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Curved orogenic belts, back-arc basins, and obduction as consequences of collision at irregular continental margins

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ABSTRACT

Continental collisions commonly involve highly curved passive plate margins, leading to diachronous continental subduction during trench rollback. Such systems may feature back-arc extension and ophiolite obduction postdating initial collision. Modern examples include the Alboran and Banda arcs. Ancient systems include the Newfoundland and Norwegian Caledonides. While external forces or preexisting weaknesses are often invoked, we suggest that ophiolite obduction can equally be caused by internal stress buildup during collision. Here, we modeled collision with an irregular subducting continental margin in three-dimensional (3-D) thermo-mechanical models and used the generated stress field evolution to understand resulting geologic processes. Results show how tensional stresses are localized in the overriding plate during the diachronous onset of collision. These stresses thin the overriding plate and may open a back-arc spreading center. Collision along the entire trench follows rapidly, with inversion of this spreading center, ophiolite obduction, and compression in the overriding plate. The models show how subduction of an irregular continental margin can form a highly curved orogenic belt. With this mechanism, obduction of back-arc oceanic lithosphere naturally evolves from a given initial margin geometry during continental collision.

INTRODUCTION

The formation of highly curved orogenic belts is still debated (Rosenbaum, 2014). Proposed mechanisms include processes associated with deformation along thinned fold-and-thrust belts (Marshak, 1988), gravitational spreading (e.g., Edey et al., 2020), or along-strike migration rate variations of plate boundaries (Rosenbaum and Lister, 2004). The latter includes processes such as plate rotation, trench rollback, or indentation during continental collision (Rosenbaum, 2014; Rosenbaum and Lister, 2004).

The Alboran and Banda arcs are two modern systems that are associated with trench rollback and diachronous continental subduction along strike (Spakman and Hall, 2010; van Hinsbergen et al., 2014). Both regions experienced trench rollback against a highly nonlinear passive margin (see Fig. 1), where continental lithosphere subducted diachronously along the trench (Pownall et al., 2016; Spakman and Hall, 2010; van Hinsbergen et al., 2014). In the Banda region, continental material initially (15 Ma) subducted at Seram (Spakman and Hall, 2010), coinciding with trench rotation, and this was followed by continental subduction (4 Ma) at Timor, in the south (Fig. 1A). Oceanic subduction is now ceased, with collision occurring along the entire continental margin. The tectonic history of the Alboran Basin in the western Mediterranean is debated (van Hinsbergen et al., 2014). Most models agree that the slab retreated westward into the narrowing Alboran embayment since 30–20 Ma (Fig. 1B).

Both regions share puzzling geological features, such as extensional basins in the upper plate (Pownall et al., 2016; Watts et al., 1993), and exhumed subcrustal continental lithosphere (Frasca et al., 2017; Gueydan et al., 2019; Pownall et al., 2014). The Banda system involves ophiolite obduction (Ishikawa et al., 2007), but this has not been identified in the Alboran arc. The world's largest subcontinental mantle exposures, the Ronda and Beni peridotites in the Alboran arc (Gueydan et al., 2019), have been suggested to have been exhumed by gravitational collapse after slab break-off (Platt and Vissers, 1989; Van der Wal and Vissers, 1993), by thrusting of an older Jurassic rifted margin (Tubia et al., 2009), or by hyperextension of the overriding plate before, and thrusting during, continental collision (Frasca et al., 2017; Gueydan et al., 2019; see Fig. 1C). It is unclear in these examples how the geologic evolution is linked to the highly arcuate geometry and irregular shape of the subducting continental margin, if at all.

The aim of this study was to examine evolving plate stresses during irregular continental subduction and the formation of curved orogenic belts, to provide a better understanding of the drivers and controls of the geologic evolution. We show that localized tensional stresses during the initial stages of irregular margin collision cause short-lived back-arc rifting or spreading centers that are rapidly inverted once continental subduction spreads along the trench.

COLLISION WITH IRREGULAR PASSIVE MARGINS

We investigated the stress regime evolution during continental collision with an irregular margin (Fig. 2) in a model space of 3300 by 3960 by 660 km in size. The finite element code Citcom (Moresi et al., 1996; Zhong et al., 2000; Magni et al., 2014) solves the conservation equations for momentum, energy, mass, and composition. We used a visco-plastic rheology including diffusion and dislocation creep, lithospheric yielding, and an upper-limit viscosity (Magni et al., 2014; van Hunen and Allen, 2011). We did not apply external forcing; all dynamics were driven by internal buoyancy forces. Subducting and overriding plates were separated by a weak zone and decoupled from neighboring plates by weak transform faults to permit toroidal mantle flow (van Hunen and Allen, 2011; Magni et al., 2014). Continental lithosphere was free to move toward the trench and collide with the overriding

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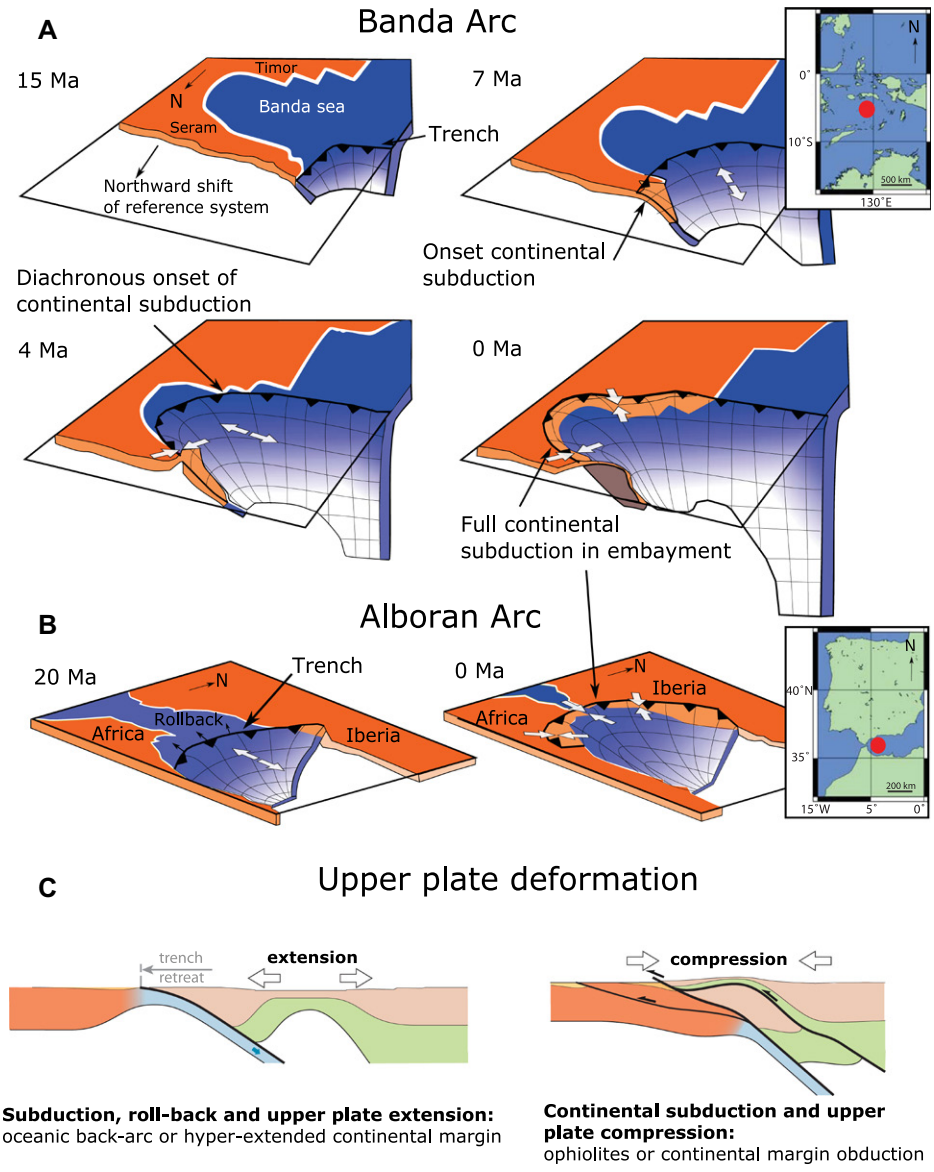

Figure 1. (A–B) Tectonic history of Banda (A) and Alboran (B) systems with diachronous continental subduction and trench rollback before continental subduction along entire trench ended subduction (overriding plate not shown in A–B). Orange—continental domains; blue—oceanic domains. (C) Mechanism for hyperextended margin obduction in Alboran domain (based on Gueydan et al., 2019), with extension in overriding plate predating continental subduction.

plate as the intervening oceanic basin closed. The continental crust was initially 40 km thick; the underlying mantle lithosphere extended from 100 km depth close to the trench to 150 km in the far-field area to mimic the thicker plate interior (Fig. 2B). The subducting passive margin included an oceanic embayment of (along-strike) width w_b and breadth (here used for the across-strike width) b_b, with flanks that sat at an angle α with respect to the convergence direction. To study the impact of this embayment on the evolving stress distribution, we varied the initial width (between 600 and 1200 km) and breadth (between 300 and 700 km) while keeping the flank orientation angle α constant. The convergence-parallel stress field, σ_{xx}, was computed to investigate the stress localization that could be responsible for the observed normal or thrust faulting.

Figure 3 depicts the model evolution of a subducting continent including an oceanic embayment $w_b = 1000$ km by $b_b = 600$ km. Initially, negative buoyancy of the oceanic slab drives subduction, trench rollback, and uniform extension in the overriding plate in all models (Fig. 3A). Initial continental subduction at the sides of the embayment locally reduces subduction velocities, and trench retreat stops (Fig. 3B). Trench retreat continues, however, at the oceanic embayment on the passive margin. With the local onset of continental collision (Fig. 3B), stresses start to vary significantly along the trench and upper plate. There is localized compression in collision regions, while extension close to the oceanic embayment continues to exert slab pull and trench retreat. The differential stresses cause yielding and rupture of the overriding plate. Opening of a back-arc spreading center (Fig. 3C) briefly increases convergence velocities by a factor of three (Fig. 4A). Shortly (~5 m.y.) after onset of back-arc spreading, the oceanic embayment is completely subducted, and this is followed by continental subduction (Figs. 3D–3F). At this point, the young oceanic back-arc basin measures only 166 km in ridge-perpendicular spreading distance (Fig. 3D). High continental crustal buoyancy reduces subduction velocities and stops trench rollback (Fig. 4A). Without local, rapid trench rollback driving the extensional stresses in the back-arc area, back-arc spreading ceases, and its stress state changes from an extensional to a compressional stress state within 10 m.y. after initial collision (Figs. 3E and 3F). During this last stage, slab break-off along the entire subducted margin ends subduction and allows the subducted continent to begin exhumation.

Subduction of smaller embayments (Fig. 4B) does not rupture the overriding plate to produce back-arc spreading, but it still thins the overriding continent; crustal thickness reduces from the initial 40 km to as little as 8 km (Fig. 4B; Fig. S1 in the Supplemental Material[1]) at the onset of collision. The embayment breadth, b_b, determines the duration of the overriding plate extensional phase, and hence the presence or absence of an oceanic back-arc basin. Too narrow embayments (small w_b) do not create sufficient tensional stresses, while the widest embayments ($w_b = 1200$ km) lack the required stress localization. When a back-arc spreading center is formed, its spreading duration varies from 1 m.y. in the smallest basin, to 5 m.y. (reference model, Fig. 3), to 10 m.y. in the models with the largest embayments ($w_b = 1200$ km and $b_b = 600$ km) and forming the largest back-arc basin (253 km cross-spreading distance).

DISCUSSION AND CONCLUSION

Subduction of an irregular passive margin allows formation of curved orogenic belts and a two-stage stress evolution from extensional to compressional states in the overriding plate during collision. Our models show how localized thinning or rupturing of the overriding plate occurs during local initial collision, while oceanic subduction continues elsewhere. Stress inversion and compressional deformation take place in a second stage, during full continental

[1]Supplemental Material. Detailed figures of the model evolution, as well as a short description of the methodology and parameters used. Please visit https://doi.org/10.1130/GEOL.S.15062262 to access the supplemental material, and contact editing@geosociety.org with any questions.

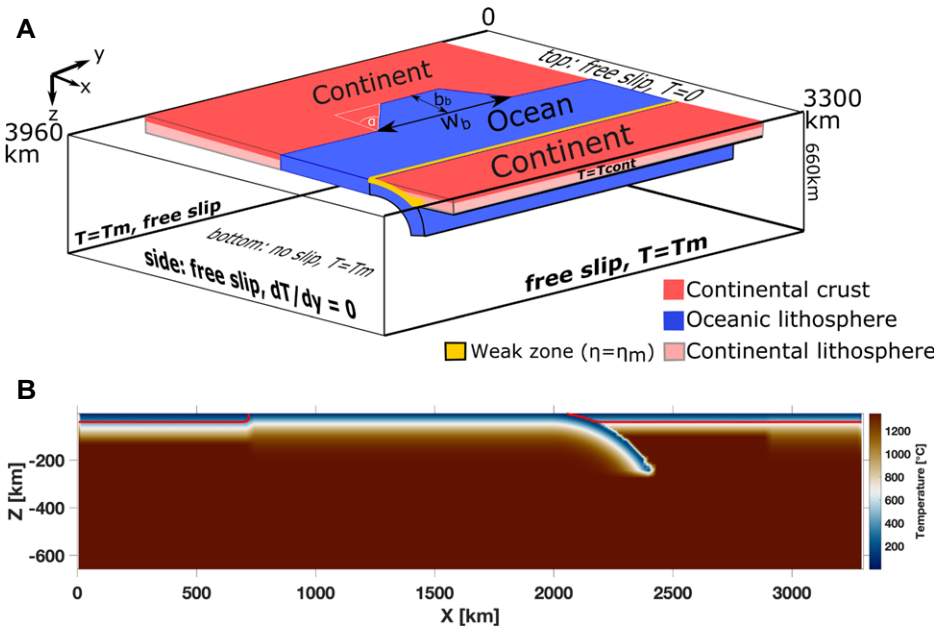

Figure 2. (A) Model setup: continental margin contains an oceanic embayment of breadth b_b and width w_b and collides with continental overriding plate after closure of oceanic basin. Overriding plate is fixed to right model boundary. Converging plates are decoupled by fixed-viscosity weak zone (yellow). See methodology for details. (B) Temperature (T) *x-z* cross section through center of model, with continental crust outlined in red. Tm—mantle temperature.

subduction. Such features are probably common in collision zones, given the naturally irregular shape of passive margins (Dewey and Burke, 1974).

Many numerical models of continental collision have focused on linear passive margins (Schliffke et al., 2019; van Hunen and Allen, 2011) or oblique collision (Bottrill et al., 2014) to study slab break-off and exhumation of subducted continental crust. Compressional stresses and resulting topography during continental collision are controlled by plate coupling (Faccenda et al., 2009), rheological flow laws (Pusok et al., 2018), and buoyancy ratios and convergence velocities (Pusok and Kaus, 2015). Also, lateral compositional variations along strike on the subducting plate can trigger the formation of back-arc spreading centers (Magni et al., 2014; Menant et al., 2016), under the precondition of nonlinear rheology (Pusok et al., 2018). Our models combined these approaches and showed that the resulting stress inversion from extensional to compressional states in the back-arc basins is similar to sequences proposed for the obduction of ophiolites (Cawood and Suhr, 1992) or hyperextended continental

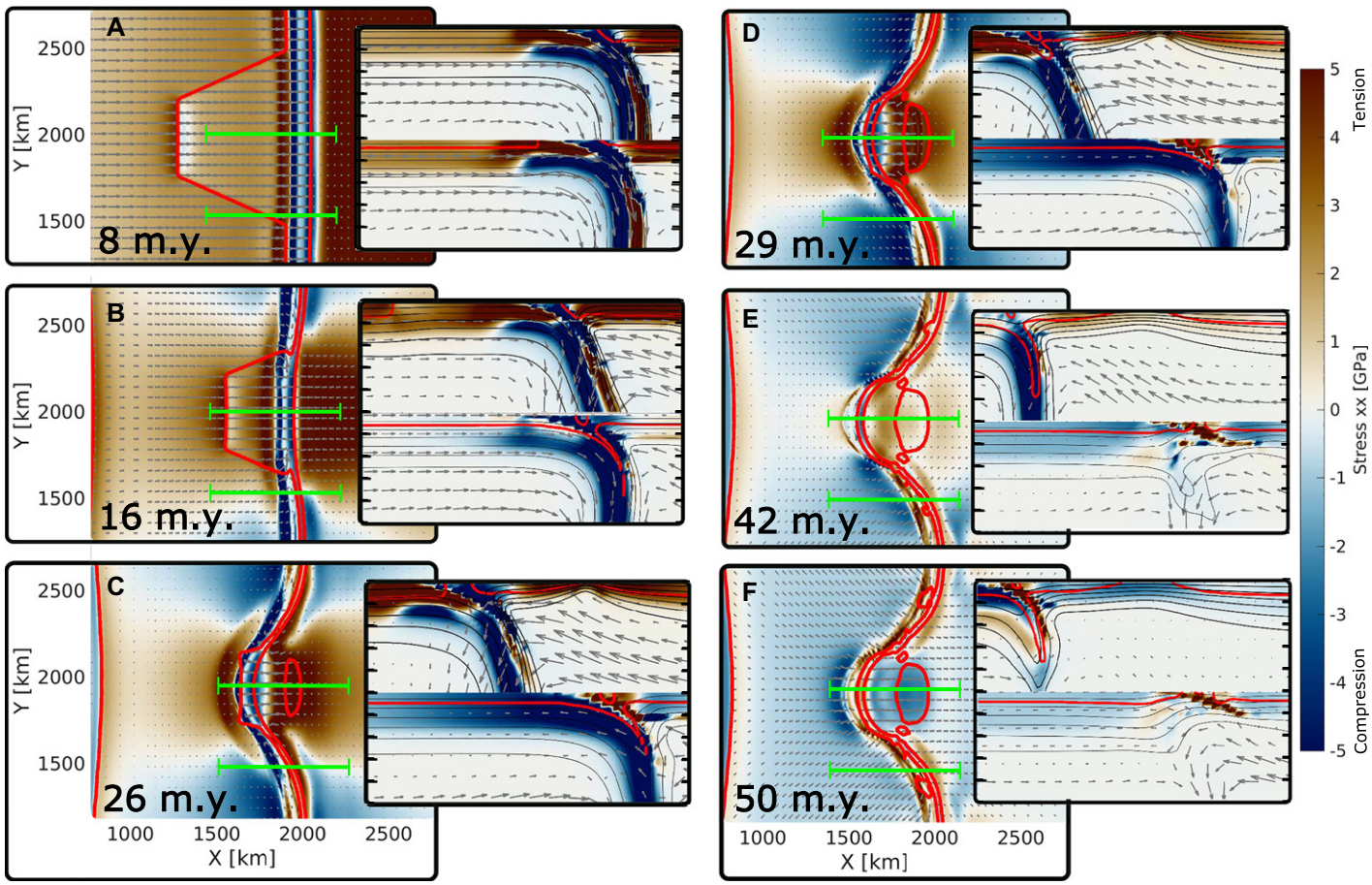

Figure 3. Stress evolution (σ_{xx}) at model surface with oceanic embayment measuring 1000 km by 600 km, where 250-km-deep vertical profiles are located at center and edge of embayment (green lines). Brown represents tension; blue is compression. Red contours outline continental crust; black lines are isotherms with 200 °C interval. Lateral onset of continental collision (A, B) triggers tensional stresses near embayment, opening back-arc basin (C, D). Once continental collision occurs within oceanic embayment, stresses in back-arc basin are inverted to shortening (E, F).

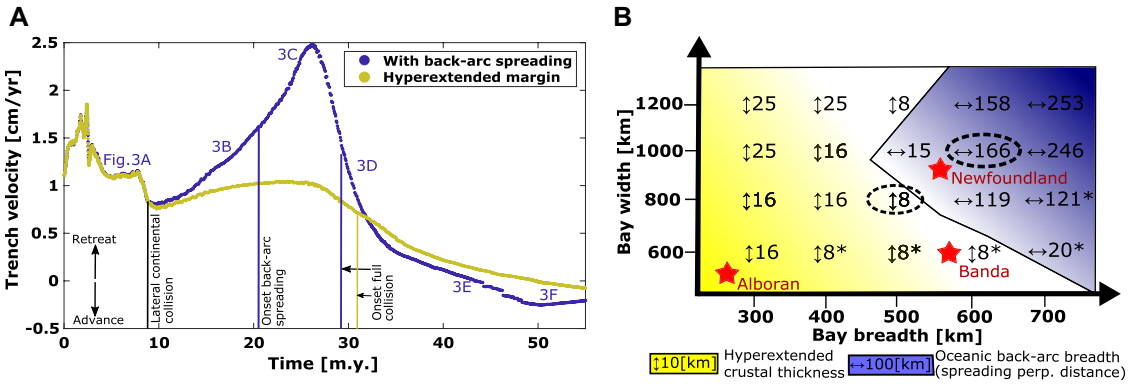

Figure 4. (A) Trench rollback velocities at center of embayment for model with back-arc spreading (blue; with corresponding time points in Fig. 3) and model with extended margin (yellow; see Fig. S1 [see text footnote 1]). The two corresponding models are circled in B. Opening of back-arc basin coincides with increase of trench rollback before it ceases with continental subduction. Without plate rupturing, trench velocities only increase marginally. (B) Thickness of hyperextended overriding crust (no back-arc basin formed) or breadth (i.e., spreading perpendicular distance) of newly formed oceanic back-arc domain upon continental subduction along entire trench for all studied parameters. Asterisks denote models where embayment becomes very narrow at its tip. In such models, slab breaks off before full subduction of oceanic embayment (see Fig. S2 [see text footnote 1]). Values for Newfoundland and Alboran and Banda arcs are estimates based on Cawood and Suhr (1992), Gueydan et al. (2019), and Spakman and Hall (2010).

margins (Gueydan et al., 2019). Previous modeling of ophiolite formation (e.g., Duretz et al., 2016; Hässig et al., 2016) typically prescribed a "thermal anomaly" (Duretz et al., 2016) or hot mantle upwelling (Hässig et al., 2016) close to a continental-oceanic margin to create a weak discontinuity. Shortening from far-field forcing then would lead to continental subduction at the weak discontinuity and emplacement of ophiolites on the continent. Our work suggests that such external factors are not necessarily required, since collision generates the weakening and compressional stresses in the back-arc area that are needed for the observed ophiolite obduction and crustal exhumation.

The interface between upper and lower plate in our models is a constant viscosity weak zone: This determines stress transfer between the two plates (Hassani et al., 1997; Faccenda et al., 2009). During the transition from subduction to collision, this coupling is likely to change (Luth et al., 2010). Lower fluid and melt percolation (Faccenda et al., 2009) or low sedimentation rates increase coupling (Burov and Toussaint, 2007) and may transfer compressional stresses into the back-arc area more efficiently than in our models. If so, the timing and degree of compressional stresses of the back-arc area in our models would be underestimated; i.e., a faster and stronger transfer of stresses should occur in nature. In contrast, too high plate coupling would end shortening and cause regional forearc uplift (Luth et al., 2010), with no shortening in the back-arc area.

Observations in the Banda and Alboran systems validate the sequence of stresses and processes predicted by our models (Gueydan et al., 2019; Spakman and Hall, 2010; van Hinsbergen et al., 2014; see Fig. 1C). Trench rollback into the Banda arc caused lateral arc-continent collision and synchronous opening of the extensional Weber Deep Basin (Pownall et al., 2016; Spakman and Hall, 2010). Subcrustal mantle lithosphere is thrusted at corners of the recent extensional basin (Pownall et al., 2014), where our models predict thinned crust and the most rapid change from extension to compression. In the Alboran arc, there is a high-velocity body along the entire arc subsurface, interpreted as a continuation of the peridotite at Ronda and Beni (Gueydan et al., 2019). A process of rifting by localized tensional stresses close to the trench during rollback (Fig. 1C), followed by compression and thrusting during continental subduction, has been suggested to trigger hyperextended continental margin obduction (Gueydan et al., 2019). In the Alboran arc, trench rollback rates have reduced from rapidly retreating (up to 8 cm/yr) during narrowing of the slab (van Hinsbergen et al., 2014) to subduction cessation coinciding with subduction of continental crust along the entire trench. Our models showing the highest retreat rates during narrowing of the oceanic basin followed by subduction cessation with full continental collision (Fig. 4A) fit these trends.

Diachronous continental subduction has further been suggested to have formed young back-arc ophiolites by thrusting onto a continental margin during large-scale continental collision in the Caledonide orogen (Cawood and Suhr, 1992; Slagstad and Kirkland, 2018). During Baltica-Laurentia collision, diachronous initial collision in Newfoundland (Cawood and Suhr, 1992) and Norway (Slagstad and Kirkland, 2018) coincided with extension, as shown by mafic layered intrusions and ophiolite creation, while obduction is dated ~15–20 m.y. later. With the onset of back-arc spreading predating continental subduction along the entire trench by ~5 m.y. and predating compression in the back-arc area by ~20 m.y. in our models, we estimate obduction within the period of 5–20 m.y. of back-arc formation, similar to the observations. The limited (trench perpendicular) ophiolite size (<100 km) in the Norwegian Caledonides and Newfoundland hints that the newly formed spreading center must have been small during onset of collision, and the spreading center was close to the main suture where the back-arc obducted.

In conclusion, collision of a continent with an irregular passive margin not only can form highly arcuate orogenic belts, but it can also cause complex geological processes in the overriding plate resulting from the associated transient stress changes. Such a setting causes localized crustal thinning, rapidly followed by crustal shortening, and it provides an ideal intrinsic mechanism for obduction of ophiolites or a hyperextended continental margin, without a need for any far-field forcing or preexisting weaknesses in the upper plate.

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