Lithospheric-scale analogue modelling of collision zones with a pre-existing weak zone

ERNST WILLINGSHOFER¹, DIMITRIOS SOKOUTIS¹ & JEAN-PIERRE BURG²

¹Faculty of Earth and Life Sciences, Vrije Universiteit Amsterdam,
De Boelelaan 1085, 1081 HV Amsterdam, The Netherlands
(e-mail: ernst.willingshofer@falw.vu.nl)
²Geologisches Institut, ETH-Zürich and Universität Zürich,
Sonneggstrasse 5, 8092 Zürich, Switzerland

Abstract: Lithospheric-scale analogue experiments have been conducted to investigate the influence of strength heterogeneities on the distribution and mode of crustal-scale deformation, on the resulting geometry of the deformed area, and on its topographic expression. Strength heterogeneities were incorporated by varying the strength of the crust and upper mantle analogue layers and by implementing a weak plate or part-of-a-plate between two stronger ones. Three (brittle crust/viscous crust/strong viscous upper mantle) and four (brittle crust/viscous crust/brittle upper mantle/strong viscous upper mantle) layer models were confined by a weak silicone layer on one side in order to contain but not oppose lateral extrusion. Experimental results show that relative strength contrast between converging plates and intervening weak plates control the location and the shape of deformation sites taken as ‘collision orogens’. If the contrast is small, internal deformation of the strong plates through fore- and backthrusting occurs early in the deformation history. However, the bulk system is dominated by buckling that nucleates on the weak plate whose antiform topography is highest; model Moho of the bordering stronger plates is deepest under these conditions. If the contrast is large, deformation remains localized within the weak plate for a larger amount of shortening and develops a root zone below a narrow deformation belt, which coincides with the locus of maximum topography. Implementing a buoyant, low-viscosity layer above the model Moho of the weak plate favours the development of asymmetric model orogens notwithstanding the initial symmetric setup. Once the asymmetry is established strain remains localized in thrust faults and ductile shear zones documenting foreland directed displacement of the model orogen. Such laterally and vertically irregular configurations have applications in continent-continent collision settings such as the Eastern Alps. First-order mechanical boundary conditions recognized from modelling to be favourable to the post early Oligocene tectonics of the Eastern Alps include: (1) subtle rather than high-strength contrasts between the Adriatic indentor and the strongly deformed region comprising Penninic and Austroalpine units to the north of it; (2) decoupling of Penninic continental upper crust from its substratum to allow for crustal-scale buckling of the Tauern Window; (3) weak mechanical behaviour of the European lower crust during collision to account for its constant thickness along the TRANSALP deep seismic transect; and (4) the direct continuation of the basal detachment underlying the fold and thrust belt in the hangingwall of the European plate with a wide ductile shear zone in the core of the orogen, which separates the European from the Adriatic plate.

The mechanical properties of the continental lithosphere are non-uniform in space and time (Ranalli 1997). This heterogeneity is primarily due to changes in composition and thermal conditions expressed in the rheological stratification of the lithosphere with the Moho being the most important discontinuity (e.g. Ranally & Murphy 1987). Laterally, the rheology of the continental lithosphere may be modified because of tectonics, leading for example to the separation or collision of continents. Such processes may result in changes of composition (e.g. continental next to oceanic rheology) and lithospheric thicknesses, both in compression as well as extension, and are usually associated with a pronounced thermal perturbation, which influences the strength of the lithosphere transiently. In that way, the thermo-mechanical age of the

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lithosphere is reset, which emphasizes the strong
time-dependence of lithospheric strength
(Cloetingh & Burov 1996). Additionally, the
lithosphere is affected by faulting and shearing
producing a number of metastable rheological
discontinuities that are prone to reactivation
(Ranalli 2000). Subsequent deformation of the
lithosphere will be steered by pre-existing
lateral strength variations (e.g. Ziegler et al.
1998) with relative strength differences among
deforming minerals (Handy 1990), rock layers
(Hudleston & Lan 1993), crustal-scale layers
(Gerbault & Willingshofer 2004), or lithospheric
plates (e.g. Molnar & Tapponnier 1975) as con-
trolling factor in terms of strain distribution,
structure and style of deformation.

Our study focuses on this strength contrast
across plate boundaries. In particular, we investi-
gate differences in the structural evolution of col-
lection zones, their deep structure, the relation-
ship to higher-level deformations and the resultant
topographic expression for conditions of contin-
nental convergence as a function of the relative
strength contrast of the colliding plates. For this
purpose we use a fully mechanical approach,
namely lithospheric-scale analogue modelling,
which is not restricted by the amount of
imposed strain and allows incorporating lateral
material transfer. We subsequently discuss impli-
cations of our modelling results on aspects of the
tectonic evolution of the Eastern Alps in Europe,
from where a wealth of surface and subsurface
data allow constraining the large-scale geometry
of the mountain range as well as its evolution
through time.

Experimental concept: why a weak zone?

When continents collide, the continental crust
thickens through complex deformation processes
involving thrusting and folding. Simplified views
invoke one plate thrust over the other. As a
consequence, the lower plate heats up in
response to thermal re-equilibration and experi-
ences Barrovian-type metamorphism. The
return of the lower plate rocks to the Earth's
surface occurs by a combination of erosion and
tectonic processes. In terms of rheology,
strengthening of the thickened lithosphere may
occur because of the underthrusting of a cold
plate. Conversely, weakening is expected due
to an increased radiogenic heat production in
the thickened continental crust and upward
advective heat transport during exhumation
(e.g. Mancktelow & Grasemann 1997), which
causes an increase of heat flow and hence a shal-
lovery of the brittle-ductile layer transition zone
(see for example Fig. 5 in Willingshofer &
Cloetingh 2003). Orogenic wedges such as the
European Alps or the Pyrenees mainly consist of
a stack of upper crustal slices detached from
the subducting mantle and lower crust (Muñoz
juxtaposes upper crust material next to that of
the lower crust or upper mantle, which are both
stronger for the prevailing temperature and pres-
sure conditions at depth. Furthermore, increased
fluid activity arising from dehydration of water-
saturated sediments during underthrusting
probably also weakens the orogenic wedge (e.g.
Mainprice & Paterson 1984). We argue that the
aforementioned processes result in significant
lateral strength variations during the progressive
evolution of collisional mountain belts in a way
that the bulk strength of the orogenic wedge is
less than that of the bordering undeformed con-
nontnal lithospheres. Such conditions are the start-
ing point for our modelling study, which
elaborates on the role of a weak zone during
subsequent collision and indentation tectonics.

Modelling setup

Materials and initial setups

In our experiments the continental lithosphere
consists of three or four layers representing the
brittle crust, ductile crust, brittle upper mantle
and ductile mantle in nature. Details of the
model setup and the properties of the materials
are given in Figures 1a to c and in Table 1. Lateral
changes of lithospheric strength are incorpo-
rated by laterally varying the thickness of the
brittle and ductile crustal layers and the
rhelogy of the brittle or ductile upper mantle
(Fig. 1). In nature, such strength variations are
expressed in the changing depth to the brittle-
ductile boundary in the crust and in differences
in the shear strength of the upper mantle as
suggested for the present-day lithospheric
strength across the European Central and
Eastern Alps (Okaya et al. 1996; Willingshofer
& Cloetingh 2003).

Dry quartz and feldspar sands, both Mohr-
Colomb-type materials (Table 1), were used as
rock analogues for the brittle crust and the
brittle upper mantle (Fig. 2a-c). Non-Newtonian
viscous layers representing the viscous crust
(silicone mix I in Fig. 1a-c) and the viscous
upper mantle (silicone mix I for the weak zone
and silicone mix II for the foreland and indenter
plates) are mixtures of Rhodorsile Gomme-
type or PDMS-type (material properties in
Weijermars 1986) silicone with barite powder
(Table 1). Additionally, we used a mixture of
Rhodorsile Gomme-type silicone with Acid Oil
Fig. 1. Initial setup for experiments A, B, and C. Location of the corresponding strength profiles are marked with A and B in the sketches of the model configuration. Strength profiles have been calculated using a failure criterion for brittle deformation in the form $(\sigma_1 - \sigma_2) = 2[c + \mu \rho g z (1 - \lambda)]/(\mu^2 + 1)^{1/2} - \mu$ where $(\sigma_1 - \sigma_2)$ is the differential stress, $c$ the cohesion, $\mu$ the coefficient of friction, $\rho g z$ the overburden pressure and $\lambda$ the ratio of water to overburden pressure (0 in our modelling); and for ductile deformation in the form $(\sigma_1 - \sigma_2) = \eta \gamma$ where $(\sigma_1 - \sigma_2)$ is the differential stress, $\eta$ the viscosity, and $\gamma$ the experimental shear strain rate. Material parameters are listed in Table 1. The black arrow indicates the direction of push. LVL, low-viscosity layer in experiment C.
Table 1. Mechanical properties of analogue materials used in this study

<table>
<thead>
<tr>
<th>Layer</th>
<th>Analogue material</th>
<th>Density, (\rho) (kg m(^{-3}))</th>
<th>Coefficient of friction*, (\mu)</th>
<th>Viscosity(^{+}), (\eta) (Pa s)</th>
<th>Power(^{\dagger})</th>
<th>Models</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brittle crust</td>
<td>Dry feldspar sand</td>
<td>1300</td>
<td>0.75</td>
<td>1.8 (\times 10^{4})</td>
<td>1.8</td>
<td>B</td>
</tr>
<tr>
<td></td>
<td>Dry quartz sand</td>
<td>1510</td>
<td>0.9</td>
<td>1.8 (\times 10^{4})</td>
<td>2.2</td>
<td>A, C, C</td>
</tr>
<tr>
<td>Ductile crust</td>
<td>Silicone mix I</td>
<td>1520</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>LVL, silicone mix</td>
<td>1100</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brittle upper mantle</td>
<td>Dry quartz sand</td>
<td>1510</td>
<td>0.9</td>
<td>1.8 (\times 10^{4})</td>
<td>1.8</td>
<td>A, B, C</td>
</tr>
<tr>
<td>Viscous upper mantle</td>
<td>Silicone mix I</td>
<td>1520</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Silicone mix II</td>
<td>1540</td>
<td></td>
<td>7.2 (\times 10^{5})</td>
<td>2.0</td>
<td>A, B, C</td>
</tr>
<tr>
<td>Asthenosphere</td>
<td>Sodium polytungstate solution mixed with glycerol</td>
<td>1800</td>
<td></td>
<td>1.2</td>
<td></td>
<td>A, B, C</td>
</tr>
</tbody>
</table>

*The coefficient of friction has been inferred with a Hubbert-type shear apparatus (e.g. Krantz 1991). Note that sand densities and \(\mu\) values have been deduced for sieved sand layers. The cohesion of the brittle materials, which is actually the resistance of the sand grains to sliding, is 40–70 Pa.

\(^{+}\)The effective viscosity of the silicone mixtures for laboratory strain rates (~1 \(\times 10^{-6}\) s\(^{-1}\)) has been determined with a coni-cylindrical viscometer under room temperature (21 ± 1°C).

as a low-viscosity layer (LVL in Fig. 1c) above the model Moho. The above named layers are floating on a dense, low-viscosity fluid simulating the asthenosphere. Mantle and crustal layers were laid down one by one atop the model asthenosphere. The uppermost layer, the brittle upper crust, has been added by sprinkling layers of coloured sand, including a 4 x 4 cm grid on the surface to facilitate strain analysis. The models were allowed to re-equilibrate isostatically for a time period of c. 16 hours.

The experiments discussed in this paper have been conducted within a 50 x 40 x 15 cm plexiglass tank. Glass walls attached to the long sides of the box reduced friction. Solid walls confined three sides of the model while the fourth, a layer of silicone putty labelled ‘Lateral Confinement’ (Fig. 1), allowed limited amounts of lateral escape. At variance with Ratschbacher et al. (1991a), we deliberately chose a deformable, yet confined lateral boundary to avoid a priori facilitated lateral extrusion. Experiment A is the reference model. During modelling, the surfaces of all experiments have been scanned with a three-dimensional video-laser every 30 min. The gathered data have then been converted to digital elevation models (DEM, Figs 2 to 4), which are crucial for the understanding of the relationships between evolving structures and topography.

The strength of the model indenter and foreland plates is the main difference between experiments A and B (Fig. 1). Experiment B contains a brittle upper mantle, which, compared to experiment A, increases the strength difference between the central weak zone and the bordering strong plates. The rheological stratification of the weak zone is the main difference between experiments A and C (Fig. 1). The buoyant low-viscosity layer placed above the model C Moho simulates a decoupling horizon between the crust and the upper mantle and aspires to explore the influence of a deep-seated rheological weakness such as partial melts or zones of high fluid activity on the structural response of the shallower crust.

Scaling procedure

Geometric and dynamic similarity is a prerequisite for a valid comparison between analogue experiments and natural prototype (e.g. Weijermars & Schmeling 1986). The experiments are geometrically scaled by applying a length ratio of 5 \(\times 10^{-7}\), equating 1 cm in the models to 20 km in nature. For both the foreland and indentor plates we assumed that the brittle crust has twice the thickness of the ductile crust, a configuration that embodies a strong lithosphere of pre-Mesozoic thero-tectonic age (Cloetingh & Burov 1996). This thickness ratio has been reversed for the pre-existing strong zone simulating young, thermally reset Alpine-type crust (e.g. Genser et al. 1996).

The dimensionless stress-scale factor for the brittle Mohr–Coulomb-type materials of grain size 100–300 \(\mu\)m used in the models is determined by the relationship

\[ \sigma^* = \rho^* g^* \ell^* \]

where \(\sigma^*\) stands for model-prototype ratios of stress, density, gravitational acceleration and thickness, respectively (e.g. \(\sigma^* = \sigma_{\text{model}} / \sigma_{\text{prototype}}\)). All experiments have been performed under normal gravity conditions, hence \(g^*\) equals 1.
\[ R_m = \frac{\rho gh^2}{\eta V} \]

and \( \rho \) and \( h \) are the density and thickness of the ductile layer, respectively, \( g \) is the acceleration of gravity, \( \eta \) the viscosity of the ductile layer and \( V \) the compression rate. To fulfill the criterion of dynamic similarity, \( R_m \) of the experiment and the prototype must be approximately the same. A summary of the scaling parameters is presented in Table 2.

### Simplifications

Analogue models simplify the complexity of nature such that only the first-order rheological stratification can be accounted for. Although ductile materials with power law rheology were incorporated to simulate creep processes, they maintained their properties throughout the thickness of the layer and hence temperature-dependent variations of the creep strength with depth were not integrated. We attempted to translate temperature and thickness variation effects into implicit mechanics. For this purpose, we (1) assigned the same rheology to the ductile crust and ductile upper mantle within the weak zone, and (2) only considered that part of the lithosphere that has significant strength. As such, the mantle layers in the experiments are thinner than in nature and lateral thickness variations possibly due to previous thickening events, which might reduce the bulk strength of the lithosphere as argued in the ‘experimental concept’ section, are neglected. By assuming the same initial thickness for all lithosphere layers in the models, gravitational effects in relation to pre-existing mass anomalies are not taken into account as large error bars are attached to the quantity of such anomalies (crustal and lithosphere thickness) for the geologic past.

The models also ignored natural recovery processes such as erosion and re-sedimentation and possible influences stemming from earlier subduction of oceanic lithosphere. In our experiments, the weak zone/strong plate boundaries are vertical, which is not likely to occur in nature at the scale of the lithosphere. However, we are confident that this simplification does not bias the main results derived from lateral strength gradients. For a concise discussion on the influence of different initial geometries of the weak zone on collision and indentation tectonics the reader is referred to Willingshofer et al. (2004).

### Modelling results

The following sections summarize and display the modelling results in chronological order. The terms ‘thrust’ and ‘backthrust’ refer to thrusting in or against the direction of push from the moving wall, respectively. The plate in contact with the moving wall is the ‘indentor’ and the distal plate is the ‘foreland’.

### Experiment A

**Surface deformation and evolution of topography.** The first structures were thrusts and backthrusts that formed simultaneously (Fig. 2a). These faults cut the model surface at or close to the boundaries between the strong and weak plates, and are clearly visible as linear features in the digital elevation model (DEM). Surface uplift of the weak plate is small and affects all the area between the thrust faults.

As convergence continued, thrusting propagated outwards with respect to the central weak plate (Fig. 2b). A sequence of backthrusts developed within the brittle crust of the indentor. After 10% bulk-shortening, thrusting flipped to the foreland side of the weak plate activating thrust 5 in Figure 2b. The weak plate folded while thrusting propagated away into the indentor and

<table>
<thead>
<tr>
<th>Layer</th>
<th>Density, ( \rho ) (kg m(^{-3}))</th>
<th>Viscosity, ( \eta ) (Pa s)</th>
<th>Layer thickness, ( h ) (m)</th>
<th>Velocity, ( V ) (m s(^{-1}))</th>
<th>( R_m )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ductile crust model</td>
<td>1520</td>
<td>( 1.8 \times 10^5 )</td>
<td>( 5 \times 10^{-4} )</td>
<td>( 1.9 \times 10^{-6} )</td>
<td>1.1</td>
</tr>
<tr>
<td>Lower crust nature</td>
<td>2900</td>
<td>( \sim 10^{22} )</td>
<td>( 1 \times 10^2 )</td>
<td>( 1.6 \times 10^{-10} )</td>
<td>1.7</td>
</tr>
<tr>
<td>Ductile mantle model</td>
<td>1540</td>
<td>( 7.2 \times 10^5 )</td>
<td>( 1 \times 10^{-3} )</td>
<td>( 1.9 \times 10^{-6} )</td>
<td>1.2</td>
</tr>
<tr>
<td>Ductile mantle nature</td>
<td>3300</td>
<td>( 10^{21-22} )</td>
<td>( 2 \times 10^3 )</td>
<td>( 1.6 \times 10^{-10} )</td>
<td>8.1</td>
</tr>
</tbody>
</table>
Fig. 2. Modelling results of experiment A for different stages of deformation are displayed as: (a–d) topview images (left panel), digital elevation models (DEM) (right upper panel), line drawings showing the structural interpretation (right lower panel), and (e, f) cross-sections. The green rectangle in (a) indicates the area extent of the DEM. The numbers in the figures indicate the sequence of the evolving structures. Black arrows in (a–f) indicate the direction of push. Note that the model surface corresponds to the uppermost coloured sand layer.
foreland plates. Consequently, the uplifted weak plate had an antiformal surface flanked on both sides by depressions (DEM, Fig. 2c, d). The foreland depression was weakly developed and originated from thrust loading. The indenter depression spanned approximately the width of the backthrust system and formed a syncline followed by an anticycle close to the moving wall. Noteworthy at this stage of convergence is the development of small piggy-back basins in the hangingwall of thrusts 1 (Fig. 2b). Such thrust-parallel basins would capture detritus derived from the raising antiformal weak plate.

After 20% bulk-shortening, outward thrust propagation stopped and convergence was taken up by out-of-sequence thrusting along thrusts 1 and smaller-scale backthrusts (9) within the weak plate (Fig. 2c). Coeval folding continued and led to surface uplift on the weak plate and subsidence in the backthrust region (DEM, Fig. 2c). Little distortion of the surface grid indicates that lateral escape was then minor.

Until the end of the experiment, shortening was mainly taken up by out-of-sequence thrusting (Fig. 2d). The topographic low at the weak plate/indenter boundary coincided with a trench in front of the main backthrust of the model orogen (DEM, Fig. 2d). This backthrust carried material over the earlier backthrusts. As a consequence, no noticeable change of surface altitude occurred within the weak plate during the last few centimetres of shortening. Minor strike-slip faulting along the weak plate–indenter boundary is a boundary effect impelled by friction along the sidewall and hence contains no implication for geological interpretation. Dextral strike-slip faulting on the lateral confinement side was related to the low resistant lateral material. The final widths of the weak and strong plates, deduced from measurements at the model surface, indicate that both strong plates have been shortened by about 45% on average. The weak zone was shortened by 17% and underwent stretching orthogonal to the shortening direction by 10%. The deformed surface grid reveals that stretching was unevenly distributed along strike with minimum stretch close to the glass wall (7%) and maximum stretch close to the lateral deformable confinement (19%).

Cross-sections. Cross-sections of experiment A document the finite state of shortening, which has arisen from interplay of thrusting and folding (Fig. 2e, f). The upper weak plate is an anticycle with a thick core of ductile crust and upper mantle. Thickening of the weak plate is due to homogeneous thickening of its ductile layers (about 90%). The anticline is slightly asymmetric, exhibiting an overturned limb facing the indenter. Thrusting led to considerable thickening of the brittle crust, especially in the backthrust area where initial thickness has doubled (Fig. 2e). These backthrusts rotated towards steeper dips to accommodate folding of the indenter. In contrast, the brittle layer of the weak plate was not faulted. Shortening was taken up within the lower crusts of the foreland and indenter plates by homogeneous thickening, which is largest (c. 25%) close to the back wall and the moving wall and least (c. 13%) at the boundary with the weak plate.

Along strike differences in the model orogen reflect differences in the strength of the confining material (glass wall versus deformable silicone putty). Shortening of the indenter close to the rigid glass wall can only be taken up by folding and backthrusting. These structures are less developed close to the lateral confinement that allowed the model to escape laterally.

Experiment B

Surface deformation and evolution of topology. A backthrust appeared at the surface of experiment B at the weak plate/indenter boundary after 1.6% bulk-shortening (Fig. 3a). Thereafter, thrusting mainly took place within the weak plate (Fig. 3b). Imbrication in its brittle layer is associated with relief growth characterized by model orogen-parallel ridges and troughs. This structurally controlled landscape is distinctly different from the smooth dome-shaped topography of experiment A (compare DEM of Fig. 2b, c and Fig. 3b, c). Deformation essentially remained within the weak plate up to 15% bulk shortening and was taken up by thrusting and backthrusting. Dextral strike-slip faults close to the lateral confinement are accommodation structures formed in response to thrust tectonics (Fig. 3c). Subsequently, convergence was accommodated by conjugate thrusts defining pop-up structures away from the weak plate and by backthrusting along fault 1 (Fig. 3c). Thrusting occurred dominantly close to both the back- and advancing walls. A pronounced increase of topography over the weak plate suggests that thickening of the ductile layers controlled the topographic evolution at this stage (DEMs of Fig. 3b, c).

At the end of the experiment, shortening was distributed throughout the model, as shown by the distribution of active thrusts and backthrusts (Fig. 3d). Folding of the indenter had a lower amplitude than in experiment A, which reflects higher strength (strength profiles of Fig. 1a
Fig. 3. Modelling results of experiment B for different stages of deformation. (a–d) topview images (left panel), digital elevation models (DEM) (right upper panel), line drawings showing the structural interpretation (right lower panel), and (e, f) cross-sections. The green rectangle in (a) indicates the area extent of the DEM. The numbers in the figures indicate the sequence of the evolving structures. Black arrows in (a–f) indicate the direction of push. Note that the model surface corresponds to the uppermost coloured sand layer. Symbols as in Fig. 2.
and b). For the same reason, loading of the indenter by the model orogen along backthrust 1 yielded a shallower trench. At the end of the deformation history, the weak zone was shortened by 44% and stretched by 9% parallel to the strike of the model orogen. Compared to experiment A, more shortening was taken up by deformation of the weak zone and less (28%) by the adjacent strong plates.

**Cross-sections.** Cross-sections of experiment B show a doubly vergent orogen and gentle folding of the adjacent stronger lithospheres (Fig. 3e, f). The brittle crust of the weak plate is truncated by several thrusts and backthrusts. Repeated activation yielded a complex pattern of partly interfering reverse faults. In both the foreland and indenter plates, thrusting within the brittle layers took place away from the weak plate, in response to stress concentration close to the distal wall and the moving wall. The initially parallelepipedic ductile layers of the weak plate were converted into elongate bodies with up to six times the original thickness due to intense horizontal shortening (Fig. 3e). The ductile crust of both the foreland and indenter plates has been thickened by 8% on average. Besides being folded with the rest of the model (Fig. 3e, f), the brittle upper mantle shows little signs of deformation but strongest thickening in the hinge zones. In section A (Fig. 3e), one single thrust developed in the foreland plate close to the contact with the weak plate. The fault plane, which lines up with the top surface of the ductile mantle (white layer) is oriented at a high angle to the main compression direction, suggesting that thrusting pre-dated folding. Homogeneous thickening of the ductile mantle material absorbed shortening of the weak plate. The ductile mantles of the foreland and indenter plates show long-wavelength folds. Thickening was highest in the fold hinges near the weak plate.

**Experiment C**

**Surface deformation and evolution of topography.** The formation of a pop-up uplift on the weak plate started soon after shortening was imposed (Fig. 4a). Surface uplift was due partly to active faulting and partly to the buoyancy of the low-viscosity layer. As shortening continued, a backthrust at the weak plate—indenter boundary became active and the weak plate got folded with the former pop-up structure lying as a tectonic klippe in a syncline between two anticlines (Fig. 4b). At the same time, material was expelled from the weak plate towards the lateral confinement, guided by a set of strike-slip faults. The evolving topography was distinctly asymmetric with the anticline close to the indenter being highest.

Dominant foreland-directed thrusting initiated at 11% bulk shortening along thrust 5, which remained active until the end of the experiment (Fig. 4c, d). Thrusting of the model orogen over the foreland plate induced flexure and hence a foreland-type basin parallel to the frontal thrust 5 (DEM, Fig. 4c). In contrast, no similar feature was observed on the indenter side. The raising topography of the weak plate cannot solely be related to thrusting because shortening also tightened folds, with normal faulting occurring along anticline crests (Fig. 4c). Such structures demonstrate internal deformation of the model orogen during thrusting over the foreland. The foreland-facing anticline became overturned and transformed into a recumbent fold-nappe during the final phases of shortening. A mature foreland basin developed while thrusting continued to load the foreland plate (DEM, Fig. 4d). The finite model configuration documents 58% shortening and 20% orogen-parallel shortening of the weak zone. Unlike the above-described experiments where shortening was distributed rather evenly over the foreland and indenter plates, distinctly more of the convergence was accommodated in experiment C on the foreland plate site (55%) through long-lasting activation of thrust 5. Shortening of the indenter plate was about 16%.

**Cross-sections.** Cross-sections of experiment C (Fig. 4e, f) show upright and overturned folds in the weak plate. Thrusts and backthrusts cut the brittle layer at inflection points of the folds and normal faults dissect the crest of upright anticlines. A major thrust (no. 5) separates the weak plate from the underthrust foreland plate bent by the load of the model orogen. This thrust links up with a broad ductile shear zone within the weak zone, which separates the foreland from the indenter plates. The brittle layers of the foreland and indenter plates are free of faulting, close to the moving wall excepted. The buoyant low-velocity layer remained trapped at depth. Part of it accumulated beneath the strong ductile mantle of the indenter and the other part was expelled towards the weak lateral confinement. The fact that the indenter only shows a small inclination towards the weak plate reflects the combined effects of the buoyancy of the accumulated low-viscosity layer underneath the indenter and the small loading by the model orogen. This former effect was probably decisive for the switch to dominant foreland-directed
Fig. 4. Modelling results of experiment C for different stages of deformation. (a–d) topview images (left panel), digital elevation models (DEMs) (right upper panel), line drawings showing the structural interpretation (right lower panel), and (e, f) cross-sections. The green rectangle in (a) indicates the area extent of the DEM. The numbers in the figures indicate the sequence of the evolving structures. Black arrows in (a–f) indicate the direction of push. Note that the model surface corresponds to the uppermost coloured sand layer. Symbols as in Fig. 2.
thrusting at 11% bulk-shortening. Short-wavelength folding accommodated shortening of the ductile mantle below the low-viscosity layer. The ductile crust and upper mantle of both the foreland and indenteror plates thickened increasingly from the boundaries with the weak plate (0–5%) towards the confining walls (<15%). The laterally escaped part of the weak plate spread into the lateral confinement and, therefore, had thinner layers and slightly lower topography.

Summary and general discussion

of modelling results

In this section we highlight similarities as well as differences of the models and discuss them in the frame of previously published modelling studies.

(1) Our modelling results are consistent with previous numerical and analogue models (e.g. Cloetingh et al. 1999 and references therein; Davy & Cobbold 1991; Martinod & Davy 1994; Burg et al. 1994) and show that folding is a primary response to shortening of the continental lithosphere. The above quoted authors have shown that homogeneous lithosphere (without strength variations) starts to fold soon after compression commences. Modelling with a weak zone within the lithosphere deviates on the following points. (a) Folding is suppressed in case of high-strength contrasts between weak and strong plates (experiment B). In such a situation, shortening is accommodated by thickening of the weak zone for a long period of time and folding of the indenter and to a lesser extent the foreland plate are late-stage features. (b) The presence of a mechanical instability within the weak zone (our buoyant low-viscosity layer in experiment C) controls the locus of folding through strain localization such that after about 10% bulk-shortening most of the remaining convergence was taken up along thrust number 5, which is interpreted to continue towards the centre of the model orogen as a ductile shear zone (Fig. 4e, f). Consequently, folding was confined to the weak zone and underthrusting of the foreland plate was the more energetic response to compression.

(2) Indentation tectonics requires strength differences between colliding plates. Strength differences are manifested by stronger deformation in the weaker plate than in the stronger plate. Depending on the regional mechanical boundary conditions, indentation causes thickening and/or activation of large strike-slip faults guiding lateral escape of crustal blocks. The stronger plate becomes deformed when thickening can no longer absorb shortening. Such an evolution has been described from various natural examples like the Anatolian region (McKenzie 1972), Asia (Molnar & Tappinon 1975) and the European Alps (Ratschbacher et al. 1991a, b). Indentation tectonics have been approached experimentally by using rigid indentors (Tappinon et al. 1982; Davy & Cobbold 1988; Ratschbacher et al. 1991a; Martinod et al. 2000; Rosenberg et al. 2004), hence forcing deformation to take place within the deformable plate. Previous work focused on lateral extrusion and deformation in front of the indenter, which are both strongly controlled by the degree of lateral confinement. Although our experiments have not been designed to specifically study this process, comparison of experiments A and B (Figs 2 and 3) with C (Fig. 4) confirms the findings of earlier works that the degree of confinement is important for extrusion tectonics. In our work the deformable indenter and foreland plates had the same properties and were subject to lithospheric-scale folding (experiments A and B), bending due to loading (experiment C), and thrusting in the brittle layers. We infer that a strong lateral confinement favoured folding of the indenter (Figs 2 and 3), whereas a weak confinement enhanced lateral material transport towards the deformable confinement (Fig. 4). The amount of orogen-parallel extension, deduced from the deformed surface grid, was in all experiments least close to the glass wall and highest in proximity to the deformable confinement. It is worth noting that shortening of the model with the higher strength contrast (experiment B) did not increase the amount of orogen-parallel extension (9%), compared to experiment A (10%). Instead, shortening resulted in a narrower and thicker weak zone.

(3) The relative strength contrast between the weak plate and the adjacent indenter and foreland plates appears to determine the sequence of deformation. High strength contrasts (experiment B) concentrated deformation within the weak plate, while low strength contrasts favoured deformation within both the indenter and
foreland plates (experiment A). These results are consistent with models that employed indenter and indented plates of equal strength (Davy & Cobbold 1988) and models using a strong, cratonic-type, lithosphere as indenter (Keep 2000).

(4) In the absence of any inherited asymmetry from preceding subduction, we ascribe the model deep structures to intracratonic processes. We infer that a low strength contrast (experiment A) between the weak and strong plates leads to buckling dominated deformation in the weak plate and, to a lesser amount in the adjacent strong plates. As a result, the topographic maximum located within the weak zone does not correspond to the deepest Moho, which is the trough syncline of the indenter. High strength contrasts (experiments B) favour the formation of crustal and lithospheric roots. In this model the topographic maximum correlates well with the locus of the crustal root. A distinctly asymmetric structure of the model orogen is only observed where a mechanical instability above the Moho within the weak zone was initially present (experiment C). Once the shape of this low density instability has become asymmetric due to the imposed kinematic boundary conditions (push from one side), it facilitated strain localization over long periods of time. As a consequence, most of the shortening was taken up by a thrust at the base and a thick ductile shear zone within the central zone of the model orogen (Fig. 4e, f). Analogue plates accommodated shortening in the ductile crust by homogeneous thickening (Models A, B, and C) thereby largely maintaining its initial geometry. These results are consistent with those of a numerical modelling study focusing on the role of lower crust rheology during collision tectonics (Gerhaut & Willingshofer 2004). The same authors suggest that a strength difference of about 20 MPa between the lower crust of the weak zone and the indenter is sufficient to favour lower crust indentation above lateral (opposite to the direction of shortening) ductile flow. In our experiments, indentation occurs on the scale of the lithosphere and interference patterns with the weak zone are simple. Conditions favourable for lower crust indentation as envisaged in the European central Alps demand decoupling at upper crust/lower crust and at Moho levels (e.g. Pfiffner et al. 2000).

Discussion of modelling results in the light of post-early Oligocene tectonics of the Eastern Alps

In the following sections we discuss our modelling results in the frame of the late Oligocene to present geodynamic evolution of the Eastern Alps. At this point we like to emphasize that our experiments have not been designed to reproduce the deformation history of the Eastern Alps in great detail. Instead, we have chosen a conceptual approach, which restricts us to the discussion of first-order aspects relevant to tectonics of the Eastern Alps and other orogens in which a weak zone has been compressed. As such, we do not present a 'best fit model', but aim at deducing favourable mechanical conditions for Late Oligocene to Neogene Eastern Alps tectonics from experiments described in the previous sections.

The Eastern Alps as geological example

The European Alps are among the youngest features of the European lithosphere and owe their origin to continent–continent collision related to the Africa–Europe convergence (e.g. Frisch 1979; Dewey et al. 1989; Schmid et al. 1996). We regard the orogenic wedge of the Eastern Alps, which existed during the early Oligocene (c. 30 Ma ago) between the Adriatic plate to the south of the Periadriatic Line and the European foreland in the north (Fig. 5) (e.g. Frisch et al. 1998), as a pre-existing weak plate or part-of-plate based on the following arguments. (1) The strong concentration of post-Eocene deformation, which affected Austroalpine and Penninic units (Fig. 5), in particular in the Tauern Window region, may be envisaged as expressing significant lateral stress variations, which controlled the locus of deformation (see also Rutschbacher et al. 1991a, b). (2) Cloetingh & Burov (1996) have shown that the strength of the European continental lithosphere strongly depends on the time elapsed since it was subject to loading and tectonic activity. Adopting this concept, it follows that the tectonic units comprising the orogenic wedge, which underwent thermal rejuvenation and loading during the Late Cretaceous (Austroalpine unit) and Eocene–Miocene (Penninic units), are weaker than the European and Adriatic plates, which have been affected by Variscan tectonics (e.g. Frey et al. 1999). (3) The radiogenic heat production, hence thermally induced softening within the orogenic wedge, which dominantly consists of continental slices (Austroalpine and
Zentralgneiss units) with minor oceanic units, was probably higher than in the adjacent plates of normal crustal thickness. (4) Alpine ductile fabrics are related to Eocene–Miocene tectonics and are confined to the Tauern Window and a small strip of Austroalpine units flanking the Tauern Window to the south. The understanding of certain aspects of Eastern Alps tectonics, therefore, may justifiably refer to first-order features identified in the analogue models. We perceive relevance on several points, namely lateral extrusion and coeval orogen-parallel extension, backthrusting on Adria, buckling in the Tauern region, and interpretation of the deep seismic structures.

**Indentation and lateral escape**

Late Oligocene to middle Miocene lateral extrusion in the Eastern Alps is interpreted to reflect the combined effects of horizontal forces related to indentation by the Adriatic plate and gravitational forces stemming from preceding crustal thickening in the Tauern Window region (Ratschbacher et al. 1991a and references therein). Eastward lateral extrusion, towards the Pannonian Carpathian region, was conditioned by contemporaneous subduction and slab retreat along the Carpathian arcs providing space for the extruding blocks (Royden 1988). Lateral extrusion ceased during the early late Miocene when subduction along the Carpathians terminated and the incoming buoyant European plate jammed the subduction process (Sandulescu 1988). Until that time, deformation of the Adriatic indenter was restricted to its northern margin, which became involved in extrusion tectonics (Frisch et al. 1998) and at the transition zone to the Dinarides (Castellarin & Cantelli 2000). Internal deformation of the Adriatic indenter commenced during the Late Middle Miocene through activation of the south vergent Valsugana thrust system, which documents southerly directed backthrusting (Castellarin & Cantelli 2000). Subsequent deformation events portray south-directed thrust propagation with active present-day thrusting in the Po Plain (Benedetti et al. 2000). In summary, significant deformation of the Adriatic indenter interior, amounting to about 50 km
north-south shortening (Schönborn 1999), post-dates the climax of lateral extrusion in the Eastern Alps north of the Periadriatic Line and subduction along the Carpathian arc.

In our study, only experiment B with the largest strength contrasts shows significant deformation of the indenter proper after about 15% bulk-shortening. This jump in deformation site takes place once the indented plate has been thickened to the point that vertical stress components hamper further thickening of this plate. As to whether the increased resistance at the lateral confinement contributes to activating deformation within inner indenter regions cannot be deduced with certainty from our experiments. The experiment with a low strength contrast (experiment A) records simultaneous deformation of the weak zone and the indenter more or less throughout the experiment, and hence exhibits little resemblance with the Eastern Alps. Strong strain localization over a long period of time (experiment C) also retards indenter deformation. From our experiments, we deduce that strength contrasts between the weak zone and the indenter are important to keep deformation localized in a particular region over a long period of time. At the same time, one needs to bear in mind that our simple geometric and kinematic model setup neglects the influence of other processes, such as subduction along the Carpathians, which is intrinsically related to the opening of the Pannonian Basin (e.g. Horváth 1993), and probably facilitated lateral extrusion in the Alps through slab pull forces, thereby reducing the demand for high strength contrasts between indenter and indented region. Model B therefore most likely exaggerates natural conditions, for which more subtle strength variations have been suggested at least for the present-day situation across the Eastern Alps (Willingshofer & Cloetingh 2003). Geometric boundary conditions adopted in this study can only account for about 20% bulk orogen parallel extension within the weak zone. Extension is not even distributed, but increases towards the deformable weak zone. Both the total amount of Oligocene to Miocene extension, which was in the order of 50% (Ratschbacher et al. 1991b; Frisch et al. 1998), and the locus of maximum extension, which was in front of the indenter dip (Frisch et al. 2000) differ from our modelling results. While the total amount of extension can be accounted for by a weaker or no lateral confinement (Ratschbacher et al. 1991a), the locus of extension might be controlled by the indenter geometry as recently suggested by analogue modelling by Rosenberg et al. (2004).

The structure of the Tauern Window

The Tauern Window is an elongate dome formed during cooling after the thermal peak at 30 Ma (Lammerer & Weger 1998 and references therein). Its internal structure is characterized by short-wavelength (several km), tight, upright folds (Fig. 6a) with NE-SW trending fold axis in the west and NW-SE trending fold axes in the east. Fold limbs are associated with sinistral shear zones in the western and dextral shear zones in the eastern Tauern Window (e.g. Behrmann & Frisch 1990; Kurz & Neubauer 1996). The western and eastern terminations of the Tauern Window against the overlying Austroalpine units are low-angle shear zones documenting normal displacement of hanging-wall units (Selverstone 1988; Genser & Neubauer 1989). Structures within and close to the Tauern Window are consistent with models employing an Oligocene-Miocene transpressional stress regime (e.g. Lammerer 1988). It has been argued, against earlier work, that the window shows similarities with extensional core complexes (e.g. Selverstone 1988) genetically linked with lateral extrusion tectonics (Frisch et al. 2000). In models A and C high-amplitude and short-wavelength buckle folds formed within the weak zone and controlled surface uplift. While buckling in model A occurred on the scale of the lithosphere, buckling in model C was confined to the crust above the buoyant low-viscosity layer, which allowed for full crust-mantle decoupling within the weak zone. Our results support the findings of Lammerer & Weger (1998), who emphasized the importance of crustal-scale buckling for the structural evolution of the western Tauern Window. Unlike the amount and spatial distribution of orogen-parallel extension accompanying folding (see previous section), buckling in front of the indenter appears to be a process independent of the indenter geometry and direction of shortening. Evidence stems from comparing our results with a straight indenter with those of Ratschbacher et al. (1991a) or Rosenberg et al. (2004), who used triangular indenters. Those works showed that contemporaneous shearing in strike-slip mode critically depends on indenter geometry and convergence direction. The strong lateral confinement in our models also suggests that at least part of the orogen-parallel component of stretch is related to the folding process itself.

Our models show that a crust weaker than its surroundings was essential in the development of high-amplitude buckle folds. Inversely, if such folds are diagnostic of weak crusts
compressed between stronger plates, the deep Tauern Window has to be treated separately in mechanical terms from either European and Adria plates to account for initial plate heterogeneity. In this respect, numerical modelling suggests that the Tauern Window region was distinctly weaker than the surrounding areas while folding and lateral escape took place (Genser et al. 1996). In accordance with the structure of the Tauern Window, we consider crustal-scale buckling as documented in model C as important processes in collisional settings described above.

The deep structure of the Eastern Alps

Similarities in structures and evolution between the eastern Alps and the models are mechanical references to interpret deep seismic profiles. Key features of the deep structure of the TRANSALP transect (Fig. 6; TRANSALP Working Group 2002) include (1) a south-dipping European lower crust of rather constant thickness, (2) a distinctly thicker Adriatic lower crust, which dips to the north, and (3) crustal-scale ramps such as the Sub-Tauern Ramp carrying Adriatic lower and upper crusts on to the underthrusting European crust and the Sub-Dolomite Ramp, which acted as a backthrust during imbrication of the Adriatic crust. The Moho of both the European and Adriatic plates dips towards the mountain root. All of our experiments resulted in lower crust geometries of the foreland plate similar to that of the European lower crust (TRANSALP Working Group, 2002), experiment C being most resembling (compare cross-sections in Figs 2, 3, and 4 with Fig. 6). In the Central Alps, geophysical evidence suggests that the European lower crust is a strong layer of approximately constant thickness (Pfiffner et al. 2000 and references therein). Analogous plates accommodated shortening by homogeneous thickening (Models A, B, and C) thereby largely maintaining initial overall geometry.

Accordingly, constant thickness does not imply that a layer is not deformed and that it has a high strength. Models A and B, where folding affected the indentor, result in Moho dipping towards the orogen as published by the TRANSALP Working Group (2002), whereas Model C shows a flat indentor Moho consistent with the Moho shape identified from receiver functions utilizing teleseismic events (Kummerow et al. 2004). The flat Moho in experiment C is a consequence of strain localization along the basal thrust, which places the model orogen over the foreland plate. This thrust (5 in Fig. 5), which can be interpreted as equivalent to the basal detachment of the northern fold and thrust belt in the Eastern Alps, joins a wide shear zone separating the foreland from the
indentor plates. Both the thrust and the shear zone remained active during the second half of the experiment and took up most of the shortening, hence hampering buckling and outward propagation of the backthrusts. The shear zone in experiment C dips towards the indentor plate and has the same function as the Sub-Tauern Ramp (Fig. 6), placing the indentor plate above the foreland plate. Inspired by experiment C, we view the Sub-Tauern Ramp as structurally continuous with the detachment at the base of the northern fold and thrust belt but, contrasting with the interpretation of the TRANSALP Working Group (2002), discontinuous with respect to the Innthal Fault.

Conclusions

Conclusions from analogue experiments for collisional settings involving a weak plate are generic and applicable to nature.

The relative strength between converging plates and intervening weak plates controls both the deformation location and style. Small contrasts favour buckling of both weak and indentor plates, whereas strong contrasts focus deformation onto the weak plate for a larger amount of shortening and result in the formation of important crustal roots. Buckling is a fundamental shortening mode of the lithosphere, even in cases with strength heterogeneities. The presence of a buoyant mechanical instability above the model Moho within the weak plate favours the development of asymmetric orogens notwithstanding the initially symmetrical setup. Furthermore, the mechanical weakness promotes strain localization and hence influences the sequence of deformations and their topographic expression.

The absence of lower crustal wedges along the TRANSALP line does not necessarily reflect a lack of strain in the lower crust and does not allow drawing unambiguous conclusions on its strength. If the lower crust is weak, as in our experiments, strain is diffusely distributed over the foreland and indentor and results in homogeneous thickening of the lower crust layer. This type of deformation is consistent with the aseismic behaviour of the European lower crust along the TRANSALP transect.

Experiment C suggests that the basal detachment, which floors the fold and thrust belt, is the northern structural continuation of the Sub-Tauern Ramp. Experiment C further suggests that short-wavelength buckling within the Tauern Window requires decoupling of the upper crust from its upper mantle and probably also from its lower crust. The simple geometric and kinematic boundary conditions used in this study show that folding is associated with orogen-parallel extension, which upon strong lateral confinement, is insufficient on its own to explain large-scale extension in the Eastern Alps.

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