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The Casanova Complex of the Northern Apennines: a mélange formed on a distal passive continental margin

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Abstract—The Cretaceous–Palaeocene Casanova Complex occurs in two thrust sheets of the eugeosynclinal Ligurids of the Northern Apennines. It is a sedimentary mélange with ophiolitic and quartzose turbidites or limestone–shale olistostrome (submarine debris flows) as matrix. Exotic blocks of ophiolite and granite, serpentinite breccias and lenticular ophiolitic breccias and olistostromes contribute to the mélange character of the complex. Deformational structures include soft-sediment slump folds (indicating a SW-dipping palaeoslope) and boudins, a gradational slumped top to the mélange, small-scale faults in chert blocks and deformation associated with the emplacement of the exotic slide blocks. The blocks were shed as rotational slides from submarine fault scarps and are surrounded by haloes of debris created by submarine weathering. The stacking pattern of the blocks, with the originally stratigraphically highest ophiolite lithologies lowest in the pile of blocks, is explained by a diverticulation model with progressively deeper erosion. Mechanical analysis shows that the blocks were stable when partly exposed resting on a soft sediment substratum. Criteria which distinguish the Casanova Complex from a tectonic mélange, and which may be of value in other mélanges, are discussed. Previous interpretations of the complex as a precursor olistostrome to northeastward nappe emplacement (the Bracco ridge model) are rejected. The mélange is believed to have formed on ocean crust as a result of turbidite and debris flow sedimentation, soft sediment deformation, block faulting, gravity sliding and submarine erosion at the distal edge of a uniformly SW-dipping continental margin.

INTRODUCTION

A MÉLANGE is a rock-body of mappable dimensions, resulting from sedimentary and/or tectonic fragmentation and mixing of rocks. It consists of heterogeneous inclusions of various lithologies generally in a finer-grained, commonly pelitic, matrix. No constraints are placed on block sizes, shapes or lithologies, nor on block proportions. Although the processes of fragmentation and mixing which contribute to a mélange are more frequently deformational rather than erosional, they may be of tectonic or (soft) sedimentary origin.

Two common misconceptions have prevailed about mélanges until recently. The first, contrary to the original definition of 'mélange' by Greenly (1919), is that mélanges are necessarily tectonic in origin, that is they are produced by deformation of lithified rocks under a considerable overburden (e.g. Hsu 1968, 1974, Mercier & Vergely 1972). Such a concept is undesirable because the distinction between tectonic and sedimentary mélanges is often difficult to make (Naylor 1978a). It also adds an unnecessary genetic qualification to a useful descriptive field term. The second misconception is that all mélanges result from processes acting at subduction zones. This stems from the frequent use of the Franciscan mélanges as type examples; these are at least partly tectonically-created subduction mélanges (e.g. Blake & Jones 1974, Ernst 1970, Hsu 1971, Maxwell 1974).

More recently, it has been recognised that mélanges may have an appreciable sedimentary component (Naylor 1978a), including soft-sediment slumps, gravity-slid lithified masses and olistostromes (submarine debris flow deposits), see for example: Bachman (1978), Cowan & Page (1975), Gucwa (1975), Horne (1969), Kleist (1974), Naylor & Harle (1976), Page (1978), Smith *et al.* (1979), Swarbrick & Naylor (1980). A number of mélange-forming environments have now been recognised, including transform faults (Moseley & Abbotts 1979, Saleeby 1977, 1978), subduction zones (e.g. Moore & Karig 1976, Jones *et al.* 1978), continental collision zones (e.g. Hall 1976) and environments less easily associated with specific plate tectonic settings—oversteepened slopes and thrusts, for example.

Because of the difficulty in distinguishing tectonic and sedimentary mélanges, it is instructive to examine the characteristic features of a well-exposed sedimentary and soft-sediment deformational mélange. These characteristics can then be compared with other mélanges of more uncertain origin. A number of mechanical and structural geological approaches can be applied to sedimentary mélanges. Thus the aims of this paper are: (1) to describe a primarily sedimentary mélange from the Apennines, examining the sedimentary and deformational processes which led to its present state; (2) to mention some new techniques in the study of mélanges and (3) to demonstrate the origin of this mélange on a Mesozoic passive continental margin.

LOCATION AND REGIONAL SETTING

Location

The Casanova Complex occurs in inland Liguria (Provincia di Genova), Italy, being named after the village of Casanova (Fig. 1). The mélange is exposed for distances of up to 28 km and 8 km parallel to and across the NW–SE regional strike, respectively. The contacts of the mélange are broadly concordant with the surrounding strata, dipping to the SW at 20–40°. The aggregate stratigraphic thickness of the mélange is difficult to estimate, but is at least 1 km, possibly 2 km. The mapped area (Fig. 2) represents the best exposed 16 km length of the mélange.

Northern Apennine geology

The Northern Apennines are generally believed to represent a 'telescoped' passive continental margin sequence, in which Mesozoic and Cenozoic carbonate and clastic sediments were emplaced from the SW to the NE as a series of thrust sheets in Tertiary time (e.g. Reutter & Groscurth 1978). From the various emplacement-related structures—such as thrusts dipping SW and cutting up-section to the NE, folds with NE asymmetry and facing directions (e.g. Reutter & Groscurth 1978, Sestini 1974, Dallan Nardi & Nardi 1975)—it is possible to make a palinspastic reconstruction of the continental margin. In this way the upward progression in the pile of nappes, from proximal platform carbonates, through calciturbidites and pelagics, to pelagic sediments resting on ophiolites (Bernoulli & Jenkyns 1974), forms a series of facies with the structurally highest being the thinnest, most distal and deepest water and originally located furthest to the southwest. Hence the continental margin is believed to have dipped to the SW. This view is supported by palaeoslope measurements in the distal sediments of the postulated passive margin (Naylor 1978a, b).

Geology of Liguria and the Casanova Complex

The unit consisting of ophiolite and pelagic sediments, the Vara Complex, is interpreted as Mesozoic ocean crust. It is tectonically one of the uppermost 'eugeosynclinal' (i.e. ophiolite bearing) units of the Northern Apennines. The complete Vara Complex occurs as allochthonous sheets in southeast Liguria (Decandia & Elter 1972, Barrett & Spooner 1977). Here it consists of Jurassic igneous rocks (serpentinites, gabbros, basalts) overlain by Lower Cretaceous pelagic sediments (cherts and limestones) and Upper Cretaceous flysch-like shaly sediments. In inland Liguria, including the area of Casanova, the flysch-like sediments are commonly detached from their ophiolite plus pelagic substratum. This forms the Middle Tectonic Unit, shown in Figs. 2 and 6, which is also known as the Monte Ramaceto nappe (Schamel 1974) and loosely as the internal Ligurids. It contains the Palombini (alternating beds of grey micritic limestone and shale interpreted as the distal calciturbidites of an abandoned fan system) and the Lavagna 'shales' (a prograding quartzose turbidite sequence) (Naylor 1978b). Because of the close lithological similarities between the Vara Complex in its type area, and the Middle Tectonic Unit or Monte Ramaceto nappe, and because of there being some ophiolite and pelagic remnants at the base of the latter unit, there is no doubt that it is simply the allochthonous upper part of the Vara Complex. The environmental interpretation of the Vara Complex as oceanic crust, and the unstacked position of the Middle Tectonic Unit indicate that the Casanova Complex which it (partly) contains must be a unit associated with the distal part of the continental margin, and was probably situated on oceanic crust.

Description of the Casanova Complex is hampered by two factors: its diachronism, and its occurrence in two major units separated by a major thrust surface (Fig. 2). The older part of the complex occurs in the Middle Tectonic Unit, and is thus intercalated with the Upper

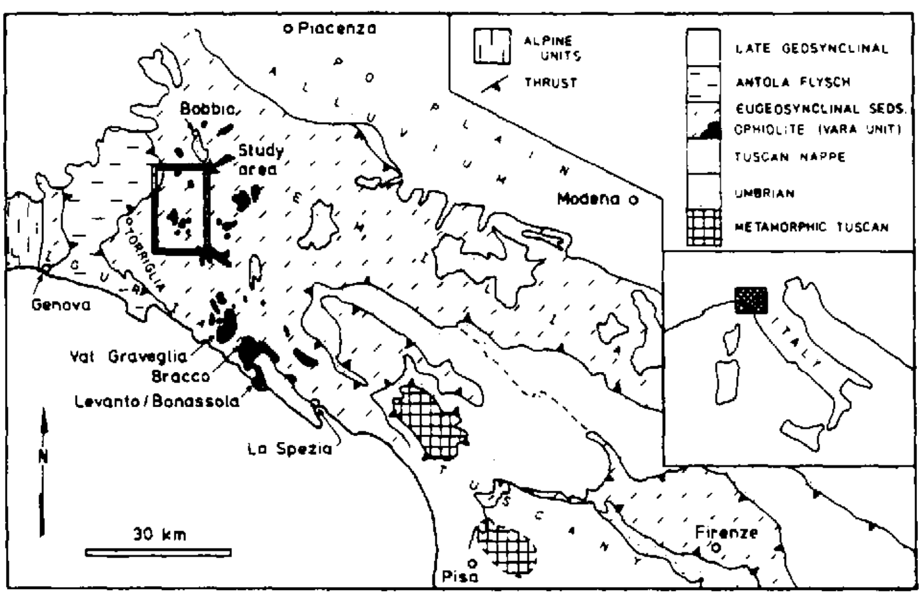

Fig. 1. Geological map of the Northern Apennines, showing the location of the Casanova area. The major tectonic units are indicated on the key in order of tectonic stacking.

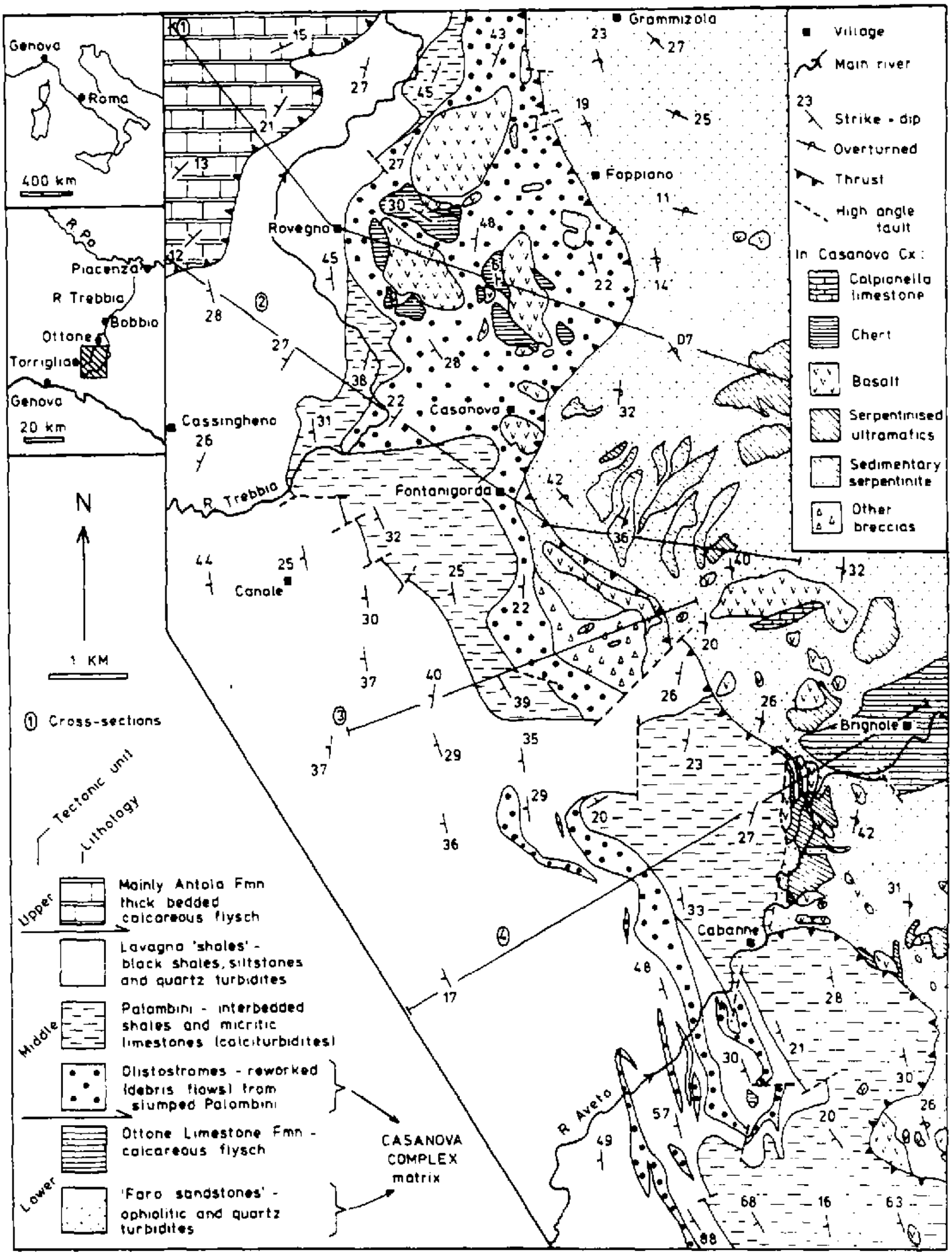

Fig. 2. Simplified geological map of the Casanova area. Representative strikes and dips are shown. Full topographic details can be obtained from topographic sheet 83 (scale 1 : 25000) of the Istituto Geografico Militare, Florence. Cross sections 1–4 are shown in Fig. 6.

Cretaceous calciturbidites and quartzose turbidites. The younger part of the Complex occurs in the wholly inverted Lower Tectonic Unit (the external Ligurids of some authors), and is of Late Cretaceous to Palaeocene age (Bertini & Zan 1974, Boni 1969). Earlier workers did not recognise the thrust dividing the Casanova Complex (e.g. Maxwell 1962, Passerini 1962), treating the whole complex as one (inverted) stratigraphically-continuous section. The term Casanova Complex (Passerini 1962) is nonetheless still a useful one. The complex represents the products of a specific set of sedimentary and deformational processes, as will be illustrated in this paper.

COMPONENTS OF THE MÉLANGE

Matrix lithologies

In the Lower Tectonic Unit (Fig. 2), the mélange matrix consists of parallel-bedded turbidites (Fig. 3a) dated as up to Palaeocene in age (Bertini & Zan 1974). The turbidites may be quartzo-feldspathic or ophiolitic with clasts of serpentinite, gabbro, basalt, glass, chert and pelagic limestone. The depositional environment has been determined by Naylor (1978b) as an outer submarine fan.

In the Middle Tectonic Unit, the mélange matrix consists of amalgamated beds of olistostromes composed of Palombini limestone clasts in a shale matrix (Fig. 3b). The olistostromes are conformably interbedded with ophiolitic turbidites at the base of the Middle Tectonic Unit. The olistostromes are interpreted as submarine debris flows reworked from slumped Palombini beds (Naylor in press), and contain slump-fold hinges as clasts and *in situ*. The olistostromes pass upwards into slumped and then non-slumped Palombini beds (Fig. 5a), indicating an upwards waning of the instability which caused slumping (Naylor in press).

Exotic blocks

Tabular to equidimensional blocks up to 140 m thick and 2 km wide or long occur throughout the mélange, and are broadly concordant with bedding. Each block is monolithologic and may be composed of: chert, basalt and serpentinised harzburgite or lherzolite. Pelagic Calpionella limestone and gabbro are rare (Fig. 4b). The lithologies and micro-faunas, basalt types, and degrees of sub sea-floor metamorphism are identical to those of the intact ophiolite of the Upper Jurassic Vara Complex in Southeast Liguria (cf. Decandia & Elter 1972, Spooner & Fyfe 1973, Barrett & Spooner 1977). Only one significant difference occurs in comparison with the Vara Complex. Many of the basalts in the mélange are rich in plagioclase, have a trachytic texture and contain abundant bladed ilmenite crystals and pink titanaugites. This suggests an alkaline to transitional chemistry, rather than the more usual tholeiitic composition of the Vara Complex.

Granite occurs as significant but volumetrically subordinate 20 m blocks. These are continental hornblende granites, much older than the Vara Complex, with radiometric ages of 309–220 Ma (Eberhardt *et al.* 1962).

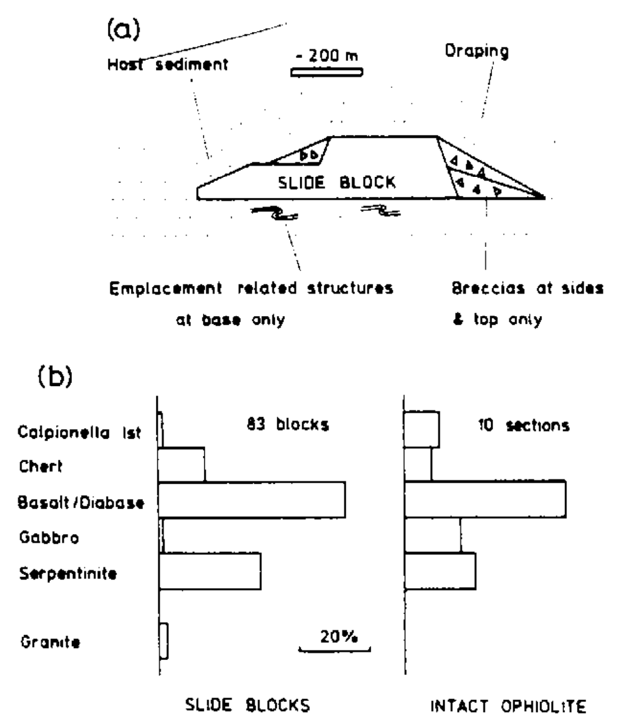

Fig. 4. (a) Diagram showing the relation between slide blocks, deformational structures and scree-like breccias. (b) Exotic lithologies present in slide blocks; each reading represents one block.

The exotic blocks are interpreted as gravity slid-masses emplaced onto unconsolidated sediment. They are thus slide-blocks, or 'olistothrymmata' and 'olistoplaka' in the terminology of Richter (1973). Origin by tectonic emplacement is precluded by the total absence of deformation at block margins, other than demonstrable soft-sediment features (see below).

Lenticular sedimentary bodies

Lenticular sedimentary bodies are of three types. At the bases of graded ophiolitic turbidite beds, ophiolitic cobble conglomerates or olistostromes may occur. Matrix-supported ophiolitic olistostromes and clast-supported angular scree-like breccias (Fig. 3c) occur as 'haloes' at the sides and tops of the exotic ophiolite blocks (Fig. 4a). They are interpreted as debris shed from the slide blocks after their emplacement, locally interfingering with the host sediments. Finally, lenses of serpentinite breccias (Fig. 3d) occur at various levels in the mélange (e.g. east of Fontanigorda, Fig. 2) and represent various types of sediment gravity flows.

DEFORMATION IN THE MÉLANGE

In this section it is demonstrated that most of the deformational structures in the mélange are of soft-sediment origin, and the heterogeneous distribution of these structures is explained.

Slumping: folding and boudinage

Intraformational slumping occurs at the top of the mélange (Fig. 5a), at the gradational contact between olistostrome and undeformed Palombini limestone-shales of the Middle Tectonic Unit. Identical slump folding occurs within the olistostrome–matrix mélange wherever thin competent sandstone beds occur. The slumping is described elsewhere (Naylor in press); pertinent points are summarised below.

Slump folds (Fig. 3e) are dominant in the volumetrically small, thin-bedded sandstones intercalated in the shales of the Palombini. They have wavelengths and amplitudes of a few cm to a few tens of cm. They are disharmonic, intrafolial folds affecting few layers; they are isoclinal, tight or close, with interlimb angles of 0–100°. Some slump folds are isolated hinges embedded in shale, whereas others are represented by trains of folds. In the latter, they have a constant SW asymmetry and vergence (Naylor 1978a, b) with NW–SE trending axes (Fig. 5b). The folds belong to classes 1A, 1B and 1C of Ramsay (1967). Folds of the same shape and orientation are developed, albeit less commonly, in the thicker limestone beds, where they have amplitudes of several metres. Whole internally folded translational slide sheets can also be mapped (Naylor in press).

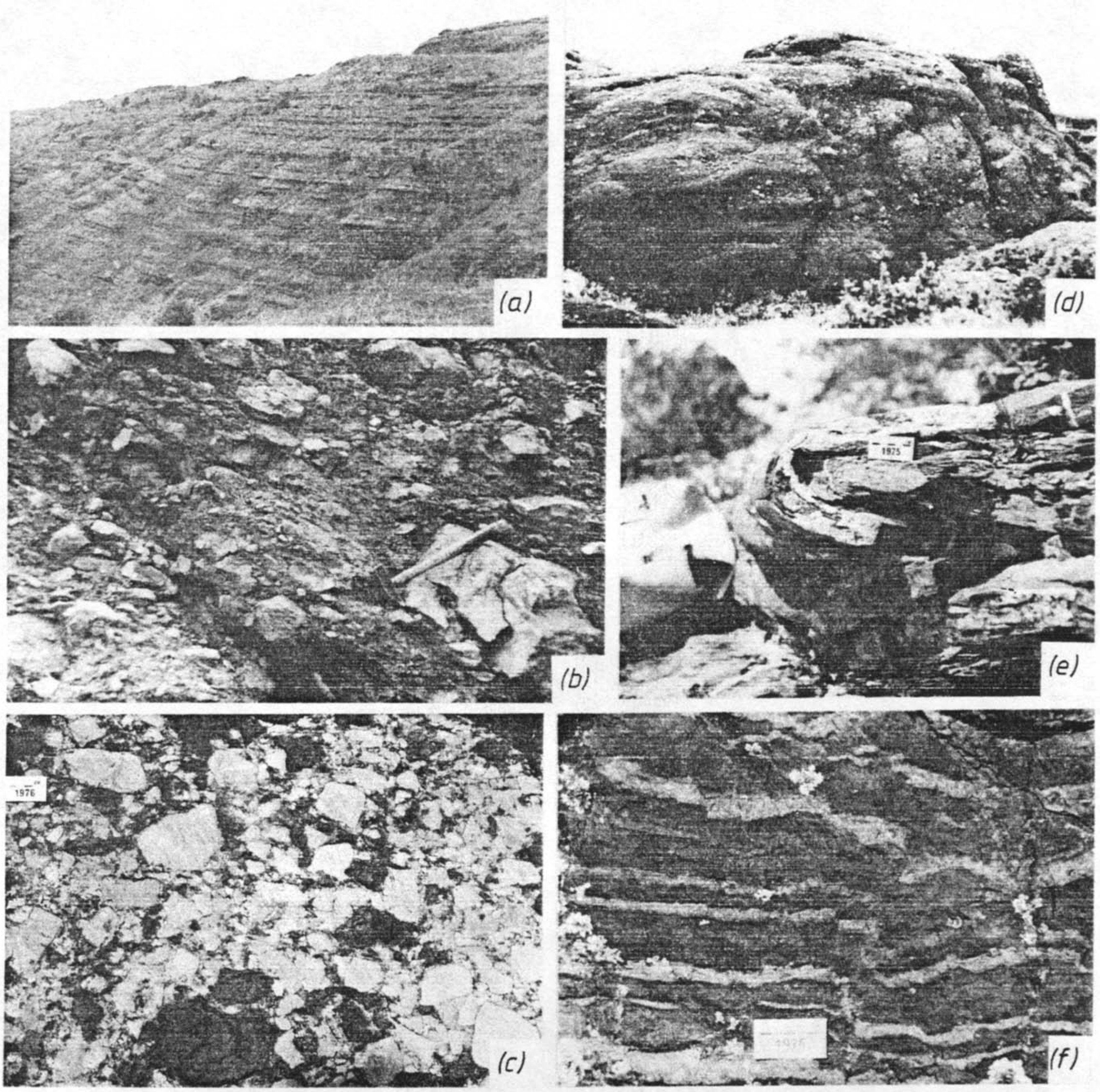

Fig. 3. Components of the Casanova mélange. (a) Faro ophiolitic and quartzose turbidites. (b) Matrix-supported limestone–shale olistostrome. (c) Scree-type clast-supported breccia of chert and pelagic limestone. (d) Serpentine breccia. (e) Typical recumbent asymmetric slump fold in sandstone bed of the Palombini. (f) Small soft-sediment faults in red–green banded chert.

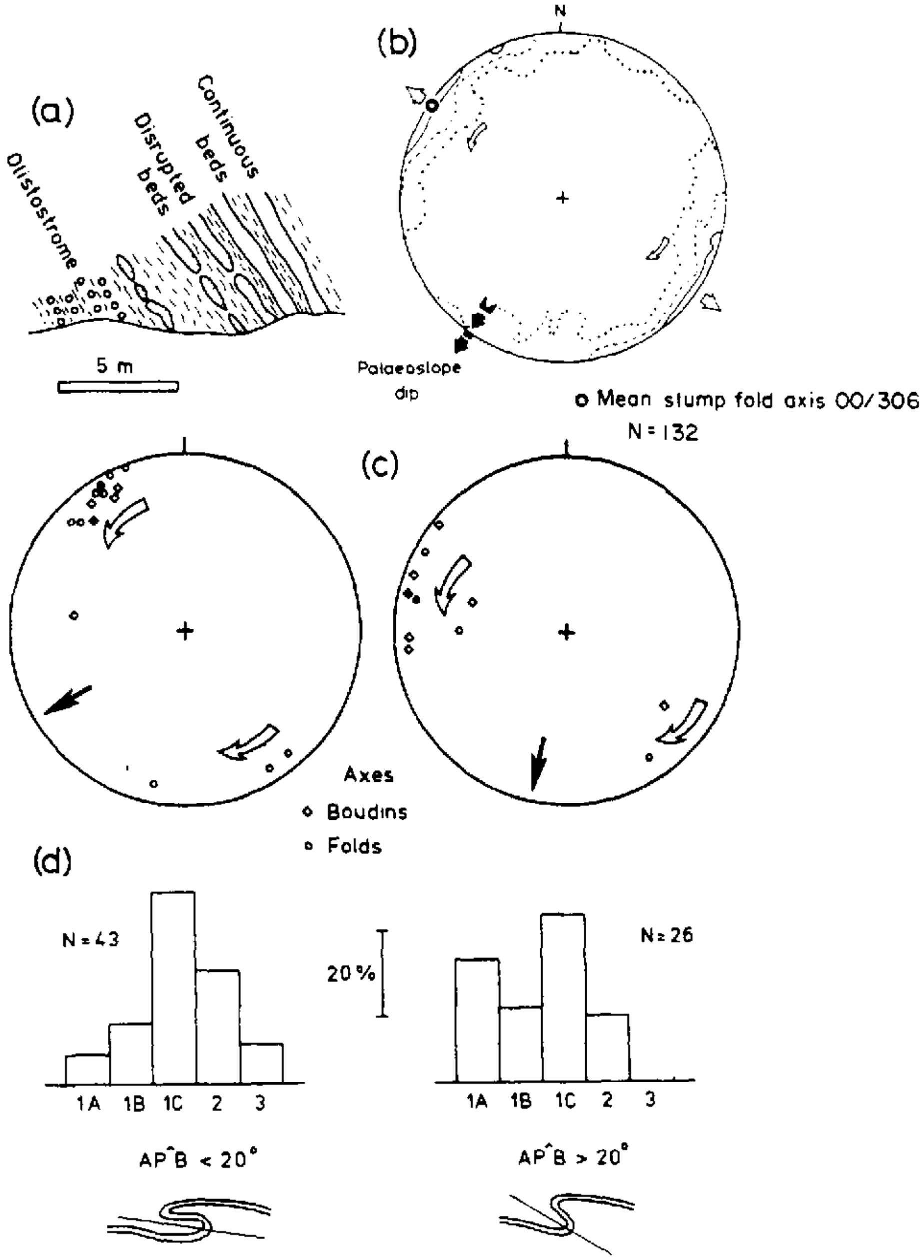

Fig. 5. Aspects of slump deformation in the mélange. (a) Olistostrome gradational through slumped and boudinaged Palombini to continuous and undeformed beds. (b) Equal-area plot of mean slump fold axis orientations at 132 localities, representing about 1000 measurements. Each mean is corrected individually for later tectonic tilting. Contours are at 1, 5 and 9% per 1% area of the net. (c) Equal-area plots showing the typical parallelism of slump-fold axes and boudin axes at individual localities. (d) Histograms showing dependence of fold-layer shape on axial plane orientation, as indicated by the angle between an axial plane and bedding (AP^B). Solid symbols, means; open arrows, fold asymmetry as viewed down fold axis plunge; solid arrows, slump movement directions calculated by the mean axis method of Woodcock (1979).

Various features identify the folds as of soft-sediment origin.

(1) Dewatering, flame and liquefaction structures in the cores of folds.
(2) Load-casting at the stratigraphic tops of folds, post-dating the folding.
(3) Sediment ponding and draping by the folds.
(4) The lack of a geometrically related mineral veining (in contrast to later, tectonic folds).
(5) The collapsed style of the folds (Fig. 3e): rounded upper limbs overlie flattened, angular lower limbs in the same asymmetric fold, implying that the sediments were weak, collapsing under the small additional load created by small-scale asymmetric folding and minor localised thickening of the sediment pile.

Slump folds in shales are commonly associated with a fine crenulation lineation, a micro-folding of the primary fissility. Its similar orientation to the slump folds at each locality suggests that it results from that same deformation episode. It may have been enhanced by mimetic recrystallisation during later regional deformation (cf. Maltman 1977).

A correlation of slump-fold layer-shape with axial plane orientation was observed (Fig. 5d). Upright folds are generally class 1A/1B; recumbent folds are commonly class 1C. The following model is suggested. Class 1B folds were developed by buckling, asymmetry resulting from simple shear due to downslope translation of the slumps (cf. Woodcock 1976). Folds then suffered compaction, which for upright and recumbent folds was respectively parallel and perpendicular to their axial planes. In this way 1B folds evolved into 1A and 1C folds, respectively (cf. Ramsay 1967). Thus the observed layer shapes support the idea that the folds are pre-compactional, and thus soft-sediment folds.

Slump boudins are also common, but restricted to limestone beds. Boudins have variable cross-sectional shapes (rectangular, barrel and tapering shapes) reflecting variable competency contrast, due to different degrees of lithification at the time of deformation (Naylor in press). Boudinage leads to a reduction in limestone bed thickness, and ultimately to the total fragmentation of the beds. Chaotic rubbly zones up to 40 m thick developed. Boudin axes, though strongly dispersed, are parallel to the slump axes (Fig. 5c).

Using the methods reviewed by Woodcock (1979), the slump fold axial orientations and asymmetry indicate a SW-dipping depositional palaeoslope. Parallelism of slump fold and boudin axes is also consistent with gravity sliding (Page 1963). The slumping generated a gradational deformational contact at the top of the mélange. Identification of the folds as soft-sediment structures with a southwestwards movement direction, distinguish this contact from a hypothetical tectonic contact due to thrusting from SW to NE during Apennine orogenesis.

Micro-faults in cherts

The chert slide blocks are commonly intensely faulted on a small scale (Fig. 3f). Irregular normal and reverse faults and breccia zones with displacements of a few mm or cm occur. These structures have subsequently been lithified in the same way as the cherts themselves, and no longer represent zones of weakness.

Lithification of the faults is evidence of their soft-sediment origin. Similar features are absent in the intact Vara Complex ophiolite. The faults in the cherts are most probably related to the deformation (uplift and faulting) which 'cut loose' the slide blocks from their source area.

Slide block attitudes

Because the ophiolite slide blocks generally form positive features (Fig. 6), their long axis orientations (Fig. 7a) can be reliably estimated by geological mapping. Although there is a weak N–S orientation, a Rayleigh test on the magnitude of the resultant vector of the distribution indicates that there is no significant preferred orientation. Strongly dispersed long-axis orientations are indicative of surficial gravity sliding rather than the more constrained simple shear deformation associated with thrusting (Dimitrijevic & Dimitrijevic 1974). The latter would be expected to give a strong alignment of axes (e.g. Escher & Watterson 1974). Furthermore, there is no evidence for deformation at the block margins (e.g. shearing, cataclasis, mixing) as would be expected if the blocks had been tectonically emplaced or affected by tectonic rotation.

Bedding can be determined in the sedimentary slide blocks, and occasionally in those of basalt. The bedding shows a clear northeastward imbrication with respect to the SW-dipping bedding of the host mélange (Figs. 6 and 7b). The bases of the blocks are interpreted as listric slide surfaces along which they were freed from the source area, and thus the imbrication reflects truncation of the original bedding. The sketches in Fig. 7(c) indicate that rotational slips from the edge of a fault-bounded scarp could give the observed imbrication. It should also be noted that the observed imbrication sense is not consistent with a tectonic origin. Under simple shear, for example in a thrust zone, an imbricated structure dipping in the opposite direction to the sense of tectonic transport is produced (e.g. Escher & Watterson 1974). In the Apennines, the sense of tectonic transport is from SW to NE (Reutter & Groscurth 1978), requiring south westward imbrication, a direction which is not observed.

Structures related to emplacement of the slide blocks

Structures spatially associated with the ophiolite slide blocks are believed to be related to their emplacement. These structures include unambiguous features of soft-sediment deformation—small translational slump sheets, dewatering and liquefaction structures, local minor unconformities, truncations of bedding and slump folds. Other less diagnostic structures include; overturned folds (0.5–1 m in size), small faults and most commonly crumpling and contortion of bedding, particularly of less competent horizons. This latter feature is a good indication of the proximity of a slide block. Apart from small faults related to differential compaction over the rigid slide blocks, most of these structures occur at the stratigraphic bases of the blocks (Fig. 4a). These features are generally absent in the areas of the Faro sandstone over which the slide blocks must have travelled. They are therefore believed to be related to the settling of the blocks shortly after emplacement.

The occurrence of deformational structures only at the bases of blocks demonstrates their origin by gravity sliding, rather than as tectonic inclusions. The distribution of ophiolitic debris and breccias only at the sides and tops of the blocks supports the gravity sliding model. It also implies that the blocks suffered submarine erosion and weathering prior to burial by the host sediments (Fig. 4a).

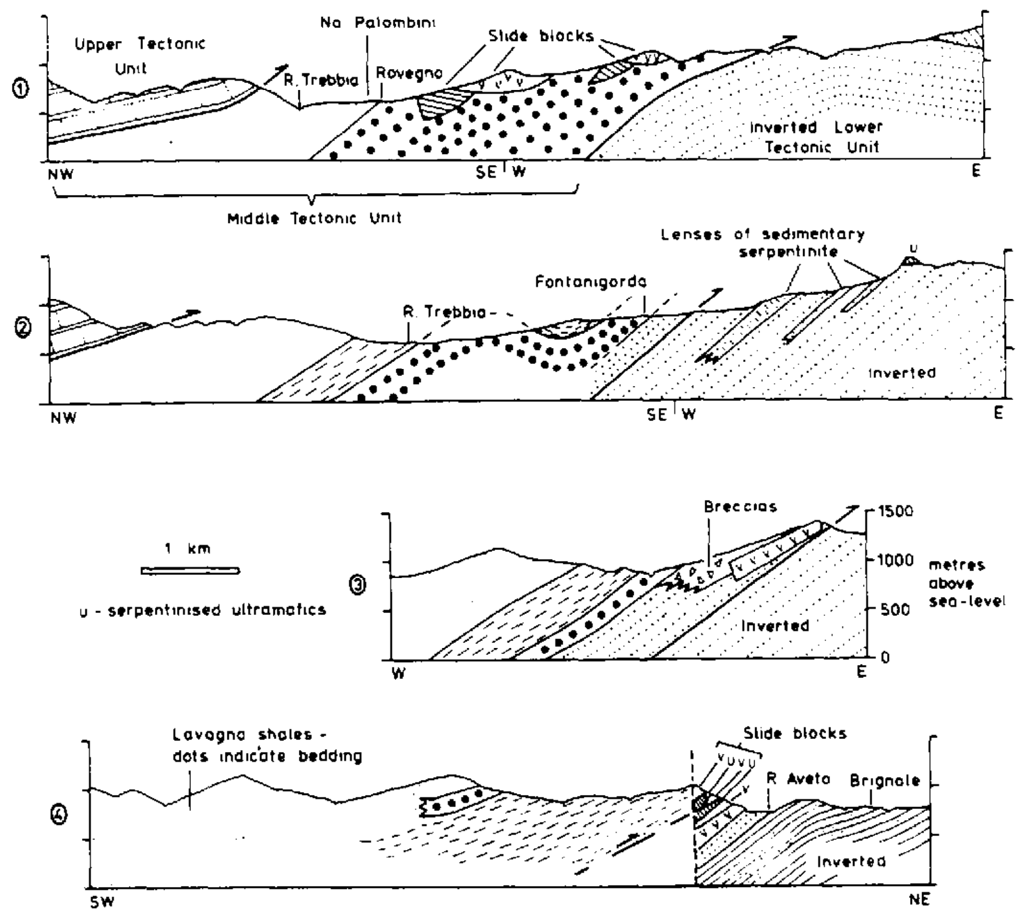

Fig. 6. Cross sections indicating the structure of the Casanova area and the stacking of ophiolite slide blocks. See Fig. 2 for key and locations of section lines.

Post-emplacement deformation of slide blocks

Post-emplacement break-up of blocks is occasionally seen. It probably relates to foundering of jointed masses of rock into the host sediments. In the basalt block near Fontanigorda (Fig. 2), four segments can be recognised on the basis of joint orientation. The mean joint dip in each segment decreases eastwards from one segment to the next, due to rotational foundering (Fig. 8). The joints in each segment have the same degree of clustering, suggesting that all joints were part of one original population. Elsewhere, fissures in slide blocks are filled with locally derived debris, again indicating small-scale break-up of the blocks.

Lenticularity and the discontinuous nature of units

An outline of the sedimentary character of many lenticular units (serpentinite breccias, polymict ophiolitic breccias and olistostromes, disrupted slumped units) has already been given. In the well-exposed Casanova Complex, there is no evidence for a tectonic contribution to the discontinuous units. In less well-exposed terrains, one might be tempted to interpret discontinuous bodies, particularly rocks superficially resembling crush breccias, as due to disruption by faulting. Others have also suggested that mélanges may have an appreciable component of sedimentary mixing (e.g. the Franciscan mélanges, Bachman 1978, Cowan & Page 1975, Beutner 1975, Kleist 1974 and other examples, Page 1978, Horne 1969).

Later tectonic deformation in the mélange

A later tectonic overprint is generally absent in the mélange. Local exceptions include:

(1) linear pebble fabrics in the olistostrome in regions of intense tight folding;
(2) planar pebble fabrics markedly oblique, or perpendicular to, bedding, associated with a corresponding inclined or vertical cleavage;
(3) strong disruption of the Palombini, attributed to superposition of tectonic deformation onto already weakened, slumped Palombini;
(4) rare folds and slickensides at chert–chert block contacts and
(5) local slickensides and imbricated cleavage and pebble fabrics in olistostromes near thrust planes.

The orientations and styles of these structures (Naylor 1978b) unambiguously identify them as the result of later regional deformation (1–3 above) or subsequent thrust emplacement (4 & 5 above).

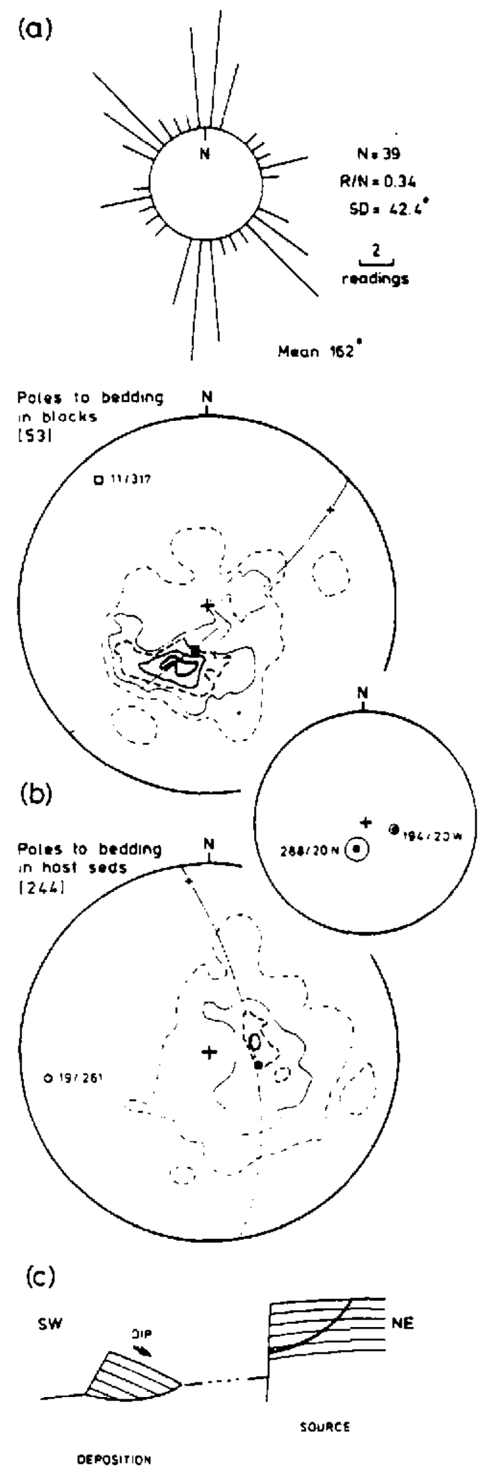

Fig. 7. (a) Long-axis orientation of slide blocks, with near random scatter (R/N is just non-random at 95% confidence level but not significantly different from random at 99% confidence level). (b) Equal-area plots comparing bedding plane attitude in chert slide blocks and in host sediment. Smaller plot shows means and 95% cones of confidence for the two datasets. By restoring the host sediment bedding to the horizontal, a mean initial attitude of slide block bedding of 326/26 NE is indicated, up the depositional palaeoslope. (c) Sketch showing development of imbrication of bedding in slide blocks freed by rotational slips at a fault scarp.

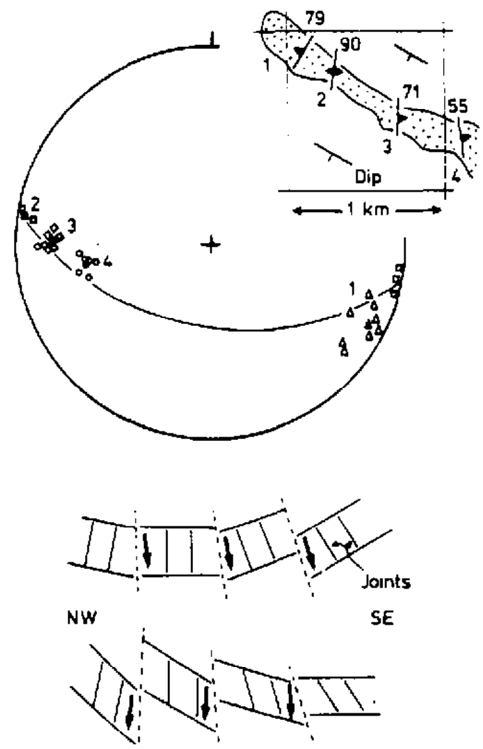

Fig. 8. Break-up into four segments of a jointed basalt mass near Fontanigorda. Each segment has a similar spread of joint orientations. Two interpretations of the data are shown, depending on whether the joints were initially perpendicular or oblique to the base of the block.

STACKING SEQUENCES OF MÉLANGE BLOCKS

Method

Transition analysis was used to quantify the upward stacking pattern of different slide blocks in the mélange. The application of transition analysis to structural sequences was discussed by Naylor & Woodcock (1977). In brief, the numbers of vertical transitions between different lithologies were counted in a number of cross-strike traverses arranged such that each slide block was crossed. The data are presented in matrix form (Fig. 9) from which the probabilities of various transitions are calculated, and typical sequences deduced. The two tectonic units of mélange, having olistostromes and sandstones as matrix, are analysed separately since they have clear differences in types and proportions of block lithologies. Chi-squared tests show that all the transition patterns discussed are significant at the 5% level, that is they are very significantly non-random sequences.

Olistostrome matrix mélange

From the transition matrix (Fig. 9a), common upwards sequences are OCDO and ODO (notation in Fig. 9); the composite sequence (defined as containing all the lithologies) is OCDSO, and has a lower probability (Fig. 9b).

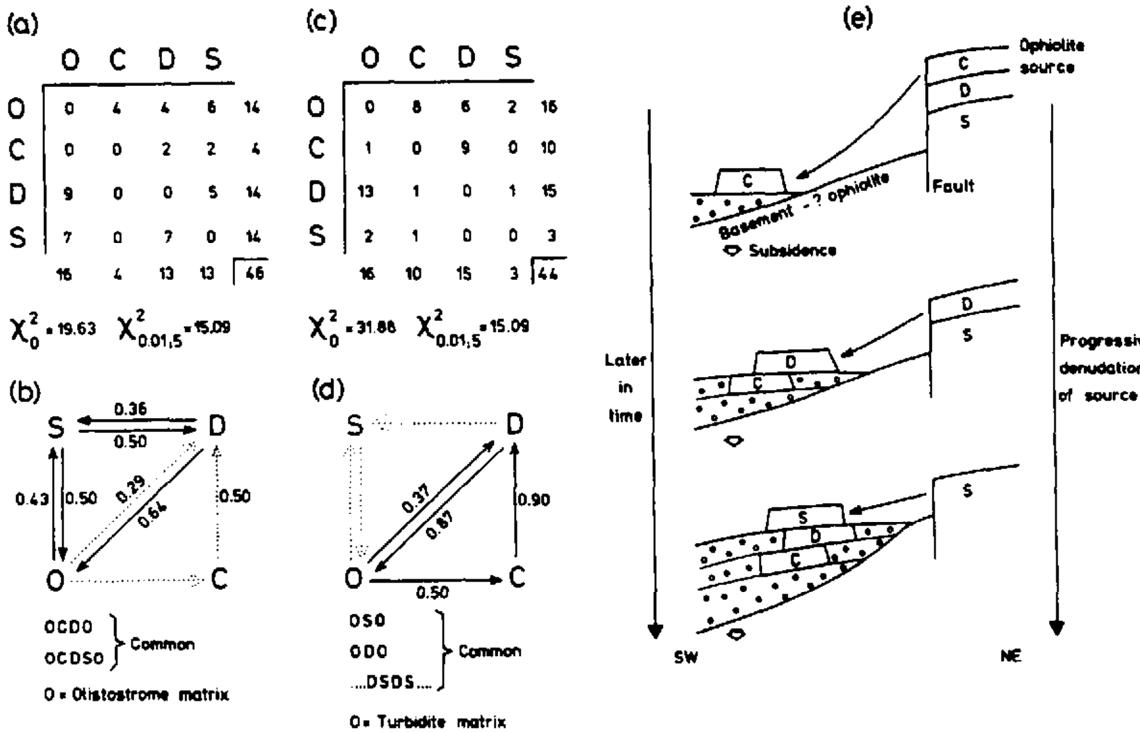

Fig. 9. (a) Upwards transition frequency matrix for olistostrome matrix mélange. (b) Diagram showing common transitions in olistostrome matrix mélange and their probabilites (equals number of transitions divided by appropriate row total of matrix). (c) Upwards transition frequency matrix for the sandstone matrix mélange. (d) Diagram showing probabilities of common transitions in sandstone matrix mélange. (e) Model of 'diverticulation', explaining inverse stacking of ophiolite lithologies in the mélange. Notation: O, matrix; C, chert; D, diabase and basalt; S, serpentinite.

Ophiolite within olistostrome would appear as OSDCO, SDC being a reasonable representation of the parent ophiolite where gabbro is absent (which is commonly the case). The observed sequences, OCDSO and OCDO (e.g. east of Rovegno, Figs. 2 and 6), contain the elements of an ophiolite but in their reverse order. The cherts, at least, can be shown from sedimentary structures to be the right way up; thus the inverse stacking sequence of the blocks cannot be due to *en masse* overturning of the ophiolite (cf. Passerini 1965) followed by its tectonic disruption within the olistostrome matrix. A diverticulation model (cf. Lemoine 1973) explains the observed sequences: blocks slide off a gradually uplifted ophiolite, with erosion reaching progressively deeper levels (Fig. 9e). Chert (C) would be the first to be stripped and accumulate, followed by basalt and diabase (D), and then if erosion reached deep enough, serpentinite (S).

Sandstone matrix mélange

Transitions involving C (chert or pelagic Calpionella limestone) are rare; isolated blocks of S or D (OSO, ODO) and alternations ...DSDS... are common (Figs. 9c & d). A composite sequence OSDO can be constructed, but is physically meaningless, being a statistical compounding of OSO and ...DSDS... alternations. The only meaningful composite sequence is OCDSO, implying that the inverse stacking model of Fig. 9(e) is still valid. Rarity of C probably implies that this lithology was absent in the source area of the sandstone matrix mélange. This absence has also been noted in parts of the the intact Vara Complex ophiolite (Barrett & Spooner 1977).

STABILITY OF SLIDE BLOCKS: A MECHANICAL APPROACH

Introduction

The purpose of this section is to apply some soil-mechanical approaches to the problem of a slide-block resting on a sediment substrate, which is mechanically identical to a foundation resting in or on a soil. It will be shown that the Casanova Complex slide blocks are mechanically at equilibrium either on, or only partly submerged by soft sediments. Sources of data for the calculations are given in Table 1 and more fully by Naylor (1978b).

There are two distinct methods used in solving soil mechanics stability problems. The first, the limit-equilibrium method, uses simple statics and finds the load at failure, assuming a given stress distribution and slip surface in the plastic soil. The second, the limit-analysis method (Chen 1975), gives upper and lower bounds to the collapse load. The upper-bound solutions are within a few per cent of the equilibrium solution, and are mathematically simpler, not requiring a step-by-step analysis of stresses and slip surfaces. Instead, the solutions

Table 1. Summary of physical properties used in modelling slide block stability

Parameter & units	Mean or value used	Range of values	Source
Block height, H m	54	10–130	Figs.
Block width, W m	211	30–1000	2,6 & F
Densities, ρ g cm^{-3}			
basalt block, ρ_b	2.77	2.22–3.14	L, 3
chert block, ρ_b	2.65	2.59–2.81	L
sea-water, ρ_w	1.93		3
sediment, ρ_s	1.55		3,5,6,9
Strengths, K g cm^{-2}			
surface sediments	35	0–320	1,2,5
at depth z m	$16 + 35z$	(15 to 71) + (4 to 35)z	6,7,8, 9,10,11
debris flows & kaolin slurries		0.30–3.60	4

(F & L, field and laboratory measurements (respectively) by author; 1, Almagor 1967; 2, Bryant *et al.* 1967; 3, Clark 1966; 4, Johnson 1970; 5, Kolb & Kaufman 1967; 6, Lee 1973; 7, Maltman 1977; 8, McClelland 1967; 9, Moore 1961; 10, Richards & Hamilton 1967; 11, Schofield & Wroth 1968).

are derived by equating external work done and internal energy dissipation. They have the advantage that published graphical solutions exist for a wide range of conditions (Chen 1975).

The limit–equilibrium solution

The relation between the maximum height (H) of a slide block stable on 'average' sediments and its width (W) is given by (Appendix):

$$\text{for no sink-in } (h = 0) \quad H = 1.03 + 0.15W \qquad (8a)$$

$$\text{for } 100\% \text{ sink-in } (H = h) \quad W = -24.38 + 4.25H - 0.08H^2. \qquad (8b)$$

These conditions are expressed as curves I and II in Fig. 10(a). The measured H and W values for the slide blocks are also plotted. Any point plotting below a given curve is stable for those conditions. A slightly simpler solution can be obtained by assuming that $\rho_s = 1.55$ g cm^{-3} (Table 1). The no sink-in curve is identical to the previous one, but the 100% sink-in curve becomes a straight line (III in Fig. 10a). It is clear (Fig. 10a) that few (8%) blocks could have been supported on the sediment with no sink-in, those below curve I. Some (11–37%) would sink in completely, those points above curves II or III. However, the majority of the blocks (89–63%) lie below curve III or II; such blocks would be in equilibrium with only partial sinking into the sediment. It was concluded earlier that many blocks were exposed at the sediment–water interface, because they shed haloes of scree-like breccias. Thus the mechanical model and geological observations are in agreement.

Graphical upper-bound solutions

Chen (1975) presents many solutions to soil mechanics problems analagous to the one discussed here. Although they require ρ_s to be constant, they are in general more realistic, involving additional variables such as internal friction of the sediment (ϕ) and base friction. H and W are first transformed into the normalised variables defined by Chen (1975):

$$q_0/K = \frac{\text{Unit load at failure}}{\text{Strength of sediment}} = \frac{(\rho_b - \rho_w)gH}{K}$$

and

$$G = \rho_s W/2K.$$

Again K is calculated at a depth $W/3$ below the base of the block. The positions of q_0/K, G pairs of values are then compared with Chen's stability curves.

The case of a surface footing (no sink-in of block) is illustrated in Fig. 10(b). The mean slide block is stable without sink-in on a sediment with an internal friction angle (ϕ) as low as 5°. Even the thickest block (largest q_0/K) is stable with $\phi = 15°$. Values for sediments often attain these values or higher. Note that a rough footing, that is with basal friction, reduces the value of ϕ needed for block stability. For the case of a shallow footing (a block with some sink-in), all blocks are stable with very low ϕ and base friction (Naylor 1978b), before the condition of 100% sink-in is reached. Very low coefficients of friction are needed (0.1) for the rough-base case. The results of limit analysis are therefore in accord with the numerical analysis presented earlier.

Discussion of the assumptions

It was assumed that the model slide block approximated to infinitely long prisms. For non-infinite bodies, the maximum load before failure may be up to 20% greater than in the model case (Chen 1975, Chapter 7); the load is spread over a larger soil volume than in the infinitely long case. Thus if a model lies just on the stability limit, a non-infinite one of the same cross-section will certainly be stable. This enhanced stability might be weakly opposed by irregularities in the bases of the blocks: for example, V-shaped wedges are less stable than blocks with flat bases. Mapping in the study area suggests that the bases of the blocks are essentially planar. It is assumed that the blocks have suffered no post-depositional rotation with respect to the horizontal, that is the measured H and W values represent the original ones. There is no evidence to the contrary.

The analyses required that the blocks rested on a horizontal surface. Because they slid into place under gravity, this may not be so. Slopes in an abyssal plain/lower continental rise environment are however only a few degrees (Naylor in press), negligible from the view point of the calculations.

The sediments were assumed to be isotropic. In fact, in normally consolidated sediments, K vertical/K horizontal lies in the range 1–2. Such anisotropy lowers the

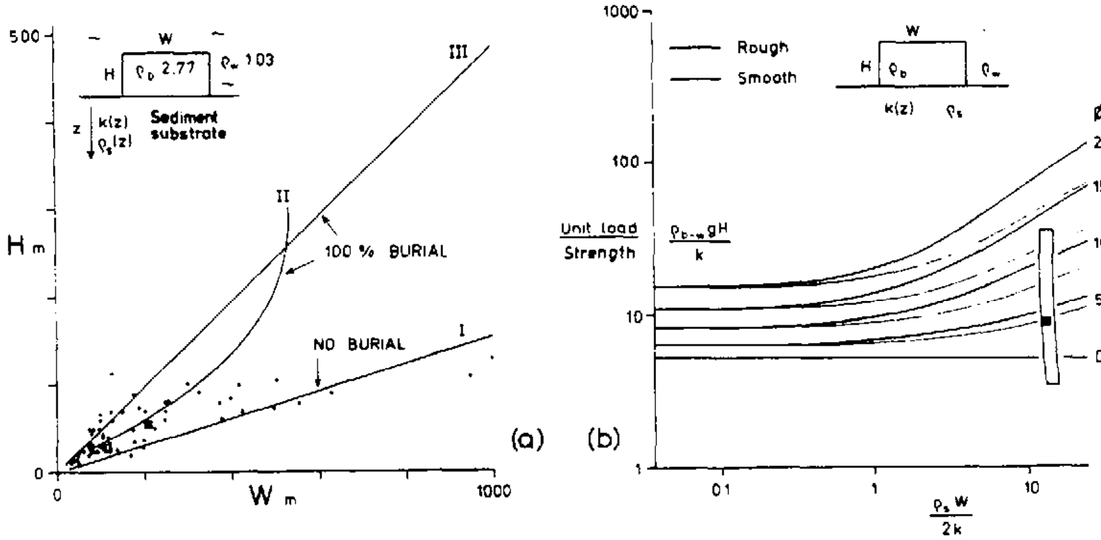

Fig. 10. (a) Graph of H vs W for slide blocks with stability curves: I, no sink-in, II, 100% sink-in, $\rho_s = \rho(z)$, III, 100% sink-in, ρ_s = constant at 1.55 g cm^{-3}. Main variables defined on inset; also see Table 1. (b) Graphical upper-bound solution for a surface footing problem (no sink-in). Range (bar) and mean (square) for slide blocks are plotted. All points below a given curve are stable for the appropriate conditions (ϕ and base roughness).

maximum failure load by about 10% (Chen 1975, p. 293). Sediment strengths used are those for clays, values which are comparable to those for debris flows (Johnson 1970, Naylor 1978b, in press). Thus for blocks in the olistostrome mélange, the strengths are valid. Those blocks in the Faro sandstone mélange rest on sandstone–shale alternations, which, because the beds are much thinner than the block dimensions may be regarded as statistically homogeneous. The effective strength will be some average of the strengths of clay and stronger sands. Again, therefore, critically-stable model blocks should certainly be stable in this real case.

The constant density models assume $\rho_s = 1.55$ g cm^{-3}, which may be too low for the olistostromes (Naylor in press). A more realistic value (1.9 g cm^{-3}) would enhance the stability of the blocks, that is some 'unstable' blocks would not sink in. The calculations were performed for diabase blocks (density 2.77 g cm^{-3}). Other lithologies are less dense and therefore would be more stable.

To summarise: the models use the lowest likely sediment strengths and densities and the highest block densities. If model blocks of given dimensions are stable under these extreme conditions, real blocks should also be stable.

No account has been taken of the pre-failure consolidation of the underlying sediment. No solution exists for the difficult problem of compaction of sediment by a load moving tangentially to its surface. As a block is emplaced by sliding, it sinks into the weak, wet surface sediment. Such sinking would by frictional resistance oppose the horizontal motion of the block. Thus when the blocks came to rest, they would already be in the partly sunk-in, or shallow footing, condition.

Thus the majority of blocks in the Casanova Complex were stable as sedimentary slide-blocks exposed at the sediment–water interface. They were buried not by sinking into the sediment, but by a combination of sediment compaction and burial by subsequent sedimentation. This conclusion is supported by geological observations (see above), verifying the validity of the method. The technique may be of value in mélanges of unknown origin, indicating whether exotic blocks could have been stable as supra-sediment bodies, or whether because of their shape (H/W ratio) they must be tectonic inclusions, supportable only by the strength of lithified sediments.

CRITERIA FOR A SEDIMENTARY MÉLANGE

In the Casanova Complex, there is a complex and chaotic assemblage of blocks, lenticular breccias, polylithologic olistostromes, turbidites and limestone–shale olistostromes. Knowing tectonic mixing to be absent, what criteria emerge for identifying sedimentary mélanges which could be used in mélanges of less clear-cut origins, including those which have been subsequently tectonised? Specifically, the following artificial question can be posed and refuted: the olistostrome–matrix mélange contains exotic bodies, has a thrust at its base and a gradational defomational top contact—could it be a tectonic mélange, that is a giant crush breccia?

Unambiguous criteria

(1) Contact relations: a thrust contact does not prove the tectonic origin of a mélange. Conformable sedimentary contacts of the olistostrome indicate its sedimentary origin, as at the stratigraphic top of the Lower Tectonic Unit and at the base of the Middle Tectonic Unit of mélange (Fig. 2). Only the gradational contact at the top of the mélange in the Middle Tectonic Unit (Fig. 5a) is ambiguous, and then only if its soft-sediment origin is unknown. A

similar slumped contact between olistostrome mélange and intact sediments has been described by Page (1978) from the Lichi mélange of Taiwan.

(2) Two mélange facies: blocks in the undeformed turbidite-matrix mélange must be sedimentary; this casts doubt on the olistostrome–matrix mélange being tectonic.

(3) Sedimentary features of the olistostrome: the olistostrome might resemble a pervasively sheared crush breccia. Debris flow features (channels and rigid plugs, Naylor in press), interbeds of thin undeformed shales and turbidites can only be sedimentary features. Intense tectonism would destroy such thin interbeds.

(4) Sedimentary serpentinites, with no tectonic contacts, are present. Criteria for their recognition were presented by Lockwood (1971).

(5) Pebble fabrics: the only observed disruption of the Palombini limestone is boudinage resulting in fragments with long axes strongly parallel to bedding (Fig. 11a). Olistostromes show a much greater range of clast long-axis orientations (Fig. 11b), and thus cannot have originated solely by this process. Resedimentation must have occurred.

(6) Slide block features: bipartite distribution of scree-like edge breccias and deformation around blocks can only result from a sedimentary origin (Fig. 4a).

(7) No matrix deformation: the mélange matrix shows no pervasive shearing. The ophiolitic sandstones, for example, are made up of detritus showing no shearing or cataclasis.

(8) Mesoscopic deformation features (folds and faults) are largely absent.

(9) No evidence for simple shear: a tectonic mélange at the base of a thrust sheet would have suffered large amounts of simple shear. This should produce a cleavage and pebble fabric imbricated with respect to the edges of the shear zone (Ramsay & Graham 1970, Escher & Watterson 1974). Such features were not observed.

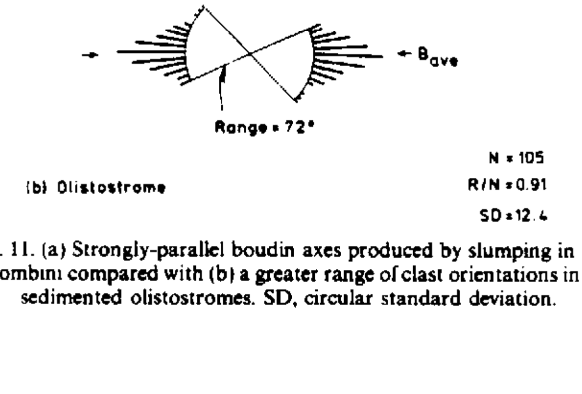

Fig. 11. (a) Strongly-parallel boudin axes produced by slumping in the Palombini compared with (b) a greater range of clast orientations in re-sedimented olistostromes. SD, circular standard deviation.

(10) Soft-sediment deformation and dewatering features generally only indicate deformation before lithification. Here the sedimentary origin can be proved because slumping and re-sedimentation are closely linked; small slumps with eroded tops associated with slide blocks also indicate a sedimentary mélange.

Supporting criteria

The following points are consistent with the sedimentary mélange hypothesis.

(1) Transition analysis: the stacking sequence of blocks is consistent with a gravity sliding model.

(2) Stability analysis confirms that blocks were stable when resting on wet sediment wholly or partly above the sediment–water interface.

(3) Source: ophiolite blocks ('exotics') have no exposed source either side of the thrust plane; a sedimentary mechanism of incorporating them into the mélange must be invoked (cf. Bruckner 1975, Naylor & Harle 1976). In view of the uncertain nature of the contacts with the thrust plane in the subsurface, such arguments must be used with caution.

ENVIRONMENTAL MODEL

Source directions

The olistostrome–matrix mélange is, on the evidence of slump folds, derived from the northeast down a uniformly SW-dipping palaeoslope. The Faro ophiolitic and quartzose sandstones are also derived from the northeast on the evidence of palaeocurrent structures (ripples, grooves, flutes and crescent marks) when these are carefully corrected for the tectonic overturning which the Lower Tectonic Unit has suffered (Bertini & Zan 1974, Naylor unpublished data). Since ophiolite was clearly available for erosion to the northeast of the site of deposition, it seems likely that the ophiolite slide-blocks were also derived from the northeast down the same palaeoslope.

Precursor olistostromes from the Bracco ridge?

Ophiolitic olistostromes such as the Casanova Complex have been previously ascribed to extrusion of an ophiolite mega-flow with associated spalling of debris and pyroclastic activity (e.g. Labesse 1962, Cortemiglia 1963), to intrusions of ophiolites into flysch (e.g. Rovereto 1939), and to tectonic brecciation at the bases of nappes (e.g. Merla 1951).

More recently, the Casanova Complex has been interpreted as a series of precursor olistostromes to large nappes, in the so-called Bracco ridge model (e.g. Elter & Raggi 1965, Elter & Trevisan 1973). The following units were believed to have a common origin: the olistostromes of Palombini limestone, ophiolitic breccias, sandstones and gravity-slid masses. The Bracco ridge, represented by the intact but allochthonous ophiolites of southeast Liguria (Fig. 1) was supposedly uplifted in the Late Cretaceous (Fig. 12), became asymmetric to the northeast and with its sedimentary cover evolved into the NE-moving Ligurid nappes. Hence the ophiolitic debris was shed northeastwards as a precursor to the arrival of the nappes, and was subsequently over-ridden and tectonised by the nappe (Fig. 12).

The model is rejected since neither the olistostromes nor ophiolitic debris came from the southwest (see above). Furthermore, these deposits are not synchronous with nappe emplacement: they were deposited during the Cretaceous and were not deformed until the Miocene (Naylor 1978b, Schamel 1974, Sestini 1974) when the Apennine thrust-pile was created.

A distal passive continental margin

Because the olistostromes are interbedded with rocks of the Vara Complex (Fig. 2), their environmental setting is that of the Vara Complex. This was shown to be a sequence of pelagic and turbiditic sediments on ocean crust (see earlier) and such an interpretation is supported by palinspastic reconstructions of the Apennine nappe pile (Fig. 13a) (Reutter & Groscurth 1978), and by the petrography and geochemistry of the ophiolite (summarised by Barrett & Spooner 1977).

Lithological comparisons indicate that the ophiolitic sands and blocks were all derived from the Vara Complex ophiolite. The association of ophiolite with Hercynian granite clasts implies that oceanic and continental crust were exposed in the source area. It is suggested that this source was the junction between oceanic and continental crust at the foot of a continental margin. This hypothesis is supported by the transitional to alkaline chemistry of the basalts, as indicated by their petrography. A similar transition from alkaline lavas at the base of a distal continental margin sequence to the tholeiitic lavas of an ophiolite has been described from elsewhere (Smith *et al.* 1975). The alkaline lavas are interpreted on the basis of stratigraphic (Smith *et al.* 1979) and geochemical evidence (Hynes 1977) as early rifting products related to the initial splitting of continental crust. As ocean-floor spreading develops, the alkaline lavas pass laterally into ocean-ridge tholeiites. If a similar relationship is valid in the Apennines, it implies that the ophiolite in the Casanova Complex, although created by spreading rather than early rifting, was derived from a part of the ocean close to the continent.

That the margin was passive rather than active at the time of mélange formation is indicated by the following observations.

(a) The lack of concurrent deformation, other than downslope slumping (cf. Bachman 1978).

(b) The simple facies pattern of the Lavagna shale turbidites (Naylor 1978b)—typical of a passive

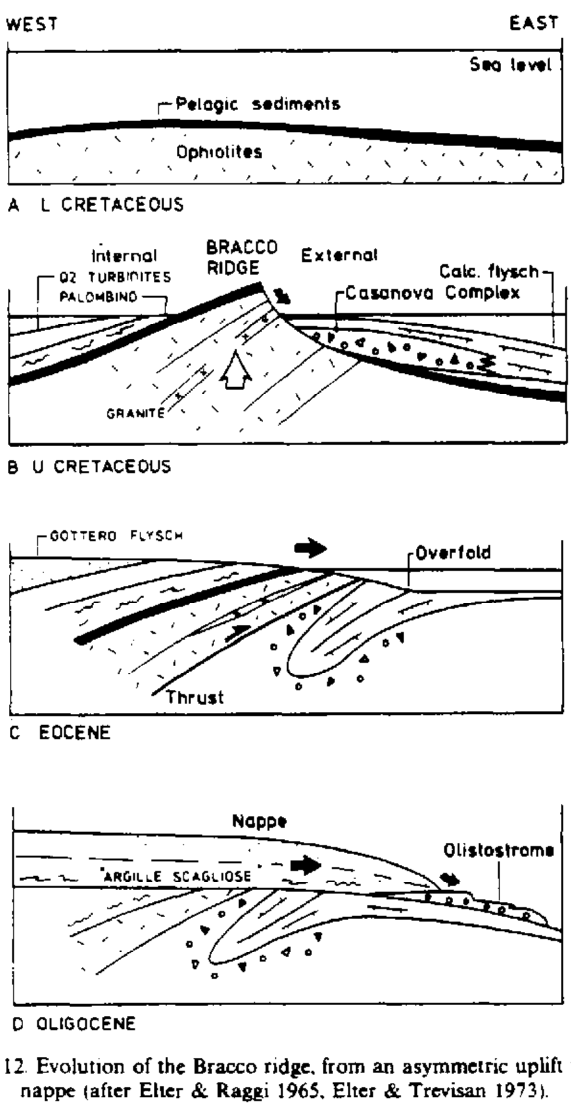

Fig. 12. Evolution of the Bracco ridge, from an asymmetric uplift to a nappe (after Elter & Raggi 1965, Elter & Trevisan 1973).

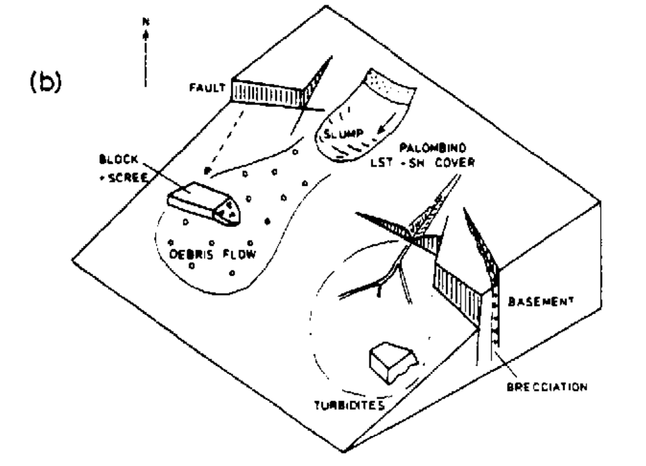

Fig. 13. (a) Palinspastic restoration and Late Jurassic-Cretaceous facies model for the Northern Apennine tectonic units, showing the continental margin, oceanic crust and location of the Casanova Complex. (b) Cartoon indicating the processes involved in genesis of the Casanova Complex.

margin submarine fan—and other units, and the absence of small elongate discrete tectonised sediment ponds (cf. Moore & Karig 1976, Bachman 1978).

(c) The mineralogically-mature nature of most of the sediments, especially the Lavagna shale turbidites (cf. Wezel 1975).

(d) The palaeoslope indicated by the slumps in the Palombini is constantly southwestwards (Naylor 1978a, b) and unidirectional, with no evidence for two source directions as in many trench systems.

(e) Palaeocontinental maps (Smith & Briden 1977) indicate that the Ligurian ocean, of which the Casanova area forms the eastern flank, was still wide and had not begun to close. The area was far from any known subduction system.

Within this distal margin, two parallel basins are postulated, the slightly older unit represented by the Middle Tectonic Unit, and a slightly more 'proximal' and younger basin represented by the Lower Tectonic Unit. In the Middle Tectonic Unit, after a brief phase of ophiolitic sandstone deposition, slumping occurred in the Palombini and olistostromes formed the mélange matrix. Later, in the Lower Tectonic Unit, when slumping waned but current erosion was apparently more active, ophiolitic and relatively coarse quartzose sandstones formed the mélange matrix. Simultaneoulsy in the more distal Middle Tectonic Unit, deposition of finer quartzose sands occurred. At all times, fault-bounded basement blocks of oceanic and continental crust existed, supplying large slide blocks of 'exotic' material to the mélange. Movement on the faults need be as little as 600 m to expose all the ophiolite lithologies. Fracturing and brecciation probably occurred during uplift, enhancing erosion and allowing the blocks, which were cut free, to shed scree-type breccias. Uniform palaeoslope and current indications suggest that the uplifts were probably fault scarps broadly parallel to the continental margin, perhaps the most distal occurrence of block faulting on the margin. The two basins represented by the Lower and Middle Tectonic Units were probably formed in tilted fault blocks. Protrusion of serpentinite along faults parallel to the margin may also have occurred (cf. Bonatti *et al.* 1973, Bonatti & Honnorez 1976), as in the case of the Iberian continental margin (Boillot *et al.* 1980). The environmental model, and the complex interplay of slumping, debris flow, sliding and turbidite sedimentation are summarised in Fig. 13(b).

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APPENDIX

Derivation of equilibrium equations for slide-blocks stability

For a simple block resting on sediment, the yield stress of the sediment, and hence the maximum load applied by the block before failure is

$$\sigma_{YS} = n\,K \tag{1}$$

(Johnson 1970), where n depends on the geometry of the block and the shape of the slip surface (n = 5.14, 4.83 for infinitely long prisms and upright cylinders, respectively on a thick substrate). The coefficient n takes into account the stress field and slip-line distribution in the substrate. Most blocks can be treated as infinite prisms (n = 5.14) because their length greatly exceeds their width (Table 1). The stress applied by the block is

$$(\rho_b - \rho_w)\,g\,H\,, \tag{2}$$

that is its submerged weight per unit base area. For the block to be stable,

$$(\rho_b - \rho_w)\,g\,H \leqq n\,K$$

or

$$H \leqq \frac{n\,K}{(\rho_b - \rho_w)\,g}. \tag{3}$$

In reality, sediment density and strength vary as a function of depth. Furthermore, considering the general case in which a block is partially submerged in sediment, then the stress applied by the block is:

$$\text{submerged weight of part in sediment } (h) + \text{submerged weight of part in water } (H-h)$$

divided by the base area of the block. Hence the applied stress is

$$g\{hI + (H-h)(\rho_b - \rho_w)\} \qquad (4)$$

where

$$I = \int_0^z \mathrm{d}z\{\rho_b - 1.11 + 0.04z + 0.0004(z-15)\}. \qquad (5)$$

The function within the integral sign represents the variation of sediment density with depth (references in Table 1).

Failure of a plastic under an infinitely long punch of width W may be considered (Chen 1975) as failure along a slip-surface of circular cross-section with radius W. The effective strength of the plastic substrate is that at a depth $W/3$ below the base of the punch (Chen 1975, Kubick, pers. comm.). Thus K in Equation (1) has to be replaced by K at the appropriate depth:

$$\sigma_{YS} = nK = n\{K_0 + \lambda(h + W/3)\}. \qquad (6)$$

Equating (4) and (6), evaluating the integral and substituting for I according to (5) gives:

$$g\{(\rho_b - \rho_w)(H-h) - 0.02z^2|_0^{15} - 0.002z^2|_{15}^{h}\}$$
$$= 5.14\{35 + 14(h + W/3)\}. \qquad (7)$$

Equation (7) can be evaluated for two extreme conditions:

$$\text{No sink-in } (h = 0) \quad H = 1.03 + 0.15W \qquad (8a)$$

$$100\% \text{ sink-in } (H = h) \quad W = -24.38 + 4.25H - 0.08H^2. \qquad (8b)$$