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Some remarks on the origin of foredeeps

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ABSTRACT

Geological observations show that there are two main types of foredeep basins on the base of their subsidence rate and their relationship with the associated orogen or accretionary wedge. High rates of subsidence up to 1600 m/Myr and a ratio lower than 1 between the area of the elevated belt and the area of the basin characterize the foredeeps associated with W-dipping subduction. Low rates of subsidence up to 300 m/Myr and a ratio higher than 1 between the area of the orogen and the area of the basin characterize the foredeeps associated with E- or NE-dipping subductions. This observation enables us to interpret the slow filling of foredeeps of the first type and the much faster filling of foredeeps of the second type. Moreover, "W-dipping" subductions have a steeper and faster monocline bounding the base of the deep foredeep in contrast to "E- or NE-dipping" subductions which have a shallower monocline. This differentiation supports the notion of the "eastward" mantle flow with respect to the lithosphere detected in the hot-spot reference frame. In this view, the foredeep depth in W-dipping subductions is mainly controlled by the roll-back of the subduction hinge pushed by the relative "eastward" mantle flow while foredeep depth in E- or NE-dipping subductions is instead mainly generated by the load of the thrust sheets and by the roll-back of the subduction hinge due to the advancing upper plate, contrasting the upward push of the mantle. The shape of the foredeeps is regularly arcuate in case of W-dipping subduction, while is linear or following the shape of the inherited continental margin in case of E- or NE-dipping subduction. Fold development in the two opposite foredeeps is significantly different: in W-dipping subduction the folds are transported down in subduction while they are forming, and consequently they are poorly eroded. Syntectonic sediments drape a preserved fold. In the E–NE-dipping subduction, folds and thrust sheets are instead uplifted and deeply eroded. This is also predicted by the envelope of the crests in the two distinct accretionary wedges: in the W-dipping case the envelope may dip toward the subduction, while it is rising toward the hinterland in the other cases.

This approach may explain the differences in terms of geometries, amount and rates of subsidence between the foredeeps around the Adriatic plate. For instance the foredeep of the Southern Alps (back-thrust-belt foredeep) is deformed in the western part by the later Apenninic foredeep development, associated with the sinistral oblique ramp of the W-dipping Adriatic plate subduction. The relationship in the frontal parts of a thrust belt between accretion and eustacy has to be analyzed in terms of when, where and how fast thrusts propagation took place with respect to the randomly distributed sea-level third-order fluctuations, and by the type of subduction which is connected to the foredeep and thrust belt or accretionary wedge. Therefore the stratigraphic patterns in foredeep basins are peculiar and casual successions that record the combination of regional tectonic evolution with independent wider-scale sea-level changes.

Introduction

Foredeeps are here considered those filled or unfilled basins located near convergent tectonic settings, both in oceanic environment (trenches) and in continental areas (foreland basins). Foredeeps analysis has made considerable progresses in the last years due to the large amount of seismic data (e.g., Leggett, 1982; Bally, 1983; Von Huene, 1986) and ODP investigations. Recent descriptions of this topic may be found for example in Allen and Homewood (1986), MacDonald (1991), Jordan and Flemings (1991) and references therein. In the last years an interesting debate emerged from a series of papers dealing with the origin of foreland basins: Beaumont (1981), Jordan (1981) and Quinlan and Beaumont (1984) demonstrated that the load of the thrust sheets may be responsible for the subsidence measured in the Rocky Mountains and the Appalachian foreland basins. On the other hand, Royden and Karner (1984) have shown in the Apennines and in the Carpathians (Royden and Horvath, 1988) that the topographic load is insufficient to explain the 8 km deep foreland basins of those two belts. This stimulating discussion has given rise to two main end members of thrust belts and related foredeeps (Uyeda, 1981; Laubscher, 1988; Royden and Burchfiel, 1989; Doglioni, 1992a). In the last decades it has been undervalued that Le Pichon (1968) noted that in the hot-spot reference frame there exist a "westward" component of the lithospheric motions. A few pioneering papers (Bostrom, 1971; Nelson and Temple, 1972) tried to unravel this observation and they noted the contrasting nature of W- and E-dipping subductions. This evidence was also in agreement with the fact that W-dipping subductions in the Pacific (Mariana type) are in general much steeper and deeper than E-dipping subductions (Dickinson, 1978; Uyeda and Kanamori, 1979; Mitchell and Garson, 1981). O'Connell et al. (1991), Ricard et al. (1991) and Cadek and Ricard (1992) confirmed the presence of the "westward" drift of the lithosphere with respect to the mantle and proposed the lateral viscosity variations at the lithosphere base in a toroidal field of degree 1 as a mechanism for this delay. Independent geologic data are in favor of a global drift of the lithosphere relative to the mantle, but following an undulate pattern (Doglioni, 1990, 1991a,b) which can explain the intriguing Mediterranean puzzle (Doglioni et al., 1991). This paper expands the considerations proposed by Doglioni (1992a) about the differences between thrust belts and brings some geological observations about the nature of foredeeps and the frontal parts of thrust belts, looking from a global perspective. Examples are taken from the Alps and the Apennines. However, further physical investigations are necessary in order to test the validity of this model: the study of the rheological properties of the crust and lithosphere (Ranalli, 1987; Sabadini and Spada, 1988; Ranalli and Murphy, 1989; Marotta and Mongelli, 1991) is helping us in better understanding the evolution of foredeeps.

Foredeeps and plate tectonics

An important parameter of foredeeps is the dip of the underlying basement which is generally ranging between 0 and 10°. The dip is determined by the bending of the lithosphere. The angle is a function of the capability to bend of the lithosphere, which is a function of its elastic or visco-elastic parameters, thickness and composition, and the velocity of the process which is trying to

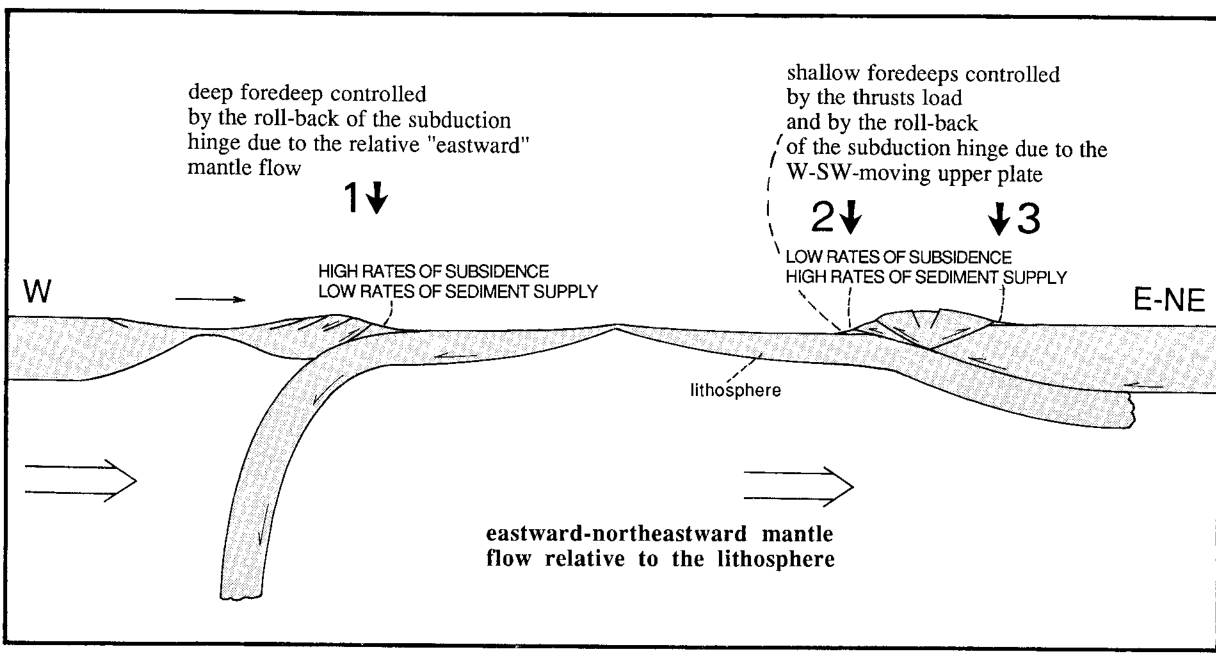


Fig. 1. Foredeeps (both trenches and foreland basins) can be differentiated on the basis of the related subduction, namely W-dipping or E–NE-dipping. Three main geodynamic settings provide foredeeps (1) at the front of a W-dipping subduction, (2) at the front of an E–NE-dipping subduction and (3) at the front of its conjugate back-thrust belt.

bend it: crustal loading or roll-back of the lithosphere. All those factors control the depth of the foredeep, the angle of its basal monocline and the velocity of forward propagation of the system. Plates converging at high rates generate fast subduction which is able to induce high rates of subsidence into the foredeep, due to the fast descent of the top of the lithosphere. The initial depth of a foredeep is also a function of the thickness and composition of the crust: thin oceanic crust will provide a 3–4-km-deep basin, even before the onset of the convergence, while a continental 20–30-km-thick crust may start to subside at a very shallow or non-submarine depths. Foredeeps are usually cold areas, with low heat flow values (see for instance the Apenninic example, Mongelli et al., 1989, 1991).

As already stated, foredeep depth is controlled by the radius of curvature of the underlying lithosphere. This value is particularly function of the

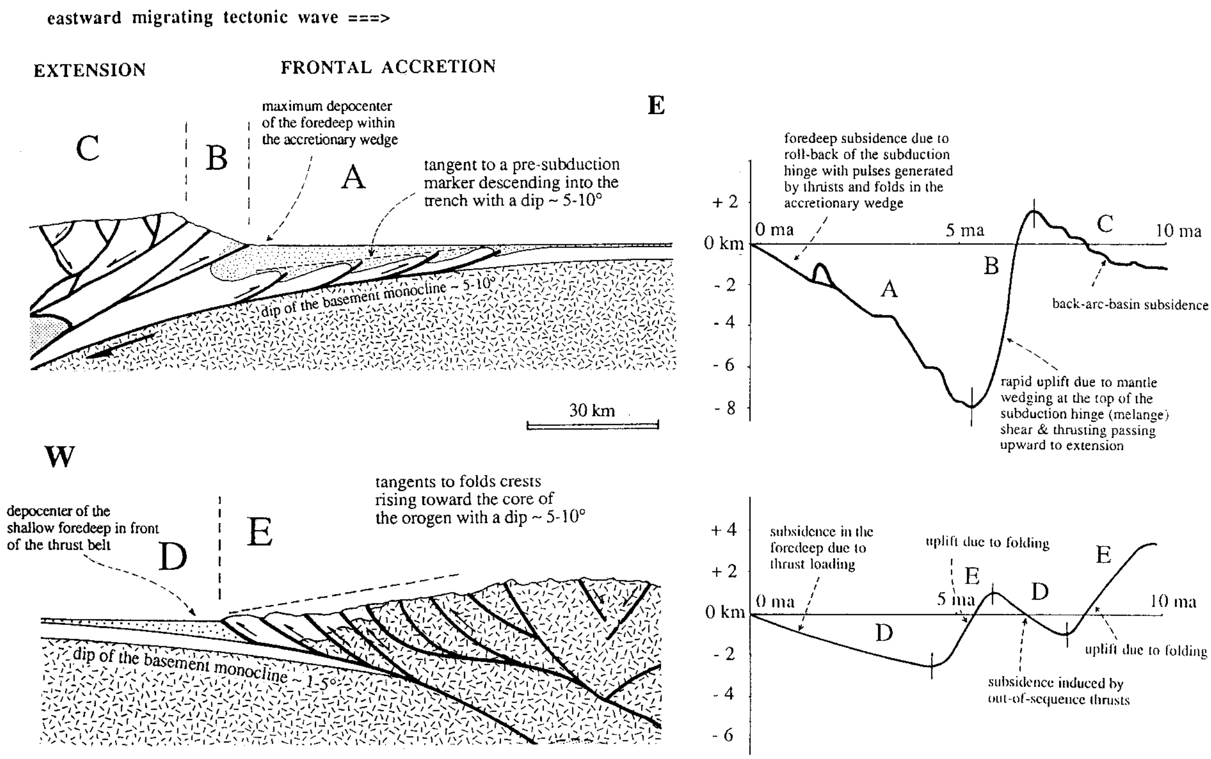
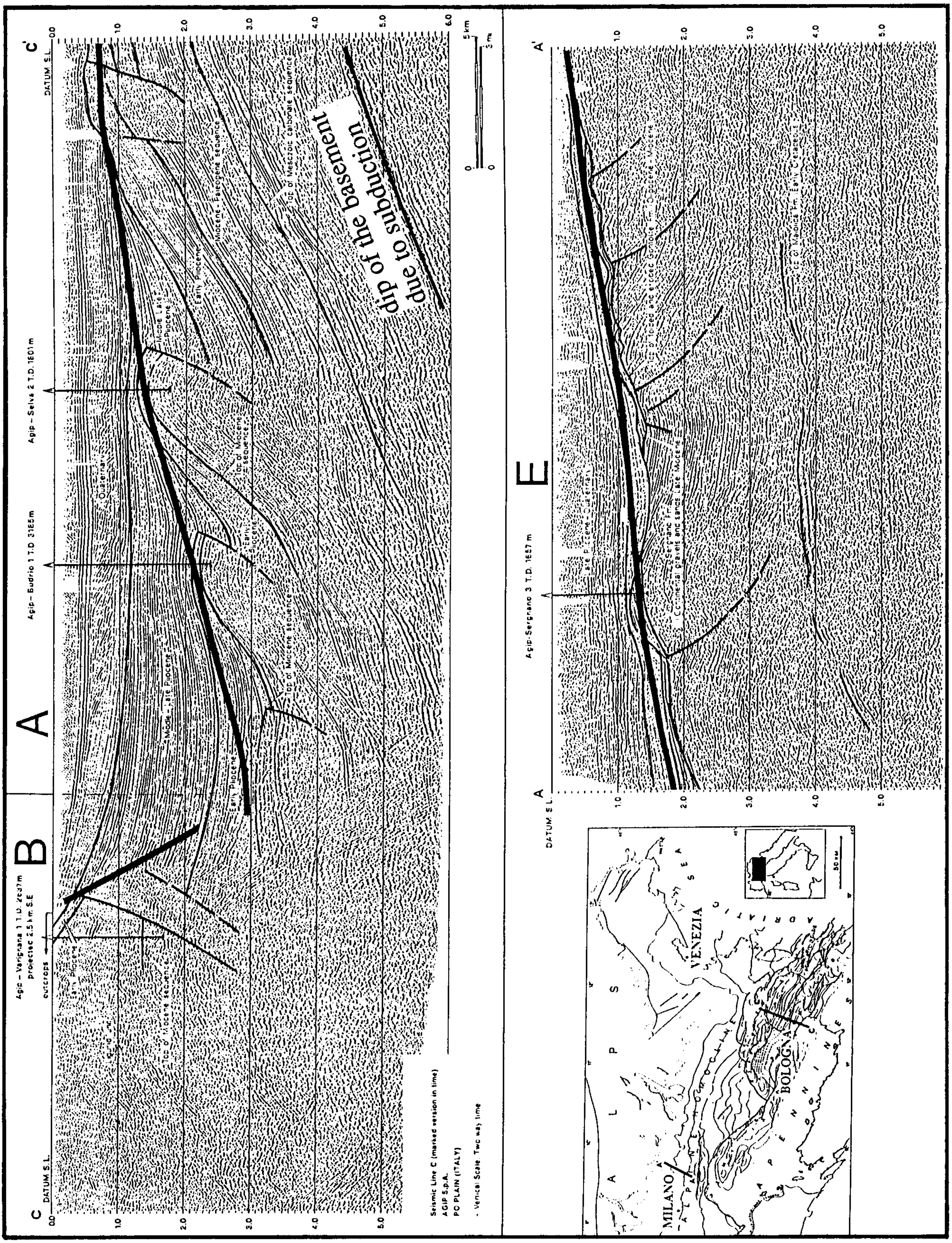


Fig. 2. Main features and structural differences between thrust belts associated to subductions opposing (W-dipping, upper left) or following (E- or NE-dipping, lower left) the mantle flow. Thrust belts related to W-dipping subductions show low structural and morphological elevation, shallow rocks involved; the tangent to the anticlines of a pre-deformation marker is descending into the trench and the depocenter of the deep foredeep basin is within the accretionary wedge (i.e. Apennines). Thrust belts related to E–NE-dipping subductions flow are characterized by high structural and morphological elevation, deep rocks involved (i.e. American Cordillera) and the tangent to a pre-deformation marker is rising toward the hinterland. The shallow foredeep is mainly located in front of the belt. The two curves in the right represent a possible elevation history of one reference point during the structural evolution of the two different end members. The upper elevation curve refers to a reference point at the surface crossed by the migration of the three main tectonic fields (A, B, C) associated to a thrust belt produced by a subduction contrasting the mantle flow. The average of subsidence in the foredeep is up to 1600 m/Myr. The lower elevation curve is relative to a reference point crossed by the migration of the two main tectonic fields (D, E) in a thrust belt due to subduction following the mantle flow. The average of subsidence in the foredeep is about 300 m/Myr. The curves are very different in shape and meaning, confirming the strong physical differences between accretionary wedges related to subductions following or contrasting the mantle flow. Note that subsidence or uplift moments in the curves are generated by different tectonic fields: for instance the initial subsidence in the foredeep related to W-dipping subduction is controlled by roll-back of the subduction hinge and this occurs during frontal accretion; during E-dipping subductions (lower section) the subsidence is generated by thrust loading and the uplift is controlled by the accretion phase (modified after Doglioni, 1992a).


C
C'
DATUM S.L.
B
A
E
A
A'
dip of the basement
due to subduction
Agip - Selva 2 T.D. 1601 m
Agip - Budrio 1 T.D. 3185 m
Agip-Sergnano 3 T.D. 1657 m
Quaternary
Early Pliocene
Middle - Late Pliocene
Top of Miocene sequence
Miocene - Paleogene sequence
Top of Mesozoic carbonate sequence
Seismic Line C (marked version in time)
AGIP S.p.A.
PO PLAIN (ITALY)
- Vertical Scale: Two way time
MILANO
VENEZIA
BOLOGNA
ADRIATIC SEA

direction of the subduction with respect to the "westward drift" of the lithosphere relative to the mantle. In an earlier paper (Doglioni, 1992a) it was proposed that foredeeps have to be differentiated on the basis of the associated subduction. In fact foredeeps related to W-dipping subductions (opposing the mantle flow, type 1 of Fig. 1) are usually 5–11 km deep, and have high subsidence rates, up to 1600 m/Myr (i.e. Apennines and Carpathians) and a steep basal monocline (up to 10°). High rates of subsidence have also been described in the Timor trough (Veevers et al., 1978), along the southern right-lateral transpressive arm of the W-dipping Banda arc subduction. Similar rates are observable all around that arc. Foredeeps associated with E–NE-dipping subduction (following the mantle flow, types 2 and 3 of Fig. 1) are shallower, ranging between 1 and 5 km for ages of 10–20 Myr, with lower subsidence rates in the order of 0–500 m/Myr based on Alpine examples, with a lower dip of the basal monocline ranging between 1 and 5° (Homewood et al., 1986; Massari et al., 1986; Pfiffner, 1986; Roure et al., 1992; and references therein). Similar rates of subsidence may be observed in the Appalachians (Colton, 1970; Viele and Thomas, 1989), Dinarides, Zagros and Himalaya foredeeps (type 2 of Fig. 1) and in the Rocky Mountains (Bally et al., 1966) foredeep (type 3 of Fig. 1), dividing the total thickness of the syntectonic sediments by the age of the clastic sediments filling the basin. This gives a maximum conservative value of subsidence; in fact high sedimentation rates may occur with very low or absent subsidence in a pre-existing deep basin. Foreland basins of back-thrust belts both compressive (e.g., Eastern Cordillera) and transpressive (e.g., Southern Alps, Balkans), are included in this type. Subsidence rates decrease when at the subduction hinge (types 1 and 2 of Fig. 1) or at the back-thrust-belt front (type 3 of Fig. 1) arrives a thicker crust or lithosphere: as more as the thickness of the lithosphere underlying the foredeep increases, as much the subsidence is inhibited. In a map view, foredeeps of type 1 (Fig. 1) are always arcuate, while types 2 and 3 are linear or following the shape of the pre-existing continental margin.

Figure 2 reports the elevation history of one point x which is cross-cut by the subduction "wave": the two curves are based mainly on Apenninic and Alpine data and they show that the dip of the Apenninic curve is much steeper, due to the different type of subduction. Moreover the different segments of the curves are controlled by different tectonic settings. Variations in composition, thickness and relative velocities between plates are additional fundamental factors that control geometry, kinematics and facies of foredeep basins.

In foredeeps associated with W-dipping subductions there are high rates of subsidence and low rates of sediment supply from the associated accretionary wedge (type 1 of Fig. 1). Higher rates of clastics may come from adjacent orogenic belts (e.g., the sediments coming from the Alps and the Dinarides into the Apenninic foredeep, Figs. 4 and 5). The depocenter is "eastward" migrating. Note that in the foredeep linked to W-dipping subductions the source area for clastic sediments is mainly coming from the extensional uplifted ridge following behind the accretionary

Fig. 3. Seismic examples from the Po Plain (northern Italy) of accretionary wedges related to subductions opposing the mantle flow (Apennines) and following the mantle flow (Alps). Note in the field A of the Apenninic section (C–C') that the envelope tangential to the fold crests is dipping toward the hinterland of the belt and the depocenter of the foredeep is located behind the thrusts (compare Fig. 2). Note that folds are transported down in subduction and poorly eroded. Syntectonic sediments drape the folds. Deformation is in sequence (the younger thrust is to the right) but Quaternary out-of-sequence thrusting coeval with the most external one occurs at the transition with field B (compare Fig. 2). Note the dip of the top of the basement which is due to subduction. In the Alpine section (A–A') the envelope to the fold crests is instead rising toward the hinterland of the thrust belt (field E) and the anticlines are strongly eroded. The foredeep (southward out of the section to the left) is shallow and located in front to the thrust belt. The Pliocene–Quaternary sedimentary wedge on top records the influence of the Apenninic subduction on the South Alpine thrust belt. Seismics from Pieri (1983), reprinted by permission of the American Association of Petroleum Geologists.

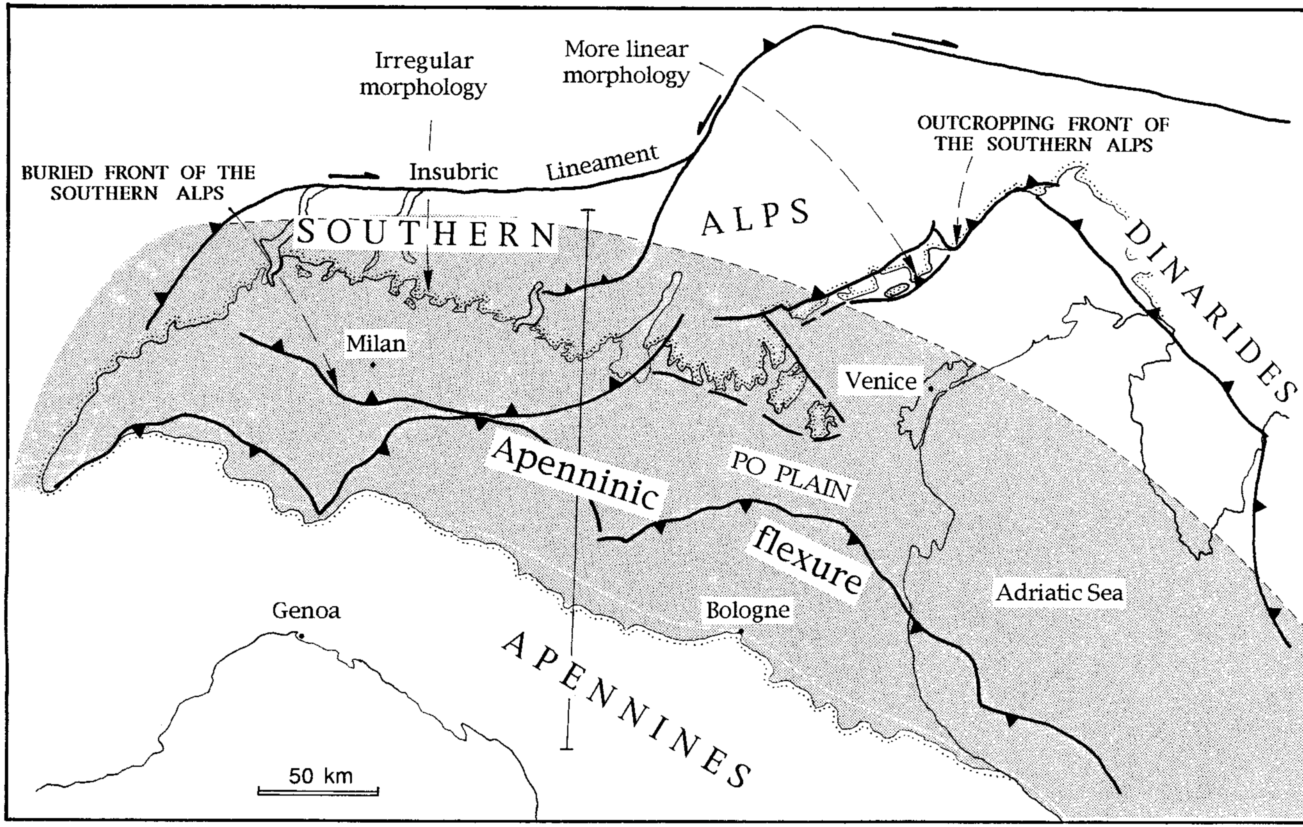


Fig. 4. The Late Miocene–Quaternary north–northeastward propagation of the northern Apenninic lithospheric flexure (shaded area) generated by the roll-back of the subduction hinge, widely involved the western side of the Southern Alps, tilting toward the south the previously formed Alpine features (S-vergent thrusts and N-dipping Late Cretaceous–Late Miocene foredeep). Only the eastern Southern Alps appear not or poorly involved by the Apenninic flexure. This could explain why the frontal thrusts and the related anticlines of the eastern Southern Alps are cropping out whereas the front of the western part is buried in the Po Plain below Plio-Quaternary sediments. In fact the irregular morphology of the western Southern Alps foothills is due to the onlap of the Plio-Quaternary over a pre-existing erosional surface. Compare Fig. 5.

wedge, and very little from the accretionary wedge itself (Fig. 2). Foredeeps associated with W-dipping subductions are part of thrust belts that show low structural and morphologic relief, and involving shallow (usually both young and superficial) rocks. The Neogene–Quaternary foredeeps (both forelands and trenches) of the Earth include the forearc basins of the Aleutian, Kurile, Japan, Nankai, Ryukyu, Philippine, Izu-Bonin, Mariana, Fiji, Tonga and Kermadec trenches, the Banda arc foredeep, the Sandwich and Caribbean forearcs, and the Apenninic and Carpathian foredeeps.

Paradoxically, foredeeps associated with E–NE-dipping subductions are in front of thrust belts that show high structural and morphologic relief, and involve deep rocks. Those foredeeps have low rates of subsidence and high rates of sediment supply. That's why in those settings foredeeps are rapidly filled, with flysch deposits passing upward to the shallow molasse. The foredeeps are usually bypassed and clastic sediments supplied from the orogen are transported far away in remote deltas (e.g., the Rhone delta for the Alps, or the Ganges and Bengal deltas for the Himalayas).

There are two main foredeeps related to subductions following (E- or NE-dipping) the mantle flow, at the front of the thrust belt (type 2 of Fig. 1) along the active subduction (trench in the oceanic environment) and in front of the backthrust belt (type 3 of Fig. 1), the internal conjugate part of the orogen (e.g., the Southern Alps, the Rocky Mountains, etc.).

The elevated ridge associated with W-dipping subduction is always lower with respect to the orogens due to E–NE-dipping subductions. Uplift rates in the two types of mountain ranges or accretionary wedges are in fact much greater for orogens due to E–NE-dipping subduction. The clastic supply is also bigger from these last orogens. Another interesting observation is that the ratio between the area of the ridge and the area of the associated foredeep is lower than 1 for W-dipping subduction, while it is greater than 1 for E–NE-dipping subduction. The consequence of that is that foredeeps due to E–NE-dipping subduction are quickly filled, while the W-dipping type may remain poorly filled for long times.

Along E- or N–NE-dipping subductions the foredeeps are deeper along transpressive belts (e.g., the dextrally transpressive central–eastern Alps, or the Chaman transform zone at the western Indian sinistral transpressive margin).

There are also foredeeps in front of thrust belts due to N–S convergence (e.g., the Pyrenees or the Southern Caribbean–Venezuela thrust belt. Those orogens appear to be controlled by rotations of plates (Iberia and South America, Doglioni, 1993) and show characters similar to Alpine or Cordillera types.

Moho depth, gravimetric and heat flow profiles across thrust belts and related foredeeps show different characters descending on whether they are associated with W-dipping or E–NE-dipping subductions.

Therefore the dip of the basal monocline of the foredeeps is controlled by a variety of factors. The final dip is the resulting combination of the former physical parameters and in particular of the contrasting forces acting on the plates: "eastward" mantle flow and lithostatic load.

The foredeeps around the Adriatic plate

The Adriatic plate is at present almost entirely surrounded by thrust belts and related foredeeps. All the main types of foredeeps are represented: the western margin is represented by the Apenninic foredeep (Pieri and Groppi, 1981; Pescatore and Senatore, 1986; Ricci Lucchi, 1986; Bally et al., 1986; Cassano et al., 1986; Mostardini and Merlini, 1986; Ori et al., 1986, 1991; Royden et al., 1987; Casero et al., 1988; Gelati and Gnaccolini, 1988; Moretti and Royden, 1988; Pescatore, 1988; Ricchetti et al., 1988; Rossi and Rogledi, 1988; Sella et al., 1988; Patacca and Scandone, 1989) due to a W–SW-dipping subduction (type 1 of Fig. 1). At the eastern margin there is the Dinaric foredeep (Dragasevich, 1983; Frank et al., 1983; Finetti, 1984; Aljinovic and Blaskovic, 1987; Finetti et al., 1987) connected to the frontal thrust belt of a NE-dipping subduction (type 2 of Fig. 1). At the northern margin there is the foredeep of the Southern Alps (Pieri and Groppi, 1981; Cassano et al., 1986; Massari et al., 1986; Bernoulli et al., 1987; Bersezio and Fornaciari, 1988; Gelati et al., 1988; and references therein) which are the back-thrust belt of the Alps (type 3 of Fig. 1). The three foredeeps have different ages, but they also have different styles and they prograded toward the internal parts of the Adriatic plate.

The Apenninic foredeep is mainly located behind the active frontal accretionary wedge which is buried in the Po Plain (Fig. 3) and in the central–eastern Adriatic Sea. It mainly formed from the Late Miocene on, forward prograding with high subsidence rates (up to 1600 m/Myr) as indicated by the 7–8.5 km thick Pliocene–Quaternary sediments. The foredeep is supplied by the neighboring chains (Apennines, Alps and Dinarides), and the basement monocline tilted the older adjacent foredeeps (western Southern Alps, Figs. 4 and 5, and Dinarides) toward the subduction zone. Only the eastern part of the South Alpine foredeep (Massari et al., 1986) seems to be less affected by the Apenninic flexure (Fig. 4); in fact in the eastern Southern Alps the frontal thrusts are cropping out, whereas they are buried beneath the Po gravels in the western part.

The Southern Alps foredeep formed between Late Cretaceous (western part) and the Pliocene. It prograded southwards with a general eastward shift of the depocenters (Massari, 1990). The Southern Alps foredeep is located in front of the thrust belt and was partly involved in thrusting. Subsidence rates have been low because they rarely exceed the 300 m/Myr, dividing the entire thickness of the flysch and molasse deposits by their duration of deposition (2–6 km of sediments deposited in 15–40 or more Myr). In the eastern part of the South Alpine foredeep, the Dinaric foredeep also interferes with its effects of subsidence since at least the Paleocene up to the Early Miocene. Taking the entire thickness of the Dinaric foredeep deposits we also obtain low rates of subsidence similar or lower to those of the Southern Alps, characterizing a shallow foredeep prograding west–southwest since Late Cretaceous times. The isopachs of the Apenninic foredeep give the dip of the Adriatic W–SW-dipping basement monocline; they clearly interfere with the Dinaric features, tilting them toward west–southwest. The differences between the Apen-

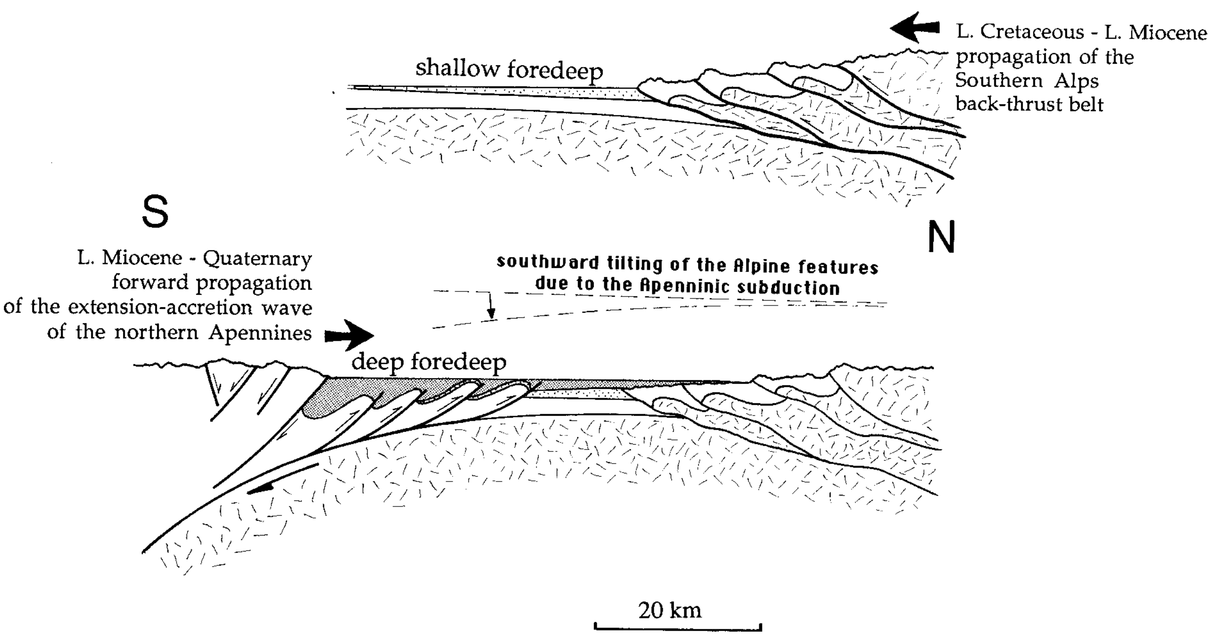


Fig. 5. Idealized cross-section between the northern Apennines and the western Southern Alps. The last ones are the S-vergent back-thrust belt of the right-lateral Transpressive arm of the Alps and formed between the Late Cretaceous and the Late Miocene. Their shallow foredeep and the thrust planes have been southward tilted by the Late Miocene–Quaternary northward propagation of the basement monocline generated by the Apenninic subduction.

ninic and South Alpine foredeeps shown on Agip cross-sections (Cassano et al., 1986; Rossi and Rogledi, 1988) may be taken as a good example of the different tectonic evolution of foredeeps related to opposite subductions.

The envelope to the crests of the frontal anticlines of the Apennines is descending into the foredeep. The folds are poorly eroded (Figs. 2 and 3) and syn-folding sediments drape the anticlines with reduced thicknesses in the hinge zones. In contrast, the tangent to the crest of the anticlines in the Southern Alps or the Dinarides is rising toward the interior of the orogen (Figs. 2 and 3) and the anticlines are deeply eroded and younger sediments onlap discordantly deep erosional surfaces.

Internal active thrusting, contemporaneous with the frontal thrusts occurs at the rear of the Apenninic accretionary wedge (between field A and B, Figs. 2 and 3). This out-of-sequence thrusting (Morley, 1988) is visible as a relic in the Gran Sasso area (Ghisetti and Vezzani, 1991) in the central Apennines and is a common feature of accretionary wedges due to W-dipping subductions: see for instance the example of Barbados (Westbrook, 1982; Casey-Moore et al., 1991, Fig. 5).

Rotations in frontal parts of accretionary wedges are different along transpressive belts whether they are associated with W-dipping subduction or E–NE-dipping subduction. For instance the large rotations (up to 140° clockwise) described along the dextral transpressive Apenninic Sicilian part (Channell et al., 1990) are not present in the dextral transpressive South Alpine belt (Channell et al., 1992). This difference may be controlled by the more superficial decollements in the accretionary wedges of W-dipping subductions (Fig. 2, e.g., the Apennines), while the decollements and thrust ramps are deeply rooted into the basement in the Alpine case, inhibiting strong rotations of the thrust sheets.

Along the strike of the foredeeps around the Adriatic plate, in any geodynamic setting, we may observe a decrease of the subsidence rates in correspondance of thicker crust (or lithosphere) underlying or in front of the foredeep: e.g., the Lessini Mountains at the toe of the Southern Alps (type 3 of Fig. 1), or the Istria Peninsula at the northern margin of the Dinarides (type 2 of Fig. 1), or the Puglia at the front of the Southern Apennines (type 1 of Fig. 1). Therefore, thickness and compositional anisotropies of the upper crust and the entire lithosphere control structural undulations along strike both of the foredeep and of the thrust belt itself. For geophysical data about the Adriatic plate the reader is referred to Panza (1984), Suhadolc and Panza (1988), Nicolich (1989), Spakman (1989), Mongelli et al. (1991) and references therein.

The Alps, in spite of the more elevated structure and topography with respect to the Apennines, have a much smaller foredeep. In front of the western Alps the Neogene foredeep is almost insignificant, while it is increasing at the front of the central and eastern Alps, where the base of the Oligocene molasse reaches 4000 m (e.g., Bigi et al., 1989). Dividing the total thickness of the sediments by their age, the average subsidence rate is of about 100 m/Myr at its maximum, at least one order of magnitude lower with respect to the Apenninic foredeep. The Alpine molasse basin from Switzerland to Germany increases moving from the western to the eastern Alps: in other words the foredeep is deeper and with higher subsidence rate along the transpressive part of the orogen. The central–eastern part of the Alps have in fact been interpreted as due to dextral transpression (e.g., Laubscher, 1983; Schmid et al., 1989).

Foredeeps: main relationships between tectonics, eustacy and stratigraphy

A few problems on the origin of foredeep basins and their relationship with sedimentation are here briefly discussed. We can subdivide this topic into two main branches: (1) the effects of tectonics in controlling depositional sequences and (2) the control operated by depositional sequences in positioning later deformation.

Effects of tectonics in controlling depositional sequences in foredeeps

Stratigraphy results from the combination of tectonics, sediment supply climate and eustacy.

The control of tectonics on stratigraphy occurs on different scales, i.e., geodynamic and regional (local). The geodynamic global scale may control first- and second-order sea-level fluctuations. Major long-term transgressions over the continents relate to fast oceanic spreading and production of new hot crust (Pitman, 1978; Heller and Angevine, 1985). New oceanic crust development is in turn generated by plate velocities which are also controlled by viscosity values in the Low Velocity Layer at the base of the lithosphere. On the other hand, plate motions over the Earth's surface produce instability of the Earth's axis and polar wandering has been demonstrated to be a possible mechanism for generation of third-order asymmetric sea-level fluctuations (coeval high stand in the Northern Hemisphere and low stand in the Southern Hemisphere, Sabadini et al., 1990). Stratigraphic differences have been noted in the Southern Hemisphere Neogene with respect to the Northern Hemisphere (Carter et al., 1991). Other possible mechanisms for the third-order non-global cycles may be regional stress changes (Cloetingh et al., 1985) or density changes of plates induced by stress (Cathles and Hallam, 1991), or chaotic mantle convection (Officer and Drake, 1985). At a smaller scale, fourth- or fifth-order sea-level fluctuations seem to be very chaotic and scattered, being controlled by the interaction of eustacy and local (never uniform) tectonics (Pirazzoli, 1988). Fjeldskaar (1989) demonstrated that deglaciation may induce a sea-level rise far away from polar regions while in Scandinavia for instance the post-glacial rebound generates a coeval widespread sea-level fall at its continental margins. In summary the different orders of sea-level fluctuations are extremely complicated and variegated in origin and as much as the wavelength is shorter, it is difficult to believe in one single global sea-level curve.

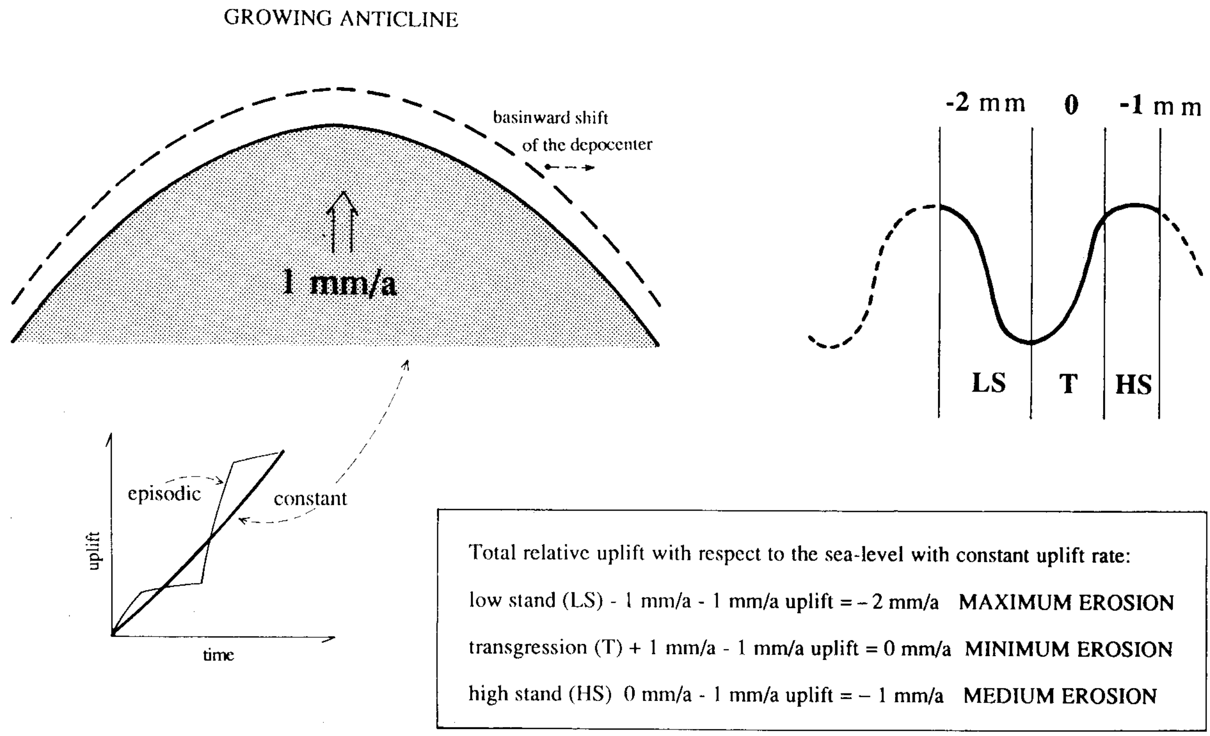


Fig. 6. The uplift of an anticline may be episodic or with a constant rate as suggested by plates motion. In this picture is considered the case of a fold growing for example at a constant rate of 1 mm/yr. The uplift rate is added to a hypothetic third order cycle on the rigth side. So during a lowstand with a fall of the sea level of 1 mm/yr, the resulting total sea-level drop will be of 2 mm/yr with the maximum rate of erosion in the anticline. During transgression at a rate of 1 mm/yr the tectonic uplift is compensated and the total relative sea-level variation should be 0 mm/yr with the minimum rate of erosion of the fold. These values are only used to illustrate the problem.

Thickness and composition of the crust determine the water depth. Old oceanic crust may be a good deep water collector of turbidites. The geodynamic environment and the related crustal nature are fundamental factors in controlling typology of sedimentation.

The different curves representing the structural elevation history of one reference point x in accretionary wedges generated by subductions following or opposing the mantle flow are presented in Figure 2. They may be ideally compared or added to an eustatic curve representing the combination of the five or six main orders of magnitude of sea-level fluctuations. The resulting new curves could be a possible interference pattern between eustacy and tectonics during a 10 Myr evolution of the two end members foredeep basins. The curves should be very different in shape whether they are the combination of sea-level rise, or sea-level drop, or the transitional point particularly of the second- and third-order cycles during the Earth history. Moreover, tectonic subsidence rates are generally of one order of magnitude greater with respect to eustacy, making the structural input more important in controlling the sedimentary stacking. Nevertheless, eustacy still has a fundamental control in orchestrating the sequence stratigraphy.

In general we do observe a foreland migration of thrusts and folds in an accretionary wedge (Fig. 7). Syntectonic depositional sequences are consequently forward migrating in time as well. Again this assumption has to be splitted between foredeeps related to W-dipping or E–NE-dipping

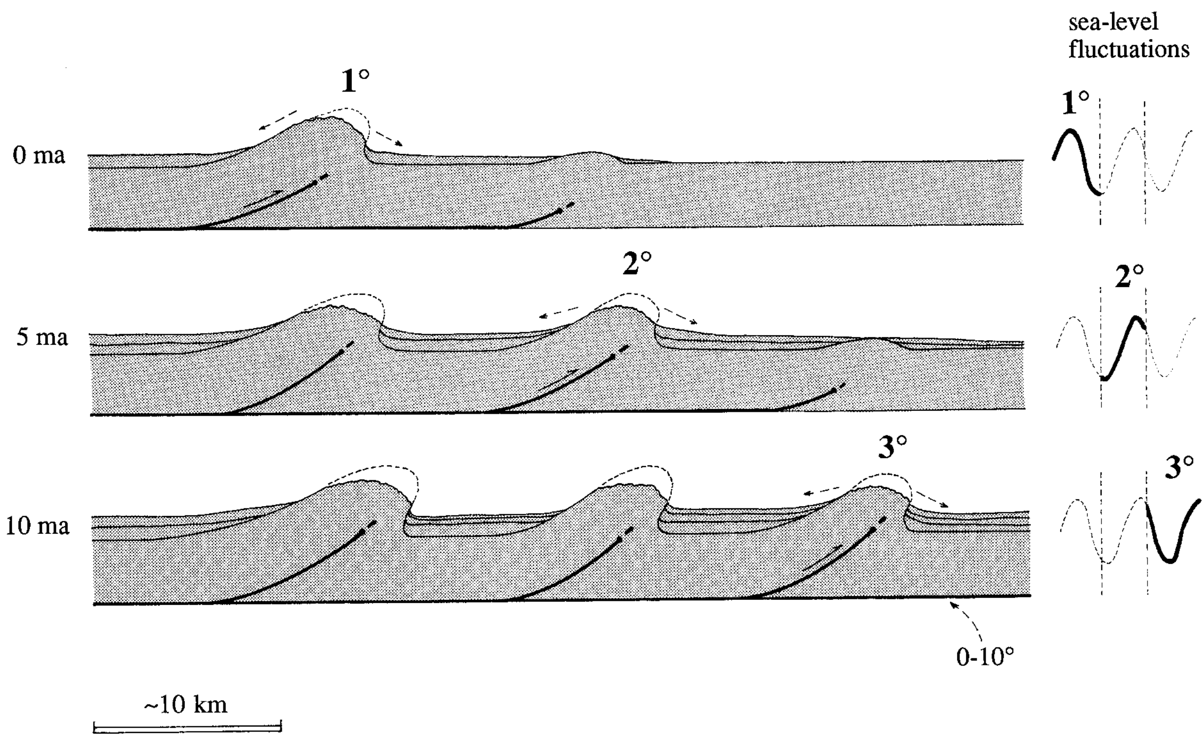


Fig. 7. Three "tectonic" cycles may occur during three different segments of the eustatic curve (a few random third-order cycles). The related depositional sequences may then be characterized by different interference patterns between tectonics and eustacy. The fault-propagation folds produce episodic and localized uplifts of a continuously foreland propagating thrust belt and consequent coeval foreland migration in age of the depositional sequences associated to the growth of each anticline. The evolution of the thrust belt is recorded by punctual information of unconformities. The basement monocline controlling the basal decollement may usually range between 0 and 5° for thrust belts related to subductions following the mantle flow (E- or NE-dipping), and between 0 and 10° for thrust belts related to subductions contrasting the mantle flow. In this last case with the steepest angles, the internal synclines may be much more filled with clastics. Along synclines both longitudinal and transversal clastic transport occurs.

subductions. In fact fold development and stratigraphy strongly differ in the two opposite members. In the W-dipping related foredeep anticlines are better preserved and draped by clastic deposition, while they are dismembered by erosion in the other case. In the W-dipping subduction the area of the foredeep is bigger with respect to the area of the mountain ridge, while it is smaller in the E–NE-dipping subduction (Fig. 8). The consequence is that W-dipping related foredeeps tend to maintain deep water environments, and E–NE-dipping related foredeeps are rapidly filled by sediments and bypassed.

Fold development is a result of plate motion which is considered to have quite constant velocity as indicated by ocean floor spreading. Folds are local episodic uplifts within a continuous foreland propagating thrust belt. One single fold of 1–3 km amplitude may develop in a time spanning 0.5–5 Myr or more. During fold growth (subaerial or submarine) syntectonic sedimentation onlaps the uplifting zone. Anticlines are in general eroded and represent the source area for the foredeep. One single anticline evolves through time, supplying sediments for deposition. The growth of the anticline may happen during any tract of a third-order cycle (low stand, transgressive, high stand) and consequently controls the sedimentation patterns as a function of the interference between uplift rates and sea-level fluctuations which may be very random (Figs. 6 and 7). Uplift rates may be considered constant for one steady-state anticline growth, but they may also be irregular, especially in transpressive realms.

If uplift rates are irregular and suddenly variable through time, the interference pattern is more complicated and difficult to unravel in terms of relative importance between tectonics and eustacy. What is important is the ratio between uplift rates and eustacy rather then the single values. If we consider one single anticline growing during an eustatic cycle (Fig. 6), uplifting for instance at 1 mm/yr, during a sea-level fall of 1 mm/yr, the sea level will relatively fall at 2 mm/yr, 0 mm/yr during the transgression and will fall again by 1 mm/yr during the high stand. This interference between tectonic uplift and sea-level fluctuations generates a greater erosion during the low stand, but during the transgression the relative still stand (0 mm of fluctuation) will produce no erosion and may be missed, as no sedimentation or condensed beds will form. Con-

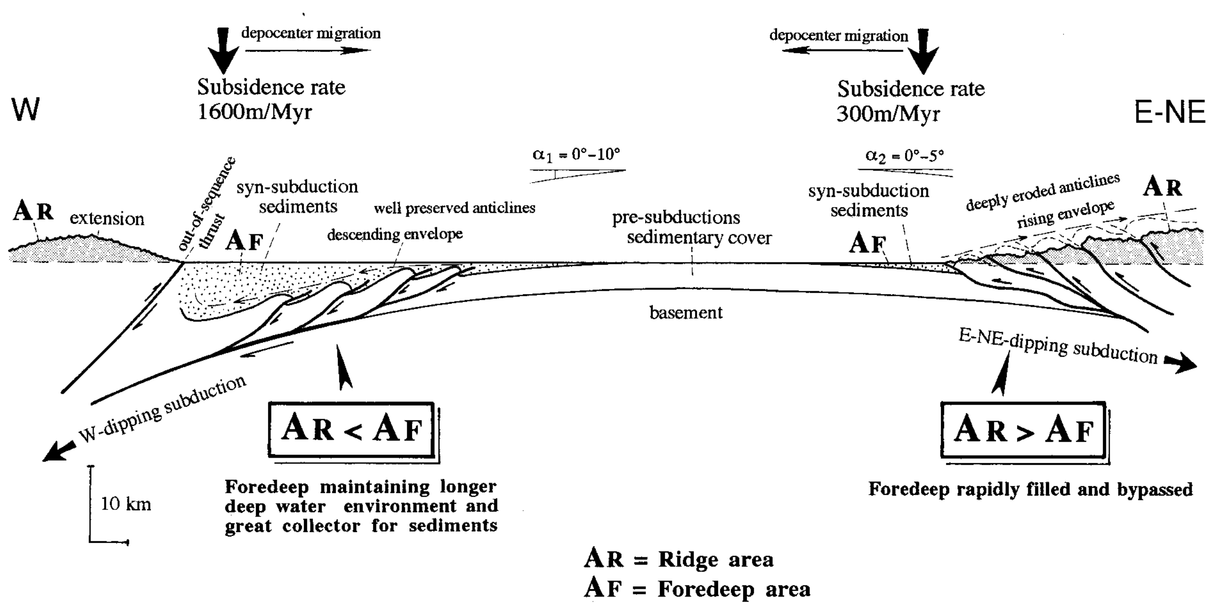


Fig. 8. A strong asymmetry of the geological parameters is observed between foredeeps (filled or unfilled) related to W-dipping subductions (left), and foredeeps related to E–NE-dipping subductions (right). In the W-dipping case the foredeep has an area wider than the mountainous ridge, while it is smaller in the opposite subduction.

sequently, because the anticline is still uplifting, the high-stand system tract may onlap discordantly the low-stand system tract, the angle amplified by the low or non-deposition of sediments during the transgression.

In the Pyrenees, there are cases in which deformation is apparently confined to transgressive systems tracts (Mutti et al., 1988; Mutti, 1991). This could be explained by the interference of the uplift rates and the sea-level fluctuations. During the low stand, rates of uplift and sea-level fall are added, incrementing the relative sea-level lowering. The erosion of the anticline is maximum during this time interval (Fig. 6). During transgression, the effects of uplift may be partly or totally compensated by sea-level rise and the relative sea level may remain constant in the limbs of one growing anticline, the erosion of the fold being a minimum. The following high stand may follow discordantly on the limbs after a time of little or no deposition along the fold limbs, representing the intermediate case in which the uplift rates of the fold are only in part or not compensated by sea-level rise, so erosion is in the intermediate case with respect to the low stand and the fast transgression (Fig. 6). So every fold has an individual history and a peculiar relation with sedimentation as a function of when, where and how it formed with respect to the two main variables: rheology of the deformed rocks and the independent sea-level fluctuations. Every fold development is then a local fact which represents only one spatial and temporal element of a greater thrust belt which has to be analyzed as a whole. Dating one fold, filtered by the effects of global sea-level fluctuations, has to be considered one recorded step of the evolution of a thrust belt (Fig. 7).

Depositional sequences may then differ in time due to the variable interference pattern between tectonics and eustacy during thrust belt evolution. The foredeep is in general located in front of the thrust belts related to E- or NE-dipping subductions, while it is behind or lateral to thrust belts related to W-dipping subductions (Fig. 8). In this last case the depocenter is often behind the accretionary wedge (forearc basin, piggy back basin) and the basement monocline is very steep (up to 10°). Once formed, a depositional sequence related to the growth of an anticline is passively transported in depth by subduction and buried by the younger clastics filling the foredeep.

In foreland basins, related to thrust belts generated by E- or NE-dipping subductions, the basement monocline is in general less steeply inclined (few degrees, 0–5°) and the foredeep is located in front of the thrust belt. Once formed, a depositional sequence related to the growth of one anticline may be thrust by the advancing thrust of a fault-bend fold, or folded or overturned by the foreland migration of a triangle zone or a fault-propagation fold. Once a fold is disactivated it may be buried and passively transported to depth by W-dipping subductions, or it may be uplifted by deeper thrusts in E-dipping subductions. In this last case, a fold may subside if out-of-sequence internal thrusts generate a new load in the foreland. An anticline stops growing when the lithostatic load is greater with respect to the horizontal shear: this may be partly influenced by the water column in submarine folds, or by the erosion in subaerial environments. Fold development is practically not controlled by sea-level fluctuations but rather by crustal thicknesses and horizontal shear stresses. In other words, a fold may grow or stop growing during any interval of a sea-level curve.

The extent of the unconformities within the molasse is a function of the thrust-belt structure and reflects areas of stronger uplift. A problem is represented by the style and timing of the orogenic evolution: are chains regularly rising in a continuum creep (Hsü, 1992) and do the unconformities record moments of sea-level fall or low stand (Vail et al., 1977), or are chains generated step by step so that the unconformities simply mark moments of tectonic crisis? In general, plates move with a regular velocity suggesting that in areas of deformation the tectonic activity should be almost constant. If the tectonic evolution generated a constant regularity of tectonic crisis with a wavelength too short or too long with respect to the eustatic sea-level changes, then an interesting problem appears in dating the thrust belt: the timing of the tectonic crisis has been tuned within the time missing at regional unconformities and correlated to the age of coarse-grained sediment supply onlapping the discontinuity (i.e. the Messinian Conglomerate of the Southern Alps). But if the unconformities recorded only moments of general low stand (in this case global or confined to the salinity crisis in the Mediterranean area) then we could argue that the chain developed more gradually and that the sea-level oscillations fixed only moments of the syntectonic sedimentary sequences. The different interpretations of the unconformities and conglomeratic supply in molasse sequences allow different tectonic reconstructions. With the sea-level change interpretation (Vail et al., 1977), thrust belts may be seen as rising constantly. If unconformities and conglomeratic supply are tectonic-related, then episodic tectonic activity existed, which is in contrast to the general regularity of plates motion.

Sequence boundaries in the frontal fold limb of thrust belts are marked by angular unconformities in syntectonic sediments, with decreasing dips toward the foredeep. Unconformities are angular only along dip where the frontal fold is perpendicular to the assumed regional maximum compressive stress and the fold presents a "cylindrical" trend. Where there are structural undulations in the fold axis (i.e. transpressive or transfer zones) the unconformities are marked by angular relationships along both dip and strike (Doglioni, 1992b). In summary, deep inherited structures control the nature of the thrust belt and consequently of the related unconformities. A growth fold, with constant horizontal axis, generated by pure compression produces angular unconformities only along dip, while a growth fold generated by transpression, or close to its periclinal termination, along a transfer zone, produces angular unconformities both along dip and strike.

Effects of depositional sequences in controlling deformation in foredeeps

This important part of the relationship between stratigraphy and tectonics is here only briefly presented. The shape of a thrust belt is controlled by several parameters: velocity of plate convergence, composition and thickness of the plates, and all the physical factors which commonly control tectonics: temperature, pressure, fluids, etc. Moreover inherited structures and lithology variations are fundamental elements in controlling undulations in a thrust belt. For instance lateral or oblique ramps form along pre-existing tensional or transtensional fault zones, or along facies boundaries, platform to basin transitions, or in the presence of lateral lithologic variations (i.e. a localized interbedded evaporitic basin). The front of a thrust belt may be of three main types of structure: triangle zone, fault-bend fold or fault-propagation fold. These three main structures may develop separately or coexist in the development of one thrust belt front and their evolution is controlled by the inherited architecture of the passive continental margin with respect to the orientation of the later compression (i.e. triangles form more likely in basinal or relatively more shaly sequences). Each tectonic environment gives a particular relationship with the coeval clastic sedimentation in the foredeep basin. Facies development and distribution control the hydrographic pattern, in other words the morphology of the thrust belt. As the geometry of the thrust belt is irregular due to pre-existing structures of the passive continental margin, the clastic deposition will follow irregular distribution. A frontal triangle zone produces a foreland dipping clastic sequence which is continuously rotated and tilted as long as the triangle works. The sediments onlapping the frontal limb are themselves progressively tilted and eroded, supplying the foredeep. In case of a frontal fault-bend fold associated with a thrust cropping out at the surface the clastic sediments of the foredeep may simply be thrusted and not supplying the foredeep itself; in this case the feeder of the foredeep is only the hanging wall of the thrust.

It is very common to observe sequence boundaries or maximum flooding surfaces acting as main decollement layers. Particularly shales or evaporitic layers of transgressive systems tracts are frequent detachment horizons. Major facies changes (i.e., a platform to basin transition, both in silicoclastic or carbonatic environment) provide good sites for thrust ramps. Lateral heterogeneities in the sedimentary cover perpendicular to the stress axis control the non-cylindricity of the deformation, generating lateral and/or oblique ramps of fault planes (both normal and reverse). In general the main control of the trajectory of faults is due to third-order cycles, in other words to depositional sequences. Flexural slip and minor internal deformations are controlled by interbedded shales at the scale of the fourth- or fifth-order cycles. Sedimentation may control tectonics even during basin formation: subsidence rates are controlled also by the load of the sediments and this load is continuously migrating through time. Thrusts move when the lithostatic load allows them to move, so thrusts may be reactivated when the erosion in the hanging wall is sufficient to push the Mohr circle across the fracture envelope, generating episodic activity. The syntectonic control due to sediments is particularly apparent when concentrations of sediments (i.e., a carbonate platform) are restricted in one particular area, producing a local gravitational instability which may enhance local greater subsidence rates. An often underestimated factor is the subsidence due to compaction of the sedimentary cover (e.g., between two growing anticlines). As much as the sedimentary cover is inhomogeneous, as much the subsidence related to compaction is irregular: this effect may introduce local depositional sequences confined to restricted areas, independent from any eustatic cycle.

Concluding remarks

Foredeeps are substantially different whether they are associated with a W-dipping or with a E- or NE-dipping subduction zone (Fig. 8), in other words, subductions opposing or following the "eastward" mantle flow in the hot-spot reference frame (Ricard et al., 1991). Subsidence or uplift rates radically change in the two main end members. As a general observation foredeeps related to W-dipping subductions have a ratio lower than 1 between clastic supply and subsidence rates, while foredeeps due to E- or NE-dipping subductions have ratio greater than 1. In fact foredeeps of "Apenninic" or W-dipping type tend to maintain deep water conditions with abundant flysch

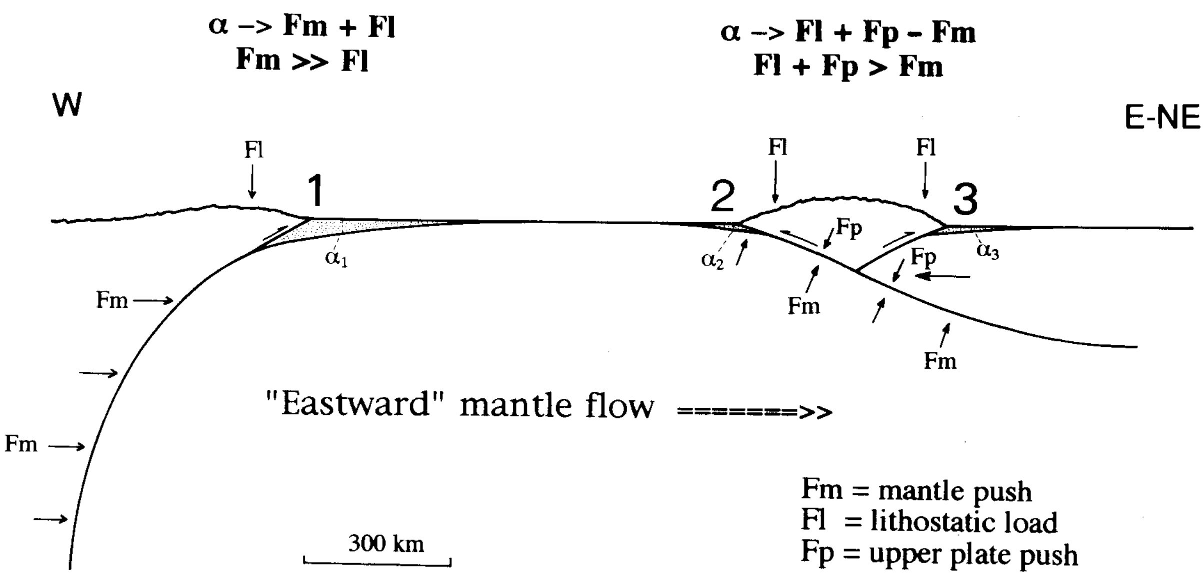


Fig. 9. The foredeep origin may be differentiated on the basis of the associated subduction. Assuming constant physical parameters (e.g., thickness, composition, elasticity and viscosity), the bending (α) of the lithosphere controlling the foredeep is due to the mantle push (*Fm*) and the lithospheric load (*Fl*) in the W-dipping subduction, where $Fm > Fl$. In the E–NE-dipping subduction the bending is due to the lithospheric load (*Fl*) and the downward push generated by the upper plate (*Fp*) moving "westward", contrasted by the upward component of the mantle push (*Fm*) which is sustaining the lithosphere, where $Fl + Fp > Fm$. Numbers *1*, *2* and *3* indicate the main geodynamic environments of foredeep development.

deposits, while "Alpine" or E–NE-dipping subduction-related foredeeps are generally shallowing upward ("molasse") and bypassed by sediments due to the insufficient rate of subsidence in comparison with the high rate of clastic supply (Fig. 8).

The origin of foredeeps in the two different thrust belts could consequently be controlled by different geodynamic factors. In the W-dipping subduction, where the topographic load is insufficient to generate deep foredeeps basins (Royden and Karner, 1984), the subsidence is permitted by the eastward retreat of the subduction hinge, mainly due to the eastward push of the mantle acting on the subducted slab (type 1 of Fig. 9). In the E–NE-dipping case the control is made by the load of the thrust sheets (Quinlan and Beaumont, 1984) opposing the upward push generated by the "eastward" mantle flow (type 2 of Fig. 9). While in the frontal thrust belt the foredeep is also controlled by the roll-back of the subduction hinge generated by the advancing upper plate (type 2 of Fig. 9), in the back-thrust belt this factor is absent (type 3 of Fig. 9). Note that the upward component of the mantle flow is maximum if the subducted slab is striking perpendicularly to it. In case of oblique and lateral subduction it should decrease allowing a bigger effect of the lithostatic load and the vertical component of the advancing upper (eastern or northeastern) plate. In fact along the transpressive belts there is an increase of the foredeep depths and of the subsidence rates. See for instance the eastward increase of the north Alpine molasse, from Switzerland to Austria in front of the central–eastern Alps, generally considered as a right lat-

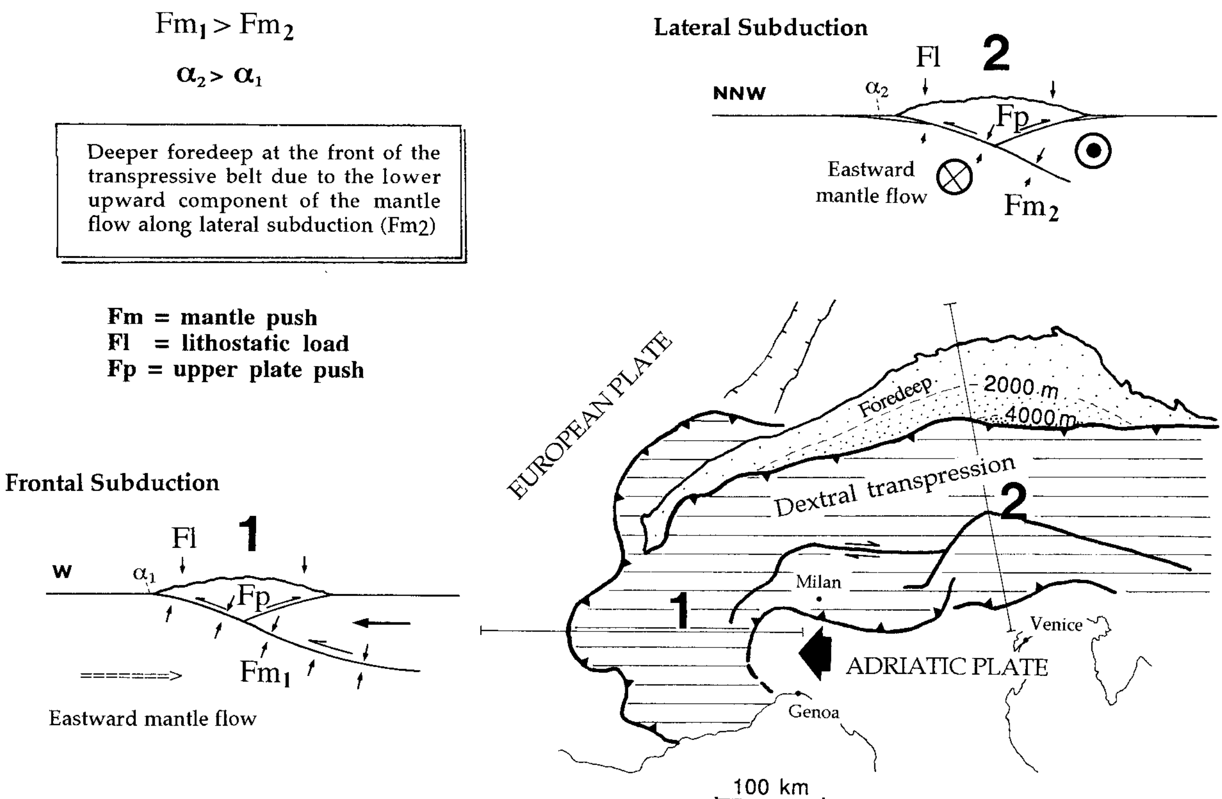


Fig. 10. The Neogene north Alpine foredeep basin is practically absent in front of the western Alps, while it increases eastwards where the Oligocene base reaches about 4000 m. This might be interpreted as controlled by the different angle of the subduction relative to the mantle flow, i.e., perpendicular in the western Alps and parallel in the central–eastern Alps where a dextral lateral subduction occurred. This should have generated a different interplay between vertical upward component of the mantle versus the vertical lithostatic load and the downward component of the eastward moving upper Adriatic plate.

eral transpressive belt (Laubscher, 1983, 1992) with a lateral subduction (Fig. 10). A similar interpretation might be applied to the Chaman sinistral transpressive belt (Jadoon et al., 1992) or along belts parallel to the mantle flow (e.g., the Pyrenees).

The basement monocline at the bottom of the foredeep is a direct indication of the angle of the hinge of the subducting lithosphere (types 1 and 2 of Fig. 9). Thus the evolution of foredeeps is controlled by the subduction type.

The stratigraphic assemblages in foredeeps are differently organized in the different types of geodynamic environment due to the different geometry and kinematics of the basins associated with W-dipping or E–NE-dipping subductions. Moreover, fold development, which may also be distinct in the two subductions, occurs with different styles, magnitudes, progradations and timing in the different tectonic settings. Eustacy, with independent and different wavelengths, overprints those variegated structural evolutions.

The division of pure eustatic control or pure tectonic control of the stratigraphic stacking in foredeep basins may be misleading. The relative interplay has to be strongly revised taking into account the magnitude and origin of the different orders of eustatic oscillations and the more random tectonic control in the geodynamic environments where they operate.

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