# Dynamic evolution of competing same-dip double subduction: New perspectives of the Neo-Tethyan plate tectonics

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## Abstract

*Same-dip double-subduction* (SDDS) systems are widely reported from present as well as past complex convergent plate tectonic configurations. However, the dynamics of their evolution is poorly understood, which is crucial to conceptualize anomalous subducting slab kinematics and associated observed geological phenomena, such as irregular trench migration rates, high convergence velocities, and slab break-off. To bridge this gap, we develop dynamic thermo-mechanical subduction models and investigate the initiation and evolution of SDDS systems, considering three different initial plate configurations: oceanic, oceanic-continental and multiple continental settings, based on Neo-Tethyan paleo-reconstructions. Each model offers new insights into the complex tectonic history of the major Neo-Tethyan subduction zones, particularly the Indo-Eurasian and Andaman convergent systems. We evaluate the slab-slab interactions, trench and subduction kinematics, inter-plate reorganization, and temporally varying mantle flow patterns involved in the dynamic evolution of these SDDS systems. The oceanic SDDS model simulations reveal that a sizable oceanic plate can initiate two subduction zones synchronously, and they evolve unequally in a competing mode, leading to exceptionally high convergence rates (~16-17 cm/year) for a prolonged duration (~7-8 Myr). This finding explains the coeval activity of coupled subduction zones in the Indo-Eurasia convergence during the Cretaceous evolution of the Neo-Tethys. We further implement a corresponding single subduction model to assess the additional effects of competing slab kinematics in an oceanic SDDS setting. The ocean-continent SDDS model, on the other hand, localizes subduction preferentially at passive margins between the oceanic plate and the continental block, forming double subduction zones that grow almost equally to form a spreading centre between the two trenches. These model results allow to reconstruct the Cenozoic evolution of the eastern Neo-Tethyan region, which ultimately led to the development of the Andaman subduction zone. We also show the post-Cretaceous evolution of the Indo-Eurasian collision zone as a consequence of the SDDS dynamics in presence of multiple continental blocks. These dynamics facilitated slab break-off, transforming the SDDS into a single subduction system in a relatively short time frame (~3 Myr). We finish with a synthesis of the paleo-reconstructions of the Neo-Tethys in the perspective of these SDDS models.

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**Keywords:** Convergent plate tectonics, double subduction systems, dynamic subduction modelling, Finite element method, slab interaction, spreading centre locations, tectonic evolution of Neo-Tethys

# 1. Introduction

Subduction zones are spectacular planetary-scale manifestations of the convergent plate tectonics, and govern a wide range of geological phenomena, such as continental crust formation, magmatism and geochemical recycling, and geophysical processes, such as earthquake localization and energy dissipation (Corbi et al., 2017; Stern, 2002; Tilling, 1996). Subduction tectonics often turn complicated due to complex configurations of multiple plates, kinematically linked with one another. Double subduction systems are a common example of such convergent settings, where two lithospheric slabs subduct synchronously along sub-parallel trenches (Fig. 1). They are widely reported from the present-day plate configurations (Fig. 1), e.g., the Mediterranean (Király et al., 2018; Vignaroli et al., 2008), the Molucca Sea (Zhang et al., 2017), Taiwan (Lin and Kuo, 2016), New Zealand (Lamb, 2015), and around the Philippines plate

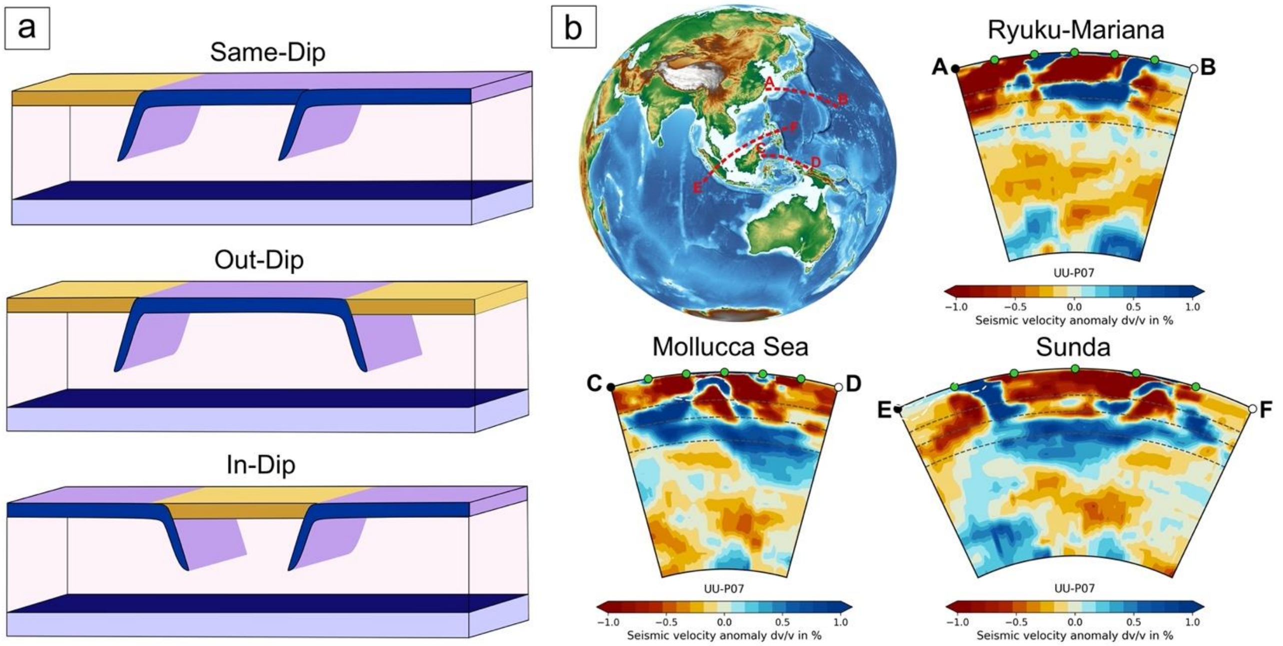

Figure 1: Illustrations showing a) the three different types of double subduction systems and b) their natural counterparts observed from seismic tomography. The P-wave model used for this analysis is UU-P07 (Amaru, 2007).

(Ryukyu vs. Izu-Bonin-Mariana trenches; (Faccenna et al., 2018; Hall and Spakman, 2002)), as well as from paleo-reconstructions, e.g., Trans-Tethyan System (Jagoutz et al., 2015), and Late Cretaceous eastern Turkey evolution (Eyuboglu et al., 2019). The double-subduction model has recently received much attention to address the problem of enigmatic slab kinematics encountered at both modern and ancient convergent zones (Čížková and Bina, 2015; Dasgupta and Mandal, 2018; Jagoutz et al., 2015). Based on seismic tomography and available plate kinematics data from GPS observations, three types of double-subduction configurations (Király et al., 2021) have been recognized: outward-, inward- and same-dipping slabs (Fig. 1). This study, however, explores the complex dynamics of same-dip double subduction with a focus on the Neo-Tethyan tectonic evolution.

Plate reconstruction studies suggest that same-dip double subduction (SDDS) systems developed repeatedly in the course of the ~150 Myr Neo-Tethyan evolutionary history (Hall, 2012; Kufner et al., 2016; Pusok and Stegman, 2020). Using quantitative models, Jagoutz et al. (2015) have shown the existence of coupled subduction zones in the Trans-Tethyan intra-oceanic system below the southern margin of Eurasia. During the Cretaceous to Early Tertiary period, this tectonic system evolved through multiple Neo-Tethyan subduction episodes, featuring the Kshiroda Plate's north-dipping oceanic slab under Eurasia (Yin and Harrison, 2000) and intra-oceanic subduction of the northern oceanic segment of the Indian subcontinent beneath the Kshiroda Plate leading to the formation of the Kohistan extensional arc. Geological, geochemical, and paleomagnetic investigations in the tectonic units, specifically the forearc, oceanic melange, and ophiolite sequences of the Himalayas (Aitchison et al., 2000; Bouilhol et al., 2013; Hall, 2012; Yin and Harrison, 2000) imply the concurrence of parallel subduction zones with similar dipping orientations. Complementing these findings, geophysical observations, such as the detection of multiple slab remnants through seismic tomography models (Van Der Voo et al., 1999), and alterations in trench kinematics and the upper plate's deformation patterns (Gürer et al., 2016; Jolivet et al., 2018), further reinforce the argument in favour of SDDS. Similarly, recent works have recognized multiple double-subduction events to reconstruct the evolution of the Andaman Subduction system (Advokaat et al., 2018; Bandyopadhyay et al., 2020; Ghosh et al., 2017). Its active tectonics now occurs at the northern flank of the Java trench where the oceanic part of the Indian Plate subducts at a low angle to the arc-trend beneath the overriding Andaman microcontinental block. Geochronological correlations and geochemical signatures of the arc-volcanism (Westerweel et al., 2019) and ophiolitic suites (Acharyya et al., 1991) indicate the activity of a second subduction zone on the eastern side in late-Miocene, continuing to present day. Also, geochemical evidence (Ghosh et al. 2017) suggests that the Andaman-Nicobar outer-arc high grew from the coalescence of two accretionary prisms linked to the double-subduction tectonics. Despite a significant advancement of their geochemical and geophysical information, there remains a necessity of dynamically self-consistent geodynamic models to conceptualize the following unresolved tectonic issues of these SDDS systems: 1) the unusually high Late Cretaceous Indo-Eurasian subduction velocity, although the Kohistan extensional zone reduced the transfer of pull force from front to rear slab, 2) reconstruction of the Andaman Convergent System in the framework of Neo-Tethyan tectonics, and 3) dynamic backdrop of the failed trench at Amirante and its relationship to the Cenozoic Indo-Eurasian subduction. This study aims to develop SDDS models, mainly to address these long-standing issues of the Neo-Tethyan tectonics.

Recent analogue and numerical modelling have dealt with the complexities in double-subduction evolution and their effects on the regional plate motion, intraplate deformations, slab interactions and topographic developments (Dasgupta and Mandal, 2018; Holt et al., 2017; Király et al., 2021; Mishin et al., 2008; Peral et al., 2018; Pusok and Stegman, 2019). It is now a well-established fact that their mode of development significantly differs from that of a single-subduction system. The additional complexity originates primarily from inter-slab interactions that greatly influence the plate bending and viscous dissipation in the mantle, coupled with additional factors, such as contrasting plate ages, plate dimensions, rheology, and inter-slab distance (Čížková and Bina, 2015; Dasgupta and Mandal, 2018; Holt et al., 2017). The initial relative arrangement of tectonic plates is another influential factor in the double-subduction evolution. For instance, the Ryukyu-IBM system has an intra-oceanic subduction zone coupled with another subduction zone at an ocean-continent convergence boundary. In contrast, the Trans-Tethyan double subduction formed with the Indian continental plate linked to the subducting oceanic counterpart. This difference in the initial plate arrangements might have led to their contrasting evolutionary paths, resulting in completely different tectonic histories of these two regions. However, how the initial plate configuration governs the dynamic evolution of a SDDS system is still an open-ended question, which the present article addresses in the context of the Neo-Tethyan subduction tectonics.

We develop geodynamic models to simulate the thermo-mechanical evolution of SDDS systems, initiated along pre-existing lithospheric weaknesses and reconstruct the tectonic history of both the Indo-Eurasian and Andaman double subduction systems. The models are designed to reproduce time-dependent and self-consistently evolving subduction for a given plate configuration. This modelling allows to address the following key questions: 1) can two adjacent same-dip subduction zones grow concurrently, and 2) if so, how do these neighbouring subduction zones mutually influence each other during their evolution, depending on their initial tectonic configurations? We also explore the following influential phenomena: a) mutual interactions of simultaneous subducting-slab motions during the evolution of commonly reported Neo-Tethyan SDDS systems, b) complexity in the associated mantle flow patterns and their role in localization of lithospheric-scale extensional zones in convergent tectonics, and c) slab detachment leading to a double- to single-subduction transition.

# 2. Numerical Modelling Approach

## *2.1.Model Design*

We use 2D thermomechanical numerical models (Fig. 2) that are inherently dynamic, implying that no external forces or velocities are applied to the overall system. Details of the numerical method and model boundary conditions are provided in Appendix A. The model domain represents a rectangular 7000 km (horizontal) x 1000 km (depth) vertical section (Fig. 2a). We introduce mechanical weaknesses in the lithospheric layer, as considered in previous studies that suggested spontaneous subduction initiation to occur preferentially at the locations of pre-existing lithospheric weakness, e.g., transform faults (Arcay et al., 2020; Zhou et al., 2018). This weak-zone perturbation for subduction initiation is a viable mechanical model in the light of damage theory proposed by Bercovici & Ricard, 2014. Our model implements this weakness as a narrow (10 km wide), low-viscosity ($10^{20}$ Pa s) zone, extending to the lithospheric base (Fig. 2). The weak zones can also characterize intraplate lithosphere-scale faults in large oceanic plates, like the

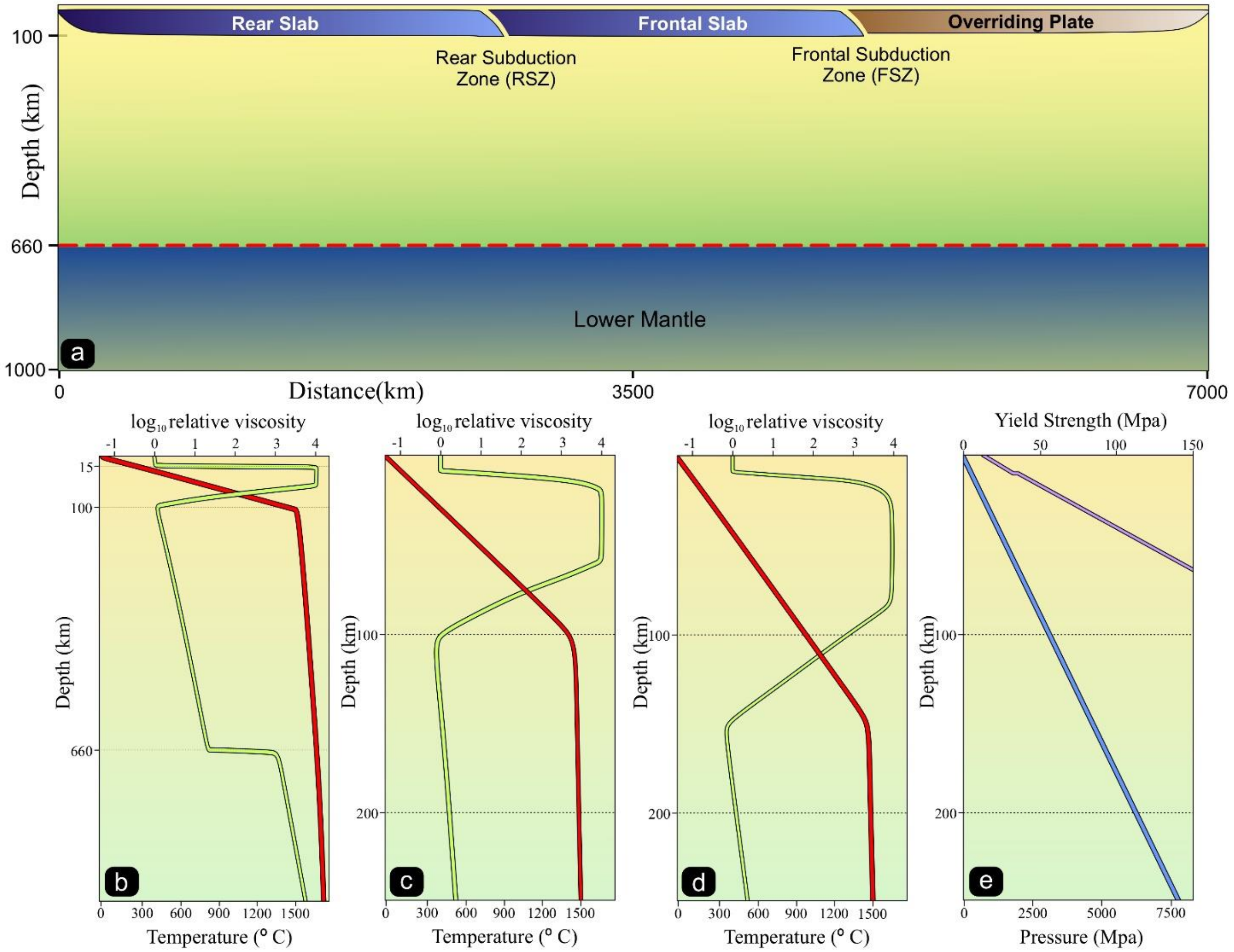

Figure 2: (a) The model setup and boundary conditions considered for CFD simulations, which shows the whole domain and initial setup of model employed for simulating same-dip double subduction in oceanic setting. Detailed illustrations representing the viscosity and temperature profiles of b) the complete model domain, c) oceanic lithosphere, and d) continental lithosphere utilized in the models. The viscosity values are relative to the asthenosphere value of $10^{20}$ Pa s. e) The strength plots exhibiting depth-dependent variations, showcasing increasing strength (purple) and pressure (blue).

present-day Wharton Basin (Stevens et al., 2020). All the rheological and thermal parameters used in this study are summarized in Table 1.

## 2.2.Model Configurations

Natural subduction zones occur in various convergent settings, which can be broadly categorized as: 1) oceanic settings (e.g., Izu-Bonin-Mariana), where two or more oceanic plates converge to each other; 2) ocean – continent settings (e.g., the Andean subduction system), where oceanic plates converge against continents; and 3) continent-continent settings (e.g. the Indo-

Eurasia collisional zone), where two or more continents approach each other. Considering Earth's present-day plate structures, SDDS systems can develop in any of these three tectonic settings, giving rise to a spectrum of their potential plate configurations. Based on the paleo-reconstructions of the Neo-Tethyan tectonics, we thus chose three different cases for the model settings, with an objective to investigate the mode of double-subduction evolution depending on the initial plate architecture. Each of these initial settings accounts for local lithospheric weaknesses in a specific arrangement with or without mechanically strong continental blocks.

The first case of our subduction modelling (referred to as *oceanic plate model*) replicates a simple oceanic plate tectonic setting, consisting of a 120 Myr old, flat lying oceanic lithosphere with a length of 5000 km (Fig. 3a). The model initially contains two similarly-dipping narrow weak zones of different thicknesses in the oceanic lithospheric plates at distances of 2500 km and 5000 km. The weak-zone configuration is chosen so that the model favours one of the weak zones to become active first to initiate an ocean-ocean subduction zone, as reported from natural SDDS, such as the Trans-Tethyan subduction system. No continent is included in the model domain to exclusively show the effects of oceanic plates on the SDDS dynamics. We also implement a corresponding single subduction model (Supplementary section S1) to find the additional effects of inter-slab interactions. The second case (referred to as *oceanic– continental plate model*) is designed to represent an ongoing subduction system, where an oceanic plate has already been subducted into the upper mantle, carrying a 400 km long and ~150 km thick microcontinental block (Fig. 4a). This configuration aims to reveal the mechanical effects of pre-existing micro-continent between two oceanic plates. The continental crust is assigned a compositional density lower than that of mantle material by 600 kg/m$^3$. The block is assumed to form a weak interface with the adjoining oceanic plate, possibly resulting from the inter-plate 3D differential stress localization (Auzemery et al., 2020; Heidbach et al., 2007). On the rear side, it is separated from an 80 Ma old oceanic lithosphere by a dipping (30º) weak zone. The third model, designed to reproduce a continent-continent collisional setting (referred to as *multiple continent model*), initially consists of a flat-lying 1200 km long, 80 Myr old oceanic lithosphere, and two continental blocks (thickness: 150 km and length: 1500 km), separated by an oceanic basin. This oceanic plate forms a narrow weak channel at the contact with the overriding strong continental plate (Fig. 5a). The continental blocks are compositionally buoyant relative to the underlying asthenosphere.

# 3. Model results

## *3.1. Oceanic plate settings*

This model setting localizes plastic yielding preferentially at one of the pre-existing lithospheric weaknesses (Fig.3a), resulting in an unstable state of the flat-lying oceanic plate to initiate the rear subduction zone (RSZ). This process accompanies the formation of a divergent spreading centre that eventually acts as an active site of oceanic plate generation. At ~6 Myr, the newly generated lithosphere constitutes a well-developed overriding plate (OP) structure in the subduction zone. The proto-slab subducts to a depth of ~300 km with a convergence rate ($V_C$) of ~13 cm/yr (relative to its adjoining overriding oceanic plate) at 8.1 Myr (Fig 3b). The active RSZ plate approaches the trench with a velocity ($V_P$) of ~+12 cm/year (+ sign denotes movement towards the right), while the trench itself retreats with a velocity ($V_{RT}$) of ~-2.5 cm/yr. The plate convergence velocity (calculated between passive markers 1 and 3; Fig. 3a) accelerates further with increasing total negative buoyancy to attain a maximum value of ~15 cm/year (Fig. 6a), and the RSZ slab dip (measured at 125 km depth) steepens to ~45°. The slab subsequently encounters the 660-km transition at 10.1 Myr, and decelerates ($V_P$≈7.5 cm/yr) due to higher viscous resistance

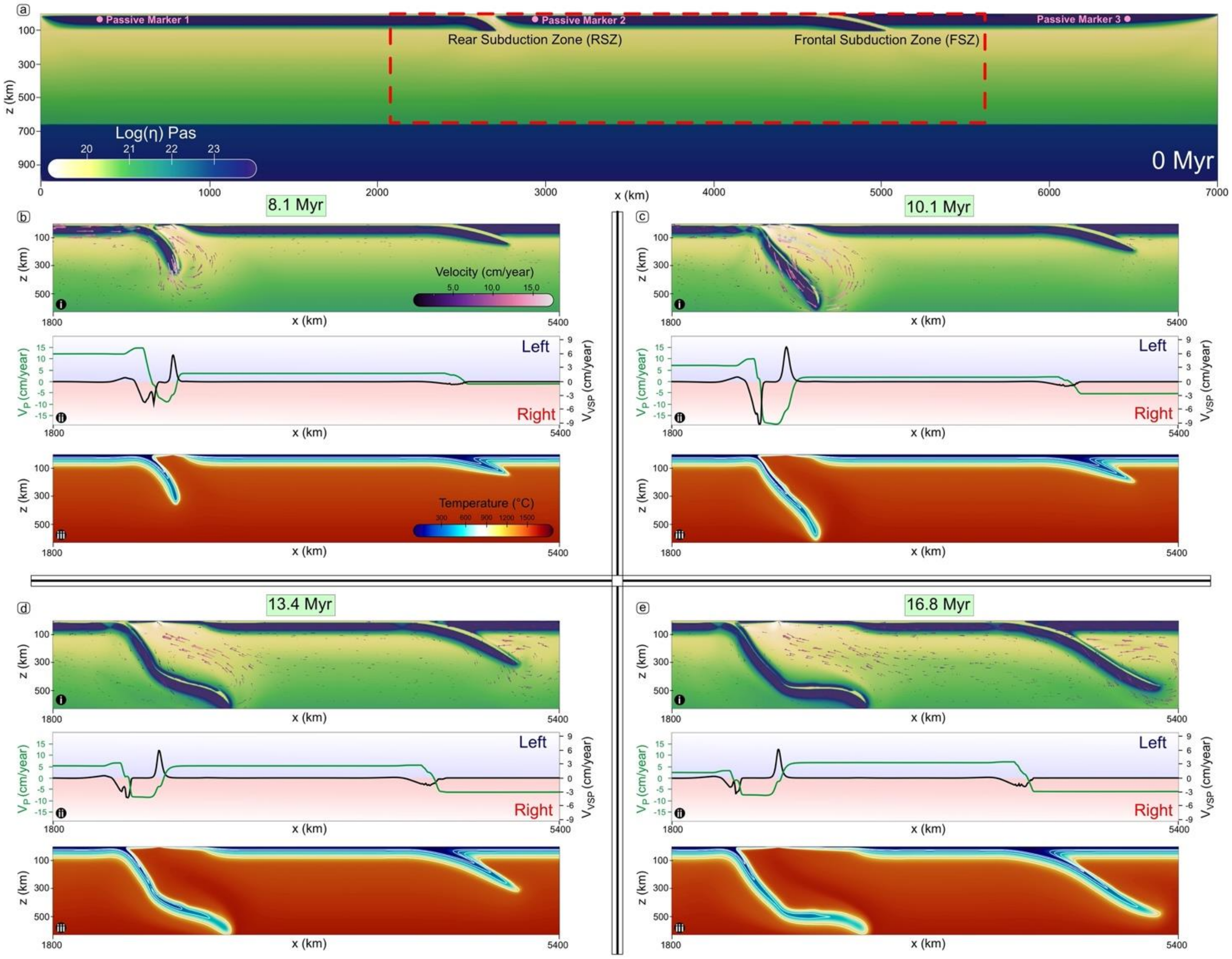

Figure 3: Evolution of the double subduction model in an oceanic setting. Panels show: (a) the initial viscosity field of the complete numerical model domain, (b)-(e) evolution of the i) viscosity and velocity fields, ii) plate velocities ($V_P$) (green) and vertical subduction-velocity ($V_{VSP}$) (black), which denote the horizontal ($V_x$) and vertical ($V_y$) component of the total velocity vector respectively, and iii) temperature field, zoomed into a region around the subduction zone for four time-steps corresponding to the back subduction initiation ($t$= 8.1 Ma), frontal subduction initiation ($t$= 10.1 Ma), frontal subduction free sinking ($t$= 13.4 Ma) and mature double subduction ($t$ = 16.8 Ma) phases.

in the stronger lower mantle. This slab-lower mantle interaction also forces the oceanic plate to slow down its motion ($Vc$≈12 cm/year; Fig. 6a), prompting the other proto-slab to activate and initiate the frontal subduction zone (FSZ) (Fig. 3c). At this stage, the tectonic setting transforms into a typical double-subduction configuration, albeit with significant asymmetry in terms of both slab geometry and kinematics, where the FSZ slab shows much shallower dip (~20°) and lower convergence velocity ($V_c^{RS}$ ≈7 cm/yr) than the rear subducting slab. Additionally, the RSZ trench retreats significantly faster ($V_{RT}$ ≈9 cm/yr) than the FSZ trench ($V_{FT}$ ≈2.5 cm/yr). The cumulative effects of double-subduction motion facilitate the convergence motion to not only attain, but also sustain extremely high velocities for a prolonged period (~6 Myr). This is in stark contrast to scenarios of single subduction, where such extreme convergence rates can prevail over short durations (<1 Myr).

The double-subduction system always maintains an active state of the spreading centre to continuously produce new lithospheric materials. At ~13.4 Myr, the RSZ attains a declining stage, marked by a significant drop in $V_P$ (~5 cm/yr), although its trench continues to retreat with $V_{RT}$ ~3 cm/yr. In contrast, the FSZ continues to remain active, with its slab sinking rapidly into the upper mantle to facilitate the convergence velocity ($V_P$≈ 5 cm/year) as well as the trench retreat velocity ($V_{FT}$≈7 cm/year). However, the SDDS system cumulatively shows a decreasing trend of the net convergence velocity ($V_c$≈11 cm/yr). At ~16.8 Myr, the rear subducting slab moves backward as $V_{RT}$ (~3.5 cm/year) exceeds $V_P$ (~2.5 cm/year) and the double-subduction evolves to a steady state condition of the RSZ in terms of $V_c$ and slab dip (~55°). On the opposite side, however, the oceanic plate associated with the FSZ acquires a maximum velocity $V_P$ ≈6.5 cm/year, significantly higher than the rear subducting plate velocity. The FSZ slab then moves through the entire upper mantle to encounter the lower mantle and eventually slows down its velocity. The corresponding trench continues to retreat with $V_{FT}$ ≈6 cm/yr, which subsequently surpasses the FSZ plate velocity in course of the subduction evolution (>20 Myr). The convergence attains a velocity of ~8.5 cm/year with a tectonically-steady configuration till the end of the model run.

During the RSZ-initiation stage, the descending-slab driven mantle flows exhibit both vertical and horizontal components, shaping a circulating flow vortex around the subducted slab. The vortex remains active till ~10Ma, gradually waning as the RSZ slab decelerates upon encountering the 660 km discontinuity. With this deceleration, the vortex's focal point shifts to the FSZ slab, accompanied by an increasing flow intensity with time. At ~13 Myr, the double-subduction configuration sets in a potent upward slab-parallel flow beneath the spreading centre, which strengthens further by merging with the counter-flow generated by the frontal subducting slab (Fig. 3e-i). Together, these components constitute a robust mechanism for channelling mantle materials towards the spreading centre. Notably distinct from typical single-subduction driven corner and wedge flows, this increased mantle-material supply emerges as a pivotal factor to facilitate the accretion rates of lithosphere.

### *3.2. Microcontinent - oceanic plate setting*

The micro-continent-bearing oceanic plate setting consists of an already initiated subduction in the front (FSZ: frontal subduction zone) and the lithospheric weakness between the microcontinent. This configuration results in initiation of the rear subduction zone, forming a double-subduction system at an early stage (~6 Myr) of the model run (Fig.4b), as evident from a convergence velocity $V_C$ =~5 cm/yr (Fig. 6b). The subduction zones, however, grow at different rates, where the FSZ subduction is much more active than the RSZ. This difference in their subduction rates causes rifting between the FSZ oceanic plate and the microcontinent. The rift subsequently acts as a spreading centre, allowing upwelling of the underlying mantle material to the surface and generate new oceanic lithosphere between the continental and the frontal oceanic

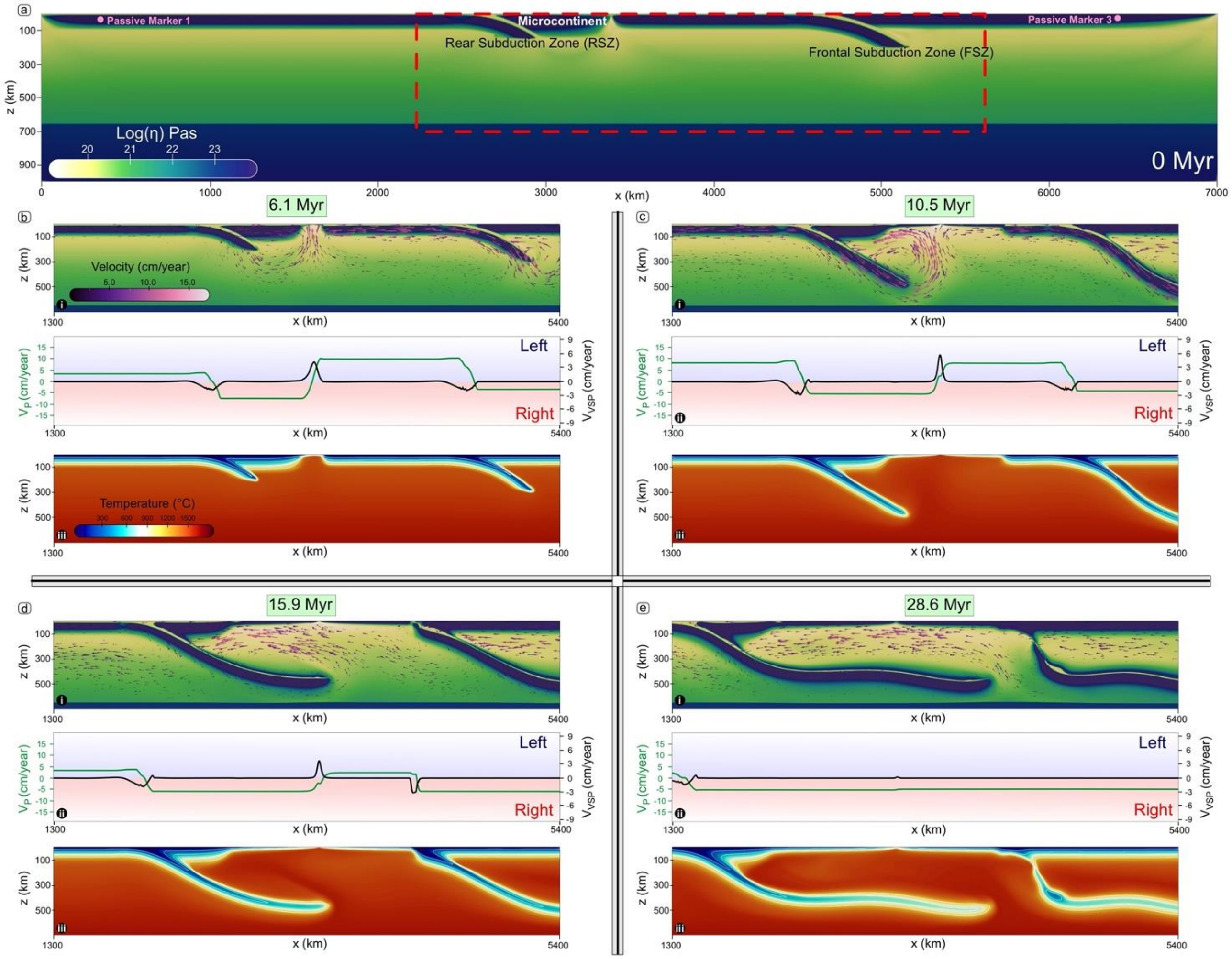

Figure 4: Evolution of the double subduction model in presence of a continental block. Panels show: (a) the initial viscosity field of the complete numerical model domain, (b)-(e) evolution of the i) viscosity and velocity fields, ii) plate velocities ($V_P$) (green) and vertical subduction-velocity ($V_{VSP}$) (black), which denote the horizontal ($V_x$) and vertical ($V_y$) component of the total velocity vector respectively, (see Supplementary section S2) and iii) temperature field, zoomed into a region around the subduction zone for four time-steps corresponding to the fore subduction free sinking (*t*= 6.1 Ma), back subduction free sinking (*t*= 10.5 Ma), mature double subduction (*t*= 15.9 Ma) and fore subducting slab detachment (*t* = 28.6 Ma) phases.

plates. At ~6.1 Myr, the RSZ proto-slab subducts to a depth of ~200 km into the upper mantle, and this subduction motion forces the continental block to act as an OP, moving trench-ward at a velocity $V_P \approx$-7 cm/year. The FSZ slab, on the other side, continues to subduct freely into the upper mantle, setting the entire plate to move at a high velocity ($V_P \approx$+10 cm/yr) (Fig. 4a). During this phase of tectonic evolution both the trenches retreat, but with contrasting velocities; the RSZ trench retreats at a significantly faster rate ($V_{RT}$ ≈-6.5 cm/yr) than the FSZ ($V_{FT}$ ≈-3.8 cm/yr). The convergence velocity between the main OP and the RSZ slab is ≈7 cm/year, which increases to a high value ($V_C$ ≈13 cm/yr) on a model run time of ~9 Myr (Fig. 6b). The frontal oceanic slab reaches the lower mantle at ~10.5 Myr and begins to slow down the velocity ($V_P$ ≈+7 cm/yr) due to higher viscous forces in the 660-km transition zone. The RSZ slab, however, continues to subduct through the upper mantle and maintains the rear oceanic plate movement at a velocity of ≈ +8 cm/year. The RSZ trench at the same time retreats, but with a reduced velocity ($V_{RT}$ ≈-5 cm/yr), whereas the FSZ trench continues to retreat at a steady rate ($V_{FT}$ ≈-4 cm/year). The convergence velocity is high ($V_C$ ≈13 cm/year), but shows a diminishing trend after this period. At ~15.9 Myr, both the subducting oceanic slabs lower their dips to become almost flat, and significantly reduce the plate velocities ($V_P$ ≈+2.5 cm/yr and ≈+3.5 cm/yr for FSZ and RSZ slabs, respectively). The relatively buoyant juvenile (~15 Myr) lithosphere, formed at the spreading centre, eventually reaches the FSZ trench and its subduction accelerates the retreat motion ($V_{FT}$ ≈+7.5 cm/year). This motion reduces its slab dip from 30° to ~26° (Fig. 6b), which starts to increase significantly after ~18 Myr. The RSZ trench retreat velocity ($V_{RT}$) increases to -5.5 cm/yr during this period. At ~25 Myr, the FSZ slab undergoes break-off at the location of maximum strain localisation developed by the negative buoyancy induced slab pull (see Supplementary S3 for detailed strain-rate plots).

In this SDDS system, the spreading centre located between the two trenches remains active, continuously adding new lithosphere to accommodate the increasing space between the continental block and the frontal oceanic plate. At ~28.6 Myr, the spreading centre, however, becomes almost inactive as the upwelling process is replaced by slab-driven horizontal flows in the mantle region between the two subducting plates. The entire newly formed lithosphere is coherently coupled with the main OP, as revealed from the plate velocity ($V_P$ ≈-5 cm/year). The double subduction system ceases to exist due to the absence of slab pull force at the FSZ trench and transforms into a single subduction system.

### 3.3. *Multiple-continental plate setting*

The model run shows that subduction localizes preferentially along the lithospheric weakness to initiate the frontal subduction zone (FSZ), leaving the trailing weak zone at the continent-oceanic plate interface almost inactive (Fig. 5b). At this stage (~4 Myr) the system follows mostly single-subduction dynamics, allowing the FSZ slab to penetrate into the upper mantle to a depth of ~200 km. The plate velocity ($V_P$) increases to ~+7.5 cm/year on a model run time of 8.5 Myr (Fig. 5b). The velocity field indicates that the lithospheric plates move coherently as a single unit to subduct beneath the FSZ trench although a lithospheric weakness is present in between them ($V_P$ ≈+5 cm/year for rear oceanic plate). This implies that the FSZ entirely determines the dynamics of the system at an early stage. The frontal oceanic plate progressively accelerates its motion to attain a convergence rate ($V_c^{RS}$ ≈8 cm/year, Fig. 6c solid blue line) with respect to the OP at ~11.5 Myr. At this stage the subduction system undergoes a remarkable kinematic transformation, leading to plate decoupling at the rear lithospheric weakness that eventually triggers mantle upwelling and accretion of new lithospheric materials to form a thin overriding plate. This decoupling decelerates the rear plate motion ($V_P$ ≈+1 cm/year) and brings it almost to a halt. In contrast, the frontal slab continues to maintain a high plate velocity ($V_P$ ≈+7.5 cm/year) and a high convergence rate ($V_c^{RS}$ =+7.5 cm/year), along with a moderate trench retreat rate ($V_{FT}$≈-3.5 cm/yr). At ~12.5 Myr, the rear subducting slab gains a significant slab-pull velocity ($V_{SP}$ ≈-10 cm/year), which consequently causes initiation of the rear subduction zone (RSZ). The proto-slab continues to move through the upper mantle, setting the convergence velocity rapidly accelerate to attain a value of ~11 cm/year at 13.8 Myr. On the other side, the FSZ slab reaches the 660-km transition zone and slows down its overall sinking velocities. The continental block in this plate reaches the trench and collides with the overriding plate, switching a reversal in the trench motion to advance at a rate $V_{FT} \approx +2$ cm/yr. The collision event also reduces the plate velocity ($V_P \approx +5$ cm/yr) as well as the convergent velocity ($V_c^{RS} \approx 4.5$ cm/yr) at ~15 Myr. Eventually, the RSZ experiences strong extensional forcing created by the FSZ, and its convergence reduces to

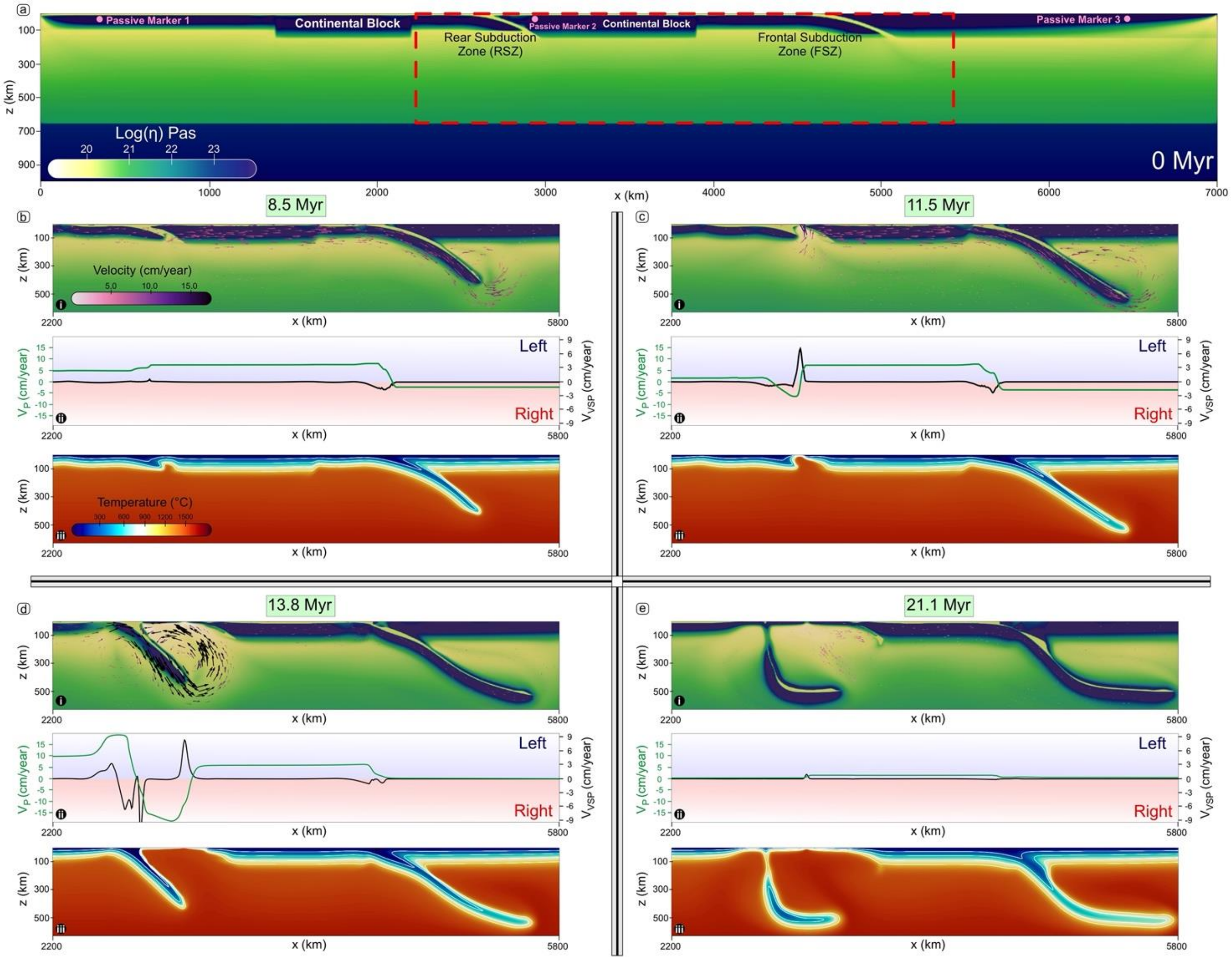

Figure 5: Evolution of the double subduction model in presence of multiple continental plates. Panels show: (a) the initial viscosity field of the complete numerical model domain, (b)-(e) evolution of the i) viscosity and velocity fields, ii) plate velocities ($V_P$) (green) and vertical subduction-velocity ($V_{VSP}$) (black), which denote the horizontal ($V_x$) and vertical ($V_y$) component of the total velocity vector respectively, and iii) temperature field, zoomed into a region around the subduction zone for four time-steps corresponding to the fore subduction free sinking ($t$ = 8.5 Ma), continental collision ($t$ = 11.5 Ma), back subduction initiation ($t$ = 13.8 Ma) and back subducting slab detachment ($t$ = 21.1 Ma) phases.

extremely low rates, which impedes the trench retreat, causing the RSZ slab to steepen (>65°) and eventually break off. The continental block plays an instrumental role to facilitate triggering the slab detachment at ~16 Myr. In course of the double-subduction evolution (~21.1 Myr), the plate velocities drastically drop to ~2.5 cm/year and the tectonic setting transforms into a single subduction system. Thus, these results show that in this SDDS system, two simultaneous subduction zones cannot be sustained and the subduction behind the collision zone ceases without attaining maturity.

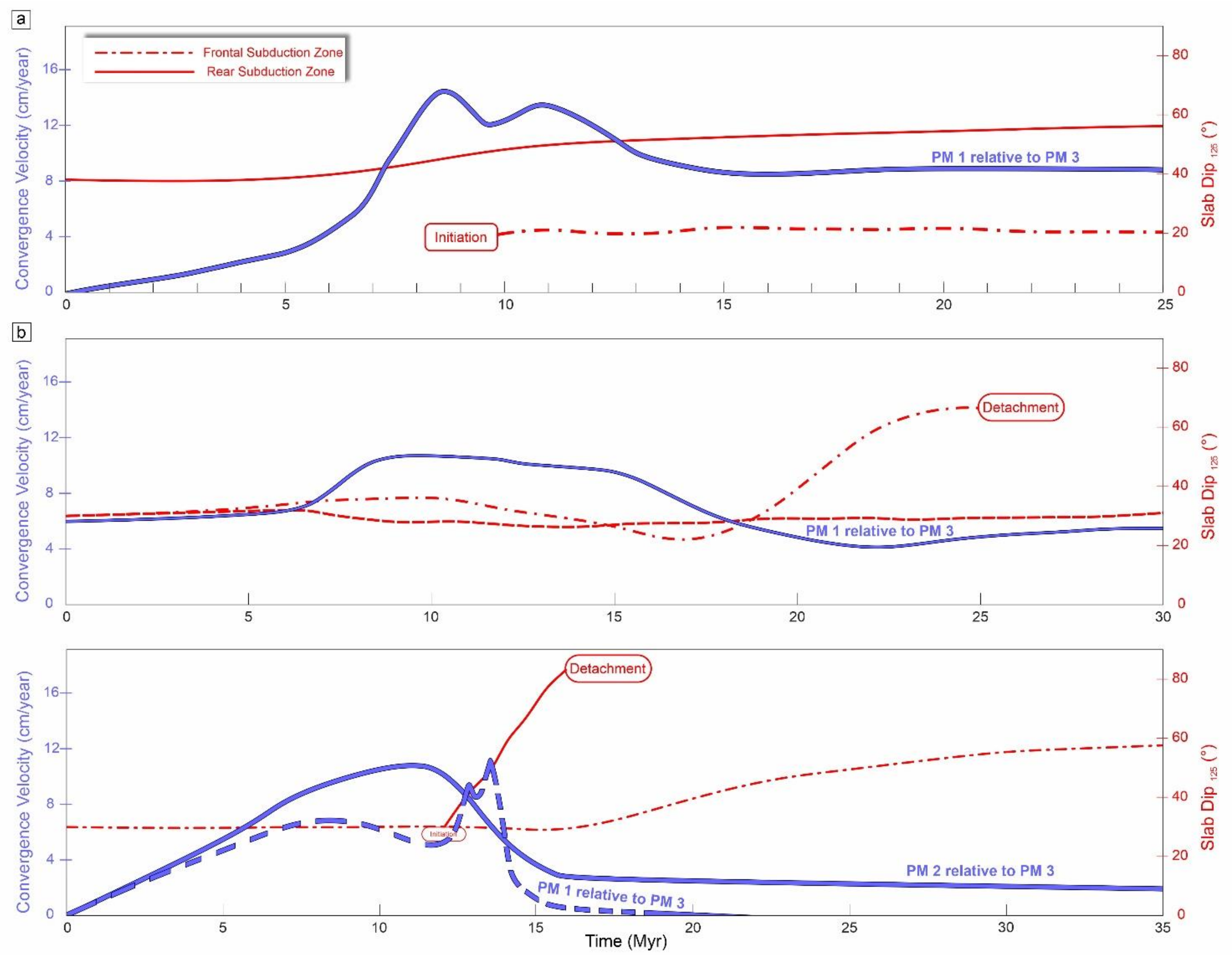

Figure 6: Temporal evolution of subduction zone convergence rates and shallow slab dips (at depth = 125 km) of fore and back subducting slabs, for double subduction in a) oceanic setting, b) microcontinent - oceanic plate setting and, c) multiple continental plate setting. PM: Passive Marker

# 4. Discussion

## *4.1. Applicability of the double-subduction models to the Neo-Tethyan systems*

Based on the available field, petrological, paleo-magnetic and geochemical information, we consider the following Neo-Tethyan subduction systems: i) Trans-Tethyan intra-oceanic system, ii) India-Andaman-Burma subduction system, and iii) Amirante-India-Eurasia to discuss their evolution in the light of the present SDDS models. These subduction systems are chosen as their tectonic settings have been well-constrained by combining geological and geophysical observations in recent studies. Additionally, seismic observations indicate that the subducted slabs in these systems rested in the upper parts of the lower mantle (Van Der Voo et al., 1999; Yang et al., 2022). To implement this slab configuration in the simulations, our modelling excludes the phase changes at the 660-km transition.

### *4.1.1. Cretaceous evolution of the Trans-Tethyan System*

The Trans-Tethyan intra-oceanic system (Fig. 7a) evolved through multiple subduction episodes in the Neo-Tethys during the Cretaceous to Early Tertiary period as discussed earlier in Section 1. The convergence history, derived from the plate reconstructions and paleomagnetic interpretations (Cande and Stegman, 2011; Copley et al., 2010; Van Hinsbergen et al., 2011) reveal that, during the Indo-Eurasian convergence the Indian Plate had a rapid northward drift moving at extremely high velocities >12 cm/year from 60 Ma to 50 Ma (Copley et al., 2010; Molnar & Stock, 2009), which subsequently reduced to 8 cm/year at ~40 Ma that marks the timing of India-Eurasia collision (Maiti et al., 2021). The oceanic plate model presented in Section 3.1 provides new insights into the geodynamic evolution of the Trans-Tethyan intra-oceanic system. The subducting oceanic plates in the model, RSZ and FSZ represent the frontal oceanic part of the Indian Plate

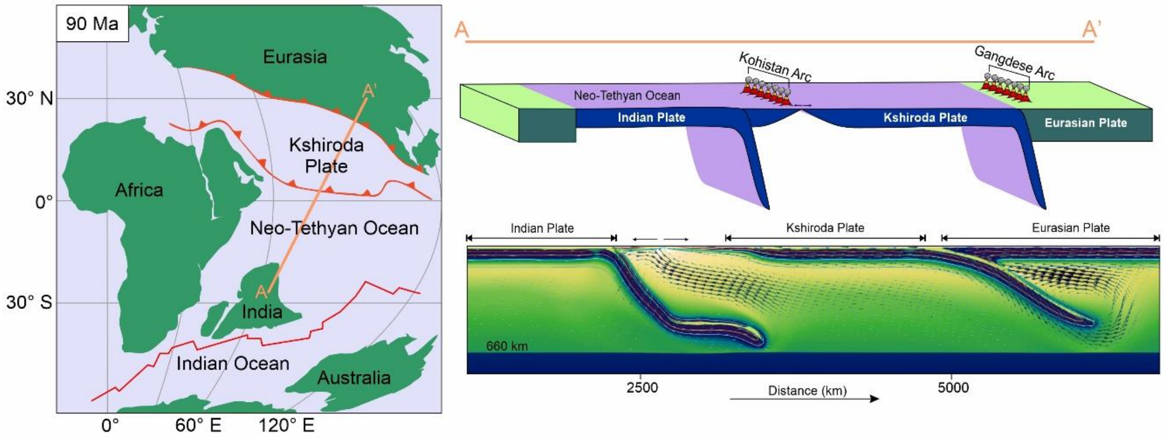

Figure 7: a) Sketch showing the reconstructions of Neo-Tethyan plate boundaries at ~90 Ma modified after Jagoutz et al., 2015 b) Cross section along AA' showing slab geometries and distances between the two prominent subduction systems giving rise to the double subduction process. c) Time snapshot (16 Myr) of the oceanic SDDS model which simulates the Cretaceous evolution of the Trans-Tethyan System.

and the Kshiroda Plate, respectively, in association with the Eurasian overriding plate (Fig 7a,b). Pre-existing lithosphere-scale weaknesses, such as transform faults or oceanic fracture zones between the Kshiroda and the Indian Plate nucleated the subduction initiation at the leading edge of the Indian oceanic lithosphere, almost synchronously accompanied by opening of a proto back-arc extensional zone. The subduction event resulted in arc formation, preserved as the Kohistan-Ladakh Arc in the Himalaya-Tibet Mountain system (Bouilhol et al., 2013). A correlation of the arc initiation timing with the paleomagnetic data reveal that the Kohistan–Ladakh Arc rocks and ophiolites in the Tsangpo-suture zone formed nearly at the equator position (Aitchison et al., 2000). After ~4-5 Myr from the intra-oceanic subduction initiation, the passive margin between the Kshiroda and Eurasian Plates became active to form a second subduction (FSZ) (Fig. 7c), which gave rise to the Gangdese-Karakoram continental arc. The paleomagnetic data of magmatic rocks indicate that the arc formed at a latitudinal position ~20°–25° N, ~2500 km from the Kohistan Arc present location, which is consistent with the present model estimates. The model convergence history reveals a period of rapid convergence with velocities >~15 cm/year for a period of ~8 Myr before reducing to ~9 cm/year in the later phase of the model run. This two-stage kinematic evolution is also in agreement with the Cretaceous convergence history of the Indo-Eurasian system prior to the Early Cenozoic slowdown (Molnar and Stock, 2009).

The existence of an extensional zone, namely the Kohistan Arc between two active subduction zones decoupled the oceanic plate, which in turn freed the rear slab from slab pull of the frontal subduction zone. According to this model, high convergence rates can originate in double subduction systems, even in the absence of any slab pull-force transfer from the frontal to the rear subduction zone. However, our model estimates yield the absolute Indian plate velocity (PM1: ~11-6 cm/year) lower than those interpreted from paleomagnetic records. This discrepancy probably results from additional influences of other geodynamic processes, e.g., push force from the Reunion plume to the Indian Plate (Pusok and Stegman, 2020), which are excluded in the modelling. This poses a limitation in the present analysis. Addition of these factors might lead to convergence velocities significantly higher than present model values. However, such positive influences were perhaps countered by the presence of Indian continent, which would act as a sort of keel to reduce the plate velocity. Further model analyses are required to pinpoint their additional effects.

### *4.1.2. Cenozoic evolution of the Andaman System*

In the case of the Andaman Subduction System (Fig. 8a, b), the active subduction occurs currently at the northern flank of the Java trench where the oceanic part of the Indian Plate subducts at a low angle to the arc-trend beneath a small overriding-continental block (Fig. 5). This continental fragment, which is a detached part of the larger Indian Plate occurs parallel to the volcanic arc, forming the Andaman Island Chain. The Pliocene-Holocene volcanism in the inner arc of the Java Trench suggests that the subduction has remained active since the late Miocene

(Acharyya et al., 1991; Sengupta et al., 1990). The Andaman Sea, a manifestation of extensional tectonics and other similar basins in this region also opened up in the late Miocene (Curray, 2005). The Andaman Ophiolites occur in a flat-lying arrangement, showing a close spatial association with a zone of negative gravity anomalies that suggest their occurrence as thin sub-horizontal bodies. Moreover, the ophiolite suits range in age between late Mesozoic and early Eocene, much older than the Andaman Sea crust. These observations indicate that the Andaman and other related ophiolites farther north in this convergent zone were derived from a suture to the east during late Oligocene time and emplaced as nappes during the middle Eocene. The geochronological order indicates the existence of a second subduction zone at the time of late-Miocene subduction (continuing to present day) and opening of the Andaman Sea (Khan & Chakraborty, 2005). Our micro-continent-oceanic plate model (Fig. 4) explains how the SDDS dynamics controlled such a tectonic evolution in the Andaman convergent belt. The FSZ simulates the mid-Eocene subduction event where the Indo-Australian oceanic lithosphere continued to subduct beneath the Burma Plate (equivalent to the stiffer OP in the model). The model setup produces an extensional spreading

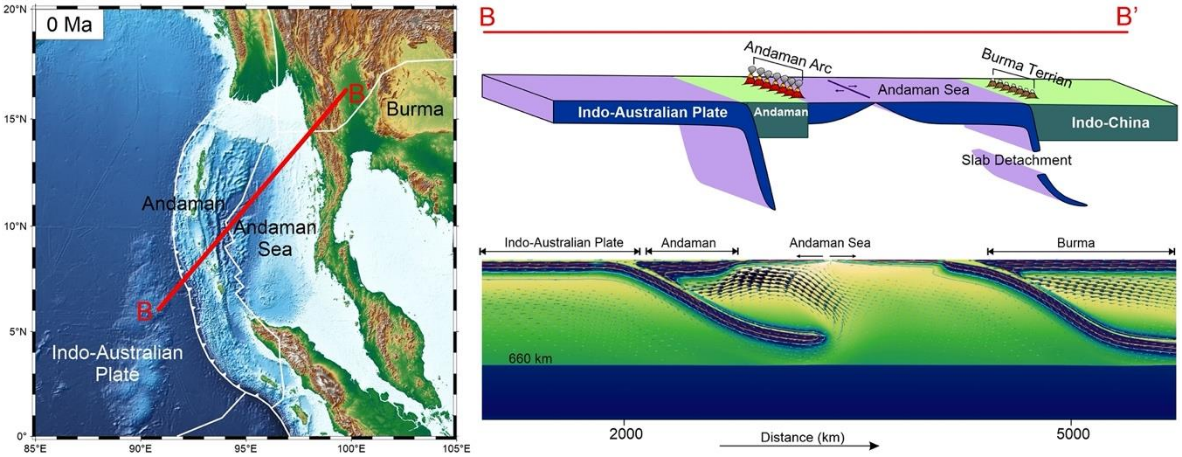

Figure 8: a) Bathymetric map of the Andaman Subduction Zone b) 3D block diagram of important components in the present-day Andaman–Nicobar subduction system along the BB’ profile showing slab geometries and spatial relation between the two subduction systems and the Andaman micro-continent. c) Time snapshot (15 Myr) of the oceanic-continent double subduction model which simulates the last 30 Myr evolution of this eastern Neo-Tethyan region.

centre as a consequence of some pre-existing lithospheric flaws, switching a remarkable change in the evolutionary course of the single-subduction system. The mechanical weakness might have developed due to the southward propagation of dextral motion in the Sagaing Fault System, which was active well before Miocene to set the mechanical conditions necessary for the Andaman Sea opening. The formation of this extensional zone eventually led to the late Miocene emergence of the Andaman Sea basin. The microcontinental fragment (*cf.* Andaman continental block) eventually decoupled itself from the ongoing subducting slab, and initiated a second subduction in the same region, as observed in our model at ~6 Myr (Fig. 8c). The newly formed subduction can be compared with the presently active Indo-Australian oceanic subduction below the Java Trench. The model suggests that both the subduction remained active for a considerable time (~24 Myr) and facilitated the Indo-Australian Plate motion relative to the Burma Plate. With time, the younger oceanic lithosphere formed at the Andaman Sea spreading centre drifted to the trench close to the Burma Plate and resulted in oceanic slab detachment as the lithosphere failed to gain density required for the mid-Eocene subduction.

### *4.1.3. Post-Cretaceous Indo-Eurasian Convergence*

The India-Eurasia convergence during the Cenozoic period had a collision between continental India and the Kohistan arc (Burg, 2011) and subsequently, with the Gangdese-Karakoram continental arc, resulting in the closure of the Neo-Tethyan basin. The opening of the Carlsberg Ridge in the Indian Ocean basin characterises the region south of the Indian continent (Fig 9c). The multiple continental plate model demonstrates the Cenozoic Indo-Eurasian collisional tectonic history very well. This model reproduces the collision of the Indian continental plate with the Eurasian Plate (~45 Ma) (Ding et al., 2005, 2016; Khan et al., 2009) when the oceanic subduction below Eurasia was active. Interestingly, the Carlsberg ridge had a fast spreading in late Cretaceous

(Merkouriev and Sotchevanova, 2003), which is found in our model that forms a spreading ridge and a second subduction in the ridge's vicinity. Based on our model results, we propose that the subduction occurred at the present-day position of the Amirante Trench- a trough-like feature,

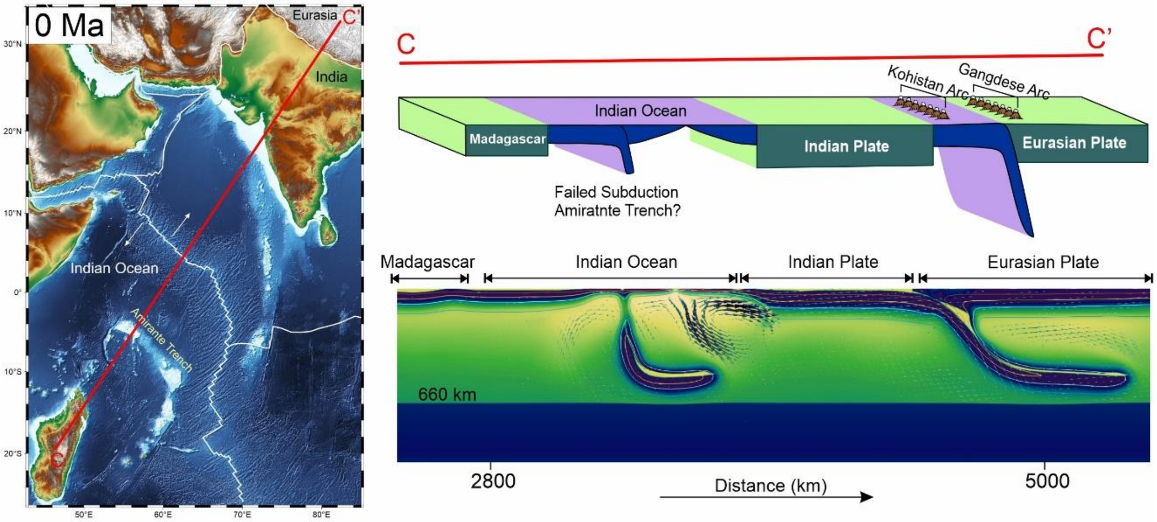

Figure 9: a) Bathymetric map of the Indo-Eurasian convergence zone and its related features b) Cross section along CC' showing slab geometries and distances between the prominent subduction features resulting in the convergence of the Indian plate with the Eurasian plate c) Time snapshot at 20 Myr of the multiple continent SDDS model which simulates the Post-Cretaceous evolution of the Indo-Eurasian collision zone.

approximately 600 km long in the western Indian Ocean (Fig. 9a). Geological and geophysical studies show evidence of partial or limited subduction within the trench (Miles, 1982). As observed in our model, the subduction was active in this region for ~4 Myr but ceased its activity following the slab detachment under the influence of a continental block (analogous to the Madagascar continental fragment in nature). The collision between the Indian and Eurasian continents led to a dramatic shift from retreating to advancing trench motion and a decrease in the Carlsberg ridge spreading rate, both of which are compatible with the model results. Our model results show that this tectonic system did not lead to the formation of a matured SDDS zones due to the influence of the continental blocks.

### *4.2. Double-subduction evolution by feedback mechanisms*

The present model experiments demonstrate that a double-subduction system can grow spontaneously from pre-existing weak zones (Bercovici and Ricard, 2014; Maunder et al., 2020) under slab-pull dynamics. The two subduction zones in such systems, however, evolve in a competing mode, one suppressing the other, ultimately leading to their unequal development on a million-year timescale. To illustrate this, consider our double-subduction model for an oceanic plate setting as applicable to the Trans-Tethyan System evolution during Cretaceous. The model shows synchronous initiation of subduction at two lithospheric weak zones (Rear and Fore Subduction Zone: RSZ and FSZ), but one of them (RSZ) preferentially becomes the most active zone (subduction velocity:~11 cm/yr), suppressing the other slab (FSZ) motion to remain significantly weak (~1.5 cm/yr). This competing double-subduction dynamics causes the second subduction to become more active once the first subducting slab encounters resistance at the lower mantle at 660 km. The reduced slab motion, coupled with a trench retreat motion in one subduction has a feedback effect on the second subduction in accelerating slab motion (~1.5 to ~6 cm/yr). To quantitatively evaluate the competing evolution of a double subduction system, we compare the SDDS model results with those from a subduction model, containing a single subduction system equivalent to the FSZ (details provided in Supplementary S1). The model comparison reveals that the RSZ activity significantly delays the FSZ initiation (~13.4 Myr), which, in contrast, begins to occur much quicker (3.4 Myr) in the single subduction model (Fig. S2). In addition, the FSZ activity, in overall becomes significantly sluggish due to its interaction with the RSZ in the oceanic double subduction system.

The other two model settings (Figs. 4,5) show similar feedback effects on the development of double-subduction systems. In the microcontinent-bearing oceanic plate setting, analogous to that of Andaman subduction system, the double subduction processes operate almost equally (Fig.

4), forming a spreading centre between the two subduction zones. However, the new lithosphere formed in the spreading centre causes the frontal subduction zone (FSZ) to undergo slab detachment, which in turn facilitates the slab motion in the rear subduction zone (RSZ). The multiple continental plate setting also shows a time-dependent interaction between the RSZ and FSZ during their evolution. The FSZ slab motion hardly allows the RSZ to become significantly active until the slab encounters the lower mantle to slow down its motion. The RSZ slab starts to actively sink when the RSZ almost reaches the upper-lower mantle boundary after a time of ~11 Myr. The feedback direction reverses as the RSZ experiences slab detachment due to resistance offered by the continent in the rear subducting oceanic plate. The paleo-trench at Amirante is an excellent example of subduction that failed to attain maturity due to the influence of the Madagascar continental plate. Thus, the model examples discussed here suggest that the SDDS systems generally evolve with a feedback relation between the two subduction zones that mutually interact with one another. To advance these findings, further studies are required to compare the model results with corresponding single-subduction model simulations and quantify the additional effects of slab-slab interaction in the SDDS systems.

### *4.3. Influence of continental heterogeneities*

Previous studies have shown that the complexity in double-subduction systems, as compared to single-subduction settings, originates mainly from the effects of inter-slab interactions in plate bending, coupled with additional factors, contrasting plate ages, plate dimensions, rheology, and inter-slab distance (Mishin et al., 2008; Cížková & Bina 2015; Holt et al., 2017). This study identifies the presence of microcontinental blocks or continental plates in oceanic plate tectonic settings as an additional influential factor in the dynamics and stability of a double-subduction system. Microcontinents are mostly surrounded by oceanic crust, and they form by extension and breakup of continental masses, followed by plate boundary relocations. Jan-Mayen in the north-east Atlantic Ocean (Gaina et al., 2009; Peron-Pinvidic et al., 2012), Andaman in the Indian Ocean (Bandyopadhyay et al., 2020) and several small, isolated islands around Australia (Gaina et al., 2003) are typical examples of microcontinents. Our microcontinent-oceanic double-subduction model suggests that such microcontinents play a vital role in localizing a spreading centre (location of new lithosphere generation) between the two subduction zones. The spreading centre, however, remains active for a specific time span, and its activity weakens following the slab detachment in the FSZ. This mechanism can be directly compared to the opening of the Andaman Sea extensional zone, synchronous to the initiation of the present-day subduction below the Andaman microcontinent.

The model results show that an oceanic plate setting in the absence of any continental blocks can readily form a double subduction system, initiated by pre-existing lithospheric weak zones in the plate setting. However, a buoyant continental fragment selectively prevents the subduction zones in front of this plate to mature with time, and forces the system to evolve unequally at the two trenches. Similarly, a multiple-continental assemblage reduces the convergence rates due to their buoyancy and continent-arc collision effects, described as *transference* by Stern (2004) to show the accretion-assisted continental growth mechanism. The reducing convergence rates result in tensile stresses in the subducting slab, eventually leading to slab detachment (Fig. 5). The model results suggest that microcontinental blocks can greatly influence the evolution of subduction processes, as reported from several natural subduction systems, such as the northern Luzon, the Puysegur (New Zealand) and the Andaman subduction zones (Bandyopadhyay et al., 2020; Zhu et al., 2023). The microcontinents provide a potential location of passive continental margin that transforms into a tectonically active region by subduction initiation as evident in the evolution

Andaman subduction system. Due to their lower mechanical strength, compared to that of large continental plates they localize subduction and facilitate the process to occur at faster rates.

### *4.4.Subduction-driven spreading centres: location of new plate generation*

Generally, single-subduction systems involve slab rollback, which is often accommodated by horizontal extension and formation of a spreading centre (back-arc basin) in the overriding plate. Double-subduction systems in an oceanic plate setting, on the other hand, forces the rear trench to retreat at fast rates as the frontal plate moves in the opposite direction, offering less resistance to the rear slab. This kinematic state results in formation of a spreading centre at the rear trench, which eventually acts as a site for new lithosphere generation. This mechanism allows the RSZ to accommodate slab roll back without involving overriding plate kinematics, as in a single-subduction setting. This model study thus brings out the importance of such spontaneous spreading centre formation in the SDDS evolution within large oceanic plates.

Our modelling demonstrates that spreading centres crucially control the subduction dynamics in transforming a double- to a single-subduction system. The microcontinent-oceanic plate model shows the development of large tensile stresses in the region between the older subducting plate and the newly formed lithosphere at the trench, which causes the older slab to experience detachment. The slab break-off eventually stops the subduction activity, leaving the other subduction active. Spreading centres also play an important role in the onset of convergent setting and promotes decoupling in a large oceanic plate (e.g., *oceanic plate model,* Fig. 3). Such tectonic processes are observed in other natural systems, for example, the Cocos-Nazca spreading centre where the spreading centre formed by splitting of the oceanic lithosphere, giving rise to the Farallon plate break up in the early Miocene.

## 5. Conclusions

The key results of this research are two-fold: firstly, it unveils the intricate mechanisms governing the development of same-dip double subduction (SDDS) systems. Secondly, it provides new insights into the role of SDDS dynamics in the Neo-Tethyan tectonic evolution. The principal findings of this study are summarized as follows: 1) SDDS are initiated spontaneously in presence of lithospheric-scale mechanical weaknesses, e.g., faults, without any aid of kinematic perturbations, in contrary to that suggested in earlier studies. Throughout the temporal progression of a SDDS system, the two subducting slabs mutually affect their kinematics and subduction-driven flow patterns in the mantle wedges as well as asthenosphere. 2) SDDS systems can remain active for a long period of time (>25 Myr) in an oceanic setting, and eventually attain exceptionally high convergence velocities, 16-17 cm/year during its extended period (~5 Myr) of activity, depending on the slab ages at the trenches. These model findings explain the development of a self-sustaining SDDS responsible for the anomalously Indo-Eurasian high-convergence velocity condition in the Cretaceous Neo-Tethyan evolution. 3) The presence of continental blocks in the initial plate setting greatly influences the SDDS dynamics, forcing the double subduction zones to localize preferentially at their passive margins with the oceanic plates. They grow almost equally, forming a spreading centre between the two trenches on a time scale of 25 Myr. This SDDS model accounts for the Cenozoic tectonic evolution in the eastern Neo-Tethyan region, which ultimately formed the Andaman subduction system. 4) In an initial plate configuration with multiple continental fragments or plates, the latter can cause one of the SDDS slabs to halt its motion, resulting in a double to single-subduction transformation, which occurred in the Amirante-India-Eurasia subduction system during the Cenozoic evolution of Neo-Tethys.

## Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this article.

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## Data Availability Statement

The authors confirm that all the data used to support the findings of this study are available within the article and as supplementary materials. All aspects of UNDERWORLD 2 (Mansour, J., et al., 2020) can be checked here (https://doi.org/10.5281/zenodo.6820562).

| Description | Symbol | Unit | Default Values |
|---|---|---|---|

| | | | |
|---|---|---|---|
| Thermal expansion coefficient | $\alpha$ | K$^{-1}$ | $3 \times 10^5$ |
| Thermal diffusivity | $\kappa$ | m$^2$ s$^{-1}$ | $10^{-6}$ |
| Reference density | $\rho_0$ | kg m$^{-3}$ | 3300 |
| Surface temperature | $T_s$ | K | 273 |
| Potential temperature | $T_m$ | K | 1673 |
| Adiabatic temperature gradient | $dT/dz$ | K km$^{-1}$ | 0.37 |
| Gravitational acceleration | $g$ | m s$^{-2}$ | 9.81 |
| Maximum viscosity | $\eta_{max}$ | Pa s | $1.0 \times 10^{24}$ |
| Minimum viscosity | $\eta_{min}$ | Pa s | $1.0 \times 10^{19}$ |
| Crust viscosity | $\eta_C$ | Pa s | $1.0 \times 10^{20}$ |
| | | | |
| **Dislocation creep (Upper Mantle)** | | | |
| Activation energy | $E$ | kJ mol$^{-1}$ | 540 |
| Activation volume | $V$ | cm$^3$ mol$^{-1}$ | 10 |
| Pre-factor | $A$ | Pa$^n$ s$^1$ | $4.1 \times 10^{15}$ |
| Exponent | $n$ | - | 3.5 |
| | | | |
| **Diffusion creep (Upper and Lower mantle)** | | | |
| Activation energy | $E$ | kJ mol$^{-1}$ | 300 (UM & LM) |
| Activation volume | $V$ | cm$^3$ mol$^{-1}$ | 4.5 (UM), 1.58 (LM) |
| Pre-factor | $A$ | Pa$^1$ s$^1$ | $1.87 \times 10^9$ (UM)<br>$1.77 \times 10^{14}$ (LM) |
| Exponent | $n$ | - | 1 |

| | | | |
|---|---|---|---|
| **Plastic yielding** | | | |
| Cohesion | $C_0$ | MPa | 20 |
| Friction coefficient | $\mu$ | - | 0.1 |
| Maximum yield stress | $\tau_{max}$ | MPa | 500 |

Table 1: List of model parameters and their corresponding values chosen for the numerical simulations.

# 6. Appendix A

## *6.1 Governing equations*

We build numerical, time-evolving, dynamically consistent thermomechanical subduction models in 2-D Cartesian domains within a theoretical framework of computational fluid dynamics (CFD), implemented by using the UNDERWORLD2 code (Mansour et al., 2020). This CFD simulation study assumes an incompressible Boussinesq fluid flow, approximated to the long-time (multi-million year) scale kinematic state of Earth's mantle. This approximation accounts for only buoyancy (body force term), treating all other effects of density fluctuations negligibly small in the momentum equation. We use the continuity and momentum conservation equations in our modelling to describe spontaneous flows in our model driven by density anomalies. The expressions of these equations are, respectively,

$$\frac{\delta u_i}{\delta x_i} = 0 \tag{1}$$

$$-\frac{\delta P}{\delta x_i} + \frac{\delta \sigma_{ij}}{\delta x_j} + \rho g_i = 0 \tag{2}$$

where $v_i$ is the velocity vector. In Eq. (2) the inertial forces are neglected, as applicable to long term flows in the mantle. $\sigma_{ij}$ can be decomposed into the isotropic (${\sigma^o}_{ij}$) and the deviatoric stress ($\tau_{ij}$) tensors as,

$$\sigma_{ij} = {\sigma^o}_{ij} + \tau_{ij} \tag{3}$$

such that

$${\sigma^o}_{ij} = P\delta_{ij} \tag{4}$$

$$\tau_{ij} = 2\eta\dot{\varepsilon}_{ij} = \eta\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right) \tag{5}$$

where $\dot{\varepsilon}_{ij}$ is the strain rate tensor. Substituting Eq. (4) & (5) into Eq. (2) gives rise to the Stokes equation with pressure and velocity as two unknown variable parameters. Applying the model boundary conditions, we numerically solve the continuity and Stokes equations in a pre-defined 2D Cartesian domain to derive the velocity and pressure in the model domain.

The thermal evolution of a subduction system is dictated by combined effects of advective heat transfer, thermal diffusion, and heat sources/sinks within the system, which we tackle with the following heat equation.

$$\rho C_p \frac{DT}{Dt} = \frac{\partial q_i}{\partial x_i} + \rho Q, \tag{6}$$

where $Q$ is the rate of internal heat production per unit volume, $T$ is the temperature, $\rho$ is the material density and $C_p$ is the specific heat. $q$ represents the rate of diffusional heat transfer, described by Fourier's law as,

$$q_i = -k\frac{\partial T'}{\partial x_i}, \tag{7}$$

where $k$ is the thermal conductivity and $\mathrm{T}'$ is the nonadiabatic temperature. Substituting Eq. (7) into (6) and expanding considering the definition of material derivative, gives

$$\frac{\partial T}{\partial t} + u_i.\frac{\partial T'}{\partial x_i} = \kappa \nabla^2 T' + \frac{Q}{C_p} \tag{8}$$

where $\kappa = \frac{k}{\rho C_p}$, which represents the thermal diffusivity. $T'$ is replaced by the adiabatic temperature ($T_a$) of the system as a function of depth ($z$) obtained from the relation:

$$T_a = T' + z\left(\frac{\partial T}{\partial z}\right) = \ T' + z\left(\frac{\alpha g T_p}{C_p}\right), \tag{9}$$

where $T_p$ is the mantle potential temperature and $\alpha$ is the coefficient of thermal expansion, which was set at a value of $3 \times 10^{-5}$ $\mathrm{K}^{-1}$. Considering $C_p$ = 1260 J/kg/K for Earth's mantle, a resultant adiabatic temperature gradient of 0.4 K/km is added to the nonadiabatic temperature of the mantle. The energy conservation equation (8) is solved using the Semi-Lagrangian Crank-Nicholson (SLCN) method (Spiegelman and Katz, 2006), built upon the potencies of the Crank-Nicolson scheme for diffusion and the semi-Lagrangian scheme for advection. The SLCN method has been found to be unconditionally stable, allowing large time-step sizes. We impose constant (Dirichlet) and zero-flux (Neumann) on the top and bottom boundaries, respectively to solve the energy equation. The model surface temperature is set initially at 0°C, whereas the initial model basal temperature at 1800°C.

We use half-space cooling profiles to constrain the thermal structures of model lithospheric plates corresponding to their assigned ages, considering a thermal diffusivity value of $10^{-6}$ $\mathrm{m}^2/\mathrm{s}$ (van Hunen and Allen, 2011), and a mantle potential temperature of 1400° C (Holt and Condit, 2021). The model density parameters are chosen as a function of the assigned temperatures, taking into account the thermal properties and their evolution following the momentum equation. For temperature dependent density variations, we adopt the equation of state,

$$\rho = \rho_r\big(1 - \alpha(T - T_p)\big) \tag{10}$$

where $\rho_r$ denotes the reference mantle density at the mantle potential temperature, which is set at 3300 kg/m$^3$.

### *6.2. Rheological considerations*

We model the mantle rheology in the framework of a composite creep law that combines diffusion creep ($\eta_{diff}$), dislocation creep ($\eta_{disl}$), and plastic yielding ($\eta_{yield}$). An Arrhenius temperature and pressure dependence of the activation volume (*V*) and the activation energy (*E*) (Hirth and Kohlstedt, 2004) is chosen to describe the creep laws for mantle silicates, which leads to the diffusion/dislocation-controlled viscosity,

$$\eta_{diff/disl} = A^{\frac{-1}{n}} \dot{\varepsilon}^{\frac{1-n}{n}} exp\left(\frac{E + P_1 V}{nRT_a}\right) \tag{11}$$

where A is a pre-factor, *n* is the stress exponent (n = 1 and 3.5 for diffusional and dislocation creep, respectively), R is the gas constant and $P_1$ is the lithostatic pressure. It is usually observed that lithostatic pressure enhances the yield strength of mantle material by different stress-limiting mechanisms. In contrast, material strength can be reduced as a consequence of brittle yielding (Kohlstedt et al., 1995) near the surface. These two yield mechanisms are combined into a simplified plastic rheology (van Hunen and Allen, 2011), and the viscosity is described as,

$$\eta_{yield} = \frac{\tau_{yield}}{2\dot{\varepsilon}} \tag{12}$$

and $\tau_{yield}$ denotes the yield strength, expressed by the following relation.

$$\tau_{yield} = min\ (C + \ \mu P,\ \tau_{max}\ ) \tag{13}$$

where $C$ (= 20 MPa) is the initial cohesion, $\mu$ (= 0.1) is the friction coefficient, and $\tau_{max}$ is the cut-off yield stress value. The harmonic mean of the three types of viscosity is considered to determine an effective model viscosity ($\eta_{\mathrm{eff}}$),

$$\frac{1}{\eta_{eff}} = \frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}} + \frac{1}{\eta_{yield}} \tag{14}$$

The activation volumes ($V$) and energies ($E$) values (Table 1) chosen in our modelling are consistent with the experimental estimates for dry olivine (Karato and Wu, 1993). We have considered the value of pre- factor $A$ for the upper mantle based on the following findings, 1) an effective viscosity in the order of $10^{20}$ Pa in the shallow part of upper mantle (Hager, 1990), and 2) seismic anisotropy implying dominantly dislocation creep in the majority part of mantle (Becker, 2006). Diffusional creep is recognized as the principal mechanism to determine the lower-mantle rheology (Karato & Wu, 1993). A range of estimates suggest its viscosity significantly higher than the upper-mantle viscosity. The calculated creep pre-factor yields a value of $3 \times 10^{22}$ at the 660 km transition (Čížková et al., 2012), which increases continuously with depth (Fig. 2).

The top crustal layer in our model is assigned a constant viscosity of $1 \times 10^{20}$ Pa s. We use Eq. (12) to simulate brittle failure in this layer, assuming a low value (10 MPa) of its cohesion. This rheological consideration aims to reproduce the mode crust with yield strength lower than that of the lithosphere. This modelling manipulation allows us to introduce decoupling of the subducting slab from the overriding plate and facilitate the plate convergence process. The mantle lithosphere strength is determined from the upper viscosity cut-off values ($1 \times 10^{24}$ Pa s), except for regions where we preferentially activate plastic yielding. For the overriding plate (OP), the upper limit of viscosity is increased to $2.5 \times 10^{24}$ Pa s to obtain a slightly higher stiffness. We model the crust and the lithosphere individually single layers without any internal rheological stratification. Considering the long-term process of our present concern, the effects of elastic deformation are completely excluded. We add passive tracers in the compositional field within the regions of interest to track the plate motion and deformation during the model run.

### *6.3. Model Setup*

We meshed the 2D domain by smaller quadrilateral elements with a mesh resolution of 512 elements in the vertical direction, which provides an element width of ~2 km, and a particle density of 50 tracers per element. These Lagrangian tracer particles used to track advecting materials and their corresponding physical properties are mapped to quadrature points with nearest-neighbour interpolation (Sandiford and Moresi, 2019). All models were subjected to $g = 9.8$ m/s$^2$, where $g$ is the acceleration due to gravity. The model has free-slip $(v_y = 0)$ conditions assigned to the bottom and top boundaries, whereas periodic boundary condition is imposed on the sidewalls.

For the temperature boundary condition, we impose constant (Dirichlet) and zero-flux (Neumann) on the top and bottom boundaries, respectively to solve the energy equation. We use half-space cooling profiles to constrain the thermal structures of model lithospheric plates corresponding to their assigned ages, considering a thermal diffusivity value of $10^{-6}$ m$^2$/s (van Hunen & Allen, 2011), and a mantle potential temperature of 1400° C (Holt & Condit, 2021). The model density parameters are chosen as a function of the assigned temperatures, taking into account the thermal properties and their evolution following the momentum equation.

### *6.4. Numerical Scaling, Mesh Refinement Tests and Model validation*

The governing equations (Section 6.1) and physical parameters are non-dimensionalized to implement them for the model simulations utilizing the UNDERWORLD2 scaling module. The model results, corresponding to the reference values of model parameters, are then scaled to their equivalent real physical units. The scaling coefficients derived from the base units of length (*Kl*), mass (*Km*), time (*Kt*) and temperature (*KT*) are set as, $Kl = 1$, $Km = \rho_{ref} \times Kl^3$, $Kt = 1 / (\frac{\eta_{ref} \times Kl}{Km})$ and $KT = T_p$, where $\rho_{ref}$ and $\eta_{ref}$ are the reference density and viscosity, respectively. The scaling strategy has been implemented to keep buoyancy force as the sole driving factor in subduction of an oceanic lithosphere, assigned a non-dimensional value of 1 to aid the solver efficiency.

We kept a mesh resolution of 1150 × 512 elements in our modelling. A set of resolution tests was performed for low (576 × 256, 288 × 128 elements) as well as for a much higher resolution (2300 × 1024 elements), with an objective to optimize the refined resolution appropriate for subduction initiation modelling and tracking the evolutionary stages of a double subduction system with thin, weak viscoplastic subduction interfaces, and a composite rheology of the upper mantle. The mesh resolutions and their effects on the trench and subducting plate (SP) velocities are detailed in the Supplementary Section S4.

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