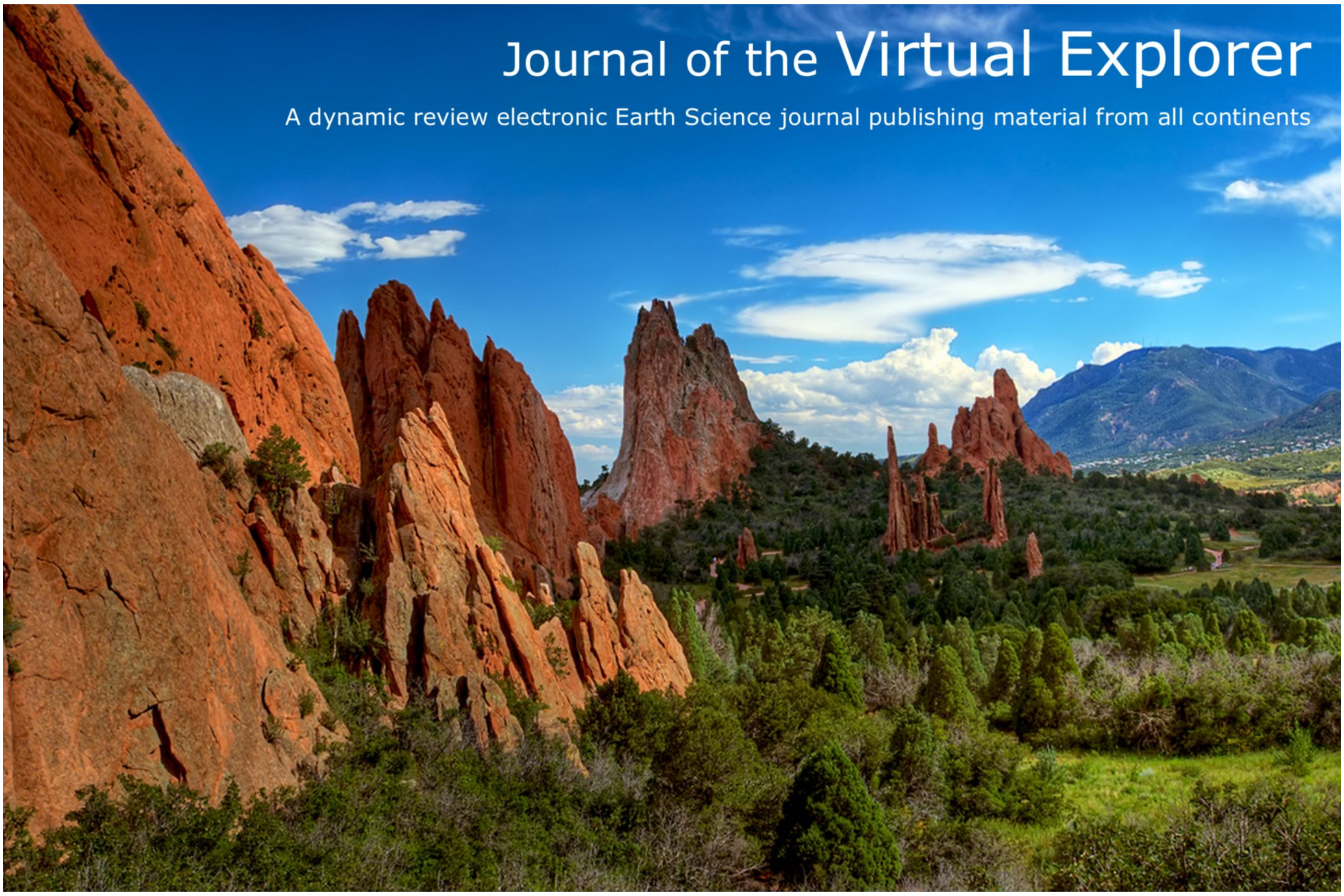

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Unraveling the geodynamics of the Central Mediterranean in a tank

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Abstract: The geodynamic setting of the Central Mediterranean is dominated by the Tyrrhenian subduction. Here, we show how laboratory models of subduction, combined with natural data, can be used as a fundamental tool to constraint the evolution of the Central Mediterranean over the last 80 Ma. We emphasize how simple tank models can help to catch the essence of topics like 1) the initiation of the subduction process, 2) subduction dynamics and their relationships with plate kinematics, 3) the episodicity of back-arc extension, and finally 4) the role played by slab-induced mantle circulation.

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Introduction

Subduction is very important to plate tectonics. Wadati-Benioff zones (e.g. Isacks & Barazangi 1977, Giardini & Woodhouse 1984, Jarrard 1986) and the distribution of tomographic anomalies (e.g. Bijwaard 1999, Fukao *et al.* 2001) provide a snapshot of present-day conditions, suggesting that the cold solid lithosphere sinks into the convective, fluid-like mantle with different dips and shapes. The long-term evolution of subduction is still uncertain and difficult to unravel due to the transient character of the process. Hence, to better understand the dynamics of subduction, seismic data have to be integrated with other constraints; these include indirect observables (geological, petrological, geochemical and structural studies of the trench – back-arc system) and numerical and laboratory models. Modeling, in fact, is the only available tool that can provide a dynamic and evolutionary picture of the slab. The sinking of a dense material in a Stokes fluid may catch the essence of the subduction process, offering a possible explanation for a wide range of natural observations (e.g., Kincaid & Olson 1987, Griffiths *et al.* 1995, Zhong & Gurnis 1997, Buttles & Olson 1998, Schmeling *et al.* 1999, Kincaid & Griffiths 2003, Schellart 2005, Husson 2006, Capitanio *et al.* 2007, Becker & Faccenna 2008, Royden & Husson 2008).

Here, we provide an overview of what can be learned about subduction with laboratory models. Because no single, comprehensive model currently exists that can simulate all scales of subduction, we focus our attention on a specific area separating different temporal aspects of the process; these include the early, intermediate and long-term evolutionary stages. For this purpose, the Central Mediterranean (Figure 1) offers an ideal regional test site. This area is characterized by a very complex structural setting. It contains a northwest dipping subduction zone with an arcuate shape, which produced the Apennine chain, and an extensional domain, the Tyrrhenian Sea, which can be thought of as the recent back-arc basin of the system (e.g., Malinverno & Ryan 1986, Royden 1993, Faccenna *et al.* 1996). In this geodynamical setting, which has been shaped by the interplay between the Eurasian and the African plates, the net convergence of the incoming African plate at the trench has always been very low (Dewey *et al.* 1989, van der Voo 1993, Ward 1994, Silver *et al.* 1998), providing a rare opportunity to preserve the clear geological remnants of past subduction on the surface. These signatures offer the possibility to constrain the evolution of the subduction process, allowing us to study the interaction between the subduction process and back-arc opening. Presently, in addition to the laboratory models that will be described in this paper, two additional approaches have been adopted to describe the main characteristics of the Central Mediterranean subduction. The first approach combines the seismic velocity structure of the subducted lithosphere with the regional tectonic history (e.g., Wortel & Spakman 2000, Faccenna *et al.* 2001a, Faccenna *et al.* 2004). The second approach uses numerical models to investigate the stress field at the surface (e.g., Bassi & Sabadini 1994, Bassi *et al.* 1997, Negredo *et al.* 1997) and at depth (e.g., Giunchi *et al.* 1994, Marotta & Sabadini 1995, Giunchi *et al.* 1996, Carminati *et al.* 1998) including analysis of the mantle circulation pattern (e.g. Ismail-Zadeh *et al.* 2009).

In this paper, we will describe how the geodynamics of the Central Mediterranean can be efficiently illustrated in a tank, using scaled experimental models. First, we briefly describe the geological context that characterizes the Central Mediterranean area. Subsequently, after an initial technical description, we describe the laboratory modeling illustrating different phases of the subduction process. Finally, the results are placed in the context of the global picture for the Central Mediterranean. The questions addressed by this work are the following:

- Which ingredients enhanced the initiation of the subduction process?

- Which dynamic mechanism produced trench migration rates up to 6 cm/yr, with convergence rates of only a few mm/yr?

- Why did the process occur episodically, causing the opening of distinct basins (Liguro-Provençal and Tyrrhenian, along with the Vavilov and Marsili)?

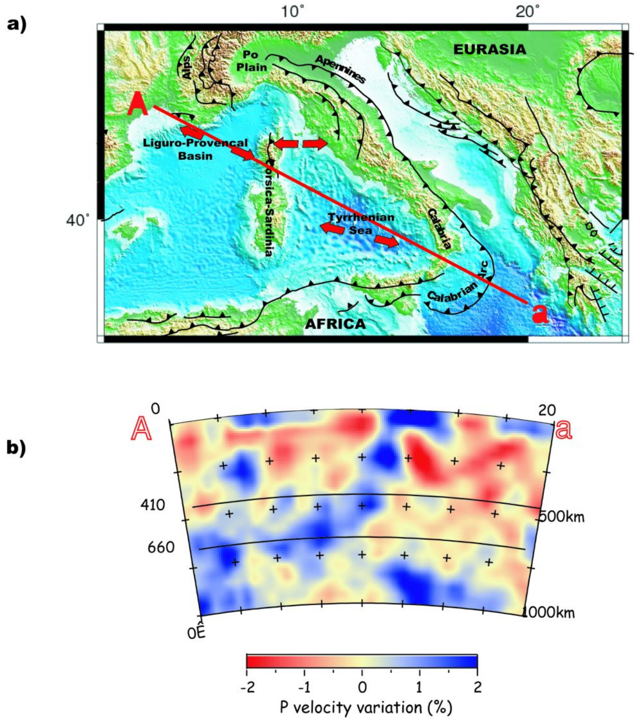
- Can we reconcile the seismological images (tomography and seismic anisotropy) of the deep mantle with the available geological record?

Answers to these questions will not only improve the comprehension of the history of the Central Mediterranean but also shed light on the general evolution of the subduction process.

Tectonic setting

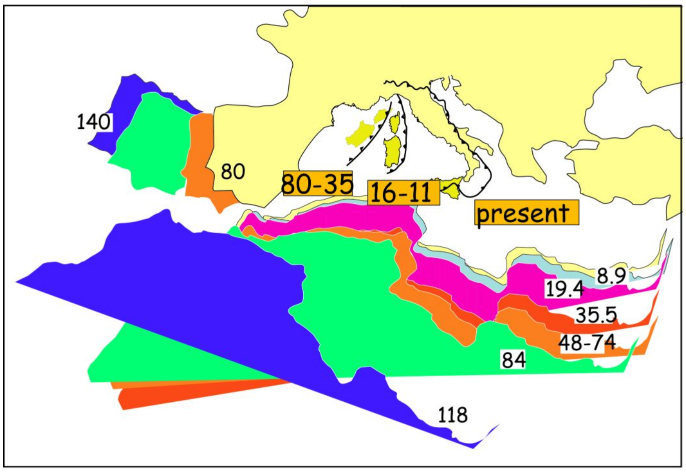
The complex lithospheric structure characterizing the Central Mediterranean is the result of long-term evolution, having as fundamental ingredient the slow convergence of the heterogeneous continental margins of Africa toward the stable Eurasia (Figures 1a, 2). This interaction, which was active over the Cenozoic at a rate of 1-2 cm/yr on average (Dewey *et al.* 1989, Jolivet & Faccenna 2000), consumed the former Tethys Ocean and caused the rise of the Alpine chain where continental collision occurred. From the Oligocene onwards (30 Ma), the scenario differentiated between the Apennininc and the Calabrian sectors. In the former area, the passive continental margin of Apulia entered the trench, leading to the subduction of the continental lithosphere (Dercourt *et al.* 1986), as proven by the inclusion of continental passive margin rocks in the Apenninic accretionary wedge (e.g., Boccaletti *et al.* 1971). Subduction progressively slowed, leading from an active subduction to a Rayleigh-Taylor-like instability as presently illustrated by tomographic images (see Faccenna *et al.* 2001b). The southern area, on the contrary, has been dominated by an impressive (>800 km; up to 6 cm yr^{-1}) slab rollback, which is enhanced by the small, subducting, oceanic Ionian basin and, additionally, by the opening of the back-arc basins (Le Pichon 1982, Malinverno & Ryan 1986). Consequently, an extensional deformation has been locally superimposed on the sites of previous continental thickening (Horvath & Berckhermer 1982). Back-arc opening occurred in discrete episodes. The first episode of extension, occurring from 30-16 Ma (Cherchi & Montandert 1982, Burrus 1984, Gorini *et al.* 1994, Seranne 1999), allowed the formation of the Liguro-Provençal Basin. The extensional process was characterized by an estimated cumulative spreading/rifting of about 400 km (Burrus 1984, Chamot-Rooke *et al.* 1999) and was accompanied by the emplacement of volcanic, basaltic-andesitic deposits (Beccaluva *et al.* 1989 and references therein). At the same time, the Sardinia-Corsica block rotated counterclockwise (van der Voo 1993, Speranza 1999). The slab should have reached a sharp bend during this phase, extending from northern Tunisia to the Apennines, probably with a tear just south of Sardinia (see Faccenna *et al.* 2004 for discussion). The subsequent episode of eastward extension shifted towards the Tyrrhenian domain (Carminati *et al.* 1998, Faccenna *et al.* 2001b) and started after a short, but clearly recognizable interval of about 3-5 Myr, during which extension and magmatism ceased. Two episodes of oceanic spreading occurred in the Tyrrhenian forming the following: (1) the Vavilov Basin during the Pliocene (4-3 Ma) and (2) the Marsili Basin further to the east, whose activity was restricted to approximately ~2-1 Ma (Patacca *et al.* 1990, Nicolosi *et al.* 2006) The pulsating Tyrrhenian backarc spreading has been related to the progressive shallow segmentation and deep disruption of the subducting Calabrian lithosphere (Faccenna *et al.* 2005, Chiarabba *et al.* 2008). This process is rather well documented over the region by the lateral shift of the foreland and the thrust belt activity from Tell towards Sicily (Casero & Roure 1994). It is also documented by the nature of volcanism characterized by the appearance of alkaline anorogenic products replacing the calc-alkaline products in the back-arc region (Maury *et al.* 2000, Faccenna *et al.* 2005). Currently, the Central Mediterranean subduction zone is only still active in the Calabrian sector, as indicated by seismological data. Seismicity is recorded along a narrow (~200 km) and steep (~70°) Wadati-Benioff plane, SW-NE striking and NW dipping, down to about 500 km (e.g., (Anderson & Jackson 1987, Giardini & Velona` 1991, Selvaggi & Chiarabba 1995). This view is supported by tomographic images of the Central Mediterranean mantle (Figure 1b), which reveal a continuous NW dipping high-velocity body extending below Calabria and lying horizontally in the upper-lower mantle transition zone (e.g., Spakman *et al.* 1993, Lucente *et al.* 1999, Piromallo & Morelli 2003). The Calabrian high-velocity anomaly creates a continuous arcuate structure that merges with the northern Apennines below 250 km (Piromallo & Morelli 2003). At shallower depths, a low-velocity anomaly, in the southern Apennines, disconnects the Calabria from the northern Apennines, which represents the seismological signature of a slab tear enhanced by the subduction of a heterogeneous continental margin. This condition not only led lateral subduction variability and slab tears but also triggered a complex mantle circulation.

Figure 1. Central Mediterranean geodynamic framework.



a) Topographic/bathymetric map of the Central Mediterranean region showing the position of the main structures. b) Tomographic cross sections Aa - Gulf of Lyon to Calabria -from model PM0.5 (Piromallo & Morelli 2003).

Figure 2. Motion of Africa and Iberia with respect to fixed Eurasia.



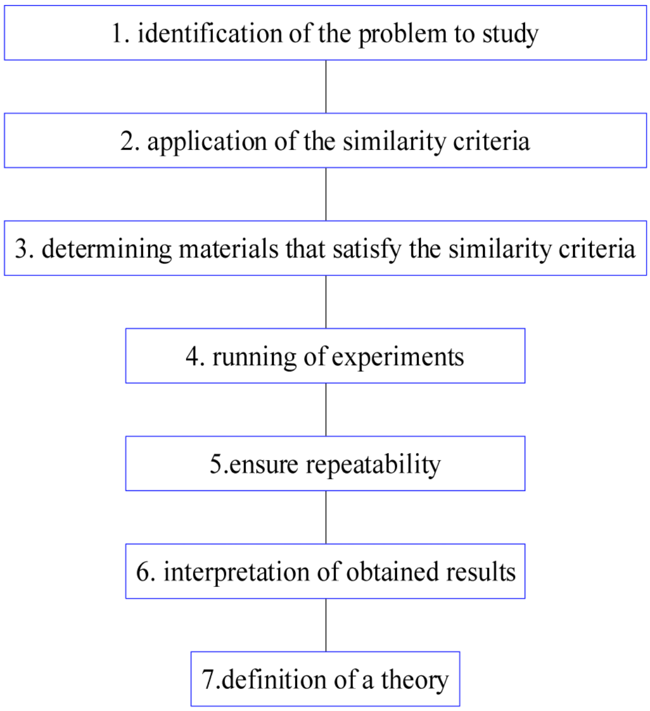
Motion of Africa and Iberia with respect to fixed Eurasia (redrawn from Dewey *et al.* 1989 and Olivet, 1996). The position of the Central Mediterranean trench is shown as reconstructed for the last 80 Ma (see Faccenna et al. 2001 a, b for details).

The pattern of shear wave splitting (SKS) measurements provides invaluable information to constrain mantle deformations. Civello & Margheriti (2004) showed that fast directions in peninsular Italy are quite often parallel to the strike of the thrust belt, while in the Tyrrhenian Sea, they are primarly E-W oriented, corresponding to the back-arc extensional direction. In addition, the N-S oriented measurements in western Sicily had a high angle with respect to the thrust front.

Experimental setup

Despite this compact description, the fundamental ingredients needed to describe the Central Mediterranean tectonic framework have been highlighted as follows (STEP 1, Figure 3): a) the presence of slow rates between convergent plates; b) oceanic vs. continental subduction within a laterally heterogeneous continental margin; and c) slab interaction with the 660-km discontinuity.

Figure 3. Flow chart of different phases of the experimental modeling.



In the following section, we describe how these observations were converted to input data to design a setup that meaningfully simulates the mechanism governing the slab dynamics, with a specific application to the studied area.

Assumptions

The assumptions underlying the design of our experimental models are listed in the following sections (for more detailed explanations see Faccenna *et al.*, 1999 and Funiciello *et al.* 2003, Funiciello *et al.* 2004):

Viscous rheology

The Earth's profile (i.e., lithosphere and mantle) is simulated using linearly viscous rheologies. A brittle ductile layering is adopted only for models that are finalized with the subduction initiation because we are specifically interested in isolating the role that the brittle and ductile strengths at passive margins play in this framework.

Self-consistent subduction

Justified by the fact that the rate of convergence at the trench has always been very low and the consumption of the lithosphere has been mostly driven by negative buoyancy, that results in trench rollback, slab pull is the only active force within the system. No external kinematic boundary conditions (i.e., plate or trench velocity) are applied. Therefore, the experimental subduction process is a self-consistent response to the dynamic interaction between the slab and the mantle. An exception is represented by models used to study the initiation of subduction, where the incoming plate is slightly pushed by means of a rigid piston that pushes at a constant horizontal velocity with reference to the box boundary.

Convectively neutral mantle

In our experiments flow is generated only by subduction. We neglect thermal convection and the effect of global (Ricard *et al.* 1991) or local background flow that is not generated by the plate/slab system.

Isothermal models

Experimental limitations lead us to neglect thermal effects during the subduction process. Hence, we assume a constant chemical density contrast throughout the model, and the roles of thermal diffusion and phase changes (Bunge *et al.* 1997, Lithgow-Bertelloni & Richards 1998, Tetzlaff & Schmeling 2000) are neglected. Consequently, the slab is thought to be in a quasi-adiabatic state. The high subduction velocity (larger than 1 cm/yr in nature) recorded in our models justifies this assumption, ensuring that advection overcomes conduction (Wortel 1982, Bunge *et al.* 1997).

Lower mantle

It is impossible to reproduce the fundamental role of the endothermic phase change at the transition zone in laboratory. Therefore, because we are interested in exploring the effects of the slab-660 km interaction in the evolution of the Central Mediterranean, the lower mantle is simulated by the increase of viscosity with depth (Davies 1995, Guillou-Frottier *et al.* 1995, Christensen 1996, Funiciello *et al.* 2003).

A viscosity increase (by a factor of 10-100) across the 660-km discontinuity has been postulated based on of models of the wavelength of the geoid (e.g., (Hager 1984, Hager & Richards 1989, King & Hager 1994) and the post-glacial rebound (e.g., Mitrovica & Forte 1997). In particular, several models assume that the bottom of the box is analogous to the 660-km discontinuity because this boundary is impermeable for a retreating slab and in the limited timescale of the analyzed process (i.e., <100 Myr; Funiciello *et al.* 2003).

No overriding plate

Except for models related to subduction initiation, the overriding plate is not modeled to simplify the experimental setting. In these cases, we assume that the fault zone between the subducting and the overriding plates is weak, having the same viscosity as the upper mantle (Tichelaar & Ruff 1993, Zhong & Gurnis 1994, Conrad & Hager 1999). We also assume that the overriding plate passively moves with the retreating trench. This choice, which does not invalidate the general behaviour of the experimental results, appears suitable to simulate the Eurasian plate. However, the rate of the subduction process can be slightly influenced (King & Hager 1990). When we are interested into a more precise kinematic assessment, we introduce a passively moving overriding plate that is positively buoyant with respect to the underlying mantle.

Reference frame

The reference frame for the entire set of models is the box boundary. This frame is considered the experimental analogue to the fixed hot spot reference frame.

Materials

A crucial requirement for experimental models is the ability to scale natural processes to the laboratory enwvironment. Similarity analysis (STEP 2, Figure 3) is the central step in the model design that helps to select the analogue materials, dimensions and rate of deformation. To scale a laboratory model to a natural process, it should be geometrically, kinematically, dynamically and rheologically similar to the prototype (Hubbert 1937, Ramberg 1981). The application of similarity theory starts with the identification of the most relevant physical parameters active into the natural system. Then, each variable (length, velocity, force and material specific parameters) is normalized by means of a dimensionless number. Each set of dimensionless parameters defines a family of equivalent solutions that only differ by a scale factor. The solution, which is characterized by the most relevant scaling, can be used to select materials and design a properly scaled analogue model (STEP 3, Figure 3).

Silicone putty (Rhodrosil Gomme, PBDMS + iron fillers) and glucose syrup/honey were used as analogues of the lithosphere and mantle, respectively. Silicone putty is a visco-elastic material, with purely viscous behavior at low experimental strain rates (Weijermars & Schmeling 1986). Glucose syrup/honey is a transparent, Newtonian, low-viscosity and high-density fluid. For the models used to study the initiation of subduction initiation, a shallow sand mixture layer was added to simulate the brittle behavior of the upper crust. These materials achieved the standard scaling procedure for stresses and were scaled down for length, density and viscosity in a natural gravity field ($g_{model} = g_{nature}$) as described by Weijermars & Schmeling (1986) and Davy & Cobbold (1991).

Table 1. Parameters used in the selected models and in nature.

PARAMETER			NATURE	MODEL
g	Gravitational acceleration	$m \times s^{-2}$	9.8	9.8
Thickness				
h	Oceanic/continental lithosphere	m	1×10^3	0.012
H	Upper mantle		6.6×10^5	0.110
Density				
ρ_l	Oceanic lithosphere	$kg \times m^{-3}$	3300	1482 - 1491
ρ_{cl}	Continental lithosphere		3190	1375
ρ_{um}	Upper mantle		3220	1380-1422
Viscosity				
η_l	Oceanic lithosphere	$Pa \times s$	10^{24}	$(1.6\text{-}3.6) \times 10^5$ (± 5%)
η_{cl}	Continental lithosphere		$\approx 10^{24}$	1.8×10^5 (± 5%)
η_{um}	Upper mantle		10^{20}- 10^{21}	$5.5 \times 10 - 4.6 \times 10^2$(± 20%)
Time				
t	$\frac{t_{nat}}{t_{mod}} = \frac{((\overrightarrow{g}\Delta\rho h)/\eta)_{mod}}{((\overrightarrow{g}\Delta\rho h)/\eta)_{nat}}$	3.1×10^{13} (1 Myr)	≈ 60	

The scale factor for length was $1.6 \cdot 10^{-7}$ (1 cm in the experiment corresponds to 60 km). Densities and viscosities were assumed to be constant over the thickness of the individual layers and were considered to be an average of effective values. The scaled density factor between the oceanic lithosphere and the upper mantle ranged from 1.05-1.07 (Molnar & Gray 1979, Cloos 1993). The viscosity ratio between the oceanic slab (η_l) and the upper mantle (η_{um}) ranged between 10^2-10^4. When continental subduction was modeled, the negatively buoyant silicone was laterally joined to a light silicone plate with decreased density over its entire thickness (see Martinod *et al.* 2005).

Considering the imposed scale ratio for length, gravity, viscosity and density (Table 1) applied to the lithosphere, 1 Myr in nature corresponded to ~1 min in the models for the purely viscous models (i.e., not including models finalized for subduction initiation). Parameters and values for nature and the experimental system are listed in Table 1.

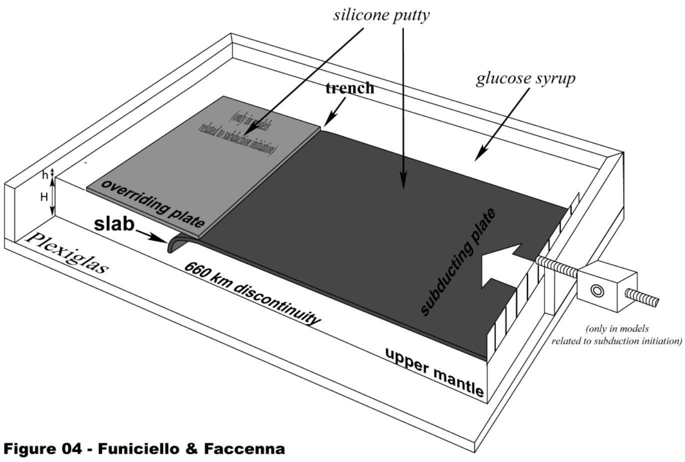
Experimental procedure

The layered system was arranged in a transparent Plexiglas tank (Figure 4). To limit the lateral box boundary effect, which could alter the results, plates were positioned as far as possible from the box sides (> 11 cm; see Funiciello *et al.* 2004). Plates were fixed at their trailing edge (“fixed ridge” sensu Kincaid & Olson 1987) to simulate a subduction zone belonging to the stable Africa. Except for the models of the subduction initiation, the subduction process was manually started by forcing the leading edge of the silicone plate downward into the glucose to a depth of 3 cm (corresponding to about 180 km in nature) and with an angle of ~30°.

Each experiment (STEP 4, Figure 3) was monitored using a sequence of digital pictures taken in lateral and top views. Kinematic and geometric parameters (trench motion, plate motion, geometry of the slab at depth, dip of the slab, and surface plate deformation) were then quantified using data analysis tools. Once planned, mantle circulation was analyzed by recording the experiment over its entire duration using two high-velocity, high-resolution, black-and-white progressive scan cameras with two lighted, orthogonal interrogation windows: the x-z plane through the centerline of the tank-slab system and the x-y plane just below the plate at a depth of about 2 cm. In this case, the glucose syrup was preliminary seeded with neutrally buoyant, highly reflecting air microbubbles used as passive tracers. These tracers negligibly influence the density and viscosity of the mantle fluid. Movies were stored on a dedicated hard disk and post-processed using the Feature Tracking (FT) image analysis technique (see Funiciello *et al.* 2006) to retrieve the circulation pattern (i.e., the mantle velocity field, mantle velocity x- and y- components and modulus, the streamlines of mantle circulation, and the mantle linear flux) highlighted by passive tracer particles.

Moreover, the models were repeated several times under the same boundary conditions to ensure repeatability (STEP 5, Figure 3).

Figure 4. Cartoon showing the experimental setup adopted in the presented models.



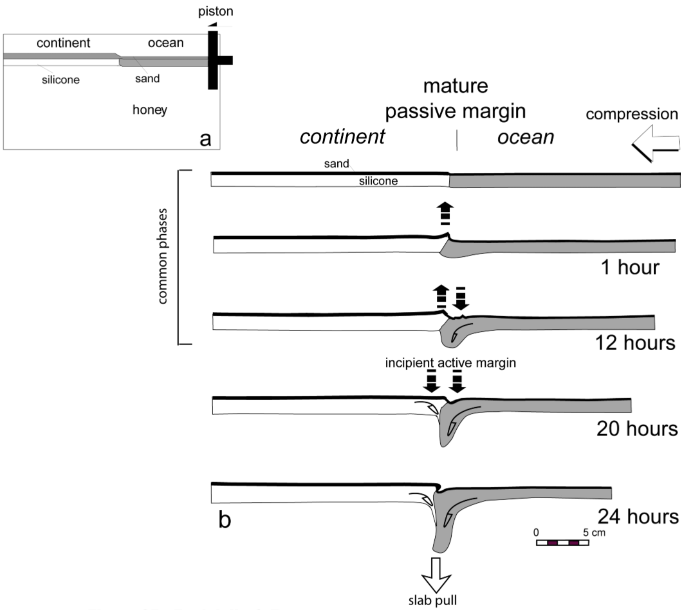
Experimental results

We present modeling results finalized to the study of the subduction, separating different aspects of the process. Here, the focus is only the physical meaning of the obtained results, and this information will be merged into a simplified global model qualitatively describing the Central Mediterranean geodynamic evolution over the last 80 Ma. It is useful to experimentally analyze the model-making process, which becomes progressively more complex to separate the contribution of single variables. In particular, our models will help with the following: a) provide an explanation of the main factors promoting subduction initiation; b) describe rapid episodic back-arc opening; c) offer a three-dimensional view of slab-induced mantle circulation; d) explain the consequences of subduction of a heterogeneous continental margin; and f) explain the opening of slab tears.

Initiation of subduction

Models simulating different configurations and rheologies of the oceanic (i.e., subducting plate) and continental (i.e., overriding plate) lithosphere have been developed to simulate feasible evolutionary scenarios for passive margins. In particular, four parameters have been tested: 1) the negative buoyancy of the subducting lithosphere (by changing the density of the silly putty); 2) the horizontal forces (by changing both the magnitude and interval of application of the horizontally pushing piston, and the density contrast between the continent and the ocean); 3) the brittle strength at passive margins; and 4) the ductile strength at passive margins (Figure 5a; see Faccenna *et al.* 1999 for additional details).

Figure 5. Modeling results of subduction initiation.



Lateral view of the experimental setup for models developed for the subduction initiation. b) Line drawing of the main evolutionary stages recognized in this set of models.

All the models share a common early behavior characterized by a rapid and localized initial uplifting of the continental plate followed by down-folding of the oceanic plate that is restricted to the passive margin zone and along the piston (Figure 5b). With an increase in shortening, the oceanic plate is affected by pervasive folding, and the passive margin undergoes bulk thickening with an indentation of the oceanic plate into the continental plate, as well as a small amount of underthrusting. Our results (see Faccenna *et al.*, 1999 for details) indicate that either an initial low compressional strain rate (operating over geological time scales) or the pressure produced by the collapse of the continental margin is necessary to trigger the development of a small instability localized at the passive margin (i.e., Rayleigh–Taylor instability). The slow growth of the instability is initially accommodated by ductile flow in the lower lithospheric layers and driven toward the margin by horizontal forces, which are not supported by the low lithosphere strength at these depths. In the subsequent phases, the instability can evolve and only allow subduction initiation under specific conditions (Figure 5b). The negative buoyancy of the oceanic plate is, therefore, the key ingredient needed to transform a margin from passive to active. A density contrast higher than 50 kg cm $^{-3}$ between the oceanic lithosphere and the underlying mantle is required to localize the compressive strain at the passive margin and to initiate self-sustained, one-sided subduction flanked by continental margin subsidence. In particular, we found that the subduction is initiated only when the negative buoyancy force overcomes the resistance offered by the ductile lithospheric level. This condition was experimentally achieved once the instability reached the length of about 2.5 cm, corresponding to about 150 km in nature. In particular, we experimentally verified that the brittle strength of the lithosphere exerts only a negligible influence on the trench nucleation when compared to the viscous strength.

In summary, the subduction initiation process described by our models is not a sudden brittle failure of a passive margin but rather a slow, continuous transition controlled by ductile flow in the lower lithosphere.

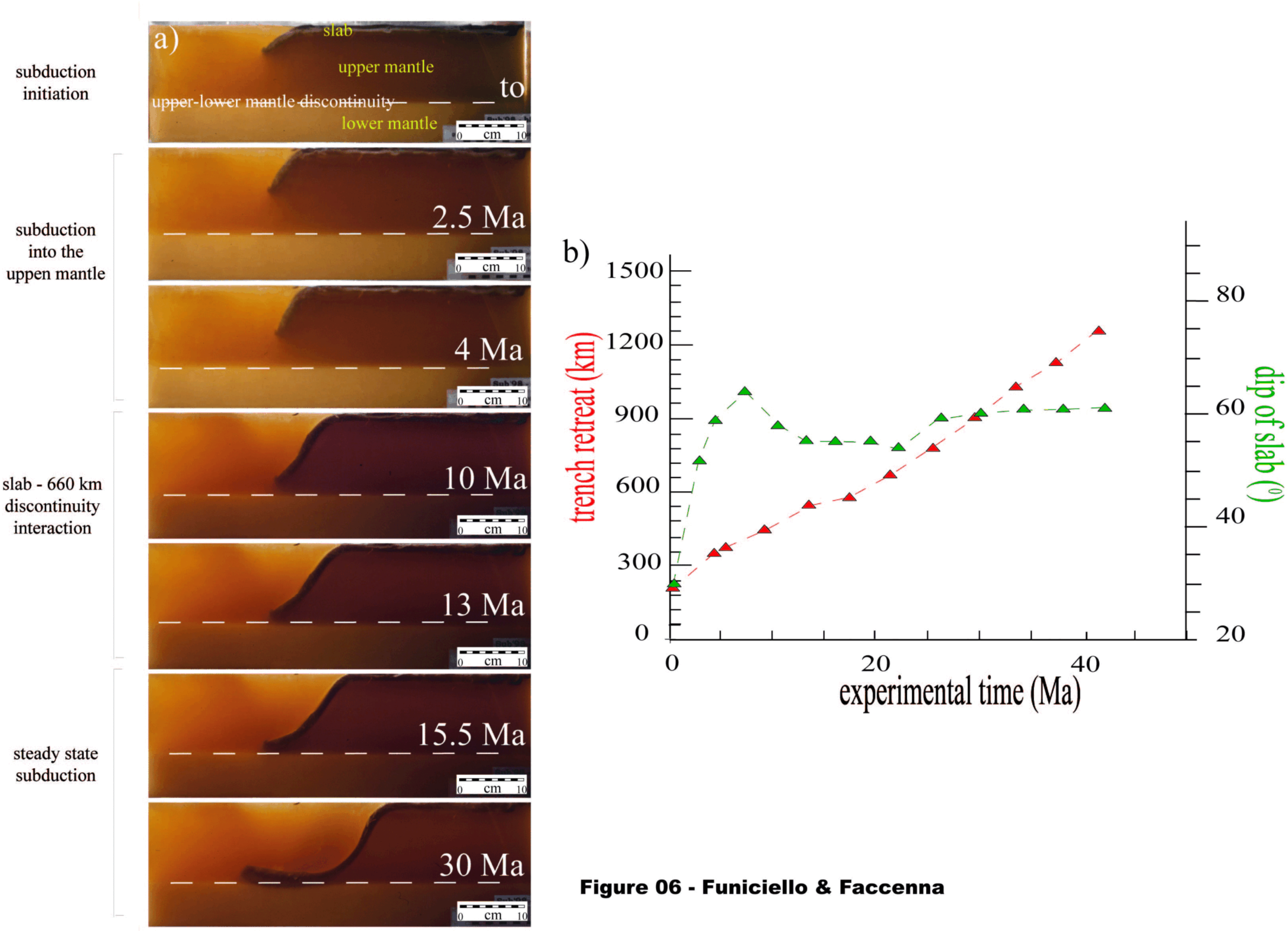
Slab into the upper mantle and interaction with the 660-km discontinuity

This section describes the models developed to study the behavior of the slab sinking into the upper mantle, and the interaction with the 660-km discontinuity (Becker *et al.* 1999, Funiciello *et al.* 2003, Funiciello *et al.* 2004, Bellahsen *et al.* 2005). First, the subduction process described in the previous section was simplified, while maintaining its usefulness, by forcing the leading edge of the lithosphere inside the mantle at a shallow dipping angle to obtain enough slab pulling force to overcome the resistance at the trench. After initiation, the slab sinks into the mantle, increasing its dip to 90o, while the trench retreats (Figure 6a, b). This process is always associated with a significant displacement of the mantle from beneath and around the slab triggered by the subducting plate (Funiciello *et al.* 2003, Funiciello *et al.* 2006). The amount of subduction, which is equal to the trench motion in this peculiar configuration characterized by a fixed incoming plate, increases exponentially with the slab length (Figure 6b). Becker *et al.* (1999) and Funiciello *et al.* (2003) found that the behavior shown in this stage is primarily controlled by the interplay between the slab pull (progressively increasing with time) and resistance related to the plate bending. However, a precise force assessment is difficult using the sole experimental approach (Funiciello *et al.* 2008), and slab-mantle interaction cannot be neglected as an additional contribution to the resisting force (e.g., Capitanio *et al.* 2007, Di Giuseppe *et al.* 2008, Funiciello *et al.* 2008, Stegman *et al.* 2010). The increasing rates of trench retreat favor the development of an extensional tectonic regime above the subducting lithosphere (i.e., overriding plate), which we assume is the analog of a natural back-arc opening.

After about 10 min (corresponding to about 10 Myr in nature), the slab interacted with the mantle discontinuity (Figure 6a). In this phase, the subduction slowed for a few minutes (Figure 6a, b), while the tip of the slab folded and deformed at depth. For our retreating configuration (i.e., fixed edge at the leading edge), this behavior is a direct consequence of the high viscosity layer representative of the lower mantle. Over the relatively short time scale of our model (< 100 Myr), advection inside the low viscosity layer was slow, and the slab could not directly penetrate into the lower mantle. We experimentally found that this behavior occurred for viscosity contrasts between the upper and the lower mantle that were higher than 10 (Funiciello *et al.* 2003). The process restarted after the reorganization of the mantle circulation, which was partially inhibited by the presence of the slab along the upper mantle depth. Trench retreat was quite constant, and the slab dip reached values of about 60° (Figure 6b), while its tip lay horizontally on top of the lower mantle. This scenario is justified by the fact that active and resisting forces were constant during this phase, which allowed for a steady-state behavior of the subduction system. Once the subduction and trench migration resumed, the locus of the back-arc extension jumped trenchward, following the new, steeper configuration.

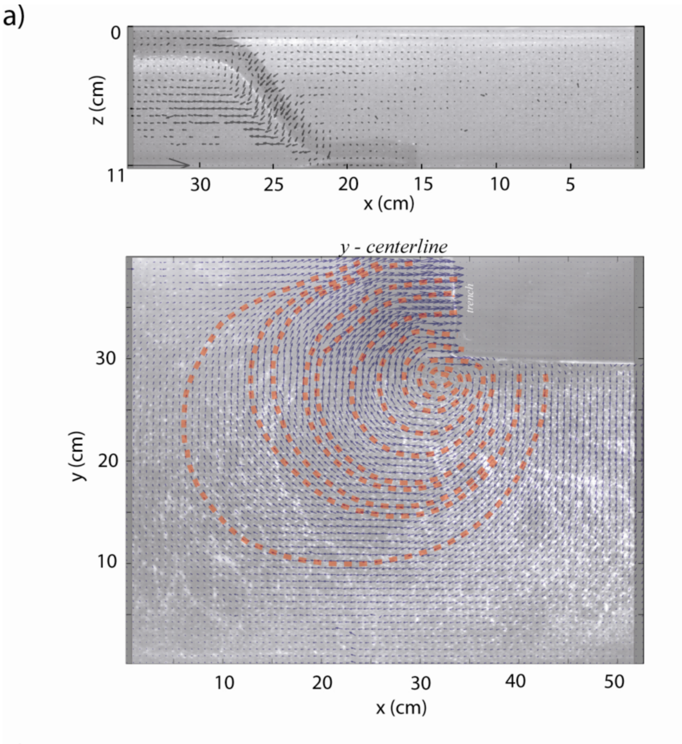
The mantle circulation pattern associated with the subduction into the upper mantle and the interaction with the 660-km discontinuity was continuously monitored and quantitatively estimated using the FT technique (see Funiciello *et al.* 2006 for additional details). The evolution of the model showed that subduction generated a complex 3-D, time-dependent mantle circulation pattern (Figure 7). The mantle velocity field was decomposed into poloidal and toroidal components. The poloidal flow corresponds to pure sources and sinks in the horizontal plane and is related with vertical mass transport in the viscously coupled slab-mantle system. Toroidal motion corresponds to vortex-like flow and rigid body rotations (e.g., O'Connell *et al.* 1991). The spatial and temporal features of mantle-induced circulation were carefully analyzed. We found that the poloidal and toroidal mantle components were both active from the beginning of the subduction process (Figure 7). The poloidal component is characterized by the presence of two cells, one at each side of the migrating slab, with different circulation regimes: the oceanic side cell is wider, has a shallower rotation center and is stable in depth. Conversely, the back-arc side cell center of rotation migrates toward the bottom of the box following the slab tip. The poloidal cells were initially not separated by the slab, allowing for a return flow beneath the tip of the slab. The mantle exchange between the two cells faded away as the slab approached the 660-km discontinuity. In particular, when the slab interacted with the upper/lower mantle discontinuity, the poloidal circulation was reduced significantly, only resuming in the third kinematic stage. In this stage the ocean and back-arc side cells were both active, but the slab represented a barrier for material exchange in the vertical direction.

Figure 6. Modeling results of subduction into the mantle and the interaction with the 660-km discontinuity.



a) Lateral view of seven stages of evolution of a model showing the slab behavior during the subduction into the mantle and the interaction with the 660-km discontinuity. The upper-lower mantle limit in this run is a viscosity increase of 30. However, all the models have an upper-lower mantle viscosity ratio higher than 10 and share a common behavior described in this picture. b) Plot of trench retreat (in red) and dip (in green) versus time.

Figure 7a. Velocity field and streamlines of mantle circulation during the subduction into the upper mantle.



Velocity field and streamlines of poloidal (upper panel) and toroidal (lower panel) mantle circulation during the steady-state phase, as obtained from the FT analysis; time-evolution of linear flow for both x and y components of the mantle circulation. Flow components are:

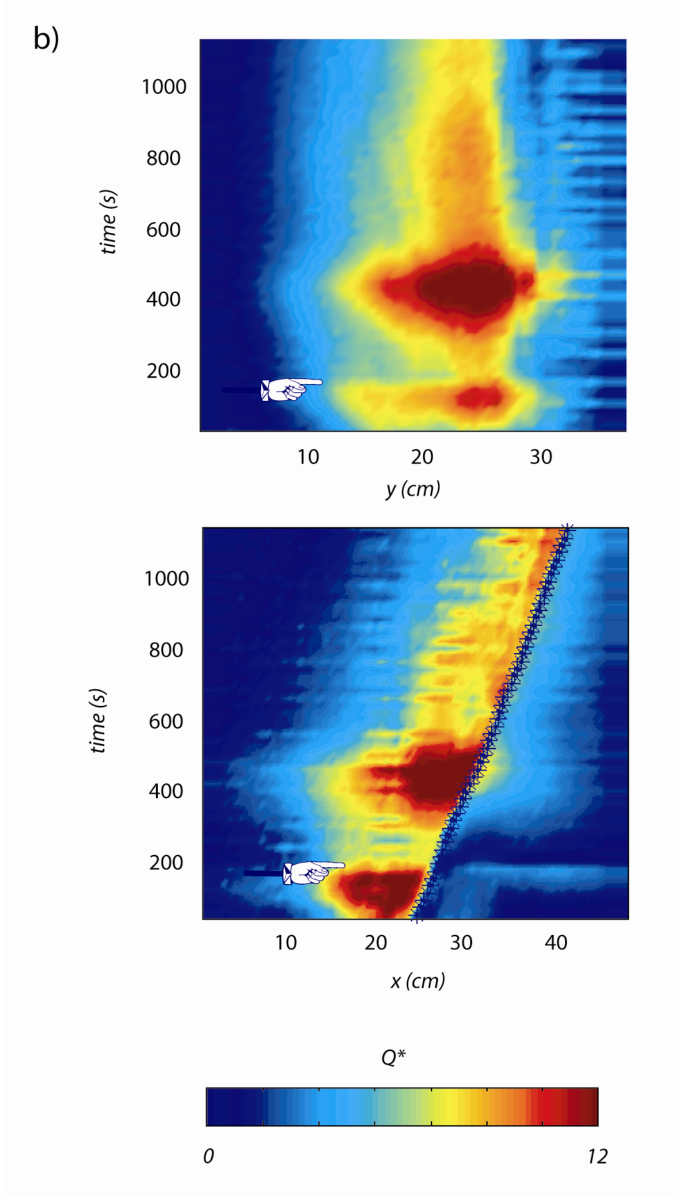
$$Q_x = \int \mid v_y dx \mid$$

and

$$Q_y = \int \mid v_x dy \mid$$

Both components are normalized for a reference flow obtained by multiplying the characteristic trench velocity of the model by the plate thickness. The time of slab-660 km discontinuity interaction is indicated in the left panel. The position of the trench motion at each time is indicated by asterisks in the right panel. This panel summarizes the evolution of toroidal flow during the entire evolution of the experiment, highlighting its strong episodicity (see Funiciello et al. 2006 for details).

Figure 7b. Velocity field and streamlines of mantle circulation during the subduction into the upper mantle.



See caption under Figure 7a.

The toroidal component, produced by the lateral slab migration in the laterally heterogeneous viscous system, was characterized by two symmetric toroidal return-flow cells centered near the plate edges (Figure 7a and b). Each of the two toroidal cells had fixed dimensions that were linearly dependent on the trench width and followed the trench migration during the entire evolution of the process (Figure 7b).

Finally, our models highlighted the episodic behavior of the mantle circulation. The mantle velocity field consistently increased with time, with trench velocity reaching its maximum before the slab interaction with the 660-km discontinuity (Figure 7b), and it reached a steady-state value only after a slowdown caused by the slab-upper/lower mantle discontinuity interaction. The velocity peak at each time was recorded in the region in front of the trench (Figure 7b) and assumed a value proportional to the trench velocity. The lateral component of flow reached its maximum velocity at a distance of about 300 - 400 km from the slab edge (Figure 7b).

Continental subduction

The subduction of the continental lithosphere/oceanic plateau can affect both the kinematics and the geometry of the process by tuning the key driving force active in the system. To experimentally study such a mechanism, we developed models containing a subducting plate simulating a dense oceanic plate, followed by a positively buoyant lithospheric block (Martinod *et al.* 2005, Espurt *et al.* 2008). Our models illustrate the typical sequence of events as described in the previous section: subduction initiation, slab interaction with the upper/lower mantle discontinuity, followed by a steady-state regime reached by the subduction of the oceanic lithosphere (Figure 10b-d). In this stage, the process was characterized by a subduction velocity and a slab dip of about 0.3 mm s^{-1} and 50°, respectively. As soon as the positively buoyant lithosphere began subducting, the subduction velocity suddenly decreased with the slab steepening (Figure 10b-d). The process stopped when ~4 cm of the light lithosphere (corresponding to ~280 km in nature) were subducted, enhancing the progressive verticalization of the slab (Figure 10b and 10d). The presence of a passively moving overriding plate does not alter the overall behavior of the system because the overriding plate passively followed the retreating trench (Figure 10b) and did not deform during the overall modeling evolution.

These experimental results show that the single ingredient represented by the subduction of a large piece of buoyant lithosphere is unable to generate a flat slab segment. This observation contradicts the popular idea, typified by the Mariana and Andes examples, that old and young plates should subduct vertically and horizontally into the mantle, respectively (Uyeda & Kanamori 1979).

Our model illustrates that part of the buoyant lithosphere can easily sink in the upper mantle, which is being pulled at depth by the lower part of the slab (Figure 10b). This can be dynamically justified because the positive buoyancy of the continent/plateau compensates for the negative buoyancy of the dense oceanic slab until an equilibrium is reached. In natural cases, the process may continue with continental/plateau subduction followed by slab break-off (Regard *et al.* 2003). However, the simplified Newtonian viscous rheology we adopted to simulate the experimental lithosphere does not favor the occurrence of this phenomenon in our models.

Opening of slab windows

In this section, we present models to investigate the long-term and transient effects of a reduction in slab width on the subduction kinematics. These models can offer insights on the consequences of slab tearing and, in turn, slab width reduction on trench kinematics.

In the first set of models (see Guillaume *et al.* in press), the subducting plate width, kept constant during the experiment, was systematically changed. As previously described, all the models consist of three phases: (1) sinking of the slab through the upper mantle, (2) transient slab/lower boundary interaction, and (3) steady-state subduction after the tip of the slab has interacted with the 660-km depth boundary. Our results confirm that trench velocity at steady state (V_t) scales inversely with slab width (w) (i.e., wider slabs are related to smaller rollback velocity) (Figure 8). These experimental results are in agreement with laboratory (Bellahsen *et al.* 2005, Funiciello *et al.* 2006) and numerical models (Morra *et al.* 2006, Stegman *et al.* 2006, Schellart *et al.* 2007, Di Giuseppe *et al.* 2008) for a free subducting plate. Our data, once scaled to nature, are also compatible with the semi-analytical results of (Royden & Husson 2006), confirming that the trench retreat velocity is controlled by the mantle’s capability to turn around the slab, producing a toroidal circulation. Larger slab width leads to larger amounts of mantle material stirring laterally around the slab edges and, in turn, slower trench retreat velocity.

Figure 8. Trench retreat velocity during steady-state subduction vs. slab width.

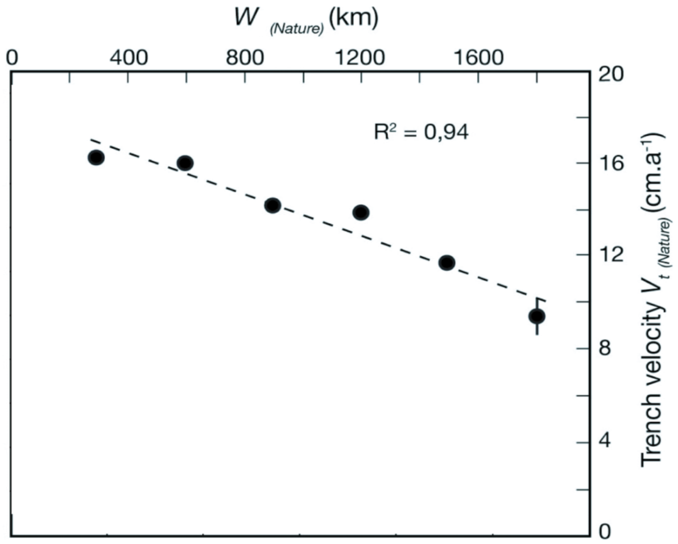
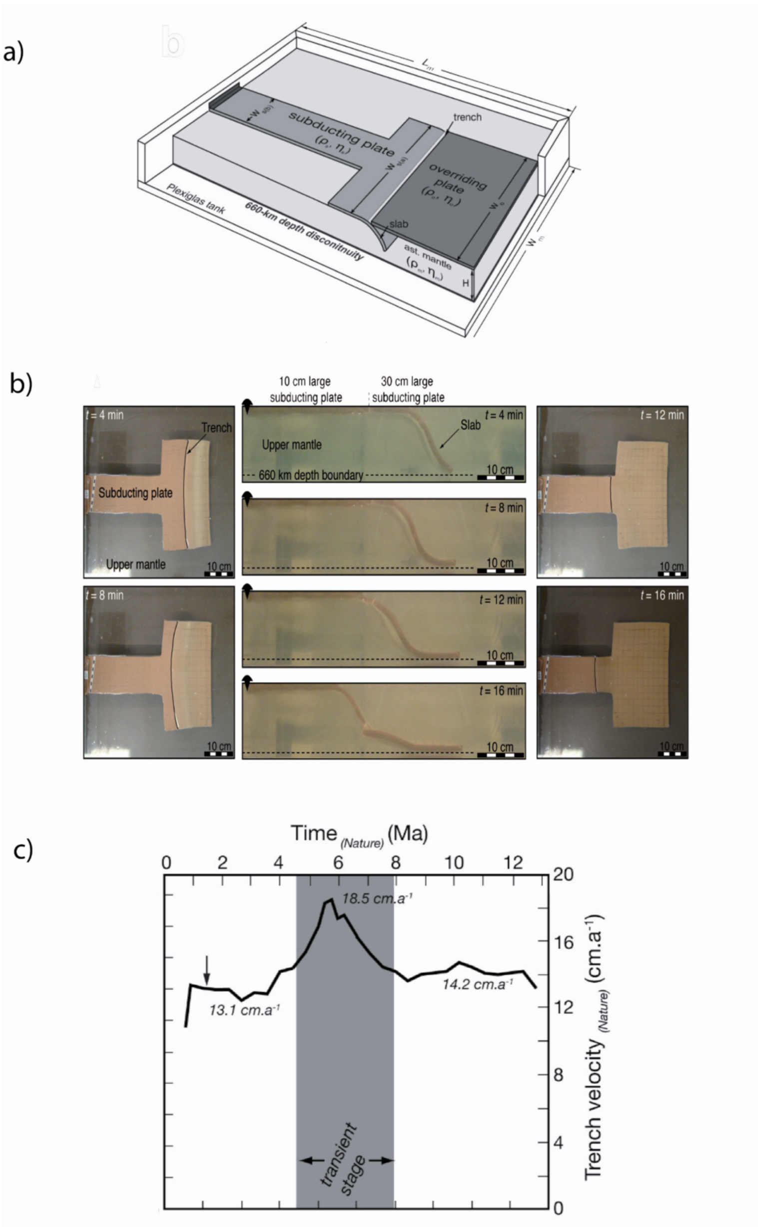


Diagram showing trench retreat velocity during steady-state subduction (V_t) vs. slab width (w) for models with a constant slab width during subduction. Lengths and velocities are scaled to natural values. The corresponding values for the natural case (considering a 72 km-thick slab, with a negative buoyancy of 77 kg.m^{-3} and a surrounding mantle with a viscosity of about 10^3 Pa s) are also reported.

A larger slab width leads to larger amounts of mantle material being stirred laterally around the slab edges and, in turn, a slower trench retreat velocity.

The second set models were developed to study the transient behavior of subduction related to slab narrowing (see Guillaume *et al.* in press for details). The models adopted a simplified configuration with an abrupt decrease of the subducting plate width (Figure 9a). The selected setup ensures that the steady-state subduction regime is reached before the narrow slab portion reaches the trench. In this way, the kinematic consequences of slab width variation can be univocally recorded. All the models had a typical sequence consisting of four phases (Figure 9a-b): (1) sinking of the slab through the upper mantle; (2) steady-state subduction of the wide part of the slab ("wide slab" stage) after the tip of the slab has interacted with the 660-km depth boundary; (3) transient stage, starting when the narrow portion of the subducting plate entered the trench. This phase is marked by a rapid slab avalanche resulting in a trench velocity acceleration of about 50 % with respect to the velocity recorded during the second phase. The trench velocity progressively decreases by about 25 % to defensively stabilize during phase (4) when, after the trailing edge of the wide slab has interacted with the 660-km discontinuity, steady-state subduction of the narrow part of the slab ("narrow slab" stage) is finally reached.

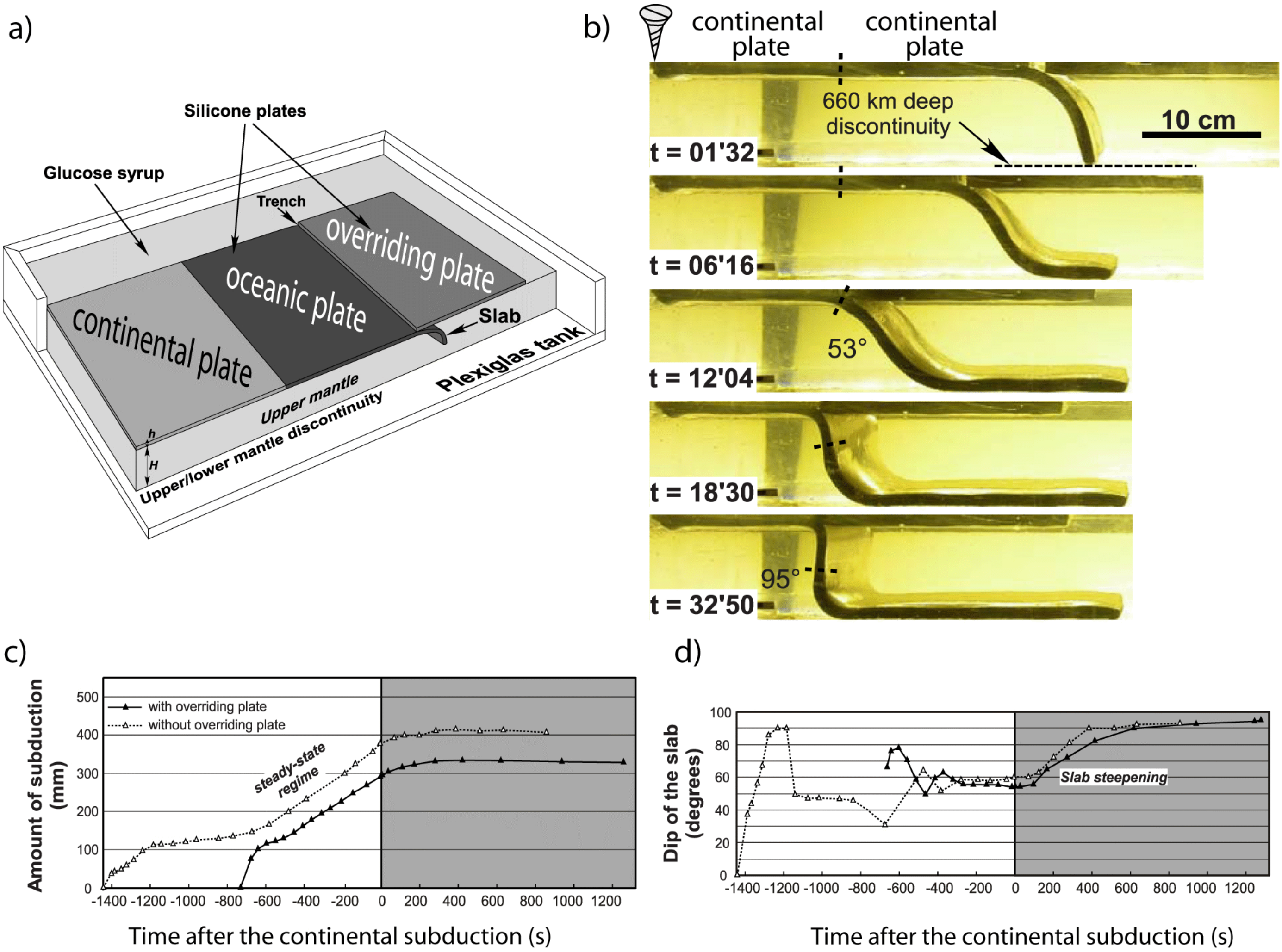
Figure 9. Modeling results to study effects related to the reduction of slab width.



a) 3-D view of the experimental set-up developed to study the transient effects of the reduction of slab width; b) Lateral and top views of four stages of evolution of a model showing the slab behavior during the subduction into the mantle and the interaction with the 660-km discontinuity while the slab width abruptly decreases. c) Trench velocity, V_t vs. t, for a reference experiment. The time it takes the slab to reach the 660-km depth discontinuity is indicated by a black arrow, and the grey frame indicates the duration of the transient stage.

In terms of slab dynamics, the peculiar peak in trench velocity recorded after the abrupt decrease in slab width may be explained by the rapid decrease in resistance force (mainly slab bending resistance and toroidal flow) related to the reduction in width, while the load of the entire slab, including the lateral detached portion, was still efficiently pulling the plate at depth.

Figure 10. Modeling results of continental subduction.

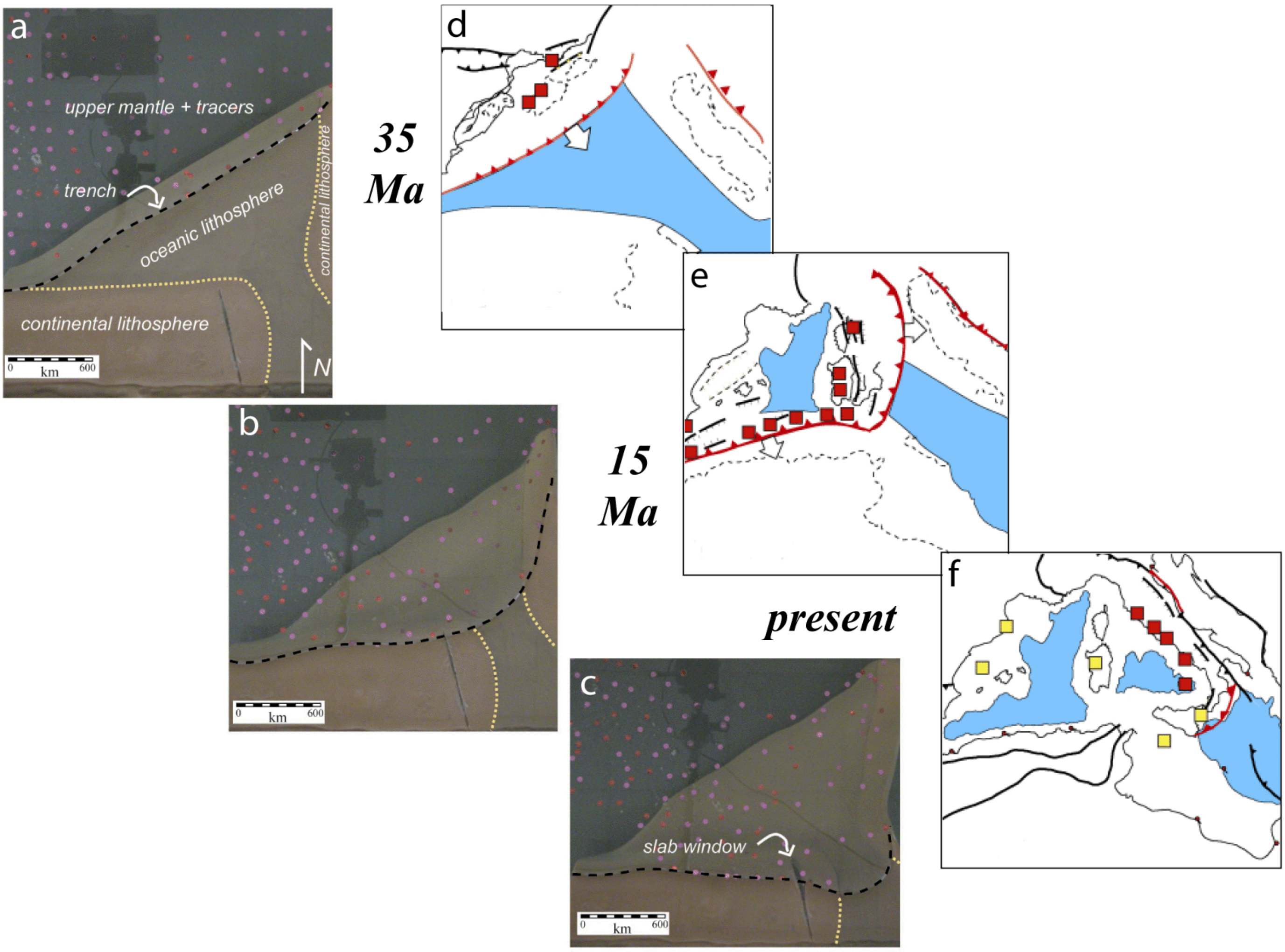


a) 3-D schematic view of the experimental set-up adopted to simulate the subduction of the continental lithosphere. The subducting plate simulates a 50-Ma-old dense oceanic plate, followed by a continental lithosphere. b) Lateral views of the model. The dashed line marks the tip of the oceanic plateau; at t = 1 min 32 s: the slab interacts with the 660-km discontinuity; t = 6 min 16 s and 12 min 04 s: steady-state subduction occurs at a constant speed, and the slab dip maintains a constant value of about 50° until the initiation of continental subduction; t = 18 min 30s: the continental lithosphere starts to subduct, decreasing the velocity of subduction and increasing the dip of the slab (see panels c and d); t = 32 min 50 s: the buoyancy of the continental lithosphere stops the subduction process. c) Amount of subduction and d) dip of the slab versus time. Results of another model developed without any overriding plate is shown for comparison (white triangles).

Discussion: towards a comprehensive picture of the Central Mediterranean history

The experimental data summarized in this paper suggest several implications for the tectonic history of the Central Mediterranean subduction zone (CMSz). It is tempting to merge them into a three-dimensional evolutionary model that will offer a “modeling” picture of the CMSz. For this purpose, five key moments in the evolution of the CMSz will be dynamically explained based on our previous ad-hoc experimental results and combined with some of the natural observables (tectonic data and reconstructions are extensively described in Faccenna *et al.* 1999, Faccenna *et al.* 2001a, Faccenna *et al.* 2001b, Faccenna *et al.* 2004, Faccenna *et al.* 2005, Faccenna *et al.* 2007). To qualitatively visualize the proposed evolutionary scenario, we additionally present a novel regional model that merges the distinct modeling components described in the previous section. In particular, this gravity-driven model evolves in a restricted mantle convection (i.e., impermeable upper-lower mantle discontinuity), and it is characterized by an initial laterally heterogeneous (i.e., non-cylindrical) geometry resembling the Mesozoic paleogeographic setting generally agreed upon for the Central Mediterranean (Figure 11 a,d), with a small trapezoidal oceanic basin trapped between continental shoulders (Le Pichon 1982, Malinverno & Ryan 1986).

Figure 11. Evolution of a regional model characterized by a laterally heterogeneous geometry resembling the CMSz Mesozoic paleogeographic setting.



Panels a-c: Top views of three stages of evolution of a regional model characterized by an initial laterally heterogeneous geometry resembling the Mesozoic paleogeographic setting generally agreed upon for the Central Mediterranean. Pinkish and brownish silly putties are representative of continental and oceanic lithosphere, respectively. The lithosphere is locally pre-cut as analog of the weak region enhancing the formation of the slab window. Panels d-f: Tectonic reconstruction of the evolution of subduction and back-arc extension in the central Mediterranean (from Faccenna et al. 2004, 2007). Three crucial phases are shown: 35 Ma, 15 Ma, present. Red and yellow squares indicate subduction-related and anorogenic volcanism, respectively.

Phase I (50?- 35 Ma): SUBDUCTION INITIATION

The first signature of subduction (not modeled in the regional experiment but extensively described above) is experimentally marked by a rapid and localized uplifting of the continental plate. The natural prototype is represented by large-scale disconformities and a sedimentary hiatus sandwiched within passive margin sequences (Cohen 1982), and it is accompanied by flysch deposition and high pressure-low temperature metamorphisms, as extensively recognized in the Central Mediterranean during the early Paleocene (e.g., Jolivet *et al.* 1998, Rossetti *et al.* 2004, Vignaroli *et al.* 2008). The age of initiation of the westward dipping subduction is difficult to establish in the Central Mediterranean and several contrasting models have been proposed. Several models, in fact, proposed that the present-day westward subduction initiated after the break-off of a previous eastward dipping subduction. The timing of this event has been put at the Cretaceous–Tertiary boundary (65 Ma; Dercourt *et al.* 1986), the Palaeocene (50 Ma; Boccaletti *et al.* 1971) or the Oligocene (30 Ma; Doglioni *et al.* 1997). Here, we will focus only the process concerning the present-day westward subduction history.

In our models, the lithospheric subduction can be enhanced by either the slow plate convergence or the lateral pressure produced by the collapse of the continental margin. Both mechanisms are present in the Central Mediterranean (e.g., Le Pichon 1982, Dewey *et al.* 1989), resulting in the critical amount of gravitationally unstable lithosphere that is dynamically necessary to overcome the resisting forces and, in turn, to actively start the subduction process. Paleo-reconstructions (Beccaluva *et al.* 1989, Faccenna *et al.* 2001b) support this interpretation and indicate that 300-400 km of cold lithosphere subducted with a rather shallow dip in this period, reaching a depth of 150 km. The slab deepening allowed the development of arc volcanism, as recognized in the Sardinia-Provençal regions, with products dated up to 38 Ma (Lustrino *et al.* 2009). Considering the few mm/yr (<1 cm/yr for the normal component of convergence), this amount of subduction should be achieved in at least 10 Ma. The north-westward subduction should then started at least 45 Ma, more probably around 50 Ma (Lustrino *et al.* 2009). In agreement with this reconstruction, the age of HP-LT metamorphism in Calabria spans over an interval of age ranging between 45-50 Ma to 30 Ma (Rossetti *et al.* 2004).

After the transition from a Rayleigh-Taylor instability to a self-sustained subduction, the NW-dipping CMSz extended for more than 1500 km from southern Iberia to the Ligurian region, consuming the land-locked Jurassic oceanic basin (Le Pichon *et al.* 1988). This geometry was used in our regional experimental model (Figure 11a-d).

Phase II (30 - 16 Ma): SUBDUCTION INTO THE UPPER MANTLE

Our modeling results predicted that just after the initiation of subduction, the rollback exponentially increased during the slab's movement into the upper mantle because of the progressively higher slab-pull driving action. This mechanism likely happened in the CMSz, enhanced by a slightly decreased African rate of motion (Jolivet & Faccenna 2000). An additional minor contribution could have been offered from the peculiar paleogeographic setting of the area, allowing for the subduction of a gradually narrower slab (Figure 11a). As we experimentally showed, trench velocity was inversely related to the slab width (see also Morra *et al.* 2006 and Schellart *et al.* 2007). Tectonic reconstructions support the hypothesis of an exponential increase in the rollback velocity for the CMSz. We propose that the maximum rates were reached when the slab arrived at the southern Sardinia margin (Faccenna *et al.* 1999 The rapid retreat of the slab favors the opening of back-arc basins (Uyeda & Kanamori 1979, Dewey 1980), and is a mechanism commonly proposed to explain the origin of the Central Mediterranean back-arc basins (e.g., Malinverno & Ryan 1986; Royden 1993; Faccenna *et al.* 2001b). A first episode of extension, which led to the formation of the Liguro-Provençal basin (Figure 1), occurred from 30-16 Ma and was accompanied by a 25-30° counter-clockwise rotation of the Corsica-Sardinia-Calabria block (van der Voo 1993).

In this phase, the continental lithosphere was localized at the shoulders of the subducting oceanic basin arriving at the trench (Figure 11; the Apenninic and the Maghrebian foreland thrust belts; e.g., Dercourt *et al.* 1986). Experimental data illustrated how, under these conditions, part of the buoyant lithosphere sank in the upper mantle that was being pulled at depth by the lower oceanic (i.e., heavier) part of the slab. This process experimentally produced the slowdown and verticalization of the slab (Figure 10c-d). We can infer that a similar mechanism occurred in the CMSz after the scraping of the crustal layers and their delamination from the denser, deeper mantle material (Serri *et al.* 1993, Jolivet 1999, Calvert *et al.* 2000). This speculation is supported by a detailed tectonic reconstruction made along an E-W cross-section from the Apennines to the Gulf of Lyon, recognizing that the subduction velocity progressively decreased in time until now (Faccenna *et al.*, 2001b). An additional constraint is offered by the seismic signature of the current northern Apenninic slab, which illustrates an activity limited to shallow depths (90 km, (Selvaggi & Amato

1992) and an almost continuous/vertical velocity anomaly dipping toward the W-SW (e.g. Spakman *et al.* 1993, Lucente *et al.* 1999, Piromallo & Morelli 2003).

The laterally heterogeneous setting is likely the main origin of the Calabrian orocline and allowed for differential horizontal movements of neighborhood lithospheric blocks (see Cifelli and Mattei in this volume for further information on the argument). In our view, it is interesting to highlight how the first smooth experimental arcs were formed in front of the oceanic seaway at the beginning of this phase (Figure 11). These data are in agreement with Gattacceca & Speranza (2002) who found that a significant counter-clockwise rotation of the Apennines occurred just after the initial opening of the Liguro-Provençal basin. However, a more significant counter-clockwise rotation of the Apennines has been recognized only from the early Pleistocene (Sagnotti 1992, Scheepers *et al.* 1993) , when, under our interpretation (and qualitatively recorded by our regional model; Figure 11), the northern block was considerably slowed by continental subduction differentiated from the southern oceanic fast-retreating lithosphere.

Phase III (16 - 10 Ma): SLAB INTERACTION WITH THE 660-KM DISCONTINUITY

Our models showed that once the slab started to interact with the 660-km discontinuity, the subduction slowed for a significant amount of time. However, this happened only when the viscosity contrast was larger than 10 and imposed between the upper and the lower mantle layers (see Funiciello *et al.*, 2003 for further details on vertical unconstrained systems), inhibiting the direct penetration of the descending slab. The restricted mantle convection could be a key reason for the peculiar behavior recorded in the CMSz between 16 and 10 Ma. In this interval, it has been recognized that no noteworthy changes took place on the surface (see Faccenna *et al.* 2001b and references therein) independently of possible kinematic variations produced by the African plate, which gradually continued its slow convergence (Figure 2). The hypothesis is further strengthened by the following: a) tomographic images showing that the deepest portion of the slab is currently stagnant on the 660-km discontinuity (e.g., Spakman *et al.* 1993, Lucente *et al.* 1999, Piromallo & Morelli 2003); b) paleo-reconstructions, which start from the present-day configuration and adopt a balancing technique constrained by natural observables, recognized that the slab should have approached the upper/lower mantle limit during this interval (Faccenna *et al.* 2001a). As an additional natural consequence of a nearly fixed subduction configuration, the Liguro-Provencal back-arc opening stopped (Bache et al., 2009), as did the associated rotation of the Sardinia-Corsica block (van der Voo 1993).

Phase IV (10 Ma - Today): SLAB NARROWING

We experimentally observed how a slab distributed along the upper mantle depth creates an important physical barrier to the mantle, influencing its three-dimensional circulation. In particular, we described how the slowed-down process can restart only after the mantle circulation is deeply re-organized, assuming a predominant toroidal form (Figure 6,7). A similar behavior should have characterized the CMSz in the Tortonian. The combination of the continental African (Tell and Apulia; e.g., Casero & Roure 1994, Chiarabba *et al.* 2008) lithospheric subduction at the shoulders of the system and a vigorous lateral flow enhanced the progressively lateral fragmentation of the Calabrian slab (Figure 11). Once broken, this small slab was free to retreat towards the SW independently of the rest of the eastern, larger, Algerian slab.

In our cylindrical models, the subduction behavior was steady-state from this time, as justified by the equilibrium between active and resisting forces. Moreover, the locus of the back-arc extension jumped trenchward following the current steeper configuration. The proposed experimental scenario partially resembles what was geologically recorded in the CMSz since the Tortonian. After the period of inactivity, a second extensional episode formed the Tyrrhenian basin, and a new spreading center jumped towards the trench.

However, the natural context has been complicated by the presence of short-lived episodes of oceanic accretion, which separated small back-arc basins during the early Pliocene (~ 4.3-2.6 Ma, Vavilov basin; Kastens & Mascle 1990) and the early Pleistocene (2-1- Ma, Marsili basin; Patacca *et al.* 1990, Nicolosi *et al.* 2006). Despite the inherent oversimplifications of our experimental set-up, these models offer useful suggestions that support the hypothesis that the episodicity of the Tyrrhenian back-arc spreading is related to the progressive narrowing of the Calabrian slab (Wortel & Spakman 2000, Faccenna *et al.* 2005, Chiarabba *et al.* 2008). In Figure 9, we show that an abrupt decrease of subducting plate width results in a pulse of acceleration of the trench retreat velocity, as the balance between driving and resisting forces acting on the slab is temporarily modified. A similar mechanism likely worked in nature, where the enlargement of the slab window operated discretely. In the late Messinian, the slab window resulting from the propagation of slab tearing shifted eastward about 300 km from the active portion of the foreland-thrust front (Argnani *et al.* 1987, Casero & Roure 1994). A further slab window enlargement of about 400 km was finally achieved at about 1-0.8 Ma ago (Faccenna *et al.* 2005). The N-S trending eastern plate boundary also experienced lateral propagation of slab detachment (Govers & Wortel 2005). Slab break-off initiated in the northern part of the subduction zone at 8-9 Ma as a result of the subduction of the continental lithosphere and then migrated in a southeastern direction (Wortel & Spakman 2000).

Finally, we have experimentally shown how a narrow slab experiences a vigorous toroidal mantle circulation. Under this condition, the sub-slab material turns around the slab edges and easily enters the back-arc domain because of the following: a) smaller lateral pathway; and b) faster mantle motion (i.e., the velocity of the mantle corresponds to the trench velocity, which is inversely related to the slab width). The main natural evidence of this behavior is represented by the SKS splitting pattern of the CMSz (Civiello & Margheriti 2004, Baccheschi *et al.* 2007, 2008). The NS-oriented fast directions, φ, in western Sicily have a high angle with respect to the thrust front and could be related to the lateral return flow around the western slab edge (e.g., Faccenna *et al.* 2005). The signal recorded at the eastern side of the Calabrian slab, on the contrary, follows the arcuate shape of the subducting lithosphere, which is analogous to the general trend of φ recorded along the belt. This pattern does not exclude the presence of a slab window below the southern Apennine, as detected by tomography (e.g., Faccenna *et al.* 2007)), because the anisotropic signal could be dominated by lithospheric deformation (Jolivet *et al.* 2009). In fact, the experimental test matched the orientation of the splitting anisotropy, justifying the orientation of the NS trend in western Sicily, but further tests are necessary to provide a quantitative comparison. The geochemical signature also agrees with this model (Trua *et al.* 2003) , indicating that the back-arc mantle has been progressively contaminated by inflow of fertile (African) mantle material. In the southern Tyrrhenian region, magmatism has preserved the imprinting of a mantle metasomatized by the subduction of crustal material before the lateral rupture of the slab occurred (El Bakkali *et al.* 1998, Coulon *et al.* 2002).After a while, the amount of mantle material flowing inside the back-arc region increased, changing the composition of the erupted magmas from predominantly calc-alkaline to alkaline (Faccenna *et al.* 2005). At the site of the initial break (northern Tunisia), the calc-alkaline rhyolite suite of Nefza and Galite islands (14-8 Ma, Savelli 2002) was replaced in the Upper Miocene (about 8-6 Ma) with the (Na)-alkaline basalt of Nefza and Mogodos. The enlargement of the slab window occurred in the late Messinian, further feeding the system. Consequently, the alkaline Aceste seamount was emplaced just west of the calk-alkaline Anchise seamount during the Early Pliocene (Beccaluva *et al.* 1994), while a new phase of anorogenic alkaline magmatism was also registered in Sardinia (Savelli 2002).

Phase V (Tomorrow): TOWARD A PASSIVE DEBLOBBING?

Despite the inherent oversimplifications of our regional model, it could be used to speculate about the future scenario that will likely characterize the CMSz in geological times. We qualitatively find that the reduction of the active portion of the slab beyond a critical dimension (<3.5 cm, corresponding to about 200 km) finally produces the locking of the trench. The slab becomes too small to actively subduct and passively steepens under its own weight. After the slab attains a vertical geometry, it evolves into a viscous deblobbing, facilitated by the adopted rheology of the system.

A similar evolution could be starting in the CMSz, as currently demonstrated by the lack of evidence for active spreading, when compared with the Neogene-Quaternary estimates of 50-70 mm/yr for the back-arc rate (D'Agostino & Selvaggi 2004). However, the low Africa-Eurasia convergence rates (not simulated in our model) provide an additional component to promote subduction, even under these conditions, and they only translate into subduction velocities.

Conclusions

In this paper, we highlighted how laboratory models can be used to study the subduction process. In particular, experimental results were used to relate the main geological observables available in the Central Mediterranean with the evolution of the slab into the mantle. The selected area is a key natural site because of its capability to maintain remnants of past subduction on the surface, offering the possibility of constraining the evolution of subduction for about 80 Myr.

We found that the ingredients necessary to resemble the Central Mediterranean history are the following: 1) the slow motion of the incoming African plate, dominating the first evolutionary phase, accumulated a critical amount of gravitationally unstable lithosphere to start the process; 2) balance between slab pull and lithospheric/mantle resisting forces, which, in the absence of an external kinematic engine (i.e., the low plate convergence has been always normal to the trench, minimizing its possible influence on the system), drove the exponential growth of the slab into the upper mantle and the opening of the Liguro-Provencal basin; 3) the restricted mantle convection regime and, in turn, the incapability of the slab to directly penetrate across the 660-km discontinuity, which allowed for the stagnation of the slab at depth and the episodic behavior of the trench motion/back-arc extension; And 4) land-locked paleogeographical scenarios, which enhanced the arcuature and progressive fragmentation of the subduction system.

Acknowledgments

This paper represents a brief summary of the experimental activity on subduction developed by the *Laboratory on Experimental Tectonics* in the last decade.

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