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ANATOMY AND EMPLACEMENT MECHANISM OF A LARGE SUBMARINE SLIDE WITHIN A MIOCENE FOREDEEP IN THE NORTHERN APENNINES, ITALY: A FIELD PERSPECTIVE

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ABSTRACT. The Casaglia Monte della Colonna (CMC) body, from the middle Miocene Marnoso-arenacea Formation of the Northern Apennines, covering more than 350 square kilometers, provides a rare opportunity to study the geometry and the internal deformation of a large, basin-wide submarine landslide. Although the head zone of the landslide has not been preserved, the body has a transitional zone, a contractional ramp cutting off more than 200 meters of the footwall stratigraphic succession, and a very wide lobe spilling over onto the adjoining basin plain deposits. The extent of the body, the depth of the ramp cut-off and the widening of the lobe are comparable with the largest present-day submarine landslides, not often observed in fossil examples.

Internal deformation structures are distributed differently in the ramp and lobe zones. The ramp zone has antiformal stacking of duplexes and steeply inclined folds. The lobe has a well-defined strain partition: the upper part of the body is affected by extensional structures such as listric normal faults and extensional duplexes; the lower part is deformed by recumbent folds, boudinage and stacking of blocks, which are compatible with flow-induced, heterogeneous simple shear. This distribution is consistent with a kinematic model of extrusion-spreading, implying rear compression in the ramp zone with thickening and shortening of the body, and spreading in the lobe with thinning and stretching. The localized buckling present in the distal part of the lobe may be related to the onset of lateral confinement, due to local topographic features of the basin plain (synsedimentary intrabasinal high).

The various degrees and styles of stratal disruption in the mass wasting body depend on the different combination of progressive simple shear, layer-parallel extension and shortening. Different structural associations occur in well-defined parts of the body due to kinematics of emplacement (spreading, flow and buckling). These associations may be diagnostic of gravitational mass wasting processes rather than shallow-level tectonic deformation.

INTRODUCTION

Sedimentary bodies from submarine landslides are abundant in both the present-day seafloor setting and in the sedimentary record of ancient basins. These bodies have received considerable attention in the scientific literature, especially in works dealing with specific academic interests, such as the distinction between different genetic processes (sliding, slumping, debris avalanches, and flows), and also dealing with how density flows (grain flows and turbidites) are triggered (see, for example, Hampton, 1972; Middleton and Hampton, 1973; Stow, 1986; Mulder and Cochonat, 1996). Moreover, these bodies can be important as markers in exhumed sedimentary successions helping to delineate basin margins and slopes, and helping to document syn-sedimentary tectonic movements (for example, Rupke, 1976; Woodcock, 1976; Ori and others, 1986; Lucente, 2002; Roveri and others, 2003). Finally, the origin of bodies from mass wasting processes of stratally disrupted rock units (mélanges), which form a considerable part of accretionary wedges (Raymond, 1984; Cowan, 1985), is gaining more and more attention in the international scientific community (Moore and others, 1976; Hibbard and Williams, 1979; Woodcock, 1979a; Brandon, 1989; Barnes and Lewis, 1991; Pini, 1999; von Huene and others, 1999; Choconat and others, 2002). Knowledge of the features of these deposits can help in distinguishing between early, surficial deformations due to mass wasting and the structures related to shallow-level tectonics in wet, non consolidated sediments (Elliott and Williams, 1988; Steen and Andresen, 1997).

There remains several cases of intense practical human interest to study, for example near-shore landslides can cut back the shoreline in populated areas, destabilize the foundation of marine engineered structures, or break submarine cables (Prior and Coleman, 1984; Hampton and others, 1996; Trifunac and others, 2002). Furthermore, large-scale submarine landslides are now considered a common mechanism for generating tsunamis (Moore and Moore, 1984; von Heune and others, 1989; Harbitz, 1992; Jiang and LeBlond, 1992; Hampton and others, 1996; Masson and others, 2002). The kinematics of submarine landslide movement and emplacement seems to control the tsunami wave-forms; thus, the study of the bodies and their relationships with the basins floor can be important for estimating the hazard from their formation (Trifunac and others, 2002; Yalçiner and others, 2002).

The overall morphology, the surface and basal attitude of mass wasting deposits on the present-day sea floors have been outlined in detail by side-scan images, multibeam bathymetry, high resolution seismic and sub-bottom profiles, (see Dingle, 1977; Embley and Jacobi, 1977; Moore, 1977; Prior and others, 1982, 1984; Normark and Gutmacher, 1988; Trincardi and Normark, 1989; Trincardi and Argnani, 1990) and more recently by 3D seismics (for example, Huvenne and others, 2002). However, only partial information on the internal structures comes from present-day examples. This fact is related to the ambiguity of interpretation (compare Gardner and others, 1999; and Lee and others, 2002) and to the presence of transparent zones (Coleman and Garrison, 1977) in seismic and acoustic profile of submarine mass wasting bodies.

By contrast, ancient mass wasting deposits provide a more detailed source of information on internal structures and depositional processes. However, it is difficult to relate the lateral and vertical changes in the internal deformation to the overall geometry of the bodies, because detailed field studies have only looked at small-scale bodies or at single outcrops of larger bodies (see Woodcock, 1976; Farrel, 1984; Gawthorpe and Clemmey, 1985; Pickering, 1987; Martinsen, 1989; Martinsen and Bakken, 1990). This lack of intensive study is because it is difficult to find large mass wasting bodies, with extensive and widely distributed outcrops that can be correlated with absolute certainty, and that can be linked to the reconstructed morphology and setting of a sedimentary basin.

This paper reports a detailed study of a set of outcrops belonging to a single, very large landslide, the Casaglia-Monte della Colonna (CMC) body. The CMC covers more than 350 square kilometers and reaches a thickness of more than 300 meters. This body is comparable with present-day mass wasting deposits (see, for example Moore and others, 1976; Trincardi and Normark, 1989) and is one of the largest on-land examples of a fossil submarine landslide (fig. 1) (Woodcock, 1979a; Macdonald and others, 1993). The CMC and other large mass wasting bodies, as well as the surrounding normal-bedded sediments, are excellently exposed in the Marnoso-arenacea Formation of the Northern Apennines, which provides a well-studied example of a well-preserved turbidite-dominated foreland basin (Ricci Lucchi, 1986, and references therein; Argnani and Ricci Lucchi, 2001; Roveri and others, 2002). In the CMC body, a wide range of structures with varied associations deforms the sedimentary layers and

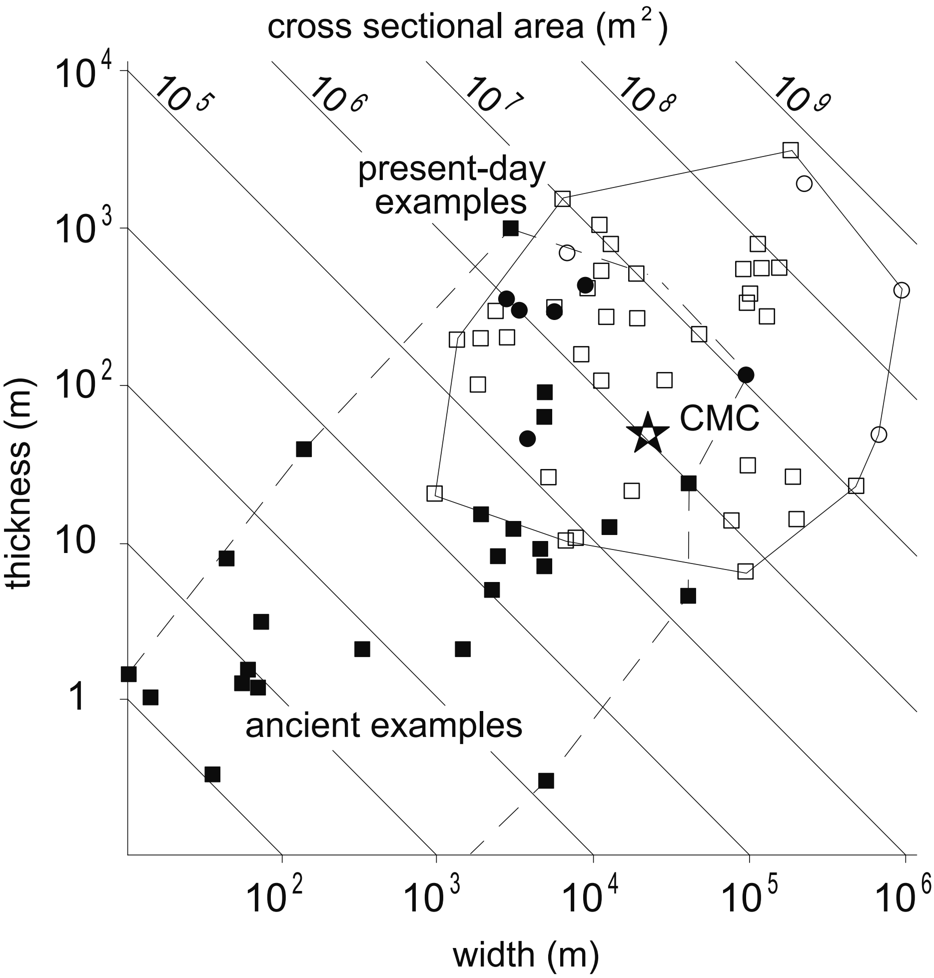

Fig. 1. Plot of thickness versus width of ancient (solid squares and circles) and present-day (open squares and circles) submarine landslides. Outlined fields are from Woodcock (1979a) modified on the base of Macdonald and others (1993). Data from Woodcock, 1979a (squares) and Macdonald and others, 1993 (circles). The star is the plot of the mean thickness and width of the CMC landslide body.

leads to different degrees of stratal disruption. The lateral/vertical transitions can be directly observed in the outcrops and are therefore easy to study in the field. The field observations and the study of the stratigraphy, coupled with the well documented understanding of the basin geometry (Ricci Lucchi, 1975, 1981; Roveri and others, 2002), lead us to propose some hypotheses that will be tested in this work:

1) It is possible to integrate the data from sub-sea remote sensing with field data from a fossil slide because the size, the external morphology and some internal features of the Marnoso-arenacea Formation and present-day large mass wasting bodies are comparable.
2) The geometry and the internal deformation of large mass wasting deposits such as the CMC body was strongly influenced by the mechanism of slope failure and the morphology of slope and basin. Thus, the distribution of the internal structures and the composition and geometry of a mass wasting body

may provide a kinematic model of slide emplacement helping to reconstruct the paleogeography of the accretionary front-foredeep system;

3) Some features of the CMC body, such as the deep-rooted ramp that cuts up through the stratigraphy of underlying basin plain deposits, can be confused with shallow-level contractional tectonics. However, the association of deformation structures and the strain partition inside the body, together with the style and the progressive changing in degree of stratal disruption, can be used to discriminate mass wasting processes from shallow-level tectonics.

NOTES ON MASS WASTING DEPOSIT TERMINOLOGY

The terms "slump" and "slide" are referred to in the literature as deposits produced from various mass gravitational processes (Woodcock, 1979a; Prior and Coleman, 1984; Gawthorpe and Clemmey, 1985; Pickering and others, 1986). Some classifications of mass wasting deposits (Nardin and others, 1979; Stow, 1986; Nemec, 1990) make a clear distinction between slide and slump mechanisms. Slide masses move on planar glide surfaces (translational movements) and show no or little internal deformation, whereas slump masses move on an upwards-concave glide surface (rotational movements) and show a wide range of internal deformation mechanisms.

However, slide-like deposits (large, practically undeformed olistoliths of either intrabasinal and extrabasinal origin) coexist alongside slump-like deposits within large, mass wasting bodies (for example, CMC), and stratal disruption (that is downslope increases of mass desegregation) causes the transformation of slumps into debris flows (for example the bodies of Susinello and Romiceto in the Marnoso-arenacea Formation, Lucente, ms, 2000). Similar associations have been recognized in present day examples (for example, Jacobi, 1976; Prior and others, 1984; Normark and Gutmacher, 1988; Masson and others, 2002). Given these associations, more general terms, such as "mass wasting deposit" and "submarine landslide", are used in this paper to refer to the Marnoso-arenacea bodies.

Another controversial term is olistostrome. It was introduced by Flores (1955) to indicate sedimentary bodies with a chaotic, block-in-matrix fabric inside layered sequences of normal marine deposits in the Tertiary succession of Sicily. This definition has been extended to similar bodies in the rest of the Apennines (for example, Abbate and others, 1970, 1981; Elter and Trevisan, 1973; Ricci Lucchi, 1975; Naylor, 1981, 1982; Pini, 1999). The definitions used do not include size (Abbate and others, 1981), and the accepted broad definition given includes the following different types of sedimentary bodies (Abbate and others, 1981; Pini, 1999; Lucente, ms, 2000; Cowan and Pini, 2001):

—bodies with blocks ranging in size from a centimeter to a meter, dispersed in a clay or sand-silt matrix (type A olistostromes, Pini, 1999);

—bodies with larger blocks (olistoliths) ranging in size from tens to hundreds of meters sustained by a matrix with the features of a type A olistostrome (type B, Pini, 1999);

—bodies almost completely consisting of olistoliths, without a matrix, such as the CMC extrabasinal component (type C olistostromes, this work).

These bodies derive from different types of gravity mass movements, such as block slides, debris avalanches, debris flows, and hyperconcentrated flows. The blocks, the olistoliths and the matrix are typically composed of extrabasinal rocks coming from paleo-Apennine accretionary wedge material, namely the Ligurian and Subligurian rocks; however, a certain amount of intrabasinal materials are found in the olistostromes of some settings, such as the Epiligurian piggyback basins (Pini, 1999, and references therein; Lucente, ms, 2000). These bodies are, therefore, made up prevailingly of rocks that are "exotic" with respect to the host successions.

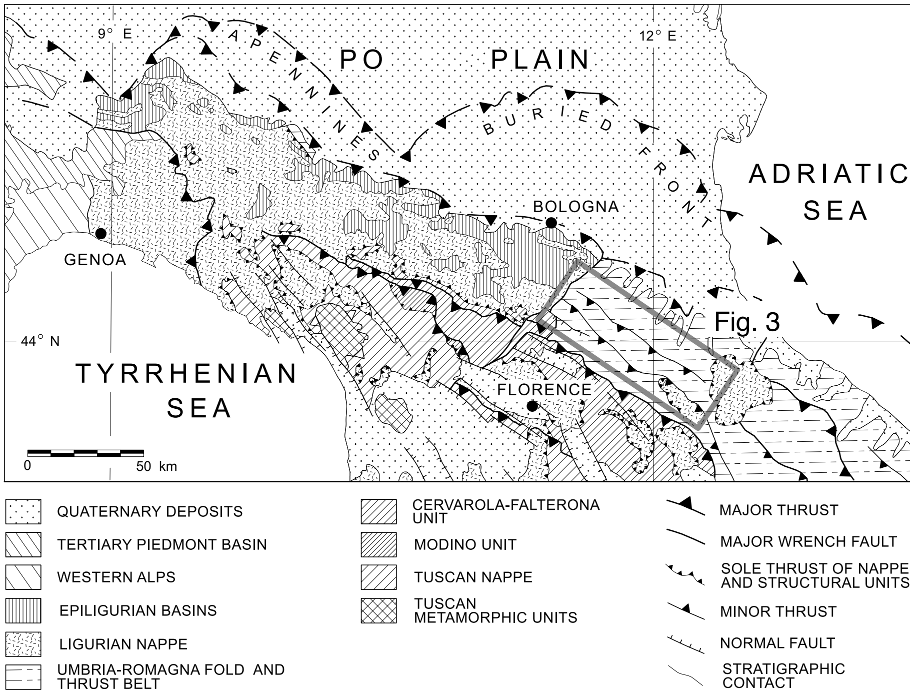

Fig. 2. Structural sketch map of the northern Apennines (from Pini, 1999, and references therein).

It is also important to point out that, even though the term mélange has been used prevailingly with the sense of tectonic product in mind by several authors in both the American and in the international literature, the term olistostrome has been applied to mélanges in many orogenic zones throughout the world, implying a possible origin from mass wasting processes (see for example, among many others, Rigo de Righi and Cortesini, 1964; Cowan, 1978, 1985; Page, 1978; Hibbard and Williams, 1979; Swarbick and Naylor, 1980; Page and Suppe, 1981; Sarwar and De Jong, 1984; Brandon, 1989; Horton and Rast, 1989; Chanier and Ferriére, 1991; Taira and others, 1992; Bader and Pubellier, 2000; Kuzmichev and others, 2001; Cartier and others, 2001; Cawood and others, 2002).

GEOLOGICAL SETTING

The Northern Apennines

The Northern Apennine chain (fig. 2) is an orogenic, northeast verging wedge composed of stacked thrust sheets and nappes. The structurally highest units, the Ligurian units, are the remnants of the Jurassic-Eocene sedimentary cover and the Jurassic basement of the Ligurian ocean, a part of the Alpine Tethys, deformed in the late Cretaceous-Eocene eo- and meso-alpine tectonic phases (Abbate and others, 1986; Vai and Castellarin, 1993; Castellarin, 1994; Marroni and Treves, 1998; Pini, 1999). The Ligurian units overlie the Subligurian units, which are deformed Cretaceous-Oligocene sedimentary cover of the westernmost part of the continental margin of the Adria microplate (Bortolotti and others, 2001). The Ligurian and Subligurian units together constitute the Ligurian Nappe, shown in figure 2, which today extends from the Tyrrhenian Sea to the southwest margin of the Po Plain. The Epiligurian successions are middle Eocene-Pliocene sediments deposited in satellite basins on top of the translating Ligurian nappe (Ricci Lucchi, 1986; Mutti and others, 1995; Cibin and others, 2001).

The Ligurian Nappe is structurally above the Tuscan units (Cervarola-Falterona and Modino units, Tuscan metamorphic units, Tuscan Nappe) and the Umbria-Romagna fold and thrust belt, the latter deriving from convergence, thrusting and accretion of different paleogeographic domains of the Adria microplate. These domains share a common stratigraphy ranging from Permian-Mesozoic continental and marine deposits, to Paleogene slope deposits, up to upper Oligocene-Miocene thick turbiditic successions interpreted as the sedimentary infillings of foreland basins (see Argnani and Ricci Lucchi, 2001, and references therein). The turbiditic successions become younger from southwest to northeast (Ricci Lucchi, 1986; Boccaletti and others, 1990; Vai and Castellarin, 1993; Costa and others, 1998; Argnani and Ricci Lucchi, 2001). Their ages range from late Oligocene-early Miocene in the southwest to mid-late Miocene (Langhian-Tortonian) for the Marnoso-arenacea Formation in the northeast. The different ages of the turbiditic deposits have been attributed to the progressive northeastward migration of the paleo-Apennine accretionary wedge and related foredeep through time (Merla, 1951; Boccaletti and others, 1971, 1990; Ricci Lucchi, 1986; Castellarin, 2001). All the foredeep successions and their underlying sequences took part in the formation of the Apennine chain and gradually became involved as structural units.

The Marnoso-arenacea Formation

The Marnoso-arenacea Formation is a composite, wedge-shaped sedimentary body, locally exceeding a thickness of 3000 meters (Ricci Lucchi, 1975, 1986). It crops out extensively over an area approximately 180 kilometers long and 40 kilometers wide as a part of the Umbria-Romagna fold and thrust belt (fig. 2). The largest part of the Marnoso-arenacea Formation is buried beneath the overthrust of older and inner foredeep deposits to the southwest (Sani, 1990; Landuzzi, 1991), the Ligurian Nappe to the northwest, and Messinian to Quaternary deposits to the north and east (figs. 2 and 3).

The Marnoso-arenacea Formation is characterized by intermittent turbiditic deposits that may be subdivided into an older and a younger stage, (fig. 4) defined as inner and outer with respect to the orogenic polarity (toward northeast). The separation occurs at the Serravallian/Tortonian boundary and is marked by an increase in the sand/pelite ratio and an apparent decrease in carbonate-clastic input (Ricci Lucchi, 1981).

The inner stage represents the period of maximum subsidence and extent of the basin, with a high development of basin plain sediments deposited by mature, low density flows (mainly facies D_1, D_2, D_3 and G, as described by Mutti and Ricci Lucchi, 1972). A paleogeographic restoration displays an elongated basin, with a northwest-southeast main axis (Ricci Lucchi, 1975). The size of the plain was intermittently reduced by the deposition of sandstone lobes advancing from the northwest (Ricci Lucchi, 1981), by uplift in some areas forming intrabasinal highs and slopes (Ricci Lucchi, 1975; De Jager, 1979) and by the accumulation of large deformed and chaotic masses due to submarine landslides (see below).

Bedding parallelism and lateral continuity are the most characteristic features of the rocks from the inner Marnoso-arenacea Formation. Some basin-wide lithostratigraphic markers have long been identified. In particular, the Contessa bed (Renzi, 1964; Ricci Lucchi and Pialli, 1973) is a siliciclastic-carbonatic hybrid megaturbidite of southwestern provenance (Gandolfi and others, 1983). The stratigraphic interval above the Contessa bed (post-Contessa succession, Ricci Lucchi, 1975) has distinctive calcareous turbidites, known as colombine layers, of southeastern provenance. The

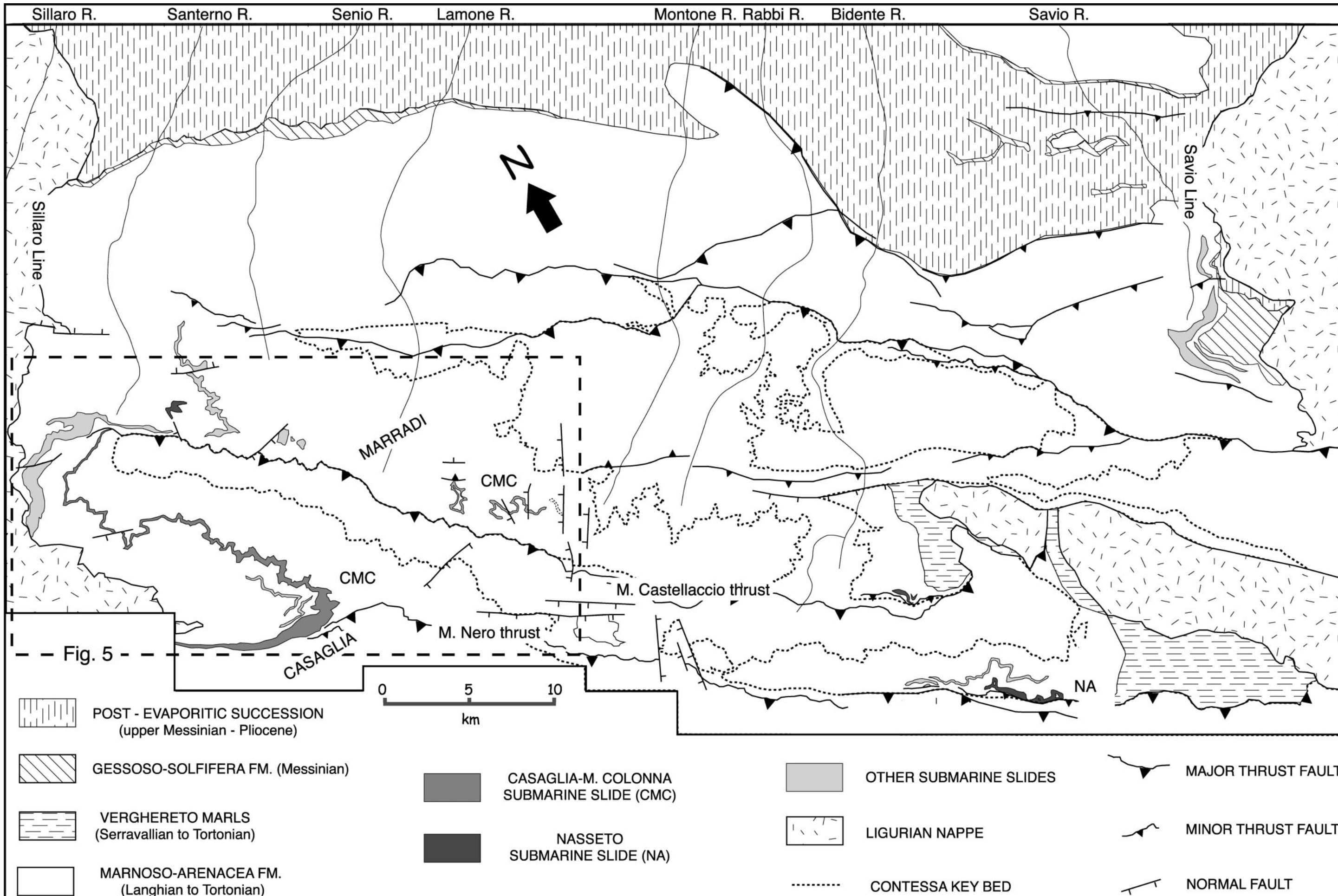

Fig. 3. Schematic geologic map of the Marnoso-arenacea Fm. between the Sillaro and the Savio valleys (Tuscan-Romagna Apennines) (from Capozzi and others, 1991; modified). Location on figure 2.

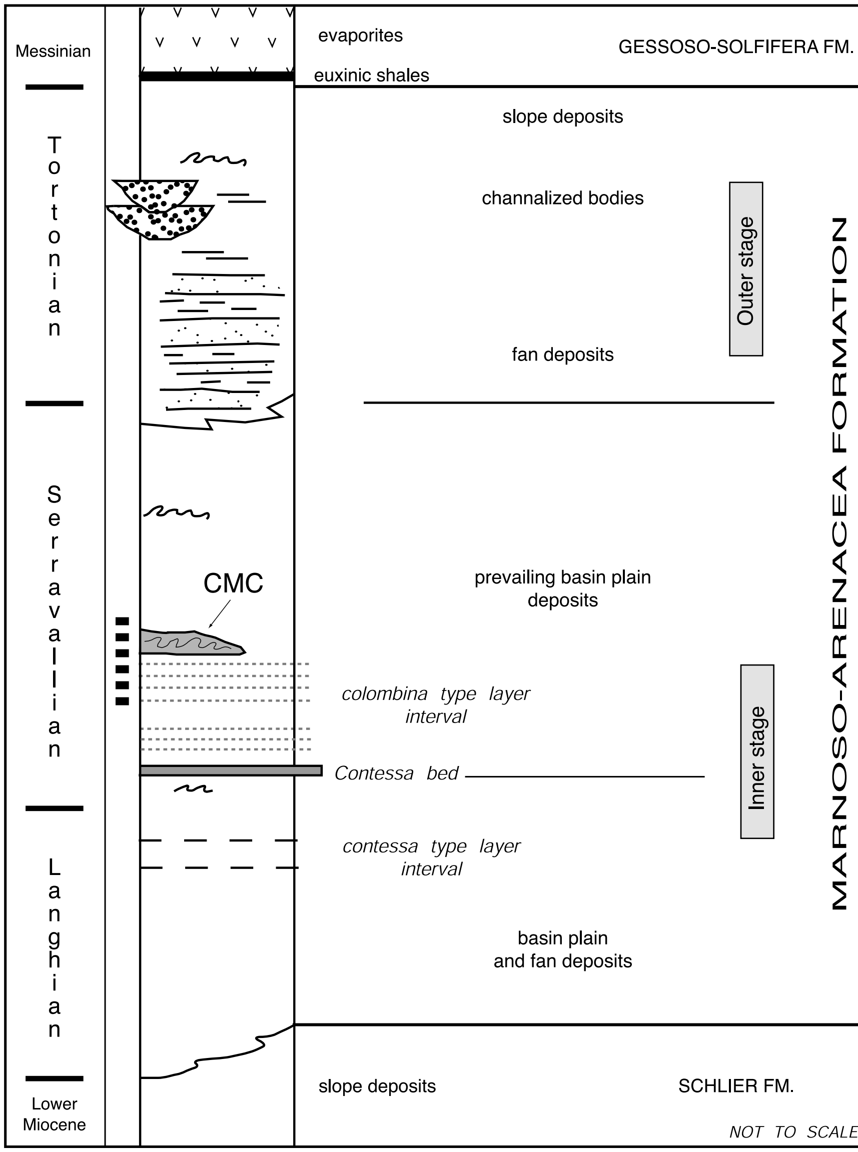

Fig. 4. Litho- and chronostratigraphic subdivision of the Marnoso-arenacea Fm. The stratigraphic interval studied in this paper is pointed out by the vertical dotted line. (From Ricci Lucchi, 1981; Lucente, ms, 2000, modified).

colombine layers alternate with the dominant siliciclastic turbidites fed by the main, northwestern Alpine source (Gandolfi and others, 1983). Due to their lateral continuity and their episodic occurrence, some colombine layers have been used as marker beds (Ricci Lucchi and Valmori, 1980).

Recently, new markers have been identified and others, already known from the literature, have been extensively correlated to outline the geometry of the Serravallian large submarine landslide bodies and the basin morphology from the stratigraphy (Lucente and Pini, 1999; Lucente, ms, 2000). The key beds are, from bottom to top:

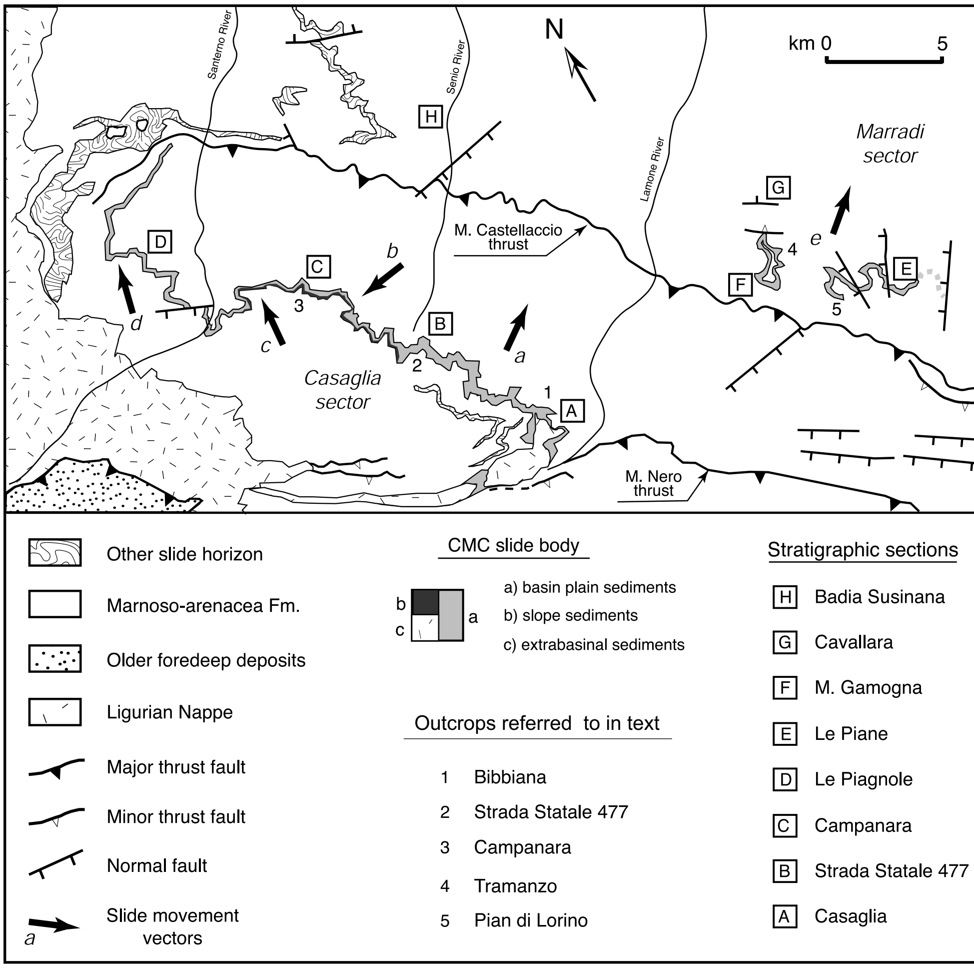

Fig. 5. Schematic geologic map of Casaglia-Monte della Colonna (CMC) slide. Numbers and capital letters indicate the location of the stratigraphic sections and of the outcrops described in this paper. (Lucente, ms, 2000). The arrows show the main vectors of the slide movement.

the Montellero layer (MT), the Crespino layer (CR), the Poggio Serra layer (PS), the Maestà delle Valli layer (MV), the Avalcelli layer (AV), the Ciriegiolo horizon (CG), and the Bibbiana layer (BB). MT, CR, MV and AV are calcareous turbiditic layers (colombina type), whereas PS, CG and BB are siliciclastic layers.

The outer stage consists of fan and channeled sand bodies (confined lobes) deposited by more immature, high-density flows (mainly facies C, B_1 and A, Mutti and Ricci Lucchi, 1972). During this stage the sedimentation occured in a narrower basin due to the tectonic break up of the previous basin plain. The main source of sediments was the Alps, but in this case a more easterly area than before (Gandolfi and others, 1983).

The Marnoso-arenacea Formation was involved in a system of northeast-verging imbricate thrusts (De Jager, 1979; Capozzi and others, 1991; Landuzzi, 1991) (figs. 2, 3 and 5) at the same time as, or immediately after, the emplacement of the Ligurian Nappe. Thrusts strike from northwest to southeast, parallel to the main direction of the

Apennines thrust belt, and are associated with fault-propagation folds (Landuzzi, 1991).

The presence of intrabasinal highs in the Serravallian Marnoso-arenacea basin due to synsedimentary growth of anticlines is an idea that has become progressively popular in the literature (De Jager, 1979; Farabegoli and others, 1990; De Donatis and Mazzoli, 1994; Roveri and others, 2002). This early differentiation of the foredeep basin has been documented by lateral facies change, onlap, and by local unconformities (see De Jager, 1979; Roveri and others, 2002). Synsedimentary detachment folds, with the sole thrust at the base of the foredeep succession, have been proposed to explain the intrabasinal highs and they are thought to have evolved in fault-propagation folds and in the thrust, which are observable in the present-day setting (De Donatis and Mazzoli, 1994). The growth of synsedimentary anticlines does not necessarily imply the existence of isolated sub-basins, but a physiographic separation controlling the deposition of turbidites, because the different thrust units share the same stratigraphy outlined by the same marker layers.

Other major structures are the Sillaro Line and the Savio Line, two tectonic lineaments striking across the Apennine chain in a northeast–southwesterly direction (fig. 3), the interpretation of which is still controversial (for example Bruni, 1973; Castellarin and Pini, 1989; Bettelli and Panini, 1992; Landuzzi, 1994; Zattin and others, 2000; Cerrina Feroni and others, 2001; Lucente and others, 2002; Zattin and others, 2002). These lineaments are the edges of the main areas of emergence of the Marnoso-arenacea Formation, separating to the west the Romagna culmination (Ricci Lucchi, 1986; Capozzi and others, 1991) from depressed zones, which are taken up by the Ligurian Nappe.

The submarine slide studied here, the Casaglia-Monte della Colonna (CMC), and other large bodies, for example the Nasseto (NA) landslide body, occur within the northwestern sector of the Marnoso-arenacea Formation, between the valleys of Sillaro and Savio Rivers (fig. 3). The slides are interbedded in the post-Contessa, basin plain deposits of Serravallian age (inner stage). From the structural point of view, the area has two main thrust faults, the Mt. Nero and the Mt. Castellaccio thrusts. The Mt. Castellaccio thrust splits the CMC into the Casaglia and the Marradi sectors (fig. 3).

The Casaglia-Monte della Colonna body

Composition and Geometry

The CMC, together with many other landslide bodies of the Marnoso-arenacea Formation, is prevailingly made up of basin plain sediments that are turbiditic layers with predominance of $T_{b\text{-}e}$ and $T_{c\text{-}e}$ Bouma sequence (mainly facies D_1 and D_2 of Mutti and Ricci Lucchi, 1972) and hemipelagic beds (facies G of Mutti and Ricci Lucchi, 1972). The CMC also includes subordinate slope deposits and extrabasinal rocks (figs. 5, 6A and B). The slope sediments are mainly characterized by hemipelagic mudstones, with rare interbedded thin-bedded turbidites (facies D_2 and D_3 of Mutti and Ricci Lucchi, 1972). The extrabasinal portion (olistostrome) is made up of Subligurian, Tuscan (De Jager, 1979) and/or Epiligurian succession rocks (Lucente, 2002). Basin plain deposits always occur at the base of the slide, whereas the intrabasinal slope sediments and the extrabasinal rocks occur at the top of the body. In particular, the extrabasinal component crops out in two lenses, in the southwestern most and upper portion of the slide body (sections A1 and A2 in fig. 6B).

The CMC is irregularly wedge shaped. The maximum thickness of 300 meters is in the southernmost part (section A1 in fig. 6B) and corresponds to the major body of extrabasinal rocks overlying a thin (few meters) level of deformed basin plain deposits. The thickness reaches 250 meters of intrabasinal sediments in section A2 and thins out to a thickness of 85 meters in section A (fig. 6A and B). From here the slide thins out

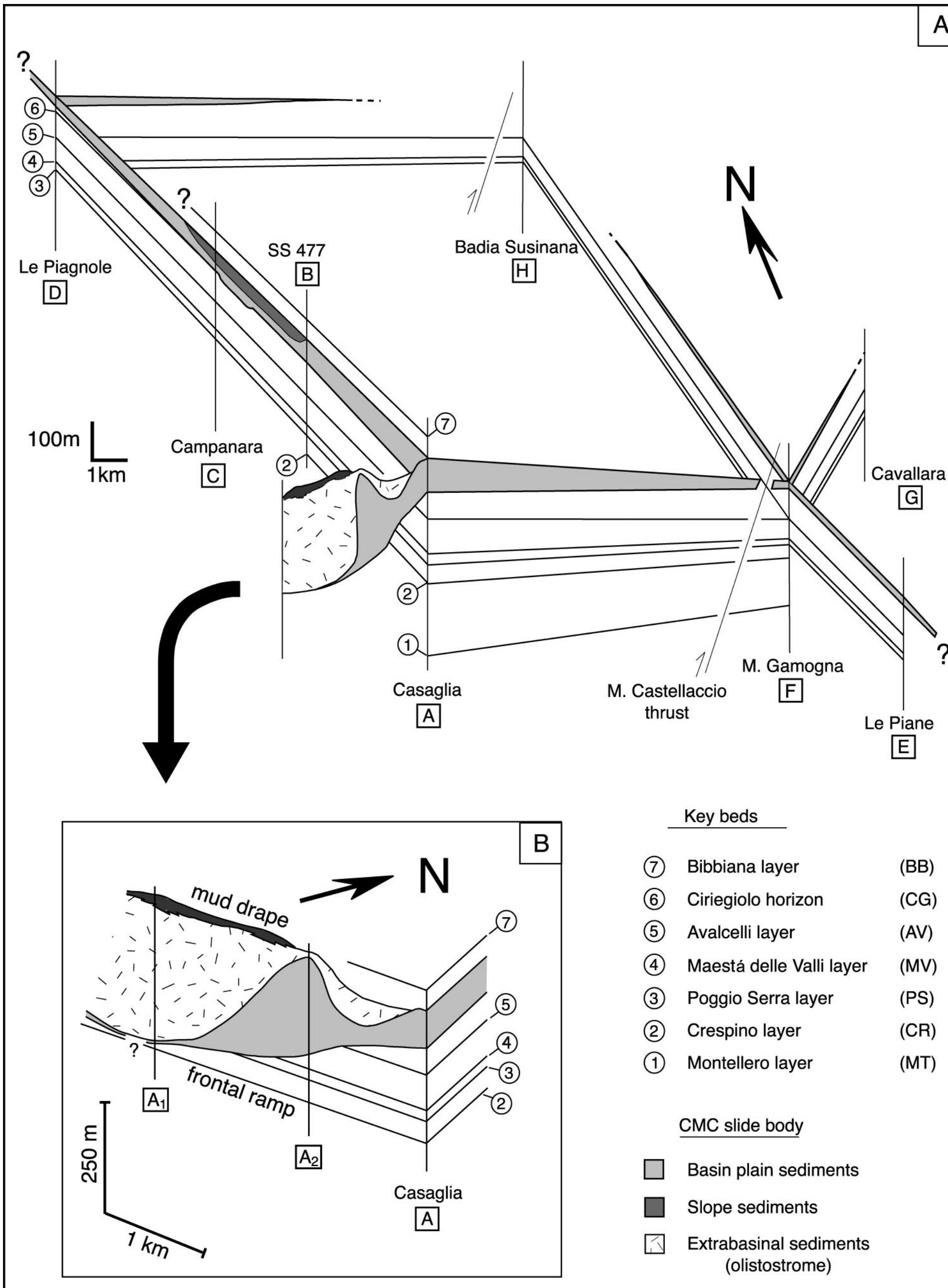

Fig. 6. Fence diagram correlating the stratigraphic sections and showing the geometry of the CMC slide. The location of stratigraphic section is in figure 5 (Lucente, 2002).

gradually both towards the northwest (25 m in section D) and towards the east (10 m in section F).

The basal surface of the body has a complex geometry with ramps and flats. Following Gawthorpe and Clemmey (1985), the ramp is a segment of basal surface cutting across bedding planes and the flat is bedding-parallel. Between sections A1 and A, the basal surface cuts off 200 meters of the footwall succession in 2 kilometers, with an angle of about 6° in a southwest-northeast direction (southwest dipping ramp). This ramp has a contractional geometry (see fig. 8 in Gawthorpe and Clemmey, 1985) cutting up the stratigraphy of footwall. Along a SE-NW transect a flat zone running on the CG key horizon from stratigraphic section B to C is in between two low-angle contractional ramps (A-B and C-D), with the only exception of a small area immediately south of section C where the basal surface cuts down and the CG horizon is involved in the sliding. Along the SW-NE transect a distal flat zone of the CMC landslide (for example Marradi sector in fig. 5; see sections E, F, G in fig. 6A) follows a low-angle contractional ramp (A-F). The low-angle contractional ramps reconstructed by stratigraphic correlations correspond to observable flats where they crop out. They can be interpreted as flats connected by small-scale ramps cutting off few beds, as suggested by the comparison with present-day examples (Trincardi and Normark, 1989).

The top of the CMC is more regular and is progressively lower toward both the northwest and northeast, (fig. 6B). Looking at outcrops, the slide top has numerous irregularities, with small culminations and depressions. All the depressions are filled with arenaceous lenses of variable thickness, up to 3 meters, with an erosional basal contact complicated by load casts and injections of sand (dikes) into the underlying slide sediments.

Summary of Deformational Structures

The CMC body has a large variety of internal deformations that are related to different structures. The more common of these structures are summarized below, and their distribution throughout the body is outlined in figure 7.

Symmetric and asymmetric folds have a very ductile style with thinning of flanks and thickening of the hinge zone. Symmetric folds are prevailingly isoclinal and recumbent, often arranged in systems of cascade folds. Asymmetric folds develop either independently or in association with other, larger scale folds (parasitic folds). The curvature of the hinge-lines is either within the axial surface and therefore almost plane (plane non-cylindrical folds, Hibbard and Karig, 1987) or it appears from the curvature of the hinge line that the axial surface has been warped by refolding of folds with type 2 and type 3 interference patterns (Ramsay, 1967). The evident expulsion of a consistent part of fold cores caused a strong disharmony of folds (intrafolial folds). Overfolds, isolated fold hinges (rootless isoclinal folds) and box folds are present in some parts of the CMC body.

Boudinage and pinching-and-swelling of beds occur dispersed throughout the CMC body. Pinch-and-swell structures are both symmetric and asymmetric (Ghosh, 1993). Boudinage is both shear fracture and lenticular (Ghosh, 1993). Isolated boudins occur frequently. The shape of single boudins is either prolate (constriction-type) or oblate (flattening-type). In some cases, the lenticular boudins are close to each other and stacked (block stacking, Pini, 1999).

Mesoscopic ductile faults are localized surfaces or millimeter sized zones offsetting beds and part of layered succession. Similar structures are present in shallow-level deformation in accretionary prisms and within slides (see deformation bands, hydroplastic faults, and so on, Pickering, 1987; Maltman, 1988; Hanamura and Ogawa, 1993; Yamamoto and others, 2000). Mesoscopic ductile faults are commonly filled by sand

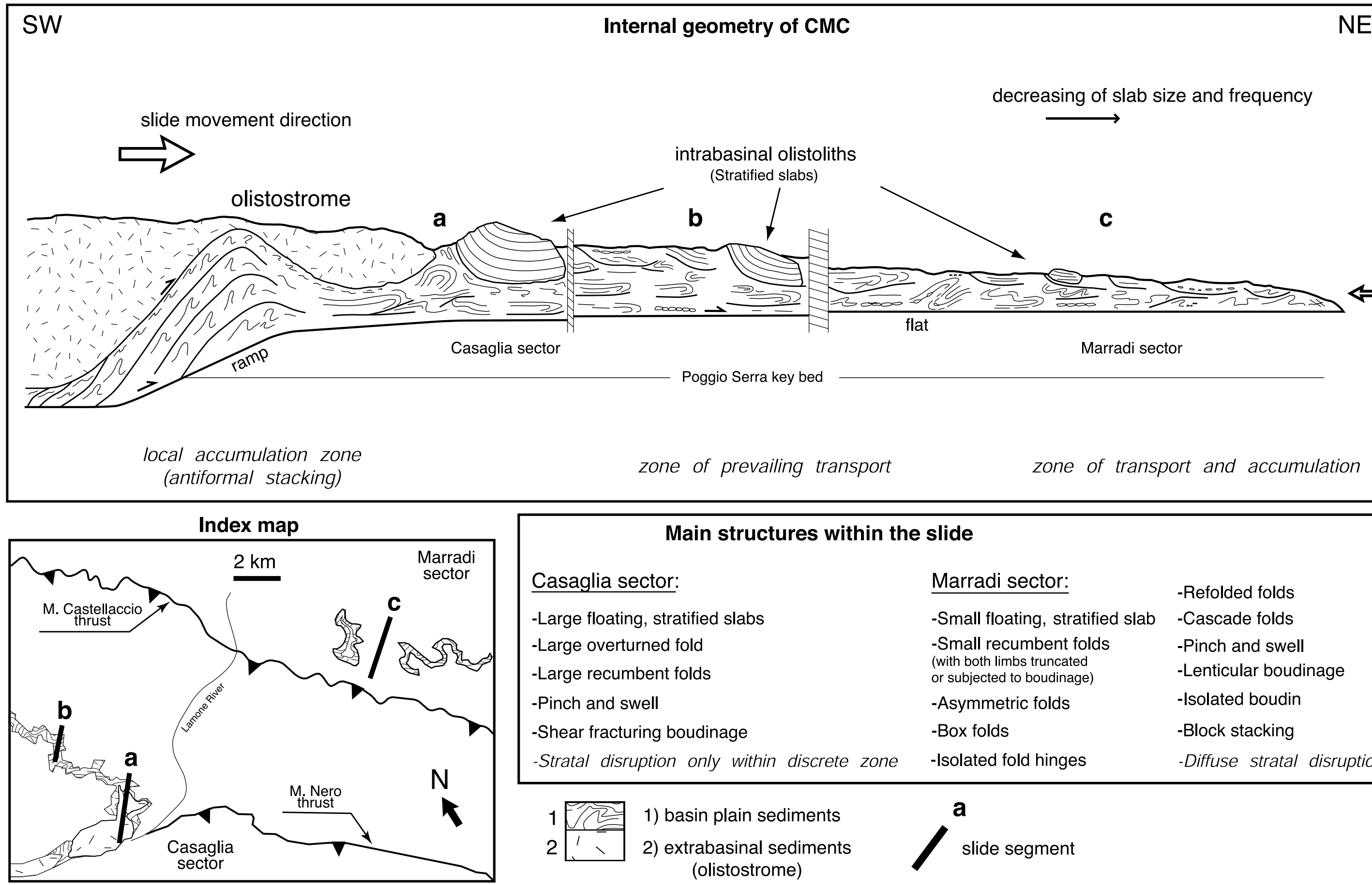

Fig. 7. Idealized transect of CMC parallel to the SW-NE direction of slide movement, showing the external geometry and internal distribution of structures. The sketch outlines three segments (a,b,c), which have been identified on the basis of the body geometry and the type and distribution of structures.

and silt less cemented than the host rocks (see for example fig. 13C) suggesting elutriation of the finer grains by escaping pore fluids (Pickering, 1987).

All these styles of deformation occur without discontinuities, such as cleavage, fractures, void veins, and secondary-mineral filled veins. The original sedimentary layers (laminations) are still observable, and have been bent and folded, or have participated in the boudinage of the beds, but have not lost their lateral continuity. Some web structures (Lundberg and Moore, 1986) have been observed, associated with both folds and boudins.

Microscopic observation and observations on polished cuts reveal that all these structures, and especially pinch and swell of boudins and shear zones, are associated with rotation and translation of grains, without grain breakage. This observation means that all these structures are related to independent granular flow, enabled by low effective confining stress (wet sediment deformations, Maltman, 1984; mesoscopic ductile structures, Cowan, 1982).

Fold Attitude

Folds may indicate the direction of slide movement assuming they verge downslope, with their axes parallel to paleoslope strike (Jones, 1940; Corbett, 1973; Woodcock, 1976; Naylor, 1981; Pickering 1982, 1987). Rotation of fold hinges, nucleation of oblique and downslope folds and the presence of sheath folds may complicate the picture (Lajoie, 1972; Woodcock, 1979b; Ghosh and Sengupta, 1984; Webb and Cooper, 1988). Thus, two different methods of slope directions computation have been proposed and discussed: the mean axis method and the separation-arc method (Hansen, 1971; Woodcock, 1979b). Bradley and Hanson (1998) suggest the usage of two different mean axis methods: the alongslope and the downslope method, in which the mean direction of the fold axes is respectively parallel and normal to the paleoslope strike.

In the case of the CMC slide, different kinds of distribution of fold axes and of asymmetry and vergence directions characterize large, well-delimited domains. Following Woodcock (1979b) and Bradley and Hanson (1998), different computational methods have been used in each domain in order to minimize the interpretational uncertainty.

The alongslope mean axis method gives the more reliable reconstructions in the domains a, b and d (figs. 5 and 8A), and the direction of paleoslope dip has been reconstructed from both asymmetry and vergence of folds because of these reasons: 1) the vergence of folds is always recognizable, because the younging direction of beds in fold hinge zone (facing) is always evident; 2) the orientation of fold axes shows a cluster distribution (fig. 9) and the direction of vergence is constant within each domain; 3) the curvature of the hinge-line in a single fold occurs within the axial surface and is moderate, up to 20 degrees, so that these folds can be treated as plane non-cylindrical folds (Hibbard and Karig, 1987) instead of sheath folds. In these domains, the separation-arc method is poorly reliable, because the fold axes are distributed in cluster rather than girdle (great circles) and the folds are plunging gently (Bradley and Hanson, 1998). However, the separation-arc method of Hansen gives results that are comparable with the alongslope mean axis method (fig. 8A).

In the Marradi sector, domain e, fold axes are distributed in a moderate developed girdle with cluster (figs. 8B and 9) and therefore show a wider dispersion. At the outcrop observation, folds are non-cylindrical, with warping of the hinge zone and axial surfaces. No sheath folds have been observed in field exposures, but their presence cannot be excluded in this sector. Thus, the alongslope mean axis method should be used with care (Woodcock, 1979b; Bradley and Hanson, 1998). However, the presence of box folds and of opposite verging folds are a problem for using the separation-arc method (Bradley and Hanson, 1998). Notwithstanding the great uncer-

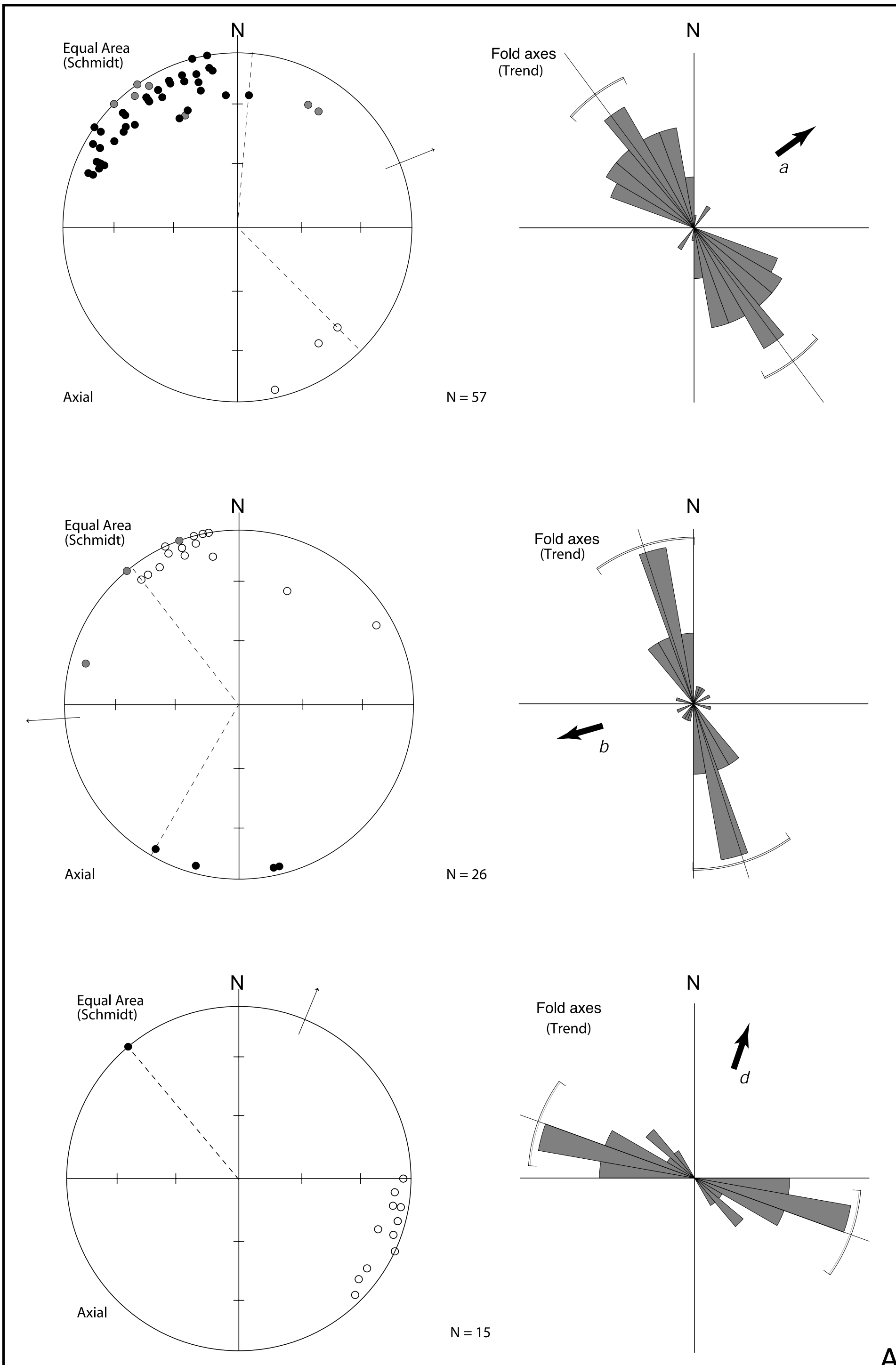

Fig. 8. Equal area stereo plots and relative rose diagrams of fold axes from domains a, b, and d (A) and from domains c and e (B). In the stereo plot, open and solid circles represent dextral and sinistral asymmetry respectively; gray circles show uncertain data. Arrows indicate the slide movements according to the separation-arc method (stereo plot) and the along-slope mean axis method (rose diagram).

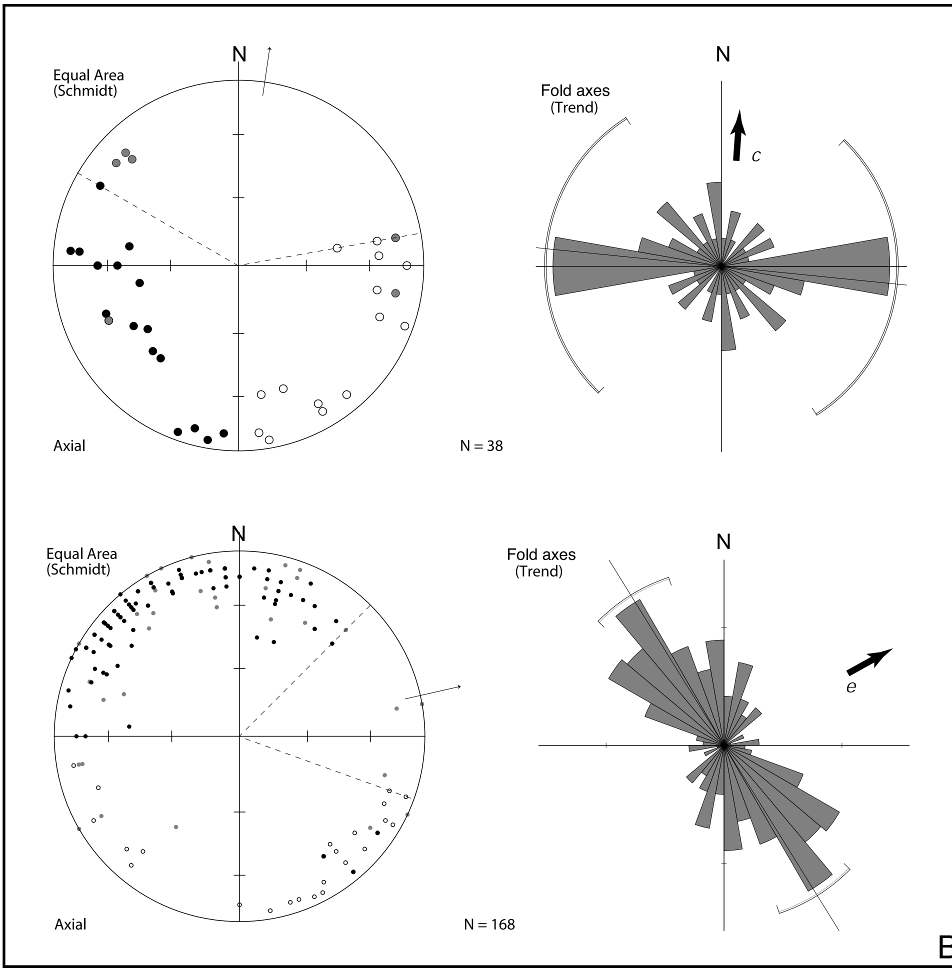

Fig. 8 (continued)

tainty on the reconstructions, the two methods give a similar, northeast direction of movement.

In the Campanara zone, domain c, the slope deposits show a weakly developed girdle distribution of fold axes (figs. 8B and 9). The separation-arc method is considered to be most reliable because of the girdle distribution of fold axes and the absence of box folds and countervergences (Woodcock, 1979b; Bradley and Hanson, 1998). However, the alongslope mean method gives results that are grossly comparable with the separation-arc method.

The inferred directions are consistent within the different domains with the paleoslope strike and the geometry of the basin reconstructed by other independent data (turbidite current direction of flow and reflections; lateral variations of bed thickness and sedimentary structures) (Ricci Lucchi, 1975; De Jager, 1979; Ricci Lucchi and Valmori, 1980; Lucente and Pini, 1999; Lucente, 2002; Roveri and others, 2002). Moreover, the slip directions reconstructed from folds are consistent with other kinematic indicators, such as thrusts, normal faults, extensional duplexes and stacking of blocks inside the landslide body.

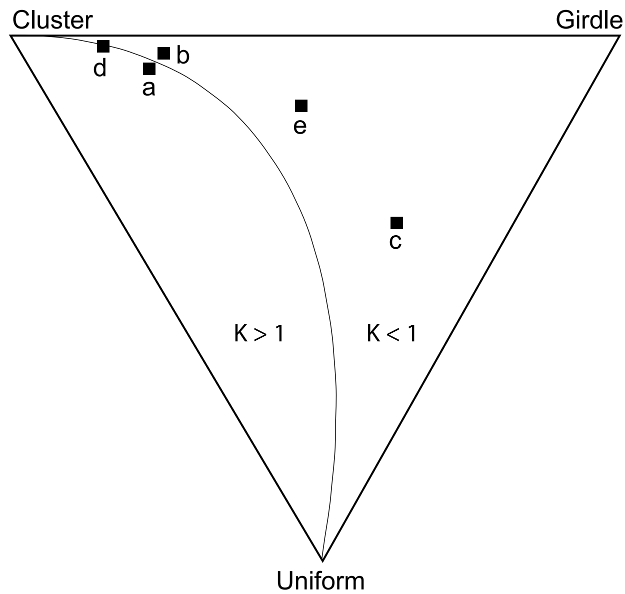

Fig. 9. Triangular fabric diagram plotting the Vollmer's components computed for each fold axis data sub-sets (domains).

Landslide Emplacement, Foredeep Geometry

The vergence of folds in the intrabasinal component of CMC slide indicates top-to-the northeast motion for the region southeast of Senio River in both the Casaglia and Marradi sectors (vector domains a and e in fig. 5) and a top to the north motion for the northern part of the Casaglia sector (vector domains c and d in fig. 5). The steep ramp of Casaglia-Bibbiana (sections A1, A2 and A in fig. 6B; outcrop 1 in fig. 5) is therefore a frontal ramp that allows the detached basin plain sediments to override the adjoining, in place sediments. This conclusion is supported by the presence of MV and AV key beds within the deformed body, immediately to the northeast of the ramp of Casaglia-Bibbiana, where these beds are cut-off. The apparent low angle ramps and the flats between sections A and D (see fig. 6A) and sections A and F and the flat of the Marradi sector (sections F-E-G) represent the base of a lobe of basin plain deposits free to spill over onto the adjoining basin plain deposits. Such external lobes have been described in modern examples of mass wasting deposits (see Prior and others, 1984), but have not been observed in fossil examples so far.

In the case of CMC, the basin plain sediments, coming from the southwest, filled the deepest part of the basin (depocenter, Lamone valley, Lucente and Pini, 1999) and went further eastwards in the southeastern sector (Senio-Lamone, Marradi sector). By contrast, in the northern part of the Casaglia sector (Senio-Santerno) the same sediments were confined to the east and have been deflected northwards (fig. 5). This observation is consistent with the gradual, eastward and northward decrease of slide thickness outlined by the stratigraphic study (fig. 6).

The peculiar characters of the CMC geometry can be related to the paleomorphology of the Marnoso-arenacea basin. Figure 10 shows a tentative reconstruction of the morphology of the basin at the time of CMC landslide. The reconstruction is based on the data from the CMC body presented in this paper, from stratigraphic data (Lucente, ms, 2000) and on the result of the study of the Nasseto (NA) mass wasting body, which is coeval to the CMC (Lucente and Pini, 1999; Lucente, ms, 2000). In this

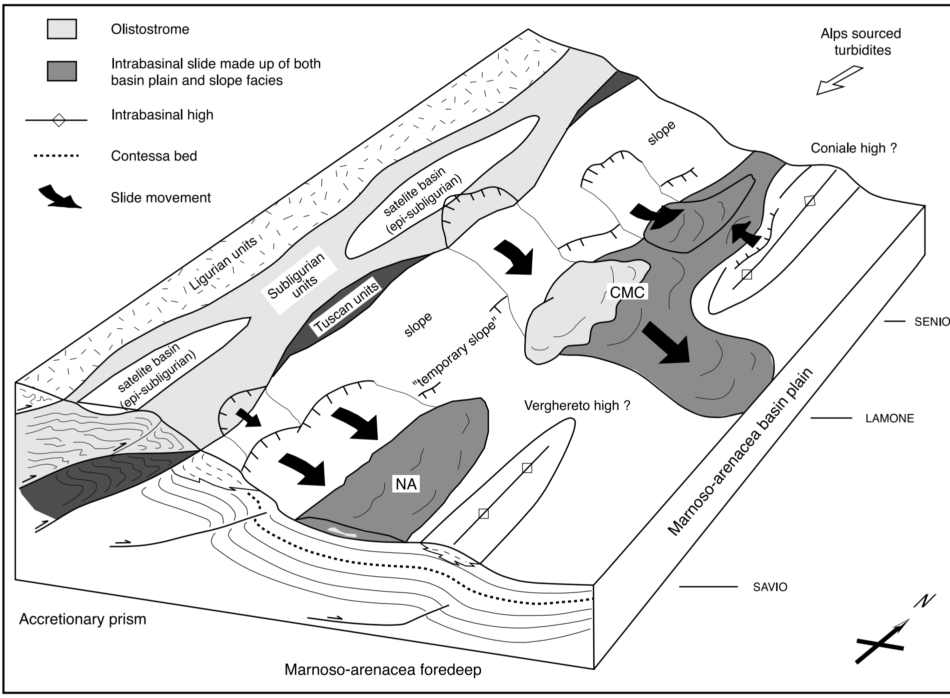

Fig. 10. Interpretative palinspastic block diagram of the CMC slide emplacement (Lucente and Pini, 2002).

area the paleo-basin axis seems to be confined by an intrabasinal high, which may correspond to the one reconstructed by De Jager (1979) and Roveri and others (2002) (Coniale high in fig. 10). The presence of this intrabasinal high is also supported by some complications that can be observed immediately south of section C (fig. 6A). Here the slide body has two distinct, differently verging portions. The lower portion, consisting of basin plain sediments, has large-scale recumbent folds with a well-defined westward vergence (vector cluster b in figs. 5 and 8A); the upper portion of slope sediments shares the general northward vergence of this part of the body. The west-verging folds could be interpreted as representing either a reflection effect of the main landslide body, or as a contribution from another minor landslide from a flank of the Coniale high. The intrabasinal high was not continuous along the basin axis as documented by the spreading of CMC in the Marradi sector.

Basin plain deposits make up the bulk of the CMC. Thus, the CMC was not simply the result of a mass transfer from the slope to the basin, as in the majority of present day and fossil submarine landslides, but a significant amount—at least 15 x 10^9 square meters—of basinal deposits were involved. Their sliding may be the result of a tilting of a basin plain segment (temporary slope, as described by Ricci Lucchi, 1978). However, slope deposits and extrabasinal rocks (olistostrome) are also present in the CMC. The sedimentation of the slope mudstones along the inner margin of the basin was probably favored by the uplift and the tilting of the basin plain, which could be related to the advance, from the southwest, of the Apennine accretionary front and the Ligurian Nappe. The occurrence of the Subligurian and Epiligurian rocks within the olistostrome in the MA basin provides evidence of the proximity of the Ligurian Nappe.

The emplacement of the slope deposits in the Senio valley and the olistostrome in the Lamone valley always occurs on top of the whole slide mass (fig. 6A). There is a clearly marked lateral boundary between the basin plain and the slope sediments; the latter seem to have filled in a depression on the basin plain slide deposits. The olistostrome has clear-cut boundary with the basin plain sediments. The anomalous culmination of failed intrabasinal sediments splitting the extrabasinal component in two lenses in section A2 (fig. 6B) may suggest that the olistostrome dragged the intrabasinal deformed sediments during its emplacement, creating an accumulation at its front, which deformed the base of the olistostrome itself.

Mechanism and Timing of Slope and Temporary Slope Failure

Because of the geometric relationship among the different CMC components, and their relative paleogeographic positions, we believe that the kinematics of the failure nucleation and CMC emplacement follows one of two scenarios:

1) Contemporaneous detachments by independent, synchronous failure events occurred in the internal (southwest) margin of the basin affecting the basin plane (temporary slope), the slope and the front of the Ligurian nappe. If this were the case, the relative position of intrabasinal and extrabasinal components in the slide body could be explained by different sliding distances, with the farthest traveled, the olistostrome, having the highest position being active immediately after the intrabasinal component emplacement.

2) Retrogressive sliding that first produced the detachment of the basin plain sediments in the temporary slope. This process made the slope deposits unstable resulting in failure cut back until the front of the Ligurian nappe began to slide. This interpretation again explains the observed relationships among the different components of the CMC. Moreover, many modern submarine slides appear to be the result of migration of the limit of slope instability upslope, as in the cases of the Hawaiian slide (Moore, 1964), the Agulhas slide complex (Dingle, 1977), the Currituck slide (Prior and others, 1987), the Storegga complex (Bugge and others, 1987), the Gela slide-debris flow (Trincardi and Argnani, 1990).

3) A third scenario of progressive sliding, where more marginal masses (that is, the olistostrome from the Ligurian nappe front) pushed ahead of the slope and, then, the basin plain sediments, cannot be excluded. If this is the case, the detachment of the basin plain sediments would be the effect rather than the cause of the slope sediments and the Ligurian rocks slide. But, it should be noted that progressive slide emplacement fails to explain two facts: a) the northward deflection of the slope lithofacies, b) the absence of any mixing between the different components of the slide body.

Independently from the mechanism of failure and emplacement—contemporaneous, regressive, or progressive—the coeval emplacement of another large submarine landslide, the Nasseto (NA) slide, documents a paroxysmal phase of basin-wide instability of the Serravallian Marnoso-arenacea. The NA body shares a similar composition and thickness with the CMC and has similar mechanism of failure and emplacement (Lucente, ms, 2000). A relevant amount of basin plain deposits occurs in the lower part of the NA body. Thus, in this phase, a significant part of the basin plain becomes a temporary slope (fig. 10) and the slope and the accretionary wedge front become very unstable. This instability is one of the largest events of basin instability in the Marnoso-arenacea foredeep during its sedimentary history from Langhian to Tortonian.

SLIDE ANATOMY AND INTERNAL STRUCTURES

We will describe now the distribution of the structures, taking into account that the CMC body shows two major directions of movement: a northeastward movement of

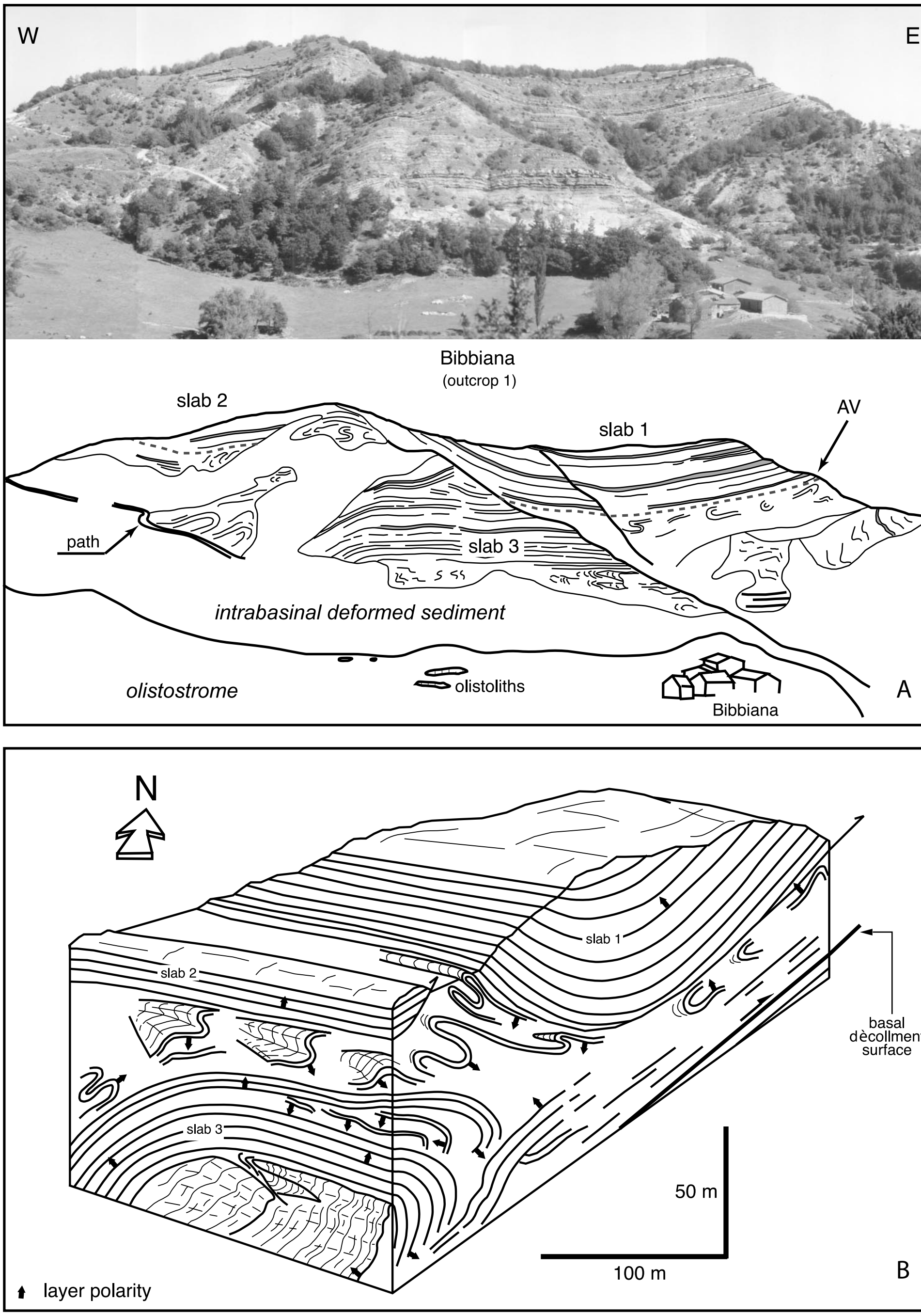

Fig. 11. Panoramic view (A) and interpretations (B) of the Bibbiana outcrop (outcrop 1 in fig. 5) of the CMC slide.

basin plain sediments to the southeast of the Senio River and a northward movement of the basin plain and the slope sediments to the northwest.

Transect Through the Top-Northeast Body

Using the association of structures as a criterion, the CMC can be divided into three different segments, which summarize the distribution of the main deformational features in the basin plain sediments that moved northeastwards (fig. 7).

The southwestern part of the first segment corresponds to the ramp zone (segment a in fig. 7) and is a zone of local accumulation complicated by the olistostrome emplacement. Portions of the slide basin plain deposits, which probably come from the southwest, below the olistostrome, have been accumulated at the front of the first lens of the olistostrome above the steepest SE-NW ramp. The portions are arranged in an antiformal stacking of duplex structures that deform previous slide structures. The ductile behavior of rocks has led to the lack of the bed continuity (stratal disruption).

In the northeastern part of segment a, the Casaglia sector (outcrop 1 in fig. 5), tens of meters thick packages of very poorly deformed beds (slabs, intraformational olistoliths) rest on strongly deformed belts that are several meters thick (fig. 11). The latter are characterized by close to isoclinal folds and by overfolds (fig. 11B). The deformed belts appear to accommodate the relative movement among the different slabs within the slide, acting as shear zones. All the olistolith successions coincide with the stratigraphy beneath the slide body, as shown by the presence of key-beds (see AV bed in figs. 6B and 11A). The olistoliths are usually gently folded (slab 1 and 2 in fig. 11), with the exception of the lowest slab, slab 3, arranged in a large scale, steeply inclined tight fold, with an overturned north-eastern limb.

The general attitude of the slabs and the arrangement and vergence of folds in the underlying shear zones suggest that the movement of the highest slabs (slabs 1 and 2) occurred in the same direction of the general slide movement, but with the geometry of listric normal faulting. No significant folds developed below and at the front of the lowest slab (slab 3); its lower, overturned limb rests on right-side-up bed packages. These packages form the base of the slide and are complicated by some small-scale overthrusts, which seem to share the same northeastward vergence of folds. The emplacement of the second lens of the olistostrome at the front of the antiformal culmination may explain the tight fold of slab 3 in the Bibbiana outcrop (fig. 11) and may also explain the vertical attitude of the folds in the shear zones between slabs 2 and 3.

Large preserved slabs of the sedimentary succession (fig. 12, outcrop 2 in fig. 5), moving independently, also characterize the second intermediate segment (b in fig. 7). The relative movements of slabs are accommodated by meter-thick shear zones, which have isoclinal folds, close asymmetric folds, overfolds, thinning of beds and asymmetric pinch-and-swell structures.

The slabs contain virtually undeformed beds, with the exception of some large recumbent folds and boudinage and truncation of beds by millimeter- to centimeter-thick mesoscopic ductile faults. In the upper part of the body, some of the ductile faults offset the beds, showing a normal displacement. Other ductile faults cause the lateral and vertical overlapping of portions of the succession and merge top and bottom in layer-parallel faults (fig. 13). The competent beds in each of the superposed elements are truncated by the ductile faults and stretched and folded in a sigmoidal-shape. The general arrangement of these elements and the distribution of the ductile faults resemble a duplex system. The lack of repetition of stratigraphic succession and the geometry of thinning and bending of the more competent beds suggest that the structures are related to layer-parallel extension and therefore to an extensional

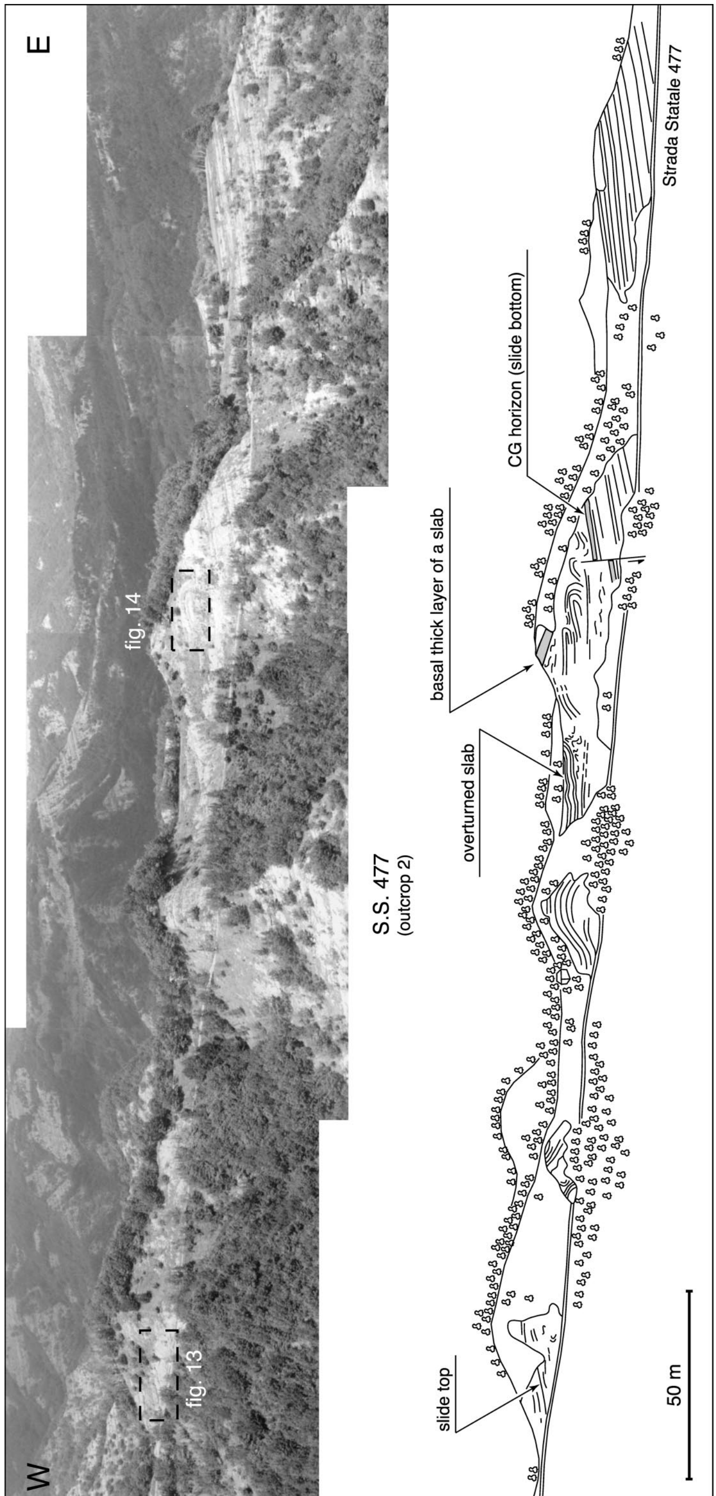

Fig. 12. Photograph and line drawing of a natural cross-section of CMC slide, exposed in road cuts along state route 477 (SS477) (outcrop 2). Location shown in figure 5.

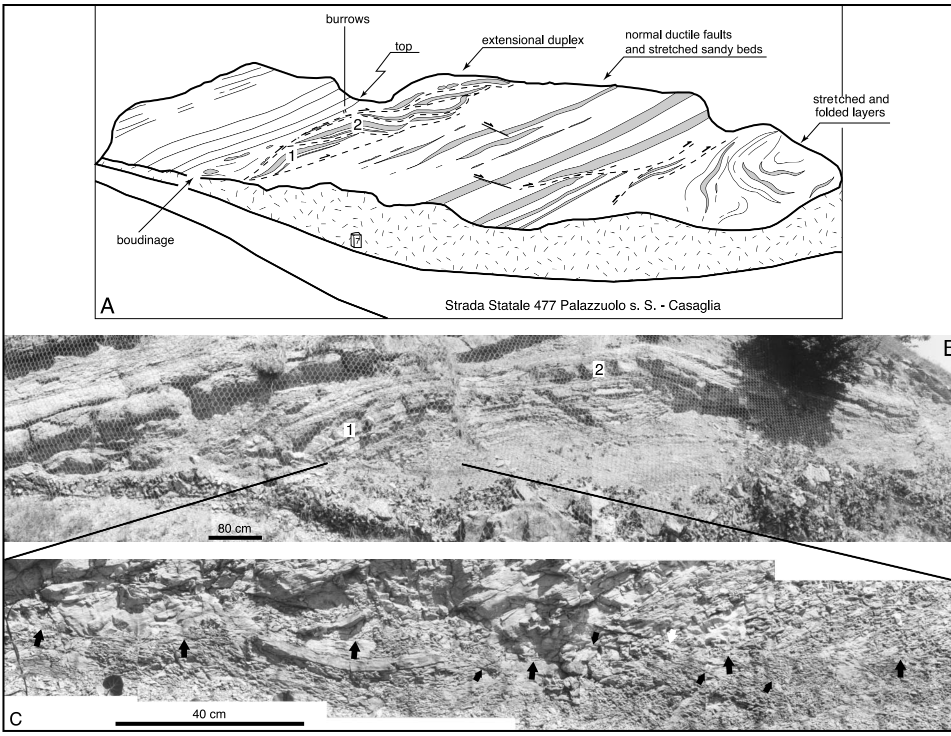

Fig. 13. (A) The uppermost slide portion of outcrop 2, along SS477. Location on figure 12. (B) Photo mosaic showing a part of the duplex system; numbers 1 and 2 refer to two distinct beds located on figure 13A. C) Detail of the lower surface of the duplex. Thick and smaller arrows mark the sole thrust and some minor, P and R planes respectively.

duplex system. Steen and Andresen (1997) described such kinds of structures in slump-deformed olistoliths of the Kalvag mélange.

Large-scale recumbent isoclinal folds are common features of the lowest part of the body. A good example is the right part of the cross-section above the slump detachment surface in outcrop 2 (fig. 14). Two isoclinal anticlines seem to affect the same bed; the associated syncline is not evident and could have been truncated by a shear surface. Thin-bedded turbidites are severely deformed to the point of stratal disruption at the front of the anticlinal folds (fig. 14B). Stratal disruption develops a characteristic triangular cross-section and increases progressively toward the anticline front (right to left in fig. 14B) by imbricate stacking of bed chunks. The stacking pattern verges in the opposite direction to the anticline and the related folding. These triangular zones may be the result of the expulsion of part of the stratigraphic succession in order to accommodate the forward movement of the anticline fronts and the elision of the flanks of the synclines. The disharmonic folding and bed stacking may be related to layer-parallel shortening triggered by the lateral expulsion. The stacked blocks are subsequently folded in an opposite direction by the dragging related to the advancing anticlines.

The third, more distal segment (see c in fig. 7) corresponds to the Marradi sector. The basin plain deposits involved in the slump body are mainly thin-bedded turbidites, with a few thicker arenaceous beds. Here ductile faults again enable the relative movement of different masses. In this segment (c in fig. 7) the faults are more frequent

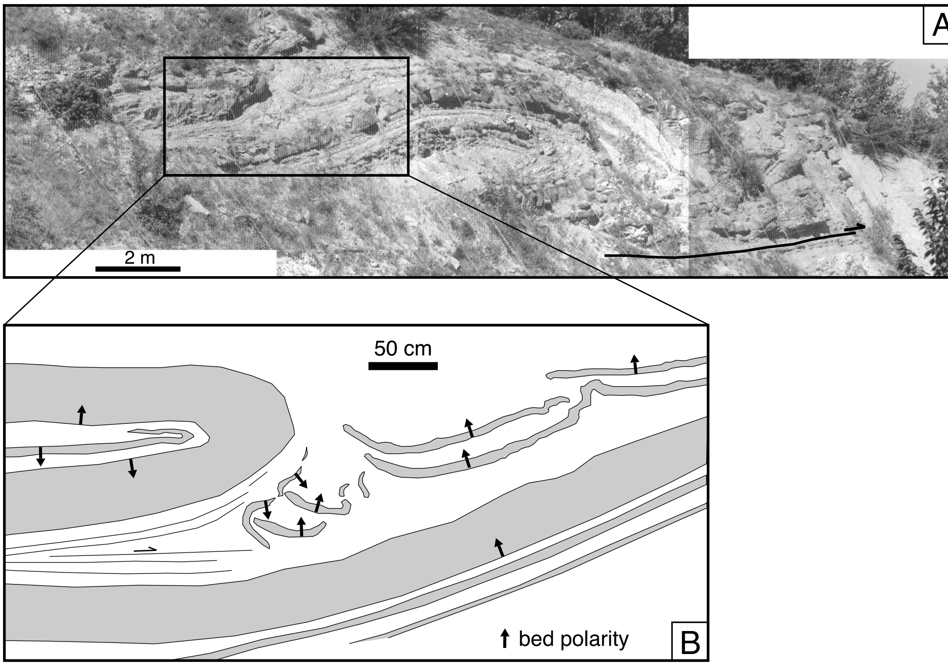

Fig. 14. (A) Photo mosaic showing the recumbent folds close to the slide bottom in the eastern edge of outcrop 2 (SS477). Location shown on figure 12. (B) Detail of the pervasive deformation at the front of a recumbent isoclinal anticline.

and the beds are much more deformed with the lateral continuity reduced to a few meters.

The upper part of the slide is again dominated by listric normal faults, which are associated with overfolds and with truncation of the bed packages. Bed stretching, pinch-and-swell and lenticular boudinage are also common features. The lower part, on the other hand, is characterized by strong contraction rather than extension. The main structures are isoclinal recumbent folds, cascade folds, asymmetric folds, box folds, isolated fold hinges and stacking of boudins (figs. 15 and 16). The folds show disharmonic behavior and the picture is complicated by parasitic, asymmetric folds (fig. 15A). Type 2 and 3 interference patterns derive from refolding of folds. Layer-parallel extension is restricted to pinch-and-swell and boudinage of beds (figs. 15 and 16). The vergence of folds and stacking of boudins is another characteristic special to this sector. Generally, isoclinal, cascade and asymmetric folds and stacking all verge toward the northeastern quadrant suggesting a general flow in a dominant simple shear regime. On the other hand, opposite verging folds and block stacking (outcrops 4 and 5 in figs. 15 and 16) are not uncommon. These structures may be related to pure shear deformation by buckling, or to the superposition of two phases of movement in opposite directions. The first interpretation is supported by the presence of box folds (outcrop 4 in fig. 15B).

Transect Through the Top-North Body

As mentioned previously, the slope intrabasinal component and the underlying basin plain sediments have a clearly northwards direction of movement, in the northern part of the Casaglia sector (Senio-Santerno). The upper part, made up of the

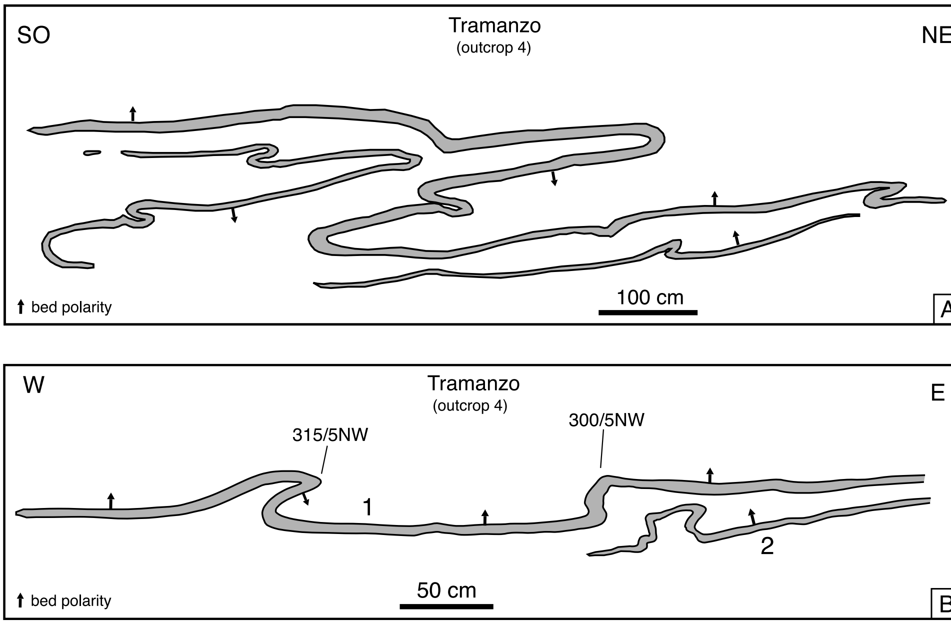

Fig. 15. (A) Recumbent, disharmonic isoclinal fold deforms a couplet of thin turbidites in the Tramanzo outcrop (outcrop 4, Marradi sector). Location shown on figure 5. (B) Two opposite verging asymmetric folds involve bed 1. In contrast, the lower bed (bed 2) has a box fold.

slope component, has a system of normal listric faults (see the Campanara outcrop in fig. 17). The upper faults systematically truncate the lower ones in turn; this relationship resembles a mechanism of regressive detachment. The relatively less deformed slabs bounded by the normal faults are very thin, just a few meters thick.

Fault movements are concentrated in millimeter-thick horizons (mesoscopic ductile faults). Boudinage of prolate ellipsoidal shape develops in the zone of major strain close to the shear surfaces and may be a consequence of the flattening (pure shear) related to the general shear caused by ductile faults (fig. 18). Asymmetric fold trains are often associated to ductile faults and boudinage and seem to propagate to the slabs. Folds have curvilinear axial surfaces that may have been rotated by progressive simple shear (fig. 18).

A horizon of pervasive deformation occurs immediately below the upper, extensional zone and marks the transition to the underlying basin-plain sediments. The deformation consists of severe boudinage, asymmetric folds and stacking of boudins. Folds and stacking indicate a certain amount of simple shear; the flattening-type boudinage may also be related to the simple shear. Deformation is generally high in this belt and the continuity of the beds is no greater than a few meters. The deformation becomes even greater in some parts causing the dispersion of isolated boudins and fold hinges in a sand-silt matrix.

STRAIN PARTITION AND MECHANICAL INTERPRETATION

A submarine landslide is considered to consist primarily of a contractional zone (accumulation) at the toe and an extensional zone (detachment) in the head. The two zones are in some cases separated by a translation zone, which is the zone of layer parallel gliding (for example Lewis, 1971; Dingle, 1977; Alvarez and others, 1985; Webb and Cooper, 1988; Trincardi and Argnani, 1990; Moore and Shannon, 1991). In

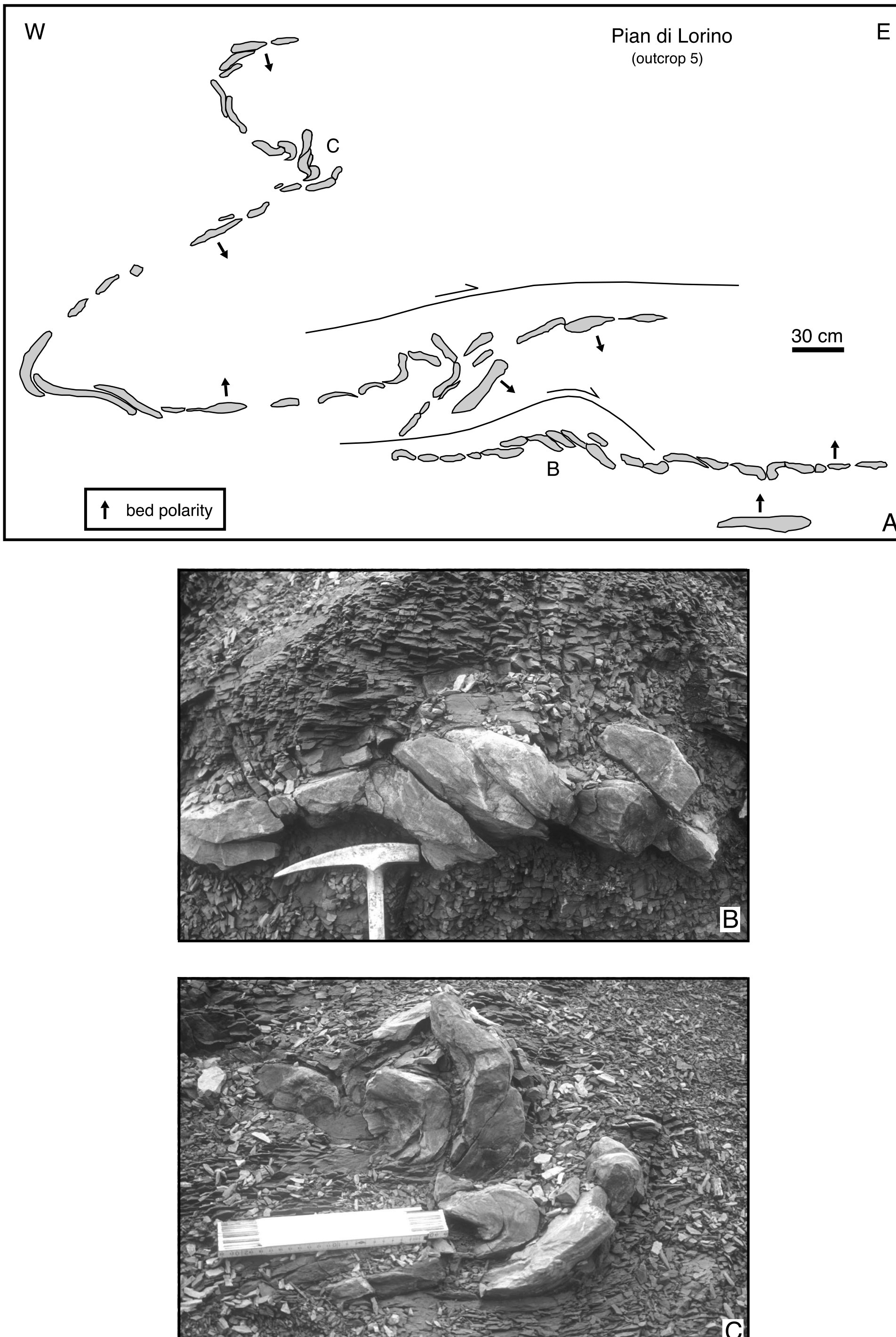

Fig. 16. (A) A thin, arenaceous bed affected by boudinage, block stacking and folding and complicated by low angle shear surfaces in the Pian di Lorino outcrop (outcrop 5). The different structures may be interpreted as the result of a progressive deformation. (B,C) Details of block stacking (B) and of folded block stacking and isolated fold hinges (C). Location shown in figure 5.

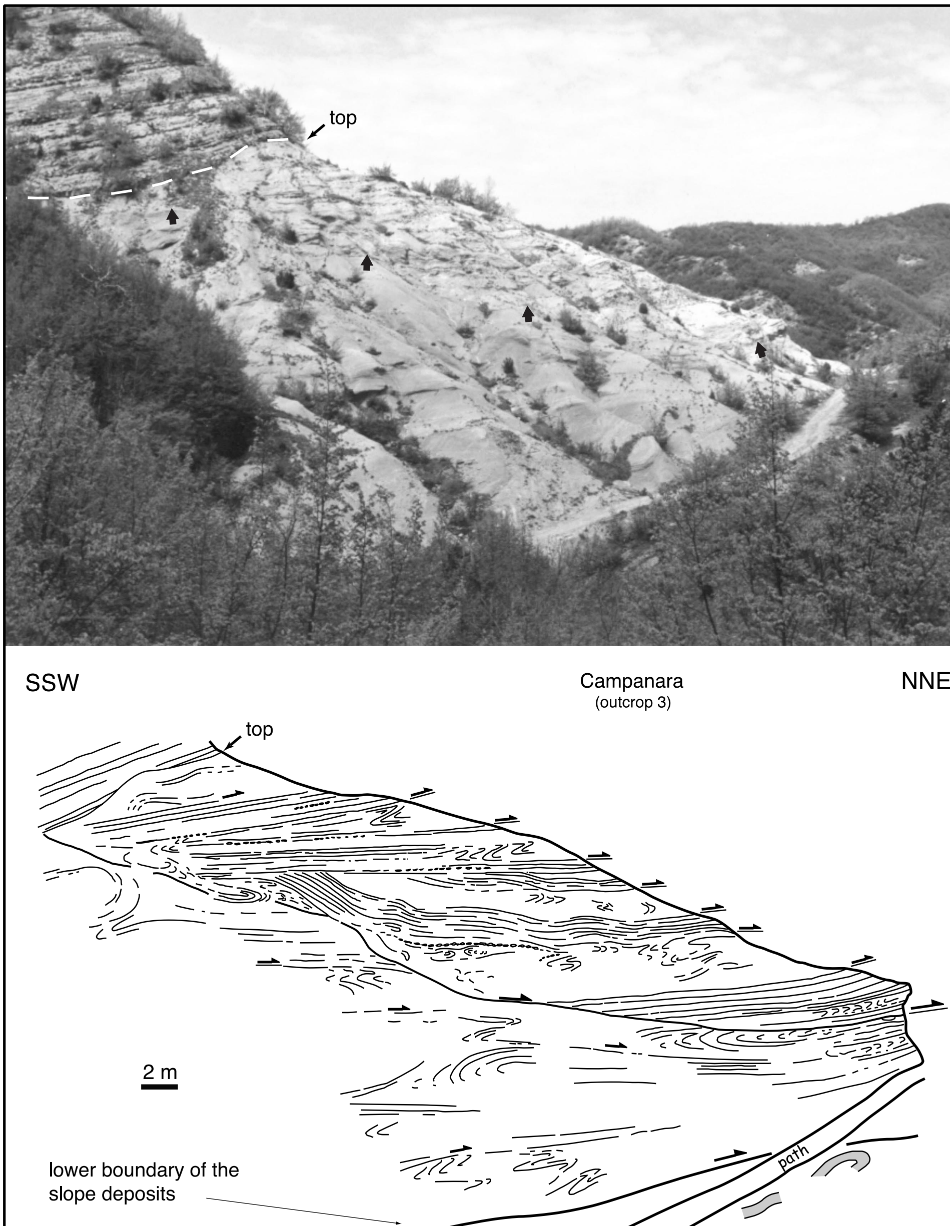

Fig. 17. Panoramic view and sketch of a part of the Campanara outcrop (outcrop 3, location on fig. 5) showing the distribution of listric, normal faults of different hierarchy, separating thin-bedded packages of the slope sediments (see figs. 5 and 6).

the CMC body the head zone has not been preserved at all. The translation zone, the contractional ramp zone and the lobe are preserved in the Casaglia area below the olistostrome, in the Bibbiana area at the front of the olistostrome and in the other outcrops respectively. The overall geometry of the CMC taken as a whole is different from geometry required for the classic slide-slump models, but has several points of

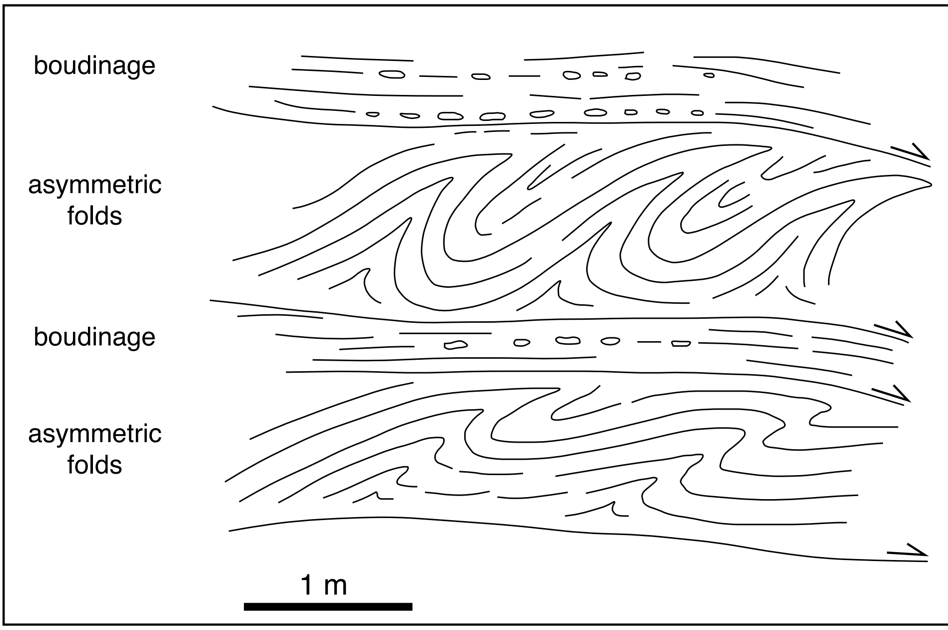

Fig. 18. Detail of figure 17 showing the vertical alternation of asymmetric, folded bands and boudinage-affected bands, which develop in relation to the discrete shear surfaces (listric, normal faults).

contact with present-day examples of submarine landslides. First at all, the CMC is dominated by slabs of barely deformed succession moving differentially (olistoliths). The differential movement of the slabs is accommodated along zones of major pervasive deformation, that is along shear surfaces and zones. The larger olistoliths are concentrated at the slide top, giving rise to a hummocky upper surface as documented in many present-day examples by seismic data (for example, Jacobi, 1976; Trincardi and Normark, 1989). The onlap and lens-shape of the turbiditic bed outline the gentle undulations of the slide top on the slabs. Larger slabs at the landslide top may imply a high competence of the flow, and/or their preservation related to the lower condition of strain.

The dimensions and frequency of slabs decrease as the slide sediments move northeastwards and northwards in lobe. The degree of internal deformation also increases as the slabs become smaller. In the southwest-northeast transect, moving from segment a to c (fig. 7), the ratio of non-deformed sediments to deformed sediments increases dramatically towards the Marradi sector where internal deformation is so pervasive that only a few relatively undeformed slabs (consisting of continuous beds several meters long) are to be found (Pian di Lorino, fig. 16). This attitude resembles that of many present-day large submarine mass-wasting deposits (Jacobi, 1976; Prior and others, 1984; Normark and Gutmacher, 1988).

The decrease in size of the slabs and their increasing deformation from southwest to northeast may be linked to the evolution of the flow towards the distal zone (see explicative examples in Gawthorpe and Clemmey, 1985). However, this interpretation does not take into account the fact that the increasing deformation is associated with changes in deformational structures. Figure 19 shows the distribution of strain in a southwest to northeast transect from the distribution of the deformational structures outlined in figure 7. The ramp zone is an accumulation zone, characterized by

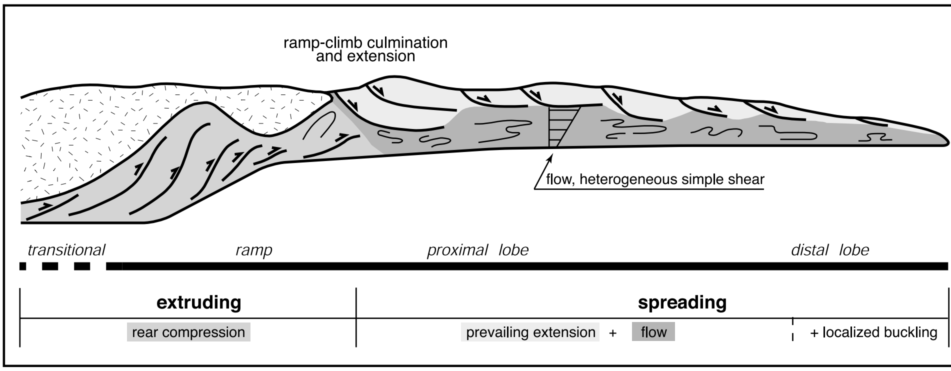

Fig. 19. Interpretative diagram of the SW-NE transect of figure 7.

contractional structures, such as systems of duplex, thrusts, and folds causing vertical thickening and horizontal shortening. The antiformal stacking of duplexes and the steeply inclined folds suggest a rear contraction promoted by the olistostrome emplacement, which dragged and pushed the slide sediments in front of it. The very thick, anomalous depth of the extensional sliding of Bibbiana can be related to a culmination caused by the extrusion of sediments above and in front of the ramp zone.

The lobe is dominated by a well-defined partition of deformation: listric normal faults and extensive duplexes are concentrated in the highest part in which vertical thinning and horizontal stretching dominate. Simple shear occurs occasionally only along ductile faults and in shear zone associated to the relative movement of discrete masses. Normal faults have been already described as leading structures in the toe zone of some ancient submarine landslide, but they are thought to be confined at the tip of slide bottom ramps (Gawthorpe and Clemmey, 1985). The upper part of the CMC lobe shares the same association of normal faults that is evident in seismic and acoustic profiles in all the upper part of present-day submarine landslides (for example, Gardner and others, 1999).

The lowest part of the lobe has flow structures, such as recumbent isoclinal folds, asymmetric folds and thrusts corresponding to simple shear deformation. Deformation by heterogeneous simple shear increases upward from the landslide base to the boundary between the lower and the upper part of the body and can be related to an increase in differential flow (fig. 19). This deformation may be caused by the dragging of the downslope movement of the upper part spreading in the basin plain.

The spreading of the upper part can be confirmed by the style of deformation. In the more proximal part (segment a in fig. 7) movement occurs along a predominant direction and the typical structures are extensional duplexes, constriction-type boudinage and non-cylindrical folds with a moderate dispersion of the fold axes. Movement develops in all directions of the plane in the more external part of the lobe (Marradi sector). It is associated with flattening-type boudinage and with an increase of fold axes dispersion.

The south-to-north verging branch of the lobe is complicated by the presence of the slope deposits. The slabs are always small (tens of cm to a few meters thick) and pervasively deformed by folds. Slabs disappear in the lowest part of the slope deposits close to the contact with the basin plain deposits due to a more pervasive and diffused deformation by progressive simple shear and flattening. The zone of pervasive deformation can be attributed to the mechanical uncoupling and the movements of the basin plain and the slope sediments in relation to each other. In this interpretation the zone of pervasive deformation is a large shear zone accommodating the differential displacement due to the faster movement of slope deposits than the underlying basin plain deposits. This movement is caused by the extension on the upper part and the simple shear flow in the lower part.

Such a kinematic model of the slide based on the field observations shows some analogies with nappe emplacement. The slide geometry resembles that of an extruding-spreading nappe (Merle, 1998) because of the characteristic ramp-and-flat attitude in the transition from the ramp to the lobe and the distribution of internal strain.

The overall geometry of the body suggests the mechanism of gravity spreading proposed here rather than gliding and/or ductile gliding (Merle, 1998; Schultz-Ela, 2001). Gliding can occur along a slope dipping in the same direction as the slide movement, with the bottom and top having the same profile and dip. In the case of the CMC distal lobe, the top dips toward the leading edge of the slide but the bottom is flat, parallel to the normal bedding and even dips at very low angle in the opposite direction to the slide emplacement.

In the Marradi area the local opposite vergence of structures and the box folds may be related to a certain amount of buckling. Buckling in slide bodies has been first pointed out by Woodcock (1976) and Farrel and Eaton (1987). These authors stated that buckling is the initial stage of fold nucleation, which can later develop into asymmetric folds and isoclinal recumbent folds. In the case of CMC, buckling is only visible in the more distal (frontal) part of the body and develops together with isoclinal and asymmetric folds in an advanced stage of deformation. In our interpretation, therefore, buckle folds can be related to a lateral/frontal confinement related to stopping of the sliding mass due to topographic constraints.

SUMMARY ON STRATAL DISRUPTION WITHIN THE CMC BODY

Attributing stratal disruption to submarine mass wasting processes or to shallow-level tectonic deformation is one of the most hotly debated problems in the geology of mélanges and accretionary wedges (Moore, 1973; Page, 1978; Naylor, 1982; Raymond, 1984; Underwood, 1984; Cowan, 1985; Brandon, 1989; Hanamura and Ogawa, 1993; Orange and Underwood, 1995; Pini, 1999; Yamamoto and others, 2000). This discussion focuses on the early stages of sediment-rich mélange formation (Brandon, 1989; Hanamura and Ogawa, 1993) and concerns the deformation of non-consolidated, water-rich sediments (for example, Elliot and Williams, 1988; Maltman, 1994a).

The perspective of stratal disruption related to mass wasting processes has been rejected for the basin plain turbidites and considered exclusive of slope deposits (Underwood, 1984). This conclusion should be reconsidered for the landslide bodies of the Marnoso-arenacea Formation because most of the deformation is in basin-plain turbiditic deposits. However, the flat-ramp-flat, thrust style of the CMC bottom is similar to very shallow-level deformation in accretionary wedges (for example Yamamoto and others, 2000). Depth continuation may be considered a distinctive feature of shallow-level thrusts versus submarine landslides assuming that slides affect only the very shallow part of the sedimentary pile, whilst the thrusts cut deeper into the stratigraphic succession, until they reach a deeper decollement level. The different vergence of structures related to slides and thrusts may also be taken as another distinguishing feature (Yamamoto and others, 2000). This study on the slides in the Marnoso-arenacea Formation show that care should be taken in using the depth of cut-off of the footwall stratigraphic succession as proof of the tectonic origin of thrust-like structures given that the ramp of the CMC body is more than 200 meters deep. The results also indicate that in a foredeep basin the direction of movement in the same body changes according to the lateral and frontal confinements provided by basin topography and intrabasinal highs.

The deformational structures within CMC are similar to those of shallow-level tectonics and it is quite difficult to differentiate a single structure outside the general context (Byrne, 1984; Maltman, 1994b, and references therein). Moreover, the finite state of the stratal disruption in CMC is the result of progressive deformation, as shown by the superposition of differential incremental stage of deformation in several outcrops (see, for example, figs. 14, 16 and 18). However, progressive deformation is one of the leading processes in slumping (Gosh and Sengupta, 1984; Farrell and Eaton, 1987, 1988), and in tectonically disrupted mélanges and broken formations (Hibbard and Karig, 1987; Cowan, 1990; Kano and others, 1991; Jeanbourquin, 2000; Cowan and Pini, 2001; Vannucchi and Bettelli, 2002). Notwithstanding these analogies, the distribution of structures and the down-flow change in degree and style of stratal disruption have all some special characters that may help to differentiate between mass-wasting and shallow-level tectonic processes. Below we consider the association of deformational structures, the degree of deformation and stratal disruption, and the change in structure morphology.

1) Association of Deformational Structures

Inside the CMC, normal faults and extensional duplexes dominate the upper part of the lobe (figs. 7 and 19) and cause thinning of beds and boudinage (fig. 13). Other structures, such as folds, overfolds and boudinage, are confined to shear zones accommodating the relative movement of normal fault limbs and/or slabs (figs. 11, 13 and 17). It is worthy to note that normal faults commonly are minor and subordinate elements in tectonically disrupted units (see, for example, Pettinga, 1982; Onishi and Kimura, 1995; Jeanbourquin, 2000).

The lower part of CMC is dominated by folds and low angle thrusts. This association is in part similar to the association of structure in the shallow-level of tectonic deformation (Yamamoto and others, 2000), however the way that the morphology of structures changes along the transport direction is different (fig. 7). Moreover, duplex structures deforming bed packages, which are an important component of shallow-level deformation in accretionary wedge (Hanamura and Ogawa, 1993; Kimura and Hori, 1993; Hirono and Ogawa, 1998), have not been observed in the slide; they only occur where the olistostrome pushed ahead the basin plain sediments. Duplex structures involving a single boudinaged bed (block stacking) are present in the Marradi sector and are morphologically similar to structures in mélanges (Kano and others, 1991).

2) Degree of Deformation and Stratal Disruption

The degree of stratal disruption increases both from top to bottom and from the proximal to the distal zone of the landslide body, with the exception of the ramp zone, which has a large amount of highly disrupted bed packages due to the olistostrome emplacement.

In the proximal parts of the CMC lobe (segments a and b in fig. 7), slabs are poorly deformed and the lateral continuity of beds is completely preserved, even when they are folded. Stratal disruption is concentrated in the shear zones and in zones of lateral extrusion at the front of anticlines (fig. 14). In some tectonically disrupted mélanges and broken formations, the disruption is predominantly concentrated in fault zones. The progressive decrease of thrust spacing, the out-of-sequence nucleation of younger thrusts, and the thickening of fault zones makes the stratal disruption extend to a larger portion of rocks (Pettinga, 1982; Moore and Byrne, 1987; Kimura and Hori, 1993; Hirono and Ogawa, 1998). Whereas, in the CMC body the stratal disruption degree increases as far as the shear zones become thinner (segment c in fig. 7). At the same time, stratal disruption becomes diffuse in the slabs by normal faulting and folding. In detail, folds increase in disharmony from concentric to intrafolial folds and fold limbs are truncated or subjected to boudinage, up to the isolated, recumbent isoclinal folds related to the complete desegregation of slabs. In this external part of the body, superimposition of folds contributes to stratal disruption.

3) Change in Structure Morphology

The last group of possible diagnostic criteria is the change in style of single structures down-flow. The shape of the fold axes and hinge zones changes from merely non-cylindrical folds to sheath-like folds. The boudinage evolution is coupled with the evolution of folds, from shear fracture boudinage in the more proximal part, to lenticular boudinage and to isolated boudins. The shape of boudins also change from prolate, constriction-type in the internal part of the lobe to oblate, flattening-type in the more external part of Marradi sector, where spreading prevails.

Box folds and opposite verging folds in distal slide sector (Marradi) suggest that a certain amount of buckling occur when the body is slowed down by lateral and frontal morphologic constraints. The association of simple shear and buckling is again a distinctive criterion, as this association seems to be uncommon in accretionary wedges.

CONCLUSIONS

The bodies of mass wasting deposits in the Marnoso-arenacea Formation are some of the best-exposed fossil examples in a foredeep succession. This paper focuses on the largest of these bodies, the Casaglia Monte della Colonna body, which covers more than three hundred square kilometers and has some unusual features. First, the body is made up of intrabasinal slope sediments, basin plain sediments and extrabasinal components from the paleo-Apennines accretionary wedge (olistostrome). The basin plain sediments dominate and suggest a temporary instability of a significant part of the foredeep basin plain. The more proximal part of the slide body preserved nowadays is a ramp zone cutting off more than 200 meters of the non-deformed stratigraphic succession of basin plain sediments. Immediately downstream of this zone, a lobe, made up of basin plain and slope sediments, spilled over onto the basin floor with flat contacts connected by small-scale ramps. This is the widest part of the body and is a new observation as far as the fossil slides are concerned.

The ramp zone has imbricate stacks of duplexes that deformed previous slide structures giving strong stratal disruption. This fact might be related to the olistostrome emplacement. A longitudinal, vertical strain partition is visible in the lobe. Extensional duplexes, listric normal faults and boudinage dominate the upper part of the body, whereas recumbent isoclinal folds, asymmetric folds and low-angle thrusts characterize the lower part. This partition of the strain and the top and bottom attitude of the landslide all suggest that gravity spreading may be the emplacement mechanism. If we combine the ramp and the lobe, the slide emplacement kinematics appears to be similar to that of extruding-spreading nappes.

Once triggered by the spreading, the landslide lobe moved in two directions, with a northward deflection of the westerly part and a northeastern movement of the easterly portion. This interpretation derives from the study of fold axes and vergence, from the distribution of normal faults, mono-directional boudinage and thrusts, and from other independent data of stratigraphy and basin analysis (Lucente, 2002; Roveri and others, 2002). The lobe deflection has been caused by an intrabasinal high, which may correspond to the Coniale high reconstructed by De Jager (1979) and Roveri and others (2002). The more proximal part of the easterly lobe portion shows a higher component of spreading suggested by dispersion of fold axes and by flattening boudinage. This part of the lobe was therefore free to spill over onto the basin plain in a wide area more to the northeast of the Coniale high (Marradi sector). Thus, the intrabasinal high should not have a lateral continuity along the basin axis. Occasionally, localized counter-vergence and buckling were probably related to some obstacle that confined the body and slowed down its movement. These structures are particularly evident in the distal part of the Marradi sector, where the lobe was probably slowed down by a basin-slope ramp. Thus, the CMC slide shows that the study of submarine slides can make a substantial contribution to a reconstruction of the palaeogeographic setting of the sedimentary basin.

At a closer view, shear zones and shear surfaces accommodate differential movements of discrete masses in both the upper and the lower part of the lobe. The differential movements cause simple shear and flattening of the shear zones, which are responsible for association of different structures. From proximal to distal part of the lobe, the structures in shear zones evolve from isoclinal folds, overfolds and thrusts to smaller scale isoclinal folds and overfolds, thinning of beds and asymmetric pinch-and-swell structures. This downflow evolution of the shear zone is coupled with the progressive increase of the internal deformation of slabs up to their desegregation.

The partition of strain in the entire body, that is extension on top and heterogeneous simple shear on bottom, and the smaller scale partition of deformation in independently moving masses make the relationships between extensional and contractional structures much more complex and frequent than thought in previous papers on mass wasting deposits. Zones of contraction and extension exist together in various degrees all along the body.

Moreover, the partition of deformation at different scales causes different styles of stratal disruption. Different styles are related to different combinations of progressive simple shear, layer-parallel extension and layer parallel shortening. The different styles and degree of stratal disruption occur in well-defined parts of the landslide body and are intimately connected to the evolution and emplacement of mass wasting deposits. Although the single structures are similar to those observed in accretionary wedges, the distribution and association of structures might be diagnostic of the distinction between mass wasting processes and shallow-level tectonic deformation.

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