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Geodynamics of the Alpine-Mediterranean system - a synthesis

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ABSTRACT

Alpine-Mediterranean orogenic structure is largely due to suture progradation after late Cretaceous to Eocene collisions. Progradation results from regional compression across the suture combined with elastic-perfectly plastic downbending of the foreland. Unbending stresses past the zone of maximum flexure cause intracrustal decollement used by regional compression to shear off and lift basement prisms. On their way up, basement nappes cross Barrovian isotherms and zones of thrusting styles. Crustal data suggest that the elastic upper parts of the foreland flexure are overlain by largely allochthonous core complexes. Along strike, folds and backthrusts in the Alpine core complex grade into a largely unrecognized Apennine thrust belt and into the Calabrian subduction. A crustal synform beneath peninsular Italy, similar to gentle-dip zones in the Canadian and Montana overthrust belt and the Chubut Andes, may record elastic loading by the thrust belt.

ZUSAMMENFASSUNG

Der alpin-mediterrane Orogen-Bauplan ist weitgehend bestimmt von krustaler Einengung und Sutur-Progradation nach eoalpinen Kollisionen zwischen Oberkreide und Eozän. Die Rettung eoalpiner Hochdruck-Paragenesen vor der späteren Diaphthorese ist nur möglich durch Deckenschub unter krustaler Einengung. Progradation ist die Übertragung des Deckenschubes auf eine jüngere Schubfläche im Liegenden und im Vorland der älteren Schubfläche. Sie entsteht durch Einengung quer zur Kollisions-Sutur, verbunden mit elastisch-vollkommen plastischer Abbiegung der Vorlandplatte. Die Abscherung und das Aufsteigen von krustenparallelen Kristallindecken können in den kinematischen Gefüge-Abfolgen und in den meso- bis neoalpinen Abkühldaten verfolgt werden. Die elastisch gebogenen Bereiche der Vorlandplatten sind von weitgehend allochthonen Kernkomplexen überlagert. Im Streichen gehen die Rückfalten und Rücküberschiebungen des alpinen Kernkomplexes über in den kaum bekannten Apennin-Schubplattenkomplex und in die kalabrische Subduktion. Eine Krustenmulde unter Peninsular-Italien ist ähnlich gebaut wie die Unterlage von schwach einfallenden Monoklinalzonen in den Rocky Mountains von Kanada und Montana und in den Anden von Chubut. Vielleicht sind diese Vorland-Strukturen durch Belastung durch den Deckenstapel entstanden.

Introduction

At the close of the International Geodynamics Project, the orogenic model of crustal convergence (ARGAND 1924) appears generally accepted for the Alpine-Mediterranean area, and it is supported by data from all earth-science disciplines. Perhaps the most significant result of applying plate tectonic concepts is the rediscovery that most of the orogenic architecture is the product of post-collisional

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intracrustal compression, whereas the subduction stages have left hardly any structural trace. Therefore, geological work after the IGP will strongly depend on established correlations between mappable geology, crustal structure, and global tectonics. The present paper is a review of such correlations in the orogen segment between the Alps and the Calabrides.

Regional surveys

The present understanding of the Mediterranean orogeny, and that on which the present paper is based, is available in geochronological summaries (FREY et al. 1974, 1976; TRÜMPY 1973), in plate tectonic models (DEWEY et al. 1973, BIJU-DUVAL et al. 1977), in analyses of seismicity and Recent plate motions (MCKENZIE 1972), and in regional crustal data (MÜLLER et al. 1976, GIESE et al. 1978, GIESE & REUTTER 1978). As an addition to the referenced syntheses, Figure 1 presents a Mediterranean plate configuration at the beginning of the Eoalpine collision, projected upon the Campanian/Maestrichtian reconstruction by BIJU-DUVAL et al. (1977). It shows the craton of the Adriatic plate fringed to the northwest by oceanic crust of the Piemontese–South-Penninic realm, containing an oceanic outer arc. Fragmentation in the foreland is isolating smaller cratons, such as the Corso–Sardinian group and the Briançonnais/Mid-Penninic, and thinned-crust belts such as the Valais or North-Penninic or Rheno-Danubian trough. The east margin of the Adriatic craton has collided, during the late Jurassic Eohellenic phase (JACOBSHAGEN et al. 1976), with the prograding foreland of the main Tethyan subduction. Thus the Eoalpine collision could be regarded as a very large increment in Trans-Tethyan prograding collision.

Eoalpine tectonism

Radiometric cooling dates of Upper Cretaceous age are grouped as Eoalpine event (DAL PIAZ et al. 1972). High pressure parageneses in South-Penninic metapelites, calcschists, metabasalts, and ultramafites suggest reemergence of deeply buried material. Its protoliths can easily be interpreted as oceanic lithosphere, marine foreland sediments, accretionary wedge, and perhaps, in some contested dates (BEARTH 1974), as continental crustal fragments of the foreland plate. These rocks outline a suture located above the Mid-Penninic craton and below the Austro-Alpine sialic "Altkristallin" and its sediment cover. The suture is not yet everywhere dated as Eoalpine, but it can be followed from the Tauern window into Alpine North Africa. The geodynamic interpretation of the Eoalpine event will influence models of the closing Tethys and its plate geometry, the kinematics and geothermics of the Alpine convergence, and the crustal structure of peninsular Italy.

The task, therefore, is to decide between three interpretations of the Eoalpine event: a) it marks the closing of the South-Penninic–Piemontese marginal sea of Tethys (BÖGEL 1976, FRISCH 1977, BICKLE & HAWKESWORTH 1978). b) The event dates a collision between an oceanic arc complex and the Eurasian foreland (ROEDER 1978, ROEDER & BÖGEL 1978). c) The event is unrelated to collision, but requires a particular mechanism of emergence during the subduction of oceanic crust (HOLLAND 1979). The choice can only be made subjectively because there is no

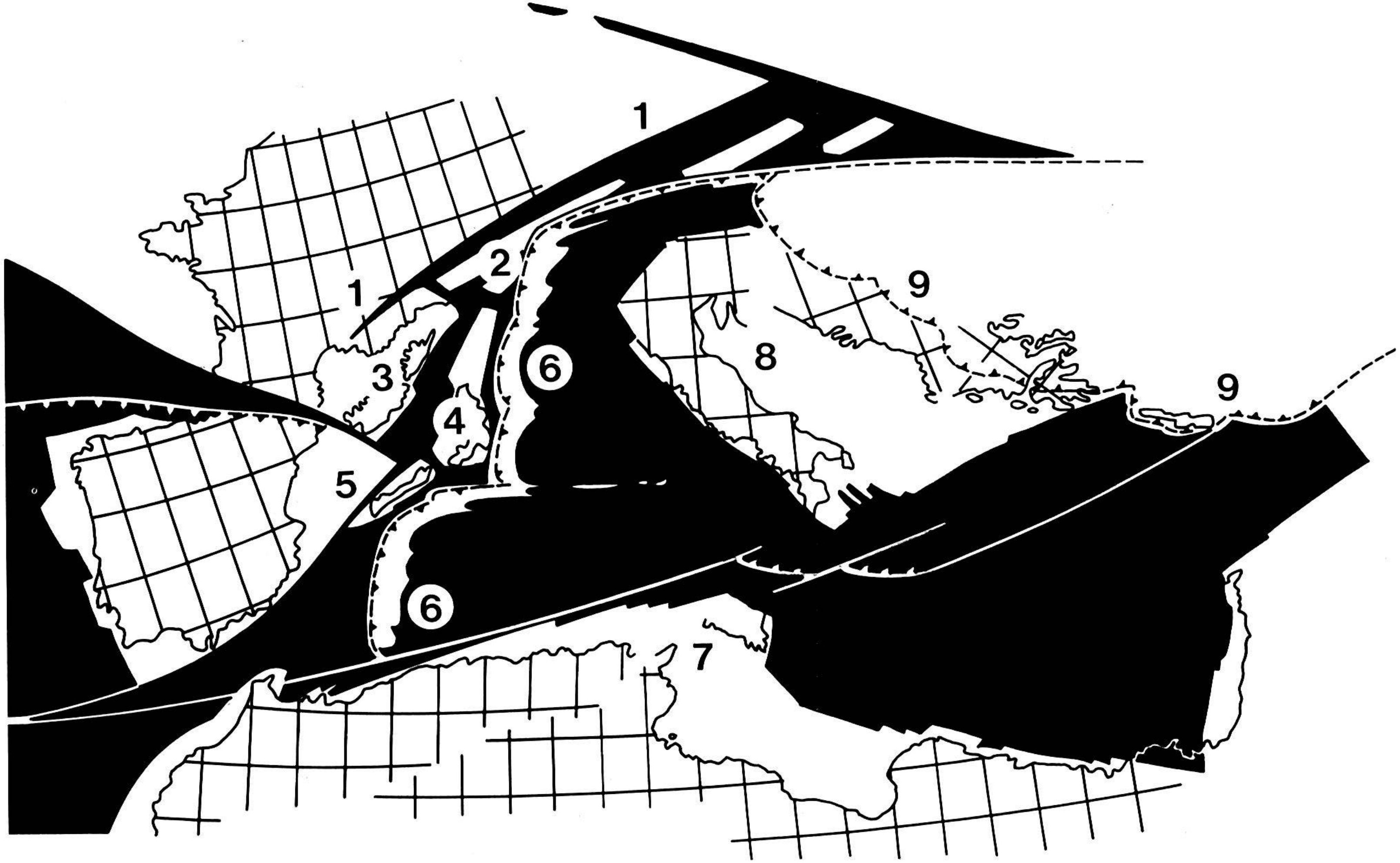

Fig. 1. Restored plate configuration, Mediterranean area, Campanian–Maestrichtian, 76–68 my. Rotations of major plates after BIJU-DUVAL et al. (1977). Outlines of some plates and rotations of smaller plates modified to suit orogenic history. *Black:* oceanic crust and, in some rift basins, mixed crust. *White:* continental crust. *Grid and outlines:* present day continents and coast lines. *Numbers:* paleogeographic areas referred to in the text: *1* = North-Penninic-Rhenodanubian, *2* = Mid-Penninic-Briançonnais *3* = Corso-Sardinian block, not yet rifted off Eurasia. *4* = Calabria and unnamed fragment in profile 5, *5* = Allochthonous Northern Sicily, *6* = Eoalpine subduction front in South-Penninic ocean, *7* = South Sicilian craton, *8* = craton of Adria plate, *9* = front of Eohellenic collision.

continuous chain of observation between petrological and structural data and the models of subduction and collision.

The Tellian/South-Penninic suture belt contains all rocks known from subduction complexes (GANSSER 1975, DIETRICH 1976). In some areas, the stacking order of thrust plates outlines a section across several zones of an oceanic arc complex (BEARTH 1974). Internal zones contain ophiolite bodies with largely serpentinized ultramafic rocks; they contain metagabbros, pillowed metabasalts, and oceanic metasediments. Great thicknesses in some sediment complexes suggest an origin in forearc or backarc basins. External zones contain oceanic sediments or accretionary wedge material in depositional contact with continental crust or its shelfal sediment cover. They are, in places, intermixed with melange-like fragments of ophiolites and undiagnostic metabasalts.

Thus the arrival of an accretionary wedge on the Mid-Penninic craton during the late Cretaceous is sufficiently documented in the Mid-Penninic and external Sub-Penninic stratigraphy. However, the plate tectonic role of the more internal units is not clear. The ophiolites may represent either remains of the upper plate or lower-plate slivers that were detached and transported with the upper plate by suture progradation (ROEDER 1979). Associated sediments show widely distributed pelagic facies, locally derived breccias, and fine-grained clastics containing chromite of contested (ROEDER 1976, TOLLMANN 1978) but remote provenance. These rocks are conceivably parts of a forearc basin with block faulting and with thin continental or oceanic crust.

Mid-Mesozoic deposits in the North-Penninic trough resemble those of the external South-Penninic realm, perhaps because the subduction complex progrades in the external direction.

High-pressure metamorphism is present in South-Penninic metasediments and metabasites, though widely obliterated in the Meso- to Neo-Alpine retrogradation. In the Mesozoic peripheral "Schieferhülle" surrounding the Central Gneiss basement of the Tauern window, the Eoalpine metamorphism is preserved, showing imbrications and polyphase folded contacts between quartz-kyanite eclogites, calcschists, and retrograde blueschists. Dependent upon the assumed role of the water, these eclogites formed at 8-11 kbar and 400-500 °C (MILLER 1974) or at 17-21 kbar and 590-650 °C (HOLLAND 1979). The resulting finite geothermal gradients of 17-24 °C/km or 12-15 °C/km reach the quartz-kyanite eclogite field of the p/T-diagram directly crossing the glaucophane field (WINKLER 1976). Therefore, the synkinematic eclogite stabilization and its retrogradation to blueschists can be explained by a single process of emergence. White-mica ages date it as Eoalpine (BICKLE & HAWKESWORTH 1978). Petrogenetically, therefore, the significance of the Eoalpine orogeny lies in the eclogite reemergence. A structural aspect of this orogeny is provided by the need to unravel the polyphase Alpine thrust stack.

Mechanisms of reemergence

The salvation of the Eoalpine Sanbagawa parageneses from later Alpine Barrovian retrogradation requires that the buried rock complexes reemerged soon enough, high enough, and fast enough. Since the reemergence implies a tectonic mechanism, the petrology of the high-pressure rocks is of general geological interest. Three mechanisms of reemergence have been proposed in the literature:

1. The high-pressure rocks rise as isolated blocks within the subduction zone, surrounded by accretionary-wedge material and moving in the opposite sense of subduction (ERNST 1972). 2. Folding, "rétrocharriage" or back thrusting, and formation of an Alpine root zone may lead to reemergence of the anticlinal core depending on the length of the front limb and on the fold interlimb angle (ROEDER & BÖGEL 1978). 3. The transfer of the plate convergence from a thrust plane above the accretionary wedge to a thrust at its base may reexpose deep-seated material. Since it requires upward movement of the upper plate relative to erosional base level, this process is limited by the view that plate convergence is powered only by gravitational subduction and not by regional compression (ELSASSER 1971).

For estimates of the time/distance parameters of reemergence, several Alpine geothermal models can be used. The "Schieferhülle" eclogites emerged from a maximum depth of 54 km (HOLLAND 1979) at least to their present depth below the restored surface of the Austro-Alpine complex, estimated at 13–15 km (FRANK 1969). However, since high-pressure rocks occur also outside the probable extent of the Austro-Alpine, estimates are made for emergence to the post-Eoalpine surface.

Subduction of surficial material causes a depression of the isotherms which is counteracted by heat flow (MINEAR & TOKSÖZ 1970). The one-dimensional effect of heat flow, therefore, is equivalent to additional burial. A high-pressure paragenesis is preserved if it emerges at least as fast as the rise of the isotherms. The Eoalpine cooling dates show, however, that the reemergence happened faster than the theoretical minimum.

A geothermal model of the Meso- and Neo-Alpine orogeny (BICKLE et al. 1975, Fig. 7) may serve to estimate the minimum emergence rate. According to this model, the Alpine heat flow would raise the isotherms by 0.09–0.14 cm/y. In order to cool, the high-pressure rocks must rise faster than these values. If reemergence is achieved by folding, recent uplift in the Western Alps may serve as a comparison; it amounts to 0.1–0.17 cm/y (GUBLER 1976). Cooling models for the Lepontine based on mineralogical and radiometric data (WERNER et al. 1976) show maximum uplift rates of 0.3–0.5 cm/y. If reemergence is achieved by thrust transport against the sense of subduction, plate models can furnish rates. BIJU-DUVAL et al. (1977) assume, for the interval between 140×10^6 and 44×10^6 y, rates across the Western Alps of 0.4–0.6 cm/y. This time interval spans the burial as well as the reemergence of the high-pressure suites. Assuming equal rates for subduction and reemergence, and assuming that the emergence component may constitute a quarter of the movement of uphill thrusting, one derives an emergence rate of 0.2–0.3 cm/y. Thus there is no rate distinction between folding and thrusting, both achieving rates sufficient to preserve and cool down the high-pressure assemblages.

For a structural interpretation of the previously believed lesser p/T-conditions (MILLER 1974), an Eoalpine backfold or root zone appeared sufficient although it could not be shown in the exposed Alps (ROEDER & BÖGEL 1978). However, if one must explain a reemergence of 54 km (HOLLAND 1979), only subduction-parallel thrusting can be considered. Also, it is compatible with described Eoalpine fabrics of melange superposed by thrusting, compositional foliation, and boudinage (Bickle & Hawkesworth 1978, Holland 1979).

Emergence by thrust transport within a subduction complex has two prerequisits (Fig. 2): 1. The thrust plane must be located above the material to be buried and below the reemerging material. The reversal of movement documented in the high-pressure rocks supposes a progradation of the main thrust contact into the foreland. 2. The emergence from a depth of 50 to 60 or more kilometers supposes an equal amount of erosion which must be maintained at the thrust front by a transport over 150 or 200 km. In the arc-trench model by Karig & Sharman (1975), the front of

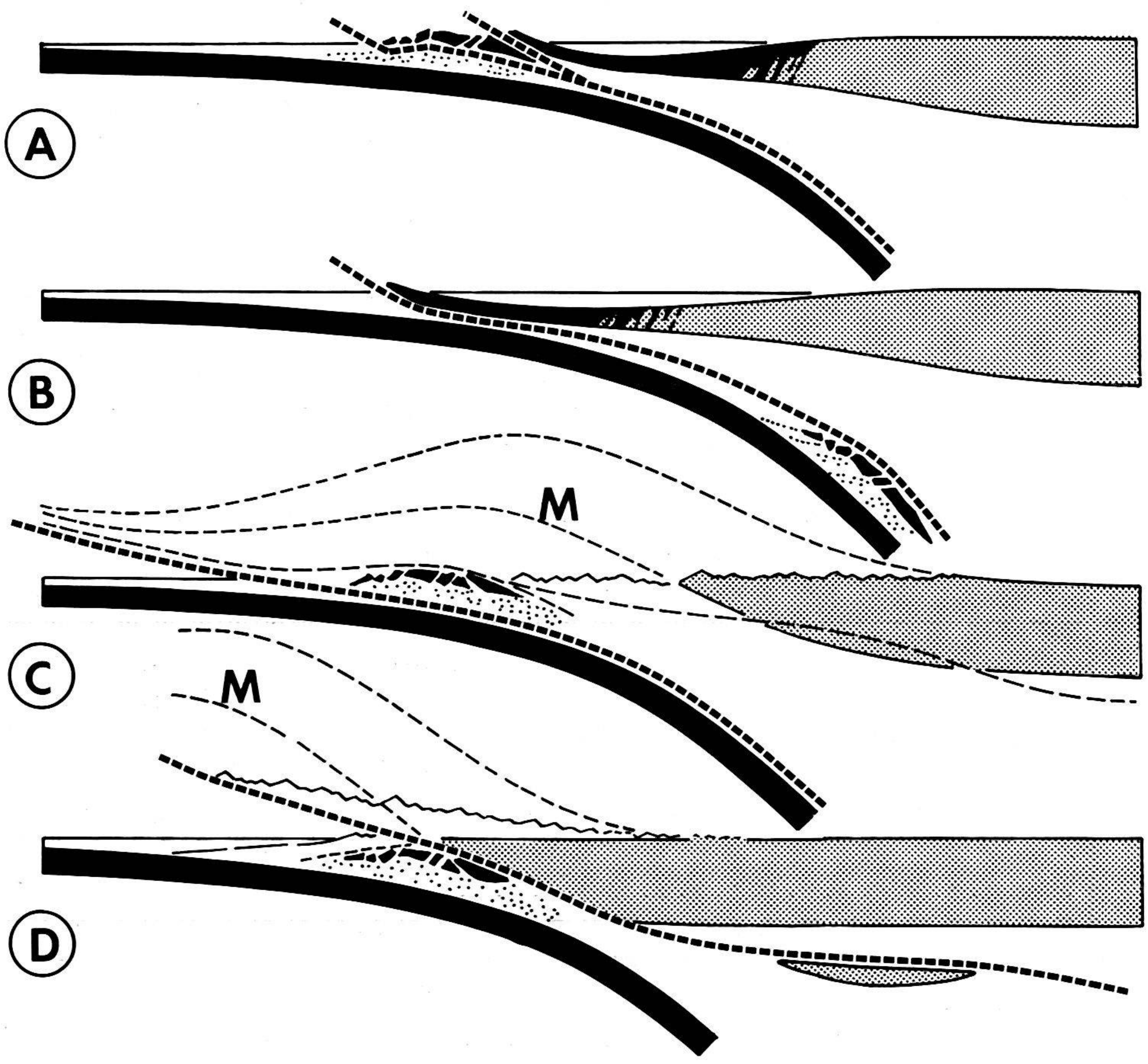

Fig. 2. Evolution of an imbricate paired belt, in schematic cross sections. *Black:* oceanic crust. *Shaded:* continental crust. *Thick dashed:* active thrust. M = Moho. A = oceanic subduction complex (or outer arc) with continental hinterland. Accretionary wedge of melange and fine clastics. B = thrust internal to outer arc and absolute downward motion of foreland subducts accretionary wedge. C = thrust external to accretionary wedge, absolute upward motion of upper plate, and commensurate erosion of plate margin, lead to reemergence of high-pressure belt. D = layer-discordant low-angle thrust internal to accretionary wedge creates tectonic contact between high-pressure belt and high crustal area or arc region of upper plate.

the outer arc is an appropriate location for the outcrop of such a thrust surface. Depending on the preceding geological history of the subduction system, this thrust brings into the realm of erosion, accretionary-wedge material, lower-plate slivers, and/or the upper plate itself. Very large thrusts required by the high-pressure rocks, are consistent neither with gravitational gliding or gravity spreading as the dominant transport mechanism (ELLIOTT 1976, CHAPPLE 1978), nor with models of purely gravitational subduction without regional compression across strike (ELSASSER 1971).

The emergence of a coherent high-pressure belt not necessarily indicates collision, as suggested by emerged circumpacific high-pressure belts (ERNST 1972, MIYASHIRO 1973) and the Recent thrust structures in the outer wall of the Peru-Chile trench (PRINCE & KULM 1975). Nonetheless, the basal zone of the Glockner nappe (peripheral "Schieferhülle", Tauern window) contains melange clasts of miogeoclinal origin. They show (FRISCH 1979) that the Eoalpine emergence was touched off by a collision between the subduction complex and the Mid-Penninic craton.

Suture progradation in the Alps

Alpine evolution after the South-Penninic collision consisted of polyphase nappe transport, folding, and extensional block faulting in the lower plate, and foredeep sedimentation. These processes can be interpreted as a progradation of the collision suture into its foreland. As suggested in plate models, the Trans-Alpine convergence continued undiminishedly. Some fixed marker points in this evolution include the following:

Tectonic contacts above the peripheral "Schieferhülle" originated during the "transport that shut the Tauern window" (OBERHAUSER 1964); they have a Paleocene or early Eocene age. The 13–15 km thick cover of the Austro-Alpine thrust masses must have been in place at least 30×10^6 y prior to the Mesoalpine temperature peak (BICKLE et al. 1975). A Paleocene faunule from the Prättigau flysch is consistent with this inference. The Austro-Alpine sole fault is discordant to the structure within the Austro-Alpine; it cuts discordantly across thrusts and fold axial surfaces in the underlying Lower "Ostalpin" (TOLLMANN 1968).

The thrust contact between Upper "Ostalpin" (above) and Rhenodanubian flysch is not older that the youngest (late Eocene) flysch. This time lag between the Tauern window and the north front of the Calcalpine, of 15 to 25×10^6 y, suggests transport below the "Zentralgneis" complex of the Tauern window. In the Alpine models of FRISCH (1977, 1979) and TRÜMPY (1973), this transport is interpreted as subduction within the North-Penninic trough.

The thrust contact between the Rhenodanubian flysch (above) and shelfal Helveticum (below) is of post-Lattorfian age and coeval with increased rates of sedimentation in the Molasse, as well as with the Meso-Alpine temperature peak in the Lepontine and in the Tauern (LEMCKE 1978). These processes suggest convergence between the Penninic crystalline core complex with its Austro-Alpine cover and the Eurasian foreland. Although they suppose thrust transport below the Tauern, a complication is provided by post-kinematic and strongly discordant Molasse deposits of Lattorfian (early Oligocene) age along the internal edge of the southwestern Alps (TRÜMPY 1973). This constrains the structural style of the Mesoalpine thrusting.

Alpine neotectonics consists of uplift or large-scale folding in the middle of the orogen, and of outward low-angle thrusting at the margins. Seismicity indicates that the 200 km wide mobile Alpine body is rather thin, 15–20 km at the south margin (WITTLINGER & HÄSSLER 1978), 5–7 km at the north front (KOCH 1979), and probably not much thicker in the middle. The structure of the equivalent compression at depth is unknown, hardly a new insight (AMPFERER 1906).

Suture progradation (ROEDER 1979) combines the processes related to the downwarp of the foreland plate with the transport and imbrication of the upper plate. Models of elastic-perfectly plastic flexure (TURCOTTE et al. 1978, FORSYTH 1979) outline a zone of strongest flexure in front of and below the collisional suture. This zone shows a layering of the plastic-ruptural deformation, consisting of extension in the upper part and of compression in the lower part (Fig. 3). The layering appears to induce plate-parallel decollement within the crust at a depth determined by the effective thickness of the foreland plate. The decollement becomes the sole fault of a crustal thrust plate which is plucked out of the lower plate, deformed, and attached kinematically to the upper plate. This new suture

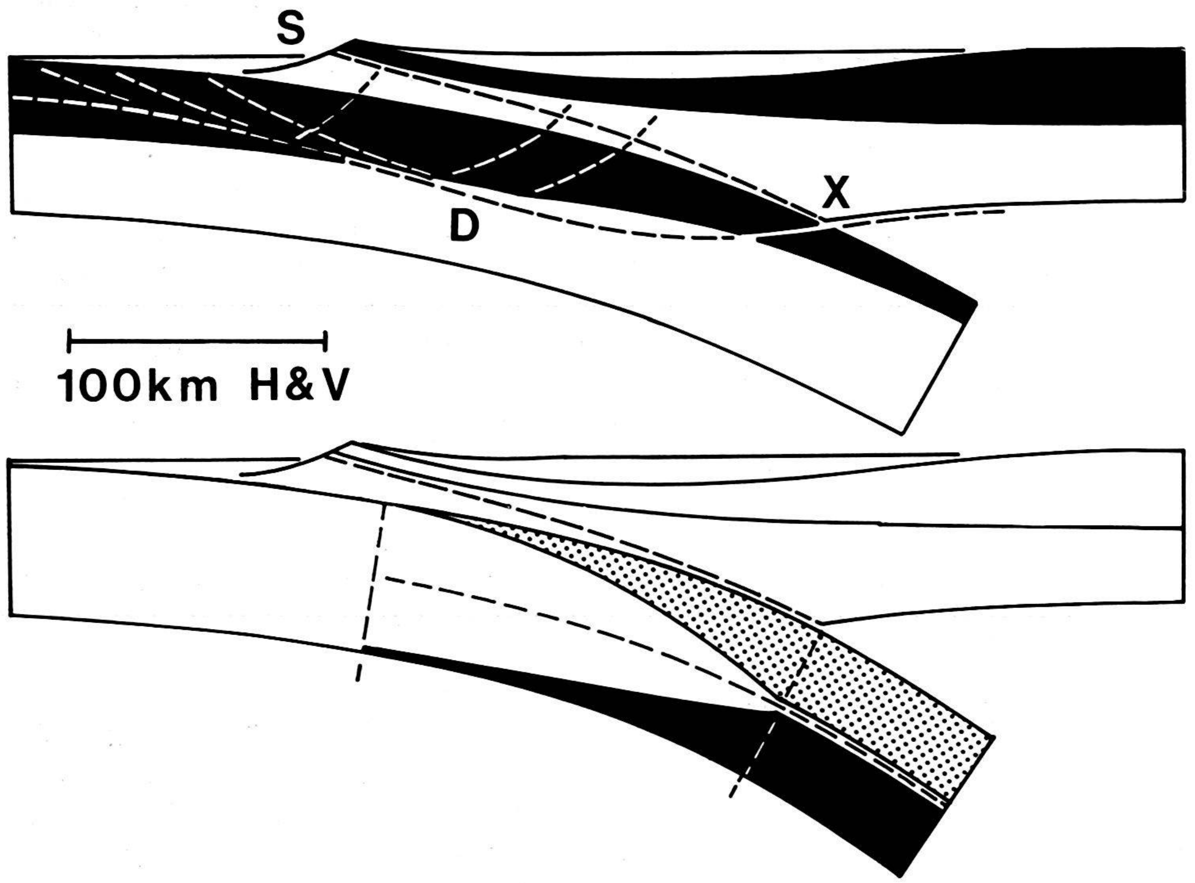

Fig. 3. Combined model of suture *S* prograding into a lower plate, in two schematic structure sections. *Above:* regional compression indents lower plate, intersects suture at *X* and mobilizes crustal decollement *D* (after ROEDER 1979). *Below:* Elastic-perfectly plastic flexing of lower plate creates layer of extension *(shaded)* above layer of compression *(black)* separated by decollement *D* (after TURCOTTE et al. 1978).

may begin by separating an upper folded layer from a lower, much less disturbed layer. Simultaneously or later transport begins, restoring the original geometry of thrusting and rising orogenic body and sinking foreland. As has been observed in other orogens (MOLNAR et al. 1976), this process is cyclical and eminently capable of combining phase-like paroxysms in a "moderately Stillean" sense (TRÜMPY 1973) with the model of rather steady-state plate motions. The geologically determined thicknesses of the plucked and mobilized lower-plate chips may vary between 10 and 50 km (ROEDER 1979); it is sufficient for placing the sole fault into the depth range of high-rank metamorphism, anatexis, and the crystalline thrust sheets of ARMSTRONG (1974).

In the Alps, the post-collisional orogenic history is documented with surface data; it can easily be interpreted as a sequence of intracrustal progradations. Three large increments are vaguely suggested; they succeeded one another at intervals of about 20×10^6 y and mobilized crustal prisms about 8 km thick.

Kinematic sequences

Kinematic sequences of tectonic deformation constitute a connection, of great future importance, between plate tectonics, crustal geophysics, and practical or mapping geology. Published kinematic sequences near the Eoalpine suture cannot yet be fully correlated. It is not even certain if there are distinct and locally repeated types of sequences in the Mediterranean orogens. Nevertheless, some common traits are recognizeable (Fig. 4):

In the western Penninic Alps, one or several generations or progressive tight to isoclinal folds with pervasive axial-plane cleavage can be correlated with the Eoalpine metamorphism (GOSSO et al. 1979) or with the thrust closure of the Tauern window (CLIFF et al. 1971). Transposed foliation in the western part of the Tauern window (ROEDER 1976) may also belong to this generation. In the Lepontine (MILNES 1978, VOLL 1976, HUBER et al. 1979), the Eoalpine or somewhat younger nappe transport is reflected in compositional layer-parallel shear planes as the oldest fabric element. Near the suture, a contact between ophiolitic and cratonic rocks is sufficient evidence of this transport.

Post-collisional folding evolves in several generations and in styles between recumbent and upright. Folds of this group of generations commonly show axial-planar crenulation cleavage. The last fold generations are chevron folds or kink bands with north-dipping axial planes; they are particularly common in the steep middle limbs of the backfolds.

The distribution of these folds may be governed by the dynamic zonation of thrust-fold belts observed in many orogens (ROEDER 1967, 1973, PRICE 1972, ELLIOTT 1976). The zonation consists of a frontal zone with compression externally and at shallow depth, and of a transport zone with transport in the extending-flow or compressing-flow mode internally and below the frontal zone. With this regional distribution assumed, the Alpine kinematic sequences appear as increments in the development of a prograding suture (ROEDER 1978, Fig. 2).

In this interpretation, the *postnappe recumbent folding* (MILNES 1974) appears as frontal-zone folding above an intracrustal decollement below and external to the

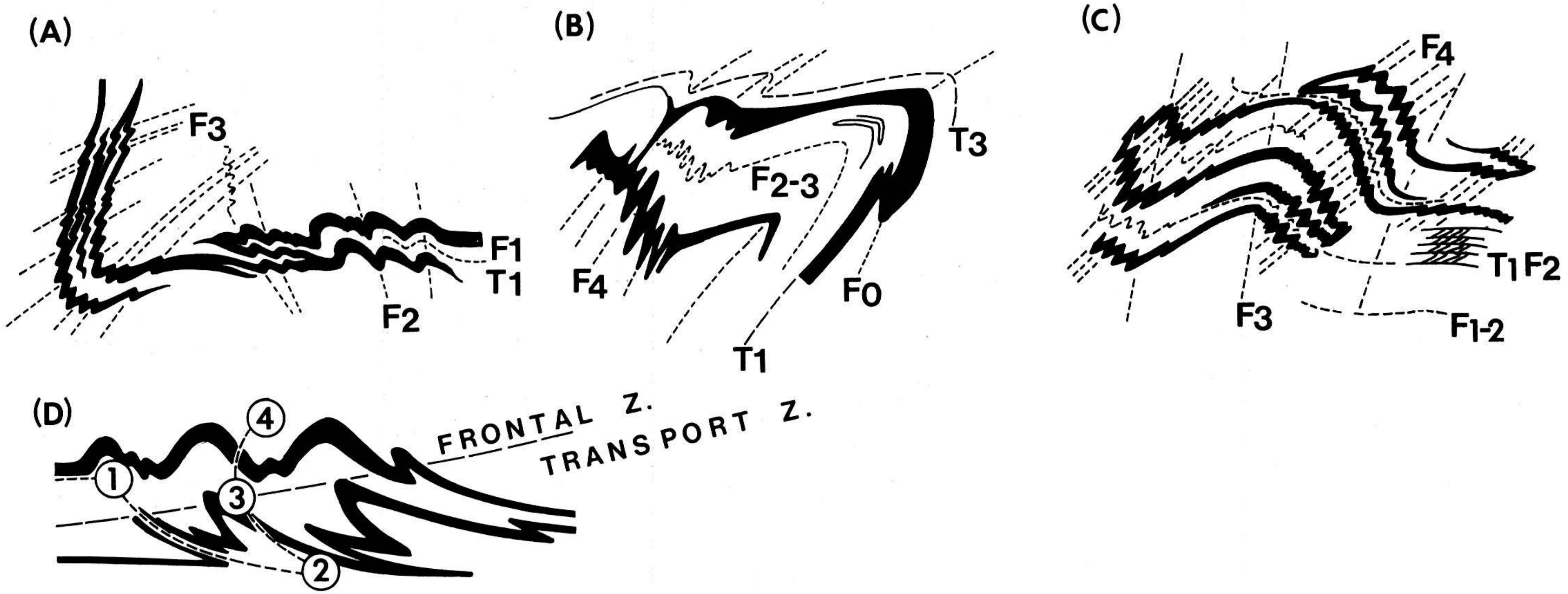

Fig. 4. Synopsis of some Alpine kinematic sequences, showing schematic cross sections drawn after published data. A = composite of sequences along north edge of Lepontin (HUBER et al. 1980). B = Penninic nappes (MILNES 1974, 1978). C = western Tauern window (ROEDER 1976). D = possible strain path of a mesoscopic realm across dynamic style zonation of nappe tectonics.

Fig. 4*a*. Location map of five structure cross sections following geotraverses in the Alpine-Central Mediterranean area. Sections are shown on Figures 5–7.

Eoalpine suture. Originally formed upright, these folds were converted into recumbent folds by the external rotation and under the increasing load of the Austro-Alpine nappe stack. The material, initially deformed in the frontal zone, thus became part of the expanding transport zone. As further acts of progradation formed new decollements deeper in the crust, the recumbent folds once more became part of the frontal zone; their material cooled and developed new generations of compressive structures.

Emergence and cooling were probably accompanied by a style change from pervasive flow to brittle deformation. This style change could be significant in the dating of the orogenic acts. Groups of radiometric dates in the Lepontine and in the Helvetic massifs show that tectonic fabrics are outlasted by crystalloblastesis (FREY et al. 1976, KLEMM et al. 1978), and that post-kinematic cooling dates coincide with periods of increased sedimentation rates (TRÜMPY 1973, LEMCKE 1978). As emphasized by TRÜMPY (1973), this relationship contradicts the correlation between tectonic and sedimentary pulses found in other orogens (PRICE & MOUNTJOY 1970). Perhaps this contradiction suggests that the cooling is post-kinematic relative to pervasive deformation, but coeval with transport and thrusting on brittle-ruptural sole faults and against erosion. This distinction would again establish a coincidence in time between tectonism, mountain building, erosion, and sedimentation.

Elastic elements in foreland deformation

One of the most important criteria for the understanding of orogeny is the geometry of the autochthonous foreland and the surface of its basement. In forelands with subsurface control (BREYER 1959, BALLY et al. 1966), this surface dips, gently arcuately, underneath the nappe stack, and it is the point of departure for any structural subsurface mapping (ROYSE et al. 1975, ROEDER et al. 1978). Therefore, methods are of interest with which one can determine the position of the basal decollement and its extent underneath the orogenic core complex.

Orogenic forelands and oceanic subductions show enough plate tectonic and structural similarities to justify their mutual use as analogies. Presently available studies (reviewed by FORSYTH 1979) show that oceanic forelands behave as elastic plates floating on a plastic substratum and being loaded at their free edge. The resulting profile of epeirogenic foreland structure is an exponentially damped cosine curve (HETENYI 1946, LE PICHON et al. 1973); it contains a peripheral upwarp and an increasingly flexed marginal downwarp. The distance between the peripheral bulge and the free edge determines the flexural rigidity as a function of the elasticity and the effective thickness of the plate. Rigidities derived from the shape of the flexure coincide poorly to fairly well with values derived from known loads (WALCOTT 1972), suggesting anelastic elements in the deformation. The vertical epeirogenic movements depend on the load. Given equilibrium conditions, it can be assumed that the elastic stress is absorbed anelastically at a time constant of 10^5 y. Since the elastically formed flexure has an instantaneous mold-and-cast relationship with the foredeep sediments and the allochthonous complexes, there is no anelastic recovery of flexural strain. With this mechanism, deformation can accumulate over geological time spans by the principle of superposition (KARIG et al. 1976), although its geometry can be described with equations of elastic deformation.

The plastic-ruptural component of deformation increases trenchward. The "free edge" of the oceanic foreland is located beneath the arc-trench complex, where the maximum flexure maintains the entire deformation in the plastic field (CHAPPLE & FORSYTH 1979).

Type and magnitude of the end load are not clear. Similarly flexed oceanic forelands have been explained by very high regional compressive stress (WATTS & TALWANI 1974), as well as solely by the moment of the subducted slab (PARSONS & MOLNAR 1976). The vertical load of the accretionary wedge contributes to the elastic flexure (KARIG et al. 1976). Even the elastic nature of deformation can be substituted by viscous flow (DE BREMAECKER 1977). In spite of these and other doubts about the physical aspects, some empirical applications seem to promise better insight into orogeny.

Stratigraphic, reflection-seismic, and crustal data of Mediterranean orogens (GIESE et al. 1978) show that the foreland basement tops and Moho segments dip underneath the core complexes, as if they were elastically flexed plates on plastic substrata. This architecture suggests that the forelands have a rheology that is fundamentally different from that of the core complexes, in contrast to views expressed about *subfluence* being a major orogenic agent (AMPFERER 1906, SCHMIDT 1976, BEHR 1978). The elastic-flexure model supports not only the inferred allochthony of the core complexes, but also the stratigraphic and regional timing of the kinematic sequences. In addition, deviating flexures beneath some core complexes appear to illustrate the conversion of a cool foreland plate into a new hot core complex.

Crustal and stratigraphic data suggest that in all Mediterranean orogenic forelands, the basal decollements dip with less than three degrees. The foreland lithosphere, therefore, remains largely in the field of elastic deformation (FORSYTH 1979). Centers of epeirogenic uplift in the orogenic forelands, such as the Vosges-Black Forest complex, the Corso/Sardinian complex, and the North Saharan line of basement arches, are interpreted as peripheral bulges of orogens. Their distribution and shape, however, are severely disturbed by three factors: 1. The flexural rigidity of the very inhomogeneous circummediterranean cratons may change locally. 2. Synorogenic plate fragmentation between the orogens and the opening Atlantic contributes to the inhomogeneity. 3. The orogenic loops lead to three-dimensional elastic foreland flexure, with domal bulges interspersed between relative depressions.

Location and geologic nature of the "free end" of the elastically flexed foreland slabs are poorly understood, but they constitute a critical element of the structural models proposed, for two reasons: 1. Its location defines the "wavelength" of the flexure and its reciprocal, λ, termed the *flexural parameter* (WALCOTT 1972). 2. The presence of a downwarped "free end" implies that any orogenic complex above the "free end" must be allochthonous. Paleogeographic data, however, show that the Mediterranean core complexes consist largely of foreland material and must have been transported over the foreland by suture progradation. Therefore, the location of the "free end" is a measure of thrust transport beneath the core complex.

Crustal data (GIESE & REUTTER 1978) show the Moho extending from the forelands to mappable edges underneath the orogenic core complexes. Although the structure of the crust above the mapped edges is unkown, the latter are assumed as "free edges" of the forelands slabs, until a better definition can be found.

In the modelled cross sections, the flexural parameter λ of the foreland crust has been derived from the distance x_0 between the "free edge" and the distal pinchout of the foredeep fill, using the equation $x_0 = \pi/2\lambda$. The crests of the peripheral bulges are difficult to pinpoint structurally or morphologically, but their distance x_{max} has been determined as $x_{max} = 3\pi/4\lambda$. The obtained distances vary between 200 and nearly 400 km; they lie between the 500-600 km typical for some major Archean cratons, and the oceanic 100-200 km (CALDWELL et al. 1976). The implied distances of transport below the core complexes are very uncertain because of the poor definition of the "free edges". A uniform distance of 100 km of transport resulted from data and assumptions.

Profile 1: Eastern Alps

This cross section (Fig. 5, above) coincides with the Geotraverse I*A* of the "Deutsche Forschungsgemeinschaft" (BÖGEL 1976). It begins at the northern Molasse pinchout near Ulm, crosses the Folded Molasse south of the Eberfing well (MÜLLER 1978), the Northern Calcareous Alps near the Vorderriss well (KOCH 1979), the west end of the Tauern window, the Periadriatic line near Pfunders, and the Alpine south front near Belluno.

The foreland is modelled as an elastically flexed plate of which only the Mesozoic shelf cover (black) is shown thinning over the Vindelician subsurface high. Based on stratigraphic data and on its assumed extent to the south edge of the mid-European Moho, the flexure has been assumed to parallel the Moho (GIESE et al. 1973). The flexural parameter is equivalent to a distance between "free edge" and peripheral bulge of 396 km, the latter being located in the southeastern Odenwald area. The model suggests that the foreland crust is overridden in full thickness, perhaps including a complete shelf-sediment cover, by the Tauern core complex. The required 100 km of thrust transport is easily achieved by plate convergence of 160 km along a northwesterly vector since the Lutetian (BIJU-DUVAL et al. 1977).

The location and attitude of the crustal break at the origin of the core complex is uncertain. If the core complex was derived from a foreland area south of the maximum bending, then the vaguely known steep segment of lithosphere (MÜLLER et al. 1980) is a likely locale. The assumed shape of the core complex is derived from the model of suture progradation. Folding after transport formed the present Tauern anticlinorium.

Profile 2: Western Alps

This section (Fig. 5, below) follows the Geotraverse Basel–Chiasso of the Swiss National Committee for the IGP. The geological profile has been taken from BÜCHI & TRÜMPY (1976) with the modification that at the buried foot of the Aare massif, a thrust with 100 km of slip extends into the fill of the foredeep. Moho data (MÜLLER et al. 1976, GIESE et al. 1978) have been translated vertically by 30 km and projected into the profile. The flexural model chosen is somewhat "stiffer" than suggested by the curvature of the Moho, but it is constrained by the reasonably known base of the Molasse basin and its depth. The difference of about 4 km below the external massifs suggests that the curvature may increase southward faster than in purely elastic deformation, or that the crustal thickness increases southward.

The flexural model has its zero deflection within the narrow Folded Jura, resulting in a flexural parameter of $\lambda = 1.066 \times 10^{-7}$ and a distance from the "free edge" below the Tonale line to the peripheral bulge north of Freiburg in the Black Forest, of 221 km. The West Alpine crystalline complexes are considered allochthonous and overlying an autochthonous cover of Mesozoic sediments.

The course of the Moho shows a 10 km deep syncline below the Lepontine (MÜLLER et al. 1976) which has not been considered in the present model. It may reflect nonelastic deformation near the "free edge", such as drag at upthrusts of the Ivrea body of the Austro-Alpine (upper-plate) crust. Alternatively it could result from the vertical load of the core complex. If the cross section of the core complex is modelled as a triangle 20 km high and 100 km wide with a sialic density of $\rho = 2.6$, it produces (KARIG et al. 1976, equation 3) an elastic flexure of its substratum of 4.7 km. A density of the core complex of $\rho = 2.8$ would increase the downwarp to 10 km.

The internal structure of the core complex is depicted as a stack of 8–10 km thick crystalline nappes with internal folds which during the last generations of south-vergent folds was compressed and uplifted. The pointed south edge suggests a cuspate shape of the initial detachment. However, the synthesis between analytical structural geology and plate tectonics, a complete unfolding of the core complex (HUBER et al. 1980) may contain structural surprises.

Profile 3: Maritime Alps to Padan Plain

This and the following two profiles are based largely upon crustal data and interpretations by GIESE & REUTTER (1978) and by REUTTER et al. (1978), with the

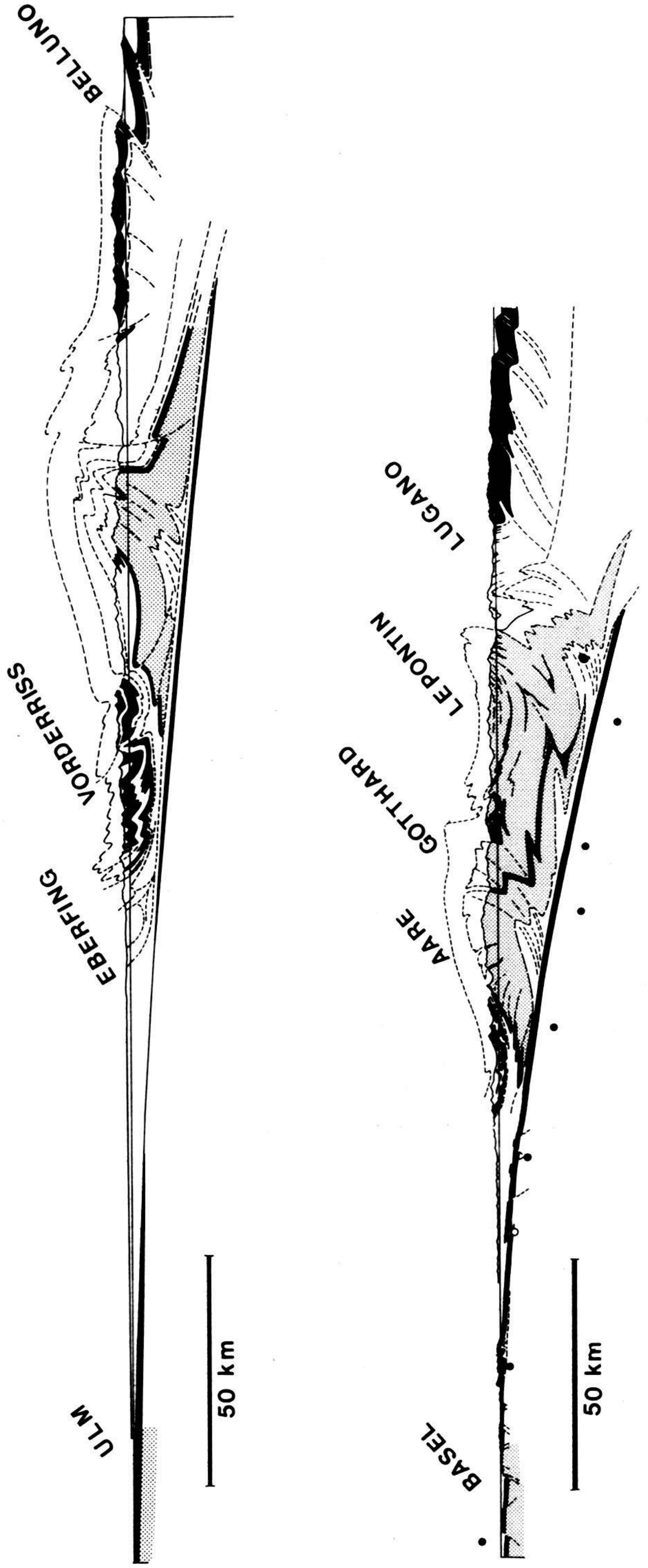

Fig. 5. Structure sections across the Alps based on models of elastic foreland flexure. Alps, German geotraverse *IA*. *Bottom:* Swiss Geotraverse Basel–Chiasso. *Black:* Mesozoic shelf sediments. *Shaded:* core complex. *Black dots:* subsurface elevations of Moho (MÜLLER et al. 1976) projected 30 km upward.

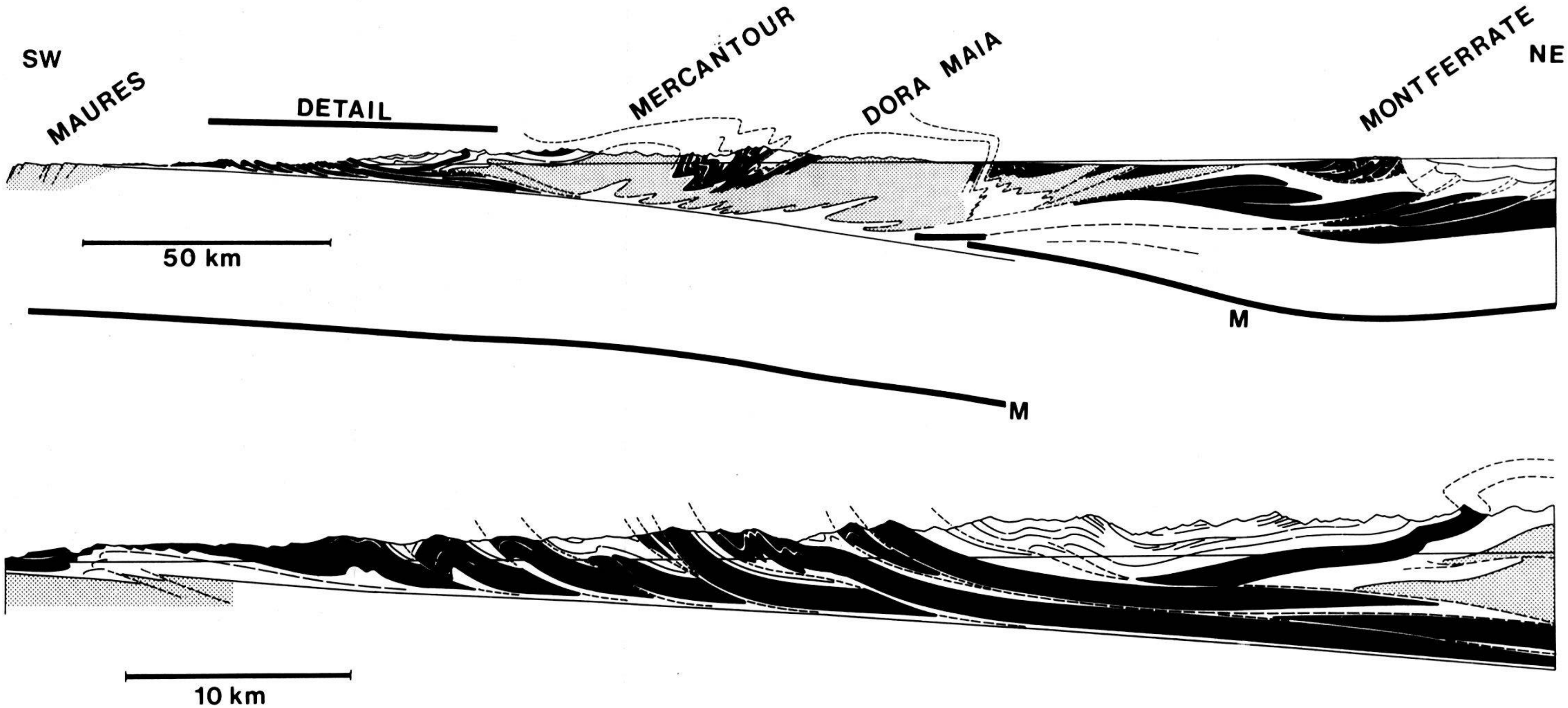

Fig. 6. *Top:* Section across Maritime Alps and Northern Apennine with shelfal Mesozoics *(black)* and allochthonous core complex *(shaded)*. *Bottom:* detailed section of pre-Alpine thrust belt based on model of elastic foreland flexure and on surface data by D. Roeder, J. Barrier, J. Gérard and R. Marty (1964–5).

assumption of elastic deformation and its structural consequences added. This modification tends to support the geodynamic interpretations by these authors.

The profile (Fig. 6) begins at the peripheral bulge of the Maures-Estérel and crosses the Prealpine thrust belt and the Alpine core complex, consisting of the Helvetic (Dauphinese) Mercantour massif, the Mid-Penninic Briançonnais and the Sub-Piemontese Dora Maia massif. Similar to other Alpine profiles, the core complex shows an antivergent internal structure which is truncated beneath an early Oligocene series (TRÜMPY 1973) and thus is older than elsewhere in the Alps. However, the northeasterly vergence continues beneath the Neogene cover of the Padan Plain, increasing to low-angle thrust sheets as exposed in the allochthonous Montferrate hills near Torino. Movement in this northernmost zone may belong to the Tuscan phase affecting the Apennine in late Miocene and younger time.

The flexure of the Eurasian plate is dimensioned by the zero deflection point at the north edge of the Maures-Estérel and by the east edge of the Eurasian Moho. One obtains a flexural parameter $\lambda = 8.48 \times 10^{-8}$ and a distance between peripheral bulge and "free edge" of 278 km.

Crustal data clearly show the synclinally warped Adria plate lying on the crust of the Eurasian plate. The contact dips at 3° to 9° underneath the Adria plate, connecting the basal decollement of the Subalpine thrust belt with the northeasterly dipping Adria Moho. This configuration is different from that of the Alps farther north, where the basal decollement in the foreland lies 5–20 km deeper than the Moho of the Adria plate. These variations can hardly be explained in another way than as conjugate thrusting at the Alpine base and in the Apennine. The importance of this *flake tectonics* (OXBURGH 1972) has been emphasized by GIESE & REUTTER (1978). The decollement beneath the Alpine core complex has been drawn without shelfal sedimentary cover in this section, because these sediments appear to be piled up in the pre-Alpine thrust belt.

Profiles 4 and 5: North and Central Apennine

These profiles (Fig. 7) cross the Apennine at a distance of only 100 km and therefore show similar crustal structure. However, they supplement one another in the details which suggest that the Apennine is a foreland-loading east vergent thrust belt, despite the differently described surface geology (ABBATE et al. 1970). The northern profile crosses the North-Tyrrhenian shelf which is probably underlain by parts of the Alpine core complex and perhaps by Ibero/Eurasian foreland. The Apuan Alps and the Tuscan structures are accessible thrust complexes (CARMIGNANI et al. 1978). A shallow refraction-seismic horizon shows that the basement is involved in thrusting. The external units are so thoroughly covered by chaotic accretionary-wedge material that even their existance is met by some doubt. The foreland structures in the Padan-Plain subsurface, however, can easily be recognized as frontal folds of post-Miocene thrust sheets (BALLY et al. 1966).

The southern profile shows a similarity in the downwarped Adria plate and in the plastic-ruptural deformation of its marginal belt. This plate is in complex contact with the Ibero/Eurasian foreland, in the form of the basement of the island of Corsica being overlain by the surface-geologically defined Alpine belt. Crustal

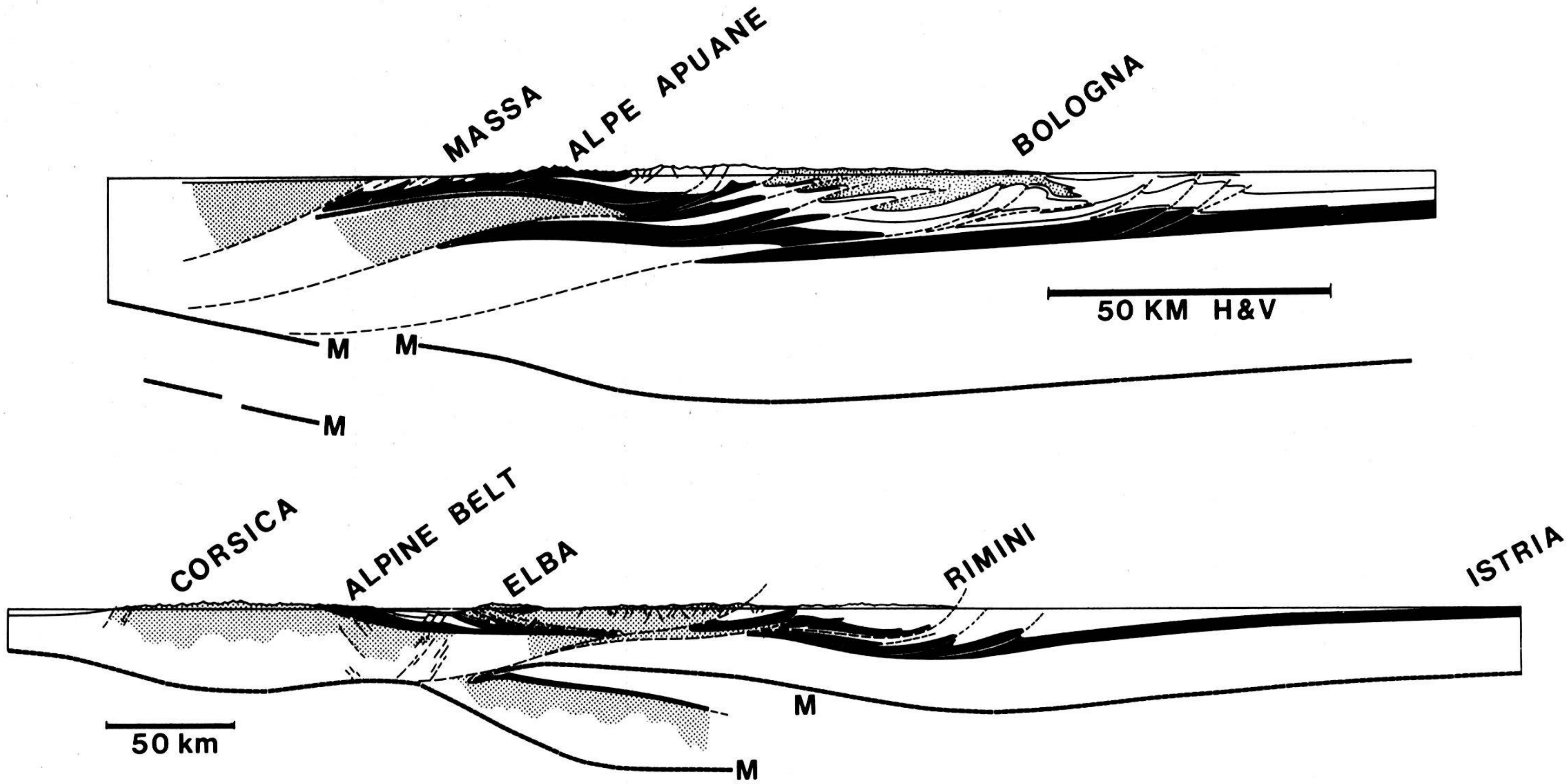

Fig. 7. Two sections across northern and central Apennine, combining crustal data (GIESE et al. 1978, GIESE & REUTTER 1978) and surface data (ABBATE et al. 1970, CARMIGNANI et al. 1978). *Black:* shelfal Mesozoics. *Shaded:* allochthonous sialic realms. *Stippled:* chaotic slump masses. M = Moho.

data show two overlapping Moho horizons underneath the Italian peninsula. The east-dipping Alpine belt contains the suture of the South-Penninic or Eoalpine collision above parautochthonous miogeoclinal Mesozoics, and overlain by a thick stack of ophiolites and Eoalpine high-pressure metamorphites, in turn overlain probably by the Eocene suture of the Ligurian–East Alpine continent/continent collision. Crustal data show that at a depth of 10 to 15 km, this belt is offset at a gently west-dipping thrust and transported 50–60 km eastward. The original continuation of the Alpine belt down dip is the duplication of the Moho. This situation clearly supports the interpretation of the Apennines as resulting from a flipped subduction (BOCCALETTI et al. 1971, GIESE & REUTTER 1978). Graben tectonics in the Tyrrhenian sea (BIJU-DUVAL & MONTADERT 1977) block the westerly thrusting in the Alpine belt of Corsica. As the grabening is coeval with easterly thrusting in the Apennine (REUTTER et al. 1978), the intersection pattern is crosscutting and not conjugate.

Crustal structure of the Apennine foreland is dominated by the median syncline with plastic–ruptural deformation in the southwest flank beneath the internal Apennine units. Simplifying, one may accept the synclinal core as the "free edge" of the foreland plate. For the northern profile, the point of zero deflection is located at the east nose of the Vicenza high, and for the southern profile, at the coast of Istria. This geometry furnishes a flexural parameter λ of 1.12×10^{-7} or 1.14×10^{-7}, and distances between peripheral bulge and "free edge" of 207 or 211 km. For the northern profile, the bulge center plots in an uncharacteristic, perhaps even allochthonous spot near the Alpine south front at Belluno. For the southern profile, the bulge center is located in the High Karst, a Dinaride thrust complex quiescent since late Oligocene time (CHOROWICZ 1977). In the orogeny model presented here, the postorogenic uplift of the Dinarides and their Istrian foreland is caused by the foreland loading of the Apennine. KLIGFIELD (1979) has presented a very similar model of the Apennine based on a comparison with the Himalaya.

The morphology of the island of Corsica and its elevation up to 3 km suggest recent uplift and perhaps eastward tilting. The culmination is too close to the subduction beneath the Apennine to constitute a normal peripheral bulge. More probable is a plastic–ruptural downwarp as is expressed in the Tyrrhenian graben tectonics (BIJU-DUVAL et al. 1977). The tilt could be caused by an east-dipping thrust which may possibly outcrop, although not proved, at the western foot of the continental shelf. Such a thrust could underlie the island at up to 30 km of depth, with movement conjugate to the transport of the Apennine.

Alpine–Tellian suture in overview

Figure 8, as a summary of this paper, is a series of crustal profiles across the Eoalpine suture between Eastern Alps and Calabria. This series of silhouettes shows the suture in its present state and the structural changes along strike, above all the change from continental to oceanic crust within one and the same plate.

The Eurasian plate is bent underneath the suture, gently and elastically in the upper part and probably more strongly and plastically in the unknown depth. Along strike, the elastic flexure goes through a maximum, accompanied by a narrow segment of foredeep, in the Western Alps. Crustal layering of seismic velocities in the foreland near this maximum may represent ruptural deformation. Decollements may reach out into the elastic realm preparing the next increment of progradation. However, the petrology of the low-velocity zones is problematic (RYBACH et al. 1977). In the third profile, the Eurasian plate appears subducted the farthest, and in the fourth, the elastic foreland flexure is obliterated by plate fragmentation. Of two continental fragments, the easterly (nameless), is subducted beneath the Adria plate and overthrust by the western segment, Corsica. This overthrust is the northern continuation of the Calabrian subduction. Farther north it is not recognized in the

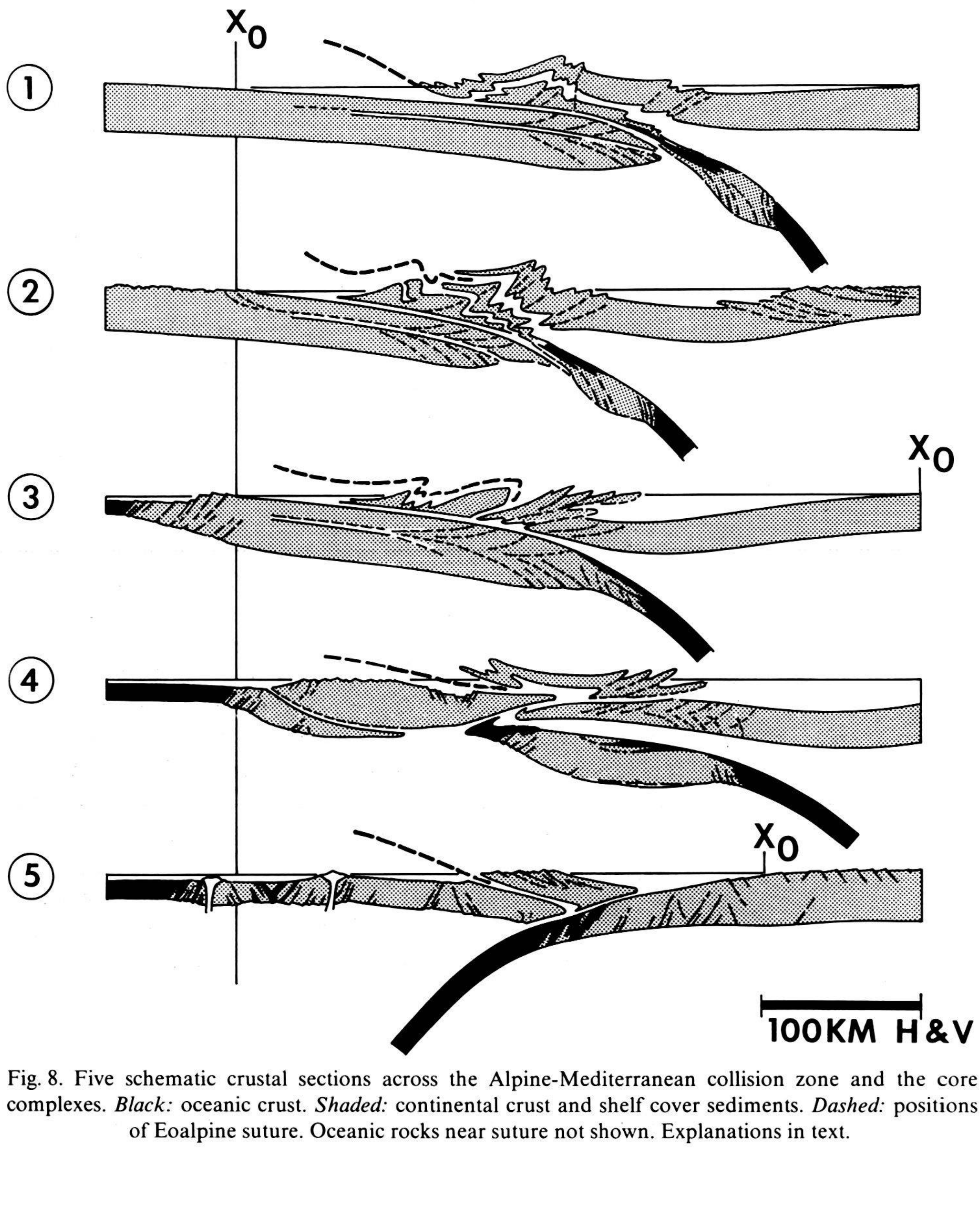

Fig. 8. Five schematic crustal sections across the Alpine-Mediterranean collision zone and the core complexes. *Black:* oceanic crust. *Shaded:* continental crust and shelf cover sediments. *Dashed:* positions of Eoalpine suture. Oceanic rocks near suture not shown. Explanations in text.

lower plate. East-dipping intracrustal overthrusting upsets the remains of elastic deformation in the western fragment. In the fifth profile, the Eurasian plate appears as backarc area in the upper plate of the Calabrian subduction, consisting of stretched and thinned, largely ocean-submerged, continental crust.

The Alpine core complex comprises, in the first two profiles, the crustal realm between foredeep and Eoalpine suture. It consists of detached foreland material. In the third profile the Alpine material is supplemented by material detached from the Adria plate, with the suture itself poorly located beneath the Padan plain. In the fourth and fifth profiles, the core complex consists exclusively of Adria plate material which toward the south becomes increasingly allochthonous. In the last profile it has become the upper plate of subduction.

The west margin of the Adria plate is affected by polyphase low-angle thrusting. Ophiolites of the Eoalpine suture zone are perhaps constituents of the Adria plate, but everywhere they show tectonic contacts with the sialic basement or "Altkristallin" in their back. In profile, the continental margin has a wedge shape possibly achieved during the Eoalpine collision.

A warped folded-fault structure in the Eastern Alps develops a swarm of south-vergent kinkfolds and backthrusts with westward increasing throw. In the Maritime Alps profile, these thrusts have increased to form the basal decollement of the Apennines with up to 100 km of transport, foreboding the Calabrian subduction.

In the southernmost profile, the Adriatic plate margin is located largely in oceanic crust showing a 350 km deep Wadati-Benioff zone (McKenzie 1972). The transition between the intracontinental North-Apennine thrust belt and the Calabrian subduction perhaps has been a strip of oceanic crust beneath the Lucanian trough (Görler & Giese 1978), at present completely subducted. The Calabrian mainland is the outer arc of the Recent subduction with an accretionary wedge beneath the Messina cone (Biju-Duval & Montadert 1977, Fig. 3, Profile 2), that is, part of the Recent Eurasian plate. However, it consists (Görler & Giese 1978, Fig. 3) of an east-dipping sialic block above the Eoalpine suture zone. This block is part of the core complex wedged between synclinally dipping thrust surfaces or subductions. It continues into North Africa (Alvarez 1976).

East of the plastically deformed zone, the Adria plate contains a wide syncline which may have formed elastically by the gravitational load of the Apennine, or by plastic–ruptural deformation of its west flank. This structure is of great interest for the deep structure of orogens, because it may mediate between subduction and thrust transport antithetic to subduction. A similarly synclinal architecture of the crust is suggested by the surface geology in Andean and Cordilleran thrust belts, such as in Montana, USA (Mudge 1970), and in Chubut, Argentina (Yacimientos Petroliferos Fiscales de Argentina, personal 1978). This aspect needs to be checked against the results by Barazangi & Isacks (1976); it recalls the tectogene as the birthplace of orogenic belts and converging lithosphere (Griggs 1939).

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