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Mélanges and mélange forming processes: A historical overview and new concepts

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Abstract

Mélanges represent a significant component of collisional and accretionary orogenic belts, and occur widely around the world. Since its first introduction and use, the term has evolved to cover both processes (tectonic, sedimentary and diapiric) and tectonic settings of mélange formation. The meaning and significance of various terms referring to the origin of ‘block-in-matrix chaotic rocks’ are still subject to debate. This study presents a historical overview of the evolving mélange concept, and investigates the relationships between mélange types and their tectonic settings of formation. We investigate the contribution of mass-transport versus contractional deformation processes at the onset of mélange formation and throughout the evolution of different mélange types, and the nature of the continuum and transition from broken formations to true tectonic mélanges. Mélange is a mappable chaotic body of mixed rocks with a block-in-matrix fabric whose internal structure and evolution are intimately linked to the structural, sedimentary, magmatic, and metamorphic processes attending its origin. Based on a comparative analysis of exhumed, ancient on-land mélanges and modern tectonic environments, where mélange-forming processes are at work, such units are classified into those related to extensional tectonics, passive margin evolution, strike-slip tectonics, subduction zones, collisional tectonics, and intra-continental deformation. Sedimentation and contractional deformation contribute significantly to mélange formation in all these tectonic environments, although the internal structure of deposits is strongly controlled and overprinted by processes that prevail during the last stages of mélange formation in a single tectonic setting. Tectonic mélanges are commonly subordinate to broken formations, and are restricted to narrow, elongated to coalescent fault zones, large-scale fault zones, and plate boundaries.

Keywords: mélanges; broken formations; olistostromes; mud diapirs; subduction processes; obduction.

Introduction

Mélanges (Fr. – mixture) are mappable units or bodies of mixed rocks including blocks of different ages and origin, commonly embedded in an argillitic, sandy or ophiolitic matrix, or more rarely in a carbonatic, evaporitic or volcanic matrix (block-in-matrix fabric). Reflecting the lack of internal continuity of strata and/or contacts due to high stratal disruption, these units are interpreted as “chaotic”. Mélanges represent a significant component of many collisional- and accretionary-type orogenic belts. Their internal fabric is a result of different processes, such as tectonic disruption, mass transport, diapirism and fluid expulsion, glacial shear and loading, and gas-hydrate generation, and their interplay. These processes commonly occur at shallow crustal depths in the presence of excessive interstitial fluid pressure, which facilitates brittle and mesoscopically ductile deformation, commonly associated with hydrofracturing (i.e. Talbot and von Brunn, 1989; Maltman, 1994; Maltman and Bolton, 2003); the latter strongly affects under-consolidated rocks (Camerlenghi and Pini, 2009). Cataclastic deformation of already cemented rocks, also with an important role of fluid overpressure and hydrofracturing, and ductile deformation due to inter-intra crystalline plasticity, are involved in those mélanges that form in deeper crustal levels within various tectonic settings (see, e.g., Wilson et al., 1989; Polat and Casey, 1995; Kusky and Bradley, 1999; Jongens et al., 2003; Ujiie et al., 2007; Federico et al., 2007).

In light of this general definition, chaotic rock bodies with different characteristics have been described around the world as mélanges. All these bodies lie within a continuum (*sensu* Raymond, 1984), ranging from originally coherent stratigraphic successions, to different degrees of stratal disruption, and finally to a totally chaotic block-in-matrix fabric totally lacking stratal continuity. Here, as stated by Raymond (1984), agreement on the meaning of the term “mélange” ends. In fact, the mélange term often substitutes or is confused with traditional terms, such as *wildflysch*, *schistes à blocs*, *argille scagliose* (scaly clays), olistostromes, olistolith field, megabreccias, agglomerates, which mostly refer to block-in-matrix rocks associated with sedimentary, tectono-sedimentary and/or diapiric origins (see Camerlenghi and Pini, 2009 and references therein). Many of these terms (i.e., *wildflysch*, *schistes à blocs*, olistolith field, megabreccias) were derived from Alpine concepts, and descriptions of the block-in-matrix rocks largely come from the supra Helvetic, Ultrahelvetic, and Penninic domains of the Western Alps and Prealps of France and Switzerland (see below). Following the Penrose conference on mélanges (1978), Silver and Beutner (1980) suggested that in a complex scenario in which the term mélange is applied, it should be used only in a descriptive way, accompanied by adjectives (tectonic, sedimentary, diapiric) indicating the relative mélange-forming processes.

In this paper, we focus on the description of different mélange-forming processes, and present a historical overview and new concepts regarding mélange types. We use only a few of numerous excellent examples of mélange occurrences from around the world. Our observations are derived from many years of detailed field studies on chaotic rock bodies in different orogenic belts. We provide a new all-inclusive subdivision and classification of mélanges distinguished on the basis of tectonic setting(s) and processes of formation. However we are certain that, in addition to the six types of

mélanges described in this synthesis, other varieties of chaotic rock bodies can be added to improve this subdivision and classification.

Some considerations about the terms “sedimentary, tectonic and diapiric”

In the following parts of this paper, we refer frequently to sedimentary, tectonic and/or diapiric processes. These terms can be ambiguous, since they refer to processes that commonly overlap, and since their original definition may have changed over time due to the progressive increase of the geological knowledge (see Alonso et al., 2006). The adjective “sedimentary” in this paper refers to the all processes of slope failure, sediment transport and deposition, with a particular attention given to the *en-mass* gravitational transport and deposition of both (non-consolidated) sediments and rocks (mass-wasting or mass-transport processes and their products, that is mass-wasting or mass-transport bodies). We include in this description also the products of *in situ* and/or very moderate translation of gravity-induced deformational processes along basin slope (e.g. creeping, gravitational spreading, etc.). These processes are currently considered as part of the term “slope tectonics”.

We follow this convention to differentiate clearly between those processes that are related to locally acting gravity and the ones that are related to contractional tectonics. In contractional deformation, the differential movement of masses leads to the systematic repetition of originally conterminous stratigraphic sequences with a result of horizontal shortening and vertical thickening. We realize that in modern conception of accretionary wedges and collisional orogens the influence of gravity is essential in controlling the wedge shape and its kinematics, and that nappes and thrust sheets are driven in part by gravity. However, in these cases the gravitational forces act at large scales, and these factors and structures fall in the definition given above.

We also focus on *mélange* occurrences associated with mud or shale diapirism. We are not referring to the classic salt diapirism, which is the driving mechanism for halokinesis. Mud and shale diapirism is a well-known and established process related to fluid overpressure that leads to the ascent of mostly muddy rocks from deep, overpressured sources toward the surface, piercing through the sedimentary column (mud diapirs) and reaching the seafloor or the topographic surface. This diapiric event commonly results in “erupting” mud ejecta (mud volcanoes).

Historical overview of the *mélange* concept

The first definition of *mélange* and the *mélange* concept, as we know it today, were introduced by Edward Greenly (1919) in describing a tectonically disrupted and internally strained phyllite-sandstone succession of Anglesey (North Wales) in England. He coined the new term to differentiate this unit of mixed rocks formed by tectonic processes from the *wildflysch*, formed by sedimentary-gravitative processes, which was largely described in the Alps after Kaufmann (in Studer, 1872; Kaufmann, 1886).

Wildflysch: the first Alpine concept of *mélange*

As reported by Şengör (see Şengör, 2003 and references therein), the concept of *mélange* was first originally discovered in the Alps (Table 1) through the observation of exotic blocks in the “Nagelfluh” conglomerates (Studer, 1825; 1834) and in the flysch of the Helvetic Nappe (see Şengör, 2003 for major details). The close association of flysch with exotic blocks and chaotic deposits of Habkern Valley (Helvetic Nappe), strongly contrasting with the well-bedded classical flysch succession, was called *wildflysch* by Kaufmann (Kaufmann in Studer, 1872; Kaufmann, 1886) after a long drawn-out process on the interpretation of the exotic blocks/flysch relationships and processes of their inclusion (i.e. Fuchs, 1877a, b; Schardt, 1898). Submarine sliding, volcanism, fluvial transport, glacial processes and *in situ* gas venting (Murchinson, 1849) have been invoked as triggering mechanisms (see Şengör, 2003 for major details). In spite of the strict association of *wildflysch* with regional shear zones at the base or within the Helvetic Nappes, nobody interpreted these deposits as a product of tectonic deformation until the introduction of the nappe concept in the Alps (Bertrand, 1884; Schardt, 1893). Following the introduction of this new concept, the exotic blocks embedded in the *wildflysch* were interpreted as (see also Camerlenghi and Pini, 2009 and Mutti et al., 2009 for major details): (1) polymictic debris formed in front of the nappes (Beck, 1912; Debemas and Kerchkove, 1973; Mercier de Lépinay and Feinberg, 1982; Weidmann et al., 1982; Piguet et al., 1998), in a similar way to the precursory olistostromes of the Apennines (Elter and Trevisan, 1973); (2) tectonic products formed in front or at the base of the nappes (Burkhard, 1988; Jeanbourquin et al., 1992; Jeanbourquin, 1994) by tectonic reworking or tectonic imbrication (Häfner, 1924; Badoux, 1967); (3) subduction-related *mélanges* that consist of tectonically deformed olistostromes, which are emplaced at the advancing front of nappe structures (Trümpy, 2006).

The scientific controversy on the meaning and interpretation of the *wildflysch* was based on the problem of the occurrence of exotic, extra-formational blocks, their emplacement (see also Mutti et al., 2009), and the intense deformation of the host shale and/or sandstone matrix. The same problem has largely affected the *mélange* literature starting from the 50's (e.g. “the problem of knockers and blueschists rocks” of the workers on the Franciscan *mélange*, i.e. Brothers, 1954; Bailey et al., 1964; Coleman, 1971; Coleman and Lenphere, 1971; Cowan, 1974; Karig, 1980 and references therein). Then, a conceptual linkage was created between *wildflysch* and *mélanges*.

Today, the term *wildflysch* seems to be used to describe chaotic deposits with different degrees of stratal disruption (Table 1). It appears to be a continuum from originally coherent and well-bedded turbidites (including olistostromes), to broken formations, up to totally disrupted formations and chaotic deposits including exotic blocks (see Mutti et al., 2009 for major details).

Mélanges and broken formations

The term *mélange*, abbreviation of “autoclastic *mélange*”, was coined in 1919 by the British geologist Edward Greenly (Table 2) while he was mapping the “Gwna Group” of the Mona Complex in Anglesey, North Wales (Greenly, 1919). However, despite its detailed description of the chaotic assemblage type (that later was acknowledged to be a “tectonic *mélange*”), Greenly’s discovery was largely overlooked and the term did not become popular until the description of the “chaos structure” (in Almargosa Range, Eastern California; Table 2) of Noble (1941) (see Şengör, 2003 for major details). The term was subsequently resurrected by Bailey and McCallien (1950, 1953) and by Gansser (1955). The first Authors adopted the term coined by Greenly in describing a tectonic mixture in Anatolia (Ankara *mélange*, Turkey) and the *argille scagliose* (“Ligurian *mélange*”) in the northern-Apennines of Italy. Gansser (1955) used the term “Colored *Mélange*” in Iran and introduced the term “ophiolitic *mélange*” (Table 2) to describe hybrid *mélanges* formed by both tectonic and sedimentary processes and composed at least in part of ophiolitic material.

During his studies on the Franciscan *mélange*, Hsü (1968) reinvented the concept of *mélange* whose usage became popular, thanks to a precise definition and a detailed discussion on the processes of their formation: fragmentation and mixing. Hsü proposed to use “*mélange*” only for tectonic *mélanges* and defined “broken formations” (Table 2) as strataly disrupted units that contain no exotic elements and that preserve their lithological and chronological identity (“tectonites” *sensu* Castellarin and Pini, 1987; “tectonosomes” *sensu* Pini, 1999). In this sense, the “Autoclastic *mélange*” of Greenly (1919) should be considered as broken formation. Moreover, Hsü (1968) suggested to use the term “olistostrome” (following the usage of Flores, 1955, see below) only for sedimentary mixtures that were not considered to be a *mélange*.

Following Hsü (1968), a long-lived debate has taken place on the definition of the term *mélange* that assumed different and in some cases controversial meanings (see Raymond, 1984 and reference therein), and that referred to both processes (tectonic, sedimentary and diapiric) and tectonic setting(s) of their formation. In the Western Alps and the circum-Pacific region (north-America Cordillera, New Zealand and Japan), this term has been used to indicate chaotic rock bodies formed mainly by tectonic processes in a subduction zone (e.g. Cowan, 1978, 1985; Aalto, 1981; Cloos, 1982; Barber et al., 1986; Brown and Westbrook, 1988; Cloos and Shreve, 1988; Onishi and Kimura, 1995; Meschede et al., 1999; Wakabayashi, 2004; Ikesawa et al., 2005; Federico et al., 2007). The virtual association between *mélanges* and subduction is closely linked in these regions (Cowan, 1985), partly because the most popular chaotic rock bodies described as *mélange* (i.e., the Franciscan Complex) has been mainly interpreted as a subduction product. In the Alps, poorly metamorphosed to eclogite, tectonic *mélanges* were largely described in different structural positions within a nappe pile (see Polino et al., 1990 and references therein), from the Eastern (e.g. Frisch, 1984; Winkler, 1988; Ring et al., 1988, 1989) and Western Alps (Caron et al., 1989) to the Piedmont ophiolitic units (Marthaler, 1984; Marthaler and Stampfli, 1989). On the contrary, in the Central-Eastern Alps, the external Alpine domains (Helvetic and Ultrahelvetic) and the circum-Mediterranean mountain chains belonging to the Alpine-Himalayan system (i.e., the Southern Alps, Apennines and Sicily), the term *mélange* has been used to indicate a larger spectrum of chaotic rock bodies formed in various and different tectonic settings (rift-drift cycles, oceanic subduction, continental collision and intracontinental deformation) by tectonic, sedimentary and diapiric processes and their mutual interactions (Şengör, 2003; Camerlenghi and Pini, 2009; Festa et al., 2009, and references therein). Then, many terms (i.e., *wildflysch*, *schistes à blocs*, olistostromes, *argille brecciate*, *argille scagliose*, olistolith fields, megabreccias, agglomerates, etc.; see Table 1) and interpretations that were derived from the description of the Alpine-Apennine *mélanges* have been applied in describing *mélanges* everywhere (see Hsü, 1968; 1974; Cowan, 1985; Camerlenghi and Pini, 2009; Festa et al., 2009). This led to a re-evaluation of the virtual association of *mélanges* with subduction zone processes (see Cowan, 1985).

During this process of re-evaluation (e.g., Raymond, 1984), the interest of the scientific community has been addressed to: (1) establish valuable criteria for distinguishing processes of *mélange* formation (e.g., Naylor, 1982; Byrne, 1984; Cowan, 1985; Orange, 1990; Clennell, 1992; Pini, 1999; Lucente and Pini, 2003) especially for polygenic *mélanges* (e.g., Gansser, 1974; Cowan, 1978; Aalto, 1981; Raymond, 1984; Saleeby, 1984; Yilmaz and Maxwell, 1984; Barber et al., 1986; Lash, 1987; Orange, 1990; Pini, 1999; Cowan and Pini, 2001; Dela Pierre et al., 2007; Festa, 2009); (2) understand the significance of *mélanges* and other chaotic rock units in the geological record (i.e., Gansser, 1974; Raymond, 1984; Cowan, 1985; Scherba, 1989; Camerlenghi and Pini, 2009; Festa et al., 2009).

Extension of the Alpine concept of *mélange*: olistostromes, olistoliths and argille scagliose

Flores (1955; 1956; 1959) defined sedimentary bodies with a chaotic, block-in-matrix fabric intercalated with well-bedded successions were defined as “olistostrome” (slide accumulation) during his studies of the Tertiary marine sedimentary units of Sicily (Table 1). These chaotic bodies, different from block-in-matrix rock types, were considered the products of gravity mass movements, encasing both scaly clays (“*argille scagliose*”, Merla, 1952) and “brecciated clays” (“*argille brecciate*”, Ogniben, 1953, or “*argille puddingoidi*”, Rigo de Righi, 1956), the distribution and origin of which were the subject of a scientific controversy (see Ogniben, 1953; Beneo, 1955, 1956a). Similar chaotic deposits were described in the Alps by the first supporters of the emerging nappe concept (e.g. Schardt, 1898; Lugeon, 1916; Tercier, 1947) as a result of submarine sliding, producing a mixture of sediments and exotic blocks originated from the front of the advancing nappes (see Mutti et al., 2009 for major details). These chaotic deposits, described as *wildflysch*, may be called as “olistostrome” or “precursory debris flow” (Trümpy, 1960; Elter and Trevisan, 1973). In this sense, a conceptual linkage exists between *wildflysch*, representing the first concept of sedimentary *mélange* in the Alps, and olistostromes.

Wildflysch, olistostrome and *argille scagliose* were subjected to similar controversies regarding the nature of their exotic blocks, the mechanisms that triggered their emplacement, and the processes of their formation (see Camerlenghi and Pini, 2009 and Mutti et al., 2009 for a review). Olistostromes are sedimentary bodies derived from different types of gravity mass movements, such as block slides, debris avalanches, debris flows and hyper-concentrated flows (Lucente and Pini, 2003). A broad definition of olistostromes includes bodies with blocks ranging in size from a centimeter to a meter, enveloped in a clayey or sandy-silty matrix (type A olistostrome, Pini, 1999); bodies with larger blocks (olistoliths) ranging in size from tens to hundreds of meters sustained by a matrix of the type A olistostrome (type B olistostrome, Pini, 1999); bodies, without a matrix, almost completely consisting of olistoliths (type C olistostromes, Lucente and Pini, 2003).

The blocks (olistoliths) and the matrix are typically polymictic and both exotic, extrabasinal with respect to the host succession (e.g. Beneo, 1956b; Rigo de Righi, 1956; Abbate et al., 1970; 1981; Elter and Trevisan, 1973; Bettelli and Panini, 1985; 1987; Labaume, 1992; Pini, 1999; Cowan and Pini, 2001; Lucente and Pini, 2003, 2008; Camerlenghi and Pini, 2009), although olistostromes composed of only intrabasinal (native) sediments were also described by Jacobacci (1963), Abbate et al. (1970; 1981) and Elter and Trevisan (1973), as related to intraformational collapse and resedimentation (see, e.g., Naylor, 1981). Elter and Raggi (1965) distinguished “endolistostromes” that contain no exotic blocks from “allolistostromes” (Table 1) consisting of both blocks and matrix derived from different depositional basins. Raymond (1978) used “endolistostromes” as sedimentary equivalent of Hsü’s (1968) broken formation and “allolistostromes” to indicate sedimentary *mélanges* (with exotic blocks).

The clayey matrix of olistostromes consists of rounded to angular polymictic millimeter-to centimeter scale clasts of claystone, randomly enveloped in a poorly compacted, “open” textured clays (brecciated matrix, see Abbate et al., 1981 and Pini, 1999 and references therein). It represents the classic “*argille brecciate*” and “*argille puddingoidi*” of Ogniben (1953) and Rigo de Righi (1956). However, a scaly fabric that may overprint but not completely obliterate the brecciated texture, is commonly present (Bettelli and Panini, 1985). Thus, olistostromes also include *argille scagliose* (scaly clays), a term introduced by Bianconi (1840) to describe the mesoscopic (millimeter-to centimeter scale) matrix of the block-in-matrix rocks of the Ligurian Units in the Northern Apennines. This term was largely adopted for almost all chaotic bodies of mixing rocks with a block-in-matrix fabric both in the Apennines (e.g. Signorini, 1946; Azzaroli, 1948; Merla, 1952; Ogniben, 1953; Maxwell, 1959; Elter, 1960) and throughout the world (e.g. Hsü, 1968; 1974; Cowan, 1985). Then, just as happened for the term *mélange*, the usage of the term “*argille scagliose*” to describe different areas characterized by different types of chaotic rock units (including both Jurassic ophiolitic units and Mesozoic sedimentary successions of both continental and/or oceanic affinity), was extended from purely descriptive to process-oriented (tectonic, sedimentary or diapiric) meanings (see Camerlenghi and Pini, 2009 for a complete review).

Today, the *argille scagliose* in the Northern Apennines have been differentiated into: “tectonites, tectonosomes or broken formations” (Castellarin et al., 1986; Bettelli and Panini, 1987; Castellarin and Pini, 1987; Pini, 1999) represented by strongly deformed units preserving a coherent stratigraphic succession and, “olistostromes or sedimentary mélanges” (Bettelli and Panini, 1985, 1987; Castellarin and Pini, 1987; Pini, 1999) represented by sedimentary, block-in-matrix bodies formed by sedimentary mass transport processes. The latter may also be formed with contributions of diapiric processes (see Camerlenghi and Pini, 2009; Festa et al., 2009). True tectonic mélanges, which are intermixed chunks of units of different ages and palaeogeographic origins with contractional contacts associated with thrust splays, have been also distinguished as belts of limited areal extent inside the *argille scagliose* (i.e., the Coscogno mélange, Bettelli et al., 2002) (see Fig. 5 in Festa et al., 2009).

Mélange and olistostrome occurrence in orogenic belts and accretionary complexes

Mélanges and olistostromes *sensu latu* occur in many collisional and accretionary orogenic belts around the world and are of tectono-stratigraphic significance because they form in different geological environments (Fig. 1). Collisional orogens display a record of various, long-lived geological events accompanied by complex processes during which mélanges and olistostromes may form. These include subduction of oceanic material and local emplacement of ophiolites during pre-collisional stages, thickening of crust and lithospheric mantle during collisional stages and gravitational instability formed as a consequence of continued plate indentation during post-collisional stages (Dilek, 2006).

The Alpine-Himalayan belt, extending from the Iberian Peninsula to southern Asia (e.g. Argand, 1916), is one of the most studied and classic collisional orogens. Mélanges and olistostromes *sensu latu*, related to subduction/obduction and/or to accretional pre-collisional stages have been widely described from the Alps, (e.g. Polino et al., 1990 and references therein; Ring et al., 1990; Trommsdorff, 1990; Hoogerduijn Strating, 1994; Federico et al., 2007), in the Alpine circum-Mediterranean orogens (Apennines, Betic Cordillera, Carpathian, Anatolia and Caucasus) (see, among many others Argnani, 1987; Dilek and Moores, 1990; Clift and Dixon, 1998; Ricou et al., 1998; Dilek et al., 1999; Pini, 1999; Kozur et al., 2000; Okay, 2000; Robertson, 2000; Pamić et al., 2002; Bettelli and Vannucchi, 2003; Pickett and Robertson, 2004; Beccaletto et al., 2005; Dilek et al., 2007; Vannucchi et al., 2008; Ghikas et al., 2009. See also Camerlenghi and Pini, 2009 and Festa et al., 2009 for a complete review), and Iranian-Afghan-Himalayan orogens (Burg et al., 2008; Dilek et al., 2009; Gansser, 1974 and references therein).

Most olistostromes and olistoliths have been largely described as related to extensional tectonics that operated during rift-drift stages or formed in front of accretionary wedges during pre-collisional stages, and at the front of or below allochthonous nappes during both pre- and collisional stages. Some examples of these olistostromes and olistoliths widely occur in the Alps (e.g. Bernoulli and Jenkyns, 1970; Castellarin, 1972; Achtnich, 1982; Frisch, 1984; Eberli, 1988; Wächter, 1987; Froitzheim and Eberli, 1990; Channell et al., 1992; Böhm et al., 1995; Ebli, 1997; Kurz et al., 1998; Gawlick et al., 1999; Mandl, 2000), the Alpine circum-Mediterranean orogens (Dilek et al., 2005; Dilek and Thy, 2006; see Camerlenghi and Pini, 2009 and Festa et al., 2009 and references therein), and the Iranian-Afghan-Himalayan orogens (Liu and Einsele, 1996; Burg et al., 2008; Dilek et al., 2009).

Similar examples of mélanges and olistostromes have been described also from other ancient collisional orogens as well as from the Appalachians (Fig. 1), where arc-forearc sequences were accreted to a continental mass, and as a consequence the origin and nature of these chaotic rock bodies were commonly obscured by multiple periods of overprinting deformation and metamorphism (e.g. Williams and Hatcher, 1983). Subduction and/or thrusting during the Taconian or Alleghanian orogenies commonly formed “olistostromal mélanges” (*sensu* Rast and Horton, 1989), tectonic mélanges, broken formations and olistostromes (e.g. Lash, 1987; Cousineau and St-Julien, 1992; Tremblay et al., 1995; Ganis and Wise, 2008). Olistostromes have been also described from syncollisional piggy-back basins (e.g. Schroetter et al., 2006). Moreover, ophiolite-bearing mélanges have been interpreted as the remnants of a Taconian suture (e.g. Pollock, 1989) or related to allochthonous terranes (e.g. Boone and Boudette, 1989).

Accretionary orogens, extensively distributed in the Circum-Pacific regions, provide some of the most outstanding examples of mélanges and olistostromes (Fig. 1). The Western US Cordillera (e.g., Franciscan complex in western California, Wrangellia Terrane and Pacific Rim Complex in San Juan and Vancouver Islands, and Hoh accretionary complex in Olympic Peninsula), Alaska (Kodiak and Mc Hugh complexes), Mexico (Middle America Trench), Japan (Shimanto, Sambagawa, Chichibu and Mino-Tamba-Ashio Belts, the Miura Group of Miura-Bozo Peninsulas and Nankai Trough) and New Zealand (Torlesse Complex, Coastal Ranges and Hikurangi margin) host some of the most

famous and extensively studied ancient and active accretionary complexes with mélanges and olistostromes. Here, ocean floor rocks that were offscraped from the downgoing plates and/or clastic sediments derived from the erosion of continental or island arcs are prominent parts (respectively) of mélanges and trench chaotic deposits (e.g. Isozaki, 1997).

Different mélanges and broken formations have been related to tectonic processes that occurred during the evolution of accretionary wedges (Fig. 1), as in the Franciscan Complex and in the Western US Cordillera (e.g., Ernst, 1965; Hsü, 1968; Cowan, 1974, 1978, 1985; Cloos, 1982; Dilek, 1989; Maekawa and Brown, 1991; Dilek and Moores, 1993; Kimura et al., 1996), in Alaska (e.g., Moore and Wheeler, 1978; Orange, 1990; Bradley and Kusky, 1992; Kusky and Bradley, 1999), in Shimanto (e.g., Ditullio and Byrne, 1990; Taira et al., 1992; Kimura, 1997; Ikesawa et al., 2005; Shibata and Hashimoto, 2005), Sambagawa (Takasu et al., 1994), Mino-Tamba-Ashio (Kimura and Hori, 1993) and, Chichibu (Matsuda and Ogawa, 1993; Takahashi, 1999) Belts, the Miura-Bozo accretionary complex (Yamamoto et al., 2009) of Japan and in the Torlesse Group (Esk Head mélange and North Island equivalents, Barnes and Korsch, 1991; Orr et al., 1991; Sunesson, 1993) and the Coastal Range of Hikurangi margin in New Zealand (Pettinga, 1982; Chanier and Ferrière, 1991). Mélanges occur in these places commonly at the base of imbricate thrust sheets (e.g. Hsü, 1973), or as tectonic erosion artifacts in accretionary prisms closer to the trench (e.g. Bailey et al., 1964; Karig, 1980).

Karig (1980) underlined the role of strike-slip tectonics in discussing the “knockers” problem in the mélange contest. “Knockers” are blocks of high-pressure metamorphic rocks enveloped by non-metamorphic sedimentary rocks, and their occurrence in tectonic mélanges poses an important question about how these high-P rocks became integrated into the low-T mélange matrix. Other models of explanation of this enigma include tectonic mixing at depth (e.g., Suppe, 1973; Blake and Jones, 1974; Platt, 1975), mixing by near surface processes (Gucwa, 1975), and large-scale differential uplift of metamorphic rocks beneath the trench slope, followed by gravitational sliding back to the trench and then the reincorporation of this mixture of rocks by subduction processes into the accretionary prism (e.g., Cowan and Page, 1975; Cowan, 1978; Page, 1978. See Karig, 1980 for a complete review). Moreover, mélanges formed in fault or shear zones, also known as “sheared mélanges” (*sensu* Needham, 1995) and/or “asymmetric mélanges” (*sensu* Hammond, 1987; Fisher and Byrne, 1987), have been described from the Shimanto belt and southern Scotland (see Needham, 1995 and Shibata and Hashimoto, 2005 and reference therein; Fig. 1 and Table 2).

Accretionary orogens are also the place for olistostrome formation (Fig. 1) that commonly occur at the front and at the base of accretionary wedges as described, for example, from the Franciscan complex and the Western US Cordillera (e.g., Hsü, 1968; Cowan and Page, 1975; Aalto, 1981, 1989; Cowan, 1985; Brandon, 1989) or from the Shimanto, Sambagawa and Chichibu Belts and the Miura-Bozo accretionary complexes of Japan (e.g., Sakai, 1981; Hisada, 1983; Taira et al., 1992; Aoya et al., 2006; Yamamoto et al., 2007; Osozawa et al., 2009), and from the Torlesse Complex (Hada and Landis, 1995) and the Coastal Range of the Hikurangi margin in New Zealand (Chanier and Ferrière, 1991; Delteil et al., 1996, 2006).

In addition, accretionary wedges may include mud volcanoes and diapirs (Fig. 1). The Barbados accretionary prism (e.g., Westbrook and Smith, 1983; Brown and Westbrook, 1988), Timor and Indonesia along the Banda Arc-Australia accretionary and collisional complex (e.g., Barber et al., 1986; Barber and Brown, 1988), the Eastern Mediterranean ridge (e.g., Cita and Camerlenghi, 1990; Kopf et al., 1998; Camerlenghi and Pini, 2009 and references therein), the Hoh accretionary complex in the Olympic Peninsula (Orange, 1990) are few of the most intriguing examples containing mud volcanoes and diapirs.

In contrast to olistostromes, mud volcanoes and diapirs are mainly submerged. They outline the submerged fold and thrust belts and accretionary complexes, whereas olistostromes outline the mature collisional orogens and the fossil fronts of accretionary prisms on land (see Camerlenghi and Pini, 2009). Although, rather rare, some fossil examples of mud volcanoes and diapirs (both mud, serpentinite, and metamorphic diapirs) have been reported (Fig. 1) from the Alps and Alpine basins (e.g., Ernst, 1965; Di Giulio, 1992; Clari et al., 2004; Dela Pierre et al., 2007; Festa, 2009), the Western US-Cordillera (e.g., Orange, 1990), the Appalachians (e.g. Lash, 1987) and the New South Wales in SE Australia (e.g. Fergusson and Friksen, 2003). Other examples may be found in the Paleozoic sequences of the North Island of New Zealand (Cobb Valley, Baloon mélange, Jongens et al. 2003). See also Kopf (2002) for a complete review.

Finally, mélanges, broken formations and olistostromes *sensu lato* commonly occur in different geological settings and are attributed either to large-scale gravity driven phenomena, or to diapiric movements and tectonic deformation. Their widespread distribution points to the importance in understanding the role of chaotic rock bodies in the evolution of

orogenic belts and accretionary prisms, and leads to a fundamental question: whether or not a genetic relationship exists between different types of mélanges and their tectonic-paleogeographic settings?

Mélange classification: tectonic environment and processes of formation

A subdivision of mélanges cropping out in the circum-Adriatic chains (Apennines, Dinarides, Albanides and Hellenides) has been recently proposed by Festa et al. (2009), based on the geodynamic setting of mélange formation. This chapter extends this classification scheme to all the other “classic” areas of mélange formation around the world and present a brief discussion on the main characters of some of the more emblematic cases.

1 - Mélanges related to extensional tectonics

Mainly angular clasts, ranging in size from decimeter to several tens of meters, and subordinate, smaller, non-angular clasts randomly distributed in a fine-grained (pelitic) matrix characterize this type of mélanges. Blocks are commonly slightly older than the matrix and its depositional age, and generally have an intra-basinal composition. Formation processes of these sedimentary chaotic bodies are consistent with *en-mass* gravitational movements (debris avalanches and flows and sliding of blocks; Fig. 2), and the bodies are also known in the geological literature as megabreccias, olistolith and olistoliths fields or swarms. These terms mostly refer to the gravitational collapse of the margins of carbonate platforms, generating carbonate megabreccias that are common in the slope and base-of-slope settings (Fig. 2). Source areas of blocks mainly consist of already cemented carbonate platform margins. The matrix is normally represented by prevalent pelagic limestones. Normal faulting associated with extensional tectonics is the triggering mechanisms affecting the margins of carbonate platforms during and after rifting (Fig. 2), although the initial geometry of carbonate platform margins may cause similar bodies to develop without the contribution of active tectonics (scaloped margins, *sensu* Bosellini, 1998).

The most notable examples are the large middle Triassic-Jurassic and Cretaceous megabreccias in the Southern Alps (e.g., Castellarin, 1972; Bosellini et al., 1977) and the early Jurassic breccias of the Northern Calcareous Alps (e.g., Channell et al., 1992; Böhm et al., 1995; Ebli, 1997; Mandl, 2000). Some spectacular examples are preserved in the Northern Apennines (e.g., Castellarin et al., 1978; Cecca et al., 1981; Bernoulli, 2001 and many others), Majella and Gargano Peninsula (see Bernoulli, 2001 and references therein) and Western Hellenides (e.g., Naylor and Harle, 1976). In the Appalachians, the well-known “Precambrian mélanges” represent some other notable examples of mélanges that formed during extensional rifting episodes (e.g., Rast and Kohles, 1986; Bailey et al., 1989).

2 - Mélanges related to passive margins and ocean floor

Passive margin mélanges (Fig. 3) are mainly represented by poorly sorted olistostromes, which consist of fine-grained carbonate and siliciclastic turbidites and mudstones and/or monomictic brecciated (matrix-supported) masses. Olistostromes commonly show a soft-sediment deformation related to *in situ* folding, boudinage and slumping. Progressive deformation from slumps to cohesive debris flow leads to a complete strata disruption evidenced by a block-in-matrix fabric, with blocks randomly distributed in a fine-grained matrix. Locally, these sedimentary rocks show a foliation and/or a planar elastic fabric formed by compaction and flow respectively (see also Naylor, 1982). Plastic deformation of clasts and fluidal structures in the matrix suggest that sediments were deformed while they were poorly consolidated or non-consolidated (Fig. 4).

The formation of this type of olistostromes is consistent with different gravity-driven processes (e.g. Mienert et al., 2003; Masson et al., 2006) of downslope transport of sediments by slides or slumps and debris flow (laminar motion) (Fig. 3B). Commonly, slides and slumps evolve dynamically during downslope movements (Fig. 3B) and they turn into debris flow and then into turbulent flows (turbidites) (e.g. Mienert et al., 2003). The processes responsible for this type of chaotic rock bodies normally operate following the main stage of continental rifting and in a passive margin setting at the edge of the thinned continental margin or at the transition to the oceanic realm.

Triggering mechanisms are various (Fig. 3) and difficult to be interpreted, but what they have in common is that the slope instability mainly depends on the shear strength of the sediments deposited on the slope. Tectonic reactivation of pre-existing faults and large earthquakes, effects of the pre-existing submarine topography (Fig. 3A), deep-ocean

processes, and gas hydrates dissociation (due to the changes in sea level or increase in bottom water temperature) represent some possibilities of triggering mechanisms (Fig. 3B).

Other important examples are described from around the world for both contemporary and ancient cases. Modern examples are well exposed along the passive continental margins of the Gulf of Mexico (e.g. McAdoo et al., 2000), Norwegian margin (e.g. Bugge et al., 1988; Laberg et al., 2000; Haflidason et al., 2001), NW-African margin (e.g. Embley and Jacobi, 1977) and many others sites (see Hampton et al., 1996; Mienert et al., 2003; Masson et al., 2006, and references therein). Some fossil examples are described, for example, from the Circum-Mediterranean area (see Camerlenghi and Pini, 2009 and Festa et al., 2009), the Appalachians, and the Himalayas. In the Northern Apennines, the lowest horizon of the Modino basal complex (De Libero, 1998; Pini et al., 2004) is characterized by different degrees of progressive stratal deformation (from *in situ* folding to boudinage and slumping, to “block-in-matrix” debris flow bodies) affecting the early Cretaceous Palombini shales (Fig. 4). Smith et al. (1979) and Shallo (1990) have described similar examples in Greece (Othris Mountains) and Albania, respectively. In the Northern Appalachians, different examples of these types of chaotic rock bodies have been described as related to the instability of passive margins during the Taconic orogeny. For example, slides and slumps triggered by changes in slope block rotation and increased seismic activity along nearby faults characterized the middle Ordovician Trenton Formation (Taconian foredeep) (Jacobi and Mitchell, 2002). In the Himalayan belt (Nepal), late Cretaceous olistostromes derived from passive (Indian) continental margin and developed at the toe of the continental slope are characterized by different horizons with a block-in-matrix fabric (Liu and Einsele, 1996). “Floating clasts” of shallow-water sandstones and limestones (from millimeter to tens of meters in size), with inner deformed structures, are enveloped in a fine-grained matrix, which represents deep-water sediments.

Open oceanic setting is an ideal place for the formation of bodies generated by failure of seamounts and submarine volcanoes and within oceanic sediments. In this case argillaceous and/or siliceous fine-grained sediments host angular blocks (ranging in dimensions from some tens of centimeter to some kilometers) of magmatic and volcanic rocks of the oceanic crust and limestones derived from reefs and carbonate buildups capping the seamounts. This origin has been suggested for mélanges, part of mélanges or isolated, large-scale blocks in some of the circum-Pacific fossil accretionary wedges, such as in the Sambagawa (Ayoa et al., 2006) and Chichibu Belts (Matsuda and Ogawa, 1993) in Japan.

Contribution of these sedimentary processes to accretionary mélanges can be extremely important from volumetric point of view, when a seamount chain arrives at a trench and subduction zone. The actualistic examples of Hawaii and Canary Islands show that some of the largest submarine landslides occur due to marked instability of volcano flanks (see, e.g., Lipman et al., 1988; Moore et al., 1989; McMurtry et al., 1999; Gee et al., 2001; Clague and Moore, 2002; Mitchell et al., 2002).

3 - MÉLANGES RELATED TO STRIKE-SLIP TECTONICS

This group refers exclusively to tectonic mélanges related to strike-slip deformation (Fig. 5), which dismembers previously coherent stratigraphic successions producing a tectonically disrupted unit. Some of the diagnostic key features of these types of mélanges include (Dela Pierre et al., 2007; Festa, 2009): (1) repetition, at different scales, of structural associations consistent with the regional stress field, (2) structurally ordered block-in-matrix fabric, (3) elongated shape of the blocks and their parallel alignment to the shear zones, and (4) decrease of strata disruption away from the fault zones. Mélanges of this group generally occur within wide from tens of meters (e.g. Festa, 2009) to kilometres (e.g. Van de Fliert et al., 1980; Jolivet et al., 1983; Karig et al., 1986) wrench-type fault zones progressive or abrupt and they are related to different geodynamic environments ranging from intracontinental and episutural basins to fore-arc and accretionary prisms.

Classic structural features of strike-slip regimes (e.g. P-C, S-C, Riedel shears and scaly cleavages) are present from micro (< 1mm) to-macro-scale, compatible with the regional stress field (Fig. 5). Blocks, ranging in size from centimeters to hundreds of meters, have elongate and lozenge shapes (Fig. 6) and show polished surfaces commonly contained by slickenlines. These blocks are aligned with the main shear zones, showing an en-échelon arrangement according to the sense of shear. Riedel shears and S-C or P-C fabrics commonly occur within the blocks (Fig. 6) that may be exotic in origin with respect to the matrix.

The origin of blocks (exotic or not) is an important point of discussion because it may differentiate strike-slip tectonics-related mélanges (with exotic blocks) from strike-slip tectonics-related broken formation (with no exotic blocks) (Fig.

5). However, the distinction between strike-slip tectonic-related broken formation and classic strike-slip shear zones is not always clear-cut and maybe it is only in the mind of workers on *mélanges*.

There are only few examples of *mélanges* associated with strike-slip tectonics in the world. Some of the most known examples of strike-slip tectonics-related *mélanges* have been described by Karig (1980) to explain the enigmatic occurrence of blocks of high-pressure metamorphic rocks (“knockers”), composed typically blueschists, eclogites and amphibolites, embedded in a matrix of non-metamorphosed rocks. Strike-slip tectonic processes of *mélange* formation as proposed by Karig (1980) present a plausible alternative model (see Karig, 1980 for major details) in explaining the mechanisms of transport and the trajectory of material associated with subduction zone kinematics in plate convergence zones (the “Knockers” problem).

Nias in the Sumatra region (Western Sunda Arc) and the Pacific margin of Baja California (in particular the San Benito Island area) are the two main examples described. In the former, blocks of garnet amphibolite are enclosed in a tectonic *mélange* composed mainly of clastic sediments (in which metamorphism is incipient at most) and formed along left-lateral splay faults (e.g. Batee fault) of the Sumatran strike-slip fault zone. Karig (1980) reported that this situation was reminiscent of the blueschist “knockers” of the Franciscan complex. In the latter example (Baja California), the emplacement of blocks of high-grade metamorphic rocks into a *mélange* matrix were related to significant amount of previous episodes of strike-slip faulting associated with oblique subduction.

In the Franciscan complex, McLaughlin et al. (1988) described another example of strike-slip tectonics-related *mélange*. This *mélange* of the Central Franciscan Belt consists of blocks of diverse lithologies and metamorphic grades embedded in a shale-matrix, which exhibits prehnite-pumpellyite facies metamorphism. Occurrence of fossils of different ages (from Tithonian to Valanginian) within the blocks and matrix of the Central Franciscan Belt may be a result of considerable tectonic recycling interpreted as the product of strike-slip activity.

Other classical examples of this type have been reported from the island-arc setting of the Philippines (Haeck and Karig, 1983; Karig et al., 1986). On the basis of the occurrence of schistosity/lineation fabric elements within a *mélange* and the Cenozoic regional stress field, Karig et al. (1986) elaborated a model that invoked strike-slip faulting in the transport of allochthonous arc terranes in the Philippines.

The *mélanges* of Hidaka Western Margin Belt and the Eastern Kamuikotan Tectonic Belt in northern Japan represent few other examples of strike-slip tectonic *mélange* (Hidaka right-lateral fault) (Jolivet and Miyashita, 1983; Jolivet et al., 1983). A gradual increase of deformation from Idonnappu greenrocks to Hidaka Western meta-ophiolites has been interpreted by these authors as strain partitioning associated with strike-slip tectonics. The *mélange* itself is a sedimentary type including blocks, from centimeters to few kilometers in size, mainly of gabbro, diabase, pillow lava, chert, conglomerate and limestone.

In the Circum-Mediterranean region some examples are described from the Northern Apennines (e.g. Abbate et al., 1970; Bortolotti et al., 2001; Marroni et al., 2001; Cerrina Feroni et al., 2002), Tertiary Piedmont Basin in NW Italy (Dela Pierre et al., 2007; Festa, 2009) and Albania (e.g. Shallo, 1990; Rassios and Dilek, 2009). For further detail on these examples, see Camerlenghi and Pini (2009) and Festa et al. (2009). Perhaps the most classic examples in this region are the Neo-Moni and Arakapas *mélanges* in Cyprus (Robertson, 1977; Krylov et al., 2005). The former is related to the Tertiary transpressional movements that reactivated the pre-existing subduction-related olistostromes (Robertson, 1977). The latter is an olistostrome produced along fault scarps related to the Cenomanian-Campanian strike-slip activity in a fore-arc basin (Krylov et al., 2005).

4 - MÉLANGES RELATED TO SUBDUCTION

This group is by far one of the most common *mélange* types observed in the Circum-Pacific region, although different examples have also been described from the Circum-Mediterranean region as the artefacts of the Eo-Alpine episodes of convergence (Late Cretaceous, see Trümpy, 1973). We have identified two sub-types of subduction-related *mélanges* on the basis of their processes of formation.

4a - Mass-transport deposits at the wedge front

These *mélanges* are characterized by a chaotic arrangement that consists of different degrees of stratal disruption (Figs. 7 and 8). Chaotic deposits generally include deformed intrabasinal sediments (Fig. 8A) and extrabasinal rocks of different

ages (commonly older than the intrabasinal component and timing of their emplacement) supplied from the front of accretionary wedge and/or wedge-top basin (Figs. 7A and 7B). Different degrees of stratal disruption are related to the state of consolidation at the stage of the slope failure, and their run-out distance. Pinch-and-swell, boudinage, slump folds and “slump balls” (Fig. 8D) represent the most common structures affecting the intrabasinal sediments. Local imbrication of beds and bed-packages, contractional and extensional duplexes, thrust systems and intrafolial, isoclinal folds are also present (see, e.g., Taira et al., 1992; Chanier and Ferrière, 1991; Yamamoto et al., 2007, 2009; Fig. 8A). As for the “passive margin-related mélanges”, deformation of these sediments seems to have occurred when they were wet and unconsolidated. The extrabasinal rocks are present as blocks (Figs. 8B, C and E), bed chunks and entire bed packages (preserving their subduction-related tectonic fabric) inside a fine-grained, commonly argillaceous matrix (see, e.g., the Forcella Pass olistostrome in the Val Lavagna Shales, Bortolotti et al., 2004, and the other Upper Cretaceous olistostromes of the Ligurian units in the Northern Apennines, Abbate et al., 1970; Pini, 1999). Aalto (1981) recognized in some of the mélanges that are spatially associated with the Franciscan Complex similar olistostromes, which were supplied by already indurated, deformed and mixed rocks from the thrust system at the wedge front.

Sedimentary mélange bodies deposited at the wedge front are bounded at the base by an erosional discontinuity surface and commonly show a lenticular shape with areal extension of tens to hundreds square kilometers. As implied by the name, their formation is consistent with different types of mass-transport phenomena as sliding, slumping, debris flow and debris avalanches and their interactions (see, e.g., Abbate et al., 1970, 1981; Chanier and Ferrière 1991; Pini, 1999; Lucente and Pini, 2003).

Triggering mechanisms depend on a combination of several factors (Fig. 7) that control the slope oversteepening and instability at the front-of-wedge (“frontal tectonic erosion” *sensu* von Huene and Lallemand, 1990). These are, for example, tectonic removal at the base of an accretionary wedge (“basal tectonic erosion” *sensu* von Huene and Lallemand, 1990), subduction of seamounts and ridges (Fig. 7A), subduction erosion (e.g. Marroni and Pandolfi, 2001; von Huene et al., 2004) and thrust faulting and folding (Fig. 7B). These mechanisms may uplift the wedge-front, promoting sedimentary instability by increasing the slope angle. Upward rise of overpressured fluids and unconsolidated sediments derived from the subduction zone (Fig. 7B) represent another probable factor to generate slope failure, especially if it is accompanied by the previously described processes.

In the Western Alps, mud diapirism has been proposed as a triggering mechanism for the origin of similar chaotic deposits (Di Giulio, 1992). During the late Maastrichtian, tectonic loading produced a diapiric flow along the decollement surface developed at the toe of the accretionary wedge and a diapiric ridge including exotic blocks of pelagic limestones, jasper and pillow lavas.

The most notable examples triggered by diapiric uprising have been described from Eastern Indonesia (Barber et al., 1986; Barber and Brown, 1988). These authors proposed that the widespread distribution of mélanges in some of the ancient accretionary complexes is incompatible with the scale and extension of modern slumping processes but is adequate with the large volume of block-in-a clay matrix produced by massive shale diapirism (Barber and Brown, 1988).

Some important examples of mass-transport triggered by oversteepening, which may result from subduction of seamounts and ridges, have been described as part of the Eo-Alpine convergence episodes. The Bocca Shales in the Northern Apennines, for example, represent a chaotic product of frontal tectonic erosion (Marroni and Pandolfi, 2001) of an early Paleocene accretionary wedge slope (i.e. subduction zone of the Ligure-Piemontese ocean basin). This erosion was produced by episodic instability due to the subduction of reactivated Jurassic faults offsetting the oceanic crust and causing relief in the lower plate. The same authors suggested that the presence of mass-transport deposits at the toe of an ophiolite sedimentary cover could be commonly seen in the oceanic units of the Apennines and the Corsican and Western-Alps (e.g. Polino, 1984; Durand Delga, 1986; Deville et al., 1992; Ducci et al., 1997; see also Camerlenghi and Pini, 2009 and Festa et al., 2009). Similar examples have been described from the Costa Rica and Nicaragua sectors of the Middle America Trench (Von Huene et al., 2004 and references therein), where the steepening of the seafloor by the uplift over subducted seamounts produced mass-transport deposits and erosion at the wedge front.

In modern subduction systems, geophysical observations and drilling evidenced many other notable examples both of subduction of seamounts and ridges (e.g. Hikurangi margin, New Zealand, Collot et al., 2001; Costa Rica continental margin, von Huene et al., 2004; Hühnerbach et al., 2005; North of d’Entrecasteaus Ridge ad Bougainville Guyot in New Hebrides Island Arc, Collot and Fisher, 1992; Collot et al., 1994; Nankai and Japan trench; Cochonat et al., 2002; Tonga Trench, Ballance, 1991) and sediments offscraped at the front of the prism (e.g. Hamilton, 1979; von Huene, 1979;

Karig, 1980; Westbrook et al., 1982, 1988; Karig et al., 1986; Ditullio and Byrne, 1990). The latter, immediately thickened at the front-wedge, produced a resulting unstable topographic slope that may have favoured abundant surficial slope deposits (commonly reincorporated into the accretionary prism).

Finally, chaotic sediments grouped in this type of *mélange* may be, at least, formed as a result of submarine mass-transport phenomena and/or mud diapirism that occur in downslope gravitational masses. These two products are commonly mixed together at the front of a wedge.

4b - Broken formation and tectonic *mélanges*

Variable physical conditions and plate interactions in subduction zone (e.g. rate of consumption, nature of downgoing plate, different degrees of progressive lithification and increasing burial temperature and pressure, nature of the offscraped and underplated material, amount of fluid flow and pore fluid, etc.) may lead to the formation of various stratal disrupted units, which can be of both sedimentary and metamorphic in nature (Fig. 9). Broken formation and tectonosomes (Hsü, 1974; Pini, 1999) are characterized by a block-in-matrix fabric in which part of the same coherent stratigraphic unit can be recognized (Figs. 9 and 10A-B), and in which no “exotic” blocks are present. Beds and bed fragments show a planar orientation (Figs. 9D-E) consistent with an internal order at the outcrop to map-scale that coincides with a structural fabric (e.g. Cowan and Pini, 2001). Various types of structures may have formed from the interaction of different subduction-related processes (Figs. 7, 9 and 10).

Fluid pressure and lithostatic loading that occur during the early stages of subduction may produce, for example, layer-parallel extension through the formation of (moderate) boudinage (Fig. 8E) and network of small scale, normal faults (e.g. Lash, 1987; Barnes and Korsch, 1991; Sunesson, 1993; Ujiie, 2002). Fluid overpressure is also responsible for the generation of layer-parallel veins and vein systems dissecting the relatively more competent beds (Bettelli and Vannucchi, 2003; Meneghini et al., 2009), and for web structures (Cowan, 1982; Byrne, 1984), up to fluidification (Fig. 10F) and the complete desegregation of rocks (see, e.g., Cosineau, 1998; Yamamoto et al., 2009; Fig. 10D).

Shear stresses are responsible for small-scale structures such as veins (Ohsumi and Ogawa, 2008) and scaly fabric (Pini, 1999; Vannucchi et al., 2003; Figs. 11 and 12), for mesoscale pinch-and-swell and (asymmetric) boudinage of beds (Figs. 9D, 9E and 9H), for isolated or intrafolial hinges of isoclinal folds (Barnes and Korsch, 1991; Ujiie, 2002), and for duplex and thrusts at all scales with related block stacking and bed transposition (e.g. Hirono and Ogawa, 1998; Pini, 1999; Yamamoto et al., 2000; Cowan and Pini, 2001; Ikesawa et al., 2005; Niwa, 2006). Transposition is also evident in the matrix by centimetre-to meter-scale interfingering of thin clayey beds, which commonly exhibit a penetrative scaly fabric (Pini, 1999; Cowan and Pini, 2001). Protracted folding associated with thrusting can cause boudinage and transposition of sedimentary layers (Vannucchi and Bettelli, 2002; Bettelli and Vannucchi, 2003).

Underthrusting of an originally coherent succession (Fig. 7C) prior to its underplating beneath an accretionary prism represents another case as described from the Shimanto belt (eastern Shikoku, southwestern Japan) by Kimura and Mukai (1991). The formation has been interpreted in terms of Riedel shear associated with layer-parallel shearing and tectonically induced loading normal to the layering (Kimura and Mukai, 1991), or in terms of layer-parallel extension by boudinage.

Even though disruption is the most prominent feature of a broken formation, bedding may be locally preserved both as a single bed and meters-to tens of meters-thick packages of a coherent stratigraphic succession (Figs. 9A-E). Generally, there is a lateral gradation from coherent bed units to broken formations (Figs. 9D-E and 10A, B) and, in the latter, from packages of coherent beds to packages of disrupted portions (see, e.g., Barnes and Korsch, 1991; Sunesson, 1993). At a regional scale, in the Northern Apennines of Italy, the same stratigraphic interval, the so-called basal complexes of the Ligurian flysch units, are cropping out as coherent to episodically and slightly disrupted units in the west (“Internal Ligurian units”, see Marroni, 1994; Meneghini et al., 2009) and moderately to highly-disrupted units (broken formations, Vannucchi and Bettelli, 2002; Bettelli and Vannucchi, 2003), and highly to completely disrupted units at the eastern border of the Apennines (tectonosomes of Pini, 1999). The style and amount of deformation vary in the same stratigraphic succession depended on the rheological characters of different stratigraphic intervals (see, Vannucchi and Bettelli, 2002; Bettelli and Vannucchi, 2003).

The key point is the relationships between broken formations or tectonosomes and tectonic *mélanges*. These relationships are unclear or ambiguous in the existing literature that we have reviewed during the course of this study.

Several units defined as *mélanges* in some studies can be more precisely defined as broken formations due to the progressive and pervasive deformation of the same stratigraphic unit, with no “exotic” blocks or mixing of rocks of different age, P-T conditions and origin. Two different schools of thought exist about (true) *mélanges* pertaining to their origin (e.g. Underwood, 1984). One of these suggests that *mélange* formation is inherited from gravitational failure of sediments occurring in trench-slope settings (e.g. Moore and Karig, 1976; Kimura et al., 1992) prior to the accretion of these sediments into accretionary prisms; the other one suggests tectonic deformation by offscraping or underplating along the subduction interface (e.g. discussion in Underwood, 1984; Cowan, 1985; Figs. 7B and C). In the latter case, the main mechanism of rock mixing is considered to be out-of-sequence thrusting of already stacked units, tectonic erosion of hangingwall and/or footwall along faults and thrusts, and thickening of fault zones, (see, e.g., Moore and Byrne, 1987; Doubleday and Trenter, 1992; Festa et al., 2009 and references therein). However, other models have also been proposed to explain the wide variety of *mélange* formation in subduction zone settings as, for example, the “subduction channel model” (see Cloos and Shreve, 1988 for major details). In summary, *mélange* and broken formation-tectonosomes display lithological and structural evidence for different degrees of mixing and deformation depending on the rheology of the rocks involved, superposition of different tectonic episodes, and different degrees of involvement in subduction processes.

Some important examples of broken formation occur in so-called basal complexes, chaotic complexes and *argille scagliose* in the Western-Alps and Apennines that represent the remnants of Upper Cretaceous eo-Alpine and Eocene meso-Alpine episodes (see Camerlenghi and Pini, 2009 and Festa et al., 2009 and references therein). For example, the Cretaceous to Eocene Ligurian and Sicilide Units and the Lower Cretaceous to Campanian basal complex (detached from the Ligurian Helminthoid flysches) correspond to broken formations and tectonosomes (e.g. Pini, 1999; Bettelli and Vannucchi, 2003; Camerlenghi and Pini, 2009; Festa et al., 2009).

In the central Appalachians, in-situ accretion-related deformation caused by tectonic loading during the early stages of subduction developed a progressive layer-parallel extension (Fig. 9E) in the Middle Ordovician trench-fill deposits of the Hamburg sequence (Lash, 1987). The occurrence of a complete and progressive deformation displayed well-bedded greywacke horizons grading into boudins and pinch-and-swell structures (Fig. 9E) indicates that the Type I *mélange* of Lash (1987) is indeed a broken formation.

Type I, II and III *mélanges* of Cowan (1985) show a gradual increase of deformation that may also be representative of different stages of disruption in the same subduction zone setting, as well as of various degrees of deformation in different tectonic settings.

Many other researchers have documented the progressive nature of deformation that produced wide spectrum of chaotic rocks ranging from broken formations to *mélanges*. For example, Ditullio and Byrne (1990) described from the Shimanto belt (Japan) a structural progression of deformation consistent with the increasing strength of rocks, decreasing fluid pressures, and increasing temperature and pressure conditions as the sediments were dewatered, lithified and integrated into the accretionary prism. In the Permian-Cretaceous *mélange* of the Mc Hugh Complex in South-Central Alaska, its matrix flowed around rheologically strong blocks (i.e. greywacke, limestone, chert, basalt, gabbro and ultramafic rocks), forming broken formation and *mélange* during deformation (Kusky and Bradley, 1999).

The *mélange* formed during subduction in the Diego Ramirez Islands (SE Chile) owes its internal structure to different episodes of deformation (Wilson et al., 1989). Stratal disruption and isoclinal folding (forming broken formations and *mélanges*) occurred during the early stages of subduction while subsequent underplating produced a new pervasive fabric through cataclastic shearing.

Other examples of subduction-related tectonic *mélanges* have been widely reported from the Alps and the Circum-Pacific region. For example, the Furgg zone (Pennine Alps) is considered a *mélange* (Froitzheim, 2001) formed during oceanic subduction and overprinted by the subsequent continental collision. It includes blocks of continental basement rocks, ophiolite, and Permian-Mesozoic continental cover rocks. A similar interpretation has been proposed for the Arosa zone (Central-Eastern Alps) that consists of a serpentinite or shaly-calcareous matrix *mélange* containing blocks of both Penninic and Austroalpine origin (Ring et al., 1990). Blocks of both continental and oceanic material are preserved in a *mélange* of the Cima Lunga and Central Adula eclogite zone (Central Alps) that has been interpreted as formed in a south dipping subduction zone (Tommsdorff, 1990).

In Western-Alps (Voltri Massif), an ophiolitic *mélange* zone, which consists of meter-scale blocks of metagabbro, metabasite, metasediments and serpentinite in a schistose chlorite-actinolite matrix, represents a possible example of an

exhumed subduction channel (Federico et al., 2007). The Osa mélangé (Costa Rica) may represent another example of an exhumed channel transporting tectonically eroded material down into the subduction zone (Meschede et al., 1999).

Although, many other examples of subduction mélanges can be given from around the world (e.g. Cowan, 1982; Moore and Byrne, 1987; Kimura et al., 1992; Orange et al., 1993), it is important to note that diapiric processes including both sedimentary rocks and serpentinites (Figs. 13 and 14) may be triggered by subduction processes producing mélanges (e.g. Ernst, 1965; Cowan, 1985; Cloos and Shreve, 1988; Westbrook and Smith, 1988; Orange et al., 1993).

5 – MÉLANGES RELATED TO COLLISION

These groups of mélanges (Fig. 15) are not well represented in the geological record. The most probable reason could be that in most collisional orogens the timing of the end of subduction processes and beginning of intercontinental deformation varies greatly throughout paleogeographic realms. Moreover, the products of the critical transition between accretion and collision are commonly involved in subsequent intracontinental deformation episodes and, are therefore extensively reworked. Thus only a few mélanges have been definitively attributed to this group (see, for example, Chang et al., 2001 and Huang et al., 2008; Fig. 15) although it is not easy to define the main characteristics of these chaotic rock bodies.

Wakita (2001) has shown that the mélanges formed in the Cretaceous accretionary collisional complex of Indonesia are of sedimentary origin in contrast to the tectonic mélanges of the Jurassic accretionary complex of Japan. The former contains allochthonous blocks of older formations representing fragments or blocks that are intercalated with coherent sedimentary successions. In the same region, the Savu mélangé (Savu Island) has been recognized as the product of the Sunda-Banda arc-continent collision (Harris et al., 2009). Blocks of indurated sandstone, limestone, and metamorphic and igneous rocks of Permian to Paleogene ages, mostly referred to as the Banda terrane, are floating in a volumetrically prevailing muddy matrix. This unit, correlated with the Bobonaro mélanges, is associated with recent mud diapirism.

The most notable example of the Circum-Mediterranean region is the Petra to Romiou mélangé (Fig. 16) in the southwestern part of the Island of Cyprus (Krylov et al., 2005) that represents the product of the Paleocene-Early Eocene collision of the Africa plate with the Cyprus microplate. It consists of a red and brown clays and mudstones matrix including blocks (up to tens of meters) of alkaline basalts and Triassic limestones (Fig. 16).

The collision between the pre-Apulia foreland and the Pindos Vourinos intra-oceanic arc, during the Eocene, produced the Advella mélangé in the Western Hellenides (Ghikas et al., 2009) that represents a polygenetic mélangé originally emplaced by sedimentary processes (rift-related mélangé). It consists of middle Triassic to Cretaceous blocks enveloped in a Cretaceous-Eocene matrix composed of shelf and turbidite deposits. In the Apennines, the base of the Epiligurian succession above the Ligurian nappe consists of Eocene mass transport deposits (breccias and olistostromes), which likely represent a collision-related mélangé.

In the Southern Appalachians, the mélangé of the Chunky Gal Mountain Complex (Lacazette and Rast, 1989) consists of blocks of metasandstone, plagiogranite, amphibolite and ultramafites included in a metasedimentary matrix. The block-in-matrix fabric is tectonic and has been related to successive episodes of thrusting and shearing during the Taconic and Acadian orogenic events that involved arc-continent and continent-continent collisions.

6 - MÉLANGES RELATED TO INTRACONTINENTAL DEFORMATION

This group is probably the most common type in ancient orogenic belts, as many examples have been described from the Alpine-Himalayan, Circum-Mediterranean and Appalachian mountain systems. Processes responsible for the emplacement of ophiolites and accretionary wedge complexes over continental crust controlled the formation of these types of mélanges. We have subdivided these mélanges into three main sub-types on the basis of their position and occurrence with respect to the allochthonous nappes: sub-nappe, intra-nappe and epi-nappe mélanges.

6a – Sub-nappe mélangé

6a1 – Precursory olistostromes

They consist of classic olistostromes and *wildflysch* commonly characterized by chaotic rock bodies in a block-in-matrix fabric formed at the front of thrust and/or nappe systems and deposited by cohesive debris flows and/or block avalanches in migrating foredeep basins (Figs. 17 and 18). The term and the main characteristics of the precursory olistostromes were originally defined by Elter and Trevisan (1973) and have been largely described from throughout the world (see also Camerlenghi and Pini, 2009; Festa et al., 2009 and references therein).

The main features of this sub-group are the same as those of the classical olistostromes (Fig. 18). The blocks/matrix ratio may be variable passing from matrix-supported blocks of centimetre-to meter size (Type A, Pini, 1999), to blocks of meters to tens of meters sustained by a matrix of Type A (Type B, Pini, 1999), to almost clast-supported larger blocks or olistoliths (Type C, Lucente and Pini, 2003). The matrix mainly consists of a brecciated matrix containing sub-millimeter-to centimeter-sized clasts that are randomly distributed. Blocks of Type A and B are fragments of commonly boudinaged single beds (with lithologies depending on the source area), weakly deformed strata composed of undeformed units and/or strongly deformed blocks of broken formation-tectonosomes (Pini, 1999). Larger blocks of Type C may preserve matrix material that fills interstices (Lucente and Pini, 2003) and, if pseudo-bedding is preserved their random distribution clearly appears. The material making up the blocks and matrix are commonly exotic with respect to the host succession and are derived from accretionary wedge and/or intrabasinal settings.

Factors controlling the formation of these types of mélanges are various, including earthquakes on active margins, oversteepening slope angles, thrusting and duplexing, but what they have in common is that each of these processes effectively increase stresses in weakened sediments and triggering failures.

In the Apennines and Alps, most notable examples of sub-nappe mélanges are well preserved in all stages of the migrating foredeep complex as a result of the collapse of the wedge-front and resedimenting of extrabasinal material in the foredeep (Fig. 18). In the Northern Apennines, the emplacement of precursory olistostromes of the Ligurian, Sub-Ligurian and Epi-Ligurian origin followed the Oligocene to middle-late Miocene migration of the Marnoso-arenacea foredeep and the Messinian to Pliocene deposition of the front-Apenninic succession (e.g. Abbate et al., 1970; Ricci Lucchi, 1986; Lucente and Pini, 2003, 2008 and references therein). Similar examples are preserved also in the Central-Southern Apennines, where precursory olistostromes of *argille scagliose* (of Sicilide origin) were progressively deposited during the eastward migration of the upper Tortonian to Messinian foredeep (e.g., Sgroso, 1988; Cosentino et al., 2002; Vezzani et al., 2004; Festa et al., 2006; 2009; Patacca and Scandone, 2007).

In the Alps, the term *wildflysch* commonly substitutes the term “precursory olistostrome” and different cases of *wildflysch* formation have been well documented in the Penninic, Helvetic and Ultrahelvetic domains. *Wildflysch* characterized by large blocks and olistoliths derived from the advancing nappe systems and emplaced into a foredeep have been described in different Alpine sectors. Hsü and Schlanger (1971), for example, described it from the Ultrahelvetic foredeep (late Eocene *wildflysch*). In the Versoyen Valley (Western Alps) a *wildflysch* (Méchanteur Formation) is exposed along a tectonic contact between the Paleozoic ophiolite of the Versoyen-Petit St. Bernard nappe on top and the Sion-Courmayeur-Tarantaise calcschist unit of the Valais zone below (Masson et al., 2008). This *wildflysch* has been considered by the same authors as the equivalent of the Visp *wildflysch* (Valais; Switzerland) which contains blocks and olistoliths of ophiolites. In the Penninic Zone of the Eastern Alps (Matrei zone, Tauern window), *wildflysch* formed in front of an advanced nappe probably triggered by a seismic activity (Frisch, 1984). Moreover, other examples have been described from the Variscan *wildflysch* of the Saxothuringan zone (NE Bavaria, Western Germany) by Behr et al. (1982) and in the Albanides by Gawlick et al. (2007) (Radiolaritic *wildflysch*).

Eardley and White (1947) suggested that the Alpine *wildflysch* is probably similar to parts of the Chany shale in the Ouachita Mountains in the Southern US. This orogenic belt preserves several notable examples of *wildflysch*-precursory olistostromes most of which were formed and emplaced during the Taconic orogeny (e.g., Bird, 1963; 1969).

One of the most intriguing examples of sub-nappe mélanges is represented by the middle-late Ordovician Hamburg sequence or Dauphin Formation (*sensu* Ganis and Wise, 2008; also known as Hamburg Klippe *Auct.*) in the Central Appalachians. The emplacement of this allochthonous unit within the middle Ordovician Martinsburg flysch was recorded in olistostromes and *wildflysches*. The Conodoguinet and Middlesex *wildflysch* formations (Root and MacLachlan, 1978) were associated with the arrival and emplacement of the allochthonous Enola Sheet and involved gravity sliding as an emplacement mode. The early Ordovician olistoliths described by Ganis et al. (2001) in the Dauphin Formation represent another evidence of this gravity driven emplacement mechanism.

6a2 – Olistostromal carpet at the base of the nappe

Olistostromes emplaced in front of an advancing nappe (Type 6a1 *mélanges*) and/or accretionary wedges (Type 4a *mélanges*) by protracted activity of mass-transport processes (debris flow and avalanches) represent the material that, when overridden, form the so-called “olistostromal carpet” (Pini et al., 2004; Fig. 17). Thus, both conceptual and genetic linkages exist between these types of *mélanges* (Lucente and Pini, 2008; Camerlenghi and Pini, 2009).

Mélanges representing an olistostromal carpet (Fig. 19) are characterized by classic *block-in-matrix* olistostromes (i.e., debris flows) associated with large slabs of variously deformed and disrupted strata from the front of a nappe (rock avalanches, block sliding) and sediments from thrust-top basins and slope deposits above the front of a nappe (debris flows, slide, slump). Also typical is the superposition of tectonic deformation and shearing due to loading and the shear stresses exerted by the nappe emplacement. Tectonic deformation and shearing may occur with various degrees of intensity up to the point of development of penetrative shear zones (Figs. 19A and C). Typical shear zone structural associations and a structurally ordered *block-in-matrix* fabric may be observed in ancient olistostromes (“sheared *mélange*” and “asymmetric *mélange*” *sensu* Needham, 1995).

Mélanges related to olistostromal carpet (Fig. 17) are, therefore, considered mixed tectono-sedimentary chaotic bodies in which exotic blocks may be supplied both by mass-transport episodes at a front of the nappe and by erosion at the base of the nappe (Festa et al., 2009 and references therein). This is, for example, the case for different interpretations of the Sestola-Vidiciatico unit in the Northern Apennines (Fig. 19A). This unit has been interpreted both as an olistostromal carpet (Lucente and Pini, 2008) and as an equivalent of an ancient subduction channel (Vannucchi et al., 2008). However, different interpretations, such as submarine landslides at the front of the Ligurian nappe or highly deformed muddy debris flows (olistostromes), are not excluded in the latter model of Remitti et al. (2007) and Vannucchi et al. (2008).

A significant question (see Alonso et al., 2006) regarding the *mélange* forming processes is whether they are purely tectonic in origin and produced by shear zones affecting nappe fragments (e.g. Bailey and McCallien, 1950; Vollmer and Bosworth, 1984; Jeambourquin et al., 1992) or whether they are the products of mass-transport detached from a nappe toe (Signorini, 1940; Page, 1962; Caron, 1966; Elter and Trevisan, 1973; Page and Suppe, 1981). A typical example is the Porma *mélange* located below a nappe stack in the Variscan foreland of the Iberian Peninsula (Fig. 19B). It consists of a highly deformed block-in-matrix succession with boudinated large exotic blocks (Fig. 19B) interpreted as the product of gravity sliding at the front of the moving nappes (Alonso et al., 2006). The largest amount of deformation in the *mélange* is related to the gravitational phases spreading, and then sliding at the top-front of the nappe but not to the subsequent nappe emplacement.

Other notable examples have been described in the literature. In the Central Alps, for example, the Habkern *mélange* (Kempf and Pfiffner, 2004) is a tectono-sedimentary *mélange* formed at the base of an advancing thrust front. This *mélange*, previously called “*wildflysch*”, consists of imbricated slices of marl and Nummulitic limestone of the Helvetic domain, intermixed with slices of a Penninic flysch and with local occurrences of large blocks of granites and Mesozoic sedimentary rocks of both Helvetic and Penninic origin.

A classic example is the The Hudson River Valley *mélange* (Central-Northern Appalachian) described by Vollmer and Bosworth (1984) and Bosworth (1989). This *mélange* formed in the Taconian foreland basin, where olistostromes and olistoliths generated by slope instability at the active emergent front of a nappe were then overridden by the thrust sheet and were incorporated into the shear zone. Shearing led to the juxtaposition and mixing of rocks of various ages and the formation of boundinage, enucleation of isoclinal folds, and phacoidal microshear cleavages (Vollmer and Bosworth, 1984; Kidd et al., 1995).

Many other notable cases have been described in the literature, from Anatolia (Bailey and McCallien, 1950, 1953; Parlak and Robertson, 2004), Othris Mountains in Greece (Smith et al., 1975), Western Alps (Jeambourquin et al., 1992), Nankai accretionary wedge (Cochonat et al., 2002), Yuwan accretionary complex in Japan (Osozawa et al., 2009), and the Middle America trench (Ranero and von Huene, 2000; von Huene et al., 2004).

6a3 – Tectonic *mélanges* at the base of a nappe

Deformation associated with large shear zones at the base of regional-scale thrust sheets or nappe systems (Figs. 17C1 and 17C2) may have involved the formation of sub-nappe mélanges and/or broken formation-tectonosomes. The related structural features in this case are similar to those of classic sheared mélanges and tectonosomes (see Needham, 1995; Pini, 1999). Fluid circulation, dewatering processes, and deformation related to overpressure are the most important phenomena acting in the shear zones. Only close to the main shear zones, the block-in-matrix fabric is structurally ordered with respect to the regional stress field only close to the main shear zones, whereas, the lack of a preferential regional orientation prevails accords to high fluid presence away from these shear zones. Overpressure conditions at the base of the thrust sheets produced brittle deformation in the hard-rock interlayers.

The Central-Southern Apennines provide spectacular examples of tectonic mélanges of this type (Fig. 20). Between the allochthonous thrust sheet and the buried Apulian units, a several hundreds of meters thick mélange zone is present (e.g. Roure et al., 1991; Butler et al., 2004; Vezzani et al., 2004; Festa et al., 2006, 2009; Patacca and Scandone, 2007) that consists of strongly deformed and overpressured deepwater successions. In Molise, for example, this mélange zone is represented by the Flysch Rosso at the base of the Molise unit (see Vezzani et al., 2004; Festa et al., 2006, 2009). It consists of a fault-bounded tectono-sedimentary body with a mainly structurally ordered block-in-matrix fabric (Fig. 20) that, far from the main shear zone, gradually passes into thick (several hundreds of metres) horizons lacking of any preferential orientation of the blocks in the matrix. Overpressured fluids may also have triggered the upward rise of fluids and underconsolidated sediments producing diapirism (e.g. Roure et al., 1991).

Bailey and McCallien (1950, 1953) provided another notable example of a tectonic mélange from the Ankara Mélange in north-central Turkey as a product of a large shear zone at the base of a regional-scale thrust sheet. However, other authors (e.g. Boccaletti et al., 1966) interpreted this chaotic assemblage as olistostromes emplaced at the front of a nappe. Yilmaz and Maxwell (1984) interpreted the Alakir Çay mélange of the Antalya complex (southwestern Turkey) as a result of compression at the base of an ophiolitic thrust sheet, but they also pointed out the role of the advancing nappe in supplying olistostromes and olistoliths in a foreland basin (see also Camerlenghi and Pini, 2009, and references therein).

These examples demonstrate that the separation between tectonic mélanges and olistostromal carpets, at the base of an advancing nappe, is not always clear cut (Fig. 17C2). Even if not in the case of a nearly continuous carpet of olistostromes, sparse olistostromal bodies and olistoliths can be always expected to collapse from the front of a nappe, thus contributing “exotic” block of the upper plate to be mixed with the indigenous component of the lower plate. This process may act together with the mechanism of tectonic erosion of the base of the lower plate.

6b – Intra-nappe mélange

These mélanges are related to deformation during nappe translation and are subdivided on the basis of the processes of their formation.

6b1 – Sedimentary

This type of mélange includes blocks of intrabasinal origin and/or sedimentary successions located at the margins of a depositional basin. Blocks and matrix mainly consist of the same composition or represent a transitional composition representative of the whole sedimentary succession in the basin. Conglomerates, breccias, megabreccias and large olistoliths mainly represent the classic blocks of this type of mélange (Fig. 21) but also shale and limestone pebble, cobble and boulder may occur.

The internal architecture of these mélanges and the internal block-in-matrix fabric document submarine gravity-driven mechanisms of emplacement, including both rock-fall and grain flow. Although many sedimentary processes are gravity controlled, tectonic events seem to be the most important triggering mechanism. Post and syn-shortening normal faulting and/or out-of-sequence thrusting may lead to erosion and downslope gravity instability that may affect coherent successions in the inner sector of a nappe and thrust sheet.

Some notable cases of this mélange sub-type are well described from the Apennines. For example, carbonate intrabasinal structural high represented the source of the Breccia della Renga (Serravallian – Tortonian, see Cipollari and

Cosentino, 1995; Critelli et al., 2007) in the Mt. Simbruini. This chaotic rock body, which is composed of carbonate breccias, intrabasinal arenites and calcilithites, were deposited by gravity-driven processes. It currently rests unconformably on the carbonate bedrock.

As for other examples in the Central-Southern Apennines (e.g., Breccie di San Massimo – Fig. 21B, Castevetere Formation and Gorgoglione Flysch, see for example Patacca and Scandone, 2007 and references therein), this type of intra-nappe sedimentary *mélange* was mainly produced during change-over of tectonic conditions from thrusting to normal faulting or out-of-sequence thrust propagation (e.g. Sgrosso, 1978; Ghisetti and Vezzani, 1998; Festa et al., 2006; Patacca and Scandone, 2007). In Southern Apennines, for example, large blocks (Fig. 21A) and olistoliths (up to hundreds of meters in size) of Cretaceous platform-derived carbonates embedded in the lower part of the Castelvete Formation (*sensu* Patacca and Scandone, 2007) represent apron deposits of intrabasinal origin, emplaced by gravity flow (Critelli and Le Pera, 1995). However, different authors have interpreted this olistolith-bearing unit as an erosional product at the front of the migrating Apennine thrust sheet (e.g. Pescatore, 1988) or as a product of deposition on thrust-sheet-top related to out-of-sequence thrust propagation (e.g. Patacca and Scandone, 2007).

6b2 – Tectonic and/or tecto-sedimentary

Large Intra-nappe shear zones produce this type of *mélange*, which differs from the “sub-nappe tectonic *mélange*” (Type 6a3) only on the basis of its tectonic setting of formation. These *mélanges* may be represented by both broken formation – tectonosomes, formed for layer-parallel shearing and slicing along the main thrust surfaces, and by tecto-sedimentary *mélanges* occurring very close to formation of olistostromal carpet at the base of the nappe (Type 6a2).

Formation of these *mélanges* and broken formation-tectonosomes commonly occurred along weakness horizons and detachment levels associated with overpressurized shear zones. Except for the case in which shearing overprinted the previously formed intra-nappe sedimentary *mélanges* (Type 6b1), tectonics is the normally main process for *mélange* formation. It occurs as a progressive deformation producing broken formation – tectonosomes characterized by poor disruption and block-in-matrix mixing; however, a recognizable stratigraphy is still commonly observable. This is the case, for example, of the deformation of the so-called “Red Beds” (*sensu* Vezzani and Ghisetti, 1998) that occurred along the main thrust surfaces imbricating the Molise Unit in the Central-Southern Apennines (e.g. Ghisetti and Vezzani, 1998; Vezzani et al., 2004; Festa et al., 2006).

In the same area, the outer front of the Apennine belt is underlined by a thick 10-15-km-wide chaotic succession, Mt. Frentani *mélange* (Vezzani et al., 2004, 2009, 2010; Festa et al., 2006, 2009; Torrente Calaggio Formation, Patacca and Scandone, 2007; Metaponto Nappe *Auct.*), consisting of a brecciated matrix locally overprinted by a scaly fabric and mesoscale shear zones, and blocks (up to hundred of meters) of the Cenozoic Molise Unit, Messinian evaporitic, and Pliocene foredeep successions. This large chaotic body represents an intra-nappe *mélange* produced by both tectonic and sedimentary processes, and is analogous in style and scale to active deformation along the contemporary Hikurangi margin in New Zealand (Ghisetti et al., 2003). Mass gravity processes during tectonic deformation are consistent with slope failure and debris avalanches at the front thrust, rapidly overprinted by thrusting and folding (Fig. 22). Overpressure conditions within the *mélange* and tectonic loading provided by the rapid deposition of late Pliocene – early Pleistocene top-thrust and foredeep succession on it might have promoted upward rise of overpressured and underconsolidated sediments (mud volcanoes and diapiric *mélanges*) as inferred from some outcrops in the area North of Atessa (Vezzani et al., 2004; Festa et al., 2006, 2009).

The Coscogno *mélange* in the northern Apennines (Bettelli et al., 2002) is composed of components from different levels of the Ligurian nappe, from the deepest Subligurian units to the deposits of the piggy-back, nappe-top Epiligurian basins. According to Bettelli et al. (2002), these lithosomes represent thrust splays with contractional contacts. It should represent a tectonic *mélange* developed deep inside the Ligurian nappe (Bettelli and Panini 1992; Capitani, 1993). The exhumation of the *mélange* and the involvement of the Epiligurian deposits occurred in the late Oligocene–early Miocene, and has been attributed to either out-of-sequence thrust dissecting the entire Ligurian nappe, transpressional activity along strike-slip faults, or to normal faulting (Capitani, 1993).

6c – Epi-nappe *mélange*

This mélange type occurs in piggy-back (*sensu* Ori and Friend, 1984) or thrust-sheet-top basins and is subgrouped into different types on the basis of the processes of their formation.

6c1 – Sedimentary

Epi-nappe sedimentary mélanges are the most common of this group and are characterized by a block-in-matrix fabric (Fig. 23), which is the product of gravitational instability along the margin of a piggy-back basin (Figs. 17A-C and 23A). The main components of both the matrix and the blocks are derived from the successions tectonically imbricated in the thrust sheet.

Triggering mechanisms of gravitational instability are various and may be related to climate change, sediment supply and subsidence rate of the basin, but mainly to tectonic processes that controlled the uplift rate of the bounding thrust surfaces (e.g. Beer et al., 1990; De Celles et al., 1991; Nemčok et al., 2005). A piggy-back basin may have developed between two growing anticlines, which strongly controlled the sedimentation as, for instance, in the Eocene-Miocene Caroni and Erin-Ortoire piggy-back basins in Trinidad or the Miocene-Pliocene Iglesia piggy-back basin in the Argentinian PreCordillera (Beer et al., 1990).

A different case is represented by the evolution of a forearc basin into a true piggy-back basin as a consequence of collision. The Oligo-Miocene mélange that formed in the Meso Hellenic Basin (Ferrière et al., 2004) covering the tectonic boundary between the external (Pindos) and internal (Pelagonian) zones of the Hellenide fold-and-thrust belt is a good example of this process. Gravity processes controlled downslope deposition of slumped, fine-grained turbidites to conglomeratic fan.

Another example is the Saint-Daniel mélange that occurs in the base of the Magog Group forearc basin in the Quebec Appalachians (Schroetter et al., 2006). This mélange was deposited during the Taconian orogeny in a piggy back basin within the forearc. Uplift, erosion and burial by enterogenous debris flow were the main processes that formed a block-in-matrix fabric successively reworked by mudflows.

Tectonic uplift of the piggy back basin margins is, one of the most common triggering mechanisms responsible for the emplacement of olistoliths into the basin (e.g. Barrême basin in Western Alps, see Artani and Meckel, 1998) in addition to formation of chaotic rock bodies.

Notable examples of an epi-nappe sedimentary mélange are represented by the middle Eocene to Pliocene Epiligurian units of the Northern Apennines (e.g. Papani, 1963; Bettelli and Panini, 1985; Pini, 1999). These “piggy back basins”, which rest unconformably on top of the Ligurian nappe, consist of mud-rich, block-in-brecciated matrix deposits, with large blocks (up to tens of meter in size), emplaced by cohesive debris flows. The areal extent of these chaotic rock bodies may be up to tens of square kilometres and their thickness up to several hundred meters, suggesting progressive and multi-phase emplacement with independent relative motion of discrete masses along shear zones (Pini, 1987, 1999).

Other mechanisms may also form sedimentary chaotic rock bodies within epi-nappe basins. This is, for example, the case of the Messinian mélange (*sensu* Festa, 2009) of the episutural Tertiary Piedmont Basin in NW Italy. Here, dissociation of gas hydrates facilitated an efficient mechanism in the disruption of the previously coherent Tortonian – lower Messinian pre-evaporitic and Messinian evaporitic successions (e.g. Dela Pierre et al., 2002; 2007; Clari et al., 2004; Irace, 2004; Festa, 2009). The upward rise of fluids triggered by gas hydrates dissociation favoured gravitational instability along the margin of the basin lowering the sediment shear strength and facilitating the formation of chaotic bodies together with other tectonic, sedimentary or riapiric processes (Dela Pierre et al., 2007; Festa, 2009).

6c2 – Tectono-sedimentary

Most of the examples quoted in the previous mélange type are commonly related to tectonic processes that mainly acted as a triggering mechanism for sedimentary disruption processes. Close relationships exist between formation of chaotic rock bodies in a piggy-back setting and temporal changes in the rate of thrusting. Reactivation during and/or after deposition of piggy-back deposits may represent another mechanism of stratal disruption that is able to produce tectono-sedimentary mélanges very close to that of Type 2, except for the geodynamic setting in which they formed.

A piggy-back basin may also experience a shift from extension to compression and *vice versa* during growth of the accretionary wedge (Platt, 1986), and/or out-of-sequence thrusting and backthrusting may lead to the formation of the so-called “mixed mode piggy-back” basin (Weltje, 1992, see also Alçiçek and ten Veen, 2008). The latter is, for example, the case of the early Miocene Acipayam piggy-back basin that rests unconformably on the Lycian mélange in southwestern Turkey (Alçiçek and ten Veen, 2008). Even if this example is not a classic mélange, it may illustrate an efficient model of the formation of epi-nappe tectono-sedimentary mélange.

No clear examples of this type of mélange exist. However, one example was described in Festa et al. (2009) from the Campobasso area (Fig. 24), in the Central-Southern Apennines, where the late Cretaceous-early Miocene Sicilide Units (“Coltre Sannitica” *sensu* Selli, 1962; Sannio Unit *sensu* Patacca et al., 1992) consist of a large intra-nappe (more than thousand kilometer square) tectonically overlying the Molise Units. Its internal arrangement varies from that of the classic olistostromes of *argille scagliose* to that of broken-formation, in which the original succession is still recognizable (Vezzani et al., 2004, 2010; Festa et al., 2006). This arrangement is consistent with the mass-transport phenomena triggered and flanked by tectonic movements. The occurrence of large blocks (up to few kilometre square in size; Fig. 24), commonly preserving the original succession, suggests also gravitational sliding.

6c3 – Diapiric

The presence of intrusive contacts, opposing shear direction on the opposite margins of a structure, zonation of deformation (Fig. 25), block-in-matrix arrangement, and occurrence of hydraulic fractures within the hard blocks are diagnostic features of diapiric type of mélange (e.g. Orange, 1990; Dela Pierre et al., 2007; Festa, 2009). This is the case, for example, of the diapiric disrupted unit (Dela Pierre et al., 2007; Festa, 2009) that, in the episutural Tertiary Piedmont Basin (NW Italy), pierces the tectono-sedimentary Messinian mélange (see type 6c1). Overpressure conditions triggered by an increase of sedimentary loading produced by the abrupt deposition of an intra-nappe sedimentary olistostrome (type 6c1, “Unit 2” of Dela Pierre et al., 2007; “Gravity-driven disrupted unit” of Festa, 2009), and by strike-slip faulting caused the upward rise of underconsolidated sediments. In the same area, Clari et al. (2004) described also Messinian mud diapirs formed by the upward rise of methane-rich fluids probably triggered by gas hydrates dissociation.

It is important to note that in some cases, the epi-nappe sedimentary mélanges are (or are suitable to be) reactivated as diapirs. Commonly, the overlying sediments sink into the sedimentary epi-nappe olistostromes (Figs. 17B and 17C). This is the case of the Epiligurian units of the Northern Apennines (Pini, 1987; 1999) in which viscous flow of the block-in-matrix rocks around the “normal” bedded strata and the almost vertical contacts of the chaotic bodies suggest a mud diapiric reactivation of the olistostromes induced by their underconsolidated state and high pressure of pore fluids. The real magnitude of the phenomenon has not been documented yet, but in some cases the diapiric rise seems to have pierced also the tectonosome units considered as the olistostrome substratum (Pini, 1987, 1999), suggesting an origin of the ascending mud from deeper structural levels inside the nappe, or from the carpet of olistostromes at the base of the nappe.

Mud diapirs piercing thrust top basin deposits have been described in the Makran accretionary wedge offshore Eastern Iran (Grando and Mc Clay, 2007) and their onset related to changes in the wedge dynamics. These bodies occasionally reach the sea-floor originating mud volcano conical apparatus and, thus, supplying sediments (mud breccias) to the thrust-top basins. At a larger scale, mud volcanoes appears as sea mounts at the edge and inside piggy-back basins in the seismic profiles transversal to the southern part of the Barbados accretionary complex (see, e.g., Moore et al., 1990). The high elevation of these bodies above the sea floor controls the sediment dispersal inside the satellite basins, as well as they also contribute with debris flows to the sediment supply (Moore et al., 1990).

Mud diapirs and volcanoes, commonly associated with fluid seepages, have been extensively observed in modern sea floor atop accretionary wedges, as in the cases of the Black Sea, the Mediterranean Ridge, the Gulf of Cadiz, the Cascadia margins, and the Sunda Arc. In some cases, such as the Black Sea and the Gulf of Cadiz, the ascending mud are supplied by extensive horizons of olistostromes (Diaz del Rio et al., 2003; Krastel et al., 2003; Pinheiro et al., 2003; Somoza et al., 2003; Huseynov and Guliyev, 2004; Camerlenghi and Pini, 2009). High pore-fluid content and a very high porosity remain inside mud-rich mass-transport bodies (olistostromes) long time after their emplacement. The

subsequent loading of a thick sedimentary pile, as well as the rapid loading of thrust sheets and nappes, may trigger fluid overpressure causing mud diapirs/volcanoes to overcome, supplied by the olistostrome.

Discussion and conclusions

Mappable units or bodies of mixed rocks, generally described as *mélange s.l.*, are largely preserved in different tectonic settings around the world displaying a record of complex interplay of tectonic, sedimentary and diapiric processes during their formation (Table 3). We have analyzed in this study different *mélange* types in a comparative fashion and have discussed the important geological processes involved in their formation. We concentrate here mainly on ancient examples of *mélanges* that are exposed on-land. The proposed subdivision and classification of *mélanges* (Table 3), following the early work of Festa et al. (2009) on the Peri-Adriatic region, allows us to investigate some of the classical *mélange* problems in further detail and in a more comprehensive approach. In particular, we address the following questions:

- (1) whether there exist some relationships between the types (internal setting and composition, fabric, genetic processes) of *mélanges* and the tectonic setting in which these *mélanges* were formed;
- (2) the relative contribution of sedimentary (mass-transport) versus contractional tectonic deformation processes at the onset and during the evolution of different *mélange* types; and,
- (3) when and how broken formations end and true tectonic *mélanges* begin for rocks whose bedding has been disrupted by means of prevailing contractional tectonic processes.

The internal structure and the regional geology of different examples of *mélanges* indicate that a close relationship exists between the types of *mélange* and the tectonic setting of their formation. The *mélanges* that formed in extensional (i.e. rift-related), passive margin, and open-ocean settings were generated mainly by gravity-induced sedimentary processes (i.e. mass-transport). Direct contribution of extensional tectonic processes, such as crushing and mixing in normal fault zones, are negligible. Tectonic activities here are instrumental in triggering the *mélange*-forming sedimentary processes only.

Tectonic *mélanges* whose fabric elements and structures were caused by contractional tectonics alone develop mainly at the base of accretionary wedges, such as along the main decollement zone, within subduction channels, in the lower plate below the main decollement, and in the zones of underplating in the accretionary wedges, including zones of protracted offscraping and stacking of thrusts, zones of steeply dipping beds and out-of-sequence thrusting, inside the nappe-stacking of collisional orogens (mostly metamorphic, high P and T *mélanges*), and in strike-slip fault zones. Sedimentary *mélanges* are associated mainly with extensional tectonic and passive margin settings, and are commonly found at the base, within and above the shallowest nappes in intracontinental deformation zones of the ancient, submarine collisional orogens. These sedimentary *mélanges* are typical of the geological record of the “Alpine” chains from the circum-Mediterranean region to the Himalayas.

The occurrence of sedimentary *mélanges* in exhumed submarine accretionary wedges is relatively minor, although it has been recently re-valuated in some on-land examples of ancient accretionary complexes in Japan, New Zealand and the West Coast of US, as well as in some remnants of the collisional accretionary wedges (“suture zones” in Alps and Apennines). The relatively lesser abundance of sedimentary *mélanges* in accretionary wedges has been confirmed by the observations from the toes of the modern circum-Pacific accretionary wedges, where only a relatively limited amount of mass-transport deposits with limited dimensions (if compared with other modern geodynamic settings) is seen on the sea floor (see Moore and Byrne, 1987; Camerlenghi and Pini, 2009). An important exception to these observations is when a localized (?) perturbation of the wedge shape is caused by the subduction of a seamount (Fig. 7A) or some asperity in the downgoing plate (reactivated normal faults? see e.g., von Heune and Scholl, 1991; Marroni and Pandolfi, 2001), or when accretion gives way to erosion (Ranero and von Huene, 2000).

Several parameters are important in different tectonic settings because they act in different ways with different magnitudes. These include the state of consolidation and the rheology of pre-deformed sediments, rate of consolidation, permeability, mineral dehydration, dewatering, temperature, overburden pressure, effective stress, strain-rate and, strain histories.

In contrast, the conditions that are necessary for the formation of diapiric *mélanges* are easily reached in different environments where both tectonic and sedimentary *mélanges* may form (Table 3). For example, both sudden loading provided by deposition of gravity-driven sediments (olistostromes and/or mass-transport deposits) and emplacement of

tectonic nappes can cause the buildup of fluid overpressure. For these reasons, diapiric mélanges are not strictly related to a particular tectonic setting but to particular physical and mechanical conditions. This phenomenon also explains as to why diapiric mélanges commonly represent the reactivation of deep-seated olistostromes and/or water-rich tectonic and/or tectono-sedimentary mélanges.

Broken formation-tectonic mélange paradox

Many mélanges, such as the Esk Head and the similar chaotic units in the Torlesse Complex and the mélanges of the Coastal Ranges in New Zealand (i.e., Pettinga, 1982; Sunesson, 1993), the youngest parts of the Shimanto Belt (Japan), a large part of the *argille scagliose* Ligurian “mélange” in the eastern Northern Apennines (i.e., Pini, 1999; Bettelli and Vannucchi, 2003), part of the *argille scagliose* (Sicilide Units) and “Red Beds” (Molise Units) in Central-Southern Apennines (e.g. Vezzani et al., 2004, 2009, 2010; Festa et al., 2006), the Franciscan Complex (i.e., Meneghini et al., 2009; Wakabayashi, 1992), Antarctica (Doubleday and Trenter, 1992), Taconic mélanges in the central and northern Appalachians (i.e. Vollmer and Bosworth, 1984; Lash, 1987; Bosworth, 1989; Kidd et al., 1995) include disrupted strata (broken formation) in which we can still see stratigraphically coherent, different units as well as well-developed block-in-matrix features making a transition to zones of slightly boudinaged beds. The true mélange zones are restricted to (1) thin, elongated and coalescent fault zones (Pettinga, 1982; Barnes and Korsch, 1991), (2) large-scale thrust fault zones (Doubleday and Trenter, 1992; Coscogno mélange, Bettelli et al., 2002; Meneghini and Moore, 2007; Meneghini et al., 2009) associated with fault thickening processes (Moore and Byrne, 1987), and (3) plate boundaries with thick (~1000-2000 meters) mélange zones (see Wakabayashi, 1992; Meneghini et al., 2009).

The transition from the area of mixing to the broken formation is commonly gradational (Barnes and Korsch, 1991), as well as the transition between broken formations and normal layered units, even if limits can be drawn and mapped through careful field observations. In describing the geology of the Southern Hawke’s Bay sector of the Coastal Ranges in New Zealand, Pettinga (1982) mapped elongated “crushing” (broken formations) and mélange belts (“bentonitic mélange with floaters”) corresponding to thrust surfaces and zones dissecting at various angles a Cretaceous-Miocene succession (Table 3; Figs. 10A and 10B).

In light of these observations, true tectonic mélanges should be considered as subordinate with respect to broken formation-tectonosomes (Table 3), and the two groups of chaotic units are somewhat subordinate with respect to the normal bedded units (as in the Esk Head “mélange” in comparison to the Torlesse Complex; Sunesson, 1993; Landis et al., 1999). Examples of more widely distributed chaotic rocks are part of the “*argille scagliose*” in the eastern Northern Apennines, recognized as the products of tectonic disruption (see, e.g., Pini, 1999; Vannucchi and Bettelli, 2002; Bettelli and Vannucchi, 2003 and previous works of the same authors, as well as the sheets of the new 1:50.000 Geological Map of Italy). These rocks are, with the exception of a restricted area of the Coscogno mélange (Bettelli et al., 2002; Festa et al., 2009), broken formation with no exotic blocks and/or mixing of units of different ages. Here, too, however, thin zones of true mélanges are represented by mixing along shear zones marking the contact between different broken stratigraphic units and by the occurrence of Upper Cretaceous olistostromes (Pini, 1999). The reason for this widespread occurrence of broken formation (over an areal extent of hundreds of square kilometers) has been explained by Vannucchi and Bettelli (2002) and Bettelli and Vannucchi (2003) through the interaction between protracted folding and thrusting.

It is important to note that a close relationship seems to exist between mélange types and geodynamic settings of their formation. It is also true that in many cases the final mélange product preserves only the artifacts of the last mélange-forming processes but that it may represent the product of a complex interaction and superposition of different processes, which in many cases are consistent with a tectono-sedimentary evolution of a mélange in a single geodynamic setting. In fact, in most of the mélange examples described in this study olistostromes and mass-transport deposits were later overprinted and deformed by tectonic processes. During their approach to an accretionary wedge and/or a subduction channel, for example, many sedimentary mélanges were gradually transpositioned from the front of nappe systems to the bottom of these nappe piles, whereby they would form the basal mélanges containing exotic blocks. Here, frictional-erosional processes, subduction erosion, folding-thrusting related to out-of-sequence thrusting strongly overprinted the previously formed fabric and structural architecture of the sedimentary mélanges. These mélanges then became tectonically deformed, sheared and mixed with other sediments and/or material derived from the subducting oceanic crust and/or the magmatic arc in the upper plate. Therefore, all mélange types described in this study demonstrate a strong contribution of sedimentary and tectonic processes during their formation. Mass-transport deposits formed by gravity at the wedge front of subduction zones and/or within the accretionary wedges could provide large

amount of material that in turn may affect the mechanisms of accretion, offscraping, and underplating (von Huene and Ranero, 2003; Yamada et al., 2009).

In closure, we define *mélange* as a mappable chaotic body of mixed rocks with a block-in-matrix fabric whose internal structure and evolution are intimately linked to the structural, sedimentary, magmatic, and metamorphic processes in their tectonic setting of formation. Thus, understanding the evolutionary history of *mélanges* requires understanding and documentation of the spatial and temporal relationships between the *mélange*-forming process and the dynamics of the tectonic setting of their origin.

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FIGURE CAPTIONS

Figure 1 – Global distribution of selected *mélanges sensu latu* related to contractional tectonics, sedimentary and (mud) diapiric processes (after Raymond, 1984, redrawn and modified) and giant submarine landslides (after Mienert et al., 2003).

Figure 2 – Conceptual model for the formation of type 1 *mélanges* associated with extensional tectonics, depicting a representative case of megabreccia accumulation triggered by normal faulting within an extended carbonate platform.

Figure 3 – Conceptual model for the distribution (A) and formation (B) of type 2 *mélanges*, associated with passive margin evolution. Relationships among compaction-related normal faulting, gas escape (hydrate dissociation?), and mass wasting processes and bodies are shown. BSR =

Figure 4 - A) Muddy debris-flow body (olistostrome) in the Ronchi lower member (“Palombini shales”, lower Cretaceous) of the Modino basal complex, Modena area in the Northern Apennines (see De Libero, 1998; Pini et al., 2004; and Lucente et al., 2006, for more details) (courtesy of C.C. Lucente). B) Close-up of the outcrop in A showing the flow-related deformational features of non-consolidated carbonate clasts in an argillaceous matrix (from Pini et al., 2004, courtesy of C.C. Lucente).

Figure 5 – (A) Conceptual model for the origin of both broken formation and tectonic *mélanges* (type 3 *mélanges*) as a result of strike-slip tectonics. Sedimentary *mélanges* (i.e., bodies related to diverse mass-transport events and processes) and mud volcanoes associated with positive (B) and negative (C) flower structures may occur.

Figure 6 – Strike-slip duplex structure along vertical strike-slip faults and related low-angle shears, enveloping a deformed block in a strike-slip *mélange* (type 3 *mélanges*) within the *argille scagliose* of Northern Apennines (West of Varzi).

Figure 7 – Conceptual models for the formation and emplacement of type 4 *mélanges* associated with subduction zone processes: A) seamount subduction; B) large double-verging wedge, with a low elevation of the backstop; C) smaller wedge, with a high elevation of the backstop.

Figure 8 – Different examples of mass-transport deposits associated with type 4a *mélanges*: A) Detachment fold and SE-verging thrust (the thrust surface in the lower part of the outcrop is pointed out by Yuijiro Ogawa) and a slump body (sb) in the Misaki Formation of the Miura Group (Kaitocho village, Miura Peninsula, Japan). A detailed reconstruction of this outcrop is available in Yamamoto et al. (2000). B) Coastal exposure of the Pahaoa olistostrome (Glendou Rocks, East Coast of New Zealand). C) Indurated clast surrounded by a fine-grained carbonate matrix of the Pahaoa olistostrome. D) Isolated fold hinge (slump ball?) inside the Pahaoa olistostrome. E) Upper Cretaceous olistostrome in the external Ligurian units, Berceto in the Parma area of the Northern Apennines.

Figure 9 – Examples of broken formations and type 4b *mélanges* in exhumed accretionary wedges. Various disrupted melange units in the Tenguyama (A) and Kawakami (B) Formations in the Chichibu Belt of the Kanto Mountains, Japan, in the Esk Head *mélange* (C), on the road from White Rock to Okuku Pass, west of Christchurch, on the South Island of New Zealand, and in the Coastal Ranges on the North Island of New Zealand (D). E- Layer-parallel extension and boudinage in the middle Ordovician trench-fill deposits of the Hamburg sequence, SW of Kempton, PA, central Appalachian, USA (see Lash, 1987). F- Coastal exposure of deformed rocks within the Rosario fault zone, San Juan Islands, western US (see Cowan and Brandon, 1994). Outcrop consists of lenses of cherts (large lenses and lighter, sigmoidal fragments) in a darker matrix composed of mudstone and cataclastic mafic volcanic rocks. G- Coastal Exposures of the Wrangellia Terrane, Vancouver Island, West Coast of US. Pinkish lenses and bands are made of sandstone, the lighter fragment in the matrix is chert, and the dominant darker matrix is mudstone. H- Lensoidal Upper Triassic pelagic limestone blocks in a heterogeneous and variously deformed matrix composed of shale, mudstone and sandstone in the Jurassic-Cretaceous Avdella *mélange* in the Pindos Mountains of northern Greece.

Figure 10 - Examples of broken formations and tectonic *mélanges* in exhumed accretionary wedges (type 4b *mélanges*): A- Coastal exposure of a fault zone related to an out-of-sequence thrust, dissecting already steeply-dipping beds, Waimarama Beach, South Hawke’s Bay, East Coast of North Island, New Zealand. B- Close-up of A showing the lozenge-shape blocks of sandstone dispersed by boudinage and shearing within a mudstone matrix displaying a pervasive scaly fabric. This kind of fault zone can be considered as a broken formation (see crushing zone of Pettinga,

1982). C- Sandstone blocks in a shaly matrix, Esk Head *mélange*, Torlesse Complex (Okuku River, west of Christchurch, South Island, New Zealand). Notice the imbrication of sandstone blocks in the center of the image. D- Close-up sandstone and volcanic rock (greenish, more elongate lozenges) clasts in a shaly matrix with fluidal features (Esk Head *mélange*, Okuku River). Some volcanic blocks are strongly deformed and plastically elongated along the matrix flow (tectonic *mélange* or sheared olistostrome?). E- Broken formation (Agios Nikolaos Formation) beneath the Vourinos ophiolite in northern Greece. Pervasively foliated mudstone, sandstone, and conglomerate include continentally derived clasts and blocks that are elongated within a mudstone-sandstone matrix. F- In-situ injection of liquefied sediments into coherent layers along a coastal exposure near the village of Kaitocho, Miura Peninsula, Japan (see Yamamoto et al., 2000, 2009 for further details).

Figure 11 – Scaly fabric in the Aptian-Albian Palombini Shales showing the indentation of variously colored clayey and shaly beds (Bologna area Northern Apennines, see Pini, 1999).

Figure 12 – Scaly fabric in the middle Ordovician Taconian Flysch (Taconic *mélange sensu* Kidd et al., 1995), Choes River, close to Waterford, NY, Northern Appalachians, USA.

Figure 13 – Duck Creek mud diapir, as exposed along the Pacific coast of the Olympic Peninsula in the state of Washington, USA. A) General view of the diapiric body (DB). B) Close-up of the internal part of the diapiric body, showing a classic mud breccia, with dispersed angular blocks. The scaly fabric, which is absent in this part of the outcrop, characterizes the more external part, close to the contact with the encasing rocks.

Figure 14 – A- Reworked serpentinite mud enveloping lozenge-shaped peridotite blocks in the Ankara *Mélange* near the city of Çankiri, Turkey. B- Close-up of the serpentinite mud matrix with various clast sizes and showing a pervasive vertical fabric, reminiscent of the internal structure of serpentine mud volcanoes in the forearc of the active Mariana convergent plate margin.

Figure 15 – Conceptual model for the formation and emplacement of type 5 *mélanges* associated with collision tectonics. Data are from Chang et al. (2001), Huang et al. (2008), and Ghikas et al. (2009).

Figure 16 – Collision-induced *mélange* examples of type 5 from the Petra tou Romiou complex in Cyprus (Krylov et al., 1993, 2005). A) Panoramic view: the white bluff (arrow) by the sea is composed of a Triassic limestone in the Mamonia complex; dark colored rocks represent pillow basalts and hyaloclastites of the Dhiarizos Group (Mamonia, DG). White rocks in the foreground are the Paleocene-Eocene Middle Lefkara Formation (LF) deposits overlying unconformably overlying the Petra tou Romiou complex. B) Coastal exposure of the Petra tou Romiou complex: Triassic limestone is in contact with a red-brown mudstone (derived from the Ayios Photios Group of Mammonia) including sparse centimetric blocks of mafic volcanic rocks. See the white chair for scale (marked by the white circle).

Figure 17 – Conceptual models for the formation and emplacement of type 6 *mélanges* associated with intracontinental deformation. A) Obduction of a nappe over the continental margin and an early stage of foredeep development. B) Single large nappe moving (by gravity and thrusting at rear) over the foredeep deposits, which are in turn subject to thrusting and folding below the nappe (inspired by the Ligurian nappe of the Northern Apennine, see Castellarin and Pini, 1987). C1 and C2) represent two different, subsequent stages of nappe stacking atop the continental crust.

Figure 18 – Field examples of precursory olistostromes as type 6a1 *mélanges*. A) A small-scale olistostrome in the Pliocene deposits at the frontal edge of the Ligurian nappe (Bologna area, Northern Apennines); B) Panoramic view of a large outcrop of the Segavecchia olistostrome, Pianaccio area, Bologna Province, Northern Apennines. The olistostrome is stratified in between the Mt. Cervarola turbidite complex, the early Miocene infill of the Apenninic migrating foredeep. Rafted blocks of the Palombini Shale (lower Cretaceous) are completely engulfed within the brecciated block-in-matrix rocks, which are attributed to muddy debris flow deposits (df). C) Debris flow bodies have been remobilized when still completely wet (from Pini et al., 2004, modified).

Figure 19 – Field examples of an olistostromal carpet as in type 6a2 *mélanges*. A) Roadcut showing some of the key features of the Sestola Vidiciatico unit (complex?), with debris bodies (block-in-matrix) in an originally sedimentary contact with the Miocene Civago Marls and the lower Cretaceous Palombini shale broken formation (Ligurian units) from the front of the Ligurian nappe (Ken-ichiro Hisada for scale). B) Darrel Cowan, Jorge Gallastegui and Angela Suarez in front of the Bodon and Forcada nappe (the mountain range at the background) and the large olistoliths (“klippe”, see arrows) inside the Porma *mélange*, Carboniferous of the Cantabrian Chain, North Spain (see Alonso et al.,

2006). C) Strongly flattened clasts of debris flow deposits belonging to the olistostromal carpet at the base of the Ligurian nappe and overlaying the Macigno foredeep complex (earliest Miocene-late Miocene), the Cinque Terre, Tyrrhenian coasts of Italy (by courtesy of Kei Ogata).

Figure 20 – Field example of a broken formation at the base of a type 6a3 *mélanges* nappe. The Flysch Rosso at the base of the Molise Units, central Apennines (SE Abruzzi, Italy), consists of a shear zone with a structurally ordered block-in-matrix fabric.

Figure 21 – Field example of type 6b1 intra-nappe sedimentary *mélanges*. A) Large block (tens of meters in size) of platform-derived carbonate embedded in the Castelvetero Formation (see Critelli and Le Pera, 1995) at Mt. Caruozzo, Southern Apennines, Italy. B) Breccias and megabreccias (Brecce di San Massimo) consisting of calcareous clasts within a yellow glauconitic matrix at the front of Matese, central Apennines, Italy (courtesy of L. Vezzani).

Figure 22 – Type 6b2 tectono-sedimentary intra-nappe *mélange*, (Mt. Frentani *mélange sensu* Vezzani et al., 2004; Festa et al., 2006) consisting of *argille scagliose* (AS) with a block-in-matrix fabric (arrows indicate the main blocks). This *mélange* represents the product of both tectonic and sedimentary (mass-transfer) processes and is superposed on the Late Pliocene sediments (LP). Trigno River, Molise, central Apennines, Italy.

Figure 23 – Field examples of type 6c *mélanges* epi-nappe *mélanges*. A- Panoramic view of a precursory olistostrome of *argille scagliose* (AS) included the Upper Oligocene – lower Miocene sedimentary rocks (OMS) of the Northern Apennines, West of Varzi, Italy. B and C- Typical brecciated matrix, as shown by the epi-nappe olistostromes of the Bologna area in the Apennines. Olistostromes are part of the sedimentary record of the Epiligurian piggy-back basins.

Figure 24 – Examples of type 6c2, epi-nappe tectono-sedimentary *mélanges* in the Molise region, Central Apennines, Italy. A- Large block (arrow) of calcareous mudstone and calcarenite with interbeds of black chert levels embedded in the *argille scagliose* (AS) of the Sannio Nappe at La Civita di Duronia (NW of Campobasso); B- Blocks that are tens of meters in size (arrows) and made of Miocene biocalcarenes are embedded in the *argille scagliose* matrix (AS), Castelbottaccio (Biferno River valley, NNE of Campobasso).

Figure 25 – Epi-nappe diapiric *mélange* (type 6c3 *mélange*) from the Moncucco T. se quarry, Tertiary Piedmont Basin, NW-Italy (late Tortonian – early Messinian marls). A few decimeter-wide marginal zone of anisotropic mud breccias, characterized by fluidal features, separates the shear zone from the host rocks (Messinian primary evaporites). See Dela Pierre et al. (2007); Festa (2009) for major details.

TABLE CAPTIONS

Table 1 – Recurrent terms in the geological literature for “olistostrome”, “*wildflysch*” and sedimentary *mélanges*. Authors who have used these terms are quoted in the text. Key to lettering: A: Austria; C: Crete; CH: Switzerland; E: Spain; F: France; GR: Greece; I: Italy; Mo: Morocco; SLO: Slovenia; T: Turkey; USA: United States of America. Modified after Camerlenghi and Pini (2009).

Table 2 – Recurrent terms in the geological literature for tectonic “*mélange*”.

Table 3 - Proposed subdivision and classification of *mélanges* on the basis of processes and tectonic settings of their formation.

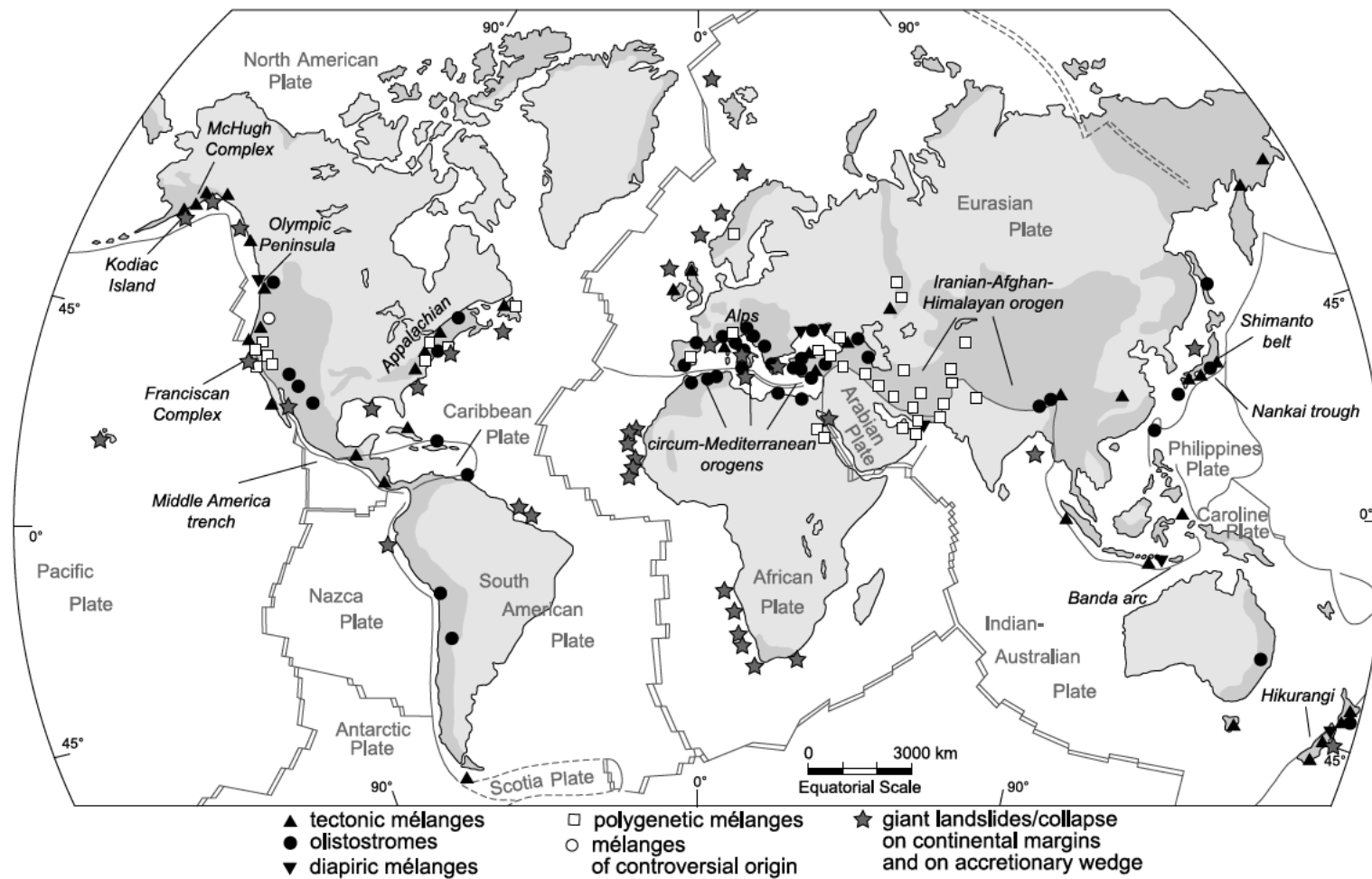


Figure 1 -

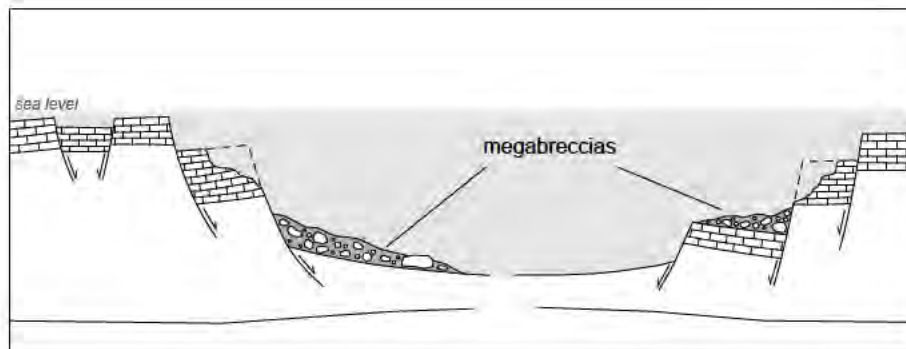


Figure 2 -

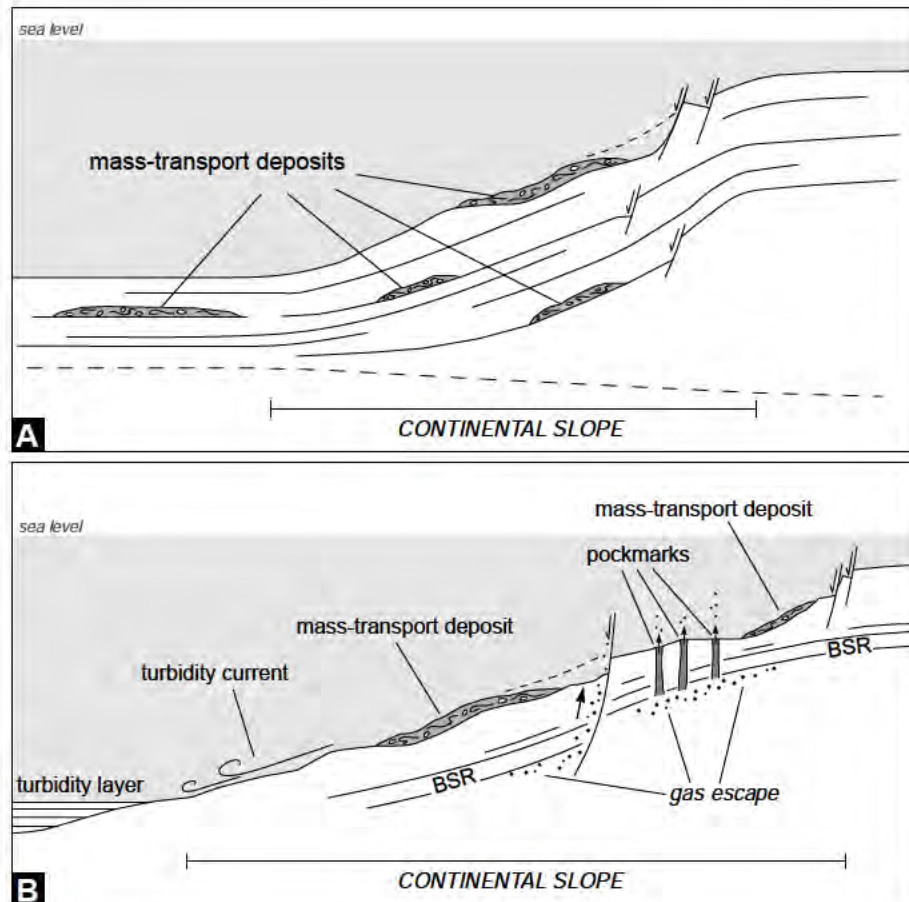
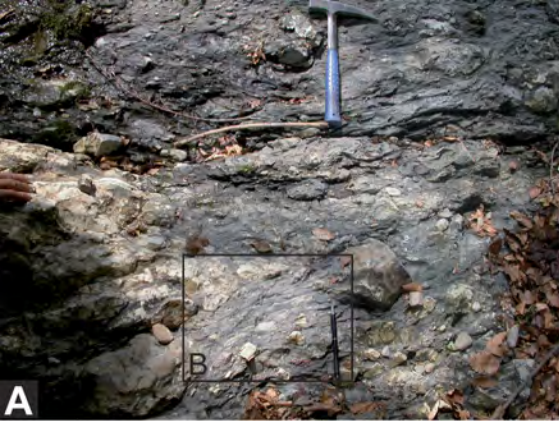


Figure 3 -



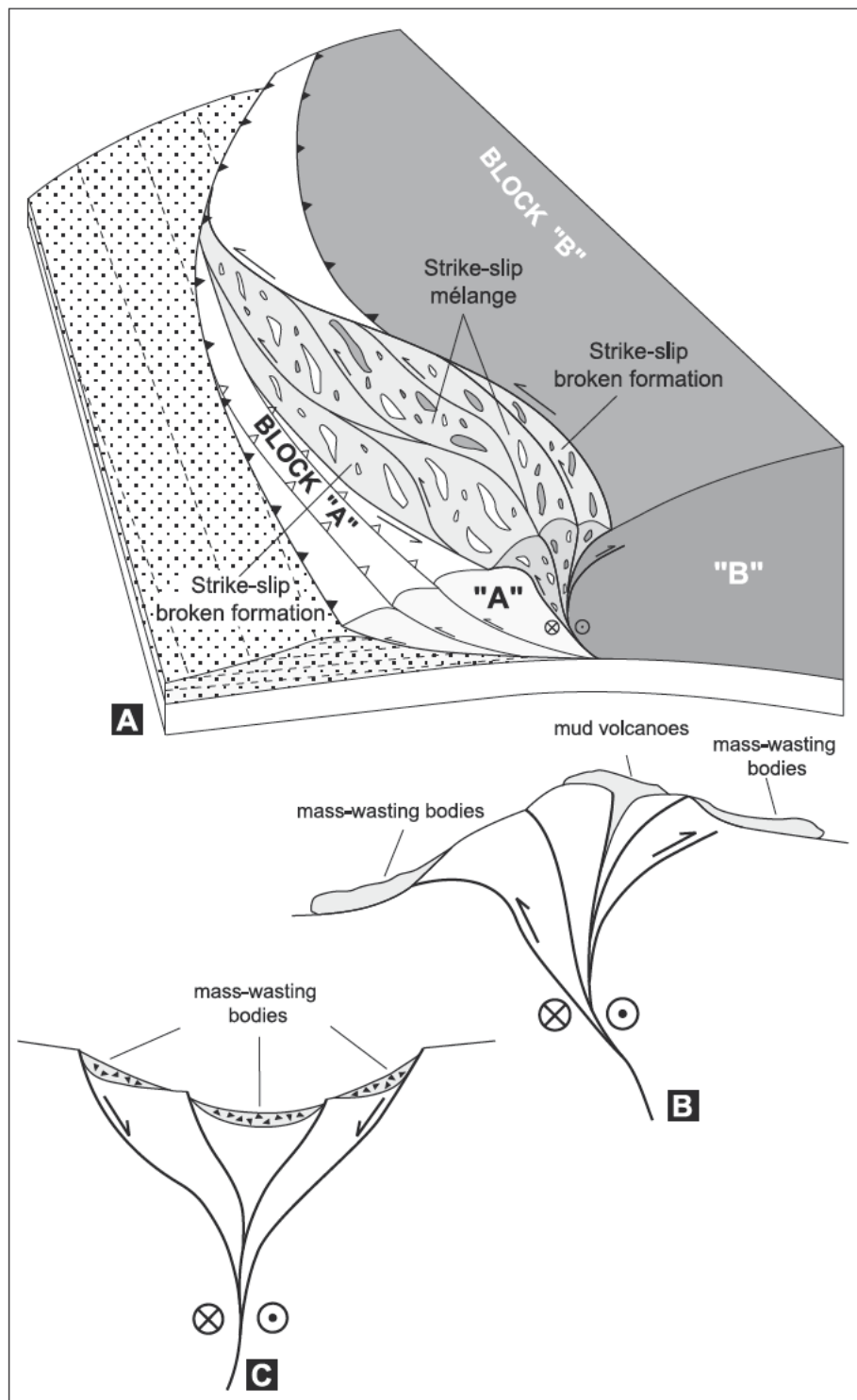


Figure 5 -



Figure 6 -

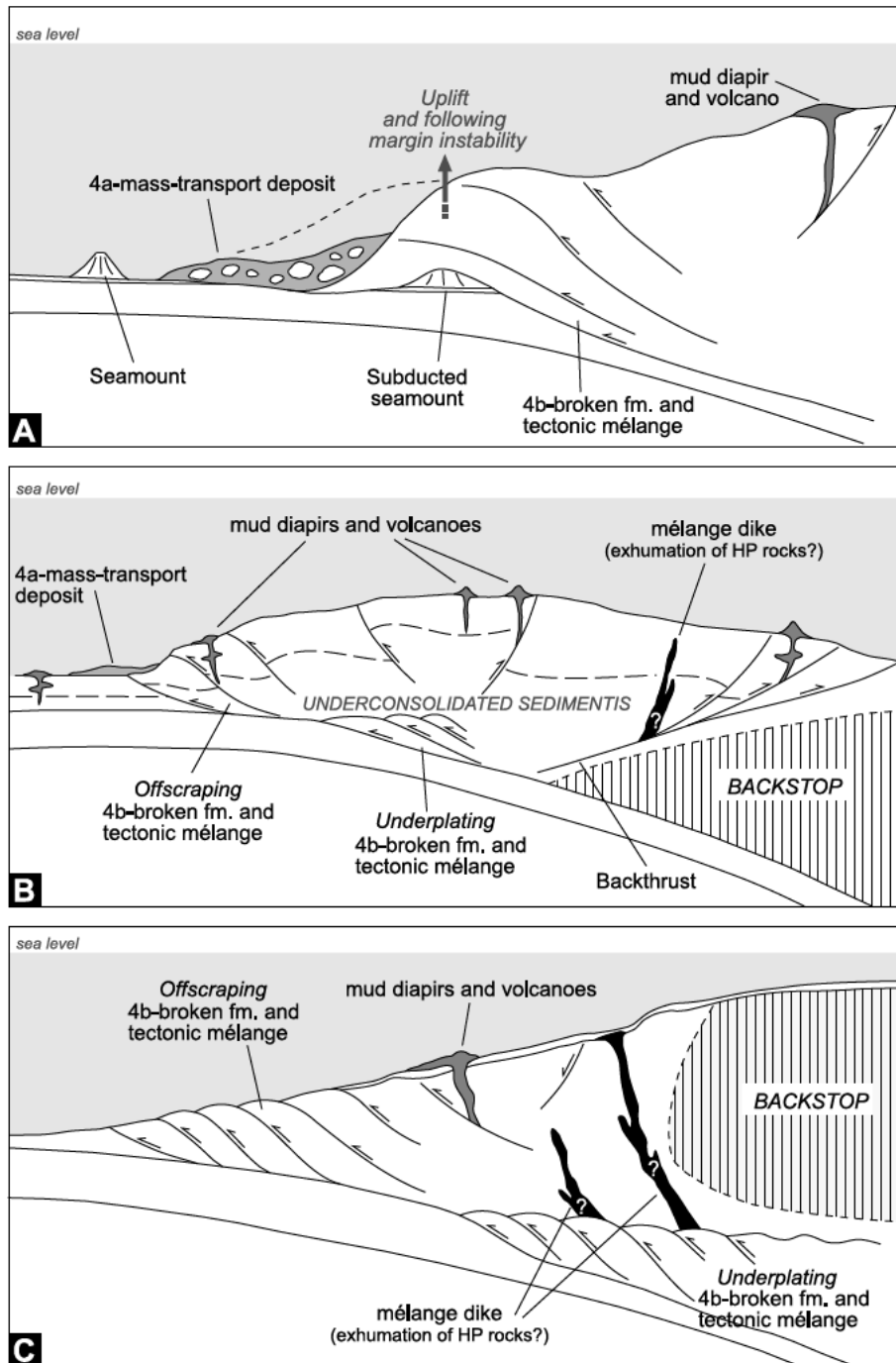


Figure 7 -



Figure 8 -

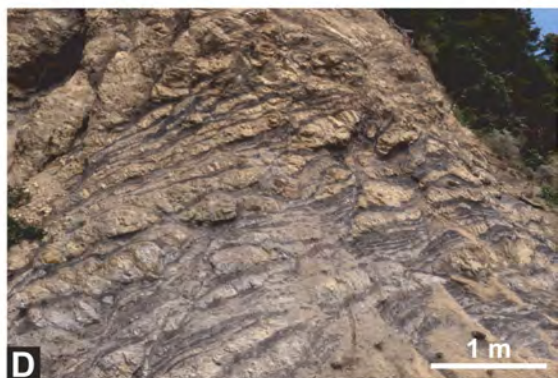


Figure 9 -



Figure 10 -



Figure 11 -



Figure 12 -



Figure 13 -



Figure 14 -

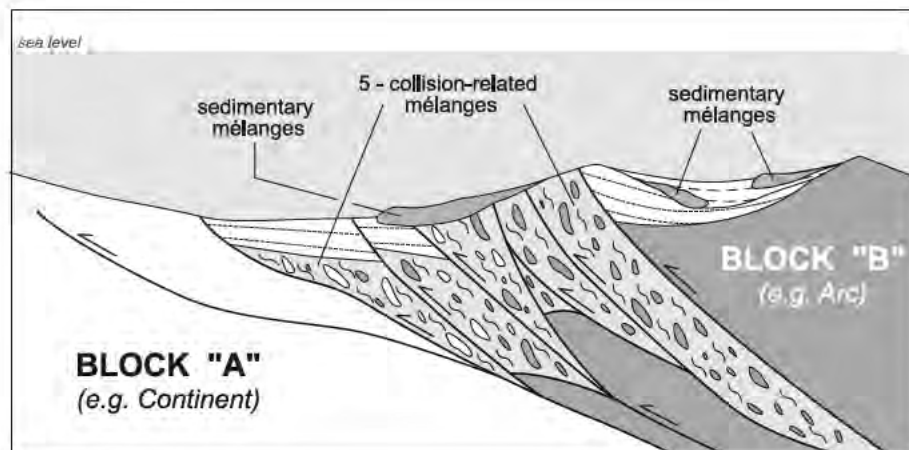


Figure 15 -



Figure 16 -

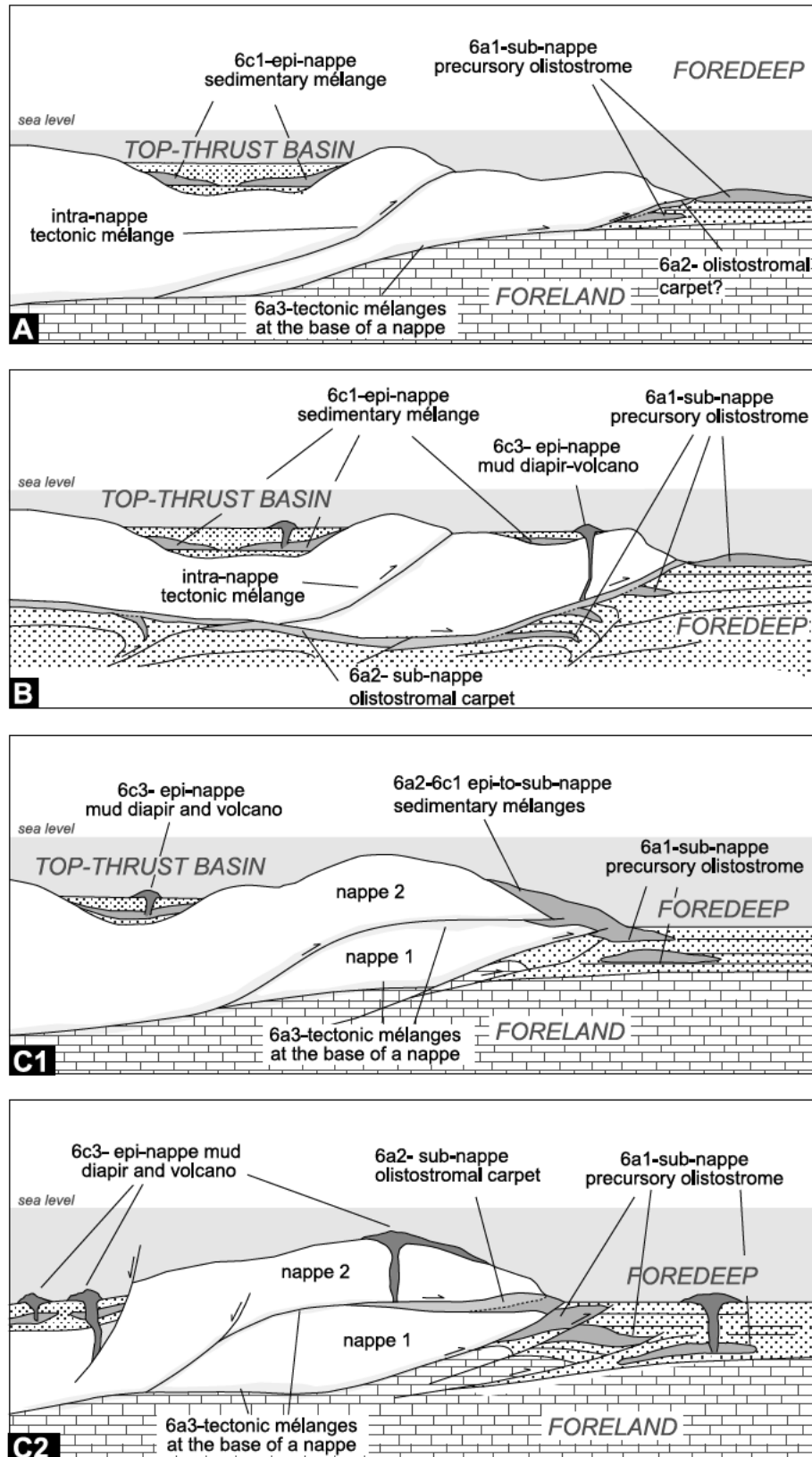


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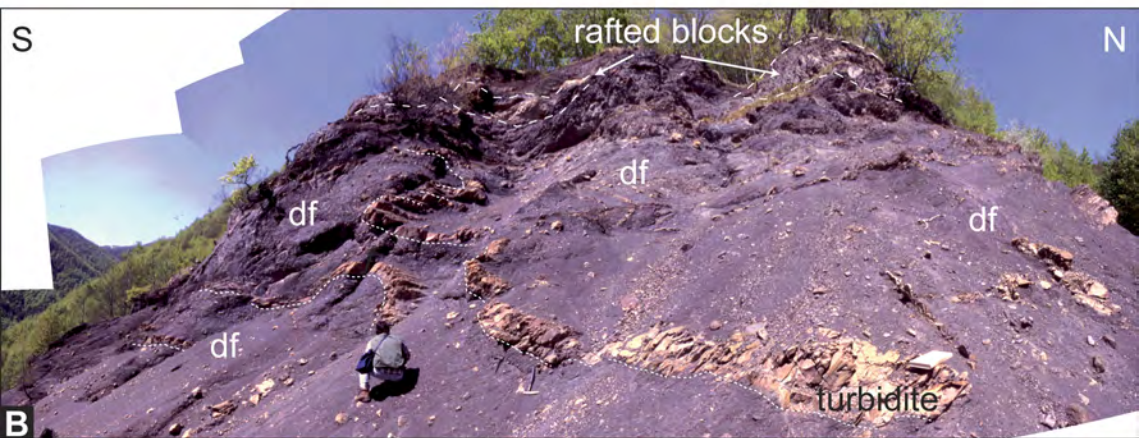
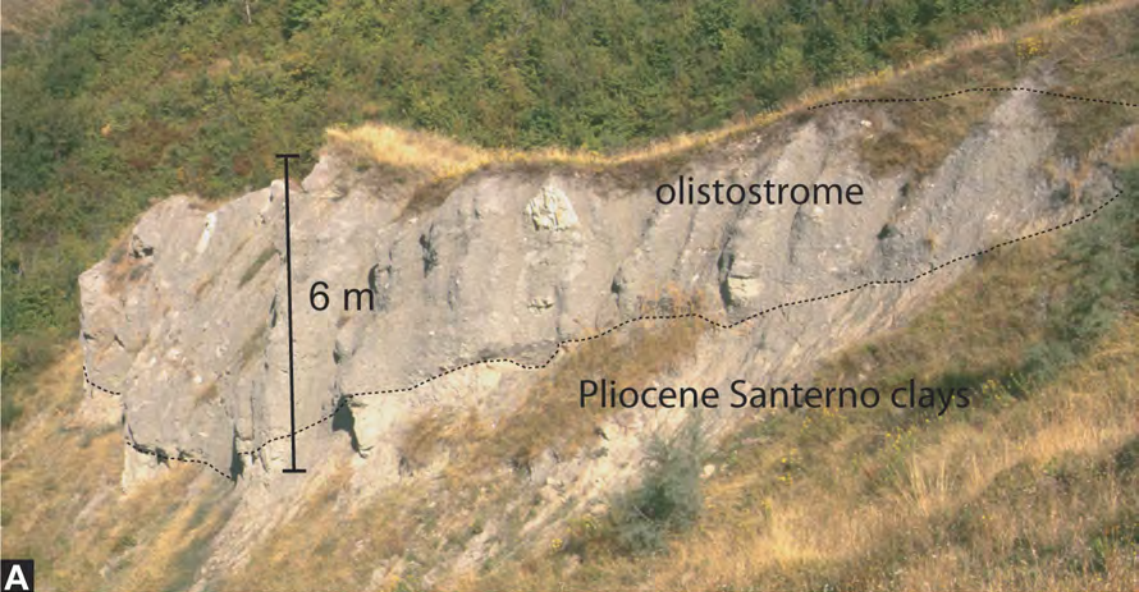


Figure 18 -

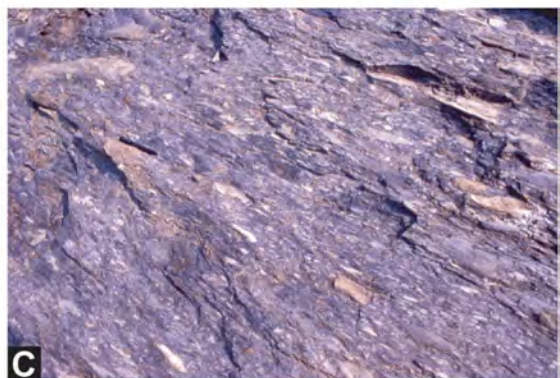
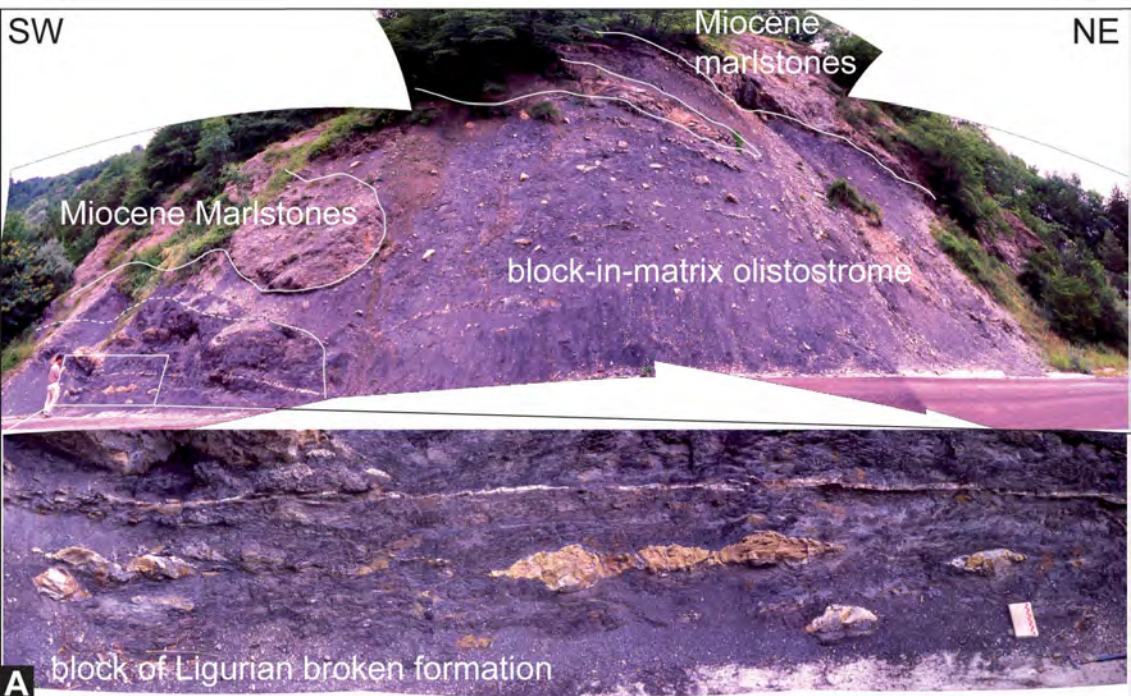


Figure 19 -



Figure 20 -



Figure 21 -

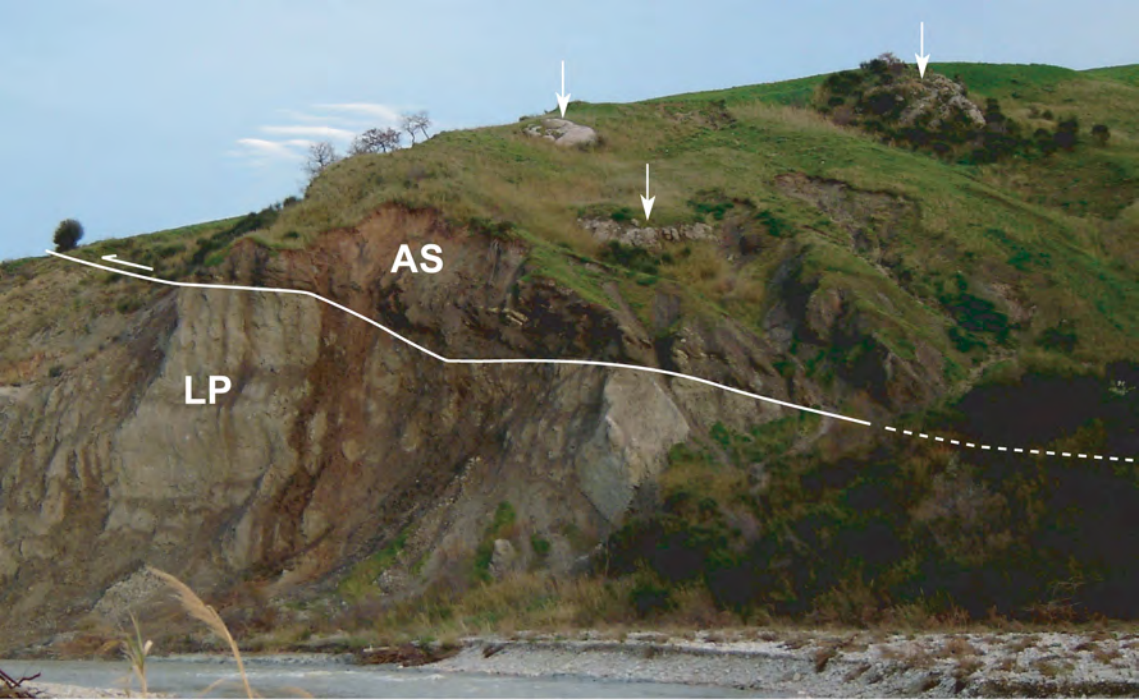


Figure 22 -

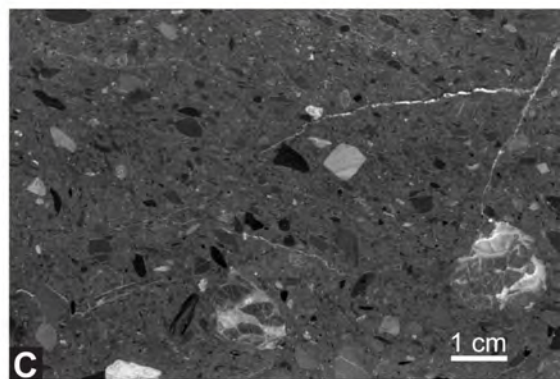
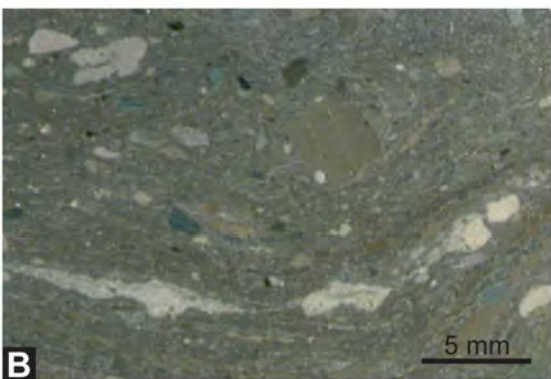


Figure 23 -

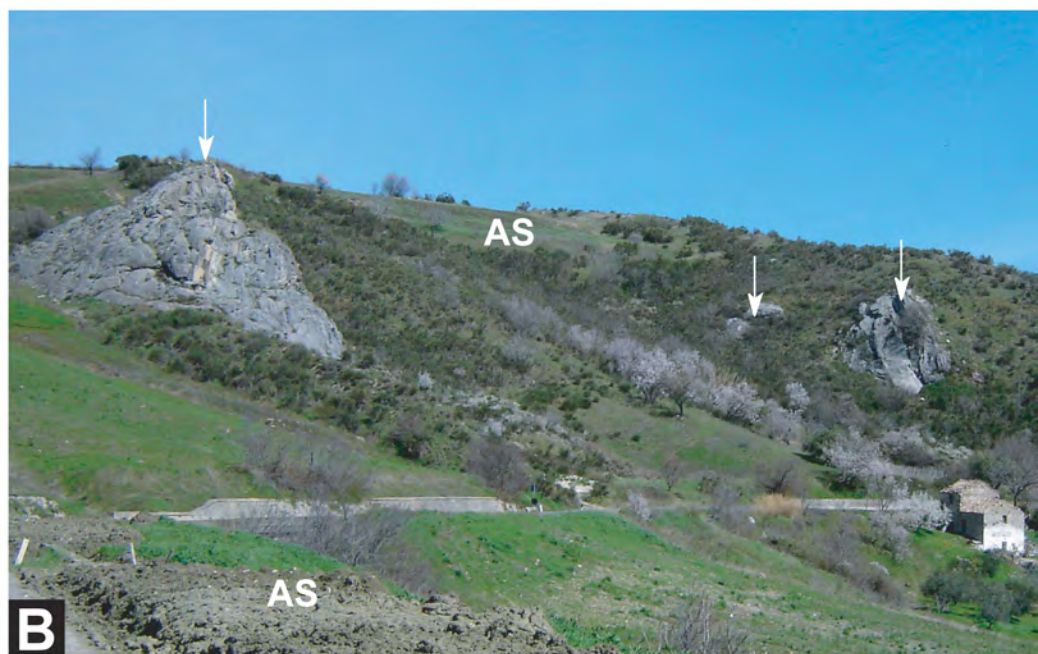


Figure 24 -

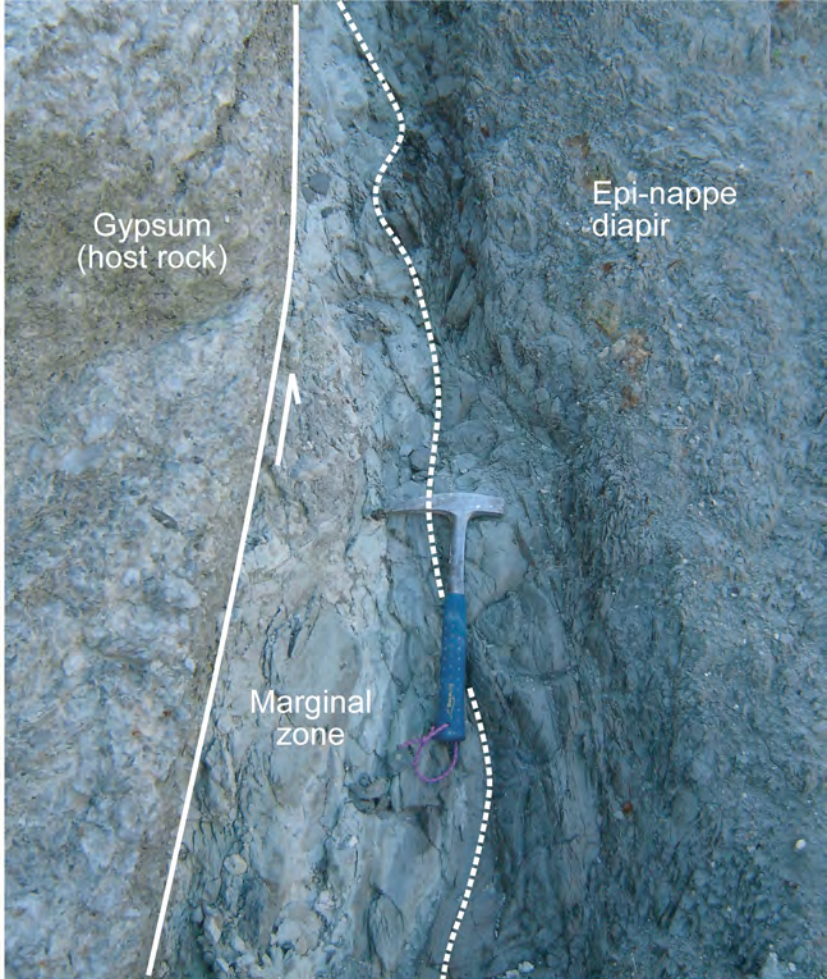


Figure 25 -

Term	Geographic location	References
Wildflysch	Habken Valley (CH) Swiss Alps (CH) Taconic region, N-Appalachian (USA) Eastern Alps, Tauern windows (A, I) Taurides, Anatolides, Pontides (T), Chios Island (GR) Western Alps, Haute Savoie (F)	Kaufmann (1886) Lugeon (1916), Trümpy (1960) Bird (1963, 1969) Frisch (1984) Papanikolaou and Sideris (1983), Groves et al. (2003) Piguet et al. (1998)
Olistostrome	Sicily (I)	Flores (1955)
Olistostromal complex/formation/unit/lens/body	Northern Apennines (I) Sicily and Sardinia (I) Betic Cordillera, Guadalquivir Basin (E) Taurides, Anatolides, Pontides (T)	Labesse (1963), Papani (1963), Sestini (1968) Coggi (1967) Berastegui et al. (1998) Yilmaz et al. (1997a, b)
Endolistostrome, allolistostrome	Northern Apennines (I)	Elter and Raggi (1965)
Megabreccia	Southern Alps, Karawanken (A, I, SLO) Western Alps, Rhone graben (F, I, CH) Southern Apennines (I) Southern Alps (I)	Castellarin (1972) Gigot (1973) Bosellini et al. (1977, 1993, 1999), Graziano (2001) Bosellini (1984, 1998)
Argilles à blocs	Betic Cordillera, Guadalquivir Basin (E)	Bourgeois et al. (1973)
Exolistoliths	Moratala, Murcia (E)	Hoedemaeker (1973)
Precursory olistostromes	Northern Apennines (I)	Elter and Trevisan (1973)
Marnes à blocs	Rif Chain (Mo)	Lespinasse (1977)
Sedimentary mélange	Northern Apennines (I)	Bettelli and Panini (1985, 1992)
Types I and II mélange	Franciscan complex, California (USA)	Cowan (1985)
Schistes à blocs	Western Alps (F, I)	Wazi et al. (1985)
Flysch à olistolites or olistolithic flysch	Montagne Noire (F)	Feist and Galtier (1985)
Argillaceous breccia; polygenetic argillaceous breccia	Northern Apennines (I)	Bettelli et al. (1994, 1996a,b)
P/PO/POA/OA type olistostromes	Central Tethyan Himalayas (Tibet)	Liu and Einsele (1996)
Olistostromal mélange	Taurides, Anatolides, Pontides (T)	Yilmaz et al. (1997b)
Olistostromal deposits	Taurides, Anatolides, Pontides (T)	Yilmaz et al. (1997a)
Olistostromal breccia	Soca Valley (SLO)	Pejovic (1997)
Olistostromal flysch	Hellenides, Aegean Islands (GR, C) Taurides, Anatolides, Pontides (T)	Clift and Dixon (1998) Atabey and Aktırmur (1997)
Agglomerates	Southern Alps, Karawanken (A, I, SLO)	Castellarin et al. (1998, 2004) and references therein
Type A, B, and C olistostromes	Northern Apennines (I)	Pini (1999), Lucente and Pini (2003)
Olistolith field	Eastern Alps, Vienna Basin (A)	De Ruig (2003)
Olistostromal carpet	Northern Apennines (I)	Landuzzi (2004), Pini et al. (2004), Camerlenghi and Pini (2009)
Blocky flysch; flysch with blocks	Taurides, Anatolides, Pontides (T)	Şenel and Aydal (1997a, b)

Table 1 -

<i>Term</i>	<i>Geographic location</i>	<i>References</i>
Mélange/autoclastic mélange	Mona complex, Anglesey (North Wales)	Greenly (1919)
Chaos structure	Franciscan complex, California (USA)	Noble (1941)
Colored mélange	Iran	Gansser (1955)
Broken formation	Franciscan complex, California (USA)	Hsü (1968)
Polykinematic mélange	Franciscan complex, California (USA)	Hsü (1968)
Ophiolitic mélange	Ankara mélange (Turkey)	Gansser (1974)
Obduction mélange	Middle East to Himalayan regions	Gansser (1974)
Dismembered formation/complex	Review of different studied cases	Raymond (1984)
Tectonic mélange	Review of different studied cases	Raymond (1984)
Type IV mélange	Franciscan complex, California (USA)	Cowan (1985)
Tectonites	Northern Apennines (Italy)	Castellarin and Pini (1987)
Asymmetric mélange	Northern Australia Kodiak Is., Alaska (USA)	Hammond (1987) Fisher and Byrne (1987)
Sheared mélange	Shimanto belt (Japan), South-Scotland	Needham (1995)
Tectonosomes	Northern Apennines (Italy)	Pini (1999)
Pre-Cambrian mélange	Appalachian (USA)	Different Authors (see for example Rast and Kohles, 1986; Bailey et al., 1989)

Table 2 -

<i>related to:</i>	<i>Types of Mélange</i>		<i>Geodynamic environments</i>	<i>Processes</i>	<i>Products</i>	<i>Minor related products</i>
1. Extensional tectonics			Rifting	Gravitational	Mass-transport deposits (megabreccias, breccias, olistoliths, olistolith fields or swarm, debris avalanches and flows, etc.)	Fault zones along normal fault?
2. Passive margins			Passive margins (after rifting)	Gravitational	Poorly sorted olistostromes (soft sediment deform.; progressive deformation from slumping to debris flows, to complete strata disruption); slides	In-situ fluidification; mud diapirs?
3. Strike-slip tectonics			Different types of collision	Tectonic	Broken formations; mélanges (exotic blocks were commonly recycled from other previously formed mélanges)	Olistostromes s.l.; mud diapirs s.l.
4. Subduction	a. Mass-transport deposits at the wedge front		Subduction (at the front of the wedge)	Gravitational	Olistostromes and mass-transport deposits (debris flows and avalanches, slumps, slides, etc.)	Mud diapirs and mud volcanoes, serpentinite diapirs
	b. Broken fms. and tectonic mélanges		Subduction (at the base of the wedge)	Tectonic	Broken formations; mélanges? (exotic blocks were commonly recycled from other previously formed mélanges)	
5. Collision			Different types of collision	Tectonic and gravitational	Broken formations; mélanges? (exotic blocks were commonly recycled from other previously formed mélanges)	Olistostromes s.l.; diapirs s.l.
6. Intra-continental deformation	a. Sub-nappe	a1. Precursory olistostromes	At the base or front of intracontinental thrust sheets or nappes	Gravitational	Olistostromes and mass-transport deposits (debris flows and avalanches, slumps, slides, etc.)	Mud diapirs and mud volcanoes
		a2. Olistostromal carpet		Gravitational and tectonic	Mélanges (exotic blocks were commonly recycled from other previously formed sedimentary mélanges), broken formations	
		a3. Tectonic mélanges		Tectonic, gravitational		
	b. Intra-nappe	b1. Sedimentary	Within intracontinental thrust sheets or nappes	Gravitational	Olistostromes and mass-transport deposits (debris flows and avalanches, slumps, slides, etc.)	Mud diapirs and mud volcanoes
		b2. Tectonic and/or tectono-sedim.		Tectonic, gravitational	Broken formations; mélanges (exotic blocks were commonly recycled from other previously formed sedimentary mélanges)	
	c. Epi-nappe	c1. Sedimentary	A top of intracontinental thrust sheets or nappes (e.g. piggy back, top thrust basins)	Gravitational	Olistostromes and mass-transport deposits (debris flows and avalanches, slumps, slides, etc.)	
		c2. Tectono-sedim.		Tectonic, gravitational	Broken formations, mélanges	
		c3. Diapiric		Diapiric	Mud diapirs and mud volcanoes	

Table 3 -