Exploring the accretion of multiple microcontinents using numerical models

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Abstract

During closure of an ocean through subduction and continental collision, bathymetric highs such as microcontinents can accrete, collide, or partially or completely subduct. Such interaction of future accreted terranes (FATs) with the overriding continent will modify the dynamics of the subduction zone, affecting its length and frictional resistance, and thus the force balance of the subduction system. Accreted microcontinents and microcontinental fragments are preserved in backarcs and collisional orogens, demonstrating that multiple terranes can accrete during a single Wilson-cycle, in what is termed accretionary orogenesis. In this study, we use thermo-mechanical numerical experiments of microcontinent-continent collision events to investigate parameters that influence whether microcontinents accrete, subduct, or collide. Our results indicate that multiple accretionary episodes are possible, but that a weak basal detachment layer within each FAT is paramount for such a scenario to occur. The introduction of a microcontinent, or FAT, in the subduction zone will affect the balance between slab-pull, far-field forces, and the subduction interface resistance. The strength (and rheological stratification) of the FATs determines the evolution of the subduction interface resistance throughout the collision event, exerting a first order control on the resulting geodynamic scenario. Collision with a strong FAT significantly increases the subduction interface resistance promoting terrane subduction and localization of deformation away from the subduction interface. In turn, collision with a weak FAT increases subduction interface resistance only mildly, allowing for multiple accretion events.
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Key Points:

• Accreting or subducting microcontinents strongly influence subduction dynamics through their interaction with the subduction interface.
• The accretion of multiple microcontinents is only possible when they are rheologically weak, incorporating a lower crustal detachment.
• Multiple microcontinents in a subduction zone can lead to either partial accretion, full subduction, or divergent double subduction.
Abstract

During closure of an ocean through subduction and continental collision, bathymetric highs such as microcontinents can accrete, collide, or partially or completely subduct. Such interaction of future accreted terranes (FATs) with the overriding continent will modify the dynamics of the subduction zone, affecting its length and frictional resistance, and thus the force balance of the subduction system. Accreted microcontinents and microcontinental fragments are preserved in backarcs and collisional orogens, demonstrating that multiple terranes can accrete during a single Wilson-cycle, in what is termed accretionary orogenesis. In this study, we use thermo-mechanical numerical experiments of microcontinent-continent collision events to investigate parameters that influence whether microcontinents accrete, subduct, or collide. Our results indicate that multiple accretionary episodes are possible, but that a weak basal detachment layer within each FAT is paramount for such a scenario to occur. The introduction of a microcontinent, or FAT, in the subduction zone will affect the balance between slab-pull, far-field forces, and the subduction interface resistance. The strength (and rheological stratification) of the FATs determines the evolution of the subduction interface resistance throughout the collision event, exerting a first order control on the resulting geodynamic scenario. Collision with a strong FAT significantly increases the subduction interface resistance promoting terrane subduction and localization of deformation away from the subduction interface. In turn, collision with a weak FAT increases subduction interface resistance only mildly, allowing for multiple accretion events.

Plain language summary

This study explores what happens when small landmasses, called microcontinents, collide over millions of years with larger continents. The research, conducted through computer simulations, focuses on understanding when microcontinents either attach to the larger continent, get pushed under it, or collide with it. We were especially interested in finding out what happens when multiple microcontinents meet a continent one after the other. The findings suggest that whether a microcontinent attaches or not depends on how strong a layer it has underneath it. A weak layer allows for attachment, while a strong one leads to subduction. This interaction between microcontinents and larger continents affects the overall force balance in the tectonic system, possibly influencing deformation of Earth’s outer layer far away from the zone of collision. Understanding these processes helps us grasp the complexities of plate tectonics and how continents have formed and shifted over millions of years.
1. Introduction

The Wilson Cycle describes oceans closing and opening along their inherited sutures and has proven to be a powerful concept in analyzing the deformational histories of plate margins [Wilson, 1966]. Its initial formulation described a process of oceanic subduction, continental collision, continental extension and formation of a new ocean for the North Atlantic Region. In its focus on the tectonic history of the continental margins, the Wilson Cycle concept considered the ocean plates as fairly simple, empty basins. Oceanic plates, however, feature multiple bathymetric highs, such as microcontinents, island arcs, seamounts, or oceanic plateaus (together termed future allochthonous terranes, or FATs by Tetreault and Buiter [2012]. We here ask the question how the convergence stage of the Wilson Cycle is influenced by accretion of multiple terranes in the form of microcontinents.

Microcontinents such as Jan Mayen in the north-east Atlantic Ocean or the Seychelles in the Indian Ocean are regions of continental crust embedded within an oceanic plate and surrounded by oceanic crust. They are common oceanic features that form during rifting events on continental rifted margins and retreating subduction margins. Microcontinent accretion during subduction is a critical geological process as it is one of the main contributors to continental crustal growth [Cawood et al., 2009; Clift et al., 2009; Stern and Scholl, 2010]. Ongoing continental growth can be observed in modern collision zones such as the Pacific accretionary belt, incorporating the collision of the Philippine microplate, and the Taiwan-Luzon-Mindoro Belt [Clift et al., 2009]. Examples of active subduction zones with a recent history of accretion can be found in the North American Cordilleras [Colpron et al., 2007; Coney et al., 1980; Gabrielse and Yorath, 1991]. Microcontinental terranes can also be preserved in orogens during the final continent-continent collisional stage as inferred, for example in the European Alps [e.g. Kissling et al., 2006; Schmid et al., 1996]. Finally – to follow the path of the Wilson Cycle – we can look at the Scandinavian Caledonides for an example where collisional orogenesis that incorporated accreted terranes was followed by continental rifting [Andersen et al., 2022; Johannes Jakob et al., 2019; J. Jakob et al., 2022; Roberts, 2003] and references therein). Multiple accreted terranes are preserved in several of these systems, demonstrating that sequential terrane accretion is possible in nature. Yet, just by looking at the geological evidence, there is no way of knowing how prevalent terrane accretion is relative to microcontinent subduction, whether traces of an accretion event can be erased, under what circumstances multiple terranes accrete or how a phase of continent-continent collision can affect the preservation of earlier terrane accretion events.
When FATs are accreted they significantly alter the subduction zone dynamics before continent-continent collision could occur. This process of FAT collision involves a balance of geodynamic forces that can result in either FAT subduction, FAT accretion, or subduction stalling (which could lead to the initiation of an altogether new subduction zone in the form of subduction zone transference [Stern and Gerya, 2018]. The most important driving forces in this system are the slab-pull force arising from the negative buoyancy of the down-going slab and the far-field force which is the result of large-scale plate-motions external to the subduction zone. These forces are counteracted (among others) by friction along the subduction interface, slab bending, viscous resistance of the mantle, and the buoyancy of the down-going FAT. The net force balance controls the overall stress-field of the overriding plate as well as the state of stress and potential deformation of any FATs embedded within the oceanic lithosphere that are not yet in the subduction zone. When multiple FATs are embedded in the subducting oceanic plate, the friction along the subduction interface can vary considerably with time. This is because the accreting FATs have a first order effect on the length and the rheology of the subduction channel, thereby controlling the interface friction. The fate of the FATs (e.g. full or partial accretion, or subduction) also affects the overall buoyancy of the slab, altering the balance of forces through the slab-pull as well.

Cloos [1993] analyzed FAT subduction or accretion based on buoyancy calculations and concluded that bodies of continental and oceanic island-arc crust with more than 15 km crustal thickness can jam a subduction zone, while oceanic plateaus are inherently subductible as they are less buoyant than the oceanic lithosphere and seamounts typically only cause temporary dents. Ellis et al. [1999] were among the first to run thermo-mechanical numerical experiments exploring the dynamics of terrane accretion. Their models on lithosphere-scale (without a mantle) focused on the formation and accretion of basement fold nappes, showing that a weak detachment is required for the accretion of a terrane. Tetreault and Buiter [2012] modelled in detail how a single FAT behaves as it enters the subduction zone confirming that accretion occurs when there is a detachment layer in the FAT, that the depth of the detachment layer controls the amount of continental crust that is accreted to the overriding plate and that terrane buoyancy does not prevent subduction of FATs unless they are very long. These results were also reproduced by Vogt and Gerya [2014], Tao et al. [2020] and Gün et al. [2022]. Here we aim to build on the work of Tetreault and Buiter [2012] and – using 2D thermo-mechanical numerical experiments – explore the roles of the structure and rheology in the accretion of
multiple microcontinents to better understand how accretionary orogenesis modifies the subduction zone and affects later continental collision.

2. Modelling method

2.1. Setup and rheology

We perform 2D numerical experiments on a Cartesian region that is 660 km deep and 2300 to 2800 km wide (figure 1) using the thermo-mechanical finite-element code SULEC v.4 [Ellis et al., 2011; Naliboff and Buiter, 2015; Naliboff et al., 2017; Tetreault and Buiter, 2012; 2018]. Details of the numerical method are provided in Supplementary material 1. The numerical resolution of the model varies along the profile as well as with depth with the highest resolution of 1 km by 1 km reached around the subduction zone (figure 1).

![Experimental setup diagram](image)

Figure 1: Experimental setup. a) The initial geometry of a model containing two microcontinents with close-ups of a representative ocean-microcontinent margin and the incoming ocean-continent margin. The black-numbers along the edges represent distance in kms measured from the top left corner. The light-gray numbers along the edges represent the numerical resolution in kms. The red arrows along the sides represent the velocity boundary conditions. The inset in the top left corner shows characteristic initial geotherms. b) Analytically calculated strength-profiles of the oceanic lithosphere and the continental lithosphere, with dashed lines showing the strain-weakened profiles. c) Analytically calculated strength-profiles of the microcontinents, with dashed lines showing the strain-weakened profiles.
Our initial geometry consists of an oceanic plate between two continents. The oceanic plate is either “empty” or one or two microcontinents are embedded within it. In the description and analysis of the model-evolutions we will refer to these microcontinents as terranes only when they accrete to the overriding plate. The materials behave either in a viscous or a strain-weakening frictional-plastic manner that simulates brittle behavior and allows for efficient localization of deformation (Table 1). The mantle has a rheology of wet olivine combined diffusion and dislocation creep [Karato and Wu, 1993; van den Berg et al., 1993]. The continental plates consist of an 80 km-thick mantle lithosphere (wet olivine dislocation creep with a scaling factor $f = 5$; Karato and Wu [1993]), a 20 km-thick lower crust (wet anorthite dislocation creep; Rybacki et al. [2006]) and a 20 km-thick upper crust (wet quartz; Gleason and Tullis [1995]). The oceanic domain consists of a 73 km-thick mantle lithosphere (wet olivine dislocation creep with a scaling factor $f = 5$; Karato and Wu [1993]), a 5 km thick oceanic crust (gabbro dislocation creep; Wilks and Carter [1990]) and a 2 km thick sedimentary cover (the same gabbro as the oceanic crust but with a lowered cohesion and internal angle of friction). A simple eclogite phase transition is implemented for subducting oceanic crust, sediments and the lower crust of microcontinents (Appendix A). Subduction is initiated with a “compositionally weak seed” between the overriding continental domain and oceanic domain representing a proto-subduction zone with a dip of 30°. The weak seed consists of oceanic crustal material overlain by a layer of oceanic sediments, underplating a sloped continental margin [Tetreault and Buiter, 2012]. The lower strength of these materials compared to the surrounding mantle lithosphere allows for deformation to localize in this region as the subducting plate is pushed toward the continent. The thickness of our initial weak seed (7 km) is based on the thickness of seismically imaged subduction channels (2-8 km) [Abers, 2005]. The margin of the trailing continent is dipping steeply at an angle of 60°. The model is initialized so that the oceanic floor sits 5 km below the surface of the continents.
Table 1 Mechanical and thermal material properties used in the experiments. 

Table 1

<table>
<thead>
<tr>
<th>Units</th>
<th>Symbol</th>
<th>Reference model</th>
<th>M1s</th>
<th>M1i</th>
<th>M1w</th>
<th>M2s</th>
<th>M2i</th>
<th>M2w</th>
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<td>20</td>
<td>8</td>
<td>10</td>
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<td>10</td>
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<td>2800</td>
<td>3170</td>
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<td>Pre-exponential factor ($\text{Pa s}^{-1}$)</td>
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<td>1.19 10$^8$</td>
<td>1.19 10$^8$</td>
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<td>3.23 10$^4$</td>
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<td>Grain-size exponent</td>
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<td>1.5 10$^{-2}$</td>
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<tr>
<td>Water content exponent</td>
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<tr>
<td>Water content exponent</td>
<td>$t$</td>
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<td>2.5</td>
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<td>2.5</td>
<td>2.5</td>
</tr>
<tr>
<td>Thermal expansion ($\text{C}^{-1}$)</td>
<td>$\alpha$</td>
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<td>0.25 10$^{-5}$</td>
<td>0.25 10$^{-5}$</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<tr>
<td>Thermal conductivity ($\text{W m}^{-1} \text{C}^{-1}$)</td>
<td>$\lambda$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
<td>1.30 10$^6$</td>
</tr>
<tr>
<td>Viscosity range</td>
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<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

The reference model $R_0$ features an “empty” oceanic domain situated between the two continents. Models $M_1s$, $M_1i$ and $M_1w$ contain one microcontinent embedded within the oceanic lithosphere while models $M_2s$, $M_2i$ and $M_2w$ contain two microcontinents. All microcontinents rise 2.5 km above the oceanic sea-floor and consist of three layers: a 7 km thick gabbroic lower crust [Wilks and Carter, 1990], a 10 km thick wet anorthite mid-crust [Rybacki et al., 2006] and an 8 km thick wet quartz upper crust [Gleason and Tullis, 1995]. The characters $s$, $i$ and $w$ denote strong, intermediate and weak in both sets. In case of the intermediate ($i$) and weak ($w$) microcontinents, the gabbroic lower crust is assumed to host preexisting weak zones and in order to mimic this state, the yield strength of this layer is lowered by the assignment of reduced values for cohesion and internal angle of friction. In models $M_1w$ and $M_2w$, the respective thicknesses of the lower and mid-crust are reversed to represent a thicker, more efficient detachment layer and an overall weaker rheology. The oceanic basin between the trench and the first microcontinent is 400 km wide, each microcontinent is 200 km wide and the two microcontinents are separated by a 300 km wide basin. The trailing continent is 250 km away from the second microcontinent (figure 1).

2.2. Kinematic and thermal boundary conditions

The kinematic boundary conditions for our experiments consist of a free top surface, free slip on the side walls and base of the model, an inflow velocity at the right continental lithosphere, and equivalently balanced outflow velocities on both side walls across the sub-lithospheric
mantle (figure 1). A weak diffusive erosion and sedimentation is applied on the top surface of the models to approximate surface processes [Culling, 1960]. All experiments are performed with a constant 5 cm yr\(^{-1}\) inflow velocity.

We assign a set of thermal parameters in our numerical experiments (table 1) and after solving for an initial steady-state, we allow the thermal field to evolve. The temperature is set to a constant 0 °C at the surface of the model and 1435 °C at the base of the model, while the side-boundaries are thermally insulated. In order to mimic active mantle convection at high Nusselt number, the thermal conductivity of the sublithospheric mantle is stepwise increased from 2.5 to 87.6 Wm\(^{-1}\)K\(^{-1}\) when above a temperature threshold of 1,300 °C. This approach has been widely used in subduction as well as rift modeling studies [Butler and Beaumont, 2017; Pysklywec and Beaumont, 2004; Tetreault and Buiter, 2012; Warren et al., 2008] as it prevents secular cooling of the model domain while maintaining a constant vertical heat flux at the base of the lithosphere and keeping the mantle close to the adiabatic gradient.

To prevent the subducting slab from interacting with the bottom boundary of the model, the slab materials are arbitrarily transformed into sub-lithospheric mantle when reaching approximately 625 km depth (at 200 GPa pressure and 900 °C temperature). This simplistic approach somewhat limits the slab-pull force but the present setup still reaches values up to \(2.5 \times 10^{13}\) Nm\(^{-1}\) (comparable to values calculated for other, similar studies; e.g. Erdős et al. [2021]; Wolf and Huismans [2019]), while keeping the computational costs of running these experiments manageable by allowing for a shallower computational domain.

### 3. Results

Here we present a brief summary of key observations from each of the 7 model experiments. The temporal evolution of all experiments is described in more detail in Supplementary material 2 (see also table 2) where Supplementary Animations (SA1 to SA7) and Supplementary Viscosity Figures (S1 to S8) complementing figures 2 to 7 are also available.

<table>
<thead>
<tr>
<th></th>
<th>Microcontinent behaviour</th>
<th>Oceanic lithosphere behaviour</th>
<th>Trench behaviour</th>
</tr>
</thead>
<tbody>
<tr>
<td>R0</td>
<td>n.a.</td>
<td>Subduction</td>
<td>Continuous, stable</td>
</tr>
<tr>
<td>M1s</td>
<td>Subduction, detachment and accretion under the overriding plate, likely affecting volcanic signature.</td>
<td>Continuous subduction with microcontinent.</td>
<td>Continuous, stable subduction.</td>
</tr>
<tr>
<td>M1i</td>
<td>Upper-middle crustal accretion by underplating. No large thrust sheets.</td>
<td>Continuous subduction despite terrane accretion event.</td>
<td>Trench transferred during terrane accretion.</td>
</tr>
<tr>
<td>Model</td>
<td>Summary</td>
<td></td>
<td></td>
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<tr>
<td>-------</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>M1w</td>
<td>Upper-middle crustal accretion by underplating. No large thrust sheets. Continuous subduction despite terrane accretion event. Trench transferred during terrane accretion.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>M2s</td>
<td>Subduction, detachment and accretion under the overriding plate of both, likely affecting volcanic signature. Divergent double subduction prior continent-continent collision. Continuous subduction that slows with new subduction initiation.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>M2i</td>
<td>Gradual subduction and underplating of the first and upper-middle crustal accretion by underplating of the second. Thickening of the brittle part of the mantle-lithosphere below the 2nd microcontinent. Trench transferred during the accretion of the second terrane.</td>
<td></td>
<td></td>
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<tr>
<td>M2w</td>
<td>Upper-middle crustal accretion by underplating of both. The frontal edges broken at depth. Continuous subduction despite consecutive terrane accretion event. Trench transferred during both accretion events.</td>
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</tr>
</tbody>
</table>

**Table 2 A summary of the unique behavior of key areas of the experimental domain for each presented model.**

In the reference model, R0, the “empty” oceanic plate subducts with deformation localized along the subduction interface before continent-continent collision occurs (figure 2). We observe that the rheological stratification of the microcontinent in experiments with a single microcontinent affects whether it becomes an accreted terrane or not. During the evolution of model M1s, the strong microcontinent embedded within the oceanic lithosphere is subducted and partially accreted at the base of the overriding plate. During microcontinent subduction, minor amounts of deformation localization are observed at the incoming continental margin well before the onset of continent-continent collision (figure 3). Conversely, in model M1i, the intermediate strength microcontinent is accreted to the overriding plate as the weak basal detachment is activated resulting in a trench-jump to the trailing edge of the microcontinent. At the end of the experiment the terrane remains sandwiched between the two colliding continents (figure 4). Model M1w shows a very similar evolution to model M1i (supplementary figure S1).
Figure 2: Reference model R0 exhibiting subduction. (a–c) Material colors (see legend of Figure 2) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$). Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure 3: Model M1s (one strong microcontinent) exhibiting terrane subduction. (a–d) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$). Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure 4: Model M1i (one intermediate strength microcontinent) exhibiting terrane accretion. (a–d) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} s^{-1}). Green triangle shows location of the “trench” defined as the deepest point in the topography.

We observe unique subduction and accretion dynamics when multiple terranes are accreted, depending again on the initial rheological stratification of the microcontinents. When two strong microcontinents are embedded within the oceanic lithosphere (model M2s) both microcontinents subduct similarly to the one microcontinent in model M1s. However, during the subduction of the second microcontinent a new subduction zone is initiated along the margin of the incoming continent creating a divergent double subduction system (figure 5).
Figure 5: Model M2s (two strong microcontinents) exhibiting subduction of both terranes and subduction initiation. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} s^{-1}). Green triangle shows location of the “trench” defined as the deepest point in the topography.

In contrast, during the evolution of the model containing two intermediate strength microcontinents (model M2i), the first microcontinent is initially accreted, progressively thinned and then eroded from the subduction zone, while the second microcontinent
experiences a phase of shortening well before it reaches the subduction zone. At the end of the experiment the second microcontinent remains sandwiched between the two colliding continents (figure 6).

Figure 6: Model M2i (two intermediate strength microcontinents) exhibiting subduction of the first terrane, intra-oceanic terrane deformation and accretion of the second terrane. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} m s^{-1}).
Green triangle shows location of the “trench” defined as the deepest point in the topography.

Finally, in the model with two weak microcontinents (model M2w) both microcontinents are accreted to the overriding plate, with the active subduction interface hopping over to each accreting microcontinent’s basal detachment during accretion. The two accreted terranes form a crustal stack between the two colliding continents at the end of the experiment (figure 7).

Figure 7: Model M2w (two weak microcontinents) exhibiting consecutive terrane accretion events. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C)
and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$).

Green triangle shows location of the “trench” defined as the deepest point in the topography.

4. Force-balance considerations

4.1. Interplay between slab-pull, tectonic boundary forcing and interface resistance

To understand the evolution of the experiments showing subduction or accretion of two microcontinents in a quantitative manner, we calculated the temporal evolution of the main driving and resisting forces in the subduction system [for examples of similar approaches see Erdős et al., 2021; Forsyth and Uyeda, 1975; Wolf and Huismans, 2019]. We consider the slab-pull force (resulting from the negative buoyancy of the slab), the integrated tectonic boundary forces (the depth-integrated horizontal deviatoric stresses within the lithosphere on both side-boundaries) and the interface resistance (using the second invariant of the deviatoric stress-tensor integrated along the interface as a proxy; see Supplementary material 3 for a detailed description of the procedures used to calculate these forces). We note that other resisting forces, like the bending resistance of the oceanic lithosphere and the viscous shear resistance of the mantle also play a role, but as they have been shown to be of second order magnitude, we will neglect them during the following discussion [see Erdős et al., 2021; Forsyth and Uyeda, 1975; Wolf and Huismans, 2019]. We have analytically calculated the integrated strength of the undeformed continental lithosphere, the oceanic lithosphere and the microcontinental lithospheres using a constant reference strain rate of $1 \cdot 10^{-16}$ s$^{-1}$. These will help us better understand which lithospheric segments might localize deformation most efficiently.

The slab-pull forces steadily increase as the slabs get longer and reach peak values of 2 to $2.5 \cdot 10^{13}$ Nm$^{-1}$ (figure 8a). The slab-pull force in models with microcontinents starts to deviate from the reference model R0 at around 12 Myr, which corresponds to the stage when the first microcontinents reach a depth of 120 km or deeper (figure 8a). This difference is the result of compositionally lighter crustal material of the microcontinent replacing cold, negatively-buoyant oceanic mantle-lithosphere. This positively-buoyant crustal material cannot subduct to great depths and therefore detaches and accretes to the bottom of the overriding plate, leading to a reduction in the overall pull-force.
Figure 8: Key driving and resisting forces. a) The temporal evolution of the slab-pull force calculated for models R0, M2s, M2i and M2w, showing minimal deviations between them. b) The temporal evolution of the slab-pull force, the integrated tectonic boundary forces and the interface resistance for model M2i, with key events marked. c) The temporal evolution of the slab-pull force, the integrated tectonic boundary force and the interface resistance for models M2s, M2i and M2w. The Integrated tectonic boundary force of R0 up to the onset of continent-continent collision is plotted as a grey line.

The most significant deviation from the continuous increase of slab-pull forces with time is the late stage drop in slab-pull force for model M2s, where the initiation of a new subduction zone reduces the force exerted by the original slab. Based on the above considerations, we argue that variations between the slab-pull forces for the analyzed three models are negligible, hence the differences in the evolution of tectonic boundary forces between the models are largely
independent of slab-pull and reflect a response to the changes in interface resistance and
changes in the integrated strength of different lithospheric bodies.

When looking at common trends within the experiments (using model \textbf{M2i} as an intermediate example) we observe that the slab-pull force briefly becomes dominant over the integrated tectonic boundary forces around the time the first microcontinent arrives at the trench (figure 8b). Subsequently, as the microcontinent enters the subduction channel the interface resistance and the tectonic boundary forces both increase (figure 8b). In all three models (\textbf{M2w}, \textbf{M2i}, \textbf{M2s}), the peak values of tectonic boundary forces are recorded when a microcontinent is in the subduction channel (see force-material co-evolution animations SA8 to SA10 in the supplements). Subsequently, the tectonic boundary forces gradually decrease, as the deformation shifts back from either the top of the subducting microcontinent (e.g. in model \textbf{M2s}), the bottom of the accreting terrane (e.g. in model \textbf{M2w}) or the broad zone of internal deformation that covers the microcontinent in the channel (e.g. in model \textbf{M2i}). We deduce that the increase in interface resistance is the result of the microcontinental crustal material replacing the oceanic crust and sediment that was lubricating the subduction channel. With increased interface resistance, more energy is dissipated along the interface, hence more force is required on the boundaries to keep the prescribed steady pace of convergence. \textit{From this observation, we can infer that a significant slab-pull and/or far-field compressional forcing is necessary in nature for subduction to continue during terrane accretion.}

Alternatively, when sufficient boundary force is not available, we predict that subduction would slow down, or even stall entirely. This phenomenon where an increase in mechanical coupling between the subducting and overriding plates leads to subduction stalling has been observed both in analogue [Duarte et al., 2015] and numerical [Afonso and Zlotnik, 2011; De Franco et al., 2008] models that do not impose fixed boundary forces. Supporting this interpretation is the previous work by De Franco et al. [2008], who have modelled the collision between an overriding continent and an incoming continental margin and found that the nature of the subduction interface has a first order role in the dynamics of the collision. Their results are in agreement with our assessment that successive accretion of multiple terranes is dynamically complex, because the rheology of the subduction channel changes with every accretion event.

The model with two strong microcontinents (\textbf{M2s}) exhibits the most pronounced changes in the competing forces, with the highest values for both interface resistance and integrated
tectonic boundary force within the set (figure 8c). In turn, as the microcontinents become weaker (i.e. experiments M2i and M2w), the increase in interface resistance diminishes, and less boundary forcing is necessary to achieve the prescribed convergence rate.

4.2. Implications of the interplay between the driving and resisting forces

The high values of integrated tectonic boundary forces in models M2s and M2i imply that the different lithospheric segments (i.e. continental, oceanic and microcontinental domains) transmit high compressional stresses towards the actively deforming areas. In experiment M2s, the interface resistance peaks during the subduction of the second microcontinent. During this phase, the interface resistance (approximately $8.5 \cdot 10^{12}$ Nm$^{-1}$) is higher than the integrated strength of the continental lithosphere ($7.6 \cdot 10^{12}$ Nm$^{-1}$). As a result, it is energetically more efficient to localize deformation within the incoming continental lithosphere than it is to keep the subduction going at the same pace. The initial diffuse, low strain-rate deformation that was present during the subduction of the first microcontinent facilitates strain-weakening during the subduction of the second microcontinent. Strain-weakening, in turn, makes the localization of deformation even more efficient, creating a positive feedback loop and resulting in subduction initiation. The deformation is effectively partitioned between the two deforming zones as the original subduction zone remains active throughout the birth of the new zone, albeit at a reduced intensity.

In experiment M2i, the interface resistance peaks during the partial accretion of the first microcontinent. During this phase, the accreting terrane is extensively deformed, resulting in an interface resistance of approximately $5 \cdot 10^{12}$ Nm$^{-1}$, which is comparable to the integrated strength of the undeformed second microcontinent ($6.3 \cdot 10^{12}$ Nm$^{-1}$). The integrated strength of the continental lithosphere remains comparatively high at $7.6 \cdot 10^{12}$ Nm$^{-1}$. As a result, the incoming second microcontinent undergoes increasingly localized shortening that starts while it is still over 200 km away from the subduction interface. It is also notable that whereas in model M1i the first and only microcontinent is accreted, in experiment M2i the first microcontinent largely subducts. The difference between the evolution of the two models starts with the onset of deformation within the second microcontinent of model M2i, when the first microcontinent reaches the subduction zone. Since strain is being partitioned between several distinct zones of deformation, the strain-rate remains lower in the first microcontinent as it lingers in the subduction channel. The lower strain-rate results in more diffuse internal deformation that prevents the efficient localization of displacement at the base of the microcontinent and ultimately leads to its subduction rather than its accretion.
In experiment M2w, the interface resistance remains much lower than the integrated strength of the undeformed lithospheric domains: the peak interface resistance is approximately $3 \cdot 10^{12}$ Nm$^{-1}$ and $5 \cdot 10^{12}$ Nm$^{-1}$ during the first and second accretion events, while the integrated strength of the microcontinents and the continents are $5.8 \cdot 10^{12}$ Nm$^{-1}$ and $7.6 \cdot 10^{12}$ Nm$^{-1}$ respectively. As a result, deformation remains localized solely in the subduction zone throughout the experiment.

**Figure 9:** Close-up from the models containing microcontinents. The first column shows convergence of one microcontinent (models M1s, M1i and M1w on subfigures a, c and e respectively). Note how the microcontinent subducts in model M1s (strong) and accretes in M1i (intermediate) and M1w (weak). The second column shows convergence of two microcontinents (models M2s, M2i and M2w on subfigures b, d and f respectively). Note the different behaviors: microcontinent subduction and subduction initiation in M2s (strong), accretion of an already deformed terrane in M2i (intermediate) and consecutive accretion of two previously undeformed terranes in M2w (weak). The slab remains continuous without a break-off event in all six models.

The experiments presented here demonstrate that a variety of tectonic scenarios can ensue during a sequence of accretion events that depend on the inherent structure (e.g. presence of a weak detachment layer) and strength of the lithospheric domains at play (figure 9). The fixed subducting-plate velocity – relative to a stable overriding plate – prescribed as a boundary condition represents areas where global plate velocities have a first order effect on the plate
movements within the system. In our particular setup, the natural sinking velocity of the subducting lithosphere generated by slab-pull forces cannot accommodate the relatively high subducting-plate velocity and additional boundary forces are required to maintain steady convergence even in the absence of continent-microcontinent collision (figure 8c). This compressional stress is transmitted efficiently through the strong oceanic lithosphere and localizes in the subduction channel. When a microcontinent enters the subduction zone the increase in interface resistance triggers an increase in the boundary forces. When another microcontinent is embedded within the oceanic plate the stress transmitted through the plate can be high enough to cause the internal deformation of the microcontinent (experiment M2i; figure 9d) and in the most extreme scenario presented here, to initiate a new subduction zone (experiment M2s; figure 9b).

5. Discussion

We observe the following large-scale behaviors of accretionary orogenesis that depend on the rheological stratification of the microcontinents: First, when the microcontinents are rheologically sufficiently weak, the subduction interface resistance is only mildly increased when a terrane enters the channel. As a result, multiple terrane accretion events can occur without significant shortening elsewhere in the system (behavior 1; experiment 2Mw). Second, when the microcontinents are stronger, the subduction interface resistance increases significantly during terrane accretion making the subduction zone less efficient at localizing shortening. Deformation will be distributed in the system as shortening localizes on either the second incoming microcontinent, embedded in the oceanic lithosphere (behavior 2a; experiment 2Mi) or along the incoming continental margin (behavior 2b; experiment 2Ms), depending on which one of these two zones has the lower integrated lithospheric strength. In the following subsections we will discuss these large-scale behaviors separately.

5.1. Terrane accretion events

The first numerical modelling study that examined terrane accretion was by S Ellis et al. [1999]. One of their findings was that a weak detachment is necessary for the accretion of a terrane. These early results on the lithosphere scale were reproduced with models that included the upper mantle by Tetreault and Buiter [2012], Vogt and Gerya [2014], Tao et al. [2020] and Gün et al. [2022]. In their study, Tetreault and Buiter [2012] examined the convergence of different types of FATs embedded in an oceanic plate that is subducting beneath a continental plate and found that (1) accretion is favored when there is a weak layer or detachment fault
within the FAT and (2) different accretionary styles are controlled by the depth of detachment in the FAT crust.

Our experiments also reproduce this behavior: in experiment **M1s**, there is no weak detachment layer within the microcontinent and it subducts before the closure of the oceanic domain (figure 9a). In contrast, in experiment **M1i**, the weak basal layer of the microcontinent acts as a detachment zone, forming a new subduction interface that allows for the efficient transfer of the trench to the oceanward edge of the accreting terrane (figure 9c). Most of the lower crust is transformed to eclogite upon reaching the requisite pressure-temperature conditions and subsequently subducts, but the middle and upper crust of the microcontinent is accreted. However, in experiment **M2i**, which has an identical microcontinent architecture to that of **M1i**, the first microcontinent subducts during the experiment, suggesting that a weak basal detachment does not guarantee accretion. In this particular experiment following an initial phase of terrane accretion, the accreted material is remobilized through a network of faults and gradually subducts. This behavior is coeval with the internal deformation of the second microcontinent, suggesting that deformation elsewhere within the model-domain has a first order effect on subduction dynamics through the partitioning of strain and strain-rate within the system.

Our models show that multiple terrane accretion events can occur in the same subduction system, and that the microcontinental crustal material is preserved as accreted terranes at the surface or at shallow depth below the tip of the original overriding plate (e.g., experiment **M2w**; figure 9f). This is possible when there is a weak layer at the base of the FATs and any already accreted terranes do not increase the interface resistance beyond the point where it would become more efficient to localize deformation elsewhere in the system than it is to keep deformation focused along the already established subduction zone.

Our models also demonstrate that terrane accretion does not necessarily result in a transfer of the subduction zone through slab break-off and the initiation of a new subduction zone; a process that has been invoked in tectonic evolution models and described in *Stern and Gerya* [2018] as well as several recent modelling studies such as *Zhong and Li* [2020] and *Yan et al.* [2024]. Instead of break-off, the slab can remain continuous as the accreted terrane is detached from its mantle-lithospheric root, transferring the trench efficiently to the oceanward edge of the terrane. This process has been demonstrated in numerical studies of single terrane accretion...
[Tetreault and Buiter, 2012; Vogt and Gerya, 2014; and Gün et al., 2022] and our models show that this event may occur multiple times within a single subduction system.

5.2. Intra-oceanic deformation

Gün et al. [2022] and Gün et al. [2024] have shown that a fixed, low subducting-plate velocity can result in microcontinent extension. Their result is substantiated by our force-balance considerations. Specifically, a fixed, low subducting-plate velocity cannot keep pace with the natural sinking velocity of the subducting slab, creating a tensional stress-field in the subducting plate. The higher the difference between the natural sinking velocities of the slab and the velocity of the incoming plate at the Earth’s surface, the stronger the resulting tensional stress-field becomes. Such a difference in velocities can result from global plate-tectonic dynamics. Extensional deformation within the microcontinent embedded in the oceanic domain only occurs when the stresses are high enough to overcome its integrated strength. Such extensional deformation of a FAT has been reported from the Ontonga Java Plateau [see and Gün et al., 2024 references therein] or the Shatsky Rise [Zhang et al., 2015] among others.

Figure 10: A pulley-system, explaining the role of the major players determining where deformation localizes. The weight marked ‘a’ symbolizes the slab-pull force, the sliding piston marked ‘b’ symbolizes the interface resistance and the material in the subduction channel that can deform internally (i.e. an accreting microcontinent), the two pistons marked ‘c’ and ‘d’ symbolize a microcontinent and the incoming continental margin that can both localize deformation if the far-field force (marked ‘e’ on the figure) is larger than their respective integrated lithospheric strength. A snapshot of the material domain...
of experiment 1Mi with the corresponding areas marked a-e is included below the pulley-system.

We argue here, that the system is more complex than the simple pulley mechanism invoked by Gün et al. [2024]. Our analysis shows that the subducting-plate velocity (relative to a fixed overriding plate), the natural sinking velocity and the integrated strength of the microcontinent are not the only controlling factors that determine whether intra-oceanic deformation can occur. The resistance along the subduction interface also plays a crucial role (figure 10). A relatively high subduction-interface resistance can dissipate a large portion of the slab-pull force, resulting in a lower effective sinking velocity for the slab, reducing the stress transmitted through the oceanic lithosphere and diminishing the potential for intra-oceanic extensional deformation.

Furthermore, if a lower effective sinking velocity is combined with a high subducting-plate velocity, the overall stress-field of the subducting plate becomes compressional. If the compressional stress transmitted through the plate overcomes its integrated strength – for example in an embedded microcontinent – shortening can occur. Such intra-oceanic shortening has been reported from the West Somali Basin [Sauter et al., 2018], from the Central Indian Basin [Bull, 1990; Delescluse et al., 2008] or from the Wharton Basin [Carton et al., 2014]. These examples generally involve the reactivation of pre-existing oceanic fracture zones that lower the integrated strength of the lithosphere, allowing for lower stresses to induce localized deformation.

5.3. Divergent double subduction

An unexpected result from this study was the divergent double subduction following microcontinent subduction. The conceptual model of divergent double subduction has been used to explain the Lachlan fold belt of southeastern Australia [Soesoo et al., 1997], the closure of the Bangong ocean and collision of the Lhasa and Qiangtang terranes in China [Zhu et al., 2016] and has been seismically imaged in the Molucca Sea in eastern Indonesia [Fan and Zhao, 2018; Puspito et al., 1993; Silver and Moore, 1978; Zhang et al., 2017]. Most of the examples of inferred divergent double subduction are Mesozoic or older in age but Soesoo et al. [1997] established a set of tectonic, metamorphic, and petrological signatures that allow for a distinction between single versus double divergent subduction zones during continent-continent collision.
The Molucca Sea is the only active example of this tectonic regime, but the collision happens between two arcs rather than two continents and there is no indication of terrane subduction along either subduction zones. Nevertheless, it is worth pointing out that the late stage evolution of our experiment 2Ms reproduces some of the observations remarkably well. In particular, two slabs of significantly different lengths are imaged by several tomographic studies of the region [Fan and Zhao, 2018; Hall and Spakman, 2002; Puspito et al., 1993; Widiyantoro and Hilst, 1997], and their trenches are buried by an accretionary complex at shallow depth [Zhang et al. 2017 and references therein].

In case of the closure of the Bangong ocean, the proposed divergent double subduction happens between two microcontinents, the Lhasa terrane towards the south and the Qiangtang terrane towards the north [e.g., Yin and Harrison, 2000]. The suture zone and its relationships with the bounding terranes provides a window into the rift, drift and accretion-related tectonism, magmatism, sedimentation, and metamorphism associated with the fragmentation of Gondwana's northern margin and subsequent accretion of the dispersed terranes onto Asia. However, the details of the assembly history of the Bangong oceanic lithosphere including subduction polarity and timing of ocean closure remain highly disputed [Zhu et al., 2016].

Recently Yan et al. [2024] investigated the amalgamation of several large, buoyant FATs using thermo-mechanical model experiments very similar to ours with the explicit goal of exploring the factors controlling the Tethyan realm. Their models produce numerous slab break-off and subduction initiation events as well as potential polarity reversals but no divergent double subductions.

5.4. Implications for crustal growth

Our results suggest that crustal recycling into the mantle during continent-microcontinent collision is linked to the crustal rheology of the microcontinent. The decoupling of the crust and mantle of a microcontinent entering a subduction zone will have repercussions on the volume of crust preserved, generated, or recycled into the mantle during subduction. Nd and Hf isotope data indicate that approximately 64% of zircons from Proterozoic orogens are from reworked crust, reflecting crustal preservation during collisional and accretionary orogenesis [Condie, 2013]. The experiments without crust-mantle decoupling (1Ms and 2Ms) can subduct felsic continental crust to sub-lithospheric depths where they are either recycled into the mantle or they can melt and generate new crust by arc plutonism. Decoupling of the crust and mantle at a pre-existing basal detachment zone within the microcontinent will preserve continental crust by transferring it to the upper plate. Mid-crustal level decoupling leads to scraping off of
small amounts of the microcontinental crust into thrust-bound nappes in the accretionary prism while still recycling a large amount of felsic material into the mantle [Tetreault and Buitier, 2012]. Lower crustal decoupling leads to underplating of crustal material onto the overriding plate at mid-crustal depths, thus preserving the greatest amount of continental crust.

A significant portion of the modern-day granites is crustal-derived, carrying inherited zircons, and interpreted as collision-related [Cawood et al., 2013], but our results indicate that terrane subduction would also lead to recycling of felsic material into the continental crust. In fact, recent studies indicate that normal subduction would not be able to produce granitic melts from the subducting mantle [e.g., Moyen et al., 2017 and references therein], so it is likely that the large volumes of granitic arc rocks are derived from accreted or subducted FATs. Whitney et al. [2009] suggest that the large amount of granitic plutonism, gneissic doming, and migmatites found at convergent margins can be explained by melting of subducted continental crust in accreted terranes, as it cannot be generated by crustal thickening. Many of our models allow for large amounts of felsic crust to subduct to sufficient depths for melting, supporting the models of Whitney et al. [2009].

5.5. Limitations of the experimental setup

Our models are a simplified representation of nature so when interpreting their results, we need to bear in mind the limitations inherent to the approach. The experimental setup is two dimensional, meaning that it neglects the effects of the finite width of the subducting lithosphere as well as the – potentially smaller – width of the microcontinents embedded in it. The toroidal return flow around the vertical edges of a slab may have a significant influence on subduction dynamics. Close to the edge of the slab the return flow can enhance slab retreat, while the central part of a wide subduction zone cannot retreat so easily [Chen et al., 2016; Funiciello et al., 2003; Schellart et al., 2007]. A narrow microcontinent embedded in a wide subduction zone will only alter the interface resistance along a limited along-strike segment creating a complex transitional zone [e.g. Moresi et al., 2014], thus causing an along-strike variation in interface frictional resistance.

The choice of a fixed velocity boundary condition means the presence of a variable far-field force that has a strong influence on model evolution. We have chosen the constant 5 cm\text{yr}^{-1} as a middle ground, as tectonic plates on Earth move with absolute velocities between 0 to 10 cm\text{yr}^{-1} [with plates with subduction zones attached to them moving faster than those without; Kreemer et al., 2014; Kreemer et al., 2003; O'Neill et al., 2005; Schellart et al., 2007]. This
choice prevents subduction stalling (that could occur in a model with a fixed stress boundary condition) but allows us to capture better some effects of a system influenced by global plate-motions such as the observed intra-oceanic deformation that might be otherwise overlooked. On the other hand, in this setup the boundary force is not a controlled parameter and can vary freely.

Furthermore, our models only include a simplistic approximation of the eclogite phase transition and neglect latent heat and all other phase changes. The inclusion of more realistic phase change rules could alter the detachment and accretion of microcontinental material at the base of the overriding plate with the introduction of higher density mineral phases at depth and might ultimately result in a higher slab-pull force but it is unlikely that it would significantly change the main dynamics described above. Melt production and migration and water and fluid flow are also not accounted for here. These would primarily affect the rheology of the initial overriding plate (potentially lowering its integrated strength) but would not alter significantly the shear-resistance of the subduction channel. We have also chosen to use a uniform aged oceanic lithosphere to keep our models general, but that also affects the density and rheology of the subducting plate.

6. Conclusions

We explore the complex plate margin deformation resulting from the convergence of multiple microcontinents within an ocean-continent subduction system. By evaluating slab-pull force, integrated tectonic boundary forces, subduction interface resistance and the integrated strength of the involved lithospheric bodies, we draw the following conclusions:

The fate of a microcontinent during subduction depends foremost on its rheological properties. Without a weak detachment at the base of a microcontinent, it tends to subduct (experiments M1s and M2s; figure 9a and b). Crustal materials from the subducted microcontinent can be entirely absent from the surface records and might only show up in the geochemical signature of backarc basins and volcanic arcs and the topographic evolution of the overriding plate. This can impact crustal growth and recycling of continental crustal material through subduction, melting and underplating.

When a weak detachment is present, the microcontinent tends to accrete through the efficient transfer of the subduction interface from the top of the terrane to the detachment at its base (experiments M1i, M1w and 2Mw; figure 9c, e and f). The notable exception of the first microcontinent in M2i (figure 9d) that subducts demonstrates that the presence of a weak
detachment is not a sufficient prerequisite of terrane accretion in and of itself. In these models, the old sutures are efficiently preserved at or near the surface.

Depending on the strength and architecture of a microcontinent colliding with the overriding continental plate, deformation either (1) remains focused in the subduction zone, resulting in microcontinent subduction or terrane accretion, or (2) it partially localizes on a weaker zone within the lithospheric bodies surrounding the subduction zone. In the latter case, (2a) internal deformation of a microcontinent embedded within the oceanic plate far from the subduction, or (2b) the initiation of a new subduction zone can occur, significantly altering the dynamics of the subduction system in the process (figure 10).

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Open research

SULEC v.4 is available from the developers Susanne Buiter and Susan Ellis upon reasonable request. Model outputs are available on Zenodo at https://doi.org/10.5281/zenodo.11144723

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Exploring the accretion of multiple microcontinents using numerical models

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Key Points:

- Accreting or subducting microcontinents strongly influence subduction dynamics through their interaction with the subduction interface.
- The accretion of multiple microcontinents is only possible when they are rheologically weak, incorporating a lower crustal detachment.
- Multiple microcontinents in a subduction zone can lead to either partial accretion, full subduction, or divergent double subduction.
Abstract
During closure of an ocean through subduction and continental collision, bathymetric highs such as microcontinents can accrete, collide, or partially or completely subduct. Such interaction of future accreted terranes (FATs) with the overriding continent will modify the dynamics of the subduction zone, affecting its length and frictional resistance, and thus the force balance of the subduction system. Accreted microcontinents and microcontinental fragments are preserved in backarcs and collisional orogens, demonstrating that multiple terranes can accrete during a single Wilson-cycle, in what is termed accretionary orogenesis. In this study, we use thermo-mechanical numerical experiments of microcontinent-continent collision events to investigate parameters that influence whether microcontinents accrete, subduct, or collide. Our results indicate that multiple accretionary episodes are possible, but that a weak basal detachment layer within each FAT is paramount for such a scenario to occur. The introduction of a microcontinent, or FAT, in the subduction zone will affect the balance between slab-pull, far-field forces, and the subduction interface resistance. The strength (and rheological stratification) of the FATs determines the evolution of the subduction interface resistance throughout the collision event, exerting a first order control on the resulting geodynamic scenario. Collision with a strong FAT significantly increases the subduction interface resistance promoting terrane subduction and localization of deformation away from the subduction interface. In turn, collision with a weak FAT increases subduction interface resistance only mildly, allowing for multiple accretion events.

Plain language summary
This study explores what happens when small landmasses, called microcontinents, collide over millions of years with larger continents. The research, conducted through computer simulations, focuses on understanding when microcontinents either attach to the larger continent, get pushed under it, or collide with it. We were especially interested in finding out what happens when multiple microcontinents meet a continent one after the other. The findings suggest that whether a microcontinent attaches or not depends on how strong a layer it has underneath it. A weak layer allows for attachment, while a strong one leads to subduction. This interaction between microcontinents and larger continents affects the overall force balance in the tectonic system, possibly influencing deformation of Earth’s outer layer far away from the zone of collision. Understanding these processes helps us grasp the complexities of plate tectonics and how continents have formed and shifted over millions of years.
1. Introduction

The Wilson Cycle describes oceans closing and opening along their inherited sutures and has proven to be a powerful concept in analyzing the deformational histories of plate margins [Wilson, 1966]. Its initial formulation described a process of oceanic subduction, continental collision, continental extension and formation of a new ocean for the North Atlantic Region. In its focus on the tectonic history of the continental margins, the Wilson Cycle concept considered the ocean plates as fairly simple, empty basins. Oceanic plates, however, feature multiple bathymetric highs, such as microcontinents, island arcs, seamounts, or oceanic plateaus (together termed future allochthonous terranes, or FATs by Tetreault and Buiter [2012]. We here ask the question how the convergence stage of the Wilson Cycle is influenced by accretion of multiple terranes in the form of microcontinents.

Microcontinents such as Jan Mayen in the north-east Atlantic Ocean or the Seychelles in the Indian Ocean are regions of continental crust embedded within an oceanic plate and surrounded by oceanic crust. They are common oceanic features that form during rifting events on continental rifted margins and retreating subduction margins. Microcontinent accretion during subduction is a critical geological process as it is one of the main contributors to continental crustal growth [Cawood et al., 2009; Clift et al., 2009; Stern and Scholl, 2010]. Ongoing continental growth can be observed in modern collision zones such as the Pacific accretionary belt, incorporating the collision of the Philippine microplate, and the Taiwan-Luzon-Mindoro Belt [Clift et al., 2009]. Examples of active subduction zones with a recent history of accretion can be found in the North American Cordilleras [Colpron et al., 2007; Coney et al., 1980; Gabrielse and Yorath, 1991]. Microcontinental terranes can also be preserved in orogens during the final continent-continent collisional stage as inferred, for example in the European Alps [e.g. Kissling et al., 2006; Schmid et al., 1996]. Finally – to follow the path of the Wilson Cycle – we can look at the Scandinavian Caledonides for an example where collisional orogenesis that incorporated accreted terranes was followed by continental rifting [Andersen et al., 2022; Johannes Jakob et al., 2019; J. Jakob et al., 2022; Roberts, 2003] and references therein). Multiple accreted terranes are preserved in several of these systems, demonstrating that sequential terrane accretion is possible in nature. Yet, just by looking at the geological evidence, there is no way of knowing how prevalent terrane accretion is relative to microcontinent subduction, whether traces of an accretion event can be erased, under what circumstances multiple terranes accrete or how a phase of continent-continent collision can affect the preservation of earlier terrane accretion events.
When FATs are accreted they significantly alter the subduction zone dynamics before continent-continent collision could occur. This process of FAT collision involves a balance of geodynamic forces that can result in either FAT subduction, FAT accretion, or subduction stalling (which could lead to the initiation of an altogether new subduction zone in the form of subduction zone transference [Stern and Gerya, 2018]. The most important driving forces in this system are the slab-pull force arising from the negative buoyancy of the down-going slab and the far-field force which is the result of large-scale plate-motions external to the subduction zone. These forces are counteracted (among others) by friction along the subduction interface, slab bending, viscous resistance of the mantle, and the buoyancy of the down-going FAT. The net force balance controls the overall stress-field of the overriding plate as well as the state of stress and potential deformation of any FATs embedded within the oceanic lithosphere that are not yet in the subduction zone. When multiple FATs are embedded in the subducting oceanic plate, the friction along the subduction interface can vary considerably with time. This is because the accreting FATs have a first order effect on the length and the rheology of the subduction channel, thereby controlling the interface friction. The fate of the FATs (e.g. full or partial accretion, or subduction) also affects the overall buoyancy of the slab, altering the balance of forces through the slab-pull as well.

Cloos [1993] analyzed FAT subduction or accretion based on buoyancy calculations and concluded that bodies of continental and oceanic island-arc crust with more than 15 km crustal thickness can jam a subduction zone, while oceanic plateaus are inherently subductible as they are less buoyant than the oceanic lithosphere and seamounts typically only cause temporary dents. Ellis et al. [1999] were among the first to run thermo-mechanical numerical experiments exploring the dynamics of terrane accretion. Their models on lithosphere-scale (without a mantle) focused on the formation and accretion of basement fold nappes, showing that a weak detachment is required for the accretion of a terrane. Tetreault and Buiter [2012] modelled in detail how a single FAT behaves as it enters the subduction zone confirming that accretion occurs when there is a detachment layer in the FAT, that the depth of the detachment layer controls the amount of continental crust that is accreted to the overriding plate and that terrane buoyancy does not prevent subduction of FATs unless they are very long. These results were also reproduced by Vogt and Gerya [2014], Tao et al. [2020] and Gün et al. [2022]. Here we aim to build on the work of Tetreault and Buiter [2012] and – using 2D thermo-mechanical numerical experiments – explore the roles of the structure and rheology in the accretion of
multiple microcontinents to better understand how accretionary orogenesis modifies the subduction zone and affects later continental collision.

2. Modelling method

2.1. Setup and rheology

We perform 2D numerical experiments on a Cartesian region that is 660 km deep and 2300 to 2800 km wide (figure 1) using the thermo-mechanical finite-element code SULEC v.4 [Ellis et al., 2011; Naliboff and Buiter, 2015; Naliboff et al., 2017; Tetreault and Buiter, 2012; 2018]. Details of the numerical method are provided in Supplementary material 1. The numerical resolution of the model varies along the profile as well as with depth with the highest resolution of 1 km by 1 km reached around the subduction zone (figure 1).

Figure 1: Experimental setup. a) The initial geometry of a model containing two microcontinents with close-ups of a representative ocean-microcontinent margin and the incoming ocean-continent margin. The black-numbers along the edges represent distance in kms measured from the top left corner. The light-gray numbers along the edges represent the numerical resolution in kms. The red arrows along the sides represent the velocity boundary conditions. The inset in the top left corner shows characteristic initial geotherms. b) Analytically calculated strength-profiles of the oceanic lithosphere and the continental lithosphere, with dashed lines showing the strain-weakened profiles. c) Analytically calculated strength-profiles of the microcontinents, with dashed lines showing the strain-weakened profiles.
Our initial geometry consists of an oceanic plate between two continents. The oceanic plate is either “empty” or one or two microcontinents are embedded within it. In the description and analysis of the model-evolutions we will refer to these microcontinents as terranes only when they accrete to the overriding plate. The materials behave either in a viscous or a strain-weakening frictional-plastic manner that simulates brittle behavior and allows for efficient localization of deformation (Table 1). The mantle has a rheology of wet olivine combined diffusion and dislocation creep \cite{KaratoWu1993,vandenBergetal1993}. The continental plates consist of an 80 km-thick mantle lithosphere (wet olivine dislocation creep with a scaling factor $f = 5$; \cite{KaratoWu1993}), a 20 km-thick lower crust (wet anorthite dislocation creep; \cite{Rybackietal2006}) and a 20 km-thick upper crust (wet quartz; \cite{GleasonTullis1995}). The oceanic domain consists of a 73 km-thick mantle lithosphere (wet olivine dislocation creep with a scaling factor $f = 5$; \cite{KaratoWu1993}), a 5 km thick oceanic crust (gabbro dislocation creep; \cite{WilksCarter1990}) and a 2 km thick sedimentary cover (the same gabbro as the oceanic crust but with a lowered cohesion and internal angle of friction). A simple eclogite phase transition is implemented for subducting oceanic crust, sediments and the lower crust of microcontinents (Appendix A). Subduction is initiated with a “compositionally weak seed” between the overriding continental domain and oceanic domain representing a proto-subduction zone with a dip of 30°. The weak seed consists of oceanic crustal material overlain by a layer of oceanic sediments, underplating a sloped continental margin \cite{TetreaultBuiter2012}. The lower strength of these materials compared to the surrounding mantle lithosphere allows for deformation to localize in this region as the subducting plate is pushed toward the continent. The thickness of our initial weak seed (7 km) is based on the thickness of seismically imaged subduction channels (2-8 km) \cite{Abers2005}. The margin of the trailing continent is dipping steeply at an angle of 60°. The model is initialized so that the oceanic floor sits 5 km below the surface of the continents.
Table 1 Mechanical and thermal material properties used in the experiments. 1Gleason and Tullis (1995); 2Rybacki et al. (2006); 3Karato and Wu (1993); 4Wilks and Carter (1990)

The reference model R0 features an “empty” oceanic domain situated between the two continents. Models M1s, M1i and M1w contain one microcontinent embedded within the oceanic lithosphere while models M2s, M2i and M2w contain two microcontinents. All microcontinents rise 2.5 km above the oceanic sea-floor and consist of three layers: a 7 km thick gabbroic lower crust [Wilks and Carter, 1990], a 10 km thick wet anorthite mid-crust [Rybacki et al., 2006] and an 8 km thick wet quartz upper crust [Gleason and Tullis, 1995]. The characters s, i and w denote strong, intermediate and weak in both sets. In case of the intermediate (i) and weak (w) microcontinents, the gabbroic lower crust is assumed to host preexisting weak zones and in order to mimic this state, the yield strength of this layer is lowered by the assignment of reduced values for cohesion and internal angle of friction. In models M1w and M2w, the respective thicknesses of the lower and mid-crust are reversed to represent a thicker, more efficient detachment layer and an overall weaker rheology. The oceanic basin between the trench and the first microcontinent is 400 km wide, each microcontinent is 200 km wide and the two microcontinents are separated by a 300 km wide basin. The trailing continent is 250 km away from the second microcontinent (figure 1).

2.2. Kinematic and thermal boundary conditions

The kinematic boundary conditions for our experiments consist of a free top surface, free slip on the side walls and base of the model, an inflow velocity at the right continental lithosphere, and equivalently balanced outflow velocities on both side walls across the sub-lithospheric
mantle (figure 1). A weak diffusive erosion and sedimentation is applied on the top surface of
the models to approximate surface processes [Culling, 1960]. All experiments are performed
with a constant 5 cm yr\(^{-1}\) inflow velocity.

We assign a set of thermal parameters in our numerical experiments (table 1) and after solving
for an initial steady-state, we allow the thermal field to evolve. The temperature is set to a
constant 0 °C at the surface of the model and 1435 °C at the base of the model, while the side-
boundaries are thermally insulated. In order to mimic active mantle convection at high Nusselt
number, the thermal conductivity of the sublithospheric mantle is stepwise increased from 2.5
to 87.6 Wm\(^{-1}\)K\(^{-1}\) when above a temperature threshold of 1,300 °C. This approach has been
widely used in subduction as well as rift modeling studies [Butler and Beaumont, 2017;
Pyšklýwec and Beaumont, 2004; Tetreault and Buit re, 2012; Warren et al., 2008] as it prevents
secular cooling of the model domain while maintaining a constant vertical heat flux at the base
of the lithosphere and keeping the mantle close to the adiabatic gradient.

To prevent the subducting slab from interacting with the bottom boundary of the model, the
slab materials are arbitrarily transformed into sub-lithospheric mantle when reaching
approximately 625 km depth (at 200 GPa pressure and 900 °C temperature). This simplistic
approach somewhat limits the slab-pull force but the present setup still reaches values up to
2.5\(\cdot\)10\(^{13}\) Nm\(^{-1}\) (comparable to values calculated for other, similar studies; e.g. Erdős et al.
[2021]; Wolf and Huismans [2019]), while keeping the computational costs of running these
experiments manageable by allowing for a shallower computational domain.

3. Results
Here we present a brief summary of key observations from each of the 7 model experiments.
The temporal evolution of all experiments is described in more detail in Supplementary
material 2 (see also table 2) where Supplementary Animations (SA1 to SA7) and
Supplementary Viscosity Figures (S1 to S8) complementing figures 2 to 7 are also available.

<table>
<thead>
<tr>
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<th>Microcontinent behaviour</th>
<th>Oceanic lithosphere behaviour</th>
<th>Trench behaviour</th>
</tr>
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<tbody>
<tr>
<td>R0</td>
<td>n.a.</td>
<td>Subduction</td>
<td>Continuous, stable</td>
</tr>
<tr>
<td>M1s</td>
<td>Subduction, detachment and accretion under the overriding plate, likely affecting volcanic signature.</td>
<td>Continuous subduction with microcontinent.</td>
<td>Continuous, stable subduction.</td>
</tr>
<tr>
<td>M1i</td>
<td>Upper-middle crustal accretion by underplating. No large thrust sheets.</td>
<td>Continuous subduction despite terrane accretion event.</td>
<td>Trench transferred during terrane accretion.</td>
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Table 2 A summary of the unique behavior of key areas of the experimental domain for each presented model.

In the reference model, R0, the “empty” oceanic plate subducts with deformation localized along the subduction interface before continent-continent collision occurs (figure 2). We observe that the rheological stratification of the microcontinent in experiments with a single microcontinent affects whether it becomes an accreted terrane or not. During the evolution of model M1s, the strong microcontinent embedded within the oceanic lithosphere is subducted and partially accreted at the base of the overriding plate. During microcontinent subduction, minor amounts of deformation localization are observed at the incoming continental margin well before the onset of continent-continent collision (figure 3). Conversely, in model M1i, the intermediate strength microcontinent is accreted to the overriding plate as the weak basal detachment is activated resulting in a trench-jump to the trailing edge of the microcontinent. At the end of the experiment the terrane remains sandwiched between the two colliding continents (figure 4). Model M1w shows a very similar evolution to model M1i (supplementary figure S1).
Figure 2: Reference model R0 exhibiting subduction. (a–c) Material colors (see legend of Figure 2) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$). Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure 3: Model M1s (one strong microcontinent) exhibiting terrane subduction. (a–d) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$). Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure 4: Model M1i (one intermediate strength microcontinent) exhibiting terrane accretion. (a–d) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} s^{-1}). Green triangle shows location of the “trench” defined as the deepest point in the topography.

We observe unique subduction and accretion dynamics when multiple terranes are accreted, depending again on the initial rheological stratification of the microcontinents. When two strong microcontinents are embedded within the oceanic lithosphere (model M2s) both microcontinents subduct similarly to the one microcontinent in model M1s. However, during the subduction of the second microcontinent a new subduction zone is initiated along the margin of the incoming continent creating a divergent double subduction system (figure 5).
Figure 5: Model M2s (two strong microcontinents) exhibiting subduction of both terranes and subduction initiation. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} s^{-1}). Green triangle shows location of the “trench” defined as the deepest point in the topography.

In contrast, during the evolution of the model containing two intermediate strength microcontinents (model M2i), the first microcontinent is initially accreted, progressively thinned and then eroded from the subduction zone, while the second microcontinent
experiences a phase of shortening well before it reaches the subduction zone. At the end of the experiment the second microcontinent remains sandwiched between the two colliding continents (figure 6).

Figure 6: Model M2i (two intermediate strength microcontinents) exhibiting subduction of the first terrane, intra-oceanic terrane deformation and accretion of the second terrane. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours (10^{-13} to 10^{-14} s^{-1}).
Green triangle shows location of the “trench” defined as the deepest point in the topography.

Finally, in the model with two weak microcontinents (model M2w) both microcontinents are accreted to the overriding plate, with the active subduction interface hopping over to each accreting microcontinent’s basal detachment during accretion. The two accreted terranes form a crustal stack between the two colliding continents at the end of the experiment (figure 7).

Figure 7: Model M2w (two weak microcontinents) exhibiting consecutive terrane accretion events. (a–e) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C)
and areas experiencing high strain rates displayed with white contours \((10^{-13} \text{ to } 10^{-14} \text{ s}^{-1})\). Green triangle shows location of the “trench” defined as the deepest point in the topography.

4. Force-balance considerations

4.1. Interplay between slab-pull, tectonic boundary forcing and interface resistance

To understand the evolution of the experiments showing subduction or accretion of two microcontinents in a quantitative manner, we calculated the temporal evolution of the main driving and resisting forces in the subduction system [for examples of similar approaches see Erdős et al., 2021; Forsyth and Uyeda, 1975; Wolf and Huismans, 2019]. We consider the slab-pull force (resulting from the negative buoyancy of the slab), the integrated tectonic boundary forces (the depth-integrated horizontal deviatoric stresses within the lithosphere on both sides of the boundary) and the interface resistance (using the second invariant of the deviatoric stress tensor integrated along the interface as a proxy; see Supplementary material 3 for a detailed description of the procedures used to calculate these forces). We note that other resisting forces, like the bending resistance of the oceanic lithosphere and the viscous shear resistance of the mantle also play a role, but as they have been shown to be of second order magnitude, we will neglect them during the following discussion [see Erdős et al., 2021; Forsyth and Uyeda, 1975; Wolf and Huismans, 2019]. We have analytically calculated the integrated strength of the undeformed continental lithosphere, the oceanic lithosphere and the microcontinental lithospheres using a constant reference strain rate of \(1 \cdot 10^{-16} \text{ s}^{-1}\). These will help us better understand which lithospheric segments might localize deformation most efficiently.

The slab-pull forces steadily increase as the slabs get longer and reach peak values of 2 to \(2.5 \cdot 10^{13} \text{ N m}^{-1}\) (figure 8a). The slab-pull force in models with microcontinents starts to deviate from the reference model \(\text{R}0\) at around 12 Myr, which corresponds to the stage when the first microcontinents reach a depth of 120 km or deeper (figure 8a). This difference is the result of compositionally lighter crustal material of the microcontinent replacing cold, negatively-buoyant oceanic mantle-lithosphere. This positively-buoyant crustal material cannot subduct to great depths and therefore detaches and accretes to the bottom of the overriding plate, leading to a reduction in the overall pull-force.
Figure 8: Key driving and resisting forces. a) The temporal evolution of the slab-pull force calculated for models R0, M2s, M2i and M2w, showing minimal deviations between them. b) The temporal evolution of the slab-pull force, the integrated tectonic boundary forces and the interface resistance for model M2i, with key events marked. c) The temporal evolution of the slab-pull force, the integrated tectonic boundary force and the interface resistance for models M2s, M2i and M2w. The Integrated tectonic boundary force of R0 up to the onset of continent-continent collision is plotted as a grey line.

The most significant deviation from the continuous increase of slab-pull forces with time is the late stage drop in slab-pull force for model M2s, where the initiation of a new subduction zone reduces the force exerted by the original slab. Based on the above considerations, we argue that variations between the slab-pull forces for the analyzed three models are negligible, hence the differences in the evolution of tectonic boundary forces between the models are largely
independent of slab-pull and reflect a response to the changes in interface resistance and changes in the integrated strength of different lithospheric bodies.

When looking at common trends within the experiments (using model M2i as an intermediate example) we observe that the slab-pull force briefly becomes dominant over the integrated tectonic boundary forces around the time the first microcontinent arrives at the trench (figure 8b). Subsequently, as the microcontinent enters the subduction channel the interface resistance and the tectonic boundary forces both increase (figure 8b). In all three models (M2w, M2i, M2s), the peak values of tectonic boundary forces are recorded when a microcontinent is in the subduction channel (see force-material co-evolution animations SA8 to SA10 in the supplements). Subsequently, the tectonic boundary forces gradually decrease, as the deformation shifts back from either the top of the subducting microcontinent (e.g. in model M2s), the bottom of the accreting terrane (e.g. in model M2w) or the broad zone of internal deformation that covers the microcontinent in the channel (e.g. in model M2i). We deduce that the increase in interface resistance is the result of the microcontinental crustal material replacing the oceanic crust and sediment that was lubricating the subduction channel. With increased interface resistance, more energy is dissipated along the interface, hence more force is required on the boundaries to keep the prescribed steady pace of convergence. From this observation, we can infer that a significant slab-pull and/or far-field compressional forcing is necessary in nature for subduction to continue during terrane accretion.

Alternatively, when sufficient boundary force is not available, we predict that subduction would slow down, or even stall entirely. This phenomenon where an increase in mechanical coupling between the subducting and overriding plates leads to subduction stalling has been observed both in analogue [Duarte et al., 2015] and numerical [Afonso and Zlotnik, 2011; De Franco et al., 2008] models that do not impose fixed boundary forces. Supporting this interpretation is the previous work by De Franco et al. [2008], who have modelled the collision between an overriding continent and an incoming continental margin and found that the nature of the subduction interface has a first order role in the dynamics of the collision. Their results are in agreement with our assessment that successive accretion of multiple terranes is dynamically complex, because the rheology of the subduction channel changes with every accretion event.

The model with two strong microcontinents (M2s) exhibits the most pronounced changes in the competing forces, with the highest values for both interface resistance and integrated
tectonic boundary force within the set (figure 8c). In turn, as the microcontinents become weaker (i.e. experiments M2i and M2w), the increase in interface resistance diminishes, and less boundary forcing is necessary to achieve the prescribed convergence rate.

4.2. Implications of the interplay between the driving and resisting forces

The high values of integrated tectonic boundary forces in models M2s and M2i imply that the different lithospheric segments (i.e. continental, oceanic and microcontinental domains) transmit high compressional stresses towards the actively deforming areas. In experiment M2s, the interface resistance peaks during the subduction of the second microcontinent. During this phase, the interface resistance (approximately $8.5 \cdot 10^{12} \text{Nm}^{-1}$) is higher than the integrated strength of the continental lithosphere ($7.6 \cdot 10^{12} \text{Nm}^{-1}$). As a result, it is energetically more efficient to localize deformation within the incoming continental lithosphere than it is to keep the subduction going at the same pace. The initial diffuse, low strain-rate deformation that was present during the subduction of the first microcontinent facilitates strain-weakening during the subduction of the second microcontinent. Strain-weakening, in turn, makes the localization of deformation even more efficient, creating a positive feedback loop and resulting in subduction initiation. The deformation is effectively partitioned between the two deforming zones as the original subduction zone remains active throughout the birth of the new zone, albeit at a reduced intensity.

In experiment M2i, the interface resistance peaks during the partial accretion of the first microcontinent. During this phase, the accreting terrane is extensively deformed, resulting in an interface resistance of approximately $5 \cdot 10^{12} \text{Nm}^{-1}$, which is comparable to the integrated strength of the undeformed second microcontinent ($6.3 \cdot 10^{12} \text{Nm}^{-1}$). The integrated strength of the continental lithosphere remains comparatively high at $7.6 \cdot 10^{12} \text{Nm}^{-1}$. As a result, the incoming second microcontinent undergoes increasingly localized shortening that starts while it is still over 200 km away from the subduction interface. It is also notable that whereas in model M1i the first and only microcontinent is accreted, in experiment M2i the first microcontinent largely subducts. The difference between the evolution of the two models starts with the onset of deformation within the second microcontinent of model M2i, when the first microcontinent reaches the subduction zone. Since strain is being partitioned between several distinct zones of deformation, the strain-rate remains lower in the first microcontinent as it lingers in the subduction channel. The lower strain-rate results in more diffuse internal deformation that prevents the efficient localization of displacement at the base of the microcontinent and ultimately leads to its subduction rather than its accretion.
In experiment M2w, the interface resistance remains much lower than the integrated strength of the undeformed lithospheric domains: the peak interface resistance is approximately $3 \cdot 10^{12}$ Nm$^{-1}$ and $5 \cdot 10^{12}$ Nm$^{-1}$ during the first and second accretion events, while the integrated strength of the microcontinents and the continents are $5.8 \cdot 10^{12}$ Nm$^{-1}$ and $7.6 \cdot 10^{12}$ Nm$^{-1}$ respectively. As a result, deformation remains localized solely in the subduction zone throughout the experiment.

**Figure 9:** Close-up from the models containing microcontinents. The first column shows convergence of one microcontinent (models M1s, M1i and M1w on subfigures a, c and e respectively). Note how the microcontinent subducts in model M1s (strong) and accretes in M1i (intermediate) and M1w (weak). The second column shows convergence of two microcontinents (models M2s, M2i and M2w on subfigures b, d and f respectively). Note the different behaviors: microcontinent subduction and subduction initiation in M2s (strong), accretion of an already deformed terrane in M2i (intermediate) and consecutive accretion of two previously undeformed terranes in M2w (weak). The slab remains continuous without a break-off event in all six models.

The experiments presented here demonstrate that a variety of tectonic scenarios can ensue during a sequence of accretion events that depend on the inherent structure (e.g. presence of a weak detachment layer) and strength of the lithospheric domains at play (figure 9). The fixed subducting-plate velocity – relative to a stable overriding plate – prescribed as a boundary condition represents areas where global plate velocities have a first order effect on the plate
movements within the system. In our particular setup, the natural sinking velocity of the subducting lithosphere generated by slab-pull forces cannot accommodate the relatively high subducting-plate velocity and additional boundary forces are required to maintain steady convergence even in the absence of continent-microcontinent collision (figure 8c). This compressional stress is transmitted efficiently through the strong oceanic lithosphere and localizes in the subduction channel. When a microcontinent enters the subduction zone the increase in interface resistance triggers an increase in the boundary forces. When another microcontinent is embedded within the oceanic plate the stress transmitted through the plate can be high enough to cause the internal deformation of the microcontinent (experiment M2i; figure 9d) and in the most extreme scenario presented here, to initiate a new subduction zone (experiment M2s; figure 9b).

5. Discussion

We observe the following large-scale behaviors of accretionary orogenesis that depend on the rheological stratification of the microcontinents: First, when the microcontinents are rheologically sufficiently weak, the subduction interface resistance is only mildly increased when a terrane enters the channel. As a result, multiple terrane accretion events can occur without significant shortening elsewhere in the system (behavior 1; experiment 2Mw). Second, when the microcontinents are stronger, the subduction interface resistance increases significantly during terrane accretion making the subduction zone less efficient at localizing shortening. Deformation will be distributed in the system as shortening localizes on either the second incoming microcontinent, embedded in the oceanic lithosphere (behavior 2a; experiment 2Mi) or along the incoming continental margin (behavior 2b; experiment 2Ms), depending on which one of these two zones has the lower integrated lithospheric strength. In the following subsections we will discuss these large-scale behaviors separately.

5.1. Terrane accretion events

The first numerical modelling study that examined terrane accretion was by S Ellis et al. [1999]. One of their findings was that a weak detachment is necessary for the accretion of a terrane. These early results on the lithosphere scale were reproduced with models that included the upper mantle by Tetreault and Buiter [2012], Vogt and Gerya [2014], Tao et al. [2020] and Gün et al. [2022]. In their study, Tetreault and Buiter [2012] examined the convergence of different types of FATs embedded in an oceanic plate that is subducting beneath a continental plate and found that (1) accretion is favored when there is a weak layer or detachment fault
within the FAT and (2) different accretionary styles are controlled by the depth of detachment in the FAT crust. Our experiments also reproduce this behavior: in experiment M1s, there is no weak detachment layer within the microcontinent and it subducts before the closure of the oceanic domain (figure 9a). In contrast, in experiment M1i, the weak basal layer of the microcontinent acts as a detachment zone, forming a new subduction interface that allows for the efficient transfer of the trench to the oceanward edge of the accreting terrane (figure 9c). Most of the lower crust is transformed to eclogite upon reaching the requisite pressure-temperature conditions and subsequently subducts, but the middle and upper crust of the microcontinent is accreted. However, in experiment M2i, which has an identical microcontinent architecture to that of M1i, the first microcontinent subducts during the experiment, suggesting that a weak basal detachment does not guarantee accretion. In this particular experiment following an initial phase of terrane accretion, the accreted material is remobilized through a network of faults and gradually subducts. This behavior is coeval with the internal deformation of the second microcontinent, suggesting that deformation elsewhere within the model-domain has a first order effect on subduction dynamics through the partitioning of strain and strain-rate within the system.

Our models show that multiple terrane accretion events can occur in the same subduction system, and that the microcontinental crustal material is preserved as accreted terranes at the surface or at shallow depth below the tip of the original overriding plate (e.g., experiment M2w; figure 9f). This is possible when there is a weak layer at the base of the FATs and any already accreted terranes do not increase the interface resistance beyond the point where it would become more efficient to localize deformation elsewhere in the system than it is to keep deformation focused along the already established subduction zone. Our models also demonstrate that terrane accretion does not necessarily result in a transfer of the subduction zone through slab break-off and the initiation of a new subduction zone; a process that has been invoked in tectonic evolution models and described in Stern and Gerya [2018] as well as several recent modelling studies such as Zhong and Li [2020] and Yan et al. [2024]. Instead of break-off, the slab can remain continuous as the accreted terrane is detached from its mantle-lithospheric root, transferring the trench efficiently to the oceanward edge of the terrane. This process has been demonstrated in numerical studies of single terrane accretion
that this event may occur multiple times within a single subduction system.

5.2. Intra-oceanic deformation

Gün et al. [2022] and Gün et al. [2024] have shown that a fixed, low subducting-plate velocity can result in microcontinent extension. Their result is substantiated by our force-balance considerations. Specifically, a fixed, low subducting-plate velocity cannot keep pace with the natural sinking velocity of the subducting slab, creating a tensional stress-field in the subducting plate. The higher the difference between the natural sinking velocities of the slab and the velocity of the incoming plate at the Earth’s surface, the stronger the resulting tensional stress-field becomes. Such a difference in velocities can result from global plate-tectonic dynamics. Extensional deformation within the microcontinent embedded in the oceanic domain only occurs when the stresses are high enough to overcome its integrated strength. Such extensional deformation of a FAT has been reported from the Ontonga Java Plateau [see and Gün et al., 2024 references therein) or the Shatsky Rise [Zhang et al., 2015] among others.

![Figure 10: A pulley-system, explaining the role of the major players determining where deformation localizes. The weight marked ‘a’ symbolizes the slab-pull force, the sliding piston marked ‘b’ symbolizes the interface resistance and the material in the subduction channel that can deform internally (i.e. an accreting microcontinent), the two pistons marked ‘c’ and ‘d’ symbolize a microcontinent and the incoming continental margin that can both localize deformation if the far-field force (marked ‘e’ on the figure) is larger than their respective integrated lithospheric strength. A snapshot of the material domain](image)
of experiment 1Mi with the corresponding areas marked a-e is included below the pulley-system.

We argue here, that the system is more complex than the simple pulley mechanism invoked by Gün et al. [2024]. Our analysis shows that the subducting-plate velocity (relative to a fixed overriding plate), the natural sinking velocity and the integrated strength of the microcontinent are not the only controlling factors that determine whether intra-oceanic deformation can occur. The resistance along the subduction interface also plays a crucial role (figure 10). A relatively high subduction-interface resistance can dissipate a large portion of the slab-pull force, resulting in a lower effective sinking velocity for the slab, reducing the stress transmitted through the oceanic lithosphere and diminishing the potential for intra-oceanic extensional deformation.

Furthermore, if a lower effective sinking velocity is combined with a high subducting-plate velocity, the overall stress-field of the subducting plate becomes compressional. If the compressional stress transmitted through the plate overcomes its integrated strength – for example in an embedded microcontinent – shortening can occur. Such intra-oceanic shortening has been reported from the West Somali Basin [Sauter et al., 2018], from the Central Indian Basin [Bull, 1990; Delescluse et al., 2008] or from the Wharton Basin [Carton et al., 2014]. These examples generally involve the reactivation of pre-existing oceanic fracture zones that lower the integrated strength of the lithosphere, allowing for lower stresses to induce localized deformation.

5.3. Divergent double subduction

An unexpected result from this study was the divergent double subduction following microcontinent subduction. The conceptual model of divergent double subduction has been used to explain the Lachlan fold belt of southeastern Australia [Soesoo et al., 1997], the closure of the Bangong ocean and collision of the Lhasa and Qiangtang terranes in China [Zhu et al., 2016] and has been seismically imaged in the Molucca Sea in eastern Indonesia [Fan and Zhao, 2018; Puspito et al., 1993; Silver and Moore, 1978; Zhang et al., 2017]. Most of the examples of inferred divergent double subduction are Mesozoic or older in age but Soesoo et al. [1997] established a set of tectonic, metamorphic, and petrological signatures that allow for a distinction between single versus double divergent subduction zones during continent-continent collision.
The Molucca Sea is the only active example of this tectonic regime, but the collision happens between two arcs rather than two continents and there is no indication of terrane subduction along either subduction zones. Nevertheless, it is worth pointing out that the late stage evolution of our experiment 2Ms reproduces some of the observations remarkably well. In particular, two slabs of significantly different lengths are imaged by several tomographic studies of the region ([Fan and Zhao, 2018; Hall and Spakman, 2002; Puspito et al., 1993; Widiyantoro and Hilst, 1997]), and their trenches are buried by an accretionary complex at shallow depth [Zhang et al. 2017 and references therein].

In case of the closure of the Bangong ocean, the proposed divergent double subduction happens between two microcontinents, the Lhasa terrane towards the south and the Qiangtang terrane towards the north [e.g., Yin and Harrison, 2000]. The suture zone and its relationships with the bounding terranes provides a window into the rift, drift and accretion-related tectonism, magmatism, sedimentation, and metamorphism associated with the fragmentation of Gondwana's northern margin and subsequent accretion of the dispersed terranes onto Asia. However, the details of the assembly history of the Bangong oceanic lithosphere including subduction polarity and timing of ocean closure remain highly disputed [Zhu et al., 2016]. Recently Yan et al. [2024] investigated the amalgamation of several large, buoyant FATs using thermo-mechanical model experiments very similar to ours with the explicit goal of exploring the factors controlling the Tethyan realm. Their models produce numerous slab break-off and subduction initiation events as well as potential polarity reversals but no divergent double subductions.

5.4. **Implications for crustal growth**

Our results suggest that crustal recycling into the mantle during continent-microcontinent collision is linked to the crustal rheology of the microcontinent. The decoupling of the crust and mantle of a microcontinent entering a subduction zone will have repercussions on the volume of crust preserved, generated, or recycled into the mantle during subduction. Nd and Hf isotope data indicate that approximately 64% of zircons from Proterozoic orogens are from reworked crust, reflecting crustal preservation during collisional and accretionary orogenesis [Condie, 2013]. The experiments without crust-mantle decoupling (1Ms and 2Ms) can subduct felsic continental crust to sub-lithospheric depths where they are either recycled into the mantle or they can melt and generate new crust by arc plutonism. Decoupling of the crust and mantle at a pre-existing basal detachment zone within the microcontinent will preserve continental crust by transferring it to the upper plate. Mid-crustal level decoupling leads to scraping off of
small amounts of the microcontinental crust into thrust-bound nappes in the accretionary prism while still recycling a large amount of felsic material into the mantle [Tetreault and Buiter, 2012]. Lower crustal decoupling leads to underplating of crustal material onto the overriding plate at mid-crustal depths, thus preserving the greatest amount of continental crust.

A significant portion of the modern-day granites is crustal-derived, carrying inherited zircons, and interpreted as collision-related [Cawood et al., 2013], but our results indicate that terrane subduction would also lead to recycling of felsic material into the continental crust. In fact, recent studies indicate that normal subduction would not be able to produce granitic melts from the subducting mantle [e.g., Moyen et al., 2017 and references therein], so it is likely that the large volumes of granitic arc rocks are derived from accreted or subducted FATs. Whitney et al. [2009] suggest that the large amount of granitic plutonism, gneissic doming, and migmatites found at convergent margins can be explained by melting of subducted continental crust in accreted terranes, as it cannot be generated by crustal thickening. Many of our models allow for large amounts of felsic crust to subduct to sufficient depths for melting, supporting the models of Whitney et al. [2009].

5.5. **Limitations of the experimental setup**

Our models are a simplified representation of nature so when interpreting their results, we need to bear in mind the limitations inherent to the approach. The experimental setup is two-dimensional, meaning that it neglects the effects of the finite width of the subducting lithosphere as well as the — potentially smaller — width of the microcontinents embedded in it. The toroidal return flow around the vertical edges of a slab may have a significant influence on subduction dynamics. Close to the edge of the slab the return flow can enhance slab retreat, while the central part of a wide subduction zone cannot retreat so easily [Chen et al., 2016; Funiciello et al., 2003; Schellart et al., 2007]. A narrow microcontinent embedded in a wide subduction zone will only alter the interface resistance along a limited along-strike segment creating a complex transitional zone [e.g. Moresi et al., 2014], thus causing an along-strike variation in interface frictional resistance.

The choice of a fixed velocity boundary condition means the presence of a variable far-field force that has a strong influence on model evolution. We have chosen the constant 5 cm yr$^{-1}$ as a middle ground, as tectonic plates on Earth move with absolute velocities between 0 to 10 cm yr$^{-1}$ [with plates with subduction zones attached to them moving faster than those without; Kreemer et al., 2014; Kreemer et al., 2003; O'Neill et al., 2005; Schellart et al., 2007]. This
choice prevents subduction stalling (that could occur in a model with a fixed stress boundary condition) but allows us to capture better some effects of a system influenced by global plate-motions such as the observed intra-oceanic deformation that might be otherwise overlooked. On the other hand, in this setup the boundary force is not a controlled parameter and can vary freely.

Furthermore, our models only include a simplistic approximation of the eclogite phase transition and neglect latent heat and all other phase changes. The inclusion of more realistic phase change rules could alter the detachment and accretion of microcontinental material at the base of the overriding plate with the introduction of higher density mineral phases at depth and might ultimately result in a higher slab-pull force but it is unlikely that it would significantly change the main dynamics described above. Melt production and migration and water and fluid flow are also not accounted for here. These would primarily affect the rheology of the initial overriding plate (potentially lowering its integrated strength) but would not alter significantly the shear-resistance of the subduction channel. We have also chosen to use a uniform aged oceanic lithosphere to keep our models general, but that also affects the density and rheology of the subducting plate.

6. Conclusions

We explore the complex plate margin deformation resulting from the convergence of multiple microcontinents within an ocean-continent subduction system. By evaluating slab-pull force, integrated tectonic boundary forces, subduction interface resistance and the integrated strength of the involved lithospheric bodies, we draw the following conclusions:

The fate of a microcontinent during subduction depends foremost on its rheological properties. Without a weak detachment at the base of a microcontinent, it tends to subduct (experiments M1s and M2s; figure 9a and b). Crustal materials from the subducted microcontinent can be entirely absent from the surface records and might only show up in the geochemical signature of backarc basins and volcanic arcs and the topographic evolution of the overriding plate. This can impact crustal growth and recycling of continental crustal material through subduction, melting and underplating.

When a weak detachment is present, the microcontinent tends to accrete through the efficient transfer of the subduction interface from the top of the terrane to the detachment at its base (experiments M1i, M1w and 2Mw; figure 9c, e and f). The notable exception of the first microcontinent in M2i (figure 9d) that subducts demonstrates that the presence of a weak
detachment is not a sufficient prerequisite of terrane accretion in and of itself. In these models, the old sutures are efficiently preserved at or near the surface.

Depending on the strength and architecture of a microcontinent colliding with the overriding continental plate, deformation either (1) remains focused in the subduction zone, resulting in microcontinent subduction or terrane accretion, or (2) it partially localizes on a weaker zone within the lithospheric bodies surrounding the subduction zone. In the latter case, (2a) internal deformation of a microcontinent embedded within the oceanic plate far from the subduction, or (2b) the initiation of a new subduction zone can occur, significantly altering the dynamics of the subduction system in the process (figure 10).

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Open research

SULEC v.4 is available from the developers Susanne Buiter and Susan Ellis upon reasonable request. Model outputs are available on Zenodo at https://doi.org/10.5281/zenodo.11144723

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Supplementary material

1. Numerical method

SULEC is a two-dimensional finite-element arbitrary Lagrangian-Eulerian numerical code that solves the plane-strain incompressible momentum equation for slow viscous-plastic creeping flows:

\[ \nabla \cdot \sigma' - \nabla P + \rho g = 0 \quad (1) \]
\[ \nabla \cdot \mathbf{u} = 0 \quad (2) \]

where \( \sigma' \) is the deviatoric stress tensor, \( P \) is dynamic pressure, \( \rho \) is density, \( g \) is gravitational acceleration (9.81 m s\(^{-2}\) in the vertical direction) and \( \mathbf{u} \) is the velocity vector. Pressure is calculated using the iterative Uzawa formulation [Pelletier et al., 1989]:

\[ P^t = P^{t-1} - f_c \nabla \cdot \mathbf{u}^t \quad (3) \]

where \( t \) signals pressure iteration number (with a cap of 75 iterations) and \( f_c \) is the compressibility factor, which is 4 orders of magnitude larger than the maximum allowed viscosity in our models. The code relies on the Boussinesq approximation of small changes when describing the temperature dependence of density:

\[ \rho = \rho_0 (1 - \alpha (T - T_0)) \quad (4) \]

where \( \alpha \) is thermal expansion and \( \rho_0 \) is reference density at temperature \( T = T_0 \).

Materials are either viscous or plastic. In the models presented here, elastic deformation is not considered. For viscous deformation, a nonlinear, thermally activated power law creep formulation is used which relates pressure, temperature and strain rate to the effective stress:

\[ \sigma'_{\text{eff}} = f \cdot A^{-1/n} \cdot \dot{\varepsilon}'_{\text{eff}}^{n} \cdot w^{-r/n} \cdot d^{m/n} \cdot e^{Q/PV \cdot R/T} \quad (5) \]

where \( f \) is an optional scaling factor, \( A \) is the power law pre-exponent, \( n \) is the power law index, \( \dot{\varepsilon}'_{\text{eff}} \) is the effective strain rate, \( w \) is water content, \( r \) is the water-content exponent, \( d \) is grain size, \( m \) is the grain-size exponent, \( Q \) is the activation energy, \( V \) is the activation volume, and \( R \) is the universal gas constant (8.314 J mol\(^{-1}\) K\(^{-1}\)). Effective stress and strain rate are defined as:

\[ \sigma'_{\text{eff}} = \left( \frac{1}{2} \cdot \sigma'_{ij} \cdot \sigma'_{ij} \right)^{1/2} \quad (6) \]
$$\dot{\varepsilon}'_{\text{eff}} = \left(\frac{1}{2} \cdot \dot{\varepsilon}'_{ij} \cdot \dot{\varepsilon}'_{ij}\right)^{\frac{1}{2}}$$  \hspace{1cm} (7)

with summation over repeated indices included.

Plastic failure is approximated with the Drucker-Prager yield-criterion:

$$\sigma'_{\text{eff}} = P \cdot \sin \varphi + C \cdot \cos \varphi$$  \hspace{1cm} (8)

where $\varphi$ is the angle of internal friction and $C$ is cohesion. This approximation is a smoothed version of the Mohr-Coulomb yield-criterion in plane-strain. Strain weakening [e.g., Pysklywec et al., 2002] is accounted for by linearly reducing $\varphi$ through a predefined plastic effective strain ($\varepsilon$) interval of $0.5 < \varepsilon < 1.5$.

Since equations (4) and (5) are temperature-dependent, we also solve the heat-transport equation:

$$c_p \rho \left(\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T\right) = \nabla \cdot k \nabla T + H + H_{sh}$$  \hspace{1cm} (9)

where $c_p$ is specific heat, $k$ is thermal conductivity and $H$ is heat production and $H_{sh}$ is shear heating where shear heating is calculated from the effective stress and effective strain rate using the following equation:

$$H_{sh} = f_{sh} \cdot 2 \cdot \sigma'_{\text{eff}} \cdot \dot{\varepsilon}'_{\text{eff}}$$  \hspace{1cm} (10)

where $f_{sh}$ is the shear heating efficiency factor, set to be 1 in these experiments.

A simplified eclogite phase-transition is implemented: when oceanic crust and overlying sedimentary material reach the appropriate temperature-pressure conditions they are instantly transformed into eclogite. We use the eclogite stability field of Hacker [1996]. In this simplified approach, only the density of the transforming material is adjusted (increased) during this transition, whereas the flow-law remains the same and latent heat is not accounted for. The same eclogite phase-transition is also applied to the lower crust of subducting microcontinents. We base this choice on the common argument that the lower crust of oceanic plateaus, submarine ridges, and island arcs is an ultramafic cumulate layer [Behn and Kelemen, 2006; Mann and Taira, 2004; Schubert and Sandwell, 1989; Tetreault and Buiter, 2014].

To solve the above system of equations (both for material flow and thermal structure) SULEC uses the direct solver PARDISO v8.0 developed for sparse matrices [Alappat et al., 2020; Bollhöfer et al., 2019; Bollhöfer et al., 2020].
SULEC v.4 employs an Arbitrary Lagrangian-Eulerian discretization method that allows for small vertical deformation of element shapes with remeshing per time step in order to achieve a true free surface [Fullsack, 1995]. The quadrilateral elements are constant in pressure and linear in velocity. We use a discretization with 1055 elements in horizontal direction and 280 in the vertical. The mesh is refined to 1 by 1 km in the domain of 800 by 900 km in the horizontal direction from the surface to 50 km depth in the vertical direction. Material properties are tracked by tracers. Initially, we use 8 tracers per element and population control keeps the elements between 4 to 16 tracers per element. Independently of the surface-process algorithm used, SULEC applies a stabilization term at the free surface that corrects numerical overshoots [Kaus et al., 2010; Quinquis et al., 2011].

2. Detailed description of the temporal evolution of the experiments
   a. Collision in the absence of microcontinents
   In our reference model experiment, R0, an “empty” oceanic basin located between two continents is subducting under the overriding continent driven by a constant 5 cm yr\(^{-1}\) boundary condition set at the far end of the incoming continent (figure 2, supplementary animation A1; for viscosity structure see supplementary figure S1). During subduction a small sedimentary accretionary wedge is formed in the subduction zone but neither the overriding continent nor the incoming continent is deformed. When the incoming continent reaches the overriding continent at approximately 11 Myr continent-continent collision is initiated (figure 2a); the incoming continent underthrusts the overriding continent continuously following the oceanic lithosphere. Distributed compressional deformation of the overriding continent begins around 14 Myr and continues thereafter. Nevertheless, the zone of most intense, localized deformation remains the continuous subduction interface until around 17.6 Myr, when the tip of the underthrusting continental crust reaches a depth of over 120 km (figure 2b). At this point a new thrust-sheet is formed in the footwall of the previous thrust, rooted in the ductile deformation zone at the bottom of the upper-crust. At 19.6 Myr model time this process is repeated, creating an outward propagating set of thick-skinned thrusts in the incoming continent. The experiment is concluded at 20 Myr model time (figure 2c).

   b. One strong microcontinent – terrane subduction
   In model experiment M1s, we have embedded a strong, 200 km wide and 25 km thick microcontinent in the oceanic plate, with no weak internal decoupling zone (figure 3, supplementary animation A2; for viscosity structure see supplementary figure S2). After
subduction initiation the microcontinent reaches the trench at 7.4 Myr and subducts cleanly by 12.4 Myr (figure 3a). During this process the overriding plate experiences a phase of significant uplift with the peak topography going from 500 m to 6000 m above sea level creating an approximately 150 km wide mountain belt. The uplift occurs largely in the absence of brittle faulting with back-thrusting only locally present at the latter stages of the subduction (see supplementary animation A2). The subduction interface remains continuous at the top of the microcontinent and no microcontinental crustal material remains on the surface. At 13.2 Myr the tip of the microcontinent is dragged down to a depth of 150 km starting a phase of significant internal deformation. During this phase the weak, buoyant upper-crust of the microcontinent is gradually detached, pooling at the LAB of the overriding plate while the rest of the microcontinent continues to subduct. At around 14 Myr the incoming continental margin starts to weakly localize deformation despite it being still over 150 km away from the trench (figure 3b). This deformation continues but remains low-key until continent-continent collision initiates at 17.4 Myr. With continued convergence the incoming continental margin underthrusts the overriding plate with oceanic and continental sediments filling in the suture zone while the remnants of the oceanic crust get buried (figure 3c). At around 19.6 Myr the buoyant middle-crust of the subducted microcontinent is rapidly detached from the sub-vertical slab and accretes at the base of the overriding plate. The experiment is concluded at 22 Myr model time (figure 3d).

c. **One intermediate microcontinent – terrane accretion**

Model experiment M1i is identical in setup with M1s except for the inclusion of a 7.5 km thick weak decoupling layer at its base (figure 4, supplementary animation A3; for viscosity structure see supplementary figure S3). This layer is designed to represent a pre-deformed mafic underplating. The model evolution of the first 7.4 Myr is identical to that of the previous model. At this point, as the microcontinent reaches the trench deformation immediately starts to localize on the basal detachment zone as well as on the top of the microcontinent. By 9.8 Myr the basal shear-zone spans the entire microcontinent reaching the surface at its oceanward margin (figure 4a). Until about 11.2 Myr the shear-zone located at the top of the microcontinent is the more prominent one accommodating the overwhelming majority of the displacement as the microcontinent under-thrusts the overriding plate. Between 11.2 Myr and 11.8 Myr the original subduction zone is gradually abandoned in favor of the detachment zone at the bottom of the microcontinent, resulting in the transfer of the subduction channel and the formation of a new trench at the oceanward margin of the newly accreted terrane (figure 4b).
subduction channel is located within the lower crust of the terrane and is largely continuous so that the middle and upper crust of the terrane remains mostly intact at a shallow depth. During the accretion of the terrane the very tip of the overriding plate experiences a gradual uplift, forming a 50 km wide 5 km high mountain belt, without any localized deformation zones within it. At 15.8 Myr the incoming continental margin reaches the new trench, starting the phase of continent-continent collision. During this process the original subduction zone located at the top of the accreted terrane is reactivated and deforms simultaneously with the new subduction zone (see the strain rate contours on figure 4c). As the incoming continent is gradually under-thrusts the accreted terrane starts a slow process of internal fragmentation. By the end of the experiment at 22 Myr most of the terrane is buried under the tip of the overriding continental crust but remains largely intact and preserved at a shallow depth (figure 4d).

d. One weak microcontinent – terrane accretion
Model experiment M1w has a 10 km thick lower crustal layer in the microcontinent as opposed to the 7.5 km thick lower crustal thicknesses in model M1s and M1i, making the potential zone of decoupling thicker and the microcontinental rheology somewhat weaker (see the strength envelopes on figure 1). The evolution of this experiment is very similar to that of model M1i as the microcontinent accretes cleanly to the overriding plate and remains preserved at shallow depth during almost 5 Myr of continent-continent collision (compare supplementary figure S8 to figure 4).

e. Two strong microcontinents – subduction initiation
Model experiment M2s is built in a wider computational domain than the previous setups and has two strong microcontinents (identical in their characteristics with the one in M1s) embedded 250 km apart in the oceanic domain (figure 5, supplementary animation A4; for viscosity structure see supplementary figure S4). From subduction initiation to the complete subduction of the 1st microcontinent the model behavior is identical to that of M1s (figure 5a). Similarly, to M1s the upper crust of the subducted microcontinent pools at the LAB of the overriding plate. The topographic evolution is also analogue to what has been observed in experiment M1s with significant topographic uplift throughout but localized back-thrusting only in the tip of the overriding plate and only at the latter part of the 1st subduction event. As in M1s internal deformation of the incoming continental margin occurs but it remains very diffuse, driven by low strain rates (below $10^{-14}$) and cumulatively below the threshold that would trigger strain-weakening. With continued convergence the buoyant middle-crust of the
subducted microcontinent gets detached and accreted at the base of the overriding plate, but this process starts at 19 Myr, slightly earlier than in M1s. At the same time, the 2nd microcontinent reaches the trench and starts to underthrust in the subduction zone (figure 5b). As a result, by 23 Myr the 2nd microcontinent is also subducted, with a sliver of oceanic crust – originally located between the two microcontinents – preserved above it (figure 5c). The subduction of the 2nd microcontinent triggers a second phase of orogeny spanning the tip of the overriding plate without triggering significant brittle faulting. The presence of the detached and accreted crustal material of the 1st microcontinent at the base of the overriding plate far behind the trench means that the positive topographic anomaly spans several hundred kms. At 23.4 Myr the incoming continental margin is still over 200 km away from the subduction zone but the strain-rates within are now in the order of $10^{-14}$ and above and deformation starts to localize. Subsequently the subduction is still active along the original interface but at a reduced rate while a new subduction zone is initiated by 26.6 Myr at the incoming continental margin (figure 5d). In this process a divergent double subduction is created on the two opposite sides of the oceanic domain and this dynamic remains active until the end of the experiment at 30 Myr (figure 5e).

f. Two intermediate microcontinents – intra-oceanic deformation

Model experiment M2i has an identical setup to M2s, but its embedded microcontinents have the characteristics of the one in experiment M1i, with a weak detachment layer at its base (figure 6, supplementary animation A5; for viscosity structure see supplementary figure S5). The model evolution is largely identical to that of M1i until 11.8 Myr. At this point the microcontinent is mostly located in the subduction channel with two active zones of deformation; one at the top of the terrane and a less active one at its base. Furthermore, compressional deformation starts to localize at the margins of the 2nd microcontinent, even though it is still approximately 300 km away from the subduction zone (figure 6a). As the convergence continues the 1st microcontinent – that is entirely in the subduction channel at this point – breaks up into several discrete blocks while the 2nd microcontinent keeps slowly accumulating internal thickening through deformation along its margins and at its base. By 17.5 Myr as the 2nd microcontinent arrives to the trench, a large portion of the 1st microcontinent is subducted and the rest is smeared in along the subduction interface as a several km wide deformation zone (figure 6b). With continued convergence the material within the subduction channel gets thinned out further and thrust deeper while simultaneously the buoyant part of the 1st microcontinent is progressively detached from the slab and accretes at the base of the
overriding plate, forming a highly deformed triangular wedge over 200 km behind the trench. In the meantime, the internal thickening of the 2nd microcontinent stops as it enters the subduction zone at around 19 Myr (figure 6c). Subsequently the 2nd microcontinent gets accreted to the overriding plate as the most intense zone of thrusting is gradually transferred from between the microcontinent and the overriding plate to the base of the microcontinent. This gradual transfer is largely completed by 22.5 Myr with the new subduction channel now firmly located at the base of the microcontinent as the trailing oceanic domain starts subducting beneath it (figure 6d). The topographic evolution of the models is largely the same as that of model M2s but with lower topographic heights and no brittle deformation within the overriding plate. At 25.9 Myr the incoming continental margin reaches the new trench starting the phase of continent-continent collision. During this phase the accreted terrane gets progressively tilted while the original subduction zone located at its roof is also reactivated. By the end of the experiment at 30 Myr most of the accreted terrane is buried under the tip of the overriding continental crust but remains largely intact and preserved at a shallow depth (figure 6e).

g. Two weak microcontinents – consecutive terrane accretion
Model experiment M2w has an identical setup to the two previous models, but the microcontinents embedded within its oceanic domain have the characteristics of the one in experiment M1w, with a thicker weak detachment layer at its base (figure 7, supplementary animation A6, see also the strength envelopes on figure 1; for viscosity structure see supplementary figure S6). The model evolution is virtually identical to the previous experiment up to 9.6 Myr, when the shear-zone at the base of the microcontinental crust reaches the surface at its oceanward margin, 0.2 Myr earlier than in model M2i (figure 7a). This shear-zone is a bit broader as the weak detachment layer is a bit thicker and it becomes dominant over the original subduction interface 0.2 Myr earlier, accreting the terrane and transferring the trench to the oceanward edge of the microcontinent. Unlike in model M2i, a large portion of the 1st accreted terrane remains intact at a shallow depth (figure 7b). At 16.8 Myr the 2nd microcontinent arrives to the trench and begins to underthrust the accreted terrane in a similar fashion in which the 1st microcontinent previously underthrust the overriding plate. By 18.6 Myr a continuous shear-zone is active at the base of the 2nd microcontinent. At this point 3 distinct zones of deformation (original subduction interface, current subduction interface, new basal shear-zone) are active simultaneously but with a variable intensity (the current subduction interface being the dominant one; figure 7c). At around 19.8 Myr the subduction interface is gradually transferred once more, from the base of the 1st accreted terrane to the base of the 2nd
microcontinent, effectively accreting the 2\textsuperscript{nd} terrane. However, multiple zones of deformation remain active up to 22.6 Myr, when this latest subduction interface at the bottom of the 2\textsuperscript{nd} accreted terrane becomes the sole zone of deformation (figure 7d). The topographic evolution of the overriding plate remains subdued throughout with narrow (app. 100 km) and low (peak elevations of 4-5 km) orogens forming during terrane accretion events. At 24.8 Myr the incoming continental margin reaches the trench and begins the phase of collision. During collision the tip of the incoming continent underthrusts the accreted terranes while previous zones of deformation are progressively reactivated. By the end of the experiment at 30 Myr both accreted terranes are buried under the tip of the overriding continental crust or a sedimentary succession but they remain largely intact and preserved at a shallow depth (figure 7e).

3. Force calculations

a. Slab-pull force

The slab-pull force ($F_{sp}$) results from the negative buoyancy of the subducting slab relative to its surroundings. We calculate the temporal evolution of this force throughout the experiments using the method described in Erdős et al. [2021]: For each time-step all cells containing slab materials within the 130 to 660-km depth range are identified. We take a reference density profile of this depth range at the left side boundary and for each cell containing slab material (i.e., sediment, oceanic crust, oceanic and continental lithosphere, microcontinental crust and serpentinites), we calculate the density difference from the reference density profile at its corresponding depth, multiplied by the area of the cell and the gravitational constant. Finally, we integrate the resulting density difference array for the entire slab.

b. Integrated tectonic boundary forces

The integrated tectonic boundary forces ($F_{tbf}$) represent far-field forces resulting from large-scale plate-motions outside the modelled domain and exert a first order control on model-evolution. Note, that the tectonic forces are the net forces resulting from the given boundary conditions, slab-pull, and resisting forces, hence they evolve through time. For every timestep, on both vertical boundaries we integrate the horizontal deviatoric stress in the lithosphere. For each cell containing lithospheric material, we extract $\sigma'_{xx}$ and multiply its value by the thickness $l$ of each cell and sum over the vertical cross-section of the lithosphere:
\[ F_{ibf} = \sum_{b=1}^{2} \sum_{i=1}^{n(b)} \sigma'_{xx}(i, b) \cdot l(i, b) \]

where \( b \) is the variable identifying the boundary column and \( n \) is the number of cells within that column. By convention, negative values correspond to compression and positive values correspond to tension. In the presented models the tectonic boundary forces remain firmly in the compressional domain at all times. On figure 8 we plot the absolute value of the sum of the left and right tectonic boundary forces for an easier comparison with the slab-pull force and the interface resistance.

c. Interface resistance

The interface resistance force (\( R_{\text{interface}} \)) is the main resisting force in the subduction system \cite{Erdos2021, Forsyth1975, Wolf2019}. Here we approximate interface resistance using the second invariant of the deviatoric stress tensor, \( \sigma'_{\text{eff}} \), assuming that the stress tensor is dominated by the shear stress component along the interface.

For every timestep we track all cells within the 600 to 930 km horizontal and 5 to 150 km depth domain that contain materials that are expected to accrue intensive deformation (strain-rate \( 10^{-15} \) or higher), hence could be construed as part of the subduction channel (i.e., oceanic crust and sediment, microcontinental crust, continental sediment and eclogites). For each timestep the following calculation is performed:

\[ R_{\text{interface}} = \frac{1}{L_{\text{interface}}} \sum_{i=1}^{n} \sigma'_{\text{eff}}(i) \cdot A_{\text{cell}}(i) \]

where \( L_{\text{interface}} \) is an arbitrary length-scale chosen to be 150 km, \( n \) is the number of cells containing interface materials at the given timestep, \( i \) is a counter running through the cell-identifier off all these cells, and \( A_{\text{cell}} \) is the area of the individual cells.
4. Supplementary figures

**Figure S1**: Model M1w (one weak microcontinent) exhibiting terrane accretion. (a–d) Material colors (see Figure 1 and legend there) at key time steps, with selected isotherms displayed by red contours (300°C, 600°C, 900°C, and 1300°C) and areas experiencing high strain rates displayed with white contours ($10^{-13}$ to $10^{-14}$ s$^{-1}$). Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S2: Viscosity plot for reference model R0 exhibiting subduction, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S3: Viscosity plot for model M1s exhibiting terrane subduction, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S4: Viscosity plot for model M1i exhibiting terrane accretion, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S5: Viscosity plot for model M1w exhibiting terrane accretion, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
**Figure S6:** Viscosity plot for model M2s exhibiting divergent double subduction, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S7: Viscosity plot for model M2i exhibiting intra-oceanic deformation, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.
Figure S8: Viscosity plot for model 2Mw exhibiting multiple terrane accretion, at key time-steps. Red contour displaying the 1300°C isotherm. Green triangle shows location of the “trench” defined as the deepest point in the topography.