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Simulation of collision zone segmentation in the central Mediterranean

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Abstract

Collision processes between the African and European plates in the Ionian Sea offshore Calabria and in Sicily take place in several ways. To the east, normal subduction of oceanic crust (underlying the Ionian Sea) occurs underneath the Calabrian arc, whereas to the west, continental collision between the Pelagian block and the Calabrian arc (which extends from Calabria to the northeastern tip of Sicily and offshore of the island to the north) gives rise to the Maghrebian thrust belt on mainland Sicily. This belt consists mostly of stacked basinal and platform carbonates of Mesozoic–early Tertiary age deposited along the Africa passive margin. No clear subduction characterizes the area north of the Hyblean–Malta plateau where it collides with the Maghrebian chain. The origin of the Hyblean–Malta plateau is unclear. Geophysical data indicate, however, that its crustal structure is different from that of the surrounding Pelagian foreland. Crustal structure variations at the edge of the subducted African plate may thus cause the observed segmentation of the collisional arc system.

A computer model of the region was created to learn about the deformation and faulting processes in the region. The simulation starts prior to the collision of the Pelagian block with the Calabrian arc. With time the Hyblean–Malta plateau collides with the northern crustal block and, as a result, a new simulated topography, a new distribution of velocities and, most important, a new distribution of fracture zones are created. A shear fracture zone is formed east of the Hyblean–Malta plateau (Malta escarpment) extending to the subduction zone at the Calabrian arc. A less active strike-slip fracture zone is created west of the Hyblean–Malta plateau. An additional E–W trending active fracture zone is creating in the southern part of the simulated area. This feature may correspond to a fault zone which runs from the Strait of Sicily to the Ionian Sea.

1. Introduction

The collision process in Sicily and adjacent areas, central Mediterranean, is controlled by the crustal inhomogeneity of the northern edge of the African plate in front of the collision zone. In the Ionian Sea the crust is probably oceanic (Hinz, 1974; Finetti, 1982) and is underthrusting the Calabrian arc, as evidenced by the seismic activity which extends down to 400 km (Gasparini et al., 1982). In the Ionian, offshore of Calabria, thrust-top basins develop upon the external Calabrian arc and a cobblestone zone marks an incipient accretionary wedge at the trench (Ben-Avraham and Grasso, 1990 and references therein).

On mainland Sicily the northern edge of the

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Africa foreland, which is made of continental crust, collides with the crystalline rocks of the Calabrian arc, which is largely submerged and forms the offshore shoals on the southern Tyrrhenean Sea (Ben-Avraham and Grasso, 1990, 1991). As a result, a large thrust system, the Maghrebian arc, was formed in Sicily during the Neogene with thrusting prograding progressively towards the Pelagian foreland. Mesozoic and early Tertiary carbonates that were deposited along the Africa paleomargin, and Oligocene–Miocene early foreland basin terrigenous deposits were strongly deformed. Despite this simple collision scheme a certain degree of complexity is introduced by the presence of a unique crustal unit, the Hyblean plateau, in southeastern Sicily. The crustal structure of this unit is different from the surrounding African foreland. This feature is part of a larger structure, which also includes the Malta plateau on the Straits of Sicily (Fig. 1). It was suggested by Ben-Avraham and Grasso (1990, 1991) that the segmentation of the collision zone in Sicily is caused by the rheological behavior of the buoyant Hyblean–Malta plateau, which resists subduction and forms a cusp in the shape of an arc.

As the result, two large shear zones have developed on both sides of the Hyblean–Malta plateau. In the west, a broad shear zone with a right-lateral movement separates the Hyblean–Malta plateau from the adjacent foredeep, partly

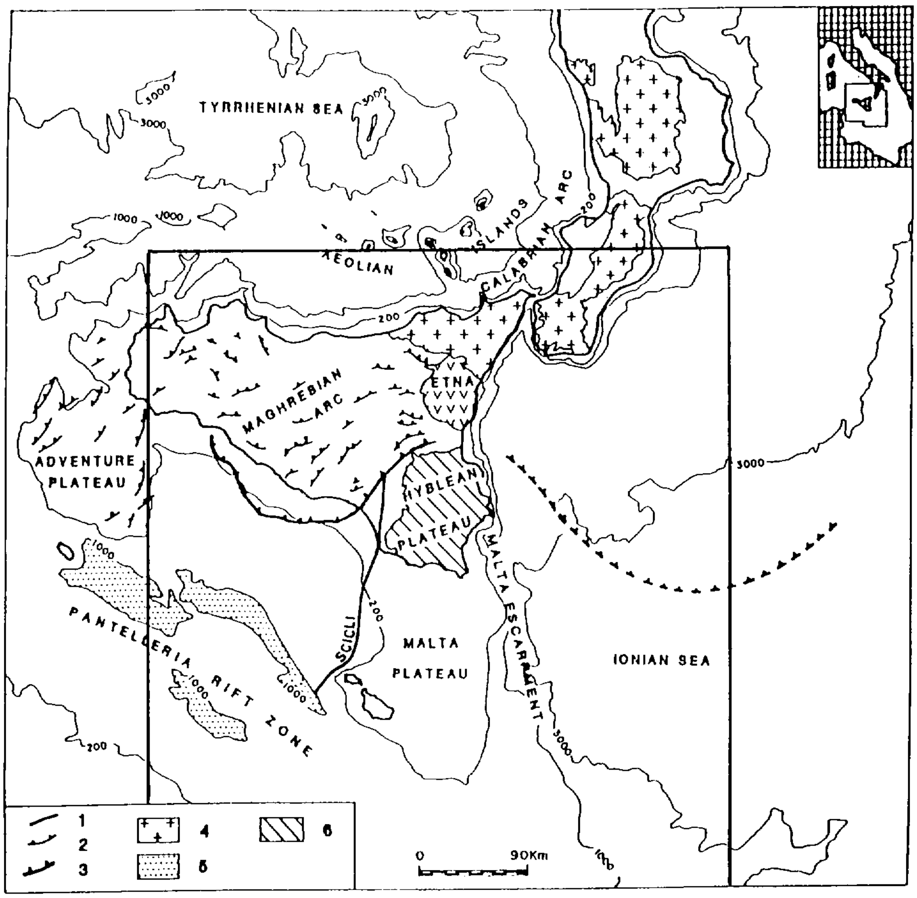

Fig. 1. Tectonic sketch map of Sicily and surrounding areas. *1* = main faults; *2* = thrust faults; *3* = main fronts of thrust belts; *4* = crystalline units of the Calabrian arc; *5* = zones of extension; *6* = Hyblean–Malta plateau. The box outlines the area of the simulation.

filled with the Gella nappe, the outermost thrust wedge of the Maghrebian arc (Butler et al., 1992). The shear system, in which the Scicli fault is the most prominent fault, connects the Pantelleria rift in the Straits of Sicily with the front of the Maghrebian thrust belt (Grasso and Reuther, 1988; Grasso et al., 1990). In the east, the Malta escarpment with large vertical displacement and minor left-lateral movement (Grasso, 1993) separates the Hyblean–Malta plateau from the Ionian basin.

In order to understand the collision processes in the central Mediterranean more clearly we simulated, using numerical analysis, the evolution of the plate boundaries in southeastern Sicily and surrounding areas in response to the existence of the Hyblean–Malta plateau. The main factor in the analysis is the variations in crustal structure in this area. Units with large differences in crustal structure; such as oceanic crust, thick continental crust and thin continental crust, are here juxtaposed with each other. In this paper we describe the principles of the analysis and present the main results of the numerical simulation.

2. Model

2.1. Thin sheet approximation

Recent studies have successfully explained many aspects of the formation and evolution of the lithosphere by application of methods of continuum mechanics (e.g., England and McKenzie, 1982, 1983; Vilotte et al., 1982, 1986; Bird, 1984, 1988, 1989; England and Houseman, 1986, 1989; Houseman and England, 1986). Fully 3-dimensional calculations of continental deformation are limited by computation resources and, therefore, only a few studies have been carried out using this technique (e.g., Ben-Zion et al., 1993; Braun, 1993). Many studies of lithospheric deformation use some approximation that reduces the mathematical problem to two dimensions; that is, thin sheet or in-plane stress approximation. This approximation assume vertically averaged rheology and thus ignores vertical variations in type, density and strength of rocks within the lithosphere. Bird (1989) recognized the importance of vertical variations of strength within the continental lithosphere and investigated the evolution of western North America using a two-layer model. The crust was split into two difference layers: upper and lower crust. This approach is computationally much more economical than full 3-dimensional calculations.

Here we assume that typical continental lithosphere consists of a number of layers. They are: a sedimentary layer, granitic and basaltic layers (upper crust and lower crust) and an upper mantle. The thickness of some of the layers in the simulated lithosphere may locally be equal to zero. Thus, a multi-layer approach makes it possible to take into consideration both continental and oceanic lithosphere with and without a granitic layer and the transition from one to the other. Under the lithosphere we assume an inviscid asthenosphere, which is a common assumption for the in-plane stress approximation.

Governing equations of the in-plane stress approximation were derived by a method of asymptotically expanded boundary-layer equations (Nayfen, 1981) for the integration of the equations of motion of the lithosphere along a vertical axis. The upper boundary (topography) was assumed to be free; that is, traction acting on this boundary is equal to zero. Tractions and velocities were assumed to be continuous on all internal boundaries. The existence of a relatively inviscid asthenosphere provides a state of zero shear tractions at the boundary between the asthenosphere and the lithosphere. The rheology of the sedimentary layer was described by a linear-viscous body; for all the other layers a visco-elastic model of a material with damage was applied, as described below. The effective viscosity of the sedimentary layer was assumed to be much smaller than the average viscosity of the crystalline crust.

Thus, we had to simulate stress distribution in two types of layers. The first type is the low-viscosity sedimentary layer and the second type is all the other layers with similar rheological properties. To describe a viscous flow velocity distribution (v_i) in the layers of the second type we used the Navier–Stokes equation. Neglecting higher order terms of small parameter $\epsilon = H/L$ (the ratio of average vertical size of the lithosphere to its horizontal size), we obtained the equations that describe the velocity and stress fields in the every layer of the model. The first approximation of the velocity components in a layer number n (v_x^n, v_y^n, v_z^n) can be represented as:

$$v_x^n = V_x(x,y)$$
$$v_y^n = V_y(x,y)$$
$$v_z^n = \left(\frac{\partial V_x}{\partial x} + \frac{\partial V_y}{\partial y}\right)(z - z_0)$$

where V_x and V_y are average velocities of the horizontal motion of the equivalent single-layer thin lithosphere with vertically averaged rheology.

Deviatoric stress components for layer number n (s_{ij}^n) are:

$$s_{11}^n = \exp\left(-\frac{\mu^n}{\eta^n}t\right)\int_{-\infty}^{t}\mu^n\left(4\frac{\partial V_x}{\partial x} + 2\frac{\partial V_y}{\partial y}\right) \times \exp\left(\frac{\mu^n}{\eta^n}t'\right)\mathrm{d}t'$$
$$s_{22}^n = \exp\left(-\frac{\mu^n}{\eta^n}t\right)\int_{-\infty}^{t}\mu^n\left(2\frac{\partial V_x}{\partial x} + 4\frac{\partial V_y}{\partial y}\right) \times \exp\left(\frac{\mu^n}{\eta^n}t'\right)\mathrm{d}t' \qquad (1)$$
$$s_{12}^n = \exp\left(-\frac{\mu^n}{\eta^n}t\right)\int_{-\infty}^{t}\mu^n\left(\frac{\partial V_x}{\partial y} + \frac{\partial V_y}{\partial x}\right) \times \exp\left(\frac{\mu^n}{\eta^n}t'\right)\mathrm{d}t'$$

with lithostatic pressure $p = -\int_{h_1}^{z}\rho g \mathrm{d}z$. η^n, μ^n are the viscosity and shear modulus of the layer number n. This form of the Maxwell constitutive relationship describes an elastic deformation for a quick processes, with respect to the Maxwell relaxation time and Newton viscous flow for a slow deformation.

These formulas allow the first approximation of the deformation of the lithosphere to be expressed by neglecting vertical gradients of horizontal velocity components in calculations of strain rate and deviatoric stress components. Using components s_{ij} and the absence of shear forces on the boundary between lithosphere and asthenosphere made it possible to achieve equations of the motion of the lithosphere:

$$\frac{\partial \tau_{ij}}{\partial x_j} + f_i = 0 \qquad (2)$$

where:

$$\tau_{ij} = \sum_{n=1}^{3} s_{ij}^n(h_n - h_{n+1}) \qquad (3)$$

and f_i = the effective internal forces caused by thickness variations in layers:

$$f_i = \rho_s g\frac{\partial h_1}{\partial x_i}(h_1 - h_5) + (\rho_1 - \rho_s)g\frac{\partial h_2}{\partial x_i}(h_2 - h_5) + (\rho_2 - \rho_s)g\frac{\partial h_3}{\partial x_i}(h_3 - h_5) + (\rho_3 - \rho_s)g\frac{\partial h_4}{\partial x_i}(h_4 - h_5) \qquad (4)$$

Eqs. (1–3) are the multi-layer analog of the one-layer approximation used by Sonder and England (1989). This type of a problem is the mathematical analog of a 2-dimensional visco-elastic flow and is solved numerically by the FLAC algorithm (Cundall and Board, 1988; Poliakov et al., 1992).

The next correction to the usual in-plane stress model is an introduction of the low viscosity sedimentary layer. The velocity of the motion in the sedimentary layer relative to the crustal velocity vector, V_i, under the sediments is (Mikhaylov, 1983):

$$v_x = V_x + \frac{\rho_s g}{2\eta}\frac{\partial h_1}{\partial x}(h_2 - z)(h_2 + z - 2h_1)$$
$$v_y = V_y + \frac{\rho_s g}{2\eta}\frac{\partial h_1}{\partial y}(h_2 - z)(h_2 + z - 2h_1) \qquad (5)$$
$$v_z = V_z + \int_{h_2}^{z}\left(\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y}\right)\mathrm{d}z$$

To take into account erosion and sedimentation during the evolution of the upper surface, the diffusion approach is applied. This approach is widely used in geomorphology and civil engineering to simulate the long-term behavior of river systems. The diffusion model has been applied to the development of river deltas in fjord and glacial lakes (Syvitski et al., 1988) and to foreland basins (Fleming and Jordan, 1989; Sinclair et al., 1991). Following Kenyon and Turcotte (1985), the equation of motion of the upper boundary (h_1) is:

$$\frac{dh_1}{dt} = v_z + \lambda \nabla^2 h_1 \tag{6}$$

The Laplace operator of h_1, with a transportation coefficient, λ, makes it possible to take into account a process of smoothing of the relief due to denudation. The value of λ was taken to be about 10^8 m^2/Ma, which corresponds to the intermediate rate of denudation.

Substituting (4) into (5) gives an equation for the partial temporal derivation of the relief:

$$\begin{aligned}\frac{\partial h_1}{\partial t} = &-V_x \frac{\partial h_1}{\partial x} - V_y \frac{\partial h_1}{\partial y} + V_z + \lambda \nabla^2 h_1 \\ &- \frac{\rho_s g\,(h_1 - h_2)^3}{3\eta_s} \nabla^2 h_1 \\ &+ \frac{\rho_s g (h_1 - h_2)^2}{\eta_s} \left[\left(\frac{\partial h_1}{\partial x} \right)^2 + \left(\frac{\partial h_1}{\partial y} \right)^2 \right. \\ &\left. - \frac{\partial h_1}{\partial x} \frac{\partial h_2}{\partial x} - \frac{\partial h_1}{\partial y} \frac{\partial h_2}{\partial y} \right]\end{aligned} \tag{7}$$

Eq. (6) is solved numerically for the known motion of the crust (V_i). The first approximation of the stress distribution in the sedimentary layer gives a pressure equal to a lithostatic pressure $p = \rho_s g (h_1 - z)$ and absence of a deviatoric components due to the low viscosity.

2.2. Rheology

The lithospheric response to applied stress depends on the rheological properties of the material and an earlier deformational history. At shallow depths in the lithosphere brittle failure is the main deformation mechanism (Brace and Kohlstedt, 1980). Brittle failure may also occur in the upper mantle of the continental lithosphere (Sawyer, 1985). England (1986) showed that, in order for continental lithosphere to stretch at a geologically relevant rate (i.e. 10^{-14}–10^{-17} s^{-1}), the temperature of the upper mantle must be such that most, if not all, of the upper mantle lies in the field of ductile deformation. Following this and other studies (e.g., Houseman and England, 1986; Sonder and England, 1989) we also use an assumption of the ductile deformation of the upper mantle and brittle deformation of the crust. Brittle failure occurs at low temperature when the differential stress reaches the yield stress of rocks (Goetz and Evans, 1979; Brace and Kohlstedt, 1980; Kirby, 1983). Byerlee (1968) found that frictional resistance of many types of rocks may be described by a simple linear fit to the normal stress. A simplified form of Byerlee's law gives a linear increase of yield stress with depth. The proportionality coefficient between yield stress and depth varies in the range 20–60 MPa/km for compression and 12–25 MPa/km for tension (Brace and Kohlstedt, 1980). This variation in the lithospheric strength due to the type of loading leads to dissimilarity in the lithospheric response to in-plane compressional and tensional forces (Karner et al., 1993). This rheology may be used for the stress field simulation of a region with a given geometry of fault zones (e.g., Bird and Kong, 1994) but fails to describe the process of new fracture zone creation.

A visco-elastic model of a material with damage (Lyakhovsky and Myasnikov, 1985; Myasnikov et al., 1990; Lyakhovsky et al., 1993) makes it possible to simulate the creation of a narrow fracture zone and strain rate localization, taking into account the rock's memory and type of loading. In order to simulate a faulting process in terms of continuum mechanics, a non-dimensional damage parameter, α, is incorporated in a visco-elastic Maxwell model. This parameter lies in the interval from 0 to 1 and describes the evolution of the medium from an ideal undestroyed solid to an absolutely destroyed material. Elastic moduli are supposed to be linear functions of α. Following the principle of maximum rate of entropy production, and by analogy to Fourier's law of heat conduction and Newtonian viscosity, the phenomenological equation for damage evolution (Lyakhovsky et al., 1993) is of the form:

$$\frac{\mathrm{d}\alpha}{\mathrm{d}t} = -C\frac{\partial F}{\partial \alpha}$$

where the free energy of a solid, $F\ (T, \epsilon_{ij}, \alpha)$, is a function of temperature, T, elastic deformation, ϵ_{ij}, and the damage parameter α; and C is a constant describing the temporal rate of the damage process.

We assume that initially the material has no cracks and is described as a linear Hookean material with elastic moduli λ_0, μ_0. The damage process starts when the shear stress, τ, attains the value defined by the Byerlee's friction law:

$$\tau = kp$$

where k is the friction coefficient and p is the normal stress. This stress distribution corresponds to the type of deformation or the ratio $\xi = I_1/\sqrt{I_2}$ ($I_1 = \epsilon_{ii}$; $I_2 = \epsilon_{ij}\epsilon_{ij}$ — the first and second invariants of the elastic strain tensor ϵ_{ij}) and:

$$\xi_0 = \frac{-2\mu_0}{\sqrt{k^2\left(\lambda_0 + \frac{2}{3}\mu_0\right)^2 + \frac{4}{3}\mu_0^2}}$$

Finally, the equation of damage evolution is of the form:

$$\frac{\mathrm{d}\alpha}{\mathrm{d}t} = -CI_2(\xi - \xi_0)$$

This equation describes the damage evolution which starts from the state of stress corresponding to Byerlee's friction law. An increase in damage causes the stress to drop. From this point of view the suggested model is an analog of the elastic–plastic model, but with a time- and stress-dependent transition from Hookean elastic behavior to the plastic regime and with rock memory.

To simulate any flow it is necessary to define an effective viscosity as a function of parameters of a solid. Power law viscosity is usually assumed for rocks in the ductile regime. Following Lyakhovsky et al. (1993), the viscosity, η, is approximated by an exponential function of the damage parameter, α:

$$\eta = A \exp(-B\alpha)$$

where A and B are constants.

Variation of α between destroyed and undestroyed zones causes changes in the effective viscosity of about 10^4 times. The viscosity of the upper crust, for instance, changes from about 10^{27} Pa · s for an undestroyed block up to 10^{23} Pa · s for a weak zone. To describe this variation the constants A and B were taken to be $A = 10^{27}$, $B = 4 \log 10$. The application of this model makes it possible to simulate the rheological transition from a ductile to a brittle regime and the process of creation and evolution of narrow zones of strain rate localization, corresponding to a high value of damage parameter; that is, the evolution of fault zones. This model has been applied for the simulation of the evolution of fracturing under tectonic stress, resulting in the appearance of seismic boundaries of a rheological nature (Lyakhovsky and Myasnikov, 1987, 1988; Mints et al., 1987) and for the investigation of a faulting process in the northern Dead Sea rift (Ben-Avraham and Lyakhovsky, 1992).

3. Results of simulation

3.1. General case

The first stage of the simulation was a numerical analysis of the collision processes of a continental fragment at a subduction zone. Two types of crust were introduced initially. These were continental crust, in the northern part of the simulated area, and oceanic crust in the south (Fig. 2). An additional block of continental crust was placed inside the oceanic crust. The continental crust is assumed to have 15 km thick granitic layer with a density of 2.7 g/cm^3 and a 20 km thick basaltic layer with a density of 2.9 g/cm^3. The density of the upper mantle is assumed to be 3.4 g/cm^3. A very thin sedimentary layer of about 0.5 km, with a density of 2.4

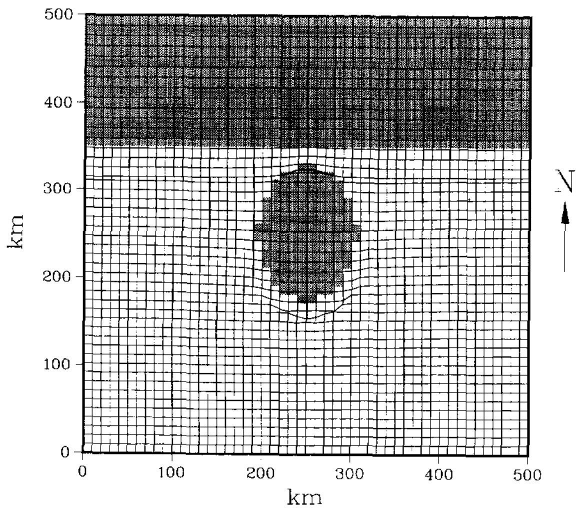

Fig. 2. Geometry of continental blocks (shaded area) used for the simulation of continental fragment accretion. The regular mesh is deformed according to the distribution of simulated horizontal velocities.

g/cm^3, initially covers the continent. The oceanic crust only has a basaltic layer about 10 km thick. A sedimentary layer about 4 km thick covers the oceanic crust. The fracture zone was initially placed at the transition between continent and ocean, where we expect subduction of the oceanic crust to occur.

The boundary conditions correspond to the northward motion of the oceanic crust. A free slip condition is enforced on the eastern and western boundaries of the simulated area. This condition permits only north–south motion along these boundaries. The northern boundary of the area is fixed and the southern boundary moves northward at a constant rate. These boundary conditions have to cause subduction of the oceanic crust. Within the framework of in-plane stress approximation it is impossible to simulate directly a downwelling motion of the oceanic plate. To make it possible for the oceanic plate to move toward the continent at a constant rate we introduced a special internal boundary condition at the subduction zone. Following the general scheme of a plate subduction we assumed a disappearance of the lower crust, which is analogous to a downwelling motion of the oceanic plate. There is no mass conservation of the lower crust in the simulated area. Less dense granitic and sedimentary layers of a terrain could not be involved into the subduction process. The upper crust and sediments are accreted to the continental crust and their mass is conserved.

The continental fragment which is embedded within the oceanic crust does not affect the stress field and the deformation around much it until the onset of its collision with the continental crust. Fig. 2 shows the geometry of the continental fragment (shaded area) and the deformation of the mesh due to the distribution of the current horizontal velocities. The deformation of the mesh in this and other figures was enlarged about 100 times to make displacements at fracture zones clearer. At some zones such a scaling results in an apparent overlap of grid points, which does not exist in a numerical scheme. With time the fragment starts to interact with the subduction zone. The accreted granitic layer produces a positive topography and stops the subduction process at the collision area. A new zone of subduction is formed at the southern boundary of the fragment, together with two large strike-slip faults east and west of it (Fig. 3).

In the next case the direction of the coastline is not east–west but at an angle to the direction

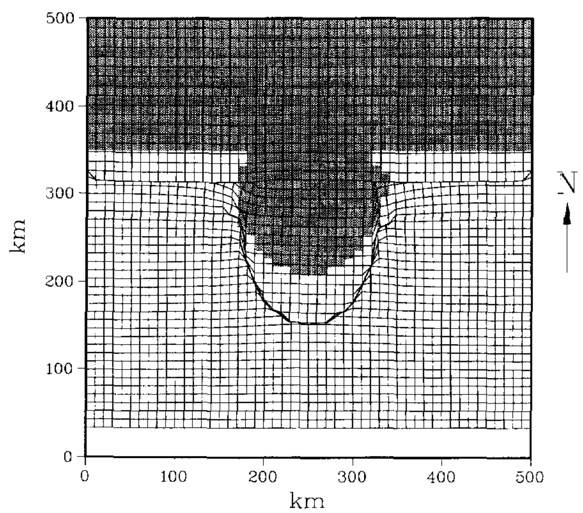

Fig. 3. Accretion of the continental fragment and formation of strike-slip faults west and east of it.

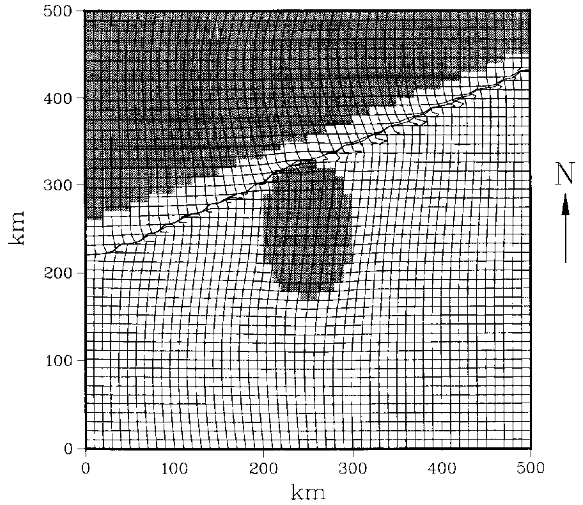

Fig. 4. Geometry of continental blocks (shaded area) used for the simulation of continental fragment accretion at a subduction zone whose trend is oblique to the direction of the northward motion of the oceanic plate.

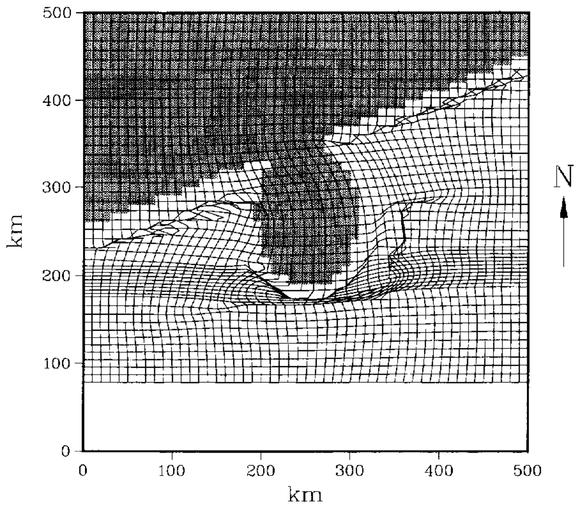

Fig. 5. Accretion of the continental fragment and the formation of zones of high deformation due to the oblique collision.

of motion of the oceanic plate. Fig. 4 shows the geometry of the continental blocks and the deformation associated with the horizontal motion at the beginning of the simulation. Shear deformation is clearly seen at the subduction zone. This shear component hardly influences the whole process. In addition to strike-slip faults, which appear west and east of the fragment at the beginning of collision, new zones of deformation extend to the east and west of the southern end of the fragment (Fig. 5). These zones are characterized by a right-lateral strike-slip motion with a small component of compression. The creation of these zones is an unexpected result of the simulation and is of great importance for understanding the collision process in the central Mediterranean.

3.2. Collision processes in the central Mediterranean

In order to simulate the structure and evolution of the central Mediterranean four lithospheric units were introduced: thick continental crustal units of the Calabrian arc and the Hyblean– Malta plateau, thin continental crust of the Pelagian block and oceanic crust of the Ionian Sea (Fig. 6). The parameters of these blocks are as follows: the thick continental crust consists of a 15 km granitic layer and a 22 km thick basaltic layer. The sedimentary layer is about 1 km thick. The thin continental crust of the Pela-

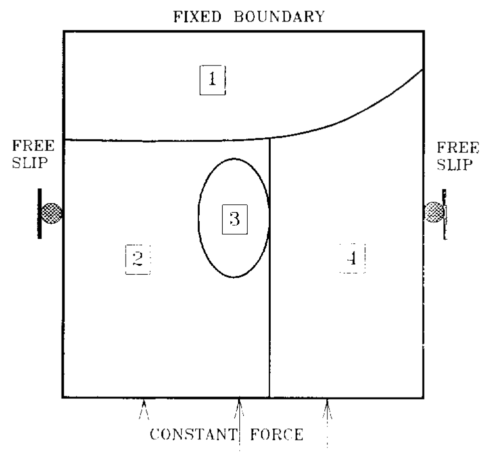

Fig. 6. Scheme of the block geometry and boundary conditions for the simulation of the collision processes in the central Mediterranean. *1* = thick continental crust of the Calabrian arc; *2* = thin continental crust of the Pelagian block; *3* = thick continental crust of the Hyblean–Malta plateau; *4* = oceanic crust of the Ionian Sea.

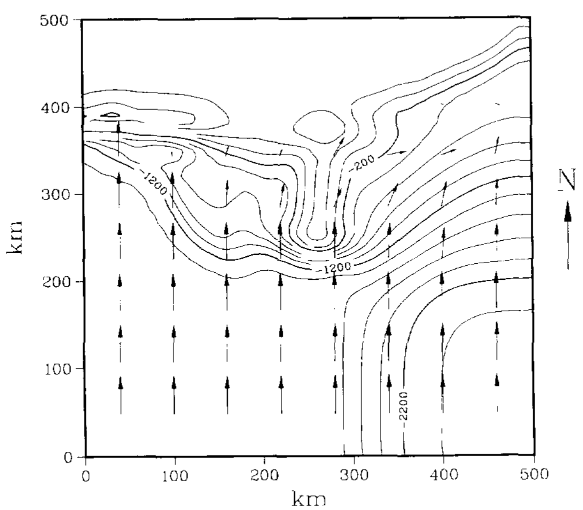

Fig. 7. Simulated topography and horizontal velocities after the collision of the Hyblean–Malta plateau with the thick continental crust north of it.

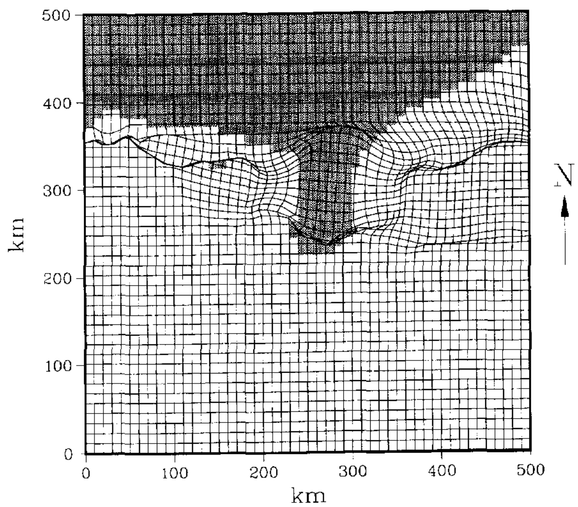

Fig. 8. Geometry of the thick continental crust (shaded area) of the Hyblean–Malta plateau and the crust north of it and the deformation caused by the horizontal velocity distribution shown in Fig. 6.

gian region has a 10 km thick granitic layer, a 10 km basaltic layer and 2 km of sediments. The oceanic crust of the Ionian Sea has no granitic layer, a 6 km thick basaltic layer and 4 km of sediments. The densities of the layers are the same as in the general case.

According to the plate tectonic model of the region (Ben-Avraham and Grasso, 1990, 1991), the boundary conditions are: fixed northern boundary, free slip condition at the eastern and western boundaries and a force acting in a northerly direction on the southern boundary (Fig. 6). These conditions and the initial geometry of the blocks cause the northward motion of the Hyblean–Malta plateau, collision of the thin continental crust west of the plateau and subduction of the oceanic crust east of the plateau. With time, the Hyblean–Malta plateau collides with the northern crustal block and, as a result, a new simulated topography and horizontal velocities are created (Fig. 7). The simulated topography shows no essential difference between the northern continental block and the accreted Hyblean–Malta plateau. The velocity distribution, shown in Fig. 7, results in the deformation shown in Fig. 8. There are two shear zones, east and west to the Hyblean–Malta plateau. The existence of these zones is typical for the process of subduction of inhomogeneous crust (Fig. 3). Apart from these zones, there are also zones of high deformation at the collision front between the Pelagian crust and the Maghrebian arc and between the Ionian

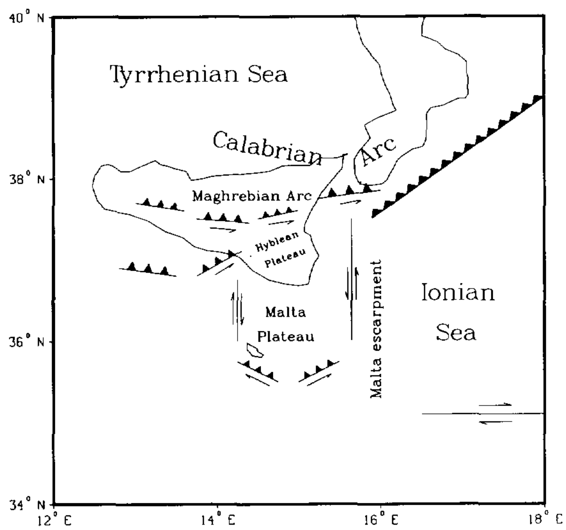

Fig. 9. Simplified map of the main tectonic features in eastern Sicily and surrounding areas based on the results of the simulation.

Sea oceanic crust and the Calabrian arc. High deformation also occurs along that segment of the Maghrebian thrust belt which is located at the transition between the continental crust of the Calabrian arc and the accreted Hyblean–Malta plateau south of it (see Fig. 1 for location). Similar to the case shown in Fig. 5, a new strike-slip zone east of the southern end of the Hyblean–Malta plateau is created.

Based on the results of the simulation, a simplified map of neotectonic collision features of Sicily and the surrounding areas is proposed (Fig. 9), which fits well other geological and geophysical data.

4. Discussion

The numerical simulation of the evolution of the plate collision zone in eastern Sicily and adjacent areas explains the origin of the main structural elements in this region. For example, the simulation clearly indicates that the formation of the western Hyblean transcurrent fault system is intimately linked with the collision of the Hyblean–Malta plateau with the Maghrebian arc. The numerical analysis also suggest that the transcurrent component along the Malta escarpment is linked with the same process.

Perhaps the most surprising result of the numerical simulation is the prediction of an east–west trending right-lateral strike-slip fault zone in the Ionian Sea, east of the southern end of the Hyblean–Malta plateau. The fault zone is also associated with compression. The existence of such a fault zone has been suggested previously on the basis of bathymetric, gravity and magnetic data (Ben-Avraham et al., 1987; Jongsma et al., 1987). Several seamounts, such as the Medina plateau and Archimedes seamount, are located along the proposed fault. These have been interpreted by Ben-Avraham et al. (1987) as push-up features along the transcurrent fault zone.

The results of the simulation of the collision processes in Sicily have implications for other plate collision zones where crustal inhomogeneities exist. As described in the general case, when a crustal inhomogeneity, such as large ocean plateau, approaches a subduction zone where the collision is oblique, in addition to large shear faults, which will develop on the margins of such a unit, a zone of deformation with a trend perpendicular to the direction of motion will be developed seaward of the unit. The deformation zone is characterized by compression and strike-slip motion. This analysis, thus, may explain the way some of the features on the seafloor in front of subduction zones have been formed. For example, the Selkirk rise, offshore South America (Mammerickx and Smith, 1980), may have formed due to the oblique collision of the Juan Fernandez Ridge with the Chile trench. Other features in the Nazca Plate may also be associated with the oblique collision of a seismic ridge with South America.

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