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Mélanges and mélange-forming processes: a historical overview and new concepts

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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. A 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. On the basis of 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 intracontinental 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 (French – 'mixtures') 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 because of high stratal disruption, these units are interpreted as 'chaotic'. Mélanges represent a

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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 (Talbot and von Brunn 1989; Maltman 1994; Maltman and Bolton 2003); the latter strongly affects underconsolidated rocks (Camerlenghi and Pini 2009). Cataclastic deformation of already cemented rocks, also with an important role of fluid overpressure and hydrofracturing, and ductile deformation because of inter-intra crystalline plasticity are involved in those mélanges that form in deeper crustal levels within various tectonic settings (Wilson *et al.* 1989; Polat and Casey 1995; Kusky and Bradley 1999; Jongens *et al.* 2003; Federico *et al.* 2007; Ujiie *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 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, and agglomerates, which mostly refer to block-in-matrix rocks associated with sedimentary, tectono-sedimentary, and/or diapiric origins (Camerlenghi and Pini 2009 and references therein). Many of these terms (i.e. *wildflysch*, *schistes à blocs*, olistolith field, and megabreccias) were derived from Alpine concepts, and descriptions of the block-in-matrix rocks largely come from the supra-Helvetic, ultra-Helvetic 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, and diapiric) indicating the relative mélange-forming processes.

In this article, 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 sections of this article, we refer frequently to sedimentary, tectonic, and/or diapiric processes. These terms can be ambiguous, as they refer to processes that commonly overlap and as their original definition may have changed over time because of the progressive increase of the geological knowledge (Alonso *et al.* 2006). The adjective 'sedimentary' in this article refers to all the processes of slope failure, sediment transport, and deposition, with 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 a basin slope (e.g. creeping, gravitational spreading, etc.). These processes are currently considered to be 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 to control 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 into 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 towards 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 definitions 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 the United Kingdom. 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 (Şengör 2003 and references therein), the concept of mélange was first 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 (Fuchs 1877a, 1877b; 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). Despite the strict association of *wildflysch* with regional shear zones at the base of 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 follows (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; Debelmas and Kerchkove 1973; Mercier

Table 1. Recurrent terms in the geological literature for 'olistostrome', '*wildflysch*', and sedimentary mélanges.

Term	Geographic location	References
Wildflysch	Habkern Valley (CH)	Kaufmann (1886)
	Swiss Alps (CH)	Lugeon (1916), Trümpy (1960)
	Taconic region, N-Appalachian (USA)	Bird (1963, 1969)
	Eastern Alps, Tauern windows (A, I)	Frisch (1984)
	Taurides, Anatolides, Pontides (T), Chios Island (GR)	Papanikolaou and Sideris (1983), Groves *et al.* (2003)
	Western Alps, Haute Savoie (F)	Piguet *et al*. (1998)
Olistostrome	Sicily (I)	Flores (1955)
Olistostromal complex/formation/unit/lens/body	Northern Apennines (I)	Labesse (1963), Papani (1963), Sestini (1968)
	Sicily and Sardinia (I)	
	Betic Cordillera, Guadalquivir Basin (E)	Coggi (1967)
		Berastegui *et al.* (1998)
	Taurides, Anatolides, Pontides (T)	Yilmaz *et al.* (1997a, 1997b)
Endolistostrome, allolistostrome	Northern Apennines (I)	Elter and Raggi (1965)
Megabreccia	Southern Alps, Karawanken (A, I, SLO)	Castellarin (1972)
	Western Alps, Rhone graben (F, I, CH)	Gigot (1973)
	Southern Apennines (I)	Bosellini *et al.* (1977, 1993, 1999), Graziano (2001)
	Southern Alps (I)	Bosellini (1984, 1998)
Argilles à blocs	Betic Cordillera, Guadalquivir Basin (E)	Bourgois *et al.* (1973)
Exolistoliths	Moratalla, Murcia (E)	Hoedemaeker (1973)
Precursory olistostromes	Northern Apennines (I)	Elter and Trevisan (1973)
Marnes à blocs	Rif Chain (Mo)	Lespinasse (1977)
Sedimentrary 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, 1996b)
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 Aktimur (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, 1997b)

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).

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 of 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 (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 1950s (e.g. 'the problem of knockers and blueschists rocks' of the workers on the Franciscan mélange; Brothers 1954; Bailey *et al.* 1964; Coleman 1971; Coleman and Lanphere 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

Table 2. Recurrent terms in the geological literature for tectonic 'mélange'.

Term	Geographic location	References
Mélange/autoclastic mélange	Mona complex, Anglesey (North Wales)	Greenly (1919)
Chaos structure	Franciscan complex, California (USA)	Noble (1941)
Coloured 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), southern 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)

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 'coloured 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 stratally 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 the use of 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 references 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 (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) have 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 (Frisch 1984; Ring *et al.* 1988, 1989; Winkler 1988) 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 ultra-helvetic) 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.* 2010 and references therein). Then, many terms (i.e. *wildflysch*, *schistes à blocs*, olistostromes, *argille brecciate*, *argille scagliose*, olistholith 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 (Hsü 1968, 1974; Cowan 1985; Camerlenghi and Pini 2009; Festa *et al.* 2010). This led to a re-evaluation of the virtual association of mélanges with subduction zone processes (Cowan 1985).

During this process of re-evaluation (Raymond 1984), the interest of the scientific community has been addressed to: (1) establish valuable criteria for distinguishing processes of mélange formation (Naylor 1982; Byrne 1984; Cowan 1985; Orange 1990; Clennell 1992; Pini 1999; Lucente and Pini 2003), especially for polygenic mélanges (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 2010), and (2) understand the significance of mélanges and other chaotic rock units in the geological record (Gansser 1974; Raymond 1984; Cowan 1985; Scherba 1989; Camerlenghi and Pini 2009; Festa *et al.* 2010).

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 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 to be 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 subjects of a scientific controversy (Ogniben 1953; Beneo 1955, 1956a). Similar chaotic deposits were described in the Alps by the first supporters of the emerging nappe concept (Schardt 1898; Lugeon 1916; Tercier 1947) as a result of submarine sliding, producing a mixture of sediments and exotic blocks originating from the front of the advancing nappes (see Mutti *et al.* 2009 for major details). These chaotic deposits, described as *wildflysch*, may be described as an '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 hyperconcentrated flows (Lucente and Pini 2003). A broad definition of olistostromes includes bodies with blocks ranging in size from a centimetre to a metre, 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 metres sustained by a matrix of the Type A olistostrome (Type B olistostrome, Pini 1999), and 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 (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 (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 a 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 millimetre-to-centimetre-scale clasts of claystone, randomly enveloped in 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 (millimetre-to-centimetre-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 (Signorini 1946; Azzaroli 1948; Merla 1952; Ogniben 1953; Maxwell 1959; Elter 1960) and throughout the world (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 Meso-Cenozoic 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 has 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 (Camerlenghi and Pini 2009; Festa *et al.* 2010). 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 Figure 5 in Festa *et al.* 2010).

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 (Figure 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 (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 (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;

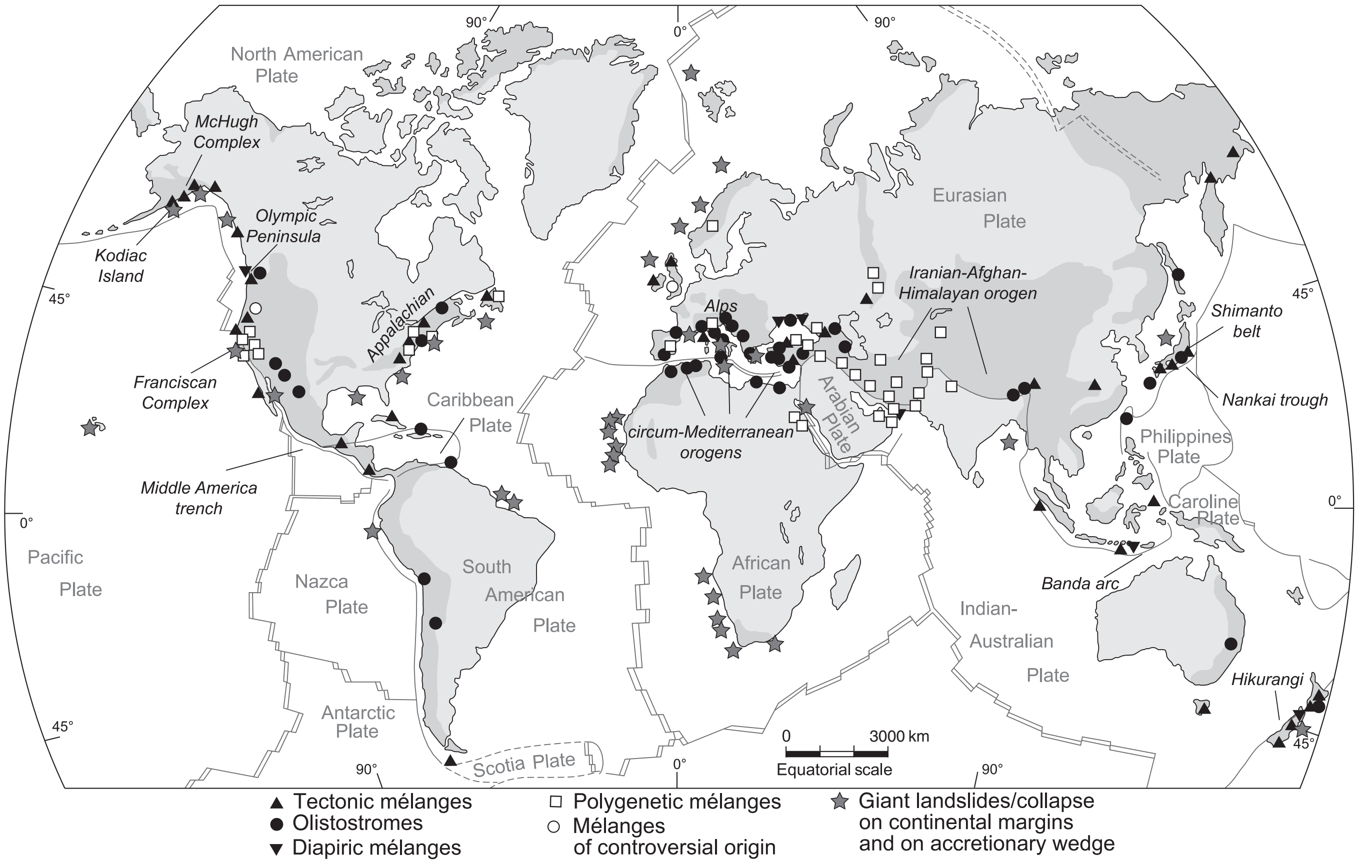

Figure 1. Global distribution of selected mélanges *sensu latu* related to contractional tectonics, sedimentary and (mud) diapiric processes (after Raymond 1984, redrawn and modified), and giant submarine landslides (after Mienert *et al.* 2003).

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.* 2010; see also Camerlenghi and Pini 2009 and Festa *et al.* 2010 for a complete review), and Iranian–Afghan–Himalayan orogens (Gansser 1974 and references therein; Burg *et al.* 2008; Dilek *et al.* 2009).

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 post-collisional stages. Some examples of these olistostromes and olistoliths widely occur in the Alps (Bernoulli and Jenkyns 1970; Castellarin 1972; Achtnich 1982; Frisch 1984; Wächter 1987; Eberli 1988; 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; Festa *et al.* 2010 and references therein), and the Iranian–Afghan–Himalayan orogens (Liu and Einsele 1996; Burg *et al.* 2008; Dilek *et al.* 2010).

Similar examples of mélanges and olistostromes have been described also from other ancient collisional orogens as well as from the Appalachians (Figure 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 (Williams and Hatcher 1983). Subduction and/or thrusting during the Taconian or Alleganian orogenies commonly formed 'olistostromal mélanges' (*sensu* Rast and Horton 1989), tectonic mélanges, broken formations, and olistostromes (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 (Schroetter *et al.* 2006). Moreover, ophiolite-bearing mélanges have been interpreted as the remnants of a Taconian suture (Pollock 1989) or related to allochthonous terranes (Boone and Boudette 1989).

Accretionary orogens, extensively distributed in the circum-Pacific regions, provide some of the most outstanding examples of mélanges and olistostromes (Figure 1). The Western US Cordilliera (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 Complex and McHugh Complex), 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 (Isozaki 1997).

Different mélange and broken formations have been related to tectonic processes that occurred during the evolution of accretionary wedges (Figure 1), as in the Franciscan Complex and in the Western US Cordillera (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 (Moore and Wheeler 1978; Orange 1990; Bradley and Kusky 1992; Kusky and Bradley 1999), Shimanto (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 (Hsü 1973) or as tectonic erosion artefacts in accretionary prisms closer to the trench (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 (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 (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; Shibata and Hashimoto 2005 and reference therein; Figure 1 and Table 2).

Accretionary orogens are also the place for olistostrome formation (Figure 1) that commonly occurs at the front and at the base of accretionary wedges as described, for example, from the Franciscan Complex and the Western US Cordillera (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 (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 (Figure 1). The Barbados accretionary prism (Westbrook and Smith 1983; Brown and Westbrook 1988), Timor and Indonesia along the Banda Arc–Australia accretionary and collisional complex (Barber *et al.* 1986; Barber and Brown 1988), the Eastern Mediterranean ridge (Cita and Camerlenghi 1990; Kopf *et al.* 1998; Camerlenghi and Pini 2009 and references therein), and the Hoh accretionary complex in the Olympic Peninsula (Orange 1990) are a 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 (Camerlenghi and Pini 2009). Although, rather rare, some fossil examples of mud volcanoes and diapirs (both mud, serpentinite and metamorphic diapirs) have been reported (Figure 1) from the Alps and Alpine basins (Ernst 1965; Di Giulio 1992; Clari *et al.* 2004; Dela Pierre *et al.* 2007; Festa 2010), the Western US Cordillera (Orange 1990), the Appalachians (Lash 1987), and New South Wales in SE Australia (Fergusson and Frikken 2003). Other examples may be found in the Palaeozoic 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 latu* 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-palaeogeographic 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.* (2010), 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 presents a brief discussion on the main characters of some of the more emblematic cases.

Type 1 – mélanges related to extensional tectonics

Mainly angular clasts, ranging in size from decimetre to several tens of metres, 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 intrabasinal composition. Formation processes of these sedimentary chaotic bodies are consistent with *en mass* gravitational movements (debris avalanches and flows and sliding of blocks; Figure 2), and the bodies are also known in the geological literature as megabreccias, olistolith, and olistolith 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 (Figure 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 mechanism affecting the margins of carbonate platforms during and after rifting (Figure 2), although the initial geometry of carbonate platform margins may cause similar bodies to develop without the contribution of active tectonics (scalloped margins; *sensu* Bosellini 1998).

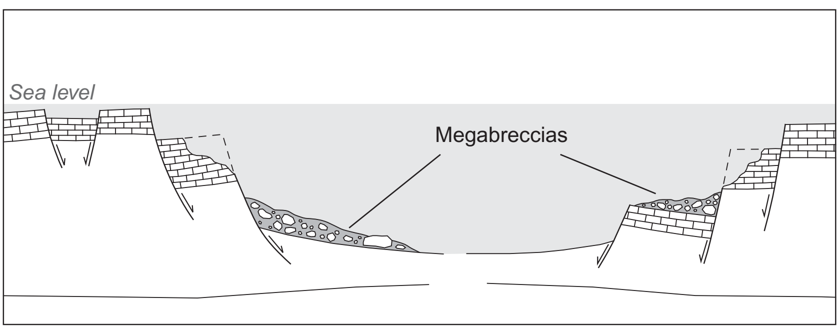

Figure 2. Conceptual model for the formation of Type 1 mélanges associated with extensional tectonics, depicting a representative case of megabreccia accumulation triggered by normal faulting within an extended carbonate platform.

The most notable examples are the large Middle Triassic–Jurassic and Cretaceous megabreccias in the Southern Alps (Castellarin 1972; Bosellini *et al.* 1977) and the Early Jurassic breccias of the Northern Calcareous Alps (Channell *et al.* 1992; Böhm *et al.* 1995; Ebli 1997; Mandl 2000). Some spectacular examples are preserved in the Northern Apennines (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 (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 (Rast and Kohles 1986; Bailey *et al.* 1989).

Type 2 – mélanges related to passive margins and ocean floor

Passive margin mélanges (Figure 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

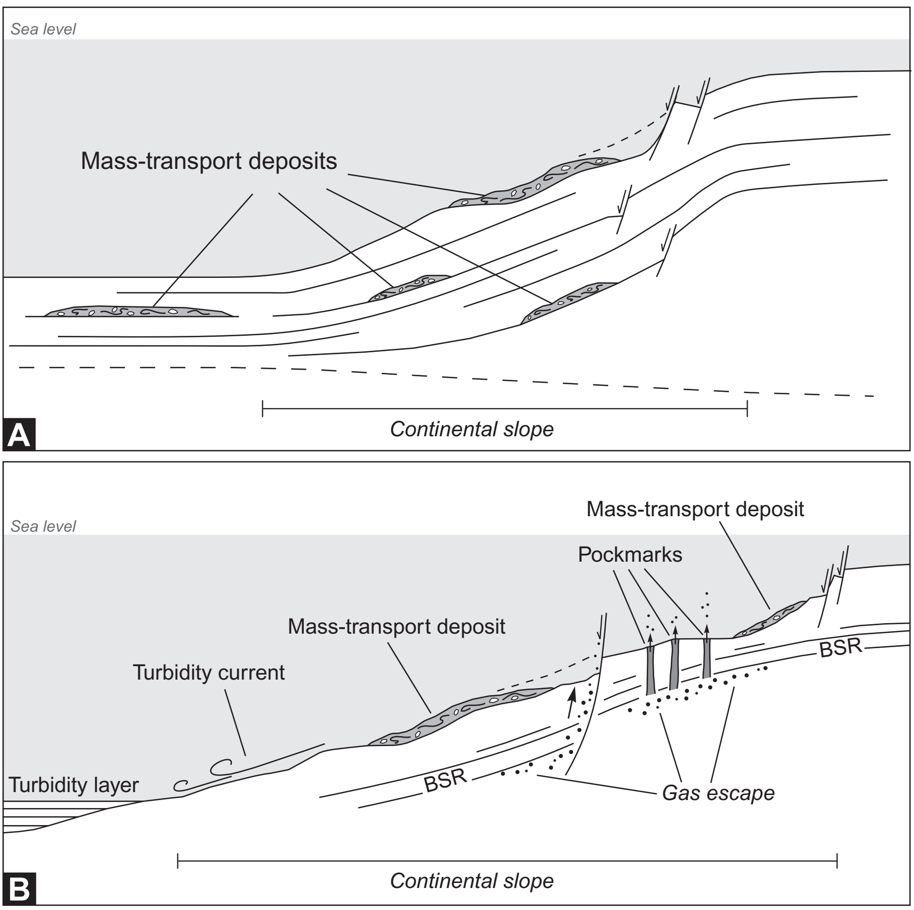

Figure 3. Conceptual model for the distribution (A) and formation (B) of Type 2 mélanges, associated with passive margin evolution. Relationships among compaction-related normal faulting, gas escape (hydrate dissociation), and mass-wasting processes and bodies are shown. BSR, bottom-simulating reflector.

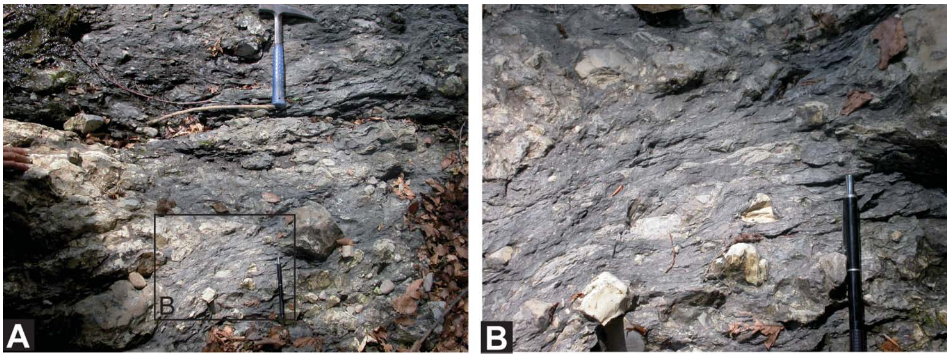

Figure 4. (A) Muddy debris flow body (olistostrome) in the Ronchi lower member ('Palombini shales', Lower Cretaceous) of the Modino basal complex, Modena area in the Northern Apennines (see De Libero 1998; Pini *et al.* 2004; and Lucente *et al.* 2006, for more details) (courtesy of C.C. Lucente). (B) 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).

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 clastic 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 (Figure 4).

The formation of this type of olistostromes is consistent with different gravity-driven processes (Mienert *et al.* 2003; Masson *et al.* 2006) of downslope transport of sediments by slides or slumps and debris flow (laminar motion) (Figure 3B). Commonly, slides and slumps evolve dynamically during downslope movements (Figure 3B), and they turn into debris flow and then into turbulent flows (turbidites) (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 (Figure 3) and difficult to interpret, 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 (Figure 3A), deep-ocean processes, and gas hydrate dissociation (because of the changes in sea level or increase in bottom water temperature) represent some possibilities of triggering mechanisms (Figure 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 (McAdoo *et al.* 2000), Norvegian margin (Bugge *et al.* 1988; Laberg *et al.* 2000; Haflidason *et al.* 2001), NW African margin (Embley and Jacobi 1977), and many other 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 (Camerlenghi and Pini 2009; Festa *et al.* 2010), 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 (Figure 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 millimetre to tens of metres 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 centimetres to some kilometres) 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 (Aoya *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 a volumetric point of view, when a seamount chain arrives at a trench and subduction zone. The actualistic examples of Hawaii and the Canary Islands show that some of the largest submarine landslides occur because of marked instability of volcano flanks (Lipman *et al.* 1988; Moore *et al.* 1989; McMurtry *et al.* 1999; Gee *et al.* 2001; Clague and Moore 2002; Mitchell *et al.* 2002).

Type 3 – mélanges related to strike-slip tectonics

This group refers exclusively to tectonic mélanges related to strike-slip deformation (Figure 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 2010): (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 tens of metres- (Festa 2009) to kilometres-wide (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 forearc 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 (<1 mm)-to-macro scale, compatible with the regional stress field (Figure 5). Blocks, ranging in size from centimetres to hundreds of metres, have elongate and lozenge shapes (Figure 6) and show polished surfaces commonly contained by slikenlines. These blocks are aligned with the main shear zones, showing an en-échelon arrangement according to the sense of shear. Riedel shears and

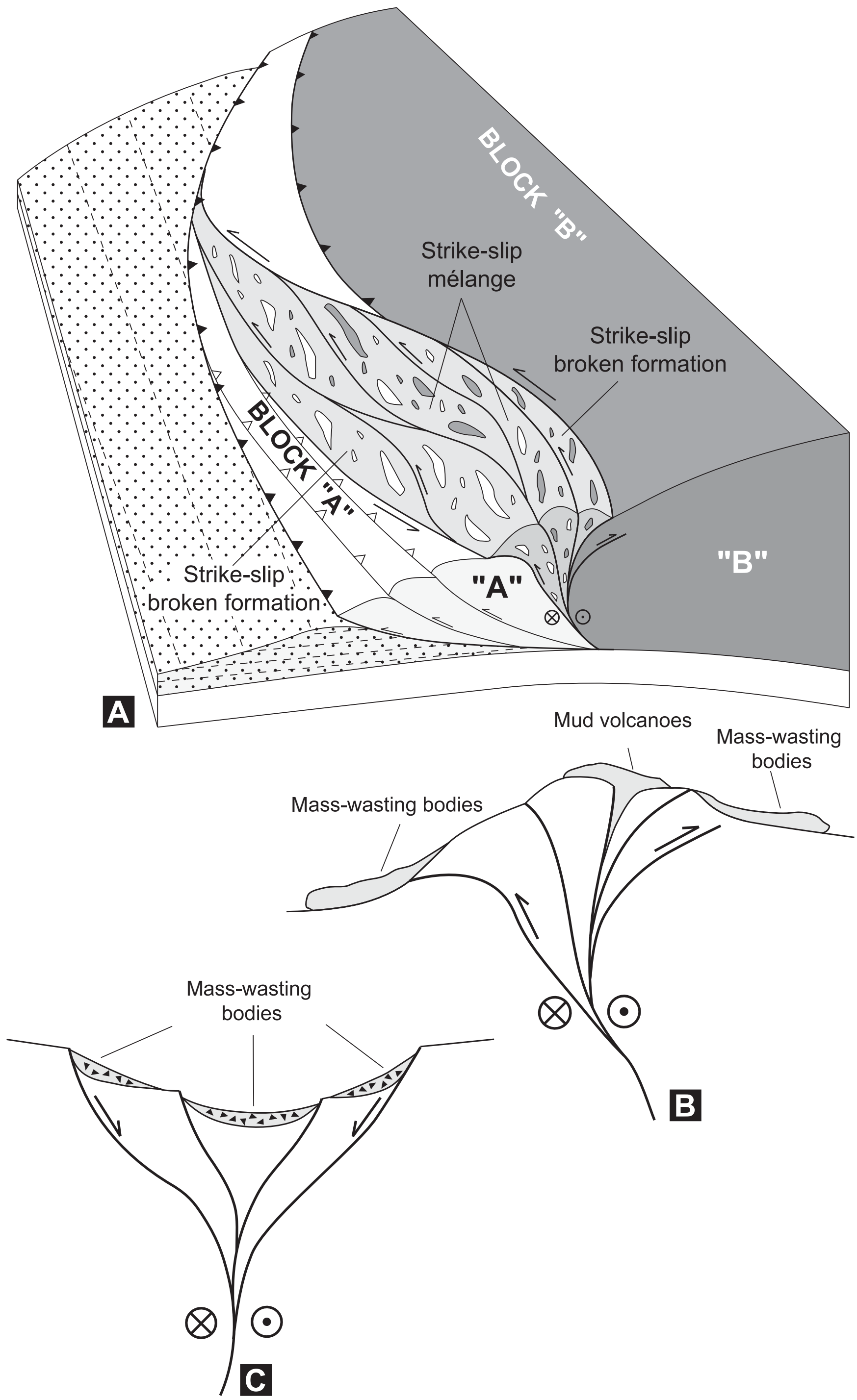

Figure 5. (A) Conceptual model for the origin of both broken formation and tectonic mélanges (Type 3 mélanges) as a result of strike-slip tectonics. Sedimentary mélanges (i.e. bodies related to diverse mass-transport events and processes) and mud volcanoes associated with positive (B) and negative (C)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

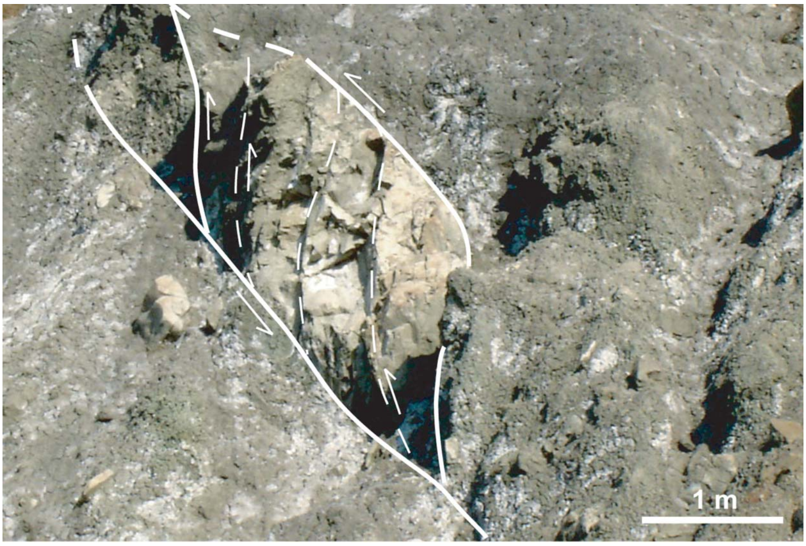

Figure 6. Strike-slip duplex structure along vertical strike-slip faults and related low-angle shears, enveloping a deformed block in a strike-slip mélange (Type 3 mélanges) within the *argille scagliose* of the Northern Apennines (west of Varzi).

S-C or P-C fabrics commonly occur within the blocks (Figure 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) (Figure 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 minds 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'), typically composed of 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 was related to significant numbers 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-pumpellyte 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 a 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 centimetres to a few kilometres 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 (Abbate *et al.* 1970; Bortolotti *et al.* 2001; Marroni *et al.* 2001; Cerrina Feroni *et al.* 2002) and the Tertiary Piedmont Basin in NW Italy (Dela Pierre *et al.* 2007; Festa 2010) and Albania (Shallo 1990; Rassios and Dilek 2009). For further details on these examples, see Camerlenghi and Pini (2009) and Festa *et al.* (2010). 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 forearc basin (Krylov *et al.* 2005).

Type 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, 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 (Figures 7 and 8). Chaotic deposits generally include deformed intrabasinal sediments (Figure 8A) and extrabasinal rocks of different ages (commonly older than the intrabasinal component and timing of their emplacement) supplied from the front of the accretionary wedge and/or wedge-top basin (Figure 7A, 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’ (Figure 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 (Chanier and

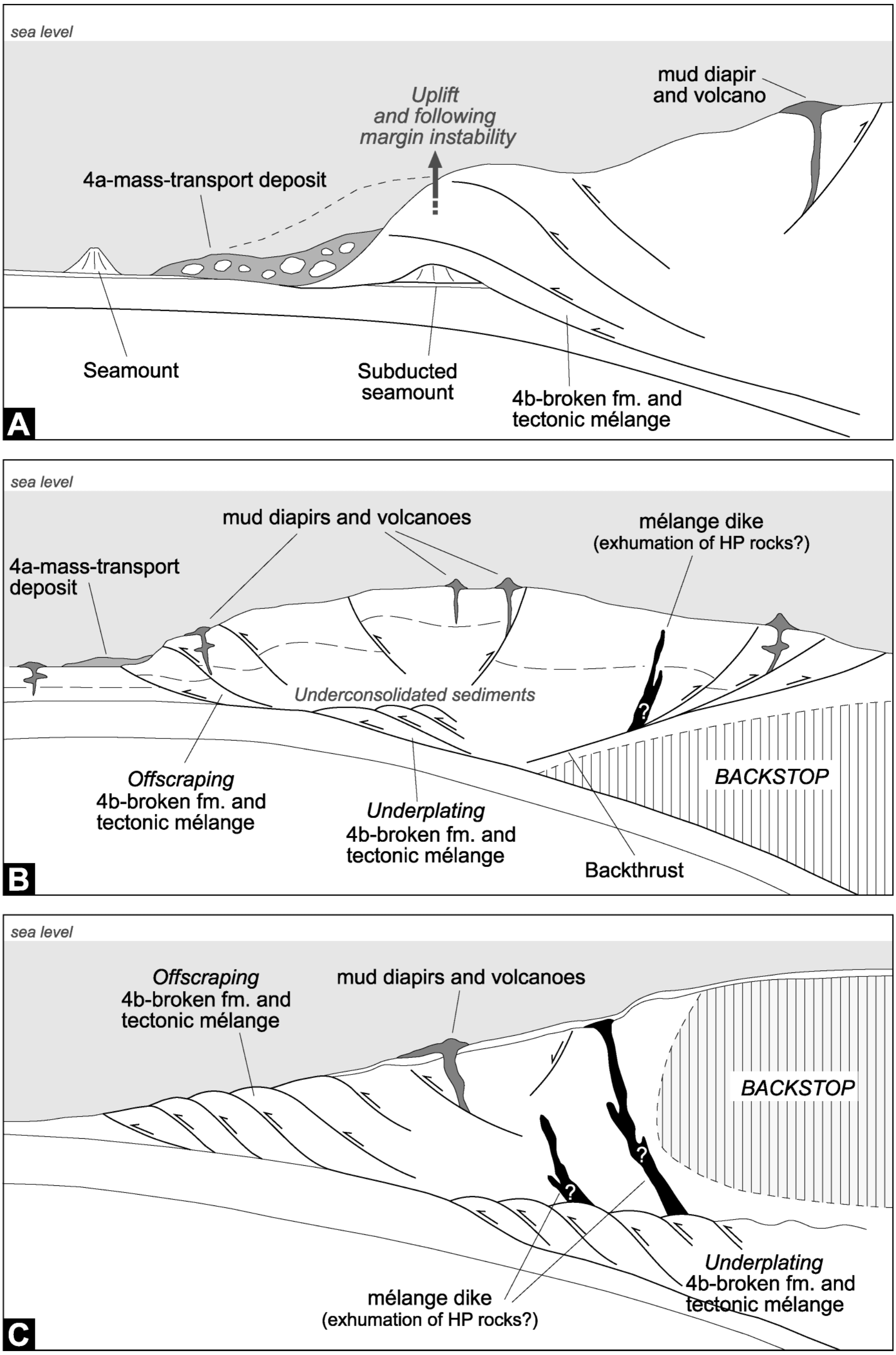

Figure 7. Conceptual models for the formation and emplacement of Type 4 mélanges associated with subduction zone processes: (A) seamount subduction; (B) large double-verging wedge, with a low elevation of the backstop; (C) smaller wedge, with a high elevation of the backstop.

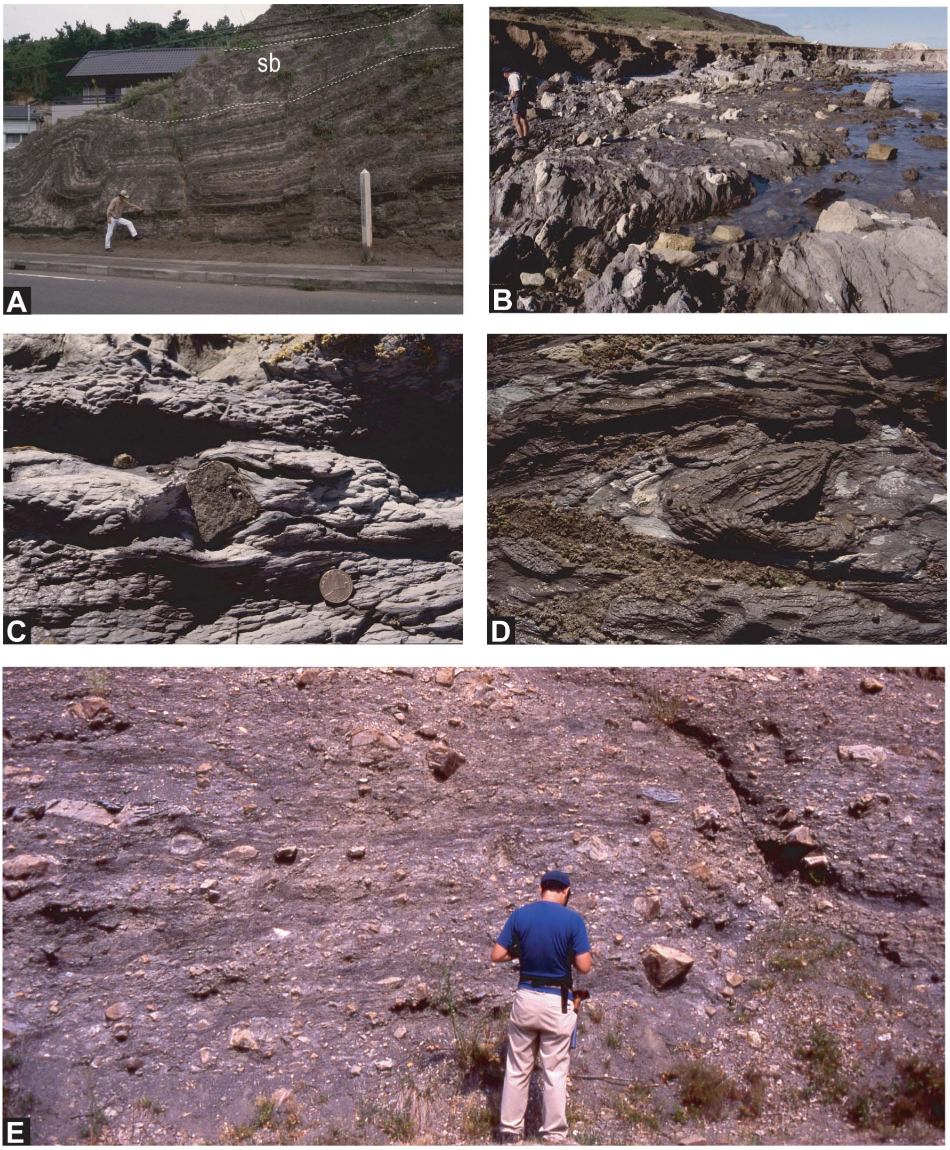

Figure 8. Different examples of mass-transport deposits associated with Type 4a mélanges: (A) Detachment fold and SE-verging thrust (the thrust surface in the lower part of the outcrop is pointed out by Yuijiro Ogawa) and a slump body (sb) in the Misaki Formation of the Miura Group (Kaitocho village, Miura Peninusula, Japan). A detailed reconstruction of this outcrop is available in Yamamoto *et al.* (2000). (B) Coastal exposure of the Pahaoa olistostrome (Glendu 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.

Ferrière 1991; Taira *et al.* 1992; Yamamoto *et al.* 2007, 2009; Figure 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 (Figure 8B, 8C, 8E), 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 of square kilometres. As implied by the name, their formation is consistent with different types of mass-transport phenomena such as sliding, slumping, debris flow, and debris avalanches and their interactions (Abbate *et al.* 1970, 1981; Chanier and Ferrière 1991; Pini 1999; Lucente and Pini 2003).

Triggering mechanisms depend on a combination of several factors (Figure 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 (Figure 7A), subduction erosion (Marroni and Pandolfi 2001; von Huene *et al.* 2004), and thrust faulting and folding (Figure 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 (Figure 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 diapirc 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 Bocco Shales in the Northern Apennines, for example, represent a chaotic product of frontal tectonic erosion (Marroni and Pandolfi 2001) of an early Palaeocene accretionary wedge slope (i.e. subduction zone of the Ligure-Piemontese ocean basin). This erosion was produced by episodic instability because of 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 (Polino 1984; Durand Delga 1986; Deville *et al.* 1992; Ducci *et al.* 1997; see also Camerlenghi and Pini 2009 and Festa *et al.* 2010). 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 of both subduction of seamounts and ridges (Hikurangi margin, New Zealand, Collot *et al.* 2001; Costa Rica continental margin, von Huene *et al.* 2004; Hühnerbach *et al.* 2005; N of d'Entrecasteaus Ridge and 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 (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 the 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 both sedimentary and metamorphic in nature (Figure 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 (Figures 9 and 10A, 10B), and in which no 'exotic' blocks are present. Beds and bed fragments show a planar orientation (Figure 9D, 9E) consistent with an internal order at the outcrop to map-scale that coincides with a structural fabric (Cowan and Pini 2001). Various types of structures may have formed from the interaction of different subduction-related processes (Figures 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 (Figure 8E) and a network of small-scale, normal faults (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 (Figure 10F) and the complete desegregation of rocks (Cousineau 1998; Yamamoto *et al.* 2009; Figure 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; Figures 11 and 12), for mesoscale pinch-and-swell and (asymmetric) boudinage of beds (Figure 9D, 9E, 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 staking and bed transposition (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-metre-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).

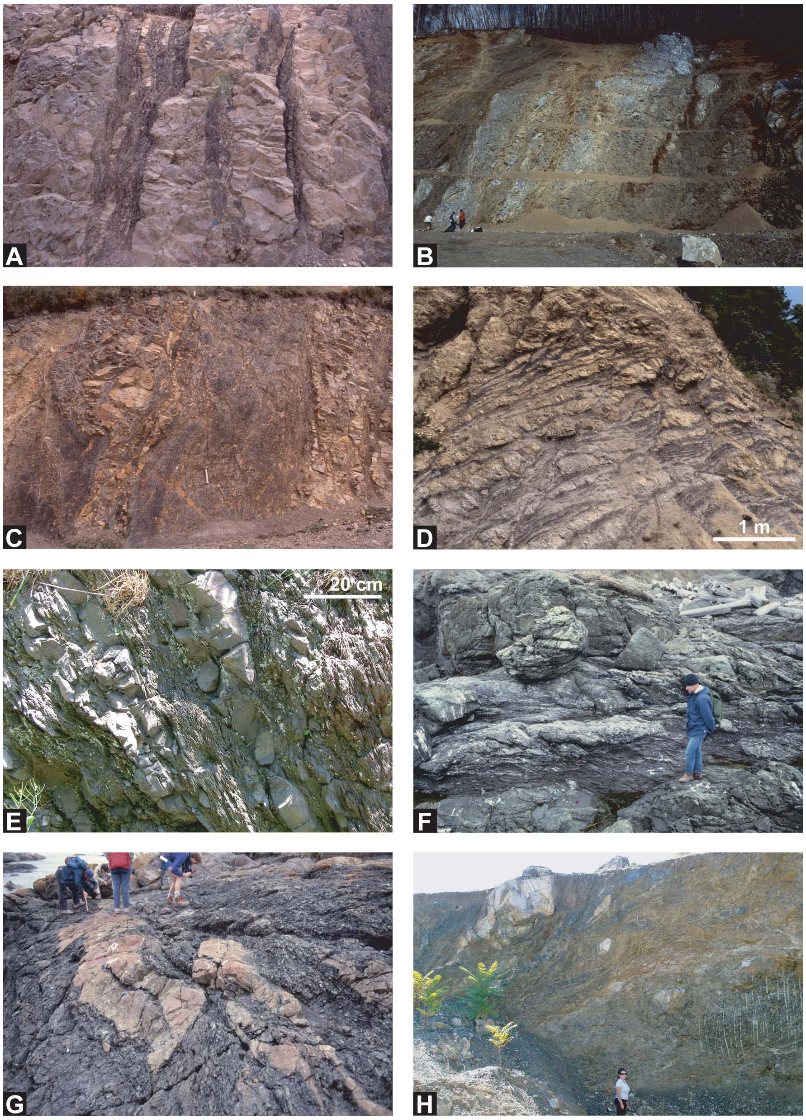

Figure 9. Examples of broken formations and Type 4b mélanges in exhumed accretionary wedges. Variously disrupted mélange units in the Tenguyama Formation (A) and Kawakami (B) Formation in the Chichibu Belt of the Kanto Mountains, Japan, in the Esk Head mélange (C), on the road from White Rock to Okuku Pass, W 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 (Lash 1987). (F) Coastal exposure of deformed rocks within the Rosario fault zone, San Juan Islands, western USA (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 Canada. 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.

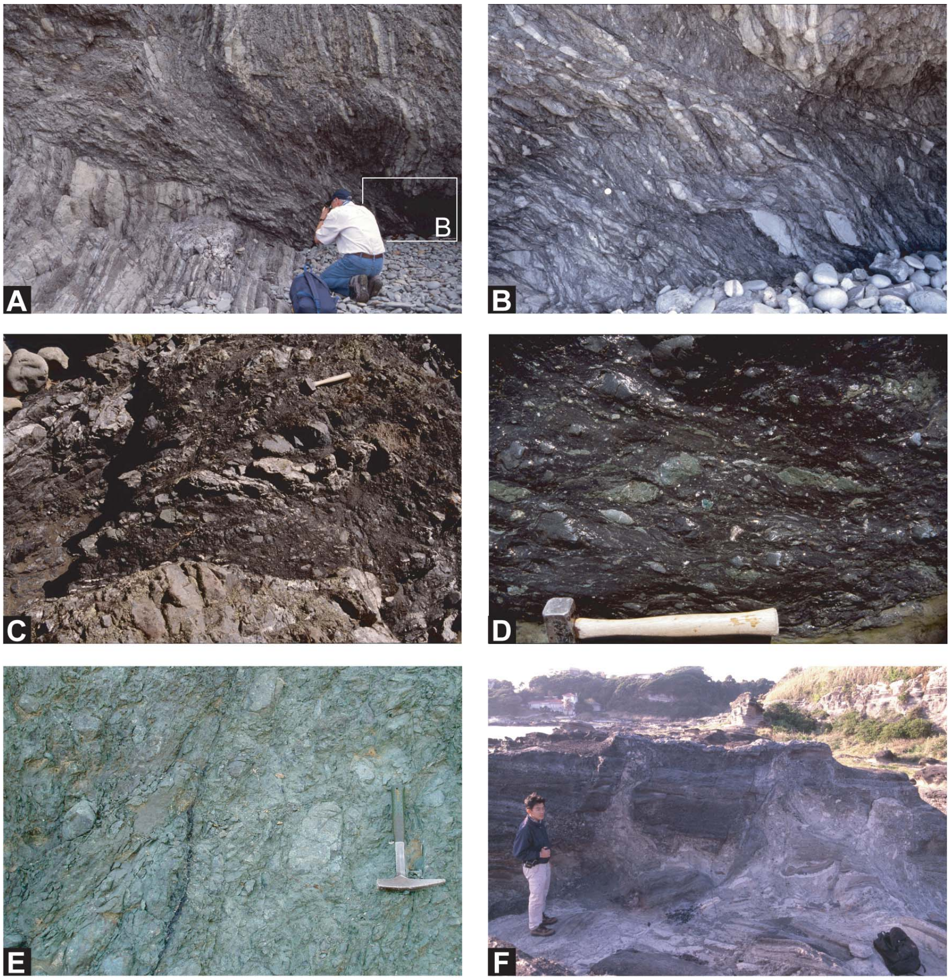

Figure 10. Examples of broken formations and tectonic mélanges in exhumed accretionary wedges (Type 4b mélanges): (A) Coastal exposure of a fault zone related to an out-of-sequence thrust, dissecting already steeply dipping beds, Waimarama Beach, South Hawke's Bay, East Coast of North Island, New Zealand. (B) Close-up of (A) showing the lozenge-shaped 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, W of Christchurch, South Island, New Zealand). Notice the imbrication of sandstone blocks in the centre of the image. (D) Close-up of 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 coloured clayey and shaly beds (Bologna area Northern Apennines; Pini 1999).

Underthrusting of an originally coherent succession (Figure 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 as metres- to tens of metres-thick packages of a coherent stratigraphic succession (Figure 9A–E). Generally, there is a lateral gradation from coherent bed units to broken formations (Figures 9D–E and 10A, 10B) and, in the latter, from packages of coherent beds to packages of disrupted portions (Barnes and Korsch 1991; Sunesson 1993). At a regional scale, in the Northern Apennines of Italy, the same stratigraphic intervals, the so-called basal complexes of the Ligurian flysch units, are cropping out as coherent to episodically and slightly disrupted units in the W ('internal Ligurian units'; Marroni 1994; Meneghini *et al.* 2009) and moderately to highly disrupted units (broken formations, Vannucchi and Bettelli 2002; Bettelli and Vannucchi 2003) and

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.

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 depending on the rheological characters of different stratigraphic intervals (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 because of 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 (Underwood 1984). One of these suggests that mélange formation is inherited from gravitational failure of sediments occurring in trench-slope settings (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; Figure 7B, 7C). 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 (Moore and Byrne 1987; Doubleday and Trenter 1992; Festa *et al.* 2010 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 the so-called basal complexes, chaotic complexes, and *argille scagliose* in the Western Alps and Apennines that represent the remnants of Upper Cretaceous Neo-Alpine and Eocene meso-Alpine episodes (see Camerlenghi and Pini 2009 and Festa *et al.* 2010 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 (Pini 1999; Bettelli and Vannucchi 2003; Camerlenghi and Pini 2009; Festa *et al.* 2010).

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 (Figure 9E) in the Middle Ordovician trench-fill deposits of the Hamburg sequence (Lash 1987). The occurrence of a complete and progressive deformation displaying well-bedded greywacke horizons grading into boudins and pinch-and-swell structures (Figure 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 a 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 McHugh 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, whereas 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-calcarous 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 (Trommsdorff 1990).

In the Western Alps (Voltri Massif), an ophiolitic mélange zone, which consists of metre-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 mélange (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 (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 (Figures 13 and 14) may be triggered by subduction processes producing mélanges (Ernst 1965; Westbrook and Smith 1983; Cowan 1985; Cloos and Shreve 1988; Orange *et al.* 1993).

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.

Type 5 – mélanges related to collision

These groups of mélanges (Figure 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 the beginning of intercontinental deformation varies greatly throughout palaeogeographic 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 (Chang *et al.* 2001; Huang *et al.* 2008; Figure 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 mélange (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

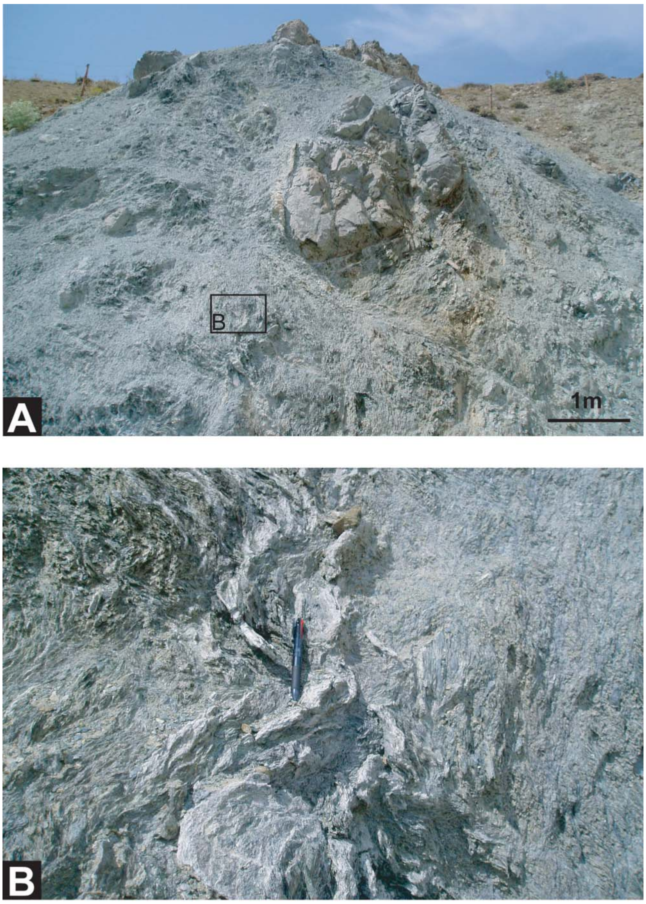

Figure 14. (A) Reworked serpentinite mud enveloping lozenge-shaped peridotite blocks in the Ankara Mélange near the city of Çankiri, Turkey. (B) Close-up of the serpentinite matrix with various clast sizes and showing a pervasive vertical fabric, reminiscent of the internal structure of serpentine diapirs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

igneous rocks of Permian to Palaeogene 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 mélange (Figure 16) in the southwestern part of the Island of Cyprus (Krylov *et al.* 2005), which represents the product of the Palaeocene–early Eocene collision of the African plate with the Cyprus microplate. It consists of red and brown clays and mudstone matrix including blocks (up to tens of metres) of alkaline basalts and Triassic limestones (Figure 16).

The collision between the pre-Apulia foreland and the Pindos Vourinos intraoceanic arc, during the Eocene, produced the Advella mélange in the Western Hellenides (Ghikas *et al.* 2010), which represents a polygenetic mélange originally emplaced by sedimentary processes (rift-related mélange). It consists of Middle Triassic to Cretaceous blocks enveloped in a Cretaceous–Eocene matrix composed of shelf and turbidite deposits. In the

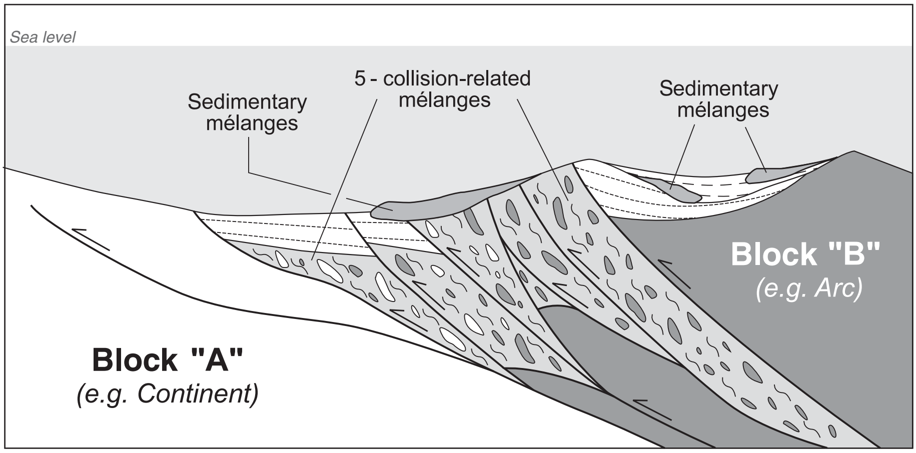

Figure 15. Conceptual model for the formation and emplacement of Type 5 mélanges associated with collision tectonics. Data are from Chang *et al.* (2001), Huang *et al.* (2008), and Ghikas *et al.* (2010).

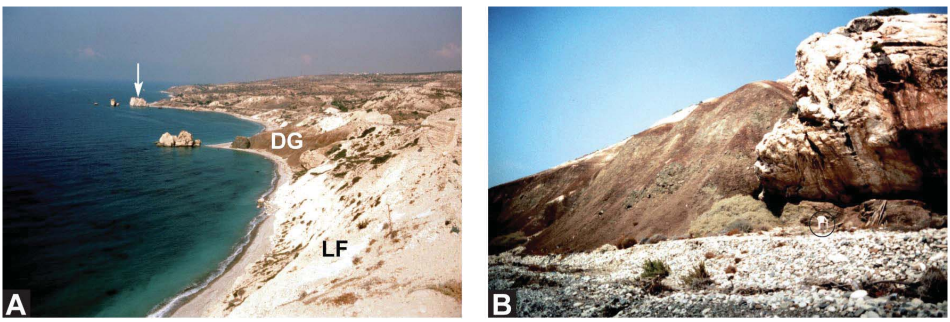

Figure 16. Collision-induced mélange examples of Type 5 from the Petra tou Romiou complex in Cyprus (Krylov *et al.* 1993, 2005). (A) Panoramic view: the white bluff (arrow) by the sea is composed of a Triassic limestone in the Mamonia complex; dark coloured rocks represent pillow basalts and hyaloclastites of the Dhiarizos Group (Mamonia, DG). White rocks in the foreground are the Palaeocene–Eocene Middle Lefkara Formation (LF) deposits 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 black circle).

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 mélange.

In the Southern Appalachians, the mélange 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.

Type 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 mélange

6a1 – precursory olistostromes. These 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 (Figures 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 throughout the world (see also Camerlenghi and Pini 2009; Festa *et al.* 2010 and references therein).

The main features of this subgroup are the same as those of the classical olistostromes (Figure 18). The blocks/matrix ratio may be variable, passing from matrix-supported blocks of centimetre-to-metre size (Type A, Pini 1999), to blocks of metres to tens of metres 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-millimetre- to centimetre-sized clasts that are randomly distributed. Blocks of Types 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 materials 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 increases stresses in weakened sediments, triggering failures.

In the Apennines and the 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 (Figure 18). In the Northern Apennines, the emplacement of precursory olistostromes of Ligurian, Subligurian, and Epiligurian 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 (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 (Sgrosso 1988; Cosentino *et al.* 2002; Vezzani *et al.* 2004; Festa *et al.* 2006, 2010; Patacca and Scandone 2007).

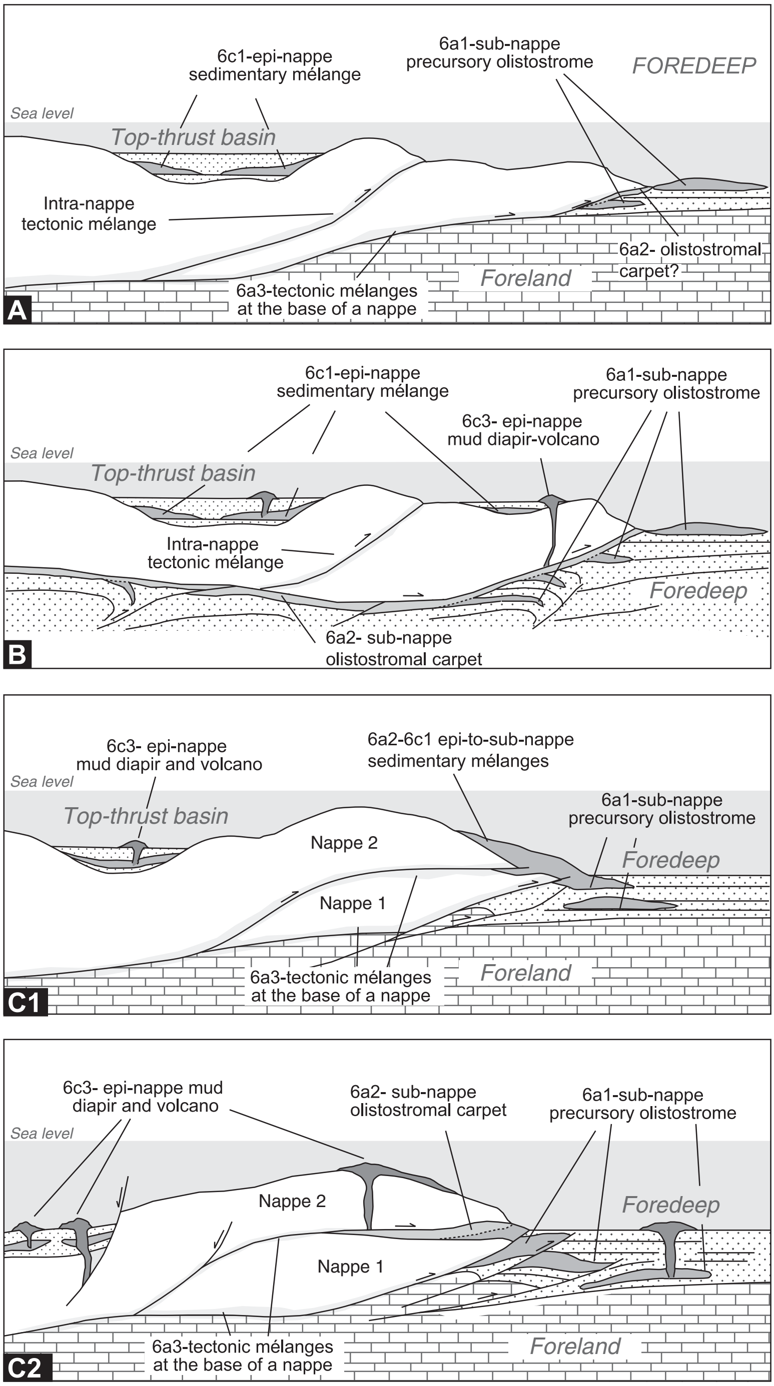

Figure 17. Conceptual models for the formation and emplacement of Type 6 mélanges associated with intracontinental deformation. (A) Obduction of a nappe over the continental margin and an early stage of foredeep development. (B) Single large nappe moving (by gravity and thrusting at the rear) over the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

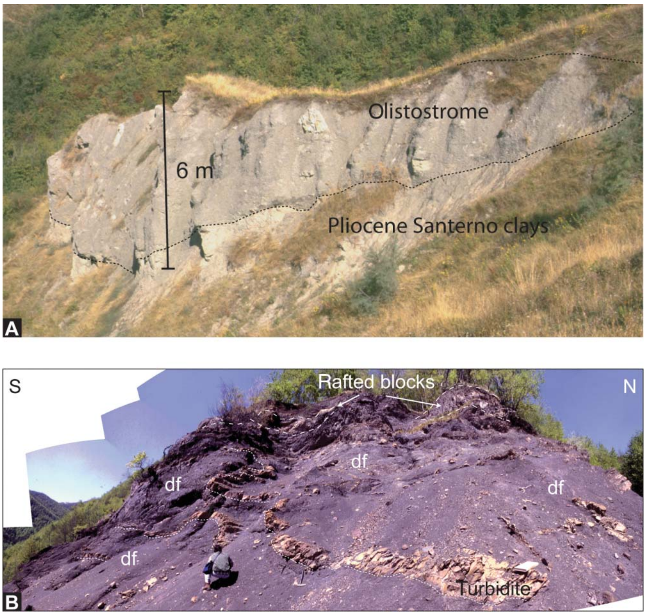

Figure 18. Field examples of precursory olistostromes as Type 6a1 mélanges. (A) A small-scale olistostrome in the Pliocene deposits at the frontal edge of the Ligurian nappe (Bologna area, Northern Apennines); (B) Panoramic view of a large outcrop of the Segavecchia olistostrome, Pianaccio area, Bologna 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). Debris flow bodies and co-genetic turbidites have been remobilized when still completely wet (from Pini *et al.* 2004, modified).

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 ultra-Helvetic domains. *Wildflysches* 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 ultra-Helvetic foredeep (late Eocene *wildflysch*). In the Versoyen Valley (Western Alps), a *wildflysch* (Méchandeur Formation) is exposed along a tectonic contact between the Palaeozoic 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 was probably triggered by a seismic activity (Frisch 1984). Moreover, other examples have been described from the Varisican *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 USA. This orogenic belt preserves several notable examples of *wildflysch*-precursory olistrostromes, most of which were formed and emplaced during the Taconic orogeny (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 piece of 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; Figure 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 (Figure 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 and slump). Also typical is the superposition of tectonic deformation and shearing because of 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 (Figure 19A, 19C). 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 (Figure 17) are, therefore, considered mixed tectono-sedimentary chaotic bodies in which exotic blocks may be supplied both by mass-transport episodes at the front of the nappe and by erosion at the base of the nappe (Festa *et al.* 2010 and refrences therein). This is, for example, the case for different interpretations of the Sestola-Vidiciatico unit in the Northern Apennines (Figure 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, highly deformed muddy debris flows (olistostromes), are considered as one of the prevailing components of the unit, even in the model of Remitti *et al.* (2007) and Vannucchi *et al.* (2008).

A significant question (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 (Bailey and McCallien 1950; Vollmer and Bosworth 1984; Jeanbourquin *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 Varisican foreland of the Iberian Peninsula (Figure 19B). It consists of a highly deformed block-in-matrix succession with boudinated large exotic blocks (Figure 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 phase spreading and then sliding at the top-front of the nappe but not to the subsequent nappe emplacement.

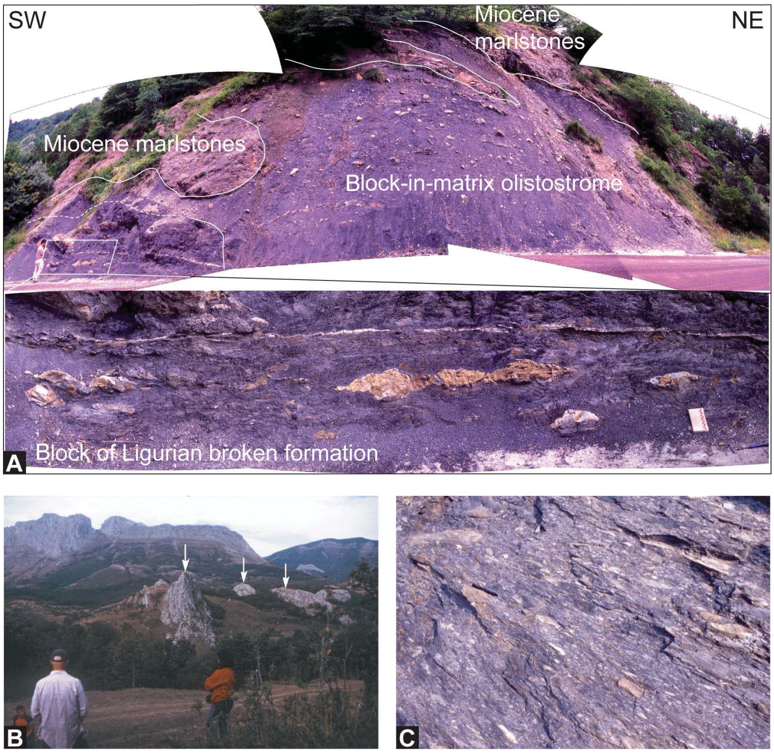

Figure 19. Field examples of an olistostromal carpet as in Type 6a2 mélanges. (A) Roadcut showing some of the key features of the Sestola Vidiciatico unit (complex), with debris bodies (block-in-matrix) in an originally sedimentary contact with the Miocene Civago Marls and the Lower Cretaceous Palombini 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 (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 (courtesy of Kei Ogata).

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 formation of 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), the Othris Mountains in Greece (Smith *et al.* 1975), the Western Alps (Jeanbourquin *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 (Figure 17C1, 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 (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 is the block-in-matrix fabric structurally ordered with respect to the regional stress field, whereas the lack of a preferential regional orientation accords with a 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 (Figure 20). Between the allochthonous thrust sheet and the buried Apulian units, a several hundreds of metres-thick mélange zone is present (Roure *et al.* 1991; Butler *et al.* 2004; Vezzani *et al.* 2004; Festa *et al.* 2006, 2010; Patacca and Scandone 2007) that consists of strongly deformed and overpressured deep-water successions. In Molise, for example, this mélange zone is represented by the Flysch Rosso at the base of the Molise unit (Vezzani *et al.* 2004; Festa *et al.* 2006, 2010). It consists of a fault-bounded tectono-sedimentary body with a mainly structurally ordered block-in-matrix fabric (Figure 20) that, far from the main shear zone, gradually passes into thick (several hundreds of metres) horizons lacking 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 (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 (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

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.

thrust sheet, but they also pointed out the role of the advancing nappe in supplying olistostromes and olistoliths in a foreland basin (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 (Figure 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 (Figure 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; Cipollari and Cosentino 1995; Critelli *et al.* 2007) in Mt Simbruini. This chaotic rock body, which is composed of carbonate breccias, intrabasinal arenites, and calclithites, was deposited by gravity-driven processes. It currently rests unconformably on the carbonate bedrock.

As for other examples in the Central-Southern Apennines (Brecce di San Massimo – Figure 21B, Castevetere Formation, and Gorgoglione Flysch, see, e.g., Patacca and Scandone 2007 and references therein), this type of intra-nappe sedimentary mélange was mainly produced during the change-over of tectonic conditions from thrusting to normal faulting or out-of-sequence thrust propagation (Sgrosso 1978; Ghisetti and Vezzani 1998; Festa *et al.* 2006; Patacca and Scandone 2007). In the Southern Apennines, for example, large blocks (Figure 21A) and olistoliths (up to hundreds of metres in size) of Cretaceous platform-derived carbonates embedded in the lower part of the Castelvetere 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 (Pescatore 1988) or as a product of deposition on thrust-sheet-top related to out-of-sequence thrust propagation (Patacca and Scandone 2007).

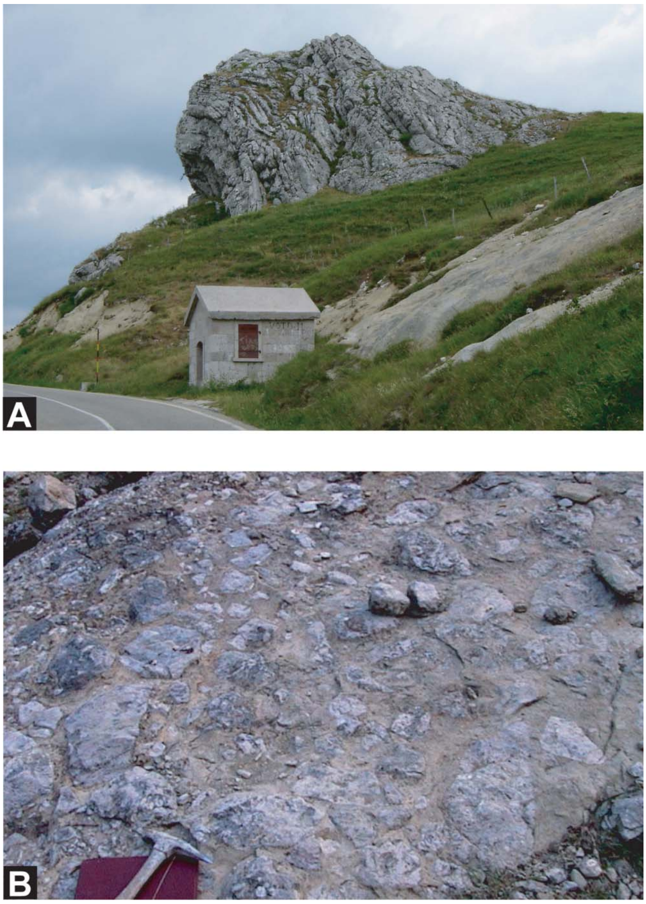

Figure 21. Field example of Type 6b1 intra-nappe sedimentary mélanges. (A) Large block (tens of metres in size) of platform-derived carbonate embedded in the Castelvetere Formation (Critelli and Le Pera 1995) at Mt Caruozzo, Southern Apennines, Italy. (B) Breccias and megabreccias (Brecce di San Massimo)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

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 the 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 normally the 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, in 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 (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, Mts Frentani mélange (Vezzani *et al.* 2004, 2009, 2010; Festa *et al.* 2006, 2010; 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 hundreds of metres) 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 (Figure 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 N of Atessa (Vezzani *et al.* 2004; Festa *et al.* 2006, 2010).

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 either to out-of-sequence thrust dissecting the entire Ligurian nappe, transpressional activity along strike-slip faults, or to normal faulting (Capitani 1993).

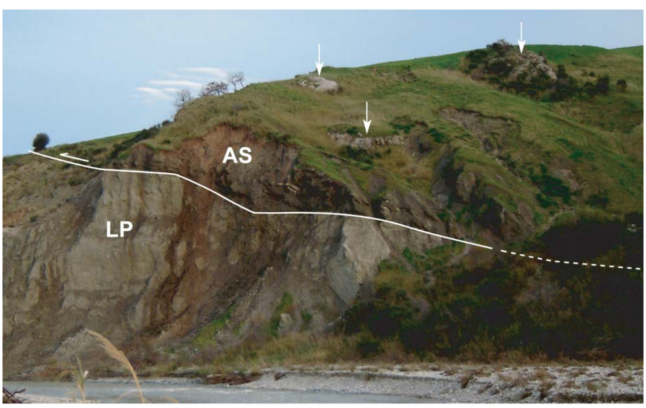

Figure 22. Type 6b2 tectono-sedimentary intra-nappe mélange (Mts Frentani mélange *sensu* Vezzani *et al.* 2004; Festa *et al.* 2006) consisting of *argille scagliose* (AS) with a block-in-matrix fabric (arrows indicate the main blocks). This mélange represents the product of both tectonic and sedimentary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

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 (Figure 23), which is the product of gravitational instability along the margin of a piggy-back basin (Figures 17A–C, 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 related to tectonic processes that controlled the uplift rate of the bounding thrust surfaces (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 Pre-Cordillera (Beer *et al.* 1990).

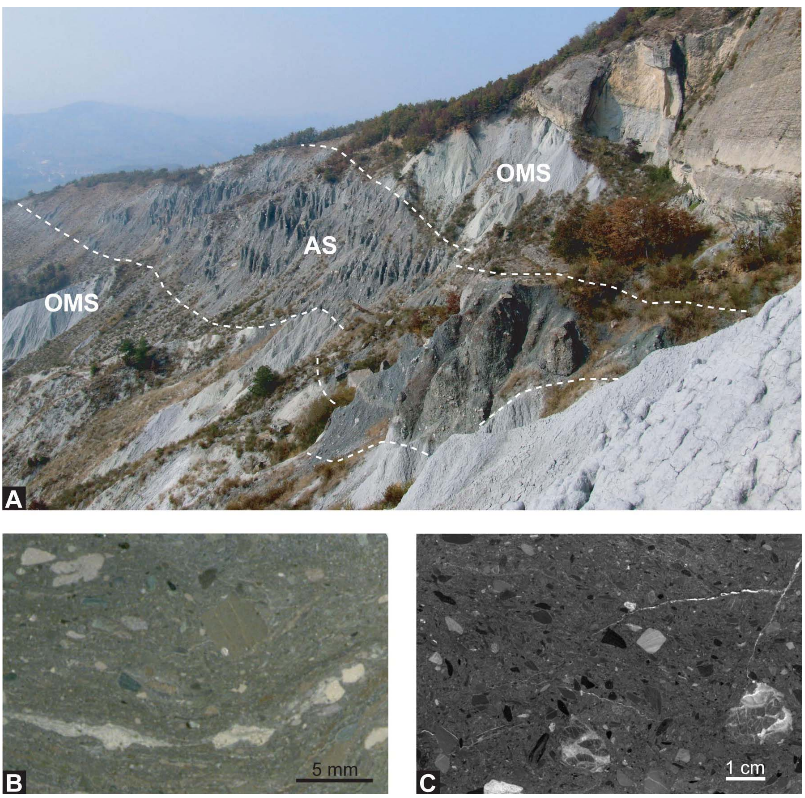

Figure 23. Field examples of Type 6c epi-nappe mélanges. (A) Panoramic view of a precursory olistostrome of *argille scagliose* (AS) including the late Oligocene–early Miocene sedimentary rocks (OMS) of the Northern Apennines, W of Varzi, Italy. (B) and (C) Typical brecciated matrix, as shown by the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

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 the 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 (Barrême basin in Western Alps; Artoni and Meckel 1998) in addition to the 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 (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 metres 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 metres, 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 2010) 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 (Dela Pierre *et al.* 2002, 2007; Clari *et al.* 2004; Irace 2004; Festa 2010). The upward rise of fluids triggered by gas hydrate 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 diapiric processes (Dela Pierre *et al.* 2007; Festa 2010).

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 the formation of chaotic rock bodies in a piggy-back setting and temporal changes in the rate of thrusting. Reactivation during and/or after the 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 the 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 soutwestern 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.* (2010) from the Campobasso area (Figure 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 a thousand kilometres 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 kilometres square in size; Figure 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 (Figure 25), block-in-matrix arrangement, and occurrence of hydraulic fractures within the hard blocks are diagnostic features of the

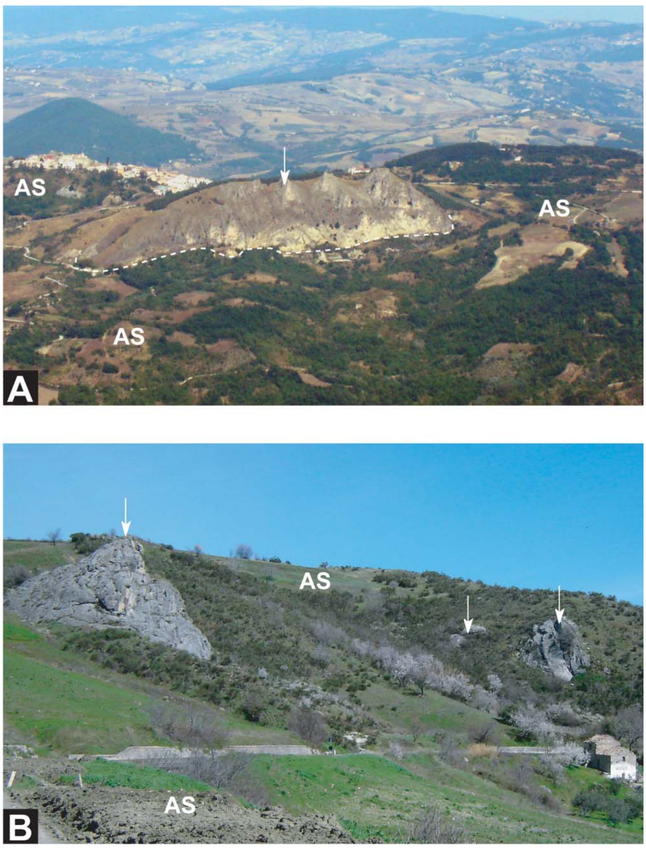

Figure 24. Examples of Type 6c2, epi-nappe tectono-sedimentary mélanges in the Molise region, Central Apennines, Italy. (A) Large block (arrow) of calcareous mudstone and calcarenite with interbeds of black chert levels embedded in the *argille scagliose* (AS) of the Sannio Nappe at La Civita di Duronia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

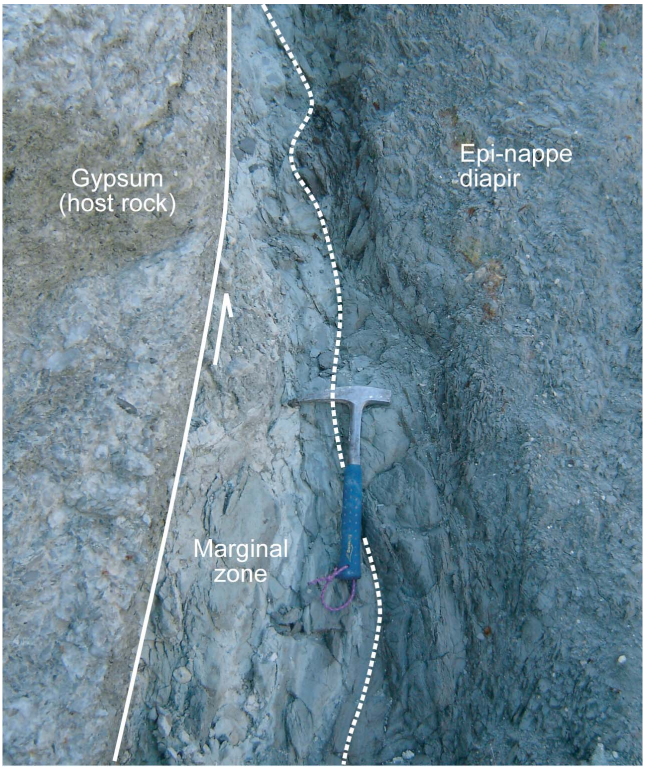

Figure 25. Epi-nappe diapiric mélange (Type 6c3 mélange) from the Moncucco Torinese quarry, Tertiary Piedmont Basin, NW Italy (late Tortonian–early Messinian marls). A few decimetres-wide marginal zone of anisotropic mud breccias, characterized by fluidal features, separates the shear zone from the host rocks (Messinian primary evaporites) (Dela Pierre *et al.* 2007; Festa 2010 for major details).

diapiric type of mélange (Orange 1990; Dela Pierre *et al.* 2007; Festa 2010). This is the case, for example, in the diapiric disrupted unit (Dela Pierre *et al.* 2007; Festa 2010) 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 2010) 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 hydrate 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 (Figure 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 yet been documented, but in some cases, the diapiric rise seems to have pierced also the tectonosome units considered to be 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 off shore of eastern Iran (Grando and Mc Clay 2007) and their onset related to changes in the wedge dynamics. These bodies occasionally reach the seafloor, originating mud volcano conical apparatus and, thus, supplying sediments (mud breccias) to the thrust-top basins. At a larger scale, mud volcanoes appear as seamounts at the edge and inside piggy-back basins in the seismic profiles transversal to the southern part of the Barbados accretionary complex (Moore *et al.* 1990). The high elevation of these bodies above the seafloor controls the sediment dispersal inside the satellite basins, and 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 the modern seafloor 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 is supplied by extensive horizons of olistostromes (Somoza *et al.* 2002; Diaz del Rio *et al.* 2003; Krastel *et al.* 2003; Pinheiro *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) a 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 analysed 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.* (2010) 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, and 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,

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

Types of mélange	Geodynamic environments	Processes	Products	Minor related products
Related to:				
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 disruprion); slides	In-situ fluidification; mud diapirs?
3. Strike-slip tectonics	Different types of collision	Tectonic	Broken formations; mélanges (exotic blocks were commonly recicled 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 avalances, 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. Intracontinental deformation				
a. Sub-nappe				Mud diapirs and mud volcanoes
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.)	
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	Gravitational	Olistostromes and mass-transport deposits (debris flows and avalanches, slumps, slides, etc.)	
c2. Tectono-sedim.	(e.g. piggy back, top thrust basins)	Tectonic, gravitational	Broken formation, mélanges	
c3. Diapiric		Diapiric	Mud diapirs and mud volcanoes	

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-evaluated in some on-land examples of ancient accretionary complexes in Japan, New Zealand, and the West Coast of the USA, as well as in some remnants of the collisional accretionary wedges ('suture zones' in the 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) are seen on the seafloor (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 (Figure 7A) or some asperity in the downgoing plate (reactivated normal faults? see e.g. von Huene 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 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 (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 (Pini 1999; Bettelli and Vannucchi 2003), part of the *argille scagliose* (Sicilide units) and 'Red Beds' (Molise units) in the Central-Southern Apennines (Vezzani *et al.* 2004, 2009, 2010; Festa *et al.* 2006), the Franciscan Complex (Wakabayashi 1992; Meneghini *et al.* 2009), Antarctica (Doubleday and Trenter 1992), and Taconic mélanges in the central and northern Appalachians (i.e. Vollmer and Bosworh 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 metres) mélange zones (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 of a Cretaceous–Miocene succession (Table 3; Figure 10A and 10B).

In the 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 with 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.* 2010), 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 kilometres) 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 artefacts 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, and 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 a 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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