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Back arc extension: which driving mechanism?

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Abstract: Several hypotheses have been advanced about the genetic mechanism of back arc basins, but at present none of them is largely accepted. This work aims at recognizing which one of the proposed models may more plausibly be reconciled with the major features of subduction zones and back arc basins in the world and with the results of numerical and analogue modelling of subduction processes. Our analysis points out that the interpretations which explain back arc extension as a side effect of subduction do not provide convincing explanations for some major evidence, as the fact that back arc extension occurs in some subduction zones and not in others, that the same process cessated in zones where subduction has remained active, that the arcs associated with back arc basins are often characterized by a strongly curved shape, that arc-trench-back arc systems do not develop along the entire length of consuming borders and that no significant correlation can be recognized between any parameter of subduction processes and the occurrence of back arc extension. In addition, modelling experiments indicate that the magnitude of the tensional stress induced in the overriding plate by subduction-related forces is significantly lower than the lithospheric strength. These problems are discussed, in particular, for three subduction-related interpretations, the 'slab-pull', the 'corner flow' and the 'sea anchor' models, which seem to be the most quoted in literature. It is then argued that possible solutions of the above problems may be provided by the extrusion model, which postulates that back arc basins are generated by the forced separation of the arc from the overriding plate, along a sector of the consuming border. This separation is generally caused by the oblique indentation of strong and buoyant structures against the accretionary belt. In this view, subduction and back arc extension are not causally linked one to the other, but rather represent simultaneous effects of the lateral migration of the arc. It is pointed out that the conditions required for the occurrence of this kind of mechanism may be recognized in most of the tectonic contexts where back arc basins developed in the wake of arc-trench migrating systems. On the other hand, in the zones where the above boundary conditions are not recognized, as in the South American subduction zones, back arc extension does not occur. It is also suggested that the stop of extension in a number of basins, as the Kurile, Japan, Shikoku, Parece-Vela, Balearic and Pannonian was caused by the interruption of the boundary conditions which determined the deformation of the respective arcs.

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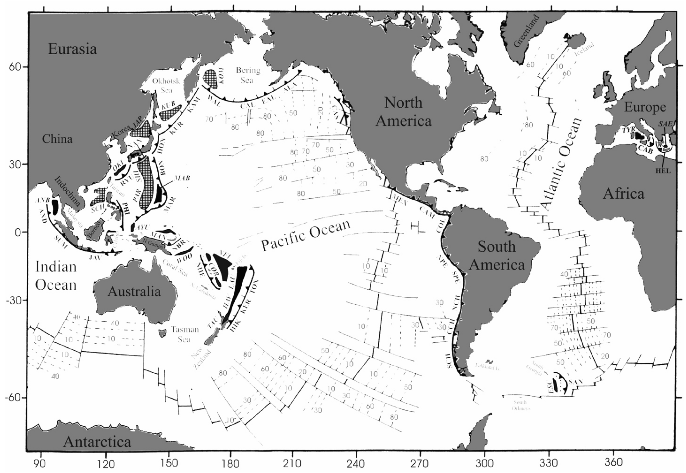
Introduction

In a number of subduction zones, the overriding plate has been affected by extensional tectonics with the consequent formation of a thinned zone, called back arc basin (e.g. Karig, 1971). The formation of this basin delineates a new structural element, here called arc, which corresponds to the portion of the overriding plate comprised between the basin and the trench. The arc is generally constituted by accretionary and/or magmatic material. In few cases, as in the western Mediterranean, the arc also included fragments of the foreland from which it detached (i.e. the Corsica-Sardinia microplate). The opening of back arc basins has repeatedly occurred in the Mediterranean area (e.g. Royden et al., 1993a,b; Mantovani et al., 1997, 2000a, 2001a), in several circum-Pacific zones and in some Atlantic zones (e.g. Karig, 1971; Uyeda and Kanamori, 1979; Taylor and Karner, 1983).

The main features of subduction zones in the world and of the possibly related extensional basins (where present) are reported in the Table. The geographical locations of these structures are illustrated in Fig. 1. In a number of cases, crustal stretching has occurred in the wake of a trench-ward migrating arc, in connection with trench retreat and slab roll back. Hereinafter, this kind of structure will be mentioned as 'trench-arc-back arc (T-A-BA) system'. In other cases, extensional tectonics has indeed occurred in the overriding plate, but the geometry of the troughs and the extensional trend do not allow to recognize them as typical T-A-BA systems. In some subduction zones, as those of South America and the Java-Sumatra one, a back arc basin never opened up.

Table 1. (View Table) Major parameters of subduction zones in the world and, where present, of the basins lying in the respective overriding plates. Some basins, as the West Philippine, Celebes, Sulu, Solomon, Coral, South Fiji, New Hebrides, Tasman have not been included in the Table since no extensional activity is clearly recognized in these zones for the last 30-40 My (e.g., Hilde and Lee, 1984; Uyeda, 1986; Tamaki and Honza, 1991) and since it is not yet clear to which subduction zones they could be associated (see Uyeda and Kanamori, 1979 and references therein). Age = Age of the subducting lithosphere. Vs = trench normal component of the absolute velocity of the slab. Vo = trench normal component of the absolute velocity of the overriding plate. For subduction zones with active back arc opening (numbers with asterisk), the motion rate of the arc is reported. A positive value indicates that the overriding plate/arc is departing from the subducting plate. Larc = length of the consuming boundary. Ls = slab length. f = slab dip. Time = time span of extensional activity. Vext = extensional rate. Text = extensional trend. The numbers reported in the two columns of references correspond to the following papers: 1) Scholz and Campos, 1995; 2) Peterson and Seno, 1984; 3) Pacheco et al., 1993; 4) Tamaki and Honza, 1991; 5) Yogodzinski et al., 1993; 6) Kusunoki and Kimura, 1998; 7) Jolivet et al., 1994; 8) Brooks et al., 1984; 9) Lee and Lavwer, 1995; 10) Uyeda and Kanamori, 1979; 11) Carlson and Mortera-Gutierrez, 1990; 12) Briais et al., 1993; 13) Tregoning et al., 1998; 14) Carlson and Melia, 1984; 15) Taylor and Karner, 1983; 16) Honza, 1995; 17) Pellettier et al., 1998; 18) Charvis and Pellettier, 1989; 19) Parson and Wright, 1996; 20) Darby and Meertens, 1995; 21) Barker et al., 1984; 22) Papazachos et al., 2000; 23) McClusky et al., 2000; 24) Giardini and Velonˆ, 1991.

Figure 1. Active subduction zones with related basins

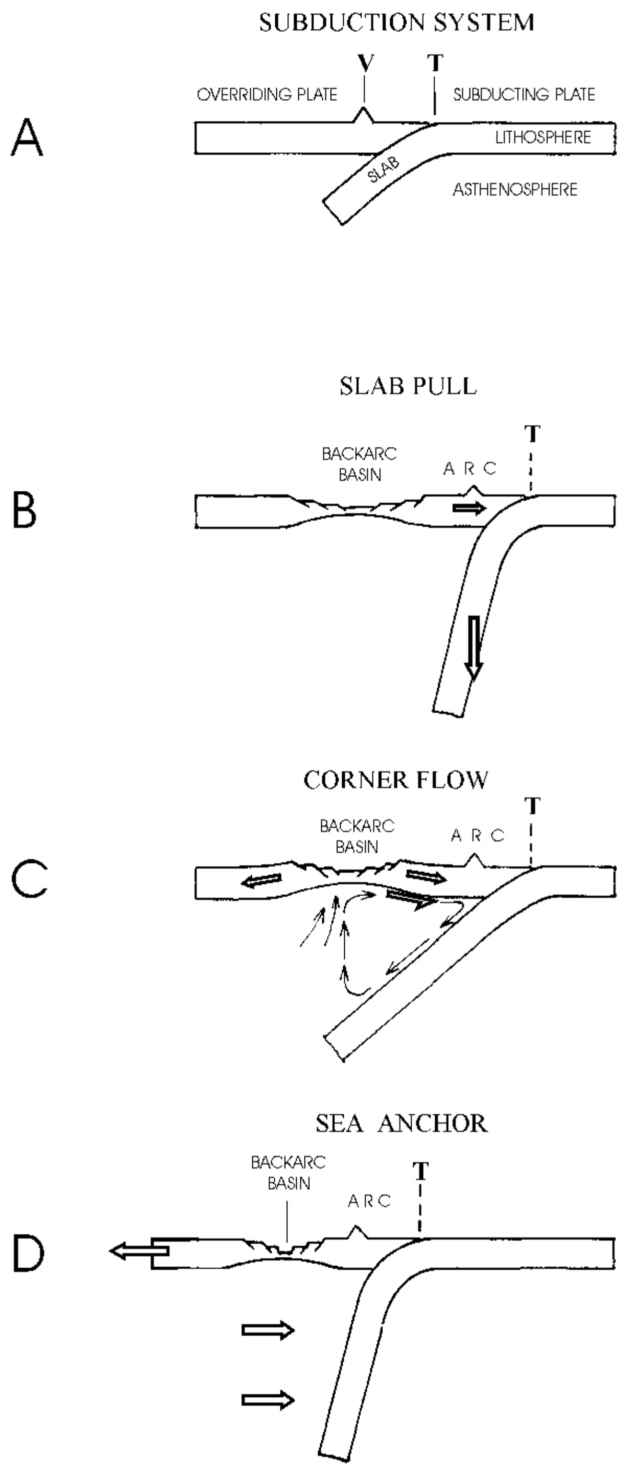


Active subduction zones (toothed lines) with possibly related basins: black indicates still active extensional basins and the grid identifies extinct basins younger than 40 My. Numbers close to the oceanic spreading axes indicate the age (My) of sea floor magnetic lineations. Abbreviations at trench zones and basins are the same used in the Table and Fig. 4. (For enlargement)

The driving mechanism of back arc opening is still matter of debate (e.g. Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Tamaki and Honza, 1991; Taylor, 1995; Mantovani et al., 1997, 2000a; Flower et al., 2001). The fact that this phenomenon is systematically associated with subduction has led many authors to believe that a causal relationship exists between the two processes and that, in particular, back arc extension is a side effect of subduction. A variety of hypotheses has been advanced about the tectonic mechanism responsible for this connection. In this work, we only focus our attention on the three of them which are the most quoted in literature, i.e. the 'slab pull', the 'corner flow'and the 'sea anchor' models.

The slab pull model (Fig. 2b) postulates that back arc extension is driven by the negative buoyancy of the subducted lithosphere with respect to the surrounding mantle (e.g. Molnar and Atwater, 1978; Dewey, 1980; Malinverno and Ryan, 1986; Royden, 1993a,b). This force would cause roll back of the slab, inducing a tensional stress in the overriding plate able to cause extensional deformation.

Figure 2. Subduction-related driving mechanisms.



Subduction-related driving mechanisms of back arc extension discussed in this work. A) Main elements of a subduction system. T = position of the trench, V = volcanic arc. B) Slab pull mechanism. The big arrow indicates the slab pull force. The arc is defined as the portion of the overriding plate which migrates trench-ward, under the action of a slab pull-induced suction force (little arrow). C) Corner flow mechanism. Thin arrows below the overriding lithosphere depict the slab-induced mantle flow. The separation of the arc from the overriding plate (and the consequent back arc extension) is driven by the gravitational collapse of lithosphere (horizontal arrows), away from the upwelled zone lying above the uprising mantle flow and by the basal drag of the arc (semi arrow), exerted by the horizontal branch of the mantle flow. D) Sea anchor mechanism. The upper arrow indicates the proposed driving force, induced by the landward motion of the overriding plate. The arc and the subduction system move slower than the overriding plate due to the resistance of the mantle to the slab's displacement (lower arrows). This differentiated motion generates the back arc basin. (For enlargement)

Slab pull model

Slab roll back and back arc extension

To interpret the generation of back arc basins as an effect of the slab pull mechanism (Fig. 2b) it is necessary to explain how the roll back of the slab may induce a tensional deviatoric stress in the overriding plate able to cause extensional deformation. A number of authors (e.g., Forsyth and Uyeda, 1975; Chapple and Tullis, 1977) invoked the existence of a trench suction force, induced by slab pull, which would cause the trenchward motion of the arc and its consequent divergence from the overriding plate. However, the results of numerical (Pacanovsky et al., 1999) and analogue (Shemenda,1993) modelling indicate that the tensional stress induced by this force in the overriding plate may range between 5 and 40 MPa, i.e. values significantly lower than the average tensional strength of the intact lithosphere, which is estimated in the range 200-260 MPa for oceanic plates (Scholz and Campos, 1995; Mueller et al., 1996) and 100-200 MPa for continents (Dunbar and Sawyer, 1989; Fadaie and Ranalli, 1990; Okaya et al., 1996). To find a tentative solution of this problem, Shemenda (1993) suggested that back arc basins could correspond to zones previously weakened by uprising of magma from mantle sources. However, this hypothesis contrasts with the fact that in most subduction zones the magmatic arc and the back arc basin are clearly separated (e.g. Sibuet et al., 1987) and that some subduction zones, as the Sunda and Chilean ones, are characterized by well developed volcanic activity, but do not present any evidence of back arc opening.

The feasibility of the slab pull mechanism is also conditioned by the length of the slab. For instance, numerical experiments (e.g. Hassani et al., 1997) have shown that the slab pull force is unable to produce extension in the overriding plate, even weakened, until the slab has reached a length of roughly 300 Km. Thus, the possibility that roll back and back arc opening are driven by a dense slab in the initial stage of its development can reasonably be excluded.

Another factor which controls the occurrence of slab roll back and of the consequent back arc extension is the mechanical coupling between the subducting and overriding plates. Numerical and analogue experiments (e.g. Shemenda, 1993, Hassani et al., 1997; Chemenda et al, 2000) point out that tensional stresses disappear from the overriding plate when the friction coefficient at the subduction fault is larger than 0.2. Considering that both observational evidence and theoretical considerations suggest that realistic values of this coefficient may range in the interval 0.4 - 0.6, due to the nearly hydrostatic fluid pressure regime in the crust (Townend and Zoback, 2000), it seems that the conditions for the occurrence of extension in the back arc zone are not generally fulfilled.

On the other hand, if extension cannot occur in the overriding plate, the roll back of the slab is inhibited as well, since the separation of the subducting lithosphere from the overriding plate is not allowed by the lithostatic pressure (e.g. Shemenda, 1993). Thus, the steepening of the slab would be the only feasible effect of the slab pull force, as the Shemenda's experiments clearly show.

Other arguments useful to evaluate the possibility that slab pull results in a simple slab steepening, instead of trench retreat and slab roll back, could be derived by comparing the energy expenses implied by these two kinds of deformation. The first pattern (Fig. 2b) would involve a number of tectonic processes, as trenchward displacement of the arc, thrusting activity at the accretionary belt, lithospheric thinning in the back arc zone, flexure of the subducting lithosphere, friction along the subduction fault, penetration of the slab into the asthenosphere and lateral displacement of asthenospheric material on both sides of the roll backing slab. Slab's steepening, instead, provides a fixed position of the trench with respect to the overriding plate and does not require any deformation of the overriding lithosphere. With respect to the first solution, this pattern involves no energy dissipation for back arc opening and for trenchward arc displacement and presumably involves less energy expense for the lateral displacement of asthenosphere at both sides of the slab. All the other sources of energy expense listed above seem to be more or less comparable in the two cases. Thus, the minimum work principle would suggest that a simple slab steepening would be the most probable effect of the slab's negative buoyancy, unless very peculiar structural-rheological conditions considerably reduce the energy budget of the tectonic processes implied by the slab pull model.

The evidence that most slabs are not vertical and that no simple relation exists between slab dip and age of subducted lithosphere (see e.g., Taylor and Karner, 1983) could imply that other forces act on the slab in addition to its negative buoyancy. These forces could be, e.g., lithosphere resistance to bending (Conrad and Hager, 1999), suction between the overriding and underthrusting lithosphere along their frictional interface (Jischke, 1975), uplift of the subducted slab due to the flow induced in the asthenosphere (Tovish et al., 1978; Turcotte and Schubert, 1982) and viscous resistance to slab migration through the mantle (Davies, 1980; Scholz and Campos, 1995). Which the resultant effect of all these (and possibly others) forces could be, is not clear yet.

Arcuate shape of arc-trench systems

Most arcs associated with back arc extension are characterized by a significant curvature (e.g. Uyeda and Kanamori, 1979; Uyeda, 1986). In the Mediterranean area, for instance, all major examples of these systems (Balearic, Tyrrhenian, Aegean and Pannonian) involved a migrating arc which started from a more or less straight initial configuration to reach a final shape where the trends of its various segments show remarkable differences, even larger than 90° (see, e.g., Dercourt et al., 1986; Royden, 1993b; Mantovani et al., 2000a). A similar behavior has been observed in several circum-Pacific T-A-BA systems (e.g. Uyeda, 1986; Lee and Lawver, 1995; Hall, 2001). In our opinion, to explain such phenomenon as an effect of slab-pull forces one must face a number of problems. For instance, one should understand why slab sinking rates are higher in the central sectors of the arc with respect to the lateral ones. This behavior could be due, e.g., to different densities in the various parts of the subducting lithosphere. However, it seems rather unlikely that such peculiar density distribution was systematically present in all T-A-BA systems of the world. Some authors argued that arc buckling might be due to the earth's sphericity (e.g. Bayly, 1982; Yamaoka, 1988). However, this hypothesis cannot explain why such effect occurred in some subduction zones and not in others (e.g. Uyeda, 1986) and, also, it can hardly account for the extremely variable curvature of arcs at consuming boundaries.

Discussions as to the plausibility of the slab pull model are also reported in other works (e.g., Taylor and Karner, 1983; Uyeda, 1986; Mantovani et al., 1997, 2000a; Flower et al., 2001).

Corner flow (or wedge flow or induced asthenospheric convection)

This model (Fig. 2c) postulates that subduction induces a convective flow in the asthenospheric wedge overlying the slab and that this flow causes extension in the overriding plate (Sleep and Toksoz, 1971; Toksoz and Hsui, 1978; Jurdy and Stefanik, 1983). This last process is an effect of two kinds of force, one is the shear traction induced by the asthenospheric flow at the base of the overriding plate and the other is the gravitational collapse of the arc away from the structural high created by the vertical uprising branch of the convective flow (Fig. 2c). Attempts at quantifying the implications of the slab-induced corner flow by numerical and analogue modelling have provided information useful to understand the reliability of this interpretation.

The uprising branch of the convective flow is expected to be located at a distance of roughly 250-300 km from the volcanic arc (Toksoz and Hsui, 1978). This result is not much consistent with the observed distances between the volcanic arc and the centre of the extensional zone, which are mostly comprised between 100 and 150 km (e.g. Taylor and Karner, 1983).

The deviatoric tensional stress induced in the overriding plate by the combination of the spreading ridge (which may rise up to 1 km above the sea floor) and the bottom shear traction is predicted in the range 10-15 MPa (Toksoz and Hsui, 1978; Jurdy and Stefanik, 1983). However, as argued earlier, these values are considerably lower than the strength values estimated for both continental and oceanic lithosphere (100-260 MPa).

Numerical models of slab-induced corner flow (Davies and Stevenson, 1992) suggest that the velocities of asthenospheric flow below the arc may only be a fraction of the imposed subduction rate. This expected pattern cannot easily explain the relative kinematics at a number of subduction zones, as in the Mediterranean area, where geological and geophysical evidence indicate a trenchward motion of arcs higher than subduction rates (Dercourt et al., 1986; Mantovani et al., 1997, 2000a) and in the Tonga zone, where geological, seismological and space geodetic observations (Fig. 8) suggest a trench ward velocity (13-16 cm/y) of the Tonga arc considerably higher than the subduction rate (9-11 cm/y) of the Pacific lithosphere.

A major problem of the corner flow model is explaining why back arc extension did not occur in several consuming boundaries and why in a number of subduction zones back arc extension has ceased, while lithosphere consumption is still going on. One could expect, in fact, that the presumed asthenospheric flow should produce similar effects at subduction zones where slabs have comparable sizes and subduction rates.

Furthermore, this mechanism cannot easily explain the curved shape of arcs and the fact that back arc extension only occurs in limited sectors of consuming boundaries. This last difficulty is particularly evident in the Izu Bonin-proto Mariana arc-trench system (e.g. Taylor and Karner, 1983) where a continuos and laterally homogeneous subduction process has been reflected at the surface by a strongly heterogeneous behavior of the arcÕs deformation. Analogous difficulties would be encountered by any attempt to apply this model to the T-A-BA systems of the Mediterranean area, where the buckling of arcs is particularly evident (Mantovani et al., 1997, 2000a).

The dynamics of the corner flow model (Fig. 2c) implies that extension occurs in the central part of an upwelling zone, like an oceanic ridge, as also predicted by numerical modelling experiments (e.g., Keen, 1985). Thus, one could try to recognize the actual occurrence of this mechanism by the analysis of the morphological features of back arc zones, also keeping in mind that extension driven by horizontal forces (as presumably occurs in passive, i.e. kinematically induced processes) provides a generalized subsidence of the back arc zone. This discrimination could be hampered by the fact that tectonic troughs are generally bounded by uplifted shoulders. However, it has been demonstrated that this effect does not imply the presence of vertical additional forces, but simply represents a normal gravitational response to the trough's subsidence (Keen, 1985; Shemenda and Grocholsky, 1994). Once removed this possible ambiguity, the morphological features of most back arc basins seem to be more consistent with passive crustal stretching (see, e.g. Sibuet et al., 1987; Park et al., 1990; Wright et al., 1996).

The corner flow model may hardly be invoked to explain the occurrence of extension in some back arc basins, such as e.g. the Mariana and East Scotia, where the sites of active opening lie beyond the seismically defined deepest limit of the subducted lithosphere (e.g. Taylor and Karner, 1983).

Discussions as to unresolved problems with this model are also reported by other authors (e.g Uyeda and Kanamori, 1979;Taylor and Karner, 1983; Uyeda, 1986; Flower et al., 2001).

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Sea anchor model

The hypothesis that viscous resistive forces in the upper mantle may oppose the lateral displacement of subducted lithosphere was initially advanced in the early stage of the plate tectonics theory (Havemann, 1972) and then reconsidered by Uyeda and Kanamori (1979) to explain the generation and distribution of back arc basins in the world. This last model was based on the hypothesis that the position of the trench is stationary, in an absolute reference frame, whatever the velocity of the subducting plate may be. Under this assumption, the deformation of the overriding lithosphere is determined by its relative motion with respect to the trench. When the overriding plate moves landward, it undergoes tensional deformation (back arc opening), since its separation from the subducting plate is prevented by the lithostatic load, that largely overcomes the tensional strength of the overriding lithosphere (Shemenda, 1993).

However, the basic assumption of this model cannot easily be reconciled with the fact that advancing, stationary and retreating subduction boundaries are now clearly recognized (Taylor and Karner, 1983; Carlson and Melia, 1984; Royden, 1993b, 1996). For instance, Uyeda and Kanamori (1979) suggested that a fully developed anchor effect is expected at the Mariana subduction system, while the fast advancing of the Mariana trench is recognized (e.g. Carlson and Mortera-Gutierrez, 1990).

The possible influence of the slab-asthenosphere interaction on back arc dynamics has been then reconsidered by Scholz and Campos (1995), who suggested that back arc extension occurs when the overriding plate moves landward, trailing the slab (Fig. 2d). The extensional stress in the back arc zone is induced by a combination of the above force and of the hydrodynamic force which resists the motion of the slab through the viscous asthenosphere (sea anchor force). The magnitude of this force depends on the trench-normal component of the overriding plate velocity, on the slab surface and on the mantle viscosity. Steady state plate motion provided by current kinematic models implies the equilibrium of forces in a subduction system, i.e. the force which drives the overriding plate must counterbalance the sea-anchor mantle resistance to maintain the observed plate velocity. Thus, the estimate of the sea-anchor effect may give information on the force that pull the overriding lithosphere landward. The proposed dynamic context may produce tensional failure in the overriding plate when the anchor force reaches the average tensional strength of the lithosphere. Under some simplyfing assumptions, Scholz and Campos (1995) estimated the sea anchor force for 29 circum-Pacific subduction zones. In the zones where back arc opening never occurred, or ceased, the values of the above force are negative (the overriding plate approaches the trench) or positive, but lower than 2.2x1012 Nm-1. In 4 subduction zones, associated with active back arc basins (Mariana, Kermadec, Tonga and Hikurangi), the above force ranges between 7.4 and 9.7x1012 Nm-1, i.e. values slightly lower than the presumed strength of the intact lithosphere (~1013 Nm-1, corresponding to a 200 MPa deviatoric stress, averaged over a 50 Km thick oceanic lithosphere). To account for the occurrence of back arc extension in the above zones, the authors supposed that back arc opening, once initiated, requires a force (3x1012 Nm-1 = 60 MPa averaged stress) considerably lower than the one acting in the initial stage. However, this does not explain how back arc extension began. Furthermore, one should consider that protracted plate thinning could strengthen, rather than weaken, the stretched lithospheric domains, due to conductive cooling of the lithospheric mantle (Sonder and England, 1989; Ruppel, 1995).

As Scholz and Campos admit, the sea-anchor model fails to explain back arc activity observed in the Ryukyu and New Hebrides subduction zones, for which the estimated driving force is close to zero. The above authors also admit that the occurrence of back arc extension cannot be related to large-scale plate motion only, but it must also depend on local factors.

The implications of this model cannot easily be reconciled with the kinematics of the Mediterranean T-A-BA systems, since the development of these zones was mainly determined by the seaward motion of arcs, with very slow convergence rates between the overriding and subducting plates (e.g. Biju-Duval et al., 1977; Dercourt et al., 1986; Mantovani et al., 1997, 2000a).

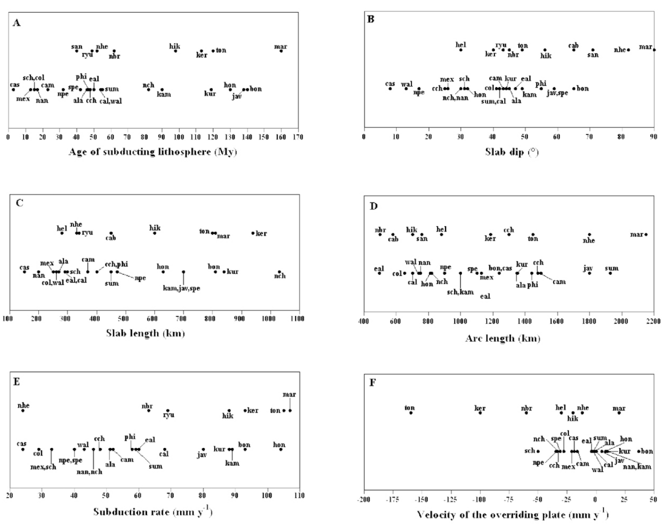
Discussions as to unresolved problems with this model are also reported by other authors (e.g. Taylor and Karner, 1983; Uyeda, 1986; Flower et al., 2001).

Observed features of subduction processes

The fact that several subduction zones are not associated with back arc basins and that in a number of consuming borders back arc extension came to an end while subduction continued for millions of years, clearly indicates that subduction is not a sufficient condition for back arc extension. To overcome this difficulty, one could try to identify some correlation between the occurrence of back arc extension and one or more of the features of subduction processes, as tentatively suggested by Uyeda and Kanamori (1979). In fact, if the subduction process was in some way responsible for the opening of the back arc basin, one could reasonably expect that the related driving force depends on the features of the slab, as suggested by the results of theoretical and modelling quantifications (e.g. Scholz and Campos, 1995; Ranalli et al., 2000).

To provide evidence about this problem we have reported in a number of diagrams (Fig. 4) the distribution of 'back arc' and 'non back arc' subduction zones with respect to the major features of the related subduction systems. It can be noted that these diagrams do not show any significant correlation between the occurrence of back arc extension and particular values of the parameters considered.

Figure 4. Active back arc opening in relation to major slab features.



Distribution of subduction zones associated (dots in the upper row) or not associated (dots in the lower row) to active back arc opening with respect to some major slab features. The papers from which the related data have been taken are reported in the Table. The average dip of the slab (B) is defined as suggested by Scholz and Campos (1995). The subduction rate (E) is the trench-normal component of the absolute velocity of the subducting plate, measured in the hot-spot reference frame. In F) velocities in the upper row refer to the absolute velocities of arcs in T-A-BA systems (see the Table). The subduction systems are identified by the abbreviations given in the Table. (For enlargement)

Non subduction-related driving mechanisms

As said before, this class of interpretations is mainly constituted by extrusion and pull apart mechanisms. The effects of these two kinds of mechanism on the structural and morphological features within and around back arc zones are generally well differentiated. The first type (extrusion) is mainly characterized by an arc undergoing a progressive deformation (buckling) and a back arc basin which progressively opens in the wake of the migrating arc (T-A-BA system). In this context the extensional trend of the basin is roughly parallel to the arc's motion trend, i.e. roughly perpendicular to the trench. The second type of mechanism, instead, requires the presence of a major shear zone between two plates and the direction of extension in the basin is more or less parallel to the relative motion of the two plates.

Extrusion model

This mechanism, whose basic features are sketched in Fig. 3, is expected to occur along a consuming border where a sector of the accretionary belt is deformed by an extrusion mechanism and partly separates from the overriding plate. This separation is accommodated by crustal stretching in the back arc zone. Simultaneously, the outward migration of the deforming belt (arc) causes the roll back of the slab lying in front of it (Fig. 3B). The tectonic context which produces the deformation of the arc may be quite variable from case to case. Most often, this phenomenon occurs when a buoyant structure enters a sector of the consuming border, with a direction of motion not perpendicular to the trench. In this oblique constrictional context, the accretionary belt undergoes a longitudinal shortening, which is accommodated by lateral bending/extrusion, at the expense of the adjacent lithosphere (Fig. 3).

The occurrence of this mechanism requires that the buoyancy of the accretionary belt is significantly higher than that of the lithospheric domain lying in front of it. Thus,for instance, the lateral extrusion of the arc is strongly favoured when it faces old oceanic lithosphere, since such kind of lithosphere is presumably characterized by very low, or even negative, buoyancy (Cloos, 1993). This might explain why even the presumably limited loading of a poorly developed arc structure, such as the volcanic arcs of the Mariana and Tonga zones, may have caused the roll back of the adjacent slabs in the related arc-trench systems. Another basic requisite for the occurrence of the proposed mechanism is a rigid behavior and a limited fracturation of the belt, which allows the formation of few relatively large crustal wedges sliding and rotating each other, as shown e.g. in Fig. 3. If this condition is not fulfilled, the highly fragmented extruding material tends to occupy the entire space available and, thus, it does not allow the separation of the arc from the overriding plate and the consequent back arc extension.

To help the description of the proposed model, some examples of T-A-BA systems, which might have formed by the extrusion mechanism, are shown in Fig. 3.

The possible importance of extrusion processes in the generation of back arc basins has been already stressed by a number of authors (e.g. Tapponnier, 1977; Tapponnier et al., 1986; McCabe, 1984; Uyeda, 1986; Lavé et al., 1996; Mantovani et al., 1996, 1997, 2000a, 2001a). The physical plausibility of this kind of mechanism has been demonstrated by analitical computations and by analogue and numerical modelling (Tapponnier et al., 1982; Peltzer and Tapponnier, 1988; Ratschbacher et al., 1991; Faccenna et al., 1996; Mantovani et al., 2000b, 2001b). In the following, we argue that the implications of this model may provide plausible solutions for the outstanding problems of the subduction related interpretations we mentioned before.

The lack of back arc extension in several subduction zones may be barely explained by the fact that the conditions required for the occurrence of the extrusion mechanism were not present at those consuming boundaries. For instance, along the south American subduction zones there is no evidence of lateral extrusion processes along the trench-arc system. This is coherent with the fact that the lithospheric domain entering this trench-zone presents quite a normal oceanic character all along the respective consuming boundaries, unlike what happens in a number of western Pacific subduction zones (see section 6.1.2) and with the fact that plate convergence is perpendicular to the consuming boundary.

The evidence that some areas of former back arc extension are now inactive even though they remain adjacent to active subduction zones, as, e.g. the Japan and Kurile basins, might be due to the fact that at a certain evolutionary stage new boundary conditions, no longer favourable to the occurrence of the extrusion mechanism, began to affect the respective arc-trench systems. This tectonic event may be caused, for instance, by the arrival of a buoyant domain in the sector of the trench zone facing the migrating arc or by the fact that the deforming arc has reached such a configuration, with respect to the dynamic boundary conditions, to inhibit any further deformation. Of course, the effective reliability of these potential explanations must be checked for each T-A-BA system (see the discussions intext).

The strong curvature of arcs in T-A-BA systems seems to be a plausible consequence of the arc-parallel compression they are supposed to undergo in the extrusion model (Fig. 3). Later in this work it is argued that this kind of kinematically induced dynamic conditions might be recognized in the tectonic contexts where back arc extension occurred.

The extrusion model may explain why back arc basins are systematically associated with subduction, without invoking a causal relationship between the two processes. In fact, this model implies that lithosphere consumption and back arc extension are both side effects of a third process, i.e. the forced outward migration of the arc (Fig. 3). Slab roll back is caused by the push and gravitational load of the advancing arc onto the margin of the subducting lithosphere, while back arc extension is produced by the divergence between the arc and the overriding plate.

Another major feature of back arc basins which may be accounted for by the extrusion model is the fact that crustal stretching and arc buckling do not occur along the entire length of a convergent plate boundary, but only develop along a sector of it (Uyeda and Kanamori, 1979). These processes, in fact, are expected to only occur in the limited zone where the arc has separated from the overriding plate (Fig. 3). The geometry of the back-arc basin is thus controlled by the kinematics and nature of the indenting buoyant block, by the original configuration of the accretionary belt and by the dimensions of the oceanic domain lying in front of the extruding arc. This last factor, for instance, had a crucial influence on the deformation pattern of the Calabrian Arc-south Tyrrhenian and Carpatho-Balkan systems (e.g. Mantovani et al., 1997, 2000a).

A major difference between the Mariana type and the Chilean type subduction zones is the dip of the slab, almost vertical in the first type and nearly horizontal in the second type (e.g. Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Scholz and Campos, 1995). This difference seems to be confirmed by the distribution of subduction zones versus dip angles (Fig. 4b), even though intermediate values of dip angles (40-60°) may characterize both types of subduction. Anyway, to explain the striking difference of slab dips beneath the extreme examples of the two types of subduction, one could consider the rather different rheologies of the mantle zone through which the slab must penetrate and eventually deform in the two cases. Various kinds of geophysical investigations have indicated that the asthenospheric layer is much more developed beneath oceanic domains than under continents (e.g. Pollack and Chapman, 1977; Artemieva and Mooney, 2001 and references therein). This implies that the viscous resistance forces which act on a slab dipping under a continental block (like the South America plate) are much higher than those acting in the 'softer' mantle underlying a T-A-BA system, like the Mariana one. Consequently, the steepening of the slab, driven by its negative buoyancy, is much less resisted beneath a Mariana type than beneath a Chilean type arc. This could explain why the Mariana slab has already reached an almost vertical configuration after a relatively short time life,whereas the Chilean slab is still dipping at a very low angle after a much longer time life.

In this regard, it is useful to point out that the key tectonic process which determines the asthenospheric environment through which the slab penetrates in the Mariana type subduction zones, is the separation of the arc from the continental plate. In fact, due to this separation, subduction must occur some hundreds of km away from the continental margin and, thus, the slab can penetrate and move through the large volume of asthenospheric material which has been attracted by the lithospheric thinning in the back arc zone.

Further insights into this problem may be gained by considering the features of the Calabrian and Hellenic subduction zones. The slabs under these arcs present quite different dip angles (65 and 30° respectively), in spite that the subducting lithosphere (the Ionian-Levantine oceanic domain) is the same in the two zones (e.g., Dercourt et al., 1986; Finetti and Del Ben, 1986). The difference of dip angles could be explained, for instance, by the different rheology of the mantle through which the respective slabs must penetrate and move. Beneath the Calabrian arc and the southern Tyrrhenian basin one can reasonably expect the presence of a soft mantle, due to the relatively large extension of the stretched zone, whose formation has certainly induced an abundant flow of asthenospheric material from the surrounding mantle. This hypothesis is strongly supported by geophysical investigations in the Tyrrhenian area, which indicate an oceanic-like lithosphere and a well developed asthenospheric mantle (Mele, 1998; Morelli and Piromallo, 2000; Martinez et al., 2000). Beneath the Aegean zone, instead, crustal stretching is very limited, being confined to the small Cretan sea, where a crustal thickness of roughly 20-25 km is estimated. In the remaining Aegean area, a continental character of the lithosphere is widely recognized (e.g. Makris, 1978; Meissner et al., 1987; Papazachos and Nolet, 1997).

The tectonic pattern implied by the extrusion model could also provide the ground for explaining another basic feature of T-A-BA systems, i.e. the weak seismicity observed at this kind of trench zone, like the Mariana one, with respect to the very strong earthquakes recorded at the Chilean type subduction zones (e.g., Uyeda and Kanamori, 1979; Scholz and Campos,1995). To this purpose, one must consider that the dimensions of the subduction fault (controlling the magnitude of decoupling earthquakes) implied by the extrusion mechanism are expected to be considerably smaller than those of Chilean type subduction zones. In fact, in T-A-BA systems the overriding arc is constituted by relatively thin crustal wedges (Fig. 3) and, thus, the frictional interface with the descending lithosphere cannot exceed few tens of km. In the Chilean type boundaries, instead, the subduction fault may involve the entire lithosphere and, thus, the magnitudes of the decoupling earthquakes may be much larger (e.g. Uyeda and Kanamori, 1979; Carlowiczs, 1995; Wang, 2000). Another factor which could contribute to mitigate seismic activity at T-A-BA systems with respect to Chilean type subduction zones is the increased role of slab pull among the forces acting on the slab. This, in fact, could reduce the coupling between the subducting and the overriding lithosphere, with significant effects on the amount of seismic energy release (Scholz and Campos, 1995).

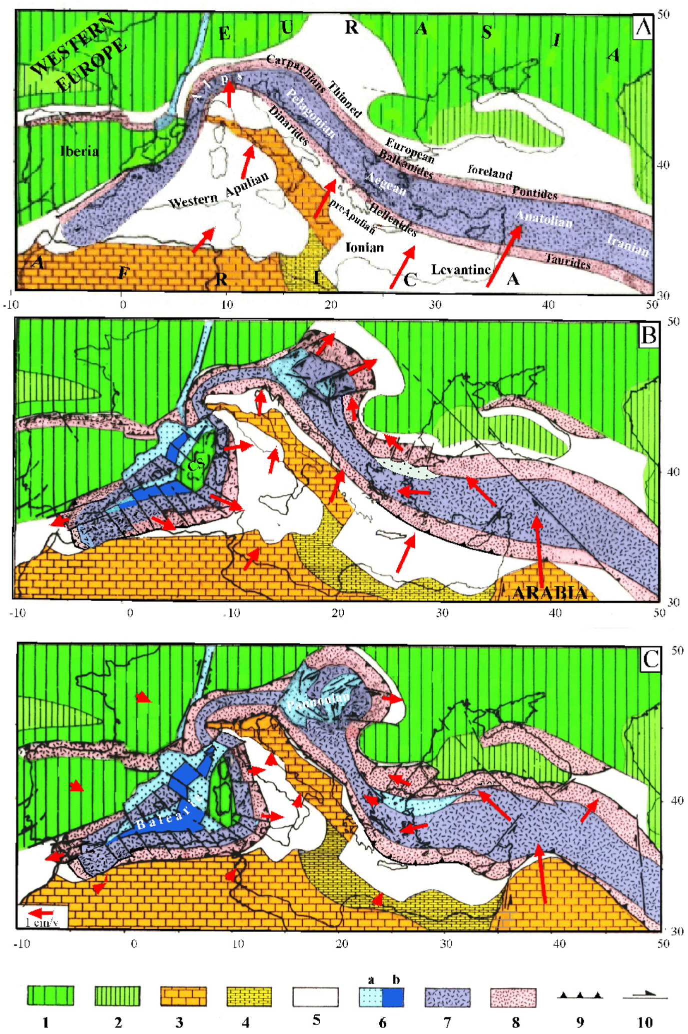
The uprise of hot asthenospheric material up to crustal levels beneath the back arc zone, implied by the extrusion model (Fig. 3), may explain the high heat flow observed in this kind of regions (e.g. Uyeda, 1986).

In this section we have discussed on how the implications of the extrusion model might account for the major features of T-A-BA systems. However, to understand the plausibility of this mechanism it would be necessary to try to recognize if the boundary conditions and structural-rheological properties required for its occurrence were present in the tectonic contexts within which back arc basins opened up. This problem is discussed in the next sections.

Mediterranean area

We start the discussion from this region since its evolution has involved a number of T-A-BA systems and also because the analysis of the deformation pattern of this area led us to believe that the extrusion model is the most plausible driving mechanism of back arc opening (Mantovani et al., 1997, 2000a, 2001a, 2002). For a detailed discussion about the large amount of evidence and arguments which may support this conviction we make reference to the above papers. Here we only point out, by the help of the proposed evolutionary reconstruction (Figs.5 and 6), that the boundary conditions and structural features implied by the extrusion mechanism might be recognized in the tectonic contexts which led to the formation of the major back arc basins in the Mediterranean area.

Figure 5. Mediterranean evolution in the Oligocene to Mid-Miocene



Tentative reconstruction of the Mediterranean evolution for the period Oligocene-middle Miocene (Mantovani et al., 1997, 2000a, 2001a). In this phase, a profound tectonic reorganization occurred in both the Western and Eastern Mediterranean regions, where the Balearic and Pannonian basins respectively opened up. Extensional tectonics also occurred in the Northern Aegean and Western Anatolian zones. 1,2,3,4) Eurasian and African-Apulian domains reported with the present configuration (corresponding to that reported in Fig. 6C). 1) and 2) respectively identify the continental and thinned parts of the Eurasian domain 3) and 4) indicate the continental and thinned parts of the African/Apulian domain 5) Parts of the Eurasian and African margins which are consumed during the successive evolutionary phases 6) Zones affected by moderate (a) or intense (b) crustal thinning 7) Orogenic belts built up by the closure of the Tethys ocean, constituted by oceanic remnants, metamorphic bodies and crystalline massifs (Tethyan belt in the text) 8) Accretionary belts constituted by units of the European or African continental margins 9,10) Compressional and transcurrent features. A) Oligocene paleogeographic setting. B) Lower Miocene, CS=Corsica Sardinia block, C) Middle Miocene. Red arrows tentatively indicate plate motions with respect to Eurasia. Arrows in orogenic belts help to illustrate the proposed deformation pattern (motion rates are merely indicative). Present geographical contours and presumed paleoposition of the present African coastal line are reported for reference in each evolutionary phase. (For enlargement)

Balearic basin

The formation of this T-A-BA system was triggered, around the upper Oligocene, by the oblique continental collision between Africa and Western Europe (Fig. 5a). After this contact, plate convergence was accommodated by the East to SEward extrusion of crustal wedges of the Alpine belt and of a fragment of the European foreland (Corsica-Sardinia microplate), at the expense of the old oceanic lithosphere in the Western Apulian zone (Fig. 5b,c). In the wake of the migrating arc, extensional tectonics developed, generating the Balearic basin. The migration of the southern branch of this arc underwent a progressive stop, as it collided with more eastern sectors (Algeria-Tunisia) of the northern African continental margin (Fig. 5c). The opening of the Balearic basin definitively ceased around the upper Miocene when the Corsica-Sardinia microplate reached its present position (e.g. Rehault et al., 1984; Dercourt et al., 1986). It is interesting to note that the morphology of the arc, suggested by geophysical investigations (e.g. Rehault et al., 1984), is constituted by a number of wedges decoupled by strike-slip faults (Fig. 5b,c), with a pattern very similar to that shown in Fig. 3b.

Pannonian basin and Aegean Miocenic extension

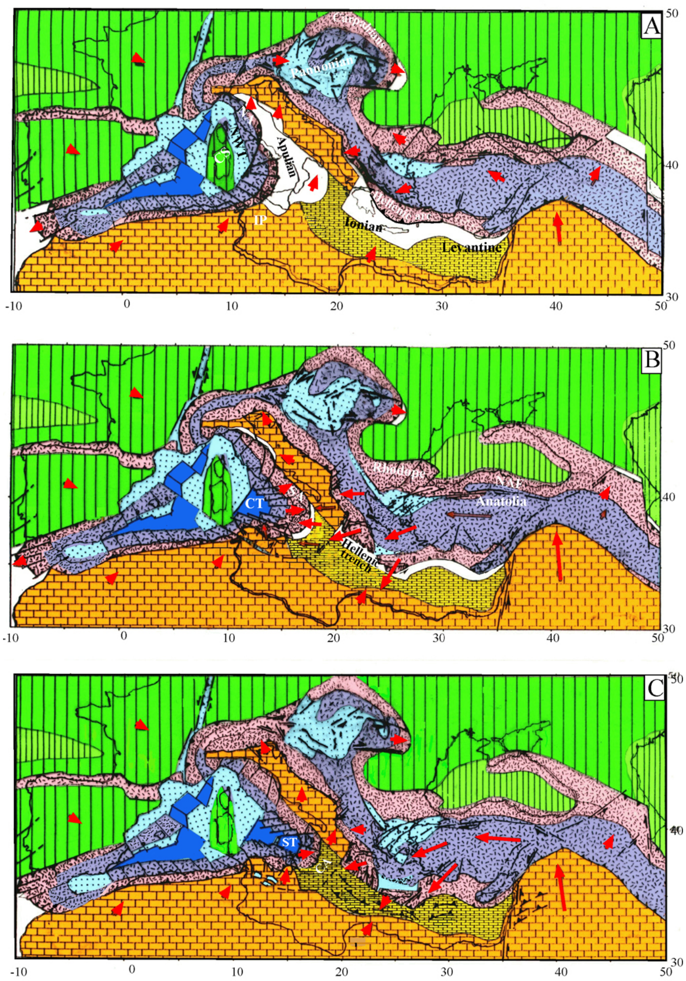
The formation of these extensional zones (almost coeval with the Balearic basin) was a side effect of the northward indentation of the Arabian promontory against the orogenic zone (Tethyan belt) created by the consumption of the Tethyan ocean (Fig. 5b,c). This belt (Fig. 5a) was constituted, in the northern part, by an accretionary chain of European affinity (Carpathians, Balkanides and Pontides), by oceanic remnants, crystallin massifs and metamorphic units in the inner part (Pelagonian, Aegean and Anatolian massifs), and by an accretionary chain of African affinity (Dinarides, Hellenides and Taurides) in the southern part.

Under the push of the Arabian indenter, the Anatolian sector of the Tethyan belt decoupled from the Iranian sector, through a system of right lateral faults (Hempton, 1987) and moved roughly NWward. This displacement was accommodated by a complex deformation of the eastern Tethyan belt, which involved a differentiated behavior of the northern chain (Carpatho-Balkan) with respect to the inner massifs and southern belt (Dinarides-Hellenides). In the Balkanides and Carpathian belt the longitudinal shortening was accommodated by the lateral extrusion of crustal wedges, at the expense of consumable zones in the southern Moesian margin and, more evidently, in the European Carpathian foreland, respectively (Fig. 5b,c). In the wake of the migrating Carpathian arc transtensional tectonics occurred in the Pannonian area. The progressive collision of the Carpatho-Balkan arc with the continental European domain caused the end of that extrusion process and of the consequent back arc extension in the Pannonian region (Fig. 5c). The inner massifs and the Dinarides belt underwent a more gentle deformation, in terms of a southward buckling, at the expense of the Ionian-Levantine old oceanic zone. The divergence between the Balkanides and the Aegean inner massifs caused the extension in the North Aegean and Northwestern Anatolian area documented by geological and vulcanological evidence (e.g. Papazachos, 1989).

Cretan basin

The evolutionary phase which led to the present configuration of the Hellenic Arc (Fig. 6) started around the late Miocene, after two major tectonic events: the continental collision between the Adriatic plate and the Aegean zone, at the outer Hellenides (e.g. Mercier et al., 1989), and the activation of the western segment of the North Anatolian fault system (Barka, 1992). These events determined a strengthening of the E-W compression on the Hellenic Arc, which accelerated its southward extrusion/buckling, at the expense of the Ionian-Levantine zone. The higher rigidity of the external belt (Hellenides) with respect to the inner massifs (Cyclades) led to the separation and opposite rotations of the eastern (Crete-Rhodes) and western (Peloponnesus) segments of the Hellenic Arc, with the consequent formation of the Cretan basin (Mantovani et al., 1997,2000a,2001a).

Figure 6. Mediterranean evolution since the late Miocene



Tentative reconstruction of the Mediterranean evolution since the late Miocene (Mantovani et al., 1997, 2000a). A profound reorganization occurred in the central and eastern regions, which respectively led to the formation of the Tyrrhenian and Aegean basins. Symbols as in Fig. 5. A) Late Miocene. NA=Northern Apennines, NWT=Northwestern Tyrrhenian basin, IP=Iblean-Pelagian zone. B) Late Pliocene. CB=Cretan basin, CT=Central Tyrrhenian, NAF=North Anatolian fault system, SA=Southern Apennines. C) Present. CA=Calabrian Arc, ST=Southern Tyrrhenian. (For enlargement)

Tyrrhenian basin

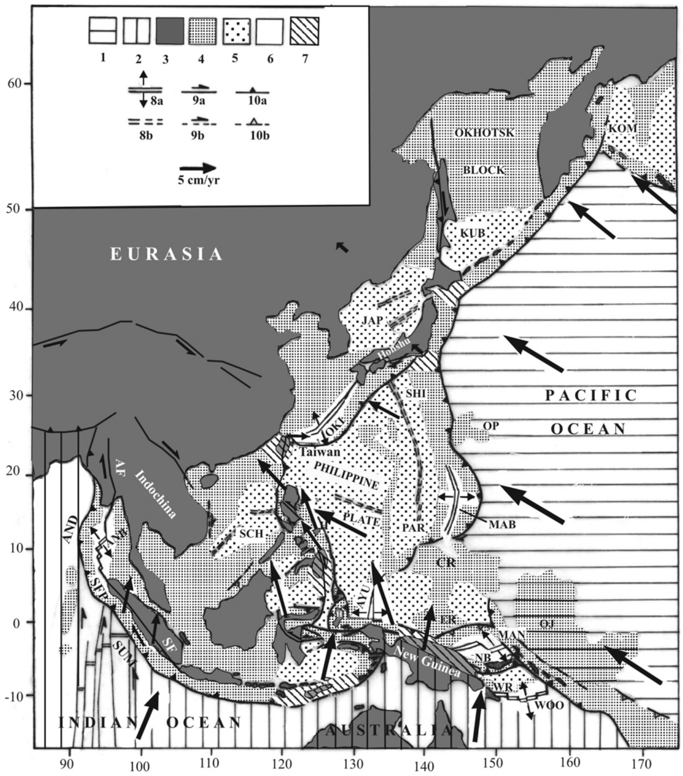
The Plio-Quaternary East to SEward migration of the Alpine-Apenninic orogenic belt, which lay east of Sardinia after the opening of the Balearic basin (Fig. 6a), was produced by an important change of the kinematics of the Adriatic plate and by its lateral effects. This change resulted from the continental collision between the Anatolian-Aegean system and the Adriatic block, occurred around the late Miocene (Mercier et al., 1989). After this contact, the Adriatic plate began a clockwise rotation, which caused the lateral expulsion (NWward) of an African fragment, the Iblean wedge (Fig. 6b). Then, in the constrictional regime induced by the convergence between the African and Adriatic blocks and the Iblean microplate, the intervening orogenic material was expulsed laterally, at the expense of the remnant part of the western Apulian zone and of the Ionian area (Mantovani et al., 1997,2000a,2001a). During the first stage (late Miocene to late Pliocene), the lateral escape of wedges was directed mainly eastward, as indicated by the extensional trend in the central Tyrrhenian basin and the features of accretionary activity in the Southern Apennines (e.g. Ortolani et al., 1979, 1992; Sartori, 1990). In the second stage (Fig. 6c) after the suture of the Southern Apennines consuming boundary (e.g. Patacca et al., 1990), the extrusion mainly involved the SEward escape of the Calabrian wedge, as indicated by the extensional trend in the southernmost Tyrrhenian basin and the features of accretionary activity in the external Calabrian Arc (e.g. Finetti and Del Ben, 1986).

The discussion on how the above geodynamic interpretations may provide plausible and coherent explanations for the complex space-time distribution of post-Eocenic tectonic events occurred in the Mediterranean area is reported by Mantovani et al., 1997, 2000a, 2001a, 2002. Numerical modelling experiments (Mantovani et al., 2000b, 2001b) have shown that a satisfactory match of the strain pattern in the central-eastern Mediterranean area, deduced by a large amount of geological and geophysical information, can be obtained by adopting the convergence of the confining blocks (Africa, Arabia and Eurasia) as the only driving mechanism of tectonic activity in this region.

Circum-Pacific zones

In this section, we describe how the tectonic conditions required for the occurrence (and the stop) of the extrusion mechanism might be recognized in the zones where T-A-BA systems developed (Figs. 7,8).

Figure 7. Tectonics and kinematics in the western Pacific region



Present tectonic setting and kinematic pattern in the western Pacific region (after Taylor and Karner, 1983; Peltzer and Tapponnier, 1988; Hall et al., 1995; Altis, 1999; Lallemand et al., 2001). 1) Pacific oceanic domain 2) Australian domain 3) land areas 4) continental margins, arcs (including the trench zone) and ridges and plateaux of oceanic domains 5) extinct basins 6) active extensional zones 7) zones of continental collision 8) spreading axes of active (a) or inactive (b) extensional zones 9) active (a) or inactive (b) strike-slip fault systems 10) active (a) or inactive (b) consuming borders. AF=Andaman fault, CR=Caroline ridge, ER=Eauripik ridge, H=Halmahera region, NB=New Britain ridge, OJ=Ontong-Java plateau, OP=Ogasawara plateau, SFP=Sumatra forearc plate, delimited by the Sumatra fault (SF) and the Sumatra (SUM) and Andaman (AND) trenches, WR=Woodlark ridge. Other abbreviations as in Fig. 1 and in the Table. Plate velocities (big arrows), in the absolute reference frame, are compatible with the kinematic models proposed by Gripp and Gordon (1990), Scholz and Campos (1995), Kreemer et al. (2000) and Michel et al. (2001). A more detailed kinematic pattern of the Pacific-Australia plate boundary, based on geodetic data, is shown in Fig. 10. (For enlargement)

Japan and Kurile basins

The formation of the Japan basin has tentatively been explained as an effect of the extrusion of the Japan arc, due to the transpressional collision between the Okhotsk block and Eurasia (Dickinson, 1978; Tapponier et al., 1982; Kimura et al., 1983; Seno et al., 1996; Kusunoki and Kimura, 1998; Altis, 1999). In this view, crustal extension occurred in the wake of the Japan arc, which was forced to bend by the above mentioned compressional boundary conditions. Altis (1999) argued that this interpretation can plausibly account for the major features of the Early-Middle Miocene deformation in Japan and surrounding zones, and that the application of an indentation-extrusion model to the Okhotsk-Eurasia collision zone allows a simpler and more coherent interpretation of the origin and development of the T-A-BA system in the Japan area, compared with the achievements of previous models. The same author also suggested that when the Japan arc began overriding the young and hot Shikoku basin (see below), in the Middle Miocene, its extrusion (and thus back arc opening) came to an end, due to the resistance that this last basin opposed to further subduction.

Other non subduction-related interpretations of back arc extension in the Japan sea have been proposed by a number of authors (e.g. Otofuji and Matsuda, 1983; Lallemand and Jolivet, 1985; Hayashida et al., 1991; Jolivet et al., 1994, 1995; Lee et al., 1999; Itoh, 2001). A discussion about the difficulties that subduction-related mechanisms may encounter in explaining the opening of the Japan sea is reported by Tatsumi et al. (1990).

The fact that the opening of the Kurile basin was more or less coeval with the one of the Japan basin (see the Table) and that the Kurile arc lay in between the Japan system and the Okhotsk block could suggest that the constrictional tectonic context which determined the formation of the Japan T-A-BA system was also responsible for the outward buckling of the Kurile arc and for the extension in the related back arc zone.

Shikoku, Parece Vela and Mariana basins

The geodynamic context which led to the coeval generation of the first two basins (e.g., Honza, 1995) might have been characterized by conditions very similar to those implied by the extrusion model:

* An arc (Izu-Bonin and proto Mariana ridges), stressed more or less parallelly to its main trend (N-S) by the convergence of the confining blocks, which were constituted, on one side, by the Australian plate moving roughly northward, and, on the other side, by the Japan Arc, extruding roughly southward (e.g., Altis, 1999).

* The presence, on the outer side of the migrating arc (Izu Bonin and proto Mariana), of a Mesozoic oceanic lithosphere (Pacific domain), playing the role of a weak lateral boundary.

Under such N-S compression, the buoyant arc was forced to migrate eastward, overriding the adjacent Pacific oceanic domain. In the wake of the migrating arc, extensional tectonics developed in the back arc zone, forming the Shikoku and Parece Vela basins (Fig. 7). This interpretation is consistent with the evolutionary reconstruction of the western Pacific area proposed by Lee and Lawver (1995). In particular, one could note that the start of extension in the above two basins coincided with the onset of bending in the Izu Bonin and Mariana arc-trench systems.

The tectonic and geologic evolution of these two regions prior to the late Miocene was essentially the same (Taylor and Karner, 1983), while in the successive evolution the Shikoku basin underwent a 25° CCW rotation, probably related to a N-S squeezing of the region (Altis, 1999) and extension has only continued behind the Mariana arc-trench system. This last change was probably connected with the collision between the Caroline ridge and the proto Mariana-Yap trench, which caused the sharp bend of the southern Mariana arc and the consequent back arc extension in the Mariana trough (McCabe and Uyeda, 1983; McCabe, 1984; Eguchi,1984). This explanation is supported by paleomagnetic, geological and geophysical observations (McCabe, 1984) and is also consistent with the evolutionary reconstruction proposed by Lee and Lawver (1995).

Okinawa trough

The formation of this extensional feature (Fig. 7) has been explained as an effect of the convergence between the Luzon arc and the East Asia continental margin at Taiwan (Letouzey and Kimura, 1985; Lee and Lawver, 1995; Huang et al., 1997). This convergence would have caused the outward migration of the Ryukyu Arc at the expense of the Pacific oceanic domain, with the consequent occurrence of crustal extension in the internal part of the arc (Okinawa trough). The evidence that the uplift of the Ryukyu Arc preceded the formation of the Okinawa trough (Sibuet et al., 1987) is consistent with the hypothesis that the above arc underwent a longitudinal shortening, in line with the extrusion model.

Ayu trough

The formation of this extensional zone has been interpreted as an effect of the divergence between the Caroline block (Fig. 7), rotating anticlockwise with respect to the surrounding zones and the northwestward moving Halmahera region. This kinematic pattern has started around the middle Miocene, when the above regions collided with the New Guinea promontory, moving northward (Lee and Lawver, 1995; Hall, 2001). A subduction-related explanation for this trough would be extremely problematic, since it can not easily be associated with any subduction zone.

Woodlark basin

Crustal extension in this zone has been interpreted as a consequence of the collision between the Ontong-Java plateau of the Pacific plate and the New Guinea continental promontory of the Australian plate (Ripper, 1982; Weissel et al., 1982; Taylor and Karner, 1983). This constrictional context was accommodated by a complex deformation pattern of the intervening zones. In particular, the most buoyant structures, i.e. the New Britain and Woodlark ridges (Figs.7 and 10), underwent bending and rotations, with the generation of extensional zones (as the Woodlark basin) in the wake of the rotating ridges and the activation of a new consuming boundary (New Britain trench) in front of the advancing arc (Lee and Lawver, 1995; Hall, 2001).

North Fiji and Lau - Havre basins

The present tectonic setting and kinematic pattern of the interaction zone between the Australian and Pacific plates is illustrated in Fig. 8. The genetic mechanism of the North Fiji and Lau-Havre basins, and of the younger Taupo rift, is still matter of debate (Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Uyeda, 1986; Hall, 2001). In literature, the North Fiji zone has been classified as a plateau, probably due to its not very deep bathymetry (e.g. Taylor and Karner, 1983), but geophysical and geological data have pointed out the recent (Plio-Quaternary) thinning of this basin (e.g., Hamburger and Isacks, 1987; Auzende et al., 1988, 1995; Honza,1995 and references therein).

\fFigure 8 full

Tectonic setting and kinematic pattern of the Tonga-Kermadec-Fiji zone

Present tectonic setting and kinematic pattern of the Tonga-Kermadec-Fiji zone (Taylor and Karner, 1983; Pelletier et al., 1998; Hall, 2001). Symbols as in Fig. 7. The numbers close to converging and diverging little arrows indicate relative plate motion rates (cm/yr) at consuming boundaries and extensional troughs, deduced by geological, seismological and geodetic data (Pelletier et al., 1998) NFFZ = North Fiji fault zone. (For enlargement)

\endfigure

Circum-Pacific zones

A number of authors (e.g. Taylor and Karner 1983; Hall, 2001) pointed out the difficulty of explaining the time-space distribution of the observed deformation in the above basins and surrounding zones as an effect of subduction-related processes. In particular, it is not clear why around the upper Miocene the New Hebrides arc started a clockwise rotation (which separated it from the Melanesian consuming boundary, also known as Vitiaz trench) up to reach its present position (Hall, 2001). This arc migration cannot be certainly associated with a subduction process, since no lithosphere consumption beneath the New Hebrides arc occurred before the onset of its rotation.

The tentative evolutionary reconstruction illustrated in Fig. 9 suggests that the generation of both the North Fiji and the Lau-Havre extensional zones might be connected with the deformation of arcs driven by plate convergence, in line with the extrusion mechanism. Fig. 9a shows the presumed tectonic setting of the zone considered in the middle-upper Miocene, mostly taken from the paleogeographic reconstructions proposed by Hamburger and Isacks (1987), Little and Roberts (1997) and Hall (2001). At this evolutionary stage, the subduction of the Pacific lithosphere at the Melanesian consuming boundary was building up an arc formed by accretionary material and volcanic products , which may be actually recognized in the Solomon-New Hebrides-Fiji-Tonga-Kermadec ridges, as suggested by Hall (2001). During this phase, the convergence between the Lord Howe and Chatham plateaux, two buoyant zones of the Australian and Pacific domains respectively, was accommodated by the consumption of the low buoyancy lithosphere comprised between them (Little and Roberts, 1997).

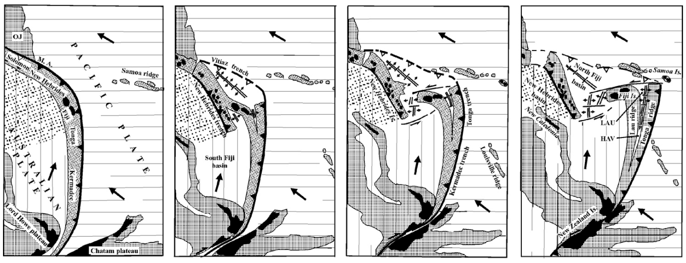
A

B

C

D

Figure 9. Deformation pattern in the old Melanesian arc



Proposed deformation pattern in the old Melanesian arc, which led to the formation of the North Fiji and Lau-Havre extensional zones, based on the evolutionary reconstructions of Little and Roberts (1997), Musgrave and Firsth (1999) and Hall (2001). Symbols as in Fig. 7. A) Configuration of the Melanesian Arc (M.A.) around the middle Miocene (Hall, 2001) B,C,D) The collision of the southernmost edge of this arc with buoyant structures of the Australian (Lord Howe plateau) and Pacific (Chatham plateau) domains caused its strong deformation, which determined the opening of the North Fiji basin, on one side, and of the Lau-Havre trough, on the other side. See text for explanations. Arrows indicate the tentative reconstruction of the kinematics of the Australian and Pacific plates. (For enlargement)

Around the upper Miocene (Fig. 9b), the incipient collision between the above domains induced an acceleration of the northward motion of the Australian plate, as suggested by the kinematic reconstruction of Gordon and Jurdy (1986). This event determined the onset of a strong deformation and disruption of the New Hebrides-Fiji-Tonga-Kermadec arc.

In the first stage (Fig. 9b), the central part of this arc (Fiji-New Hebrides) underwent a lateral extrusion (southwestward), at the expense of the low buoyancy lithosphere lying in front of it (New Hebrides basin), which caused the separation of the arc from the trench zone (Vitiaz) and the consequent formation of the North Fiji basin. The proposed deformation pattern of the Fiji-New Hebrides arc involved a double bending, which determined its horizontal delamination and the flexural separation of its inner sector (Fiji-Lau ridge) from two lateral slats (southernmost part of the New Hebrides ridge, on one side, and Tonga ridge, on the other side). Extensional deformation occurred in the zones of separation between the inner Fiji-Lau ridge and the two lateral slats (Fig. 9b), triggering the generation of the Lau sphenocasm. Evidence of Late Miocene extension in the other sphenocasm, which opened up between the southernmost New Hebrides arc and the Fiji ridge, is provided by magnetic lineations (Honza, 1995).

In the second stage (Fig. 9c), the Fiji-Tonga-Kermadec arc moved roughly eastward (under the push of the Australian block) and detached from the New Hebrides arc. This divergence was responsible for the opening of troughs in the southern part of the North Fiji basin and for the evident counterclockwise torsion of the Fiji segment of the arc. The separation between the Lau-Fiji and Kermadec ridges has then continued, with the consequent opening of the Havre trough (Fig. 9d). This divergence might also be a consequence of the fast lateral escape of the Tonga ridge, guided by the North Fiji fracture zone. A significant role in this extrusion mechanism might be played by the presence of a subducted lithospheric body under the Vitiaz trench. In particular, the shallowest part of this slab could represent an obstacle against the northward motion of the Tonga wedge, which, consequently, could prefer to extrude eastward ,at the expense of the thin Pacific lithosphere. In this regard, it could be noted that seismological investigations (Fisher et al., 1991) indicate a severe slab contortion beneath the northernmost Tonga trench.

Of course, one must be aware that the proposed evolutionary pattern only represents a working hypothesis. However, it must be pointed out that it provides a possible coherent interpretation for the very complex space distribution and time succession of tectonic events in this zone.

Scotia basin

The fact that this extensional zone has developed along the transpressional boundary between the South American and Antarctica plates, could induce to interpret this event as an effect of a pull apart mechanism. However, such hypothesis cannot easily account for the fact that the extensional rate observed in this basin (70 mm/y, Carlson and Melia, 1984; Barker, 1984) is higher than the relative transcurrent motion between the two confining plates (about 30 mm/y). This would suggest the presence of an additional driving force. Such force could be connected with the strong buckling of the South Sandwich arc (Barker, 1984; Royden, 1993b), which emphasized extensional activity in the Scotia basin, in agreement with the main concepts of the extrusion mechanism (Fig. 3).

Pull apart (or leaky transform) mechanism

This kind of mechanism has been recognized as responsible for crustal stretching in some circum-Pacific basins.

Komandorsky basin

Crustal stretching in this basin (Fig. 7) has been interpreted as an effect of a pull apart mechanism developed along the transcurrent decoupling zone between the Pacific and North American Plates (e.g. Yogodzinski et al., 1993 and references therein). The boundary conditions which caused the beginning of the above strike-slip motion were created by an important change of the Pacific motion trend (Gripp and Gordon, 1990). After this change, the Pacific plate started an highly oblique convergence with the accretionary belt (Western Aleutians) lying along the North American margin. This induced a differentiated motion between the external (trenchward) and internal sectors of the western Aleutians, causing the opening of the Komandorsky basin. This mechanism stopped around 15 My, when the migrating arc collided with the North American continental domain at Kamchatka (Yogodzinski et al., 1993).

South China basin

A number of authors suggested that the formation of this basin (Fig. 7) was a side effect of the collision of India against Eurasia and of the consequent lateral escape of Indochina (Tapponnier et al., 1982; Letouzey and Kimura, 1985; Kimura and Tamaki, 1986; Briais et al., 1993; Jolivet et al., 1994; Lee and Lawver, 1995; Chung et al., 1997). In particular, crustal extension developed in the above basin by a sort of pull apart mechanism along the Red River fault, in response to the forced separation between the Sunda block and China (Briais et al., 1993; Lee and Lawver, 1995). The end of crustal extension in the South China basin was caused by the continental collision between the Australian plate and the Sunda block, which stopped the separation of this latter microplate from China (Briais et al., 1993). During the opening of the South China basin, other troughs (as the Sulu basin) might have opened up inside the Sunda block, in response to the relative motion between microplates (e.g. Lee and Lawver, 1995).

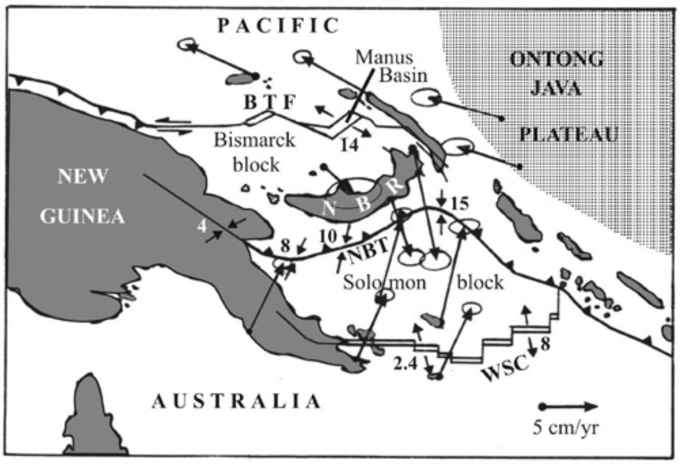
Andaman basin

Most authors (e.g. Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Lee and Lawver, 1995) recognize that this extensional feature (Fig. 7) has developed along a leaky transform segment of the megashear zone (Andaman fault) between the Indo-Australian domain and the Sunda-Indochina block. This old shear zone acted as a western strike slip guide for the extrusion of the Indochina block (50-20 My, Tapponnier et al., 1986) in response to the indentation of the Indian plate. Then, the collision of Indochina with the Sunda land and Australian blocks caused the stop of the above extrusion process. After this event, the Andaman fault system, recently prolonged through the Sumatra zone (Sumatra fault), reactivated, due to the lateral escape of the Sumatra forearc sliver plate (Fig. 7), as an effect of the oblique convergence with the Indo-Australian plate (Lee and Lawver, 1995).

Manus basin

Several authors (e.g. Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Eguchi et al., 1989; Taylor et al., 1991; Lee and Lawver, 1995; Tregoning et al., 1998) suggested that the extensional tectonics observed in the Manus basin, inside the Bismarck sea, has developed by a pull apart mechanism, along the left lateral mega-shear zone between the New Guinea (Australian plate) and the Ontong-Java and Caroline ridges of the Pacific domain (Fig. 7). This interpretation seems to be reasonable, since extension occurs along a releasing sector of a well recognized strike-slip fault system. However, it cannot account for the fact that the extensional rates inferred for the Manus basin from magnetic anomaly data (130 mm/yr, Taylor, 1979) and geodetic data (141 mm/yr, Tregoning et al., 1998) are much higher than the relative motion rate (70-100 mm/yr) between the confining plates (Figs.7 and 10). To overcome this difficulty, one could consider that the New Britain arc, being squeezed between the Ontong-Java and the Australia (New Guinea) blocks, is undergoing a southward extrusion accompanied by clockwise rotation, as indicated by geodetic data (Fig. 10). This process might contribute to emphasize the extensional deformation at the releasing sectors of the Bismarck fault system, providing, thus, a possible explanation of the observed extensional rates.

Figure 10. Boundary between the Australian and the Pacific plates



Local microplate kinematic pattern at the transpressional boundary between the Australian (New Guinea) and the Pacific (Ontong-Java plateau) plates inferred from geodetic measurements (Tregoning et al., 1998). Big arrows (with error ellipses) indicate absolute motions. Couples of converging and diverging little arrows and the nearby numbers indicate relative plate motion rates (cm/yr) at consuming boundaries and extensional troughs, respectively. NBR= New Britain ridge, NBT= New Britain trench, BTF= Bismarck trans-tensional fault system, WSC = Woodlark extensional centre.

Conclusions and discussion

In the literature, there is a widespread tendency to believe, or to admit the possibility, that back arc basins are causally linked to subduction. In this work we argue that this hypothesis is poorly supported by observational evidence and by the quantifications of the effects of subduction processes. The major questions which still lack convincing answers are:

a) why does back arc extension occur in some subduction zones and not in others?

b) why in a number of basins has subduction preceded of My the beginning of back arc extension and why in other basins has back arc activity ceased much earlier than subduction?

c) why are marginal basins often associated to strongly curved arcs?

d) why do T-A-BA systems not develop along the entire length of the convergent boundary, but only occur along a fraction of it?

e) why is the occurrence of back arc extension not correlated with any of the major parameters of subduction systems?

Furthermore, it has been argued by several authors that the implications of subduction related driving mechanisms cannot easily be reconciled with the observed space-time deformation pattern of T-A-BA systems in the circum-Pacific (e.g. Uyeda and Kanamori, 1979; Taylor and Karner, 1983; Uyeda, 1986; Tamaki and Honza, 1991; Flower et al., 2001) and Mediterranean regions (Mantovani et al., 1997, 2000a, 2001a). Then, the attempts to quantify the effects of subduction, by numerical and analogue modelling, have shown that the tensional stress induced in the overriding plate by the subduction related driving mechanisms here discussed (slab pull, corner flow and sea-anchor models) is not strong enough to cause back arc extension in the lithosphere, unless it includes zones of weakness and the interplate coupling is very low.

In general, the basic concept suggested in this work is that the slab pull force, or any other subduction related force, cannot produce the detachment of the arc from the overriding plate and, consequently, slab roll back cannot occur. This implies that the only effect of these forces may be the steepening of the slab, without trench retreat. In addition, it is argued that this last effect may only be achieved when the mantle surrounding the slab has relatively low viscosity, as occurs, e.g., beneath the Mariana arc. When, instead, this condition is not fulfilled, as under the Peru-Chile subduction zone, the slab maintains a very low dip angle, due to the high resistance of viscous forces in the mantle.

In our view, the detachment of the arc from the overriding plate may only be produced by external forces, induced for instance by the oblique indentation of a strong and buoyant structure against an accretionary belt (Fig. 3). This mechanism may provide plausible and coherent answers to the questions mentioned above and has not to face conceptual difficulties, since its plausibility can be demonstrated under given tectonic conditions. The main problem is demonstrating that such conditions have actually occurred in the zones where T-A-BA systems developed. Unfortunately, this demonstration is not easy, since there is no direct evidence on the paleo-kinematics of plates and on the rheological properties of the structures involved. These features may only be tentatively inferred and reconstructed by the analysis of as many as possible geological, geophysical and, where available, geodetic observations, which, in addition, are often affected by a poorly known uncertainty. In spite of this, we think that the presently available evidence is sufficient to consider the extrusion model an important candidate as genetic mechanism of back arc extension. For the Mediterranean area, an accurate study of the post-Eocenic deformation pattern, and of its compatibility with the implications of the various driving mechanisms so far proposed (Mantovani et al., 1996, 1997, 2000a, 2001a, 2002) led us to grow an high confidence on the hypothesis that T-A-BA systems were generated by the interaction of buoyant structures, driven by plate convergence. As regards the circum Pacific regions, we noted in literature a growing attention on the idea that back arc extension is closely connected with extrusion processes. In the text, we point out that for most of the T-A-BA systems reported in the Table it is possible to find in literature an extrusion-related interpretation which may provide plausible explanations of the observed features. When this is not possible, as for the Lau-Havre, Shikoku and Parece Vela basins, we have discussed the compatibility between the observed tectonic context and the conditions required for the occurrence of the extrusion model.

Another interpretation of back arc opening, which can be classified as non-subduction related, has been recently proposed (Flower et al., 1998, 2001). This model suggests that back arc extension in the western Pacific could be an effect of the extrusion of asthenospheric mantle, induced by the closure of the Tethys ocean. In particular, the occurrence of this mechanism has been discussed for the Izu Bonin-Mariana T-A-BA system (Flower et al., 2001), where it is recognized as a major cause of slab roll back and basin opening. This idea is mainly based on petrological evidence and seismic anisotropy data, which would indicate an eastward flow of asthenospheric material from the India-Eurasia collision zone towards the western Pacific area. However, one should also consider other possible interpretations of the above evidence. For instance, the eastward roll back of slabs, predicted e.g. by the slab pull (Fig. 2) and extrusion (Fig. 3) models, would recall asthenospheric material from the surrounding zones, causing an eastward asthenospheric flow like the one inferred for the eastern Asian area. Furthermore, one should also understand if the presumed mantle flow can produce the complex time-space deformation patterns of the West Pacific arc-trench systems, with particular regard to the different evolution of the Izu Bonin and Mariana arcs. In our opinion, explaining such features as effects of the interaction between lithospheric structures, characterized by laterally heterogeneous buoyancy and mechanical strength, seems to be less problematic.

Acknowledgments

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