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Earth and Planetary Science Letters 212 (2003) 31–45

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Subduction versus accretion of intra-oceanic volcanic arcs: insight from thermo-mechanical analogue experiments

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Received 13 February 2003; received in revised form 25 April 2003; accepted 26 April 2003

Abstract

We perform thermo-mechanical laboratory experiments designed to explore the behaviour of the volcanic arc during intra-oceanic arc–continent collision following oceanic subduction and subsequent back-arc opening. The overriding oceanic lithosphere is made of two layers representing the oceanic crust and the lithospheric mantle. This lithosphere carries a volcanic arc and is thinned and weakened beneath both the arc and the back-arc basin. The subducting plate contains three parts: one-layer oceanic and two-layer (crust and mantle) continental lithosphere with a continental margin between them. When the continental margin reaches the trench and starts subducting, the overriding plate undergoes growing horizontal compression and finally fails in the vicinity of the back-arc spreading centre, which is the weakest part of this plate. The failure can result in subduction of the whole arc plate comprised between the trench and the back-arc spreading centre. During subduction of the arc plate, the mantle part of this plate subducts completely, while the behaviour of the arc crust depends on its thickness and strength, which is a function of composition and temperature. We tested four cases with different arc crust thicknesses and composition (rheology), with total lithosphere thickness in the arc being constant. Three types of tectonic evolution have been obtained: complete arc subduction, complete arc accretion, and partial arc subduction/accretion. The result is largely controlled by the crustal thickness of the arc. A thin arc (thickness equivalent to ~16 km in nature) made of the same strong material as the oceanic crust subducts completely without leaving any trace at the surface. On the contrary, a thick arc (equivalent to ~26 km in nature) made of the same material is scraped off and accreted to the overriding plate. The lower crust of such an arc is hotter, therefore its strength at 'Moho' depth and coupling between crust and mantle are small. In addition, the thick arc has a high isostatic relief and hence a greater mechanical resistance to subduction. Therefore, the arc is scraped off. If the arc is made of a weaker 'continental-like' material or contains a weak layer/low friction interface, it is completely or partially scraped off even if it is small. When there is no back-arc opening before collision (no thin and weak lithosphere in the rear of the arc), the overriding plate fails in the arc area, which may result in a complete fore-arc block subduction, with the volcanic arc remaining at the surface. The obtained models are compared with mountain belts with nearly no trace of arc activity (Oman), with accreted arc (Kohistan), and with small remnants of subducted arc (southern Tibet).

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doi:10.1016/S0012-821X(03)00239-5

Keywords: physical modelling; geodynamics; arc–continent collision; continental subduction; volcanic arc; back-arc closure; litho thermo-mechanics

1. Introduction

The general sequence of tectonic events leading to the formation of collisional mountain belts includes closure of an ocean through oceanic subduction, subduction of the continental margin (arc–continent collision and/or obduction) and subduction/deformation of the continental lithosphere (continent–continent collision). Oceanic subduction typically results in the formation of magmatic arcs. Therefore, collisional mountain belts should normally contain remnants of the arcs. However, such remnants are less systematically found than expected. For example, the lack of arc-related rocks in the Alps leads most of the authors to consider that Cretaceous subduction of the Tethys Ocean beneath the Adriatic plate was amagmatic [1,2]. One can suppose that subduction in this area was relatively flat as, for example, in Costa Rica or Central Chile [3], or that the amount of the subducted oceanic lithosphere was too small to initiate the magmatic activity [2]. An alternative is to suppose that an intra-oceanic magmatic arc has existed and is not found because it has been entirely subducted. The India–Asia collision has been preceded by long-lived intra-oceanic subduction of the Tethys lithosphere which therefore should have resulted in the formation of the island arcs. Indeed, there are large Kohistan–Ladakh arcs accreted in the western part of the Himalayas [4–6]. On the other hand, further to the east the southern-Tibetan part of the suture (Tsangpo Suture) was reputed to include no intra-oceanic volcanic arc, with subduction producing only the active continental margin preserved in the Transhimalaya Belt [7,8]. The recent discovery of calc–alkaline volcanic and volcaniclastic rocks in this area [9–11] indicates that an island arc did exist and was active in the Cretaceous, but then has almost entirely disappeared. The arc could not have been 'erased' by erosion [12]. Therefore, the most plausible mechanism of arc disappearance is its subduction. Arc subduction has been obtained in both experimental and numerical models [13,14]. This process is part of a larger-scale phenomenon, the subduction of the whole arc plate comprised between the trench and the back-arc spreading centre. This process was shown to occur when collision follows oceanic subduction of extensional regime (with back-arc opening). With the onset of continental margin subduction, the convergence system enters into a compressional regime: extension in the back-arc basin ceases and is then changed by compression. Increasing compression causes failure of the overriding plate in the vicinity of the extinct back-arc spreading centre, which can lead to the subduction of the whole arc plate.

In the previous lithospheric-scale modelling [13] the overriding lithosphere has been presented by a single layer. The oceanic and volcanic arc crusts were not integrated and hence their deformation could not be studied. This deformation is the subject of the present study. We have performed new thermo-mechanical laboratory modelling of arc–continent collision. The overriding oceanic plate now consists of the lithospheric mantle, the oceanic crust and the arc crust. Below, we first summarise the present-day knowledge on the structure and properties of active intra-oceanic island arcs. This information is then used to define the experimental set-up. The experiments have revealed two end-member behaviours of the arc during subduction. They mainly depend on the thickness of the arc crust and the presence of décollement surfaces within this crust: a small and strong arc can be entirely subducted into the mantle, while a thick and/or weak arc is only partially subducted. These results are applied to modern and ancient collisional systems.

2. Arc structure

The structure of island arcs is ill-constrained because of very limited modern controlled-source data and a lack of seismic velocity resolution at depth. The eastern Aleutian [15,16], the Izu-Bonin

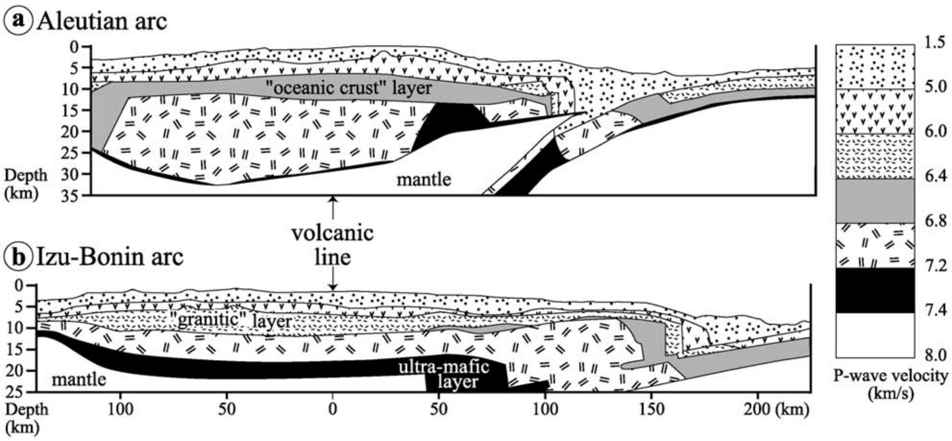

Fig. 1. P-wave velocity structure of the Aleutian (a) [16,20] and Izu-Bonin (b) [17,20] arcs.

[17,18] and to some extent the Mariana [19] are the best-studied island arcs and are essential references for the following summary.

Arcs have various crustal thicknesses: 30 km in the Aleutian arc, 20–22 km in the Izu-Bonin arc, and about 16 km in the Mariana arc [17–19]. As usual, seismic velocities increase with depth, but they vary considerably from one arc to another, suggesting different crustal composition. For example, the deepest layer in the Aleutian arc has P-wave velocities of 6.8–7.2 km/s and is interpreted as a 15 km thick sequence of mafic to ultramafic cumulates (Fig. 1a). The middle crustal layer has thickness and velocity corresponding to the oceanic crust [15]. The lower crust of the Izu-Bonin arc is more 'rapid' (there is a layer with velocities of about 7.3 km/s, which is absent in the Aleutian arc) and thinner (∼10 km; Fig. 1b). Moreover, the middle, ∼5 km thick layer of this arc has velocities around 6 km/s, believed to correspond to a granitic layer [17]. The Izu-Bonin arc appears therefore as more mature than the Aleutian arc, although both have approximately the same age of about 50 Ma (55–50 Ma for the Aleutian [21] and 48 Ma for the Izu-Bonin [17] arc). The age thus does not directly control the arc crustal structure. What are these controlling factors? An important one is the arc rifting (initial stage of back-arc opening) which, on the one hand, reduces the crustal thickness and, on the other, removes the part of the arc that becomes a remnant arc after the back-arc basin opening. Rifting also changes the thermal history (hence mineralogical composition) of the arc and is commonly accompanied by mafic and ultramafic intrusions [22]. The Izu-Bonin and Aleutian arcs have different rifting histories. The Izu-Bonin arc has experienced one cycle of rifting and back-arc spreading and is presently being rifted again [17], while the Aleutian arc has never been split. It is therefore reasonable to assume that extension has impelled some differences between these two arcs. We conclude that the thickness of the arc is not indicative of its maturity and temperature (lithosphere thickness). A young arc may be thin and cold (i.e. with a thick lithosphere). Yet, a thin arc can be old and have a thin lithosphere (be hot) if it has undergone recent rifting. In terms of modelling set-up, we must thus test hot versus cold and thin versus thick arcs with little consideration of age, because any thickness/temperature combination may have happened.

2.1. Constraints on the models

The first parameter to be tested is thus the arc crust thickness, which for the chosen examples ranges between ∼16 km for the Mariana and ∼30 km for the Aleutian arcs. The effective strength of the arc is another controlling parameter. It depends on the arc composition, which is heterogeneous and changes from ultramafic/mafic in the lower crust to more felsic (andesitic) in the upper crust [15]. Since island arcs have nearly the same bulk composition as the oceanic crust [16], the model arc in most experiments was made of the same material as the model oceanic crust. A weaker material was used in some experiments to test the behaviour of predominantly felsic arcs or those weakened by various faults, low-strength layers or décollement surfaces that may be present within the arc.

The arc strength also depends on the thermal gradient and/or the total lithosphere thickness. Tomographic data reveal well-developed low-velocity zones beneath volcanic arcs, which are interpreted as caused by ascending mantle flow reducing the overriding plate thickness [23,24]. The arc lithosphere is thus very thin (e.g. <30 km in NE Japan [24]). We adopted approximately this value to calculate the arc lithosphere thickness in the model.

Coupling between the crust and the mantle is one more parameter that should affect the arc behaviour during subduction. Early ideas professed that volcanic rocks are progressively accumulated, resulting in a passive growth of the arc upon the surface of the oceanic crust covered by the sedimentary (lubricating) layer. In reality, magmatic underplating reworks the lower crust of the arc, the oceanic crust and the underlying mantle, which provides a rather continuous transition from deep ultramafic to top andesitic crust. Therefore, a sharp interface between the crustal arc and its mantle is unlikely and it seems reasonable to assume that crust/mantle coupling is strong in the sense that there is no décollement layer. This is especially true for arcs that have undergone rifting and that are somehow reinforced by strong mafic and ultramafic intrusions. Coupling between the arc and its substratum will thus be defined by the lower crust strength, which depends on temperature and hence on the crustal thickness (assuming the total lithosphere thickness to be constant). Therefore, in experiments we apply strong coupling between the arc crust and its basement (the oceanic crust and lithospheric mantle) before subjecting the models to a thermal gradient.

However, we also tested a low coupling interface between the arc crust and the underlying oceanic crust to cover supplementary eventualities.

3. Modelling set-up

The models include two ‘lithospheric’ plates resting on the liquid ‘asthenosphere’ (Fig. 2). The overriding plate corresponds to the two-layer oceanic lithosphere with coupled oceanic crust and lithospheric mantle. The oceanic crust material is stronger than the continental crust (Table 1 and Fig. 3), while the mantle layers of both plates have the same composition and hence properties.

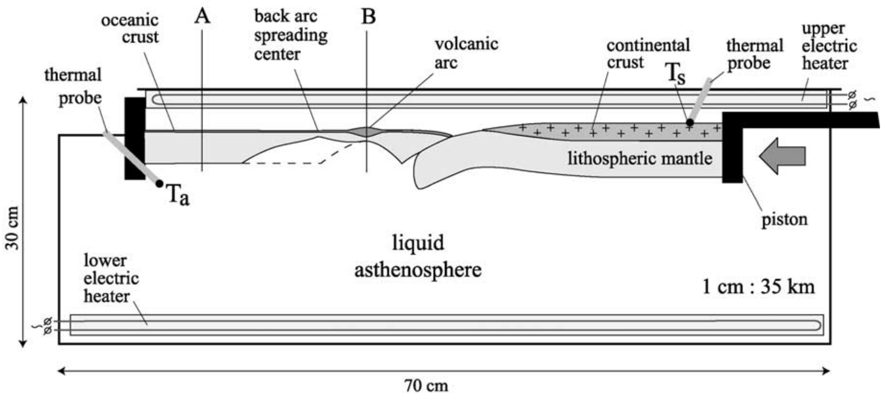

Fig. 2. Scheme of the experimental set-up. Dashed line in the lower part of the overriding plate shows the base of this plate in experiment 1 where young back-arc basin is absent. T_s and T'_a are the temperatures of the lithosphere surface and the asthenosphere, respectively. Sections A and B correspond to the strength envelopes in Fig. 3.

Table 1
Parameter values adopted for the model and scaled to nature

Parameter	Symbol	Prototype	Model
Average yield limit of the oceanic crust	σ_{oc} (Pa)	6.8×10^7	60
Average yield limit of the continental crust	σ_{cc} (Pa)	2.62×10^7	23
Average yield limit of the continental lithospheric mantle	σ_l (Pa)	3.12×10^7	27
Density of the oceanic crust	ρ_{oc} (kg/m^3)	2.8×10^3	0.86×10^3
Density of the continental crust	ρ_{cc} (kg/m^3)	2.8×10^3	0.86×10^3
Density of the lithospheric mantle	ρ_l (kg/m^3)	3.37×10^3	1.02×10^3
Density of the asthenosphere	ρ_a (kg/m^3)	3.3×10^3	10^3
Thickness of the oceanic crust	H_{oc} (m)	7×10^3	2×10^{-3}
Thickness of the continental crust	H_{cc} (m)	2.8×10^4	8×10^{-3}
Thickness of the continental lithospheric mantle	H_l (m)	5.25×10^4	1.5×10^{-2}
Convergence rate	V	3 cm/yr	3×10^{-5} m/s
Thermal diffusivity of the lithosphere	κ (m^2/s)	10^{-6}	8×10^{-8}
Time	t	1 Ma	4.6 min

Because our materials are temperature-sensitive and due to the thermal gradient imposed in the model, the strength of the materials decreases with depth in each layer. In this table we indicate the strengths averaged over the layer thickness, which are half of the sum of the maximal and minimal values.

The overriding plate carries the volcanic arc made of the same material as either the strong 'oceanic' or the weak 'continental' crust. To get the above-mentioned strong coupling between the arc crust and oceanic crust, both are made as one single body. In experiment 3 presented later, these two units are separated by a low coupling (friction) interface. The overriding plate in all experiments is thinned under the volcanic arc (equivalent lithospheric thickness here is 30 km, as indicated above). In most of the experiments this plate has also been thinned in the rear of the arc (Fig. 3) to simulate young back-arc lithosphere formed just before the arrival of the continental margin into the subduction zone. The lithosphere at the back-arc spreading centre is very young, thin and weaker than in the arc. Subduction/collision is driven by a piston moving at a constant rate throughout the experiment and the pull force generated by the subducted oceanic lithosphere and the continental lithospheric mantle, both being slightly denser than the asthenosphere (Table 1). In similar, purely mechanical experiments reported by Chemenda et al. [13] it was shown that during subduction of the continental lithosphere the overriding plate fails near the spreading centre. Failure can occur along either of two possible oppositely dipping directions depending in particular on the distance L between the trench and the back-arc spreading centre. Failure along the arc-verging fault is followed by subduction of the arc plate, whereas failure along the arc-dipping fault results in subduction reversal. Similar results have been obtained when the overriding plate contains only one weak zone in the arc area (no young back-arc lithosphere). In this case failure occurred along a fault dipping under the arc in either of two possible directions, resulting either in subduction reversal or fore-arc block subduction [13,14]. In the present paper we are interested in arc subduction. Therefore we have chosen an L value ($L=8.8$ cm, corresponding to ~ 300 km in nature) that provides an arc-verging failure plane, synthetic to the main subduction. The principal difference between the present experiments and those reported earlier [13] is the structure of the overriding plate, which now contains both oceanic and arc crust. New models are thermo-mechanical. The model lithosphere is made of temperature-sensitive materials (hydrocarbon compositional systems [25]), and the whole model is subjected to a vertical thermal gradient. Temperature T_s at the model surface and T_a in the 'asthenosphere' ($T_a > T_s$) are imposed and controlled using high-precision thermal probes, an auto-adaptive thermo-regulator and electric heaters (Fig. 2). In the experiments presented, $T_s=38$°C and $T_a=42$°C. The thermal gradient within the lithosphere before deformation (subduction) is mainly defined by its thickness as the

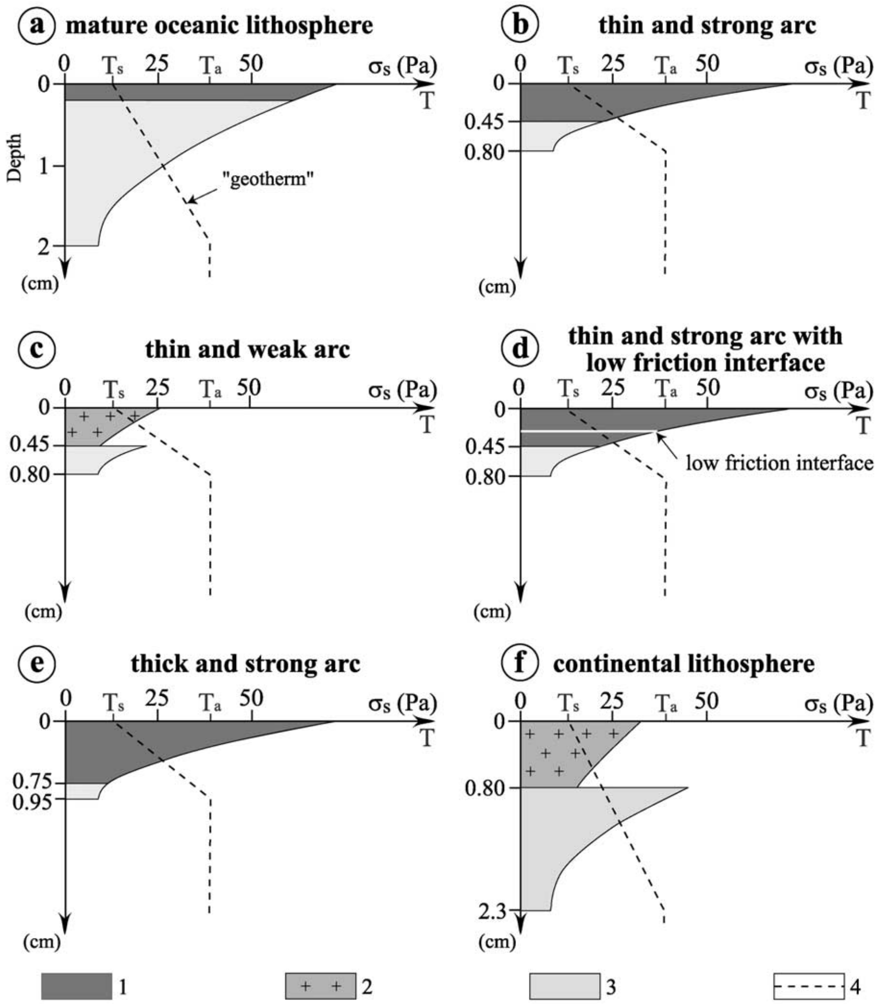

Fig. 3. Strength envelopes of the model lithosphere. The strength of each model material has been measured experimentally (shear tests) for temperatures between T_s and T_a. The thermal gradient in various models was also measured. The strength envelopes presented were calculated using these experimental data. (a) Mature oceanic lithosphere corresponding to section A in Fig. 1; (b–e) Thinned oceanic overriding lithosphere in the volcanic arc area (section B in Fig. 1): (b) thin (4.5 mm, equivalent to 16 km in nature) and strong arc crust made of the same material as the oceanic crust; (c) thin and weak volcanic arc made of the same material as the continental crust; (d) the same as in panel b with low-friction interface inside the arc; (e) thick (7.5 mm, equivalent to 26 km in nature) and strong volcanic arc; (f) continental lithosphere. 1, oceanic crust; 2, continental crust; 3, lithospheric mantle; 4, 'geotherm'.

thermal diffusivity of the model crust and the lithospheric mantle are close (Table 1). Thermal and mechanical structures of the 'normal' oceanic and continental lithospheres as well as four tested variants of the arc lithosphere properties are presented in Fig. 3. Note that the deformation of the subducting continental lithosphere is not the subject of this study.

4. Similarity criteria

The similarity criteria in the present experiments are the following [26]:

$$\sigma_{oc}/\rho_{oc}H_{oc} = \text{const},\ \sigma_{cc}/\rho_{cc}H_{cc} = \text{const},$$

$$\sigma_1/\rho_1 g H_1 = \text{const},\ \rho_1/\rho_a = \text{const},\ \rho_1/\rho_{cc}$$

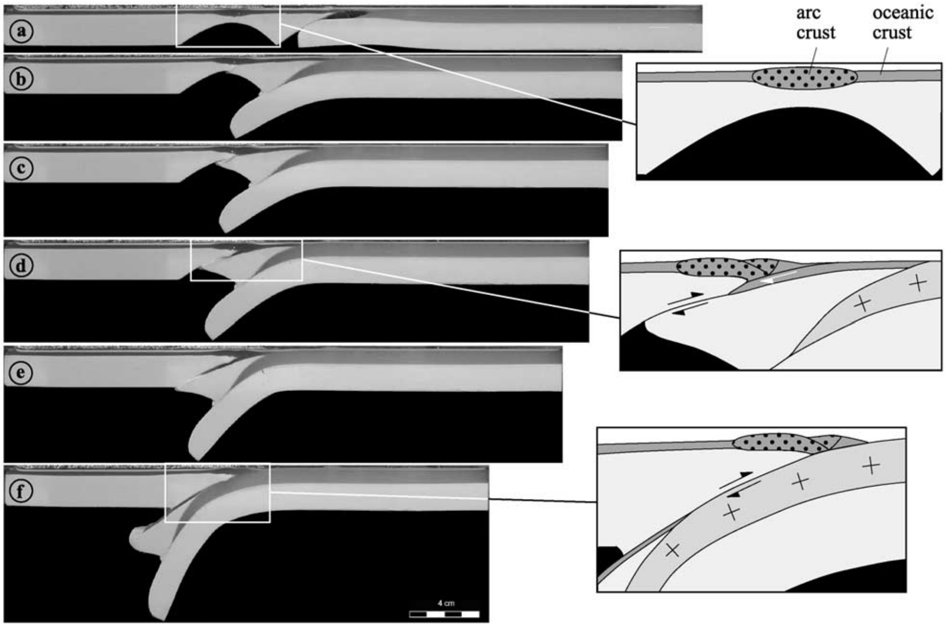

Fig. 4. Successive stages of experiment 1 with zoom-in of some parts of the deformed model. There is no young back-arc basin; therefore the overriding lithosphere contains only one thin zone beneath the arc. The lithosphere fails below the arc crust (panels b and c), resulting in the fore-arc block subduction (panels d–f). Only small parts of the fore-arc oceanic crust are scraped and accreted (zoom-ins in panels d and f). The arc crust is slightly deformed and remains in its place.

$= \text{const},\ \rho_l\rho_{oc} = \text{const},\ H_l/H_{cc} =$

$\text{const},\ H_{cc}/H_{oc} = \text{const},\ Vt/H_l = \text{const},$

$$VH/\kappa = \text{const}; \quad (1)$$

where σ_{oc}, σ_{cc} and σ_l are the average yield limits under normal loading of the oceanic crust, the continental crust, and the lithospheric mantle, respectively. As the strength of the materials decreases downward in each layer due to the temperature increase, the yield limits used for the similarity criteria are averaged over the layer thickness. H_{oc}, H_{cc} and H_l are the thicknesses of the oceanic crust, the continental crust, and the mantle of the continental lithosphere, respectively. ρ_{oc}, ρ_{cc}, ρ_l and ρ_a are the densities of the oceanic crust, the continental crust, the mantle, and the asthenosphere. V is the convergence rate; t is the time; κ is the thermal diffusivity of the lithosphere and H is the total thickness of the continental lithosphere. The parameter values assumed for the prototype and the model which satisfy the similarity conditions 1 are given in Table 1.

5. Results

A total of 33 experiments have been performed. We present the five most representative experiments, which demonstrate end-member scenarios for arc deformation during arc–continent collision.

Experiment 1 (Fig. 4): The overriding plate contains only one weak (thinned) zone in the arc area; there is no young back-arc lithosphere (the dashed line in Fig. 2 shows the base of this plate). The strength envelope of the arc lithosphere is presented in Fig. 3b: the model arc is made of a 4.5 mm thick 'oceanic' crust, which scales up to 16 km in nature. The trench/arc-axis distance is 6.5 cm (~230 km in nature), which provides failure of the overriding plate along a continent-verging fault dipping under the arc (Fig. 4a), see Tang et al. [14]. Failure is

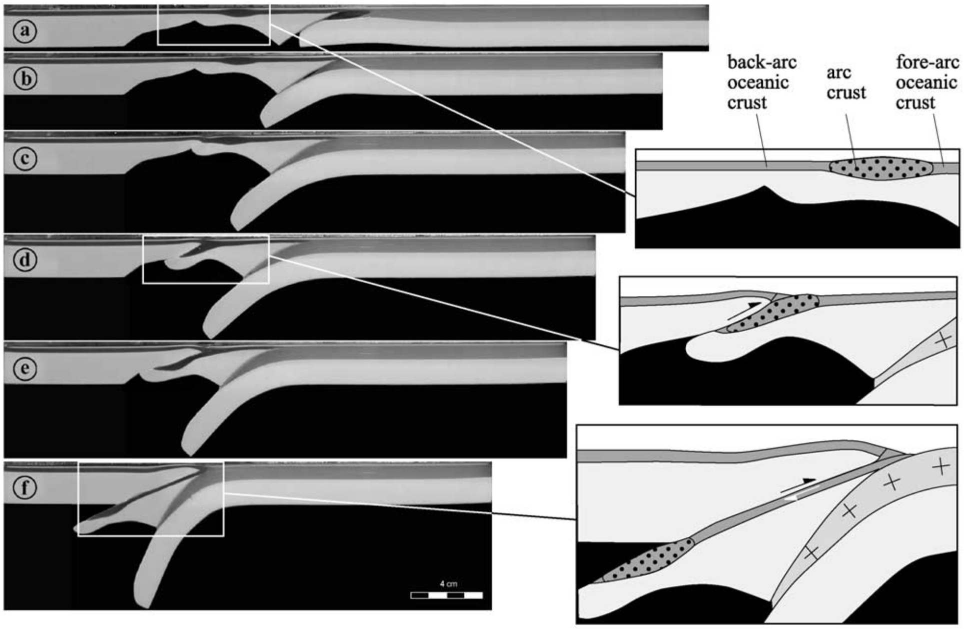

Fig. 5. Successive stages of experiment 2. The arc is thin and strong (corresponding to Fig. 3b). The arc subducts almost completely, leaving on the surface only small blocks of scraped off and accreted back-arc and arc crust (zoom-ins in panels d and f).

followed by complete subduction of the fore-arc block (Fig. 4d–f). The arc crust and a scraped and accreted slice of the oceanic crust remain at the surface (Fig. 4f). In the alternative scenario with subduction reversal, the arc also remains at the surface and new subduction develops under the arc [14,27].

Experiment 2 (Fig. 5): The overriding plate is the same as in experiment 1 but now includes a young 'back-arc basin lithosphere' (see Fig. 2). The arc is strong and thin (Fig. 3b). The initial stages of this test (Fig. 5a–c) are similar to those of purely mechanical experiments with a one-layer overriding plate [13]: subduction of the continental margin causes an increase in the horizontal compression of the overriding plate which fails near the 'extinct back-arc spreading centre'. A subduction jumps to this new location. At the initial stages, the new subduction accommodates almost all convergence (the initial subduction zone is practically inactive) and 'consumes' the back-arc segment of the arc plate. This process then slightly slows down when the arc enters the new subduction zone, resulting in reactivation of the continental margin subduction (Fig. 5d). The whole arc plate then subducts into the 'asthenosphere', which leads to complete disappearance of the arc from the surface at the end of the experiment (Fig. 5f). Continental subduction then resumes to absorb further convergence, and the oceanic lithosphere of the back-arc basin is obducted on the subducting continental lithosphere. The experiment was stopped at that stage.

Experiment 3 (Fig. 6): The same set-up as in experiment 2 is applied, but the arc is weaker (made of the 'continental crust'; Fig. 3c). As in the previous experiment, failure occurs near the back-arc spreading centre (Fig. 6b,c), resulting in arc plate subduction (Fig. 6c–f). The weak volcanic arc is scraped off with part of the fore-arc oceanic crust (Fig. 6d). The arc undergoes large deformation, being jammed between the overriding plate and the fore-arc block which continue to subduct (Fig. 6d). At the same time, the continental margin continues slow subduction under the fore-arc block until stage *e*. After the arc plate is consumed, there remains only one subduction zone. A small amount (about 20%) of the arc is

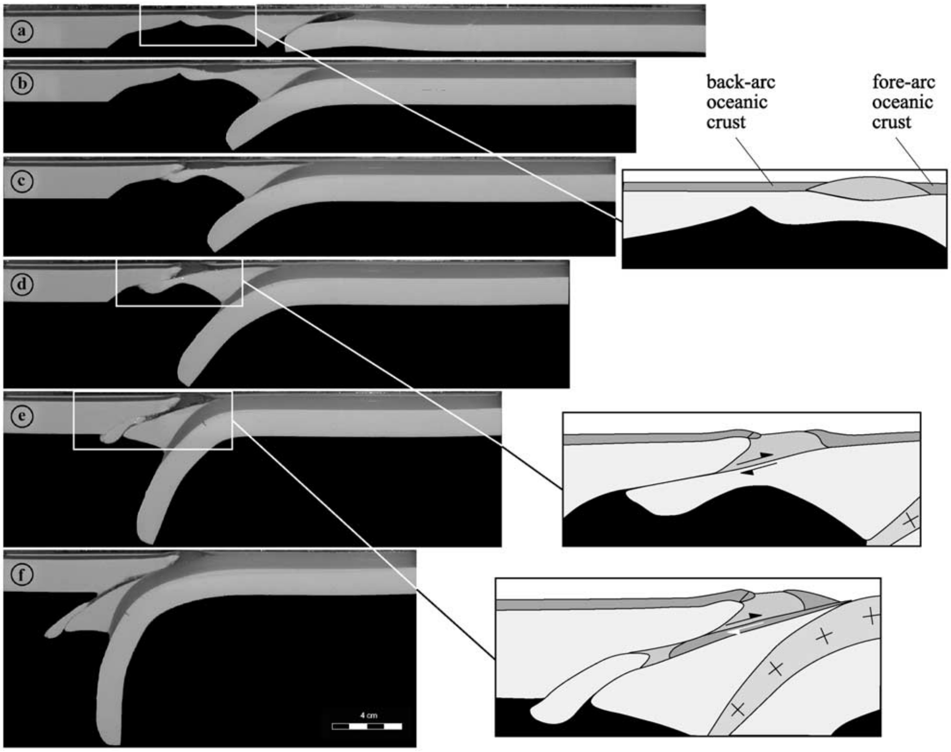

Fig. 6. Successive stages of experiment 3. The thin and weak 'continental-like' volcanic arc (corresponding to Fig. 3c) is deformed and accreted along with small amounts of the fore-arc oceanic crust (zoom-in in panel c).

dragged into the subduction zone (Fig. 6e), but its main part is accreted at shallow depth. Relatively small volumes of the back-arc and fore-arc oceanic crust are also accreted to the overriding plate and obducted onto the continental lithosphere (Fig. 6f). During subduction of the lithospheric mantle of the arc plate, a second failure occurs in the thinned plate that was the former arc mantle (Fig. 6e).

Experiment 4 (Fig. 7): As in experiment 2, the arc is thin and strong, but also comprises a low-friction interface at the surface of the 'initial oceanic crust' (Fig. 3d). Cohesion along this interface is 20–30% lower than the shear strength of the crust. In this case, the upper part of the arc is shifted from its substratum during subduction (Fig. 7d) and is accreted at very shallow depth (Fig. 7e,f). The convergence is almost entirely accommodated by the arc plate subduction, the continental lithosphere subducting together with this plate. The last stage of this experiment (Fig. 7f) shows that the accreted upper arc becomes part of the overriding plate and is obducted on the subducting continental lithosphere.

There was no need to test the same decoupling conditions with a weak crust because it is obvious that the low-strength, 'continental-like' crust will also be, and even easier, scraped off and accreted.

Experiment 5 (Fig. 8): In this experiment, the arc crust is made of the same strong material as the oceanic crust, but the arc crust thickness is 7.5 mm (equivalent to 26 km in nature). The strength envelope of such an arc lithosphere is presented in Fig. 3e. After failure of the overriding plate at stage *c*, the arc undergoes considerable deformation when it enters the new subduction. During this stage (Fig. 8c,d), the convergence is absorbed by subduction of both the arc plate and the incoming continental lithosphere. After subduction of a few tens of equivalent kilometres of the arc plate, the arc crust is scraped off (Fig. 8d). The thin and hot (hence, very weak) arc lithosphere

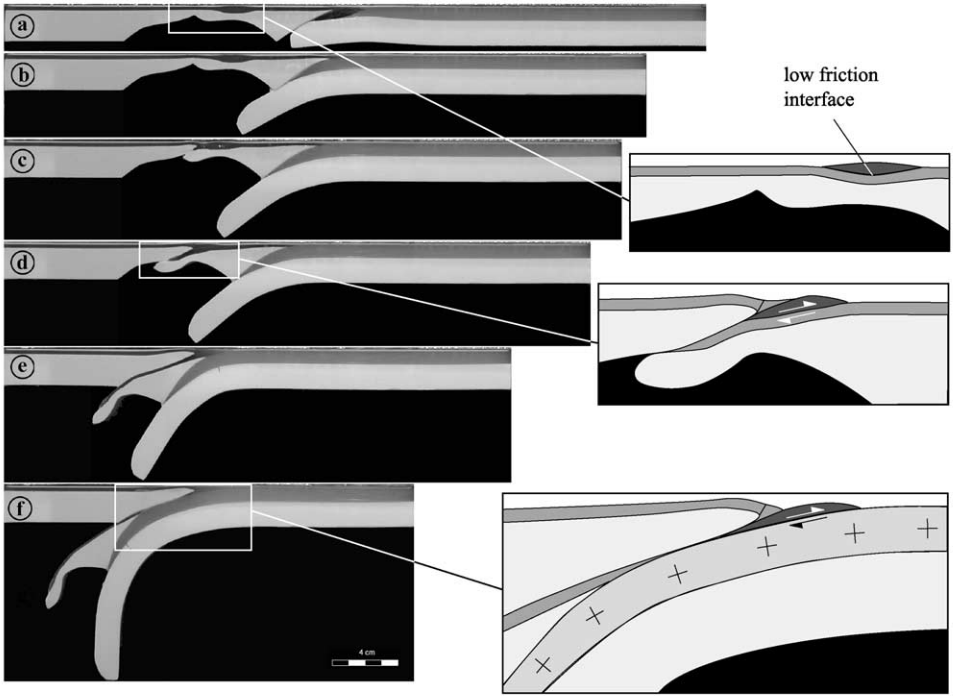

Fig. 7. Successive stages of experiment 4. The arc is strong, but there is a low-friction interface within it (zoom-in in panel a), which is a décollement surface along which the upper part of the crust is scraped off just at the onset of the arc subduction (panels d–f).

fails at depth (Fig. 8e) and remains under-accreted against the overriding plate, while the fore-arc block continues subducting. The whole mantle part of the arc plate disappears in the asthenosphere, whereas a large part of the deformed arc crust remains at shallow depth and at the surface, being obducted on the continental lithosphere (Fig. 8f).

6. Discussion

Cloos [28] argued that the low density of an intra-oceanic island arc should prevent its subduction and that the plate bringing an arc into subduction should be choked and blocked. Our experiments, where arc subduction/accretion occurs during back-arc closure (subduction of the arc plate), confirm this statement insofar that solely the arc crust can be blocked, while the underlying lithospheric mantle keeps subducting. Yet the arc blockage and accretion occur only if the arc crust is thick (> ∼15–20 km thick). This result does not at first glance match the experiments on continental subduction [26,29–31], where even thicker (35 km thick) and weaker continental crust was subducted to ∼100 km equivalent depth. Three main reasons explain the scraping-off of the scaled-to-26-km-thick arc: First, the lithosphere in the arc area is thin and hence the lower crust is hot and weak. Strong decoupling thus takes place at this level. Second, the thick arc has a high isostatic relief and hence resists subduction, being submitted to high pressure while it is driven in the interplate zone. Third, the buoyancy force generated during subduction of the low-density arc crust also opposes this process. The coupling between the arc crust and its lithospheric mantle is only defined by the strength of the lower arc crust, which in the case of a thick arc crust is low. Since the basal surface of the arc is also small, the integral shear coupling force between the arc crust and the lithospheric mantle is low and scraping off is easy. The coupling of the continental crust

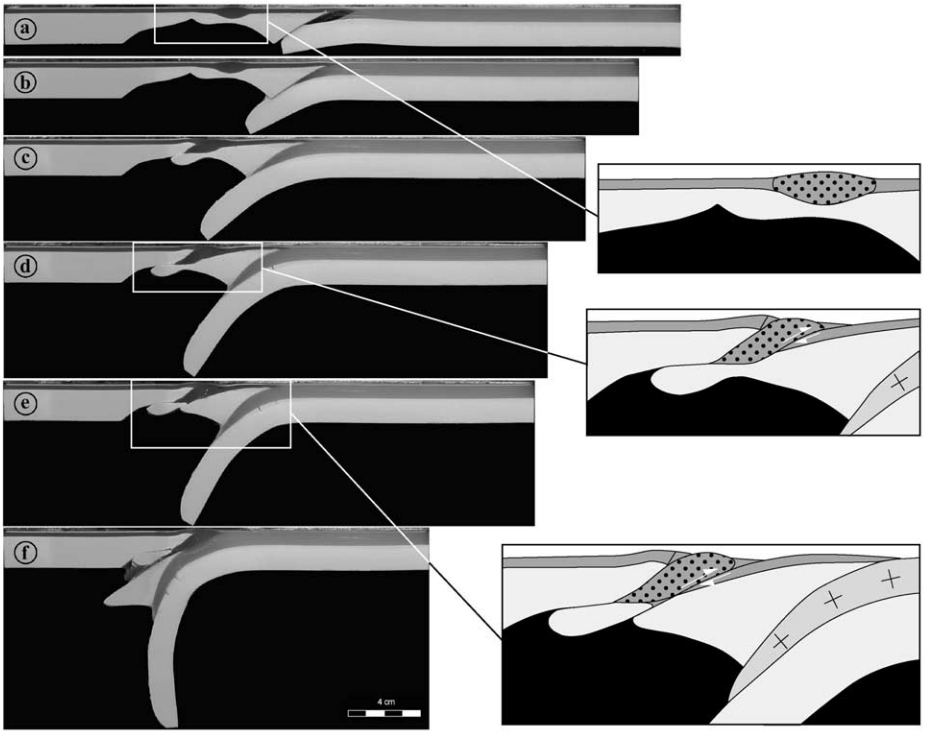

Fig. 8. Successive stages of experiment 5. The arc is thick and strong (Fig. 3e). It subducts to a few tens of kilometres-equivalent depth (panel d), but then it is scraped off (panels e and f) and one-third to one-half of it remains exposed at the surface along with the remnants of back-arc and fore-arc crust.

with its mantle substratum is also weak, but the surface of the crust is very large and hence the integral shear coupling force between the crust and the mantle is high. Accretion of the thick arc crust does not prevent the rest of the arc plate including the fore-arc and part of the back-arc lithosphere, from being subducted. These leave small remnants near the surface. Experiments show (Figs. 5 and 7) that slices/blocks of the young and hot back-arc oceanic crust as well as hot oceanic crust from the arc area (Fig. 6) are more easily scraped and accreted because coupling to the mantle is small compared to the cold fore-arc block. This result is consistent with the geochemical information, which points out that most ophiolites are likely to be derived from arc and back-arc lithospheres [32,33].

Reduction of the arc thickness will obviously facilitate arc subduction. A thin arc is stronger than a thick one (compare Fig. 3b,e) and forces resisting subduction, such as buoyancy and pressure due to the subduction of an elevated topography, are smaller for a thin arc than for a thick one. All this favours complete thin-arc subduction. We cannot tell what is the threshold arc thickness that controls the transition between the two end-member behaviours of the arc because the modelling gives qualitative rather than quantitative results. Moreover, the experiments show that any weak layer (zone) within the arc, whatever the arc thickness, strongly affects the result: the weak layer (interface) favours accretion of the upper (felsic) crust, while the lower (mafic and ultramafic) crust can be subducted.

Our experiments were focussed on intra-oceanic island arcs. On the other hand, the arc can include a fragment of the continental lithosphere separated from the main continent by a back-arc basin (e.g. Japan). In this case, the arc is thicker (> 30 km) and weaker (made on a continental crust basement) than an intra-oceanic arc. Experiments

3 and 5 show that such a 'continental' arc has little or no chance to completely subduct.

The experimental result on which we want to elaborate is that subduction may absorb the whole arc as part of a larger-scale process, which is subduction of the entire arc plate. To know whether plate convergence has entirely swallowed an arc in a particular collision belt, experimental scenarios of the whole arc plate subduction must be tested against the data. For example, we can come back to the Tsangpo Suture, along which there are only faint traces of a within-Tethys Cretaceous island arc [9–11]. The experiments suggest that this volcanic arc, the eastern continuation of the Kohistan and Ladakh arcs of the western Himalayas, is presently nearly absent because it has been subducted below the remaining ophiolites, which corresponds to the fore-arc of the Asian active margin [7,8]. This arc should thus be rather thin, either because it was young or because it has undergone back-arc rifting. Conversely, accretion of the Kohistan and Ladakh arcs corresponds to a thick volcanic arc model. Although arc rifting is documented in the Kohistan [34], we speculate that extension there may have been insufficient to reach the critical thickness under which the whole arc plate would have been subducted. The topic requires discussing many geological and geochronological details, which are outside the scope of this paper. The existence of a second subduction zone at the rear of the back-arc basin (the South Asia active margin) should also be taken into account for the realistic modelling of Kohistan Island arc accretion. This preliminary application only sets the stage of ongoing research that will be presented in a later paper.

Now we demonstrate how the experimental results shed some light on other geological examples. Taiwan is a present-day example of ongoing fore-arc block subduction [27,35], but we do not know any example of present-day arc plate subduction. In the Aegean region the African margin front enters into subduction but the back-arc lithosphere is still under tension [36]. We can predict that the next stage of this arc–continent collision following intra-oceanic subduction and back-arc opening will be the change from extension to compression of the overriding plate and failure of this plate in the back-arc area. The same evolution should occur in Papua New Guinea, where the Australian continental lithosphere is subducting beneath the Huon-Finisterre arc, while the Bismarck Sea is still opening in the back-arc area [37,38].

In the Oman Mountains there are disputed traces of arc activity [39,40]. It has already been argued that the arc plate carrying small volcanic arcs has been subducted in this region in Cretaceous times [13], a conclusion that experiment 2 supports. A scenario of complete arc plate subduction (including the arc itself) is consistent with the available data (see [13] and references therein). Assuming that arcs can disappear in the asthenosphere revives the question on the existence of a volcanic arc before the Adria–Europe collision in the Alps. Most of the authors consider that there never was any [1,2]. However, calc–alkaline volcanism is documented in clastic sediments [41] and produced volcanoes now buried below the Po plain [42,43]. The volcanic belt extends into the eastern Alps [44] and, seen from the experimental perspective, could represent the remnants of a subducted arc.

Traces of pre-collisional setting in older orogenic belts are often scarce and discrete but they find a possible interpretation based on the presented model experiments. For example, the Kudi ophiolites of the western Kunlun Mountains in northwestern China represent the upper mantle and crustal section of a back-arc basin [45]. These rocks show two distinct episodes of arc activity, the second one witnessing interaction with the back-arc basin lithospheric mantle. This event has been interpreted as the closure of the back-arc basin. We conclude that in this region the pre-collisional setting included intra-oceanic subduction with back-arc opening that evolved into back-arc closure with arc plate subduction before continental collision. Other, old sutures such as along the Tilemsi arc of northern Mali [46] expose only the accreted volcanic arc, whereas back-arc and fore-arc rocks are rare. Experiment 3 with arc accretion and subduction of the rest of the arc plate or experiment 1 with only fore-arc block subduction and the arc remaining in its place offer plausible scenarios for the evolution of this area.

7. Conclusion

Subduction of oceanic and continental lithosphere is a key element of plate tectonics. Modern studies are concentrated on thermo-mechanical and mineralogical evolutions of both overriding and subducting plates, their interaction with the surrounding mantle and the geophysical signature of these processes. Our experiments show that the basic subduction scheme can be considerably modified at the beginning of continental subduction when subduction can jump into the back-arc region. During a few million years there will be two active subductions, resulting in complete disappearance of a large lithospheric unit, the arc plate. This process will drastically modify the thermo-mechanical regime of the lithosphere and the underlying mantle [13]. Does this process occur in nature? We can judge this only based on the traces that possible arc plate subduction could leave at the surface. When the mountain belt underwent intra-oceanic subduction and now contains accreted arc, but instead of fore-arc and back-arc lithosphere there are only small remnants of oceanic crust and possibly sedimentary units, the solution consisting of subduction of the arc plate with arc accretion seems to be obvious; at least it is difficult to find a credible alternative. We obtained this process in the physical models when the arc is thick ($> \sim 20$ km thick). A good natural example of this case is the Kohistan arc. Another scenario obtained is complete or almost complete arc subduction corresponding to a thin arc ($<$ 15–20 km thick). This scenario can be applied to the orogens that do not contain an arc. The alternative solution suggested for such cases is that there was no arc at all because the intra-oceanic subduction was too short. This is a possible hypothesis. However, when not only the arc is absent, but also the fore-arc and back-arc, such a solution is no longer valid. The subduction of the whole arc plate is a more plausible scenario, which fits well to the history and structure of the Oman Mountains [13] and possibly the Alps (this paper). Of course, intermediate regimes with partial arc subduction/accretion are also possible. A good example of arc subduction leaving only small accreted remnants is presented by the Indus–Tsangpo Suture in southern Tibet.

Acknowledgements

This work has been supported by the IT 2002 (CNRS/INSU) Programme, the Swiss National Fonds (Project 20-61465.00) and the ETH (Project 0-20884-01). It is UMR Geosciences Azur – CNRS – UNSA contribution No. 576.***[VC]***

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