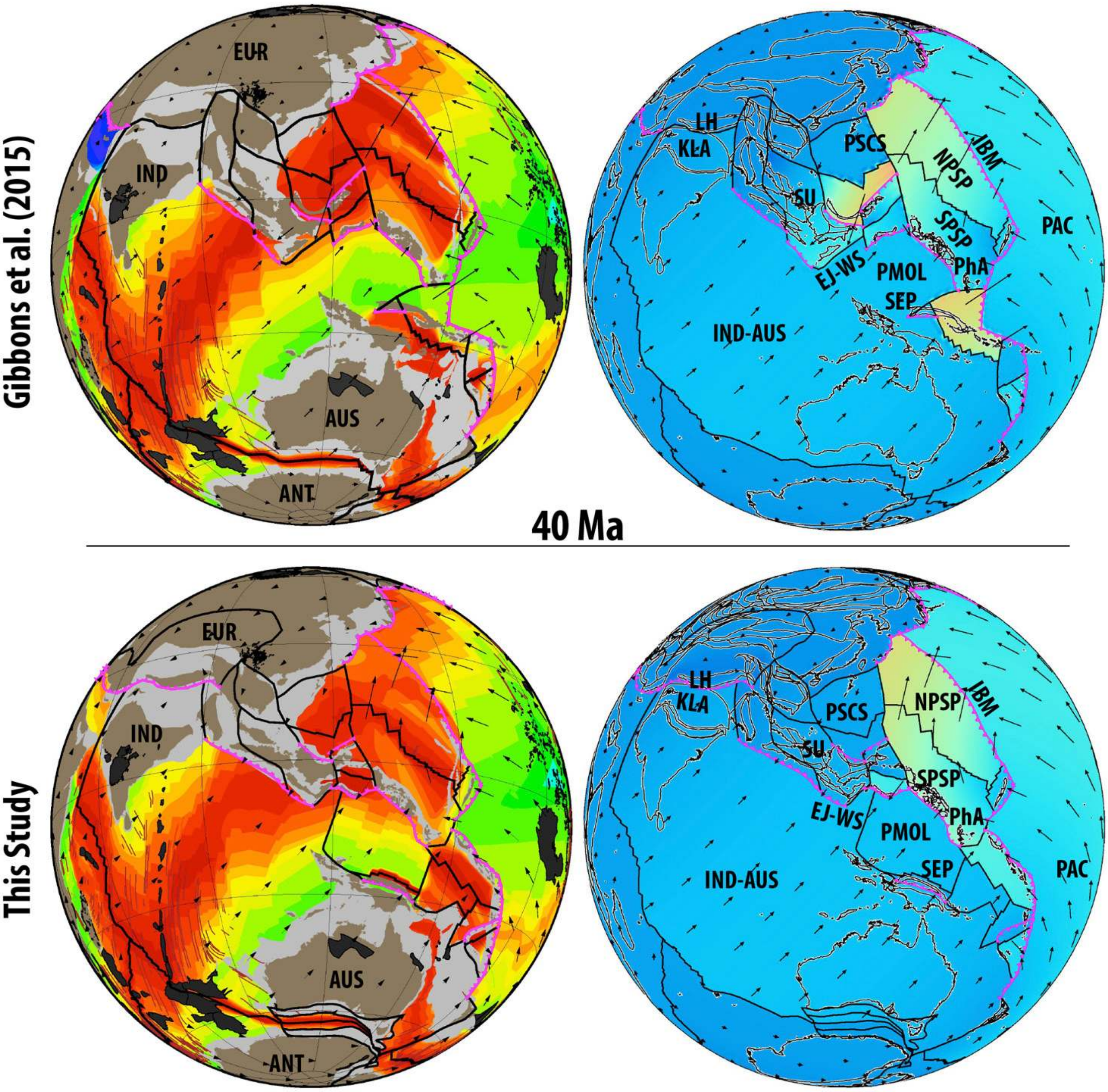

Fig. 17. Continent-continent collision between India and Eurasia likely initiated by ~47 Ma, leading to final closure of the Yarlung-Tansgpo and Shyok suture zones (see Fig. 1). The rollback of the Izanagi slab opens the Proto South China Sea and transfers the Semitau and Mindoro continental fragments from the South China margin onto northern Borneo, leading to a mid Eocene collision. The Sepik oceanic gateway is almost consumed along a north-dipping subduction zone, north of which the Proto Molucca Plate (PMOL) is consumed contemporaneously along the Philippine Arc.

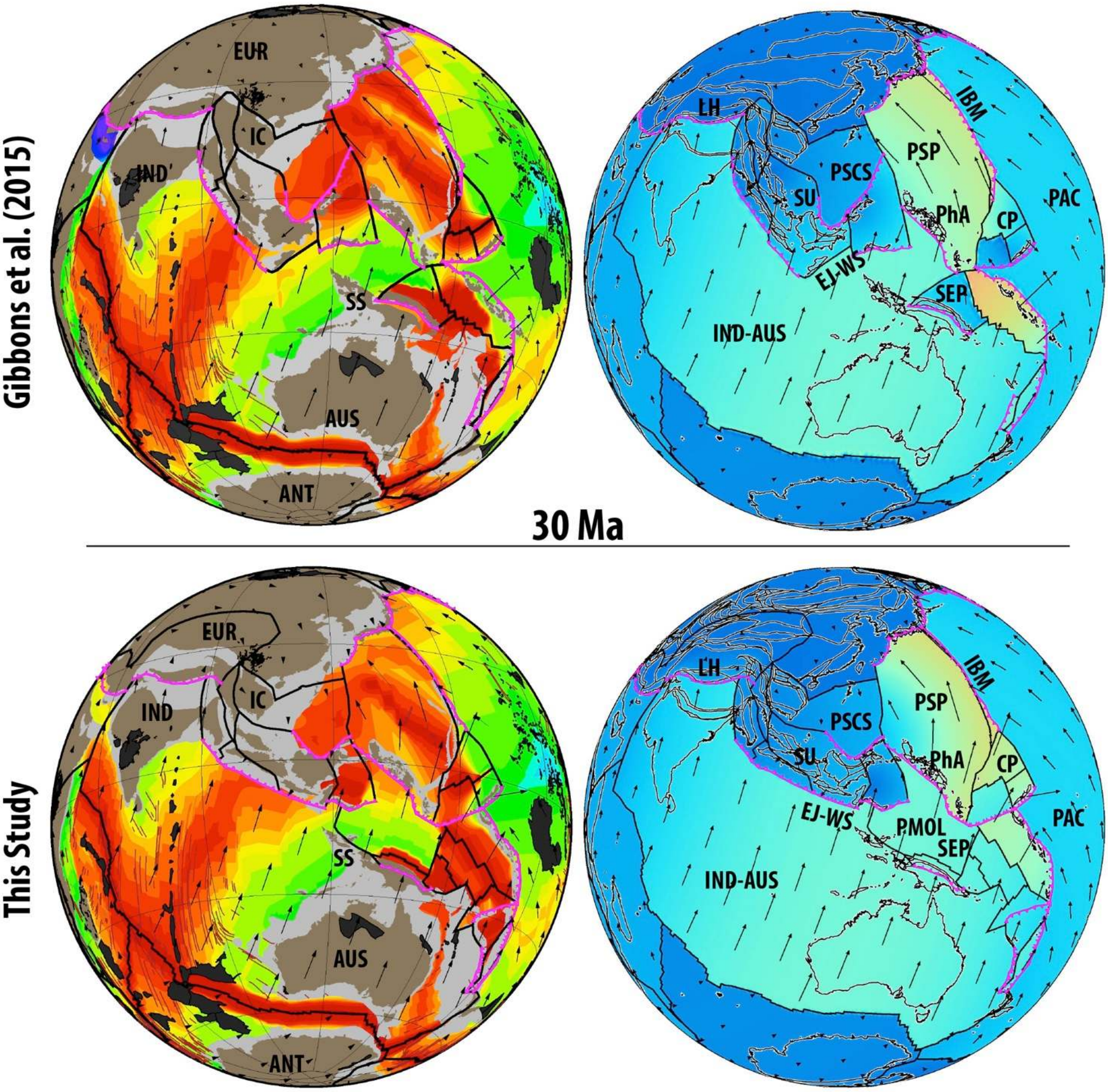

Fig. 18. India's continued, although slowed, northward advance results in the clockwise rotation and lateral extrusion of Indochina (IC), leading to the first stages of oroclinal bending in western Sundaland. The Sepik Terrane docks with the New Guinea margin, and the Sula Spur (SS) continental promontory on the northern Australian margin approaches Sundaland. CP – Caroline Plate, PSP – Philippine Sea Plate.

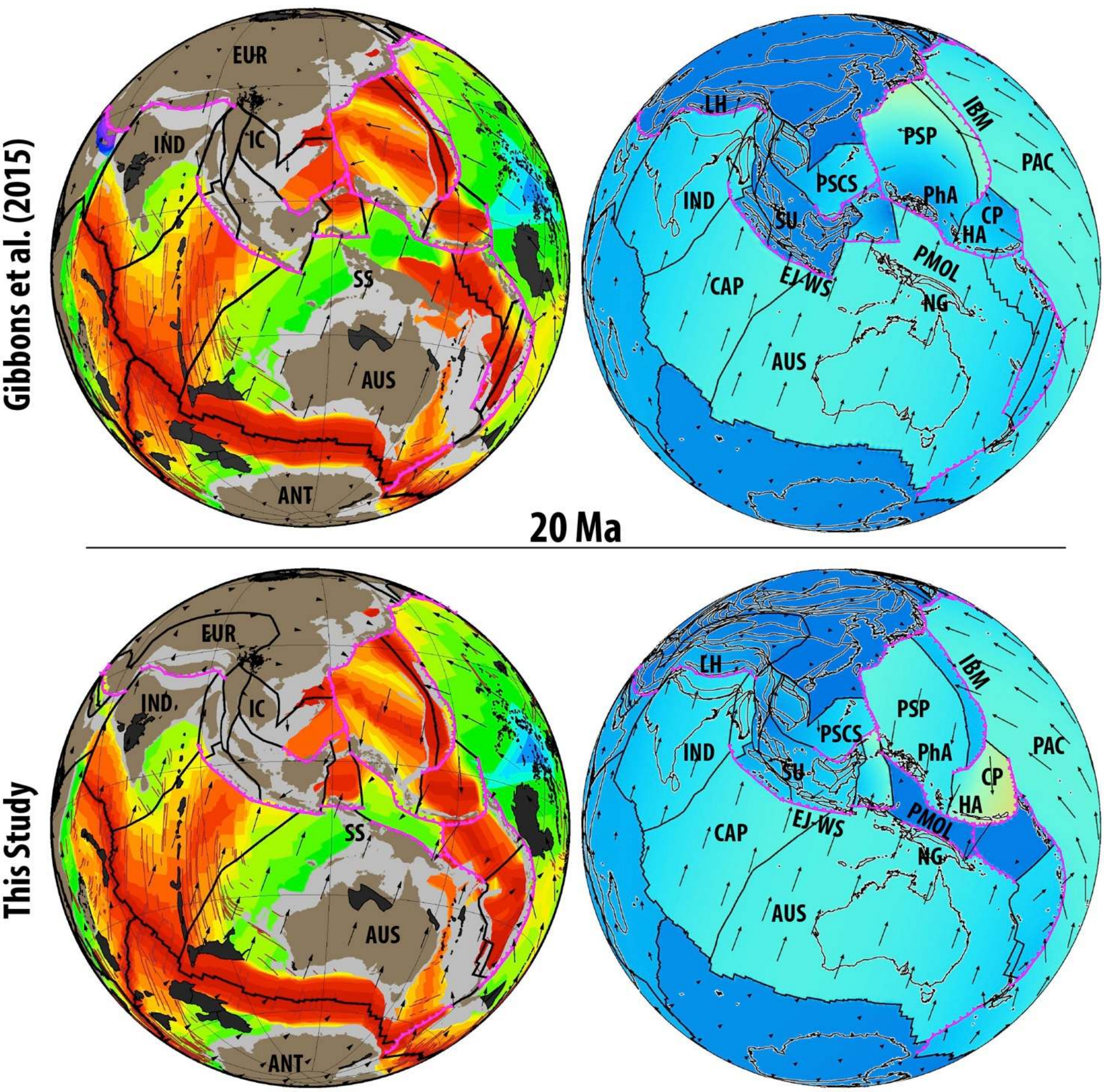

Fig. 19. The extrusion of Indochina due to India's northward motion, together with the collision between Sula Spur and West Sulawesi results in the oroclinal bending of Sundaland, resulting in major counterclockwise rotation of Borneo relative to Sumatra and the Malay Peninsula. South-dipping subduction initiates by ~20 Ma to account for the Maramuni Arc volcanics on New Guinea, with coeval north-dipping subduction of the Proto Molucca Plate (PMOL) accommodating southward motion of the Caroline Plate (CP) and the Halmahera Arc (HA).

3.4.2 Northern Sundaland collisions

The geological record on the Sundaland continental promontory captures the geodynamic interaction between the Tethyan and proto-Pacific oceanic domains, and holds clues as to how the present-day complexity of plate boundaries developed. Most notably, an ongoing topic of interest relates to how the east Asian margin transitioned from purely Andean-style subduction (Fig. 20) in the Late Cretaceous (Shi and Li, 2012), to one that is presently dominated by a labyrinthine network of intra-oceanic active margins connected by splayed transforms, ridge segments and diffuse plate boundaries (Bird, 2003). Although much of the proto-Pacific plates have been recycled into the mantle, the preserved flanks of the seafloor spreading history have been used to restore the lost plates, assuming seafloor spreading symmetry (Seton et al., 2012). However, the location and evolution of subduction zones is difficult to constrain, with the only clues coming from paleomagnetic constraints, arc volcanics and present-day mantle structure (Hall and Spakman, 2002; Miller et al., 2006; Queano et al., 2007; Zhao et al., 2007).

Although a key component of the intra-oceanic system is the Philippine Sea Plate, which is discussed in Section 3.5.1, the transition from Andean-style to intra-oceanic subduction north of Sundaland is most likely controlled by back-arc basin opening processes in the Late Cretaceous (Morley, 2012a). In the model proposed in Zahirovic et al. (2014), and adopted here, the emplacement of the Fukien-Reinan massif (Fig. 20) from Andean-style subduction ceases in the Late Cretaceous (Jahn et al., 1976), and was replaced with extension and back-arc basin opening (Li, 2000) due to rollback of the Izanagi slab (Figs. 13-14, 16). Such a scenario is consistent with the onset of Late Cretaceous tectonic subsidence in East Asian basins (Yang et al., 2004), as well as the crustal and biogeographic affinity between continental fragments wedged in northern Borneo and the Philippine Archipelago, namely the Semitau and Mindoro blocks, and their likely origin on the South China continental margin (Fig. 20) in the Late Cretaceous (Zahirovic et al., 2014). The rollback induced extension in the overriding plate (Schellart and Lister, 2005) likely progressed to back-arc basin opening, following the analogue of the Tyrrhenian back-arc system in the Mediterranean that detached and carried continental blocks to eventually collide as allochthons with a distant margin (Doglioni, 1991; Jolivet et al., 1999; Rehault et al., 1987). In the case of the Proto South China Sea, the Semitau and Mindoro fragments were likely detached from the East Asian margin by ~65 Ma, based on the onset of tectonic subsidence (Yang et al., 2004) and the ~59 Ma emplacement of supra-subduction zone ophiolites on Mindoro (Yumul et al., 2009). The continued rollback transferred Semitau and Mindoro southward, resulting in an Eocene collision with the northern Borneo margin to produce the Sarawak Orogeny (Cullen, 2010; Fyhn et al., 2010; Hutchison, 1996; Hutchison, 2004), after which southward subduction consumed the Proto South China Sea to emplace widespread volcanism on northern Borneo (Soeria-Atmadja et al., 1999).

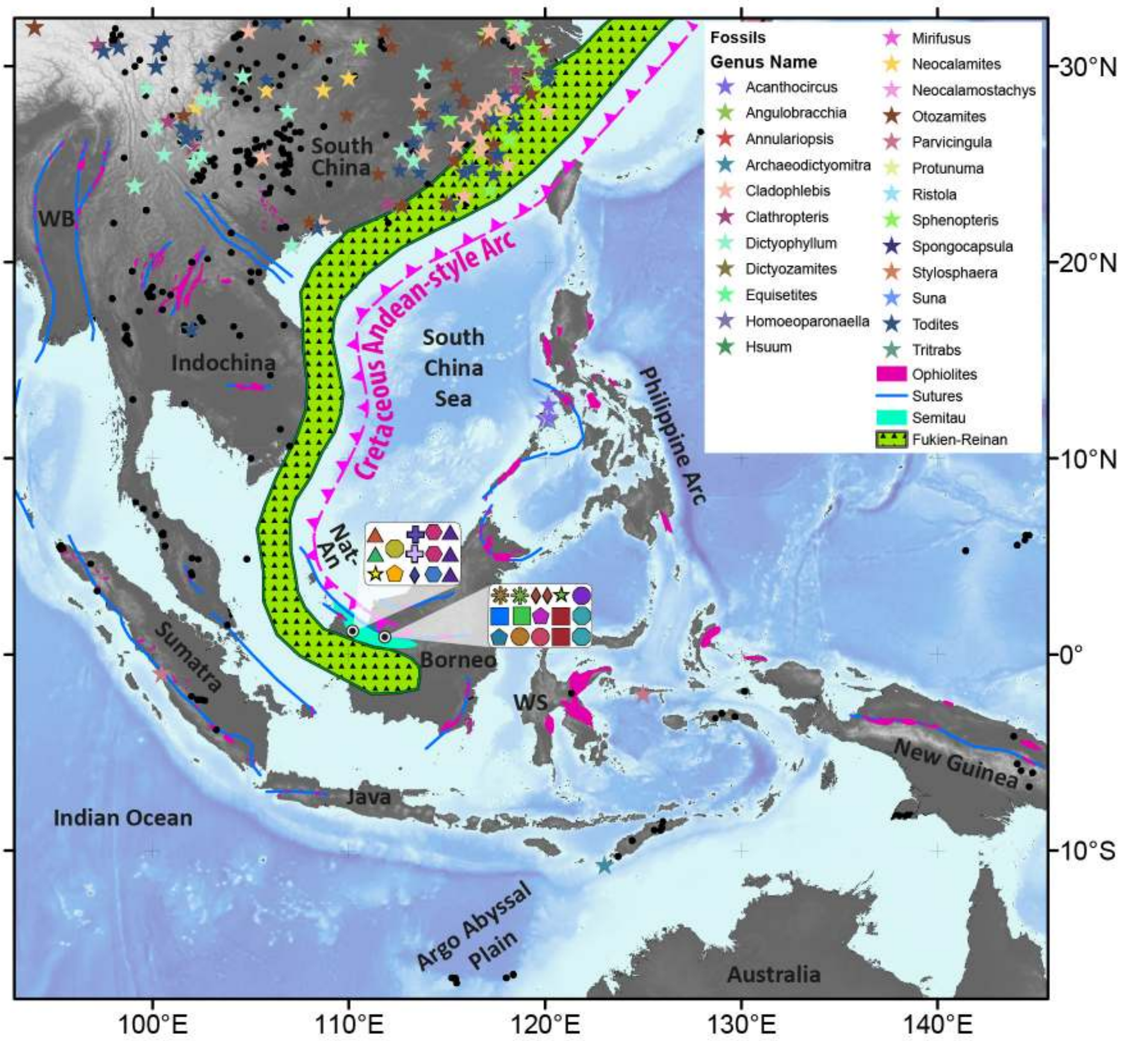

Fig. 20. Triassic and Jurassic fossil occurrences from the global Paleobiology Database (now Fossilworks) from Semitau (northern Borneo), represented by coloured symbols. The same fossils are also found elsewhere in Asia, with the strongest biogeographic affinity with mainland South China. The curved Fukien-Reinan massif (green hatched region) represents the Cretaceous Andean-style east Asian margin, which was replaced with an intra-oceanic setting in the Late Cretaceous. The curvature of the Andean-style magmatic arc also supports strong oroclinal bending of Sundaland in post-Cretaceous times. Nat-An – Natuna–Anambas Arc, WB – West Burma, WS – West Sulawesi. Figure adapted from Zahirovic et al. (2014).

The slab pull from south-dipping Proto South China Sea subduction along northern Borneo, along with the clockwise (CW) extrusion of Indochina resulting from the India-Eurasia collision (Fuller et al., 1991; Tapponnier et al., 1982), may have led to significant adjustments in the plate boundary forces acting on Sundaland. The ~32 Ma onset of seafloor spreading along the South China margin (Fig. 18) detached the Dangerous Grounds-Reed Bank continental blocks to open the South China Sea (Lee and Lawver, 1994; Lee and Lawver, 1995), with collision of the continental fragments with northern Borneo and South Palawan at ~15 Ma resulting in ophiolite obduction and the Sabah Orogeny (Hutchison, 2004; Hutchison et al., 2000), as well as choking the north Borneo subduction system and shutting down seafloor spreading in the South China Sea (Briais et al., 1993). This southward collision was wedged between the India-Eurasia collision from ~47 Ma (see Section 3.2) and the collision of the Australian northern margin with eastern Sundaland from ~25 Ma (Bergman et al., 1996; Hall, 2002). This arrangement of plate boundaries, and the driving forces, presumably had significant consequences for the rotational history of Borneo and the deformation of Sundaland.

3.4.3 Sundaland oroclinal bending

The large counterclockwise (CCW) rotation of Borneo, relative to stable Sundaland, in the Cenozoic has drawn a range of interpretations and led to a number of competing models (see discussion in Zahirovic et al. (2014)). Each model of Borneo rotation has consequences for the deformation history on Sundaland (in particular, the basins of the Sunda Shelf and Java Sea), as well as understanding the mechanism that led to the 90° CCW rotation of Borneo relative to Sundaland in the Mesozoic, including up to 50° CCW rotation since 25 Ma (Fuller et al., 1999). In the absence of large transform faults, such as the Red River Fault bounding northern Indochina, within the Java Sea or the Sunda Shelf, Hutchison (2010) proposed a model of oroclinal bending for the rotation of Borneo as a mechanism to explain the deformation of the Sundaland continental promontory.

Hutchison (2010) synthesised the paleomagnetic evidence, as well as observations of curved lineaments observed in the gravity anomalies of Sundaland and the curvature of the Natuna and Anambas Cretaceous paleo-arc (Fig. 20) to infer that wholesale bending of Sundaland accommodated the CW rotation of Indochina and the CCW rotation of Borneo (Fuller et al., 1999; Fuller et al., 1991). The curved lineaments (Fig. 21) are most likely to be successive generations of ancient volcanic arcs (Hutchison, 2010), with the most obvious example being the curved arc belonging to the Middle to Late Triassic tin belt granites on Bangka and Billiton islands, as well as the previously-mentioned Natuna-Anambas Cretaceous Arc. Zahirovic et al. (2014) expanded on the work of Hutchison (2010) and used filtered Bouguer anomalies (Balmino et al., 2012) to extract geometrical constraints on the oroclinal bending (Fig. 21), and constructed a kinematic oroclinal bending model that accounts up to ~78° CCW rotation of Borneo since ~50 Ma.

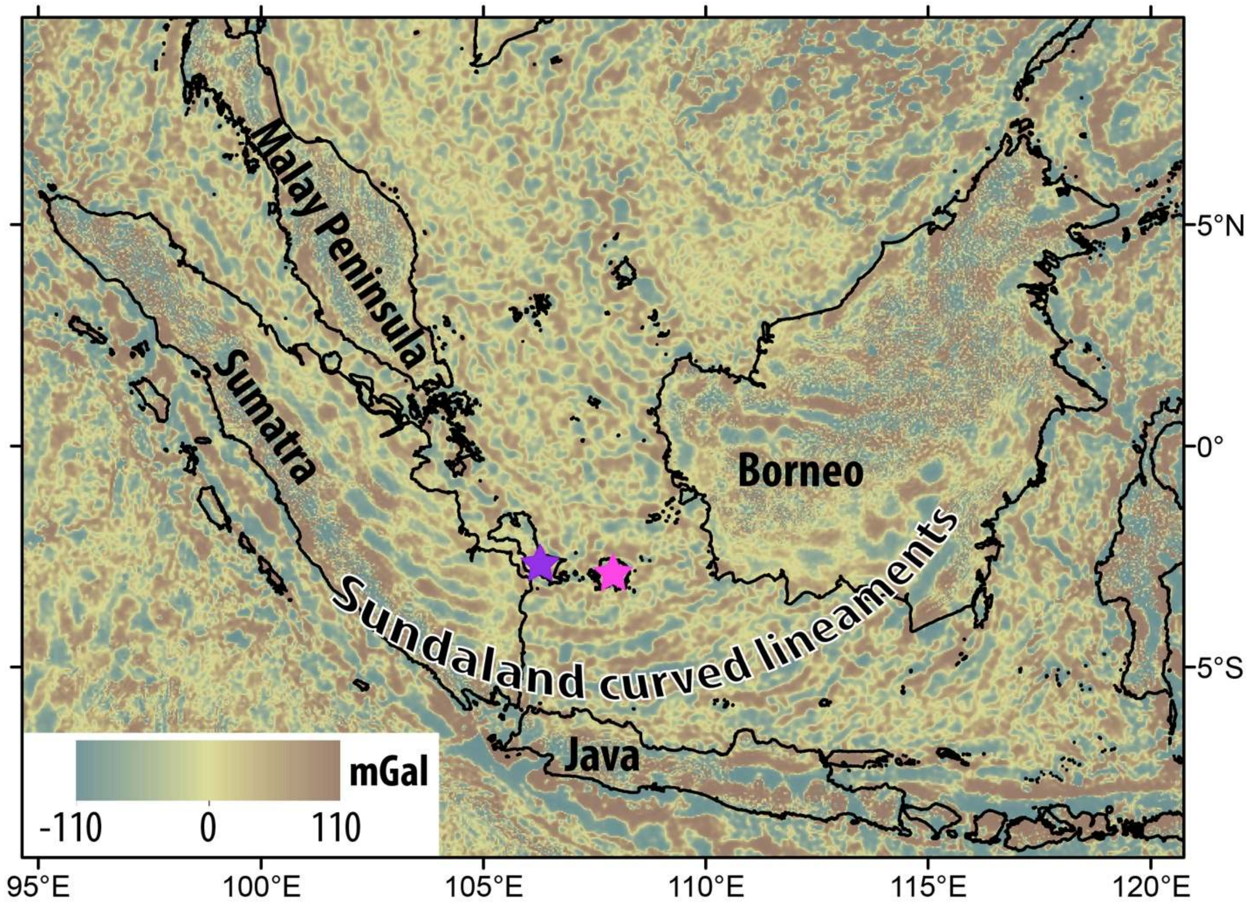

Fig. 21. The band-pass-filtered (150 to 10 km) Bouguer gravity anomalies from the 1 min World Gravity Map (Balmino et al., 2012) highlights the large-scale structures and the curved lineaments on Sundaland, resulting from oroclinal bending as proposed by Hutchison (2010). Bangka Island – purple, Belitung (Billiton) Island – pink.

One important distinction between the oroclinal bending of Sundaland and classical models of oroclinal bending largely relates to the tightness of the oroclinal folds and the deformation experienced by the continental crust (Carey, 1955; Eldredge et al., 1985). In the case of Sundaland, the hinge of the orocline is likely to be in the Sunda Shelf west of Borneo, with a wide region of bending rather than tight oroclinal bends that are typically reported for Kazakhstan (Abrajevitch et al., 2008) and the Mediterranean (Rosenbaum, 2014). Although the boundaries of the Sundaland continental promontory experienced compression during oroclinal bending, the Java Sea and Sunda Shelf were dominated by extension from Eocene times to the mid Miocene, after which a major phase of basin inversion dominated the tectonic regime of Sundaland to the present (Doust and Sumner, 2007; Pubellier and Morley, 2014).

3.5 New Guinea and the Philippines

3.5.1 Origin and evolution of the Philippine Archipelago

Tectonic reconstructions of the transfer of Gondwana-derived terranes following the breakup of Pangea are limited by the lack of preserved seafloor, but are typically supplemented with high-quality and well-constrained onshore geological data that helps reconstruct the synthetic seafloor spreading histories. However, the region east of Sundaland, which includes the Philippines and New Guinea are considerably more complicated, as they straddle the Tethyan and (proto-) Pacific tectonic domains, resulting in a complex interaction dominated by back-arc basin formation processes and multiple phases of collision, obduction and subduction that consumed them.

One early synthesis of the tectonic evolution of the post-Eocene West Pacific was carried out by Jolivet et al. (1989), who modelled the plate motions in six stages (56, 43, 32, 20, 12 and 3 Ma), and importantly, provided finite rotation poles that define their time-dependent plate circuit. In the reconstructions of Jolivet et al. (1989), and subsequent models (Hall et al., 1995a; Lee and Lawver, 1995; Pubellier et al., 2003; Queano et al., 2007; Zahirovic et al., 2014), the Philippine Sea Plate develops in near-equatorial southern latitudes during the Eocene (Hall et al., 1995a; Hall et al., 1995b; Richter and Ali, 2015), and is isolated from the surrounding plate circuits by a network of plate boundaries (including subduction zones and transforms) for much of the time. This tectonic isolation, and lack of preserved seafloor spreading linking the Philippine Sea Plate directly to the Pacific, Eurasian or Australian plates leads to difficulties in reconstructing the absolute and relative plate motions of this region that links the Pacific with the Indian oceans. However, the seafloor spreading history within the Philippine Sea Plate itself has been well-documented, including the opening of the West Philippine Basin between ~55 and 33 Ma (see Hilde and Chao-Shing (1984), Deschamps and Lallemand (2002), and references therein) and the back-arc opening of the Shikoku and Parece Vela back-arc basins between ~29 and 15 Ma from Philippine Sea Plate rotation (Sdrolias et al., 2004) and Izu-Bonin-Mariana trench rollback (Kobayashi, 2004).

Although the seafloor spreading history of the Philippine Sea Plate is confined to post-Eocene times, the Philippine Arc has recorded a much longer history of subduction, with the earliest supra-subduction zone (SSZ) rocks from the Late Jurassic. The SSZ ophiolites from the Philippine Arc have ages of 156.3 ± 2.0 Ma and 150.9 ± 3.3 Ma (Lagonoy Ophiolite), and 142 ± 4 Ma (Gag Island, Halmahera) from the synthesis of Encarnación (2004). They are discussed at length in Zahirovic et al. (2014). Recent work by Deng et al. (2015) reported mid-Cretaceous, 126 ±3 Ma and 119 ± 2 Ma (U-Pb), SSZ volcanics from Cebu Island. These ages are consistent with the minimum 99.9 ± 7.0 Ma (Ar-Ar) age of the Calaguas Ophiolite (Geary et al., 1988; Geary and Kay, 1989), and the 100 ± 4 Ma arc rocks reported from Obi Island on Halmahera (Hall et al., 1995b), suggesting continuous Early Cretaceous subduction along the Philippine Arc. To reconcile the likely southern hemisphere origin of the Philippine Arc and the Late Jurassic-Early Cretaceous temporal similarity with north Gondwana rifting events, Zahirovic et al. (2014) proposed a SSZ origin in the vicinity of New Guinea, recently independently suggested by Deng et al. (2015). The multiple generations of ophiolites may suggest a tectonic scenario similar to the current multi-generation opening of back-arcs along the Izu-Bonin-Mariana system, and may explain the origin of the Daito and Oki-Daito ridges as paleo-arc features in the north West Philippine Basin.

3.5.2 Nature of the New Guinea margin since the Late Jurassic

To accommodate the northern Gondwana rifting episode in the Late Jurassic, Zahirovic et al. (2014) placed the East Java-West Sulawesi continental fragments along New Guinea as the simplest end-member of transferring blocks north towards Sundaland, but acknowledged that an

Argo Abyssal Plain origin (NW Australian shelf) would also be possible due to the lack of preserved seafloor spreading histories to constrain a pre-drift fit. In this study, we implement the Argo origin end-member scenario, which is consistent with the zircon age spectra linking East Java to the NW Australian shelf (Sevastjanova et al., 2015; Smyth et al., 2007). The 158-137 Ma ages of mafic rocks, some of which are associated with pillow basalts, on West Sulawesi (Polvé et al., 1997) are consistent with the oldest oceanic crust (155 ± 3.4 Ma) in the Argo Abyssal Plain on the NW Australian Shelf (Gradstein and Ludden, 1992). By shifting these continental fragments west along northern Gondwana, the New Guinea margin can accommodate the source of the Philippine Archipelago to have formed along its margin. The benefit of such a scenario is that it accounts for the origin of (likely) Jurassic age SSZ ophiolites within the Central Ophiolite Belt in New Guinea (Monnier et al., 2000). However, the Late Jurassic-Early Cretaceous (~157 ± 16 Ma) and Late Cretaceous (66 ± 1.6 Ma) are unpublished ages from Permana (1998), reported in Pubellier et al. (2003), and require further corroboration. What is known is that at least part of the New Guinea margin was an active margin in Early Cretaceous times, as indicated by the Early Cretaceous volcanism and the Kondaku Tuffs (Dow, 1977; Rickwood, 1954), and likely represents the continuation of the long-lived east Gondwana active margin.

To what extent the Late Jurassic-Early Cretaceous active margin extended west into the Indonesian portion of New Guinea remains poorly constrained. In the east, the protolith of the Bena Bena metamorphics is partly Late Triassic (221 Ma) in age, and is intruded by Jurassic granite of 172 Ma in age (Davies, 2012). In the west, the Bird’s Head region experienced granitoid emplacement in the Early Jurassic with one sample yielding an age of 197 ± 3 Ma (K-Ar) (Pieters et al., 1983), 205 ± 5 Ma in the P’nyang-1 exploration well in western Papua New Guinea (Valenti, 1993), and the 210 ± 25 Ma Mangole Volcanics on Banggai-Sula (Charlton, 2001). These results suggest that the trench along western New Guinea may have undergone rollback by the Late Jurassic to explain Early Cretaceous volcanics confined only to eastern New Guinea. Such a scenario is also consistent with the sedimentary history that records syn-rift sedimentation in the

Early-Middle Jurassic, followed by a post-breakup unconformity and the formation of a diachronous passive margin along much of New Guinea (Pigram and Panggabean, 1984), with the exception of continued Early Cretaceous arc volcanism in the east.

Although the Early-Middle Jurassic rift-drift sequence is well preserved on New Guinea, few constraints exist to identify which (if any) continental terranes rifted from this margin (Hill and Hall, 2003; Pigram and Symonds, 1991). Apart from the East Java and West Sulawesi blocks (Zahirovic et al., 2014), parts of the Sepik Terrane may represent a para-allochthon that detached from the margin in the Jurassic, as invoked in this study, to open a somewhat-narrow oceanic basin and form the Late Jurassic SSZ ophiolites (Permana, 1998; Pubellier et al., 2003) exposed along the Central Ophiolite Belt in New Guinea (Fig. 3). Even though the Sepik Terrane is the largest accreted block on the New Guinea margin, the composite nature of the Sepik crust – with both continental and intra-oceanic arc fragments (Klootwijk et al., 2003) – leads to an enigmatic tectonic evolution. The New Guinea margin experienced at least two collisional phases; one in the late Eocene (Hall, 2002) to mid Oligocene (Crowhurst et al., 1996; Pigram and Symonds, 1991), and another major collision responsible for compressional deformation in the Mobile Belt during the late Miocene (Hall, 2002; Hill and Hall, 2003; Hill and Raza, 1999). However, the collisional history of the Sepik Terrane remains controversial in terms of whether the terrane first collided with one or more intra-oceanic arcs and subsequently welded to New Guinea, or whether the converse is true.

Although the size of the oceanic basin that separated the Sepik Terrane from mainland New Guinea remains uncertain, the longevity of the oceanic basin can be inferred from subduction-related metamorphics that are distributed along the Central (Irian) Ophiolite Belt (Fig. 3), and eastward into the April Ultramafics and the Marum Ophiolite. The ~68 Ma high-temperature metabasites and ~44 Ma blueschists in the West Papuan Ophiolite (Weyland Overthrust) indicate that subduction of the Sepik oceanic basin was underway (Davies, 2012), which is consistent with ~45 to 40 Ma glaucophane (K-Ar) and 28 to 25 Ma (K-Ar) phengite ages (Baldwin et al., 2012) in the April Ultramafics. The Balantak Ophiolite on the East and Southeast Sulawesi Arm records ages of ~96-32 Ma (Monnier et al., 1995), with a paleo-latitudinal constraint of 17 ± 4°S at 80 Ma (Mubroto et al., 1994) suggesting that these ophiolites formed somewhere between Sundaland and the New Guinea margin. Such a scenario is consistent with north-dipping subduction of the Sepik oceanic basin, which may have generated supra-subduction zone ophiolites that were subsequently obducted onto Sulawesi. The Maastrichtian (~71 to 66 Ma, stratigraphic correlation and dating using foraminifera) Emo volcanics (Worthing and Crawford, 1996) were likely emplaced in a back-arc setting from north-dipping subduction along the Sepik Terrane, with final docking likely taking place by ~30 Ma (Findlay, 2003; Zahirovic et al., 2014), based on the 35 to 31 Ma (Ar-Ar) amphibolite age of the Emo metamorphics (Worthing and Crawford, 1996) and the cooling histories of exhumed blocks (Crowhurst et al., 1996).

Following the docking of the Sepik composite terrane, south-dipping subduction was likely established (Figs. 19, 22) to account for the ~18 to 8 Ma Maramuni Arc volcanism (Hill and Hall, 2003; Page, 1976), followed by post-collisional volcanism to at least ~1 Ma (Holm et al., 2014; van Dongen et al., 2010). The approaching Halmahera Arc, attached to the southern portion of the Caroline Plate, likely collided with the northern New Guinea margin by ~14 Ma (Figs. 22-23), leading to a major compressional phase in the Mobile Belt, that has been inferred from apatite fission track geochronology (Hill and Raza, 1999; Kendrick, 2000). Although the New Guinea margin is a key component of the Australian, Pacific and Eurasian convergence zone, more work is required to resolve competing tectonic scenarios for this margin (van Ufford and Cloos, 2005). However, additional insights can be made from interpretations of mantle structure from seismic tomography, as well as testing end-member scenarios using coupled plate kinematic and numerical mantle convection modelling of the New Guinea margin.

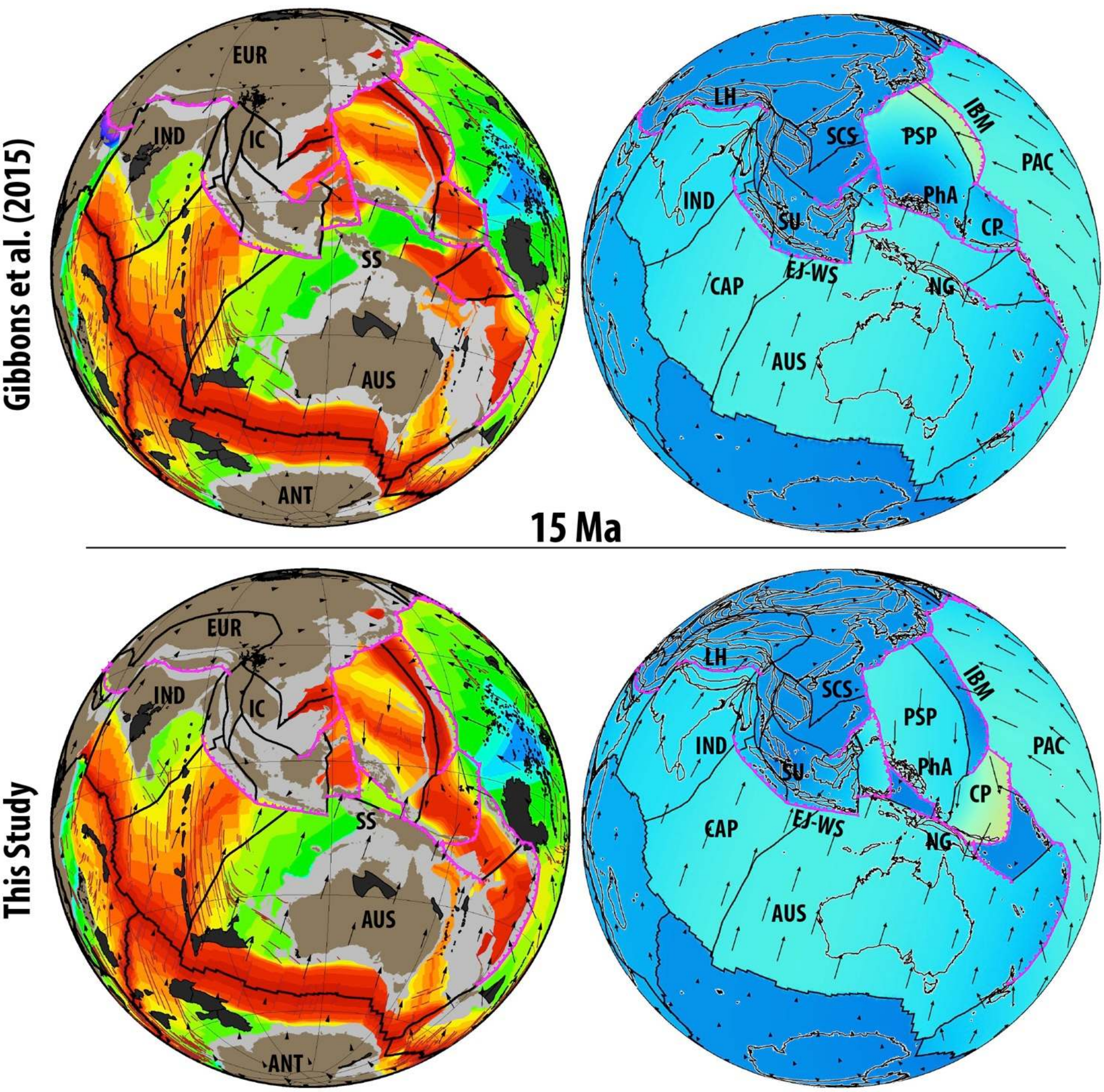

Fig. 22. The 15 Ma timestep records the transition to compressional tectonics on Sundaland and New Guinea. The arrival of the Dangerous Grounds-Reed Bank continental fragment shuts down Proto South China Sea subduction along Borneo, and results in ophiolite obduction in Palawan and orogenesis on Borneo. In the refined reconstructions, the Halmahera Arc collides with New Guinea at ~15 Ma to result in major compression in the New Guinea Mobile Belt.

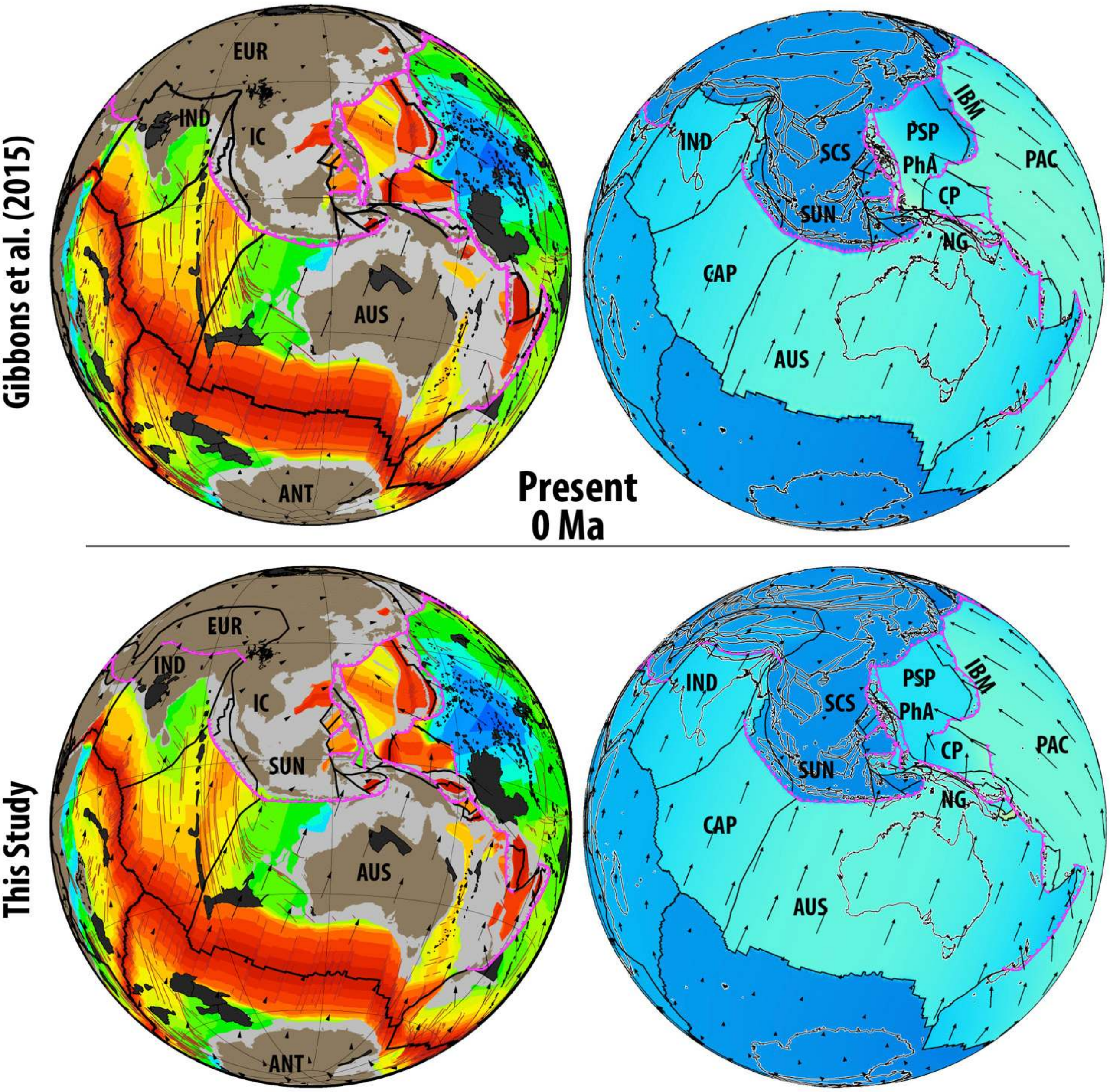

Fig. 23. The present-day tectonic configuration of Southeast Asia is the result of long-lived Indo-Australian, Eurasian and Pacific convergence accommodated by the subduction of Tethyan ocean basins and back-arcs. The northward motion of India is significantly slower than in the early Eocene, with intra-plate diffuse deformation in the Capricorn (CAP) Plate since ~20 Ma.

4 Insights from age-coded slabs in seismic tomography

In light of the complex tectonic evolution of Southeast Asia and the Tethyan-Pacific oceanic linkages, we interpret high velocity seismic anomalies from the P-wave model of Li et al. (2008) to obtain insights into the subduction history (Fig. 24). Although assuming constant and vertical slab sinking is a simplification, it is arguably a reasonable assumption for the late Cenozoic where large lateral slab advection would be limited, as indicated by previous estimates of less than ~1-2 cm/yr of mid-mantle lateral flow in the Tethyan realm (Becker and Faccenna, 2011; Zahirovic et al., 2012). We compare the plate reconstructions with age-coded depth slices of high velocity seismic anomalies, applying a sinking rate of 2 cm/yr in the lower mantle, and end-member estimates of 3 and 8 cm/yr in the upper mantle (see Methods). The proposed lower mantle slab sinking rates for the Tethyan realm are larger (~2 cm/yr) than estimated global averages (1.2-1.3 cm/yr in Butterworth et al., 2014; van der Meer et al., 2010), and may reflect the suction exerted by deep slabs in this slab graveyard area (Conrad and Lithgow‐Bertelloni, 2004).

The plate reconstructions in our base model (Zahirovic et al., 2014) were calibrated for the Philippine Sea Plate and Sundaland using a similar method (assuming sinking rates of 3 and 1.2 cm/yr in the upper and lower mantle, respectively). However, in the refined reconstructions, we do not modify the Sundaland oroclinal bending model, but modify the position of the Philippine Sea Plate since ~30 Ma to ensure collision of the Halmahera Arc with New Guinea at ~15 Ma, to account for the onset of widespread compression in the New Guinea Mobile Belt (Hill and Hall, 2003). Consequently, the match between Sundaland subduction zones and age-coded slabs from tomography is not surprising. Modifications to fit the Philippine Sea Plate to surface geology since ~30 Ma, rather than seismic tomography, presents a case study to test whether both geological and seismic tomographic constraints can be accommodated simultaneously.

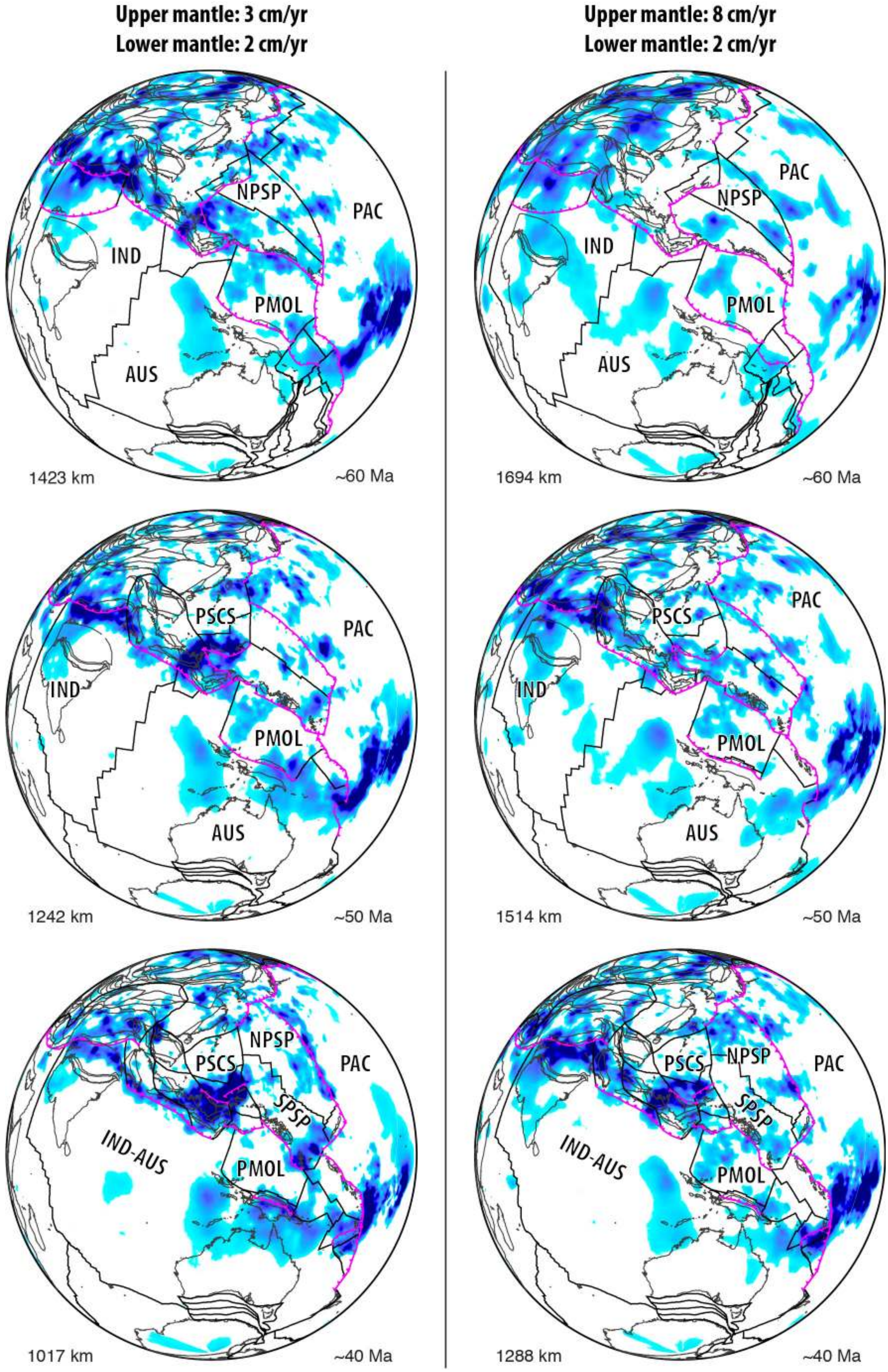

Fig. 24. Fast seismic velocity anomalies from the Li et al. (2008) P-wave seismic tomography model compared to our refined plate reconstructions. Ages are attributed to depths assuming that

the average vertical sinking rate of slabs is ~2 cm/yr in the lower mantle, and 3 cm/yr (left) and 8 cm/yr (right) in the upper mantle. Additional polygons in Australian-Antarctic and Lord Howe-Tasman Sea regions represent areas of deforming continental crust. See Supplementary Animations 5 and 6.

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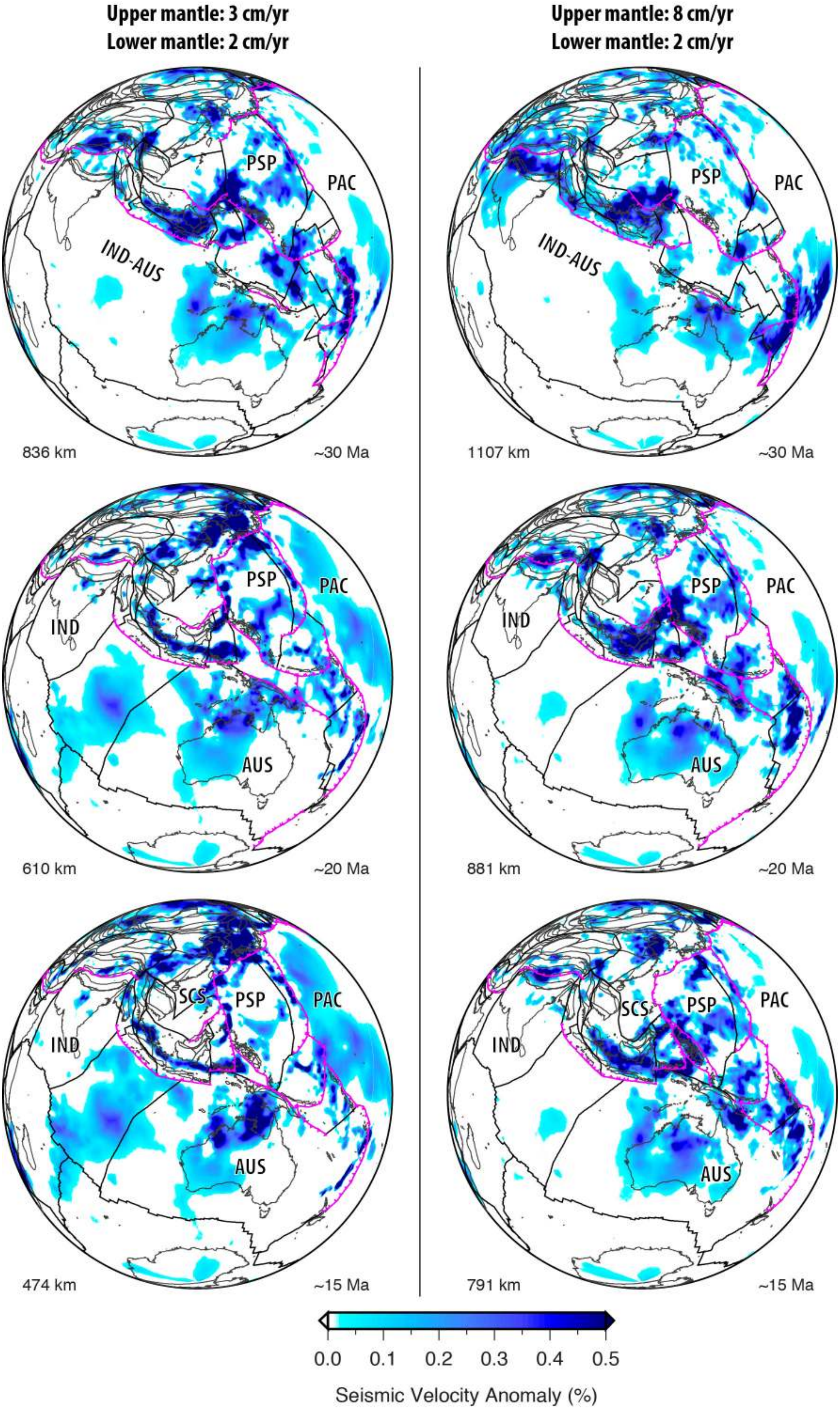

Upper mantle: 3 cm/yr
Lower mantle: 2 cm/yr
Upper mantle: 8 cm/yr
Lower mantle: 2 cm/yr
PSP
PAC
IND-AUS
836 km
~30 Ma
1107 km
~30 Ma
IND
AUS
610 km
~20 Ma
881 km
~20 Ma
SCS
474 km
~15 Ma
791 km
~15 Ma
0.0
0.1
0.2
0.3
0.4
0.5
Seismic Velocity Anomaly (%)

Fig. 24. (continued)

At ~60 Ma (Fig. 24), the scenario invoking a slower sinking rate in the upper mantle better reproduces the Sunda slab, as well as the subduction of the Proto Molucca Plate (PMOL) beneath the Philippine Arc and the rollback-induced opening of the Proto South China Sea. The match with the Sunda and Philippine slabs is not surprising, as a slower sinking rate was also used to calibrate the position of these blocks in our base plate motion model (Zahirovic et al., 2014). Both slab sinking scenarios reproduce the Andean-style subduction along southern Eurasia consuming the Kohistan-Ladakh and Woyla back-arc basins, as implemented in the reconstructions based on the near-equatorial latitudes from paleomagnetic estimates. At ~50 Ma, the slower sinking rate matches the oroclinal bending of Sundaland and subduction of the Proto South China Sea, which is again expected due to calibration of the reconstructions with tomography. However, the match to subduction of the Sepik oceanic basin north of New Guinea is not imposed, and suggests the large E-W slab presently beneath Australia is likely sourced from this subduction system (Schellart and Spakman, 2015). Interestingly, the gap in the Sunda slab along Sumatra in both sinking rate end-members, coinciding with the modelled location of the Wharton Ridge, supports the slab window scenario proposed by Whittaker et al. (2007).

The 40 and 30 Ma timesteps in the age-coded seismic tomography depth slices support ongoing subduction along northern Borneo (Fig. 24), and waning subduction in the India-Eurasia segment of the active margin. The ~20 and 15 Ma reconstructions (Fig. 19 and 21) highlight the requirement of south-dipping subduction along New Guinea to account for the Maramuni Volcanics (Fig. 24), as well as contemporaneous north-dipping subduction along the Halmahera Arc, which is terminated after ~15 Ma following the arc-continent collision on northern New Guinea. The collision of Dangerous Grounds-Reed Bank with northern Borneo at ~15 Ma also choked the north Borneo subduction zone, and likely resulted in Proto South China Sea slab breakoff. The ~15 Ma reconstruction using the faster sinking rate, and corresponding to a 791 km depth slice, shows a discrete slab volume along northern Borneo that we interpret as the Proto South China Sea slab (Fig. 24).

The seismic tomographic interpretation highlights that the refinement of the New Guinea margin (namely the Maramuni subduction zone) and the adjustment to the Philippine Sea Plate (namely the position of the Halmahera Arc) since ~30 Ma can accommodate both the geological and the tomographic constraints. Neither sinking rate scenarios produce consistent matches throughout the plate reconstruction timeframe, likely due to the complex time-varying slab sinking rates and regional interactions of slabs in a spherical mantle shell. However, the assumption of vertical sinking of slabs is likely to be an acceptable estimate of trench locations in the Cenozoic for slabs that are still attached to the subducting plate, or slabs that have experienced little stagnation or folding in the mantle transition zone. The numerical computations described in the following section provide a more consistent approach to tracking slabs in the mantle resulting from the complex subduction history in the Tethys, east and Southeast Asia, and New Guinea.

5 Numerical modelling results

5.1 Large-scale post-Jurassic mantle evolution of the Tethyan tectonic domain

We present the first synthesis of post-Jurassic Tethyan plate reconstructions and geodynamics in a 4D (space and time) global context. We ran five cases of coupled plate kinematic and geodynamic numerical experiments, mainly to test end-member plate reconstructions, and present 3D snapshots of two experiments that compare the Zahirovic et al. (2014) model with refinements for the Neo-Tethys, Philippine Sea Plate and New Guinea presented in this study (Fig. 25). Although the mantle convection models are initiated at 230 Ma during the time of Pangea stability, we present only the post ~160 Ma timeframe applicable to the refined plate reconstructions. At ~160 Ma, the dominant feature of the mantle is the circum-Pangea subduction girdle, as well as the southern Eurasian active margin consuming Meso-Tethyan oceanic lithosphere (Fig. 25a). The tectonic scenario presented invokes the northward continuation of East Gondwana subduction along New Guinea and connecting to the East Asian subduction of the Izanagi Plate. By ~140 Ma the rollback of the Lhasa trench opens the Kohistan-Ladakh back-arc basin, with a slower opening and southward position of ~10°N in our base model by 100 Ma (Zahirovic et al., 2014), compared with the equatorial position implemented in this study following Burg (2011) and Gibbons et al. (2015).

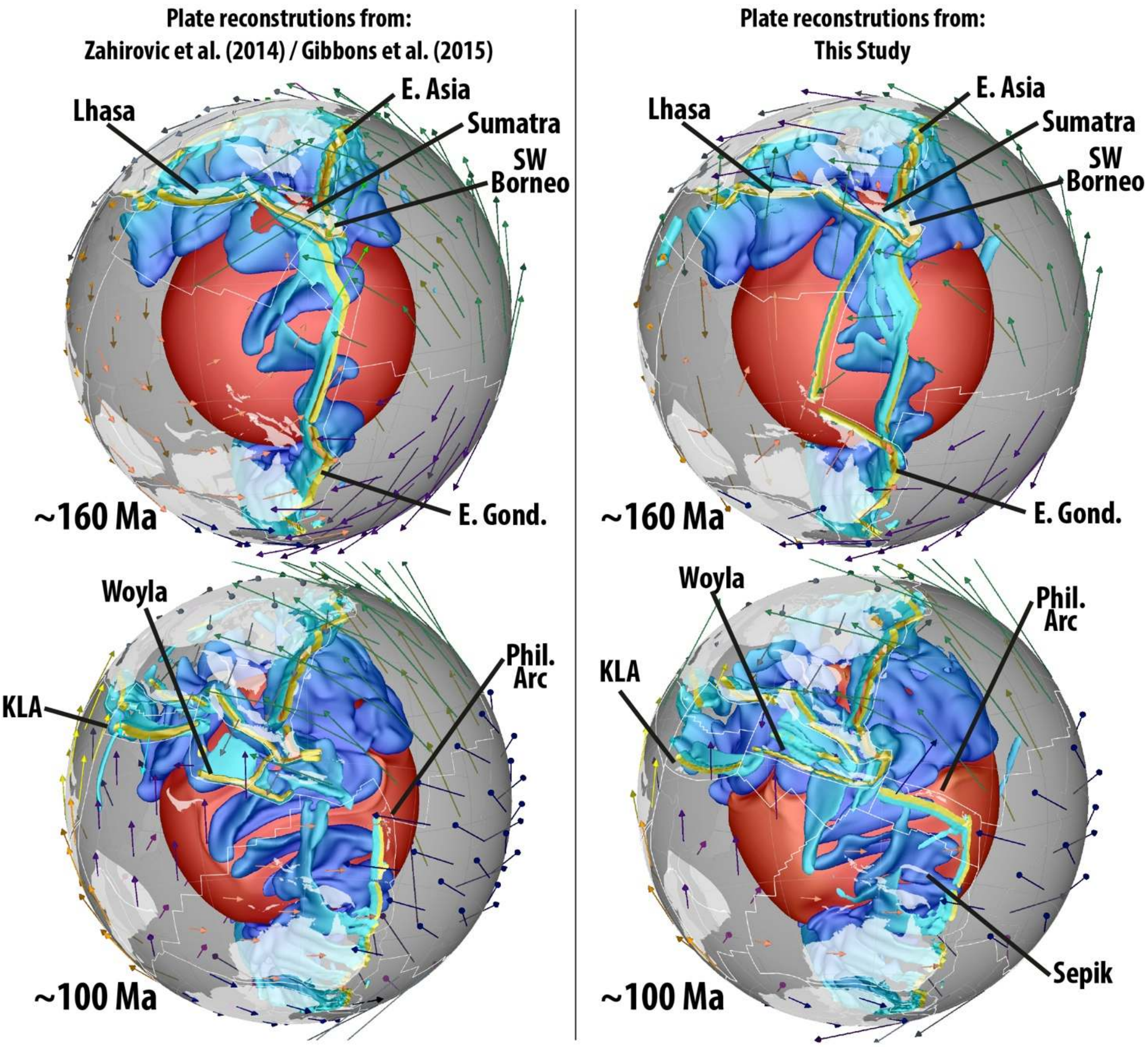


Fig. 25a. Snapshots of mantle structure including sinking slab volumes (blue) and thermochemical upwellings (red) from the core-mantle boundary are visualised in GPlates using the mantle temperature predictions from the CitcomS numerical experiments of mantle flow depicting Case 4

(left) and 5 (right). The snapshots compare the large-scale mantle evolution in the latest Jurassic and Early Cretaceous from Tethyan subduction along Lhasa and the Kohistan-Ladakh Arc (KLA), as well as recycling of proto-Pacific lithosphere along the East Asian, East Gondwana and Philippine Arc subduction zones. Plate boundaries (white), velocities (coloured arrows) and reconstructed present-day coastlines (translucent white) are plotted. Subduction zones are yellow regions, and slab colouring is a function of depth from light blue (shallow slabs) to darker blue (deep slabs). These snapshots highlight the global nature of our numerical experiments, with complex interactions of slabs as they sink in the mantle shell. The experiments allow us to track the sinking trajectory (vertical and lateral) of the slabs to identify their source from the present-day mantle prediction, which are compared to the mantle structure imaged using P- and S-wave seismic tomographic techniques. Central co-ordinate is 10°S, 115°E.

As our modelling domain is spherical, and because the flow is constrained to follow surface velocities that include net rotation of the lithosphere, lateral mantle flow may influence the trajectory of sinking slabs. As subducting slabs sink in the mantle, the core-mantle boundary becomes draped with older slabs that sweep the hotter material into the large-scale Pacific and African upwellings (Bower et al., 2013; McNamara and Zhong, 2005) (Fig. 25). In addition, mantle flow advects slabs laterally, with notable southward (and somewhat westward) translation of the Paleo-Tethyan slabs, and eastward advection of the east Asian slabs (Fig. 25). India's collision with the Kohistan-Ladakh Arc ceases intra-oceanic subduction by ~50 Ma in our model, resulting in the Andean-style subduction of the Kohistan-Ladakh back-arc basin along southern Lhasa (Fig. 25b). The ~47 Ma continent-continent collision temporarily shuts down subduction, causing a slab break-off event, followed by ongoing subduction of the Greater Indian mantle lithosphere (Capitanio et al., 2010). Australia's northward motion results in the northern margin, including New Guinea, overriding the Southeast Asian slab graveyard from ~30 Ma following the docking of Sepik, and the initiation of south-dipping Maramuni subduction from ~20 Ma.

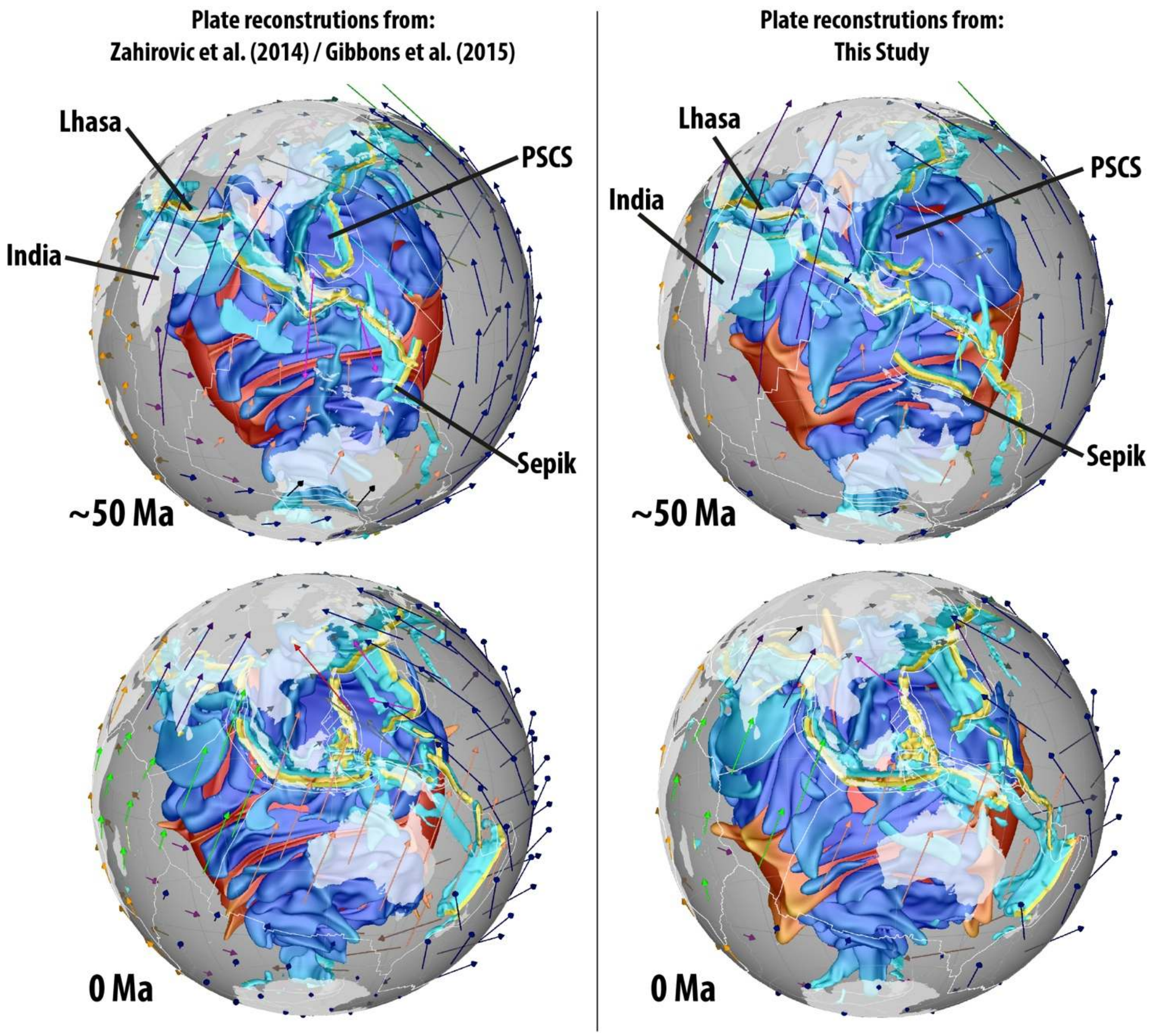


Fig. 25b. Coupled plate reconstructions and mantle flow models in the Eocene and present-day, highlighting the draping of subducted slabs along the core-mantle boundary and the self-organisation of the African and Pacific large-scale upwellings as a result of post-Pangea subduction. These models are interrogated regionally using vertical profiles in Figs. 26-29. Centre co-ordinate is 10°S, 115°E. See Supplementary Animation 7 consistent with right panels.

The large-scale evolution of slab sinking and lateral advection, as well as the evolution of the large-scale upwellings, can be depicted in 3D hemispherical views of the mantle, while regional cross-sections of the numerical experiments provide a more detailed approach to interrogating the spatio-temporal geodynamic evolution of key subduction zones from post-Jurassic plate reconstructions.

5.2 Regional interpretations of mantle evolution

To capture the detailed evolution of subduction, vertical cross-sections of the mantle are presented in a plate frame of reference (i.e., fixed to the overriding plate, Figs. 26-29). Such time-dependent sections help understand the sinking trajectory of subducted slabs, as well as interpreting sinking rates and lateral mantle flow resulting from the post-Jurassic plate reconstructions. For the geological reasoning underpinning the reconstructions, please refer to Section 3 and Table 3.

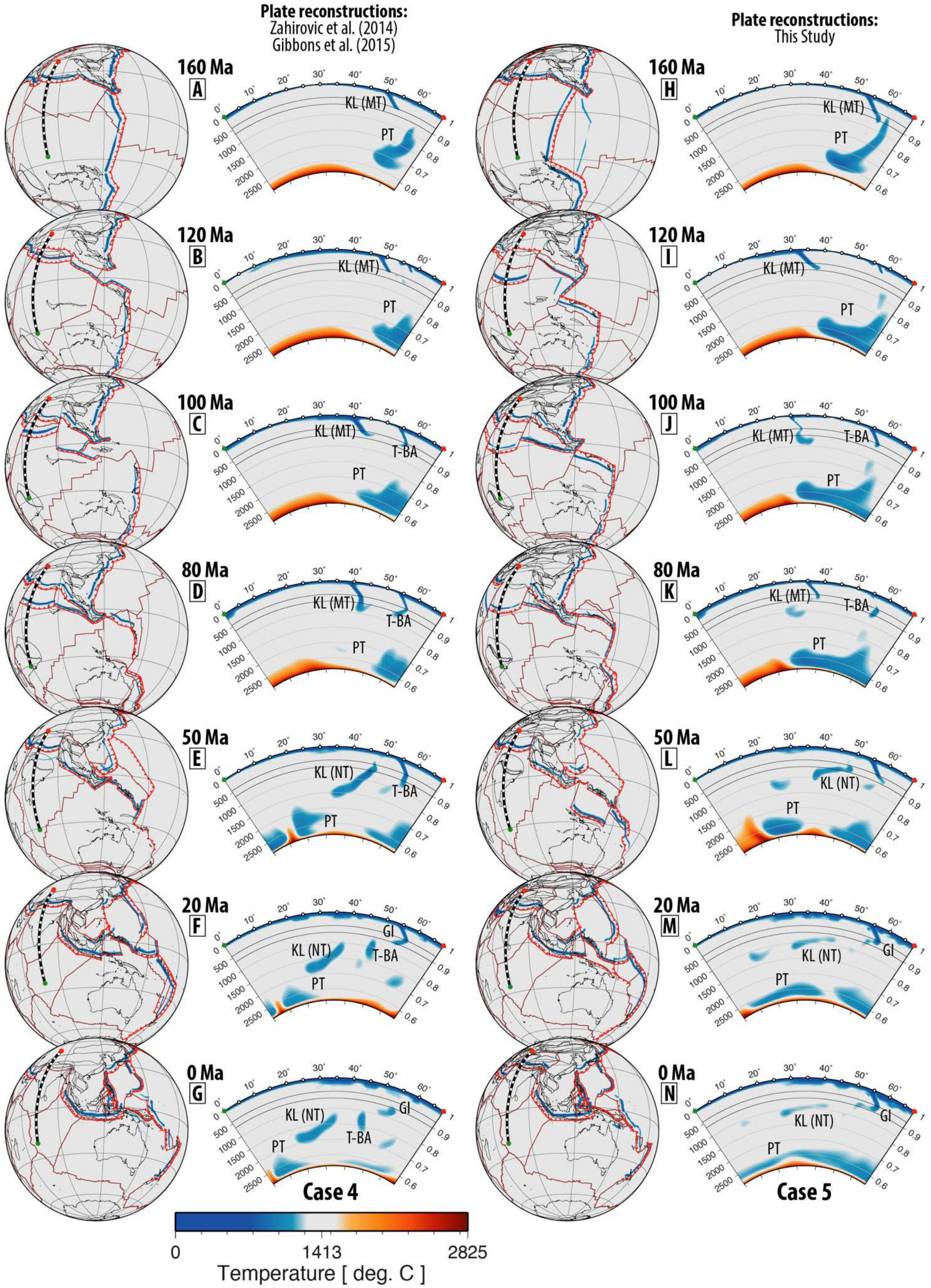
5.2.1 India-Eurasia convergence

The India-Eurasia segment is best represented by a largely north-south profile at present-day that is reconstructed with Lhasa (Fig. 26). At ~160 Ma, the Paleo-Tethyan slab has detached and is sinking through mid-mantle depths at ~1.5 cm/yr (Fig. 26a,h), while the Meso-Tethys is being actively consumed northward beneath Lhasa in both plate reconstruction scenarios. Both tectonic scenarios include southward slab rollback and the establishment of a Kohistan-Ladakh (Tethyan) back-arc basin, reaching to ~10°N in the base model (Zahirovic et al., 2014) and the equator at ~100 Ma in this study (Fig. 26c,j). Using the base reconstructions, the Meso-Tethyan slab only penetrates the mantle transition zone at ~100 Ma and begins sinking into the lower mantle by ~90 Ma (Fig. 26, Supplementary Animation 7). In the refined reconstructions the slab enters the mantle transition zone by ~120 Ma, and enters the lower mantle by ~110 Ma. This reflects greater convergence rates due to the combined effect of greater slab rollback in the refined reconstructions, as well as continued seafloor spreading north of India during this timeframe. In the base reconstructions, the seafloor spreading north of India is abandoned by ~120 Ma, leading to lower convergence rates across the Kohistan-Ladakh Tethyan trench (Fig. 10). This results in the subduction of larger volumes of older, and therefore thicker, oceanic lithosphere, while in the refined reconstructions subducted volumes along the Kohistan-Ladakh Arc system in the Early to mid-Cretaceous are smaller because the oceanic lithosphere associated with the Neo-Tethyan seafloor spreading north of India by ~100 Ma is younger and thinner (Fig. 26j). Once the Tethyan slab has entered the lower mantle, the sinking rate in the refined reconstruction is only ~1.4 cm/yr (between 100 and 89 Ma), while it is ~2.5 cm/yr in the base reconstructions (between 89 and 79 Ma) likely due to the larger subducted volumes.

The intersection of the Neo-Tethyan mid-oceanic ridge with the Kohistan-Ladakh subduction zone in the mid-Cretaceous would likely lead to slab breakoff and the formation of a slab window. However, our model does not capture the complexity of a subduction hiatus that would be associated with a slab window along Kohistan-Ladakh in the mid-Cretaceous. Perhaps due to the arrival of buoyant oceanic crust at the intra-oceanic subduction system, north-dipping subduction becomes established along Lhasa and begins to consume the Kohistan-Ladakh back-arc basin, eventually resulting in two Late Cretaceous north-dipping subduction zones in the Neo-Tethys (see Section 3.2). The mid-ocean ridge from the Kohistan-Ladakh back-arc is subducted in the Late Cretaceous in both reconstruction scenarios, with no interruption in subduction assumed in the base reconstructions. In the alternative reconstructions we impose a subduction hiatus along Lhasa from 80 to 65 Ma, which leads to slab breakoff. This slab window may be linked to adakitic volcanism at ~80 Ma (Wen et al., 2008a), followed by a ~75-60 Ma magmatic gap, in the Gangdese Batholith (Chung et al., 2005; Ji et al., 2009; Wen et al., 2008b).

In both reconstruction scenarios, Greater India collides with Kohistan-Ladakh by ~50 Ma, inducing Neo-Tethyan slab break-off at ~5-10°N. Since the Kohistan-Ladakh Arc is at equatorial latitudes in the refined reconstructions in the mid-Cretaceous, one may expect the collision with India to occur by ~60 Ma. As the magmatic chemistry change is much later, at 52 Ma (Bouilhol et al., 2013), the model includes some advance of the intra-oceanic subduction system between ~60 and 52 Ma. In the base reconstructions, the Kohistan-Ladakh Arc is closer to Eurasia at precollision times, meaning that relatively little trench advance is required. However, in both instances, the Tethyan slab is anchored in the lower mantle, leading to India overriding the sinking slabs. Andean-style subduction of the Kohistan-Ladakh back-arc along southern Lhasa is temporarily shut off by the ~47-40 Ma continent-continent collision, after which subduction of Greater India (continental) mantle lithosphere continues to present-day in the refined reconstructions. The Meso- and Neo-Tethyan slabs are predicted at present to be approximately at mid-mantle depths (~1000 to 2000 km), with a latitudinal range of ~0 to 35°N, using the base model plate reconstructions (Fig. 30, IND-EUR). More generally, the base plate reconstruction in modelled Case 4 reproduces a number of discrete slabs at mid-mantle depths, with a large latitudinal range, which is consistent with the interpretations of the mantle structure by van der Voo et al. (1999b) (Fig. 5).

Although the numerical model that uses the refined reconstructions presented in this study has the same radial viscosity profile, the Meso- and Neo-Tethyan slabs are predicted to be shallower at ~800 to 1500 km depths, as opposed to ~1000 to 2000 km depths predicted using the base reconstructions (Fig. 26g,n). This is likely due to the subduction of smaller volumes of younger Tethyan oceanic lithosphere north of India in the Cretaceous, resulting in less negative buoyancy. In addition, the required trench advance also results in folding of the Tethyan slab at the mantle transition zone, leading to generally shallower penetration into the lower mantle. In the post-collision timeframe, the Tethyan slab sinks at a rate of ~0.65 cm/yr (from 38 to 0 Ma) using the base reconstructions, and ~0.25 cm/yr (from 39 to 0 Ma) using the refined reconstructions where the slab is almost stagnant at ~1000 km depth (Supplementary Animation 7). The time-varying sinking rates in the lower mantle highlight the role active subduction has in adding negatively buoyant slab volumes into the mantle, and the role of thermal diffusion of slabs in reducing negative buoyancy of subducted lithosphere. A slightly deeper depth range (~1000 to 2000 km) provides a better match to the equivalent P- and S-wave tomography slice (Fig. 30, IND-EUR), as is obtained with the base plate reconstructions (Gibbons et al., 2015; Zahirovic et al., 2014), with the potential that slabs may extend further south of the equator based on the S-wave model.



1690 **Fig. 26.** Time-dependent evolution of the India-Eurasia convergence zone, with a representative

1691 vertical slice reconstructed with Lhasa to capture the evolution of the Kohistan-Ladakh (KL) and

Tethyan (T-BA) intra-oceanic subduction zones modelled in Zahirovic et al. (2014) and Gibbons et al. (2015) (left), and compared to the subduction histories implied in the revised plate reconstructions presented in this study (right). The cross-sections depict the temperature field from the numerical mantle flow models, and the globes show the position of the vertical slices through time, the plate reconstruction and the predicted mantle temperature field at ~400 km depth. The background mantle temperature is ~1413°C, and the small tick marks on the temperature scale represent temperature intervals of 250°C. Great circle angular distance along the vertical profile is shown on the x-axis. The left y-axis represents depth in kilometres, and on the right represents non-dimensional Earth radius. The plate reconstructions are plotted in an Orthographic projection with centre co-ordinate of 15°S, 115°E. GI – Greater India (continental) mantle lithosphere, MT – Meso-Tethys slab, NT – Neo-Tethys slab, PT – Paleo-Tethys slab. See Supplementary Animation 8.

5.2.2 Woyla and Sumatra active margin evolution

The Sumatra segment of the Sunda margin accommodates northward subduction of the Meso-Tethys in the Late Jurassic, with rollback of the slab opening the Woyla back-arc to near-equatorial latitudes (Fig. 27), similar to the development of the Kohistan-Ladakh Arc further to the west. In the base reconstructions, the rollback imposed is faster and the maximum southward extent of subduction is ~0-10°S (Fig. 11). This leads to a smaller volume of subducted slabs folded in the mantle transition zone. Although the base reconstruction maintains convergence across the Woyla subduction zone, there is significant trench advance between ~100 and 75 Ma, leading to a similar smearing effect of slabs in the transition zone. Although trench advance occurs at present-day along the Izu-Bonin-Mariana Trench (Becker et al., 2015; Carlson and Mortera-Gutiérrez, 1990; Mathews, 2014), the modelled values in our base reconstructions are likely excessive, with a more geodynamically reasonable evolution of trench migration in the refined reconstructions.

Subduction continues along Sumatra to consume the Woyla back-arc basin, and is interrupted for ~10 Myr between accretion of the Woyla Arc onto the Sumatran margin between ~75 and 65 Ma in the base reconstructions. Due to the convergence required between the Tethyan-Indian Ocean and Eurasia, we impose a shorter ~5 Myr hiatus in subduction between 75 and 70 Ma to induce slab breakoff that may have occurred due to Woyla Terrane accretion. Our assumption of slab breakoff is simplistic and based on the magmatic gap, and more realistic slab breakoff timings of 5-10 Myr after collision (Li et al., 2013; van Hunen and Allen, 2011) will need to be considered in future refinements of the model in this region.

In both tectonic reconstructions, subduction at the Sunda Trench is initiated at ~70-65 Ma, and persists to present-day. The slab is predicted to have penetrated the lower mantle by ~50 Ma, after which the collision of India with Eurasia, and subsequent rotation of Indochina and much of Sundaland from ~30 Ma leads to a kink in the slab in the mantle transition zone (410 to 660 km, Fig. 27g,n). Although the kink results from the constant slab dip imposed in the slab assimilation, this slab kink is imaged by the P- and S-wave seismic tomography analysed here, and also recently discussed in Hall and Spakman (2015). The numerical experiments of mantle flow also reproduce the latitudinal range of the subducted slab (Fig. 30, SUM), as well as a gap in the slab at depths greater than ~1500 km, consistent with earlier interpretations of the Sunda slab (Widiyantoro and van der Hilst, 1996). The mantle convection models predict the Woyla/Meso-Tethys slab at ~1500 to 2000 km depths at ~10°S along the Sumatran vertical slice (Fig. 30, SUM), and ~20°S along the Java-Borneo Sundaland slice (Fig. 30, SUN), which is likely to be equivalent to fast seismic velocities in P- and S-wave tomography at ~1500-2000 and ~1200-1600 km depth along the Sumatran and Java-Borneo vertical slices, respectively.

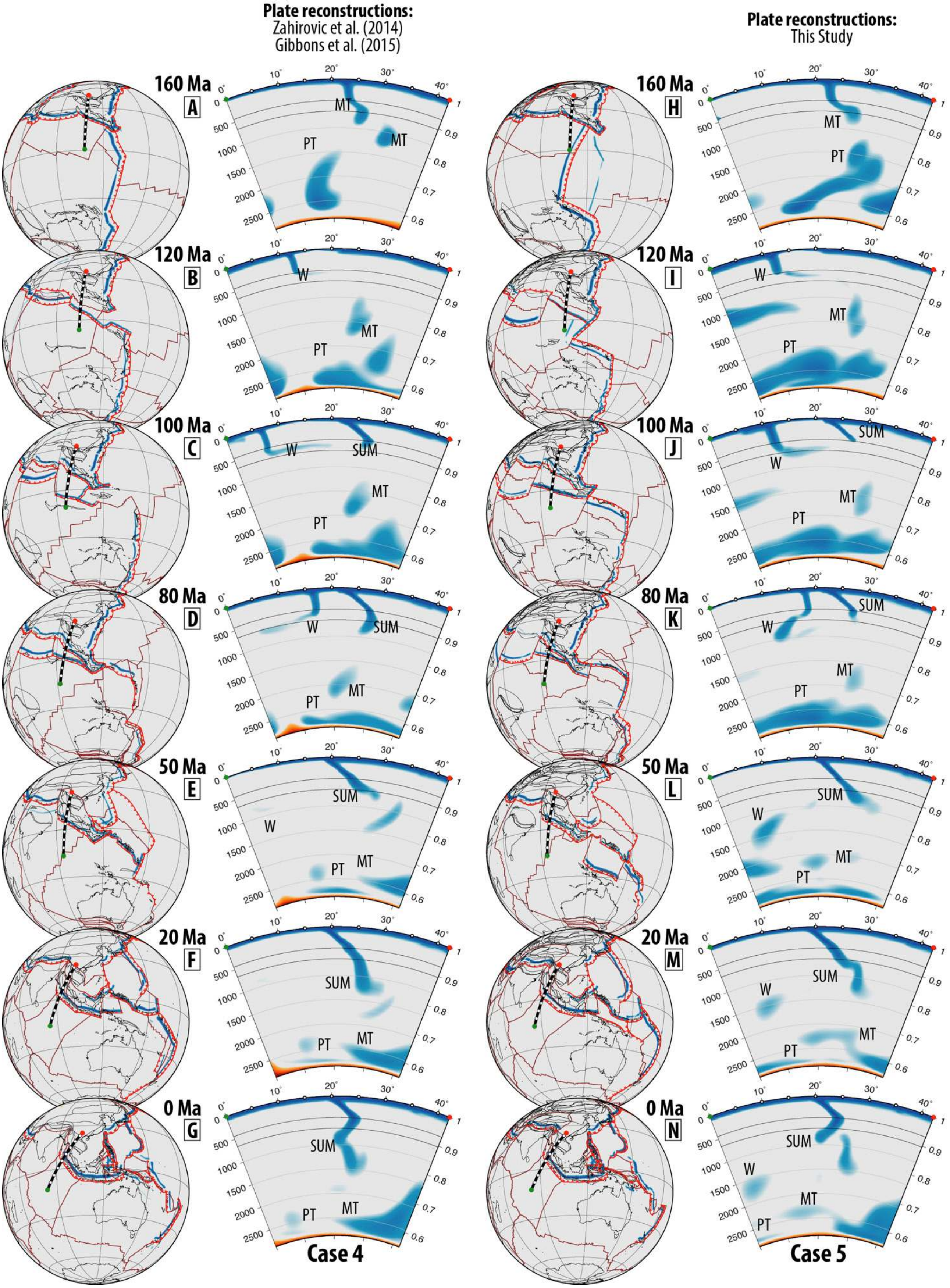
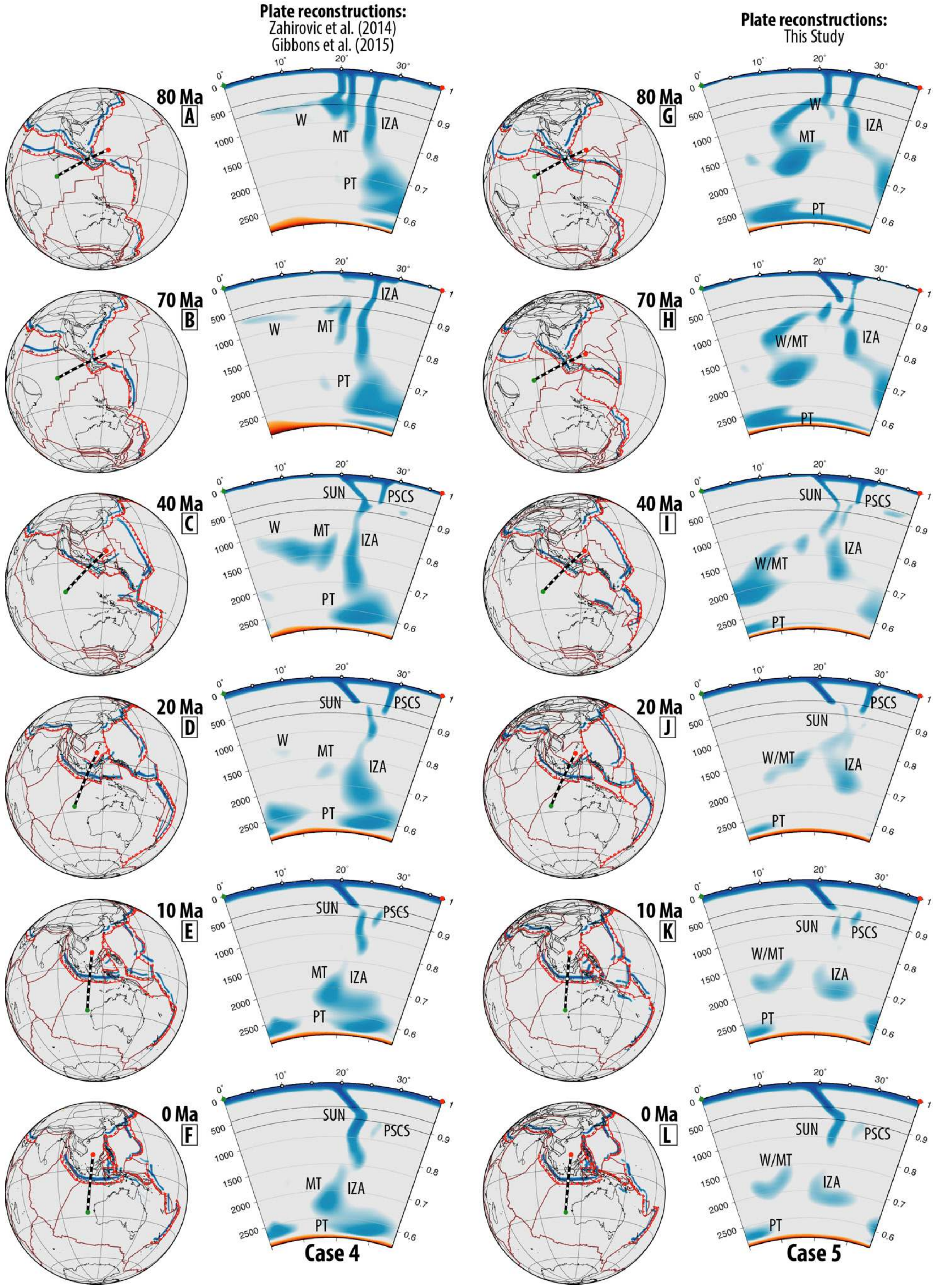


Fig. 27. Reconstructed vertical profiles across northwest Sumatra highlighting the evolution of the Woyla intra-oceanic and Sunda subduction zones through time, with both numerical experiments

predicting a significant kink in the Sumatran portion of the Sunda slab (SUM) when the slab dip is held constant during the clockwise rotation and extrusion of Indochina. MT – Meso-Tethys slab, NT – Neo-Tethys slab, PT – Paleo-Tethys slab, W – Woyla slab. See Supplementary Animation 9.

5.2.3 Java and Borneo subduction history

Similar to the Sumatra margin, the Java segment of the Sunda Trench accommodates subduction of the Meso-Tethys and the Woyla back-arc basin during the Cretaceous (Fig. 28). However, as this segment represents the Sundaland continental promontory, south-dipping subduction of the Izanagi Plate is contemporaneous to the Tethyan subduction history. As a result, the mid- and lower mantle slabs are likely to be a mixture of Pacific- and Tethyan-derived slabs. The accretion of the Woyla Arc temporarily shuts off subduction in this segment in the Late Cretaceous at ~70 Ma in the base reconstructions (Fig. 28b,h), followed by the accretion of the Semitau continental fragment and Proto South China Sea Arc onto northern Borneo at ~45 Ma in both reconstruction scenarios (Fig. 28c,i). The late Eocene is dominated by renewed north-dipping Sunda subduction and south-dipping subduction of the Proto South China Sea. Although the Sunda subduction continues to the present-day, the Proto South China Sea subduction is interrupted at ~15 Ma with the docking of the Dangerous Grounds-Reed Bank continental fragment along northern Borneo, which leads to the abandonment of the South China Sea seafloor spreading. The refined reconstructions imply a longer-lived Meso-Tethyan Plate that is completely consumed by ~45 Ma, leading to much younger oceanic crust and thinner oceanic lithosphere subducted at the Sunda Trench than in the base reconstructions. This leads to the subduction of smaller slab volumes between ~60 and 30 Ma for the refined reconstructions that predict a smaller and shallower slab that penetrates to ~1200 km depth at present. In contrast, the base reconstructions lead to a larger Sunda slab at depths of ~1500 km (Fig. 30, SUN), which is consistent with the interpretations of P- and S-wave seismic tomography. The kink in the slab observed in the Sumatra segment (Fig. 30, SUM) is much less pronounced in the Java region (Fig. 30, SUN), especially when compared to the results using our base plate reconstruction. A gap in the slab is also reproduced for depths greater than ~1500 km, with older Tethyan and Izanagi slab fragments reproduced near the core-mantle boundary when comparing to the S-wave seismic tomography (Fig. 30, SUN). The Proto South China Sea slab is predicted at ~600-1000 km depths, while P- and S-wave tomographic images indicate a slab stagnating at the base of the 410-660 km mantle transition zone.



1776

1777 **Fig. 28.** Reconstructed vertical slice through eastern Sundaland, capturing the subduction of the

1778 Meso- and Neo-Tethyan, as well Indian Ocean, basins. The Sunda slab is predicted to reach a maximum depth of ~1500 km along southern Sundaland in the base reconstructions, and ~1200 km in the refined reconstructions, while a small Proto South China Sea slab is predicted just beneath the 660 km upper-lower mantle transition. See Supplementary Animation 10.

5.2.4 New Guinea margin evolution

Further east along the New Guinea Tethyan segment, the Early Cretaceous Sepik oceanic basin is consumed at a north-dipping subduction zone from ~40 Ma (Fig. 29b) in the base model reconstructions, while in our refined model subduction starts earlier at ~71 Ma (Fig. 29g) to account for the ~71 to 66 Ma Emo volcanics (Worthing and Crawford, 1996) that likely formed in the back-arc of this subduction system. North-dipping subduction along the Sepik Terrane is interrupted at ~30 Ma in both plate reconstruction scenarios, based on the timing of docking of the composite terrane at the New Guinea margin. We impose slab breakoff during the collision, leading to a slab that is entrained in the upper part of the lower mantle (660-1000 km depths) for both reconstruction scenarios. In the base model, north-dipping subduction is then accommodated along the Halmahera Arc, which forms the southern boundary of the Caroline Plate, and is accreted to the New Guinea margin diachronously from west to east by ~5 Ma. In the refined reconstructions, a south-dipping subduction zone is implemented (Fig. 29j) to account for the ~18 to 8 Ma Maramuni Arc volcanics (Hill and Hall, 2003; Page, 1976), as well as simultaneous north-dipping subduction along the Halmahera Arc. Both subduction zones are abandoned progressively from ~15 Ma, resulting from the collision of the Halmahera Arc with the New Guinea margin.

The numerical experiments of mantle flow assimilating the base plate motion model predict two slabs at depths between ~500 and 1000 km (Fig. 30, PNG), with the southernmost slab at ~20°S belonging to the Sepik oceanic basin, and the northern slab at 0 to 5°S resulting from the subduction of the Solomon Sea along the Halmahera Arc. With the addition of the Maramuni subduction zone in the refined plate reconstructions (~20 to 10 Ma, Fig. 29), an additional slab is predicted at slightly deeper depths of ~900 to 1500 km at ~15°S. The Sepik oceanic basin slab is predicted to be further south in the refined plate reconstruction scenario, at depths of ~700 to 1000 km and latitude of ~30°S (Fig. 30, PNG). This difference in latitude is largely due to the earlier onset of Sepik oceanic gateway subduction at ~71 Ma in the refined reconstructions, leading to the slab entering the lower mantle at more southerly latitudes, as opposed to the younger age of ~40 Ma subduction onset using the base reconstructions (Fig. 29b). The refined plate reconstructions result in a much better fit with the mantle structure than the base model, with the ~30°S position of the Sepik oceanic gateway slab corresponding to the fast seismic anomaly interpreted beneath Lake Eyre in eastern Australia, recently interpreted in Schellart and Spakman (2015). However, a slab at ~1300 to 1800 km depths at present-day (PNG, Fig. 30), and latitudes between ~10°S and the equator, is not accounted for in either model of mantle flow – suggesting the Cretaceous plate reconstruction needs additional refinement.

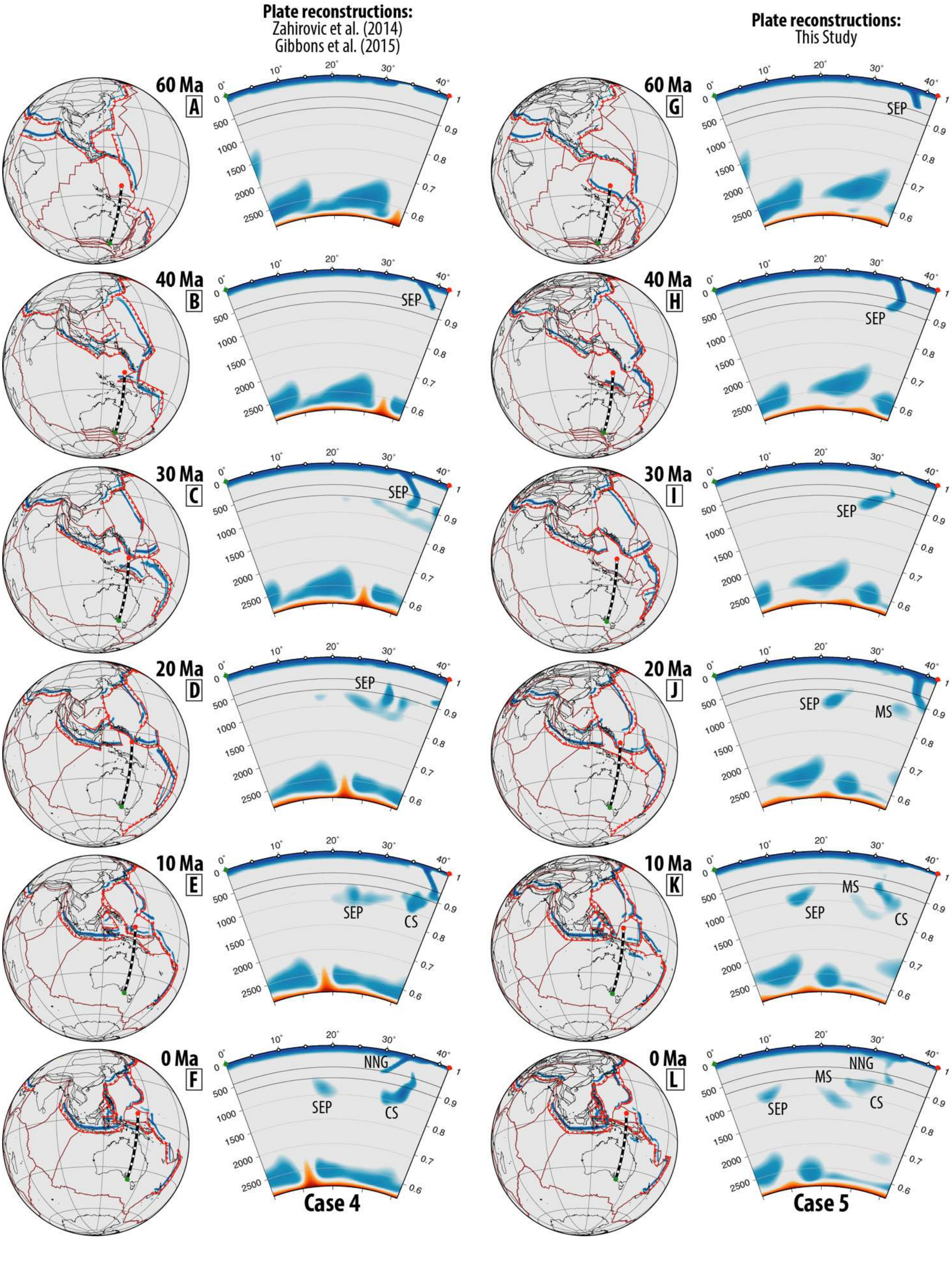
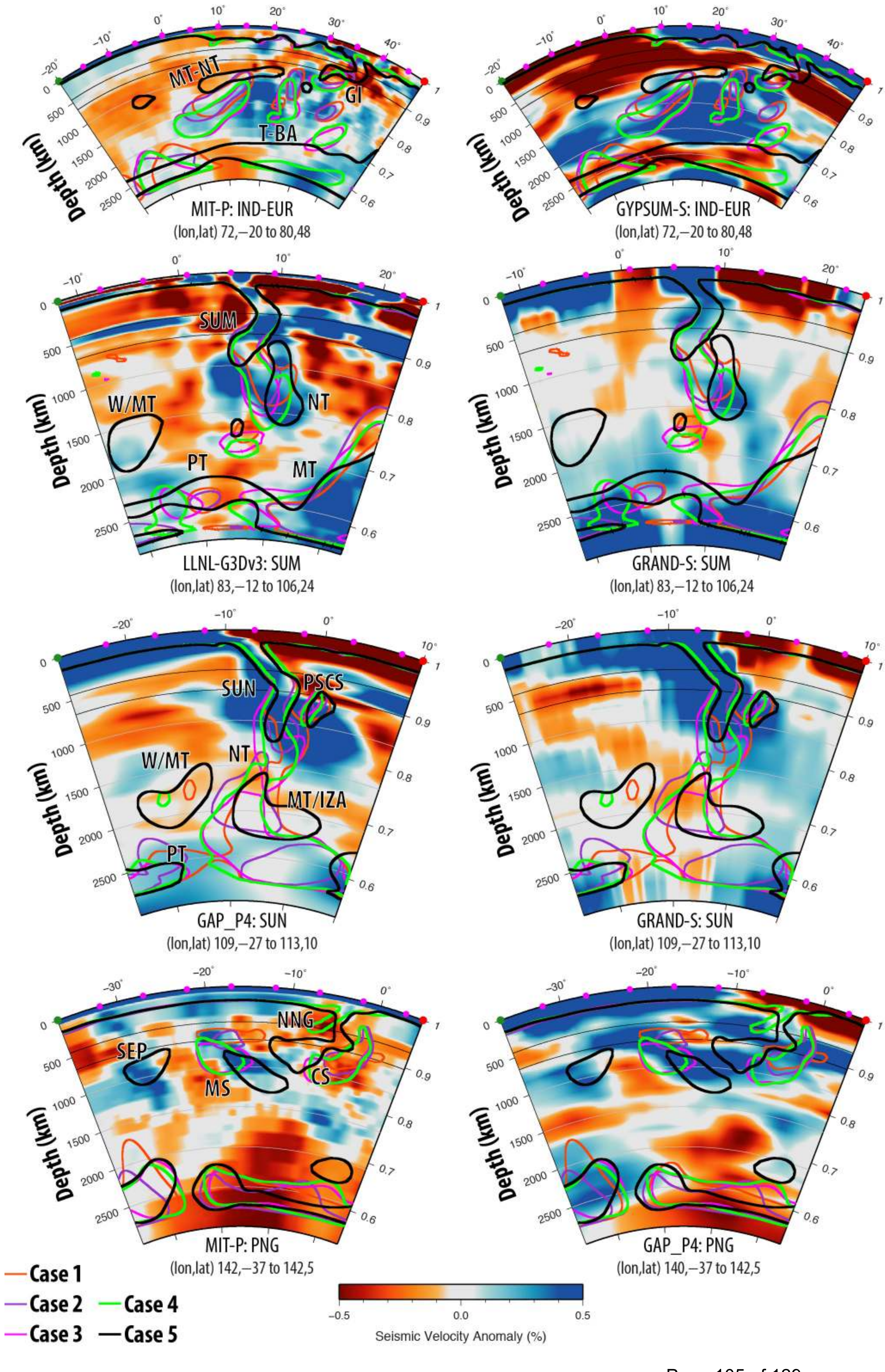


Fig. 29. Reconstructed representative profile through Australia and New Guinea, highlighting that the revised plate reconstructions account for additional slab volumes above mid-mantle depths. The southernmost slab is related to the subduction of the Sepik oceanic gateway (SEP) between ~71 and

30 Ma, while Maramuni subduction (MS) has taken place since ~20 Ma, coeval with north-dipping subduction along the Halmahera Arc to produce the Caroline/Proto Molucca slab (CS). See Supplementary Animation 11.


MT-NT
GI
T-BA
MIT-P: IND-EUR
(lon,lat) 72,−20 to 80,48
GYPSUM-S: IND-EUR
(lon,lat) 72,−20 to 80,48
SUM
W/MT
NT
PT
MT
LLNL-G3Dv3: SUM
(lon,lat) 83,−12 to 106,24
GRAND-S: SUM
(lon,lat) 83,−12 to 106,24
SUN
PSCS
W/MT
NT
MT/IZA
PT
GAP_P4: SUN
(lon,lat) 109,−27 to 113,10
GRAND-S: SUN
(lon,lat) 109,−27 to 113,10
NNG
SEP
MS
CS
MIT-P: PNG
(lon,lat) 142,−37 to 142,5
GAP_P4: PNG
(lon,lat) 140,−37 to 142,5
Depth (km)
Case 1
Case 2
Case 3
Case 4
Case 5
Seismic Velocity Anomaly (%)

1826

Fig. 30. Present-day mantle structure along vertical slices through India (IND-EUR), northwest Sumatra (SUM), eastern Sundaland (SUN) and Australia-New Guinea (PNG) using P- and S-wave seismic tomography models, superimposed with the predicted slabs (coloured lines) from five computations of mantle flow. Case 1 to 4 uses the Zahirovic et al. (2014) plate reconstruction, but varies the radial viscosity profile of the mantle. Case 5 uses the plate reconstruction presented in this study, and the preferred viscosity structure used in Case 4. Slabs from the numerical models are defined as regions 10% colder than the background mantle temperature. The P-wave seismic tomographic models used are the MIT-P (Li et al., 2008), GAP_P4 (Obayashi et al., 2013) and LLNL-G3Dv3 (Simmons et al., 2015). Both P- and S-wave models from Simmons et al. (2010) are used, as well as the S-wave model from Grand (2002). The x-axis of the vertical section represents latitude. Table 4 lists the differences between Cases 1 to 5. The start and end coordinates for each profile are included on each cross-section, and plotted geographically in Figs. 26-29 and Supplementary Fig. 2.

6 Discussion

We have demonstrated the strength of using coupled plate tectonic reconstructions and numerical models of mantle flow to test competing kinematic scenarios in the absence of preserved seafloor spreading histories. In addition, the global nature of the models removes the edge effects associated with Cartesian box models of mantle convection, and allows us to track the origin and trajectory of sinking slabs, and therefore their sinking rates, which can then be compared to the mantle structure interpreted from P- and S-wave seismic tomography (Fig. 30).

6.1 Intra-oceanic subduction in the Meso- and Neo-Tethys

In the India-Eurasia segment of the Tethyan margin, a number of important geodynamic implications arise from the Neo-Tethyan seafloor spreading history and the evolution of intra-oceanic subduction zones along southern Eurasia. Although early plate reconstructions of the Tethys incorporated intra-oceanic subduction and an initial collision between Greater India and the Kohistan-Ladakh Arc at or before ~53 Ma (Patriat and Achache, 1984), this two-stage India-Eurasia collision scenario was abandoned based on subsequent work that argued that Kohistan and Ladakh first collided with Eurasia in the Late Cretaceous along the Shyok Suture Zone (Clift et al., 2002; Debon et al., 1987; Treloar et al., 1996). However, recent work requires near-equatorial position of the Kohistan-Ladakh Arc in the Late Cretaceous (Burg, 2011; Chatterjee et al., 2013; Zaman et al., 2013; Zaman and Torii, 1999), and an initial arc-continent collision between Greater India and the Neo-Tethyan intra-oceanic arc sometime between ~60 and 50 Ma (Aitchison et al., 2007; Bouilhol et al., 2013; Khan et al., 2009). Our results favour a ~60 Ma arc-continent collision if the near-equatorial paleo-latitudes of Kohistan-Ladakh are robust, as a younger collision requires more significant advance of the Kohistan-Ladakh intra-oceanic trench before Greater India enters the subduction zone, terminating the subduction of oceanic lithosphere. The Kohistan-Ladakh back-arc basin was then subducted along southern Lhasa at an Andean-style margin, with final Shyok suturing occurring by 40 Ma (Bouilhol et al., 2013; Gibbons et al., 2015; Zahirovic et al., 2014).

The geodynamic implications of a well-established intra-oceanic system in the Neo-Tethys, suggest a scenario that is much like the present-day Izu-Bonin-Mariana Arc in the west Pacific. The paleo-latitudinal position of the intra-oceanic arc largely determines the timing of Neo-Tethyan ridge subduction, as well as the plate driving forces acting on the Indian Plate. In our reconstruction, the Neo-Tethyan mid-oceanic ridge is consumed at the Kohistan-Ladakh subduction zone from ~105 Ma (Figs. 11 and 26c,j). Subduction of the southern Neo-Tethyan flank of the spreading system from ~100 Ma would have been associated with progressively strengthening northward slab pull, to which we attribute the change towards largely northward convergence with

Eurasia that is best represented by the ~110-90 Ma fracture zone bends in the Wharton Basin (Gibbons et al., 2015; Matthews et al., 2011; Matthews et al., 2012).

The role of two coeval north-dipping subduction zones in the Neo-Tethys (Fig. 26) were suggested to have contributed to the ~80 Ma acceleration of India in Gibbons et al. (2015), which has recently been proposed as a mechanism for India's rapid northward advance using numerical techniques quantifying plate driving forces in Jagoutz et al. (2015). Although the arrival of the Reunion Plume head south of India at ~65 Ma possibly played a role in India's acceleration (Cande and Stegman, 2011; van Hinsbergen et al., 2011), the effects were likely short-lived (~5-10 Myr), and post-date by 15 Myr the initial acceleration of India from ~80 Ma. India's northward acceleration resulting from greater slab pull (and slab suction) may have induced stronger large-scale mantle return flow, possibly triggering the ascent of the Reunion Plume from the margin of the lower mantle African super-swell (Fig. 25). The recent data compilations of the surface geology, as well as new plate reconstructions and numerical approaches, suggest that a two-stage collision between India and Eurasia is more likely than a single continent-continent collision, and that the Tethyan tectonic evolution was punctuated by generations of back-arc basins and intra-oceanic subduction systems more similar to the present-day West Pacific, than a simpler long-lived Andean-style margin.

6.2 Southeast Asia and New Guinea

Southeast Asia, and in particular Sundaland and New Guinea, played an important role in the convergence history of Australia, Eurasia and the Pacific. In this study we have shown that our plate reconstructions are compatible, at least to the first-order, with fast seismic anomalies imaged by seismic tomography. In particular, the numerical methods suggest that the extrusion and clockwise rotation of Indochina from ~30 Ma is likely responsible for the Sunda slab kink beneath west Sumatra. The models reproduce the depth of the Sunda slab beneath Sumatra and Borneo, which supports subduction initiation from ~65 Ma, rather than from ~45 Ma (Hall, 2012). If the 1600 km deep Sunda slab represents subduction since 45 Ma (Hall and Spakman, 2015), then an average whole-mantle sinking rate of 3.5 cm/yr is required, while our post-65 Ma subduction history would require average sinking rates of 2.5 cm/yr, which is more consistent with previous studies of sinking rates in numerical models (Butterworth et al., 2014; Steinberger et al., 2012). The constraints from the subduction-related volcanic history of Sumatra (McCourt et al., 1996) result in a predicted slab that is consistent with P- and S-wave tomography. This highlights that segmentation of the Neo-Tethyan and Indian Ocean plates across pre-existing structural fabric (Hall, 2012) is not required to account for the subduction history recorded on the Sumatra-Java Sundaland margin. Although Hall and Spakman (2015) invoke a leaky transform in the Neo-Tethys at ~90°E (Fig. 4c) to explain a possible ~90-45 Ma subduction hiatus, the mantle discontinuity linked to this interpretation is much further east at 110°E. There is a clearer discontinuity in slab structure east of ~120°E (Fig. 24), which represents the complex subduction history of New Guinea that is possibly linked to the evolution of the Philippine Sea Plate and the Pacific, rather than Sundaland.

The relatively small slabs predicted at ~600 to 1000 km depth beneath northern Borneo in our models roughly correspond to the interpreted Proto South China Sea slab (Zahirovic et al., 2014) imaged in seismic tomography at shallower depths in the mantle transition zone (~410 to 660 km). This suggests that Proto South China Sea subduction along northern Borneo may have started later than 45 Ma, which would be consistent with a shallower slab, and perhaps linked to a ~32 Ma onset in seafloor spreading of the South China Sea (Briais et al., 1993). However, a major phase of volcanism along northern Borneo from ~50 Ma (Soeria-Atmadja et al., 1999) might instead indicate earlier subduction initiation. In this case, stagnation of the Proto South China Sea slab in the mantle transition zone could play an important role in the depth mismatch between our numerical experiments and the seismic tomographic constraints. Recent backward-advection modelling by Yang et al. (2016) suggests that the large volume of subducted slab beneath

Sundaland stagnated in the mantle transition zone before ~30 Ma, and entered the lower mantle as a slab avalanche in the Miocene from ~20 Ma. This work highlights the time-varying slab sinking rates in the region, but more importantly, demonstrates that a slab avalanche resulted in dynamic subsidence of Sundaland and flooding (Yang et al., 2016) that was asynchronous with global eustasy (Haq et al., 1987). In addition, the slab avalanche was likely responsible for Miocene basin inversions (Doust and Sumner, 2007) by propagating stresses acting on the lithosphere. Since Proto South China Sea subduction ceased at ~15 Ma, recorded by cessation of seafloor spreading in the South China Sea (Briais et al., 1993), it is therefore likely that the Proto South China Sea slab is in the upper mantle or transition zone when considering the role of slab stagnation in this region. This interpretation is in contrast with that of Hall and Spakman (2015) who argued for a lower mantle (~1200 km deep) position of the Proto South China Sea slab.

Further east on New Guinea, the complexity of the surface geology has led to competing plate tectonic reconstruction scenarios (van Ufford and Cloos, 2005), some of which are discussed in this study. Our plate reconstructions and numerical experiments of mantle flow require Sepik oceanic gateway subduction in the Late Cretaceous, likely from ~70 Ma, which accounts for the present-day slab at mid-mantle depths at ~30°S beneath Lake Eyre in northern South Australia, consistent with recent interpretations (Schellart and Spakman, 2015). However, the slab our mantle flow model predicts is smaller, which raises the possibility that the Sepik oceanic basin was larger than modelled in our plate reconstructions. The refinement to the plate reconstructions and inclusion of south-dipping Maramuni Arc subduction along New Guinea from ~20 to 6 Ma, with coeval north-dipping subduction along the Halmahera Arc, improves the fit between predicted slab distributions and the mantle structure inferred from seismic tomography. We interpret the presently-inactive Trobriand Trough as the Maramuni Arc subduction zone (active ~18 to 8 Ma). The Maramuni subduction may have caused the dynamic subsidence and progressive flooding inferred for the northern Australian shelf since the Oligocene (DiCaprio et al., 2009; DiCaprio et al., 2011; Sandiford, 2007; Spasojevic and Gurnis, 2012). Although our results reproduce the Sepik and

Maramuni slabs, more work is required to account for a consistently-imaged near-equatorial slab at ~1500 km depths that is not reproduced by either of the plate reconstruction scenarios presented in this study. Further work using numerical techniques is required to improve the understanding of the complex tectonic linkage between Southeast Asia and the Pacific through New Guinea.

6.3 Relevance to global plate reconstructions and geodynamics

Our coupled plate kinematic and numerical geodynamic approach has wider implications for understanding the long-term evolution of the plate-mantle system. One important outcome is that numerical models testing alternative plate reconstruction scenarios that are compared to mantle structure from seismic tomography should consider the regional and global plate tectonic evolution. The evolution of Neo-Tethyan intra-oceanic subduction along the Kohistan-Ladakh and Woyla arc systems also has wider geodynamic implications. Although we implemented subduction initiation at the passive margin of the back-arc systems, a more complicated geodynamic mechanism may be required, such as the inversion of a mid-oceanic ridge to become a subduction zone in order to accommodate convergence and explain ophiolite obduction (Hébert et al., 2012; Shemenda, 1993). Due to the paucity of data constraining the nature and location of subduction initiation of the Woyla back-arc basin, a south-dipping subduction zone as proposed by Morley (2012a) will also require testing in future work. However, India's Late Cretaceous northward acceleration from two coeval and coupled north-dipping subduction zones (Jagoutz et al., 2015) may also require two north-dipping subduction zones in the Woyla segment of the Neo-Tethyan active margin, which highlights the prevalence of intra-oceanic subduction in the Neo-Tethys.

7 Conclusions

This study shows the power of considering the coupled plate-mantle system to study the geodynamics of the Tethyan tectonic domain that is dominated by long-term Eurasian, Indo-Australian and Pacific convergence following Pangea breakup. The reconstructions, used as boundary conditions in mantle flow simulations, consider intra-oceanic subduction along the entire south Eurasian active margin from ~160 Ma in the Neo-Tethys. We suggest that the Neo-Tethyan ridge was likely consumed along the Kohistan-Ladakh intra-oceanic arc from ~105 Ma, followed by northward subduction of the Indian Plate that significantly modified India's plate motion direction. For Sundaland, a tectonic scenario with Woyla Arc accretion at ~75-70 Ma, followed by a ~10 Myr subduction hiatus, and renewed subduction along the south Sundaland margin by ~60 Ma places the Sunda slab at the same depth as in P- and S-wave seismic tomography. In addition, our results suggest that a slab beneath northern Borneo, which is likely stagnant in the mantle transition zone, could be a remnant of the Proto South China Sea. Further east along New Guinea, the plate reconstructions coupled to geodynamic experiments are consistent with north-dipping subduction along the Halmahera Arc coeval with the ~20 Ma onset of south-dipping Maramuni subduction along New Guinea. The Late Cretaceous (~71 Ma) onset of Sepik oceanic basin subduction, followed by the docking of the Sepik composite terrane to southern New Guinea by ~30 Ma, produces a mid-mantle slab imaged in tomography beneath Lake Eyre in Australia, as discussed in Schellart and Spakman (2015), due to the combination of southward mantle flow and Australia's northward advance towards the Southeast Asian slab burial grounds.

We present testable and reproducible plate reconstructions with regional refinements and improvements to the understanding of post-Jurassic eastern Tethyan geodynamics. The reconstructions may form the basis of future work to better understand the tectonics of the Tethyan domain, and could also be used to study oceanic circulation, long-term climate change and biogeographic dispersal pathways. In addition, our work highlights the need for testing competing plate reconstruction scenarios using numerical modelling approaches in a global and geodynamic framework.

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