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Mechanisms and processes of stratal disruption and mixing in the development of mélanges and broken formations: redefining and classifying mélanges

[1] Festa A., [2-1] Dilek Y., [3] Pini G.A., [1-4] Codegone G., and [5] Ogata K.

[1] Dipartimento di Scienze della Terra, Università di Torino, 10125 Torino, Italy

[2] Department of Geology and Environmental Earth Science, Miami University, Oxford, OH 45056, USA

[3] Dipartimento di Scienze della Terra e Geologico-Ambientali, Università di Bologna, 40127 Bologna, Italy

[4] Dipartimento di Ingegneria dell‟Ambiente del Territorio e delle Infrastrutture, Politecnico di Torino, 10129 Torino, Italy.

[5] Department of Arctic Geology, UNIS (University Centre in Svalbard) – Pb. 156, 9171, Longyearbyen, Norway

*Corresponding author:

Andrea Festa

E-mail: andrea.festa@unito.it

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ABSTRACT

The terms *mélange* and *broken formation* have been used in different ways in the literature. The lack of agreement on their definition often leads to confusion and misinterpretations. An evaluation of the various uses of these terms allows us to consider several types of chaotic rock bodies originated by tectonic, sedimentary and diapiric processes in different tectonic settings. Our review of stratal disruption and mixing processes shows that there exists a continuum of deformation structures and processes in the generation of mélanges and broken formations. This continuum is directly controlled by the increase of the degree of consolidation with burial. In tectonically active environments, at the shallow structural levels, the occurrence of poorly consolidated sediments favors gravitational deformation. At deeper structural levels, the deformation related to tectonic forces becomes gradually more significant with depth. Sedimentary (and diapiric) mélanges and broken formations represent the products of punctuated stratal disruption mechanisms recording the instantaneous physical conditions in the geological environment at the time of their formation. The different kinematics, the composition and lithification degree of sediments, the geometry and morphology of the basins, and the mode of failure propagation control the transition between different types of mass-transported chaotic bodies, the style of stratal disruption, and the amount of rock mixing. Tectonically broken formations and mélanges record a continuum of deformation that occurs through time and different degrees of lithification during a progressive increase of the degree of consolidation and of the diagenetic and metamorphic mineral transformation. Systematic documentation of the mechanisms and processes of the formation of different broken formations and mélanges and their interplay in time and space are highly important to increase the understanding of the evolutionary history of accretionary wedges and orogenic belts.

Key words: Tectonic and sedimentary mélanges, diapiric mélanges, broken formations, mélange forming processes, stratal disruption and mixing of rocks, mass-transport deposits and processes.

1. Introduction

Mélanges and broken formations represent a significant component of most convergent margins and orogens around the world (Fig. 1), and the details of their block-in-matrix character reflect a close relationship between the processes and the tectonic setting of their formation (Suzuki, 1986; Festa et al., 2010a). However, the lack of agreement on the definition of mélange (e.g., Silver and Beutner, 1980; Rast and Horton, 1989; also compare Şengör, 2003 with Pini, 1999; Cowan and Pini, 2001; Festa et al., 2010a; Wakabayashi, 2011) has lead to some confusion and misinterpretations in the literature. At shallow structural levels in tectonically active environments, sediments are subject to small-scale deformation immediately after deposition at rates and in ways dependent on the interplay between gravitational deformation and tectonic burial (e.g., Byrne, 1994; Maltman, 1994). The downward increase in both the consolidation and lithification of buried sediments and tectonic forces controls the progressive increase in deformation and, in cases, stratal disruption (Maltman, 1994 and references therein; Onishi and Kimura, 1995; Yamamoto et al., 2012a). The result of these conditions is a continuum of development of structures in the originally coherent stratigraphic successions via stratal disruption and mixing processes, which play a major role in the genesis of broken formations and mélanges (e.g., Hsü, 1968; Raymond, 1984; Cowan, 1985).

Time-progressive evolution of deformation structures in chaotic rock units, such as broken formations to mélanges, has been rarely described in the literature (see e.g., Smith et al., 1979; Raymond, 1984; Cowan, 1985; Needham, 1995; Harris et al., 1998; Lucente and Pini, 2003; Ogata et al., 2012a; Pini et al., 2012; Yamamoto et al., 2012a). This is in part due to the fact that collisional and post-collisional shortening, magmatism, extensional deformation, and strike-slip tectonics may have obscured or strongly remodified the structural evidence for the pre-existing continuum. Nevertheless, a careful examination of the rock record and the internal fabric of the chaotic rock bodies, together with their contact relationships with the country rocks, reveals important clues about the larger-scale processes that occurred in different tectonic settings and at shallow structural levels during mélange formation. The most important mélange forming process at deeper structural levels is thought to have taken place in subduction channels (e.g., Cloos, 1982; Cloos and Shreve, 1988a, 1988b; Ogawa, 1998; Gerya et al., 2002; Guillot et al., 2004; Ernst, 2006; Federico et al., 2007; Blanco-Quitero et al., 2010; Malatesta et al., 2012), in which high degrees of mixing of rocks (including ultramafic rocks) with differing P-T-t histories and metamorphic grades may occur.

This paper is aimed at streamlining the existing discussions on the mechanisms and processes of stratal disruption and mixing in the development of mélanges and broken formations, and at redefining and reclassifying the mélanges and related rock units. In the first part of the paper, we briefly review and discuss the definitions of the terms mélange and broken formation and re-define these terms in light of recent observations and interpretations made by the international scientific community. We also discuss the origins of the chaotic rock masses (tectonic, sedimentary, diapiric, and polygenetic) and review their global occurrences, expanding on the tectonic-genetic classification of chaotic rocks we proposed earlier (see Festa et al., 2010a). In the second part of the paper, we present several models for the formation of various types of mélanges and broken formations at shallow structural levels in accretionary wedges and orogenic belts (where metamorphism is very low grade or absent). Here, we review and synthesize the existing data, and demonstrate that a continuum of stratal disruption and mixing processes operates across different structural levels or depths of burial in various tectonic settings. The nomenclature we propose here and the continuum of stratal disruption and mixing described at shallow structural levels provide a useful and coherent framework for future studies in mélange terrains. Redefinition and more systematic, process-oriented classification of mélanges should also be highly insightful for the recognition of these chaotic rock bodies in the Precambrian greenstone belts (Dilek and Ahmed, 2003; Dilek and Polat, 2008).

2. Mélange and broken formation terminology

“Mélange” is a descriptive, non-genetic term that must be used only in describing a mappable (at 1:25,000 or smaller scale) body of internally disrupted and mixed rocks in (or rarely without) a

pervasively deformed matrix (Berkland et al., 1972; Wood, 1974; Silver and Beutner, 1980; Raymond, 1984; Cowan, 1985). Yet, the debate and discussions on the mélange concept continue after nearly four decades of extensive studies of mélanges and related rock units around the world (Silver and Beutner, 1980; Raymond, 1984; Rast and Horton, 1989; Wakabayashi and Dilek, 2011). We refer the reader to Hsü (1968), Cowan (1974, 1985), Raymond (1984), Suzuki (1986), Rast and Horton (1989), Pini (1999), Şengör (2003), Camerlenghi and Pini (2009), Festa et al. (2010a), Vannucchi and Bettelli (2010), Wakabayashi and Dilek (2011) and Ogata et al. (2012b) for various discussions on the conflicting uses of the term *mélange*.

The term *mélange*, in its classical descriptive and non-genetic definition (Berkland et al., 1972; Wood, 1974; Silver and Beutner, 1980; Raymond, 1984; Cowan, 1985), does not restrict the nature of lithological units involved (sedimentary, metamorphic or igneous); contact relationships between these diverse lithological units can be tectonic, stratigraphic or intrusive, depending on the process of mélange formation (Tab. 1). This definition implies that the term *mélange* can be used only, at least in part, as a synonym of *complex* (see Salvador, 1994). However, the controversial definition of *complex*, as a lithodeme subunit (NACSN, 2005), suggests a mostly tectonic origin of its contacts (see also Pasquarè et al., 1992; Vannucchi and Bettelli, 2010). The term *complex,* however, is also a formal lithostratigraphic term, defining any rock body that is characterized by complicated deformation patterns and bounded by primary (i.e. stratigraphic) contacts (see also Pasquarè et al., 1992; Salvador, 1994). Only "sedimentary mélanges" (olistostromes) are compatible with the classic principles of stratigraphic superposition, whereas many mélange occurrences in nature do not follow these principles (Silver and Beutner, 1980) because they are not bounded by stratigraphic contacts.

The classical definitions of the term *mélange* (Hsü, 1968; Berkland et al., 1972; Silver and Beutner, 1980) portray these rock bodies to "*commonly*" include a *"pervasively deformed matrix"* (Silver and Beutner, 1980) or a *"fragmented matrix of finer-grained material"* (Raymond, 1984). Some researchers have avoided in these definitions any specification of the origin of that matrix as tectonic, sedimentary or diapiric. This matter is particularly important in studying sedimentary mélanges in that it restricts the use of the term *mélange* only to mass-transport deposits, which display a chaotic internal arrangement and mixing of exotic and native blocks in a deformed matrix (debris flows, hyperconcentrated flows, blocky flows, Mutti et al., 2006; Ogata et al., 2012a). It also excludes other sedimentary deposits such as turbidites. Glacial till (e.g., Hoffmann and Piotrowski, 2001) or the Martian chaos (e.g., Kargel et al., 2007) may be included in this definition of mélange because they, too, form by mixing of different blocks as a result of slope failure, mass-transport processes, gas outburst by clathrate dissociation, mud volcanism, and bolide impacts on the surface of the planet.

Mixing of rocks is clearly stated in the definition of mélange by Hsü (1968) and Silver and Beutner (1980), and is addressed by Raymond (1984) as one of the two fundamental mélange-forming processes, but the significance and the amount of mixing is not clearly defined in these earlier definitions. In addition, the meaning of the terms “exotic” and “native” blocks is ambiguous (Tab. 1) mainly because the concept of an exotic origin changes dramatically in different tectonic settings, structural levels, and according to the different origins of mélanges. A restrictive usage of the term “exotic” is consistent with a virtual association between exotic blocks and subduction settings (e.g., Bailey et al., 1964; Cowan, 1978, 1985; Aalto, 1981; Cloos, 1982; Barber et al., 1986; Brown and Westbrook, 1988; Cloos and Shreve, 1988a, 1988b; Onishi and Kimura, 1995; Meschede et al., 1999; Wakabayashi, 2004, 2011; Ikesawa et al., 2005; Federico et al., 2007; Malatesta et al., 2012) partly because subduction channel is the most popular setting where exotic blocks (e.g., HP eclogite blocks) become encased in epizonal metamorphic binders (e.g., Cloos, 1982; Cloos and Shreve, 1988a, 1988b; Ogawa, 1998).

At shallow structural levels in subduction zone settings or in other tectonic environments (e.g., continental to intracontinental deformation settings, strike-slip tectonics, extensional settings), the meaning of “exotic” must be extended to a wider range of blocks. Hsü (1968) defined an exotic component as a *“tectonic inclusion detached from some stratigraphic rock units foreign to the main body of mélange”*, whereas native components are *„disrupted brittle layers interbedded with the ductiley deformed matrix“*. Berkland et al. (1972) clearly distinguished between “exotic” and “tectonic” blocks, whereas Hsü (1968) considered them to be synonymous. These authors defined “exotic” blocks as *“variably sized masses of rock occurring in a lithological association foreign to that in which* *the mass formed".* "Tectonic" blocks are then considered more restricted in origin because they consist of only blocks "*transported through the operation of tectonic processes".* Then, not all "tectonic" blocks are "exotic", and not all "exotic" blocks are tectonic in origin (Berkland et al., 1972).

Sedimentary and diapiric processes and a combination and superposition of them with tectonic processes have been widely accepted to play a major role in the incorporation and mixing of exotic blocks during mélange formation (e.g., Hsü, 1968; Berkland et al., 1972; Cowan, 1974; Cowan and Page, 1975; Aalto, 1989; Harris et al., 1998; Erickson, 2011; Osozawa et al., 2011; Wakabayashi, 2011). Exotic component can be "foreign" at different levels with respect to the "native" component of the main body, varying from simply extra-formational (e.g., Panini et al., 2002; Codegone et al., 2012b), to an extra-basinal origin (e.g., Lash, 1987; Ogata et al., 2012c), up to having been derived from different structural units (Abbate et al., 1970, 1981; Alonso et al., 2006; Lucente and Pini, 2008 and references therein), paleogeographic domains, tectonic settings or structural levels (P-T conditions, diagenetic/metamorphic degree) (e.g., Cloos, 1982; Cowan, 1985; Cloos and Shreve, 1988a, 1988b; Dilek, 1989; Harris et al., 1998; Ogawa, 1998; Dilek et al., 1990, 1999, 2007; Wakabayashi, 2011, 2012; Ukar, 2012).

The processes of exotic block incorporation into mélanges can provide important information about the mélange genesis. Fragmentation and dismemberment may exceed the strength of a given lithostratigraphic unit (or formation), and the rocks that formed in different geological environments at different times may involve extensive mixing (Raymond, 1984). If fragmentation and dismemberment does not exceed the strength of a given lithostratigraphic unit (or formation), we must then use the term "broken formation" (*sensu* Hsü, 1968) to describe a stratally disrupted unit, which contains no exotic blocks but only "native" components. These broken formations preserve their lithological and chronological identity ("„tectonosomes" *sensu* Pini 1999). Here, stratal disruptions and fragmentation occur without mixing (Hsü, 1968; Cowan, 1985) (Tab. 1) and broken formations show a gradual transition from a bedded succession to a strongly disrupted block-in-matrix fabric (Lash, 1987; Barnes and Korsch, 1991; Sunesson, 1993; De Libero, 1998; Festa et al., 2010a; Codegone et al., 2012a) representing the intraformational equivalent of mélanges (Tab. 1). In this sense, this definition embodies "broken" and "dismembered units", as described by Raymond (1984), that were proposed to differ from each other in the degree of stratal disruption and from mélanges by the lack of exotic blocks. Cowan (1985) chose not to use the exotic block requirement in defining mélanges, and the four types of mélanges he defined included the broken formation of Hsü (1968) and Raymond (1984).

In this article and as our common practice (Codegone et al., 2012a, 2012b; Ogata et al., 2012b), we consider mélanges and broken formations two end-members, which differ from each other in terms of the nature of blocks (exotic vs. native) and the mechanisms of their formation (mixing plus stratal disruption vs. only stratal disruption; e.g., Hsü, 1968; Harris et al., 1998). They can also both form by tectonic, sedimentary or diapiric processes (Tab. 1) or through a combination and superposition of these processes. Independently from their deformational path and origin, we favor to define a disrupted rock body without exotic blocks and rock mixing as a broken formation and not as mélange. Following Raymond (1975), we define mélange as a body of mixed rocks, containing blocks (exotics and native) that are derived from different stratigraphic units or sequences, different tectonic units, various paleogeographic domains, and/or dissimilar metamorphic zones.

The scale of observation is highly important in characterization of mélanges. Although extension of the mélange term to bodies that are non-mappable at 1:25,000 or smaller scales renders the term mélange useless (e.g., Raymond, 1984), it is not unusual to find this term used in describing small-scale or meso-scale mélanges (e.g., Bosworth and Vollmer, 1981; Bradley and Kusky, 1992; Wakita, 1988, 2000; Fukui and Kano, 2007). To avoid any confusion, we agree to the use of the terms "small-scale mélanges and broken formations" (see Codegone et al., 2012b) or "meso-scale mélange" (see Bradley and Kusky, 1992) in order to indicate not-mappable (at 1:25,000 scale) mélanges and broken formations, whereas "chaotic or disrupted (rock) units" (*sensu*, e.g., Yamamoto et al., 2009; Festa, 2011) must be considered a general term to indicate bodies apart from the nature of the embedded blocks. These terms do not define micro-scale mélanges and broken formations, nor do they suggest applying these terms to chaotic bodies at the scale of sedimentary layers/beds. However, they are to be used in describing chaotic rock units mappable on a scale that is larger than

1:25,000 according to the requirement of geological map databases and GIS technologies used for the production of modern geological maps.

3. Types, processes of formation and triggering mechanisms of broken formation and mélanges

The re-defined terms *mélange* and *broken formation* and their clarified meaning could be extended to a large number of bodies of mixed rocks formed at different structural levels and in various tectonic settings (Fig. 1, and Tab. 2; see Suzuki, 1986; Festa et al., 2010a, 2010b and reference therein). It is important, however, to distinguish mélange-forming processes from triggering mechanisms in each of these tectonic settings (Tab. 2; Moore and Wheeler, 1978; Cloos, 1982; Saleeby, 1984; Barber et al., 1986; Raymond et al., 1989; Orange, 1990; Festa et al., 2010a; Festa, 2011).

3.1. Mélanges and tectonic settings of their formation

When we compare some of the exhumed, ancient chaotic rock bodies and their modern analogues that developed as a result of different tectonic processes in different geodynamic environments, we realize that several examples of "tectonic mélanges" described in the literature do not include exotic blocks (see, e.g., Cowan, 1974, 1985; Vollmer and Bosworth, 1984; Lash, 1987; Brandon, 1989; Wakabayashi, 1992, 2011; Harris et al., 1998; Onishi et al., 2001; Vannucchi and Bettelli, 2002) (Tab. 2). Instead, they consist of variably disrupted units or well-developed block-in-matrix units corresponding to broken formations (*sensu* Hsü, 1968), which are transitional to slightly boudinaged beds and coherent layered units. These broken formations were formed by stratal disruption of the original coherent successions in various tectonic environments, mainly related to subduction zone processes (Type 4b in Tab. 2, and Figs. 2B and 2D), arc-continent and continent-continent collisions (Type 5 in Tab. 2, and Fig. 2E), intra-continental deformation (Types 6a3, 6b2 and

6c2 in Tab. 2, and Fig. 2E) and strike-slip tectonics (Type 3 in Tab. 2, and Fig. 2F). Some of the most salient examples of broken formations include the Ligurian mélanges in the eastern Northern Apennines (Fig. 3C) (e.g., Pini, 1999; Vannucchi and Bettelli, 2010), part of the *Argille Varicolori* and "Flysch Rosso" (Figs. 3A and 3B) in the Central-Southern Apennines (Festa et al., 2010a, 2010b; Vezzani et al., 2010), the Costal Range mélange of Hikurangi margin, and the Esk Head (Fig. 3D) and similar chaotic rocks in the Torlesse Complex in New Zealand (e.g., Pettinga, 1982; Barnes and Korsch, 1991), part of the Franciscan Complex in California (e.g., Wakabayashi, 1992, 2011; Meneghini et al., 2009), the youngest section of the Shimanto Belt in Japan (Ditullio and Byrne, 1990), the Bobonaro Mélange of the active Banda arc-continent collision (Harris et al., 1998; Harris, 2011), and the Taconic mélanges in the Central-Northern Appalachians (e.g., Vollmer and Bosworth, 1984; Lash, 1987; Codegone et al., 2012a).

The origin of exotic blocks and the nature of processes responsible for their emplacement and mixing within a mélange are a subject of long-lasting debate and controversy (e.g., Bailey et al., 1964; Hsü, 1968; Coleman and Lanphere, 1971; Berkland et al., 1972; Cowan, 1974, 1985; Raymond, 1984; Suzuki, 1986; Aalto, 1989; Harris et al., 1998; Şengör, 2003; Osozawa et al., 2011; Wakabayashi, 2011). Tectonic mélanges with exotic blocks mixed solely by tectonic processes (Tab. 2) are predominant in shear zones (Figs. 3E and 3F, see also Type 3 in Fig. 2F), and occur at different scales in (1) narrow, anastomosing and coalescent fault zones (Fig. 3F; see, e.g., Coleman and Lanphere, 1971; Suppe, 1972; Cowan, 1974; Pettinga, 1982; Kimura et al., 1996; Hashimoto and Kimura, 1999; Codegone et al., 2012b), (2) crustal-scale thrust fault zones (e.g., Moore and Sample, 1986; Doubleday and Trenter, 1992; Kusky et al., 1997; Meneghini et al., 2009), (3) plate boundaries (e.g., Wakabayashi, 1992; Ogawa, 1998; Onishi et al., 2001; Vannucchi et al., 2008; Meneghini et al., 2009; Kusky and Jianghai, 2010; Kimura et al., 2012) and transform fault or fracture zones (e.g., Moseley and Abbotts, 1979; Suzuki, 1986; Dilek, 1989; Saleeby, 1989; Dilek et al., 1991; Shervais et al., 2011), where they may be as thick as 1000-2000 meters, and (4) subduction channels (e.g., Cloos, 1982; Federico et al., 2007; Blanco-Quitero et al., 2010; see Type 4b in Fig. 2C) where flow-mélanges form (e.g., Cloos, 1982; Shreve and Cloos, 1986; Ukar, 2012). In these settings tectonic processes incorporate exotic blocks into the mélange matrix by offscraping, underplating, sinking of roof thrust rocks, and tectonic slicing (see Type 4b in Figs. 2B, 2C and 2D). In view of similar observations, Cowan (1974) suggested that tectonic mélanges are structurally equivalent to faults, along which the tectonic dislocation *"has expanded from a plane (i.e., fault) to a zone of several members to kilometers in width (i.e., tectonic mélange)".* However, mélanges with exotic blocks originated from sedimentary (e.g., Hsü, 1968; Cowan and Page, 1975; Abbate et al., 1981; Naylor, 1982; Cowan, 1985; Liu and Einsele, 1996; Burg et al., 2008; Erickson, 2011; Wakabayashi, 2011; Cieszkowski et al., 2012; Codegone et al., 2012b; El Bahariya, 2012; Hitz and Wakabayashi, 2012; Pini et al., 2012) and diapiric processes (e.g., Maxwell, 1974; Cloos, 1983; Becker and Cloos, 1985; Maekawa et al., 1993; Fryer et al., 1999; Camerlenghi and Pini, 2009) are common in other tectonic settings (Tab. 2). Both of these mélange types may subsequently be overprinted and structurally reworked by tectonic processes such as shearing and tectonic mixing when placed in an accretionary wedge or in a subduction channel (e.g., Cowan and Page, 1975; Cloos, 1982; Cowan, 1985; Cloos and Shreve, 1988a, 1988b; Medialtea et al., 2004; Dilek and Thy, 2006; Burg et al., 2008; Osozawa et al., 2009, 2011; Cowan and Brandon, 2011; Wakabayashi, 2011; Fig. 2C; see Type 4b in Tab. 2; see also Figs. 4A and 4B), or by thrusting and folding in a collisional belt (e.g., Brandon, 1989; Pini, 1999; Dilek, 2006; Camerlenghi and Pini, 2009; Osozawa et al., 2009, 2011; Festa et al., 2010a, 2010b; Codegone et al., 2012b; Ogata et al., 2012b; Fig. 3A; see Types 6a2 and 6b2 in Tab. 2 and Fig. 2E). These mélanges (with exotic blocks) mainly represent "polygenetic" mélanges, in which the occurrence of exotic blocks in their matrix is commonly due to different types of mass-transport (slides, debris flows and avalanches, etc.) or diapiric processes (Fig. 4C) rather than due solely to tectonic processes.

Although subsequent tectonic processes commonly affect and overprint the existing sedimentary and diapiric mélanges (Cowan and Page, 1975; Osozawa et al., 2011), it is not uncommon for tectonic mélanges and broken formations to be, in turn, reworked and overprinted by later sedimentary or diapiric processes (e.g., Aalto, 1981). The occurrences of mud diapirs, which reworked some previously formed tectonic mélanges and broken formations, have been described from the wedge-top succession of the Tertiary Piedmont Basin (Dela Pierre et al., 2007, Festa, 2011) and from the frontal part of the ancient External Ligurian accretionary complex of the Northern Apennines (Codegone et al., 2012b) in NW Italy. Large-scale sedimentary processes reworking tectonic mélanges and broken formations are responsible for the development of basin-wide olistostromes in the wedge-top and foredeep Tertiary basin of the Apennines (Abbate et al., 1970, 1981; Pini, 1999; Lucente and Pini, 2003, 2008; Cavazza and Barone, 2010; Vezzani et al., 2010; Remitti et al., 2011). Diapiric processes overprinting sedimentary mélanges occur, for example, in the Hoh accretionary complex in the Olympic Peninsula (e.g., Cowan and Brandon, 2011), in the Hamburg Klippe of central Pennsylvania (Lash, 1987; Codegone et al., 2012a), in the Timor region of the Banda arc (Harris et al., 1998) and in several offshore cases (Camerlenghi and Pini, 2009 and references therein). Diapiric mélanges and shale diapirs reworked by sedimentary processes occur in the accretionary complex of Timor in Indonesia (e.g., Barber et al., 1986; Barber and Brown, 1988; Harris et al., 1998).

Sedimentary mélanges (Figs. 4D, 4E and 4F) may occur in many tectonic environments, but they prevail particularly in extensional (Type 1 in Tab. 2, and Fig. 2A; e.g., Bernoulli, 2001; Alonso et al., 2008) and passive margin settings (Types 2a and 2b in Tab. 2, and Fig. 2A; e.g., Naylor, 1982; Liu and Einsele, 1996; Dilek at al., 2005; Camerlenghi and Pini, 2009; Robertson et al., 2009; Ghikas et al., 2010; Bonev et al., 2012), where the direct contribution of extensional tectonic processes (e.g., crushing and mixing in normal fault zones) is negligible. Sedimentary processes commonly take place at the front and atop of a nappe stack in collisional and intra-continental deformation zones (Types 5, 6a1 and 6a2 in Tab. 2, and Fig. 2E; e.g., Dilek et al., 1999; Remitti et al., 2011; Hernaiz Huerta et al., 2012; Ogata et al., 2012a, 2012b) of ancient, submarine collisional orogens as in the “Alpine-Himalayan” chains (e.g., Abbate et al., 1970; Smith et al., 1979; Liu and Einsele, 1996; Marroni and Pandolfi, 2001; Burg et al., 2008), and modern ones as in the Gela Nappe in the Sicily Channel (Tricardi and Argnani, 1990; Minisini et al., 2009) and Adriatic Sea (Tricardi et al., 2004; Argnani et al., 2011). Their occurrence in exhumed accretionary wedge and in subduction settings (Type 4a in Tab. 2, and Figs. 2A, 2C, and 2D) is relatively minor, although it has been recently re-evaluated in some on-land examples of ancient accretionary complexes (e.g., Collot et al., 2001; Burg et al., 2008; Yamamoto et al., 2012b). These mélanges occur at the wedge front of subduction settings (Yamada et al., 2010) and are mainly related to subduction of seamounts and to the reactivation of normal faults (Fig. 2A) in a downgoing plate (see e.g., Marroni and Pandolfi, 2001; Martinez Catalan et al., 1997). Some may also form when accretion is replaced by tectonic erosion at a convergent margin (e.g., von Huene and Lallemand, 1990; Ranero and von Huene, 2000; von Huene et al., 2004; Remitti et al., 2011). Sedimentary processes might also have been responsible for the formation of different types of sub-aerial mélanges (Type 7a in Tab. 2), such as debris flow and avalanches, alluvial fan deposits, talus breccias (scree deposits) and megabreccias, block falls, and glacial till (see Hoffmann and Piotrowski, 2001).

3.2. Tectonics as a prominent triggering mechanism

The occurrence of different types of sedimentary mélanges in most geodynamic environments (Tab. 2) could be simply related to the fact that sedimentary processes are more efficient in terms of conservation of kinetic energy in comparison to tectonic and diapiric ones. These processes may also play a prominent role in maintaining the dynamic equilibrium in active tectonic settings (e.g., frontal erosion in accretionary complexes, slope failure on steep margins of carbonate platforms or passive margins). However, tectonic processes, rather than sedimentary or diapiric ones, constitute the most effective triggering mechanisms (both directly and indirectly) (see Tab. 2). Hence, they play a primary role in controlling the processes and mechanisms of stratal disruption and mixing, and in the formation of tectonic, sedimentary or diapiric mélanges (Fig. 5).

The direct role played by tectonics is achieved mainly by seismic faulting associated with strike-slip or contractional deformation (Fig. 5). Faulting is an effective mechanism of disruption of a coherent stratigraphic succession (e.g., Cowan, 1974, 1985; Vollmer and Bosworth, 1984; Karig et al., 1986; Needham, 1995; Rassios and Dilek, 2009; Ghikas et al., 2010; Festa et al., 2010a). The superposition of displacements along innumerable subparallel, meso-scale shear faults and fractures develop zones of distributed shear from several meters to kilometers in width (Coleman and

Lanphere, 1971; Suppe, 1972; Pettinga, 1982; Moore and Sample, 1986; Kimura et al., 1996; Kusky et al., 1997; Ogawa, 1998; Meneghini et al., 2009; Bradbury et al., 2011). This process represents an effective mechanism of formation of tectonic mélanges and broken formations with different end-members, from brittle broken formations to flow mélanges according to the tectonic setting and structural level of their formation.

In poorly- or non-consolidated sedimentary successions, faults represent the preferential pathways for the upward rise of overpressured fluids (Fig. 5) that facilitate *in situ* stratal disruption, diapiric deformation and related processes. These fluids are able, in turn, to increase the driving forces along the slope, inducing gravitational processes and the formation of sedimentary mélanges.

Tectonic processes can also play an indirect role in triggering stratal disruption and mixing in most geodynamic settings (Fig. 5). Tectonic activities can trigger mass-transport processes by both (1) reducing the shear strength of sediments (e.g. higher sedimentation rates, gas hydrates dissociation, etc.) and, thus decreasing the resisting forces along the slope, and (2) magnifying the effect of other driving mechanisms, processes and events along the slope (e.g., failure by slope oversteepening, mud diapirism and mud volcanism, sea level fluctuation, etc.).

Sedimentary instability may be caused by the upward rise of over-pressured fluids from a subduction zone. The upward rise of these fluids is commonly related to tectonic loading (Fig. 5) along the decollement surface developed at the toe of an accretionary wedge (e.g., Brown and Westbrook, 1988; Brown, 1990) or to fluids pumped-up along strike-slip faults (e.g., Dela Pierre et al., 2007). The abrupt emplacement of mass-transport chaotic bodies can strongly increase the magnitude of sedimentary loading, causing overpressure and consequent sediment liquefaction (Figs. 4G and 4H), which in turn induces diapiric processes forming diapiric mélanges (Fig. 5). The emergence of diapiric bodies (e.g., sedimentary diatremes, mud volcanoes or diapirs) may create instability in unconsolidated material and then gravitational movement along the slope, forming mass-transport chaotic bodies (e.g., Barber et al., 1986; Camerlenghi and Pini, 2009). This complicated interplay of different processes induced mainly by tectonics is strongly controlled, in each tectonic setting, by the physical conditions (e.g. water content, overpressure, P-T conditions, etc.), the nature and state of consolidation of sediments, and the burial depth or structural level at which broken formations and mélanges form (see below).

4. Mechanisms of stratal disruption, mixing and related chaotic products

At shallow structural depths in different tectonic environments, the final structural texture and the fabric of chaotic rock units are commonly achieved through progressive deformation of originally coherent stratigraphic successions (stratal disruption), and through series of interacting or overlapping mechanisms (Fig. 6). This progressive deformation is directly controlled by the increase of the degree of consolidation with burial, or with the increasing depth of the structural level in which these processes commonly operate (see Tab. 2, and Fig. 7). Consolidation controls the change of mechanical strength of sediments from deposition to progressive burial (e.g., Lash, 1989; Jones et al., 1991); it is time-dependent and closely related to changes in pore-volume, expulsion of pore-fluid, and interaction and packing of grain particles (e.g., Maltman, 1994; Maltman and Bolton, 2003). Then, the occurrence of poorly consolidated sediments in the shallow part of accretionary prisms or sedimentary piles favors gravitational deformation, whereas with the downward increase of consolidation at depth, the deformation related to tectonic forces becomes gradually more significant (Fig. 6, see also Maltman, 1994). However, tectonics and related stress conditions may greatly affect this linear relationship between consolidation and structural or burial level (see below), changing the local physical properties of sediments (e.g., permeability, strength; see Maltman, 1994; Michiguchi and Ogawa, 2011).

4.1. Sedimentary stratal disruption and mixing

At shallow structural levels, sedimentary mass transport processes are the most efficient causes for stratal disruption (Figs. 6A and 6B), occurring both inside of a sliding body of rocks (e.g. via partial disaggregation of still stratified blocks) and outside (e.g. within the uppermost portion of the overridden substrate) during its downslope motion. This kind of deformation commonly involves poorly-lithified or loose material, and results in the formation of a broad spectrum of structures (Figs. 6A and 6B) ranging from folded and boudinaged successions (e.g. slump deposits) to block-in-matrix bodies (e.g. debris flow deposits; Figs. 6A and 6B). These products are characterized by the occurrence of a strongly mixed, liquidized matrix (i.e. hyper-concentrated suspension *sensu* Mutti, 1992) enveloping disrupted layers and blocks. Those layers or blocks may show different degrees of lithification, and represent the final artifacts of progressive down-slope, soft sediment deformation (e.g., Maltman, 1994; Ogata, 2010). The latter deformation is enabled by the relative movement (i.e., fast vs. slow) of unlithified masses with progressive flow transformation, stratal disruption, or both, of the partially-to largely lithified sediments (Pini et al., 2012).

The different velocities of movement, the composition and lithification degree of sediments (related to the stratigraphic level of the rupture surface), the geometry and morphology of the depositional setting, and the mode of failure propagation (progressive vs. regressive collapse) commonly control the nature of the transition between different types of mass-transport chaotic bodies (Fig. 8A). Pini et al. (2012; see also Lucente and Pini, 2003) distinguished three main types of mass-transport chaotic bodies, representing the end-members of a continuum of chaotic products and displaying different characteristics. First, viscous-flow, which is dominated by shearing in fine-grained sediments, is responsible for the movement and emplacement mode of classic olistostromes (Figs. 8A and 8B; see also Figs. 4D, 4E and 4F), which are characterized by centimeter-to meters sized hard blocks that are randomly distributed in a mud-rich, brecciated matrix (Fig. 8C) (e.g., Swarbick and Naylor, 1980; Abbate et al., 1970, 1981). Commonly, at the base of these bodies, a shear zone may form accommodating the flow of sediments (e.g. Pini, 1999; Ogata et al., 2012a) and deforming poorly consolidated blocks (Figs. 8A and 8D). These bodies may assume different shapes depending on the deformation style (flattening vs. simple-shear) and strain magnitude (see Type 1 MTC of Pini et al., 2012). Second, overpressure of fluids sustaining mud-silt-sandy sediments, controls the down-slope movement of hyper-concentrated suspension (*sensu* Mutti, 1992) characterized by a block-dominated part overlying a matrix-dominated one (Fig. 8A; see Ogata, 2010; Ogata et al., 2012a, 2012b). Third, narrow and over-pressured shear zones (millimeters-to decimeters thick, see Dykstra,

2005) allow the emplacement of chaotic, sandy sediments displaying folds, boudinage, extensional and contractional duplexes, and showing a gradual downward increase of stratal disruption (Figs. 8A, 8E and 8F; see Pini et al., 2012). Localized zones of liquefaction of sandy sediments can also be locally related to the emplacement of these chaotic bodies (see Lucente and Pini, 2003 for detail).

In all these types of mass-transport chaotic bodies, the mixing of rocks and the incorporation of exotic blocks are controlled by two fundamental factors. The first one is the depth reached by the slope failure and its propagation toward the basin margins. Mixing of the rocks derived from the basin margins with those sediments in the basin is a common process in the formation of mass-transport bodies (Page and Suppe, 1981; Callot et al., 2008; Ogata et al., 2012c). During deposition, these "exotic" rocks were extraformational, extrabasinal, older and much more consolidated than the basin sediments. The second factor involves the exhumation and uplift and the subsequent reworking of older rocks. The emplacement of submarine nappes can supply extrabasinal blocks of different size, centimeters to hundreds of meters, to mass-transport deposits in the foredeep basins. Originated from different structural units, these blocks are composed of rocks that were completely consolidated, tectonically deformed and metamorphosed at the time of deposition. Classic examples include the precursory olistostromes in the Apennines of Italy (see, e.g., Abbate et al., 1970, 1981; Elter and Trevisan, 1973; Lucente and Pini, 2008 and references therein), the Porma mélange in the Cantabrian chain (Alonso et al., 2006), the "wildflysch" of the Alps (Trümpy, 2007) and the "klippen zones" in the Carpathians (Camerlenghi and Pini, 2009 and references therein; Cieszkowski et al., 2012; Ślączka et al., 2012). The same scenario could occur in oblique subduction (Hernaiz Huerta et al., 2012) and transpressional tectonic zones (Marroni et al., 2001). In the Franciscan Complex, the HP (and medium pressure) blocks, hosted by a low metamorphic epizonal matrix, can be explained as clasts (see Cowan and Page, 1975) eroded from exhumed blueschists facies rocks and deposited as debris flows and avalanches in the accretionary prism front (Erickson, 2011; Wakabayashi, 2011). Exhumation of subduction channels, such as in Timor and Taiwan, represents the source of exotic blocks of different metamorphic grades and mantle origins (e.g., Guillot et al., 2009; Ota and Kaneko,

2010) that may be mixed together by mass-transport processes forming sedimentary mélanges or supplying UHP-HP "knockers".

4.2. Diapiric stratal disruption and mixing

Liquefaction of sediments is a primary factor in controlling the downslope mobilization of unconsolidated or incompletely lithified sediments (Maltman and Bolton, 2003), but it is also highly important for *in situ* stratal disruption processes characterized by relatively limited transport of material. At shallow crustal-levels, injectites and seismites (Figs. 4G, 4H and 6B) may develop in response to slope tectonics and seismic shocks (at least at the micro-and meso-scales). Yamamoto et al. (2009) described some notable examples of these structures from the Miura-Boso accretionary complex that formed during the early stages of accretion (Fig. 4G; Central Japan) and Codegone et al. (2012a) from the Hamburg Klippe in the central Appalachians (Central-eastern Pennsylvania). These authors described some good examples of injectites and seismites intruding the overlying and underlying sandy layers, and showing a randomly oriented ”block-in-homogeneous sandy matrix”. These examples lack sedimentary features such as lamination, grain-size grading or small-block preferred orientation (Fig. 4G). Another important diagnostic feature of injectites and seismites is the lack of a basal erosive surface and internal slip planes. The injectites and seismites in other chaotic bodies are characterized by a liquefied matrix and constitute hyper-concentrated density flows (Lucente and Pini, 2003; Ogata, 2010).

With an increase of the consolidation degree and the rheological contrast between the layers of the stratigraphic succession, an abrupt increase of tectonic or lithostatic loading, gas hydrates dissociation, density inversion, and diagenetic transformation (Kopf, 2002) may cause the overpressurization of fluids at relatively deeper structural levels. Over-pressured fluids may then result in the development of sedimentary diatremes (e.g., Borgia et al., 2006), mud-volcanoes (e.g., Kopf, 2002; Camerlenghi and Pini, 2009) and diapiric bodies (Fig. 4C; e.g., Barber et al., 1986; Orange, 1990; Festa, 2011) of unconsolidated sediments. These structures may show a great diversity stemming from the origin of the fluid phases (e.g., Kopf, 2002; Camerlenghi and Pini, 2009). Diapiric processes are widespread as subordinate processes in most tectonic environments (see Tab. 2, Fig. 2) and occur where the necessary physical and mechanical conditions (such as fluid overpressure) exist (e.g., Brown and Westbrook, 1988; Kopf, 2002; Dela Pierre et al., 2007; Camerlenghi and Pini, 2009; Festa, 2011). The mixing of exotic blocks is mainly achieved by a combination of hydrofracturing processes and the progressive incorporation of the wall-rock material (collapse and assimilation of the roof and margins of the structure) and flow. Notable examples of mud volcanoes with exotic blocks of blueschist rocks contained in serpentine-dominated muds have been described from the forearc region of the active Mariana subduction zone (Maekawa et al., 1993; Fryer et al., 1999). These exotic blocks, originated from the metamorphosed subducted plate, were entrained in rising serpentine mud diapirs (up to 30 km wide and 2 km high), and were then extruded from the mud volcanoes onto the sea floor.

Blocks may be also derived from the previously formed underlying mélanges (Camerlenghi and Pini, 2009). Cyclic diapiric reactivation of the previously formed sedimentary or tectonic mélanges (Fig. 4C) may also occur when the physical conditions are sufficient, leading to the formation of a complex polygenetic mélange (Barber et al., 1986; Henry et al., 1990; Brown and Orange, 1993; Cronin et al., 1997; Camerlenghi and Pini, 2009; Festa et al., 2010a; Festa, 2011; Codegone et al., 2012b).

4.3. Tectonic stratal disruption and mixing

Immediately after deposition, sediments may start undergoing deformation due to the interplay between gravitational forces and tectonic stresses during progressive burial (see Maltman, 1994). There exists an overlapping zone at shallow structural levels, where the block-in-matrix fabric of broken formations shows a strong convergence of fabric with sedimentary mélanges.

In general, layer-parallel extension occurring in all directions records a coaxial strain history compatible with sedimentary processes on gently dipping slopes (Fig. 9A and 9A1; e.g., Lash, 1987, 1989; Cowan, 1985), whereas layer-parallel shearing records a non-coaxial strain history commonly related to extensional slicing across a basal shear zone and underthrusting (e.g., Cowan, 1985;

Byrne, 1984; Fisher and Byrne, 1987; Hashimoto and Kimura, 1999). The development of layer-parallel extension and shearing may change as a function of the relationships between the consolidation degree, dewatering processes, and the magnitude of strain occurring in the tectonic setting, whereas tectonic loading may produce layer-parallel extension with coaxial strain on sub-horizontal bedded succession.

Tectonic loading related to pre-thrusting deformation during the advancement of an accretionary wedge or a continental nappe stack is an effective mechanism in triggering dewatering and fluid expulsion (Breen et al., 1986; Harris et al., 1998) (Fig. 6C). These fluid-driven processes result in layer-parallel disruption and in the development of boudinage or dismemberment of the most lithified layers due to hydrofracturing and fluid overpressure (e.g., blocky veins, see Meneghini et al., 2009; web-like fragmentation, see Kimura et al., 2012; brecciation, etc.). Depending on the degree of consolidation and rheological contrasts within a stratigraphic succession (see Bettelli and Vannucchi, 2003), layer-parallel disruption can evolve into an incipient foliation, formation of a scaly fabric in the less competent layers (e.g., claystone, limestone, mudstone), and development of a progressive boudinage structure in competent layers (Figs. 9A; e.g., sandstone; see Lash, 1989; Kimura and Mukai, 1991; Onishi and Kimura, 1995). Pinch-and-swell structures and irregular boudinage features defining ellipsoidal-shaped blocks are commonly related to coaxial strain that induced heterogeneous flattening in all directions (Figs. 9A and 9A1; Harris et al., 1998; Pini, 1999). Regular boudinage features form as a result of non-coaxial strain and may produce lozenge-to sigmoidal shaped blocks (Figs. 9B and 9B1).

With the gradual increase of consolidation, conjugate extensional fractures may develop into symmetrical brittle boudinage structures, which may turn into asymmetric shear planes with increased shearing (Figs. 9F, 9F1, 9G, 9G1 and 9G2; see also Kimura et al., 2012). Boudinage structures may also form due to the propagation of Y, P, R, R‟ shear surfaces (Figs. 9G, 9G1 and 9G2; see also Needham, 1995; Pini, 1999). In lithified sediments, extensional veining (Figs. 9C and 9C1), cataclastic deformation and brecciation at necks and tails of boudins (Figs. 9D and 9D1), and asymmetric veins (Fig. 9E) may develop as a result of a sequential process of cataclasis, fracturing, and Riedel shearing leading to boudinage formation (see Kimura et al., 2012). The style and degree of block fragmentation may change as a function of block aspect ratio (see Needham, 1995), whereas at seismogenic depths the increase of diagenesis and metamorphic grade may also change the shape of boudinaged blocks from elongate and oblate to a spherical shape (Kimura et al., 2012).

With the increase of shear and consolidation, mixing process in fault/shear zones cause mechanical crushing of the hanging and footwall rocks, which then becomes progressively incorporated into an evolving tectonic mélange (Figs. 4E and 6D; Cowan, 1974, 1985; Pettinga, 1982; Barnes and Korsch, 1991; Onishi and Kimura, 1995; Ogawa, 1998). Flattening of subcreted thrust sheets, as they are detached from the footwall, also represents an effective mechanism of disruption of fractured layers into broken strata and mélange, as for example along the Sonnebad Disruption Zone in the Timor region of Indonesia (Harris et al., 1998). Tectonic mélanges formed in these ways commonly display (Tab. 2; Fig. 2) a pervasive, scaly fabric, which is most pronounced in fine-grained lithologies (Figs. 3C and 3D). At relatively deeper structural levels in an accretionary prism, faulting (e.g., Cowan, 1985; Wakabayashi, 1992; Pettinga, 1992; Needham, 1995; Ogawa, 1998) and folding (e.g., Moore, 1973; Onishi and Kimura, 1995; Kusky and Bradley, 1999; Vannucchi and Bettelli, 2002; Bettelli and Vannucchi, 2003) are the main mechanisms of stratal disruption (Figs. 6E1, and 6E2). Deformation here is concentrated within fault and shear zones (Figs. 3F), and tectonic thickening occurs due to duplexing, antiformal stacking (Figs. 6D, and 6F) and out-of-sequence thrusting (e.g., Pettinga, 1982; Needham, 1995). This phenomenon explains why tectonic mélanges with exotic blocks occur exclusively in different scale shear zones, as discussed earlier. Localized fault zones (Fig. 3F) are strictly responsible for large-scale tectonic mixing processes (e.g., Cowan, 1974; Festa et al., 2010a).

Although in this study we focus only on mélanges formed at shallow structural levels, at deeper levels diagenetic and metamorphic processes significantly influence the deformational style by enhancing the competence contrast between different layers through the formation of new mineral phases and by dehydration of the clay minerals (e.g. increasing pore pressure). Diagenetic and metamorphic processes collectively favor the occurrence of mixing processes, forming tectonic mélanges (Fig. 2C). Brittle or semi-brittle deformation becomes gradually replaced by ductile deformation with the progressive increase of the temperature and pressure (e.g., Blanco-Quintero et al., 2011). Within a subduction channel (Shreve and Cloos, 1986), for example, tectonic processes facilitate the formation of flow mélanges, and control the upward trajectories of exotic rocks (blueschist or serpentinized peridotite) (Cloos and Shreve, 1988b; Federico et al., 2007; Blanco-Quintero et al., 2011) and large-scale serpentinite diapirism (e.g., Maekawa et al., 1993; Fryer et al., 1999). Here, the buoyancy of subducting sediments affects the flow, and the deformed sediments become incorporated into the upper plate of the subduction channel (Fig. 2C). The block-in-matrix fabric and the pattern of underplating depend on the shear stress distribution along the hanging-wall of the subduction channel, the sediment supply along this channel, the geometry, properties and permeability of the overriding units, and particularly the nature of the back-stop (Shreve and Cloos, 1986; Cloos and Shreve, 1988a, 1988b).

4.4. Small-scale and localized horizons of stratal disruption and mixing

All the examples described so far show that a “continuum” of stratal disruption and mixing may exist (Figs. 6, 8 and 9), recording the history of progressive burial and shear strengthening at shallow structural levels (see also Needham, 1995). However, the development of tectonic surfaces, the occurrence of impermeable barriers or strong rheological contrasts within stratigraphic successions, or both, can define thin horizons of deformation zones (up to tens of meters thick) that can affect the progressive increase of stratal disruption related to the burial conditions (Fig. 3F; e.g., Bosworth and Vollmer, 1981; Bosworth, 1989; Lash, 1989; Byrne, 1994). Within these horizons, fluid pressure can increase up to overpressure point, a level of which drives the sediments toward a critical state condition (Maltman and Bolton, 2003), promoting stratal disruption and mixing processes that produce broken formations and mélanges. The physical superposition of these horizons, as well as that of different mechanisms and processes, further complicate the above described “continuum” of stratal disruption and mixing, and can favor the formation of polygenetic mélanges. Although the last pervasive process commonly obliterates the products of the previously formed ones (e.g., Raymond, 1984; Raymond et al., 1989; Ogawa, 1998; Dela Pierre et al., 2007; Festa et al., 2010a; Festa, 2011; Codegone et al., 2012b), polygenetic mélanges may display a continuum of stratal disruption and mixing shown by the superposition of different products formed by tectonic, sedimentary and diapiric processes. The understanding of the interplay and superposition of these different processes is of primary importance in understanding the mechanisms of broken formation, mélange, and polygenetic mélange development as well as the evolution of the tectonic setting in which they formed.

5. Conclusions

A redefinition of the terms *broken formation* and *mélange* and a clearer distinction between them allow us to extend these definitions to a more diverse occurrence of chaotic rock bodies developed in different tectonic settings. Tectonic events represent the most prominent triggering mechanism inducing, directly or indirectly, different processes of stratal disruption and mixing that produce a broad spectrum of chaotic rock bodies. The block-in-matrix arrangement in these chaotic rock bodies is mainly controlled by a linear relationship between the degree of consolidation (including tectonically-induced compaction) and progressive burial (Fig. 10). As a result, a continuum of processes and deformation structures gives rise to gradual disruption and mixing processes that are significant for the development of broken formations and mélanges.

At shallow structural levels in tectonically active environments the occurrence of poorly consolidated sediments favors gravitational deformation. Sedimentary (mass-transport) and diapiric chaotic products record punctuated and instantaneous stratal disruption features, which provide important clues about the physical conditions of their formation (e.g., consolidation, fluid pressure, changes of pore-volume, expulsion of pore-fluid and strength of sediments), and about the evolution of their tectonic setting of formation (Fig. 10). In the geological record, the occurrence of these sedimentary and diapiric chaotic bodies and their tectonic or stratigraphic relationships with other chaotic bodies and coherent stratigraphic successions allow us to better constrain the changes in the dynamic equilibrium in a geological setting. A good example of the control exerted by these processes on the dynamic equilibrium would be the switch from an accretionary tectonics to an erosion tectonics at the wedge front of accretionary complexes (e.g., von Huene and Lallemand, 1990; von Huene et al., 2004; Remitti et al., 2011).

With the downward increase of consolidation at depth, the deformation related to tectonic forces becomes gradually more significant. Tectonically broken formations and mélanges record a continuum of deformation that occurs through time and different degrees of lithification during a progressive increase of the degree of consolidation and of the diagenetic and metamorphic mineral transformation (Fig. 10). At shallow structural levels the sediments are affected by a brittle to more ductile deformation that follows their progressive dewatering and strengthening as a result of burial. At deeper structural levels, diagenetic and metamorphic mineral transformation accompanies deformation patterns that are controlled strongly by the increase of P-T conditions. Several tectonic, chaotic products may record a deeper progressive evolution in a continuum of deformation that is related to several mechanisms of stratal disruption and mixing (i.e., *in situ* stratal disruption, faulting, shearing, thrusting and faulting).

The superposition of different mechanisms and processes of disruption and mixing of rocks in some tectonic settings may lead to the reworking of existing mélange products and to the formation of polygenetic mélange types. The previously formed chaotic products may then change their block-in-matrix arrangement according to the last deformation style, strain rate, stress direction, alternating coaxial and non-coaxial strain paths, and variations in consolidation degrees. Polygenetic mélanges may thus provide useful information on their multiphase evolution (e.g., Festa, 2011; Osozawa et al., 2011), the spatial and temporal relationships between the physical conditions (e.g., burial loading, porosity and fluid pressure), and the mechanisms and processes that acted in the depositional environment or within the sedimentary succession where they formed.

None of the geological processes forming mélanges operates in isolation. They commonly interact in a continuum of stratal disruption and mixing processes (Figs. 5 and 6). These processes and their mode are also strongly controlled by the balance between the hydrological activities and the rate at which the fluids are produced by burial-related consolidation, mineral dehydration mechanisms, diagenesis, and metamorphism in any given tectonic setting (e.g., Byrne, 1994).

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CAPTIONS TO TABLES AND FIGURES

Table 1 – Deterministic characters of broken formations and mélanges, representing two end members involving the nature of blocks (native vs. exotic) and mechanisms (*in situ* stratal disruption vs. mixing).

Table 2 – Subdivision and classification of mélanges and broken formations on the basis of their geodynamic setting of formation, processes, triggering mechanisms, products and mesoscale characteristics (modified after Festa et al., 2010a). Acronyms are listed at the bottom of the table.

Figure 1 – Global distribution of mélanges and mélanges terrains.

Figure 2 – Conceptual model for the formation and emplacement of mélanges associated with (A) extensional tectonics (type 1 mélanges), passive margin (type 2a mélanges), ocean-continent transition settings (type 2b mélanges) and convergent margins (type 4 mélanges). Different models and cases of subduction settings are shown: (A) open-double verging wedge with a low elevation backstop; (B) obduction of ophiolites (modified after Rassios and Dilek, 2009); (C) close wedge and subduction channel (modified after Cloos, 1982); (D) close and smaller wedge with an high elevation of the backstop; (E) collisional tectonics (type 5 mélanges; modified after Huang et al., 2008; Ghikas et al., 2010; Festa et al., 2010a), intra-continental deformation (type 6 mélanges), and (F) strike slip tectonics (type 3 mélanges).

Figure 3 - Different examples of broken formations and mélanges: (A) layer-parallel extension in the *Argille varicolori* displaying lozenge-shaped boudins of red clayey marl enveloped in greyish matrix (broken formation) (Aventino valley, Abruzzi region, Central Apennines of Italy; photograph by E. Malerba); (B) progressive stratal disruption of well bedded units (Flysch Rosso) forming lozenge-shaped boudins of mudstone in a clayey marl matrix (broken formation) (Aventino valley, Abruzzi region, Central Apennines of Italy); (C) progressive stratal disruption of well bedded units (Subligurian

Eocene Canetolo Complex) forming broken formation (Corniglia, La Spezia, western coastal exposures of the Northern Apennines of Italy); (D) lozenge-shaped boudins of sandstone within a mudstone matrix displaying a pervasive scaly fabric (broken formation), due to transposition of upright beds in a fault zone related to an out-of-sequence thrust (Waimarama Beach, South Hawke"s Bay, East Coast of North Island, New Zealand); (E) phacoidal Upper Triassic pelagic limestone blocks in a heterogeneous and variously deformed matrix composed of shale, mudstone, and sandstone in the Jurassic-Cretaceous Avdella mélange (Pindos Mountains, Northern Greece); (F) narrow, anastomosing and coalescent fault zone including exotic blocks of sandstone and mudstone in a shaly limestone matrix (Taconic mélange) (Hoosic River at Schaghticoke Gorge, eastern NY, Central Appalachian - USA)

Figure 4 – Different examples of polygenetic mélanges, diapiric and sedimentary mélanges: (A) sedimentary mélanges overprinted by tectonic deformation forming a polygenetic mélange in the footwall of the Taconic Allochthon (Northern Appalachians, USA). Exotic blocks (with respect to the shaly limestone matrix) of sandstone, mudstone and chert show a lenticular shape resulting from tectonic shearing (Hoosic River at Schaghticoke Gorge, eastern NY, Central Appalachians – USA); (B) Exotic blocks of sandstone and volcanic rocks showing an elongated shape within a shaly matrix with fluidal feature (tectonic mélange or sheared olistostrome?) (Esk Head mélange, Okuku River, New Zealand); (C) small-scale diapiric body overprinting a previously formed broken formation (polygenetic mélange) in the footwall of the Taconic Allochthon. Red lines bound the margin of the diapiric body. Note the vertical reorientations of blocks enveloped in a fluidal scaly fabric (Poestenkill Gorge at Troy, eastern NY, Central Appalachians – USA); (D) olistostrome of the uppermost portion of the Oligocene Macigno Costiero Formation (precursory olistostrome) cropping out in the Cinque Terre area (La Spezia, westernmost Northern Apennines of Italy); (E) Upper Cretaceous olistostrome in the external Ligurian units, flattened and slightly deformed by compaction and tectonics (Berceto, Parma area of the Northern Apennines of Italy); (F) Lower Miocene olistostrome (Val Tiepido – Canossa olistostrome) of the wedge-top Epiligurian successions. Note the random distribution of hard

block in a marly matrix (Costa del Vento, Montalto P.se area of the Norhern Apennines of Italy); (G) and (H) liquefied sediments into coherent layers by *in situ* injection (Kaitocho, Miura Peninsula, Japan).

Figure 5 – Diagram showing the direct and indirect role of tectonics as a major triggering mechanism in the formation of mélanges.

Figure 6 – Composite diagram showing the continuum of processes of stratal disruption. Arrows indicate the genetic link and the continuum of dismemberment processes from shallow to deeper structural domains and from sedimentary to tectonic processes. Left and right pictures in Figure A are modified from Yamamoto et al. (2009) and Cowan (1985), respectively. Pictures in B, C, E1 and E2, modified from Yamamoto et al. (2009), Meneghini et al. (2009), Cowan and Pini (2001), and Bettelli and Vannucchi (2003), respectively.

Figure 7 – Schematic diagram showing the progressive increase of the consolidation degree (and decrease of fluid production) with depth. Note that consolidation is time-dependent. Modified after Collison (1994) and Brown (1994).

Figure 8 – (A) Progressive transition of stratal disruption and mixing processes in mass-transport chaotic complexes (modified after Mutti et al., 2006; Ogata et al., 2012a; Pini et al., 2012). Three main types of chaotic bodies are formed during the progressive increase of lithification and are characterized by a gradual decrease of matrix amount from debris flows to block flow and slide/slump bodies (see text for explanation). (B) Example of debris flow (Val Tiepido – Canossa olistostrome at Mt. Penola, Val Curone, Northern Apennines of Italy). The arrows indicate the erosive basal surface that is characterized by a decimeters-thick shear zone accommodating the flow of sediments (D). Away from the basal shear zone, the hard blocks are randomly oriented within the clayey matrix (C). (E) and (F) examples of slumping and related boudinage in the *Argille varicolori* of External Ligurian

Units (Montalto P.se) and Marnoso arenacea Fm. (Passo dei Mandrioli) in Northern Apennines of Italy.

Figure 9– Progressive stratal disruption forming different types of broken formations by layer-parallel extension (see text for explanation). Stratal disruption is controlled by the degree of consolidation and lithification and the increase of shear (see vertical and horizontal arrows). (A) At shallow structural levels where sediments are non- to poorly-lithified, pinch-and-swell and boudinage structures are formed by coaxial strain. Deformation acts in different ways on the basis of the rheology and nature of the bedded succession, inducing heterogeneous flattening. (A1) Photograph showing an example of heterogeneous flattening (Marnoso arenacea Fm. at Passo dei Mandrioli, Northern Apennines of Italy). (B) With the increasing shear, non-coaxial strain forms lozenge- to sigmoidal-shaped blocks as show in the *Argille varicolori* of photograph (B1) (Monteu da Po, Tertiary Piedmont Basin, NW Italy). The increasing amount of lithification is coupled by different mechanisms of stratal disruption as, for example, (C and C1) veining, (D and D1) brecciation in the neck and tails of blocks, (E) veining along the border of the blocks (modified after Pini, 1999), and (F and F1) extensional fracturing. The increasing shearing (G, G1, and G2) forms asymmetrical brittle boudinage with the development of Y, R, R" and P shear surfaces. Photograph localities are: (C1) Taconic flysch at Schaghticoke Gorge, eastern NY, Central Appalachians – USA, (D1) "Messinian mélange" (see Festa, 2011) in the Tertiary Piedmont Basin, NW Italy, (F1) Broken formation in the Hamburg Klippe of Eastern Pennsylvania (south of Albany, Berks County, USA), (G1) Taconic flysch at Schaghticoke Gorge, eastern NY, Central Appalachians – USA, (G2) *Argille varicolori* in the External Ligurian units at Brusasco, Tertiary Piedmont Basin, NW Italy.

Figure 10 – Schematic diagram, showing a conceptual difference between depositionally (gravitational), diapirically and tectonically induced deformation with respect to the consolidation. Sedimentary and diapiric chaotic bodies may record only instantaneous and episodic events that punctuate the consolidation history, whereas tectonic chaotic bodies may record different stages of

deformation that persist through time and different degrees of consolidation and lithification (modified after Byrne, 1994).

Table 1

<table>
<tr><th>Products</th><th>Nature of blocks</th><th>Mechanisms</th><th>Lithological unit involved</th><th>Contacts with host rocks</th><th>Processes</th></tr>
<tr><td>Mélange</td><td>Exotic and Native</td><td>Mixing</td><td>Sedimentary
Metamorphic
Igneous</td><td rowspan="2">Tectonic
Stratigraphic
Intrusive</td><td rowspan="2">Tectonic
Sedimentary
Diapiric</td></tr>
<tr><td>Broken Formation</td><td>Intra-formational</td><td>Stratal disruption</td><td>Sedimentary
Metamorphic
(?)</td></tr>
</table>

Table 1 – Festa et al.

Table 2

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Types of Mélange related to:	**Geodynamic environments**	**Processes**	**Triggering mechanisms**	**Products**	**Mesoscale characteristics**	**Minor related products**
1. Extensional tectonics	Rifting	MTP (debris avalanches and flows, etc.)	Tectonic	MTD (megabreccias, breccias, olistolith fields, debrites, slide blocks, etc.)	Chaotic angular clasts (cm to >10 m) in fine-grained (pelitic) matrix	Fault zones along normal fault?
2. Passive margin **a.** ***Downslope mass-transport deposits***	Passive margins (after rifting)	SSD and mass-wasting related progressive deformation from slumping to debris flow, to complete strata disruption	Tectonic, sedimentary	MTD, poorly sorted olistostromes (olistoliths, slide blocks)	Chaotic monomictic brecciated (matrix-supported) masses	*In situ* fluidification: mud diapirs?
b. ***Mass-transport deposits at the ocean-continent transition (OCT)***	Ocean-continent transition	SSD and MTP with related progressive deformation from slumping to debris flows, to gravitational sliding	Tectonic, sedimentary	MTD, olistostromes with continent rock olistoliths (tens of meters to several km slide blocks) in a matrix of oceanic origin	Chaotic polymictic brecciated (matrix-supported) masses (including native, extra-basinal and/or exotic blocks)	
3. Strike-slip tectonics and transform setting	Different types of collision	TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic	BrFm; mélanges (exotic blocks being commonly recycled from other previously formed mélanges)	Structurally ordered BIM fabric (parallel orientation of blocks and matrix features – i.e. pseudo-bedding)	Olistostromes *s.l.*; mud diapirs *s.l.*
4. Convergent margins and oceanic crust subduction **a.** ***Mass-transport deposits at the wedge front***	Subduction (at the front of the wedge)	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (including native, extra-basinal and/or exotic blocks)	Mud diapirs and mud volcanoes, serpentinite
b. ***Broken fms and tectonic mélanges***	Subduction (at the base of the wedge) and subduction channel	TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation); tectonic mixing	Tectonic	BrFm; mélanges (exotic blocks being recycled from other previously formed mélanges or formed by subduct. channel processes)	Structurally ordered BIM fabric (parallel orientation of BIM features – i.e. pseudo-bedding)	
5. Collision	Different types of collision	TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic	BrFm; méanges? (exotic blocks being commonly recycled from other previously formed mélanges)	Mainly structurally ordered BIM fabric (that in some cases overprinted chaotic BIM fabric)	Olistostromes *s.l.*; mud diapirs *s.l.*
6. Intracontinental deform. **a.** ***Mass-transport deposits at the wedge front*** a1. Precursory olistostromes	At the base or at the front of intra-continental thrust sheets or nappes	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (from matrix-supported cm-to m in size blocks to clast supported >10 m blocks and olistoliths)	Mud diapirs and mud volcanoes
a2. Olistostromal carpet		TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic, sedimentary	Mélanges (exotic blocks being commonly recycled from other previously formed sedimentary mélanges); BrFm	Chaotic BIM fabric overprinted by tectonic deformation and shearing	
a3. Tectonic mélanges			Tectonic		Structurally ordered BIM fabric	
b. ***Intra-nappe*** b1. Sedimentary	Within intra-continental thrust sheets or nappes	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (blocks of intra-basinal origin)	Mud diapirs and mud volcanoes
b2. Tectonic and/or tectono-sedimentary		TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic	BrFm; mélanges (exotic blocks being commonly recycled from other previously formed sedimentary mélanges)	Structurally ordered BIM fabric (parallel orientation of blocks and matrix features – i.e. pseudo-bedding)	
c. ***Epi-nappe*** c1. Sedimentary	A top of intra-continental thrust sheets or nappes (e.g. piggy back, top thrust basins)	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (originated from the succession tectonically imbricated in the thrust-sheet)	Mud diapirs and mud volcanoes
c2. Tectono-sedim.		TSD (overprinting previous mass-wasting-related deformation)	Tectonic, sedimentary	BrFm, mélanges	Structurally ordered BIM fabric	Olistostromes *s.l.*; mud diapirs *s.l.*
c3. Diapiric		Extrusion of non-to poorly consolidated sediments	Tectonic, sedimentary	Mud diapirs and mud volcanoes	Zonation of deformation from core to margins	Olistostromes *s.l.*
7. Sub-aerial deformation **a.** ***Sedimentary***	On the Earth and/or other planets surface	MTP and glacial processes (debris flows, avalanches, slides, etc.)	Sedimentary, glacial, tectonic	MTD (debris flows/avalanches, alluvial fan dep., talus breccias/megabreccias, block falls, glacial till, etc.)	Chaotic BIM fabric (block of intra-basinal origin)	
b. ***Impact of bodies***		Impact of bodies on planet surfaces	Impact processes	Ejected breccias and megabreccias	Chaotic breccias in a fluidal matrix	

BIM – Block-in-matrix; BrFm – Broken Formation; MTD – Mass-transport deposits; MTP – Mass-transport processes; SSD – Soft sediment deformation; TSD – Tectonic stratal disruption

Table 2 – Festa et al.

Figure 1
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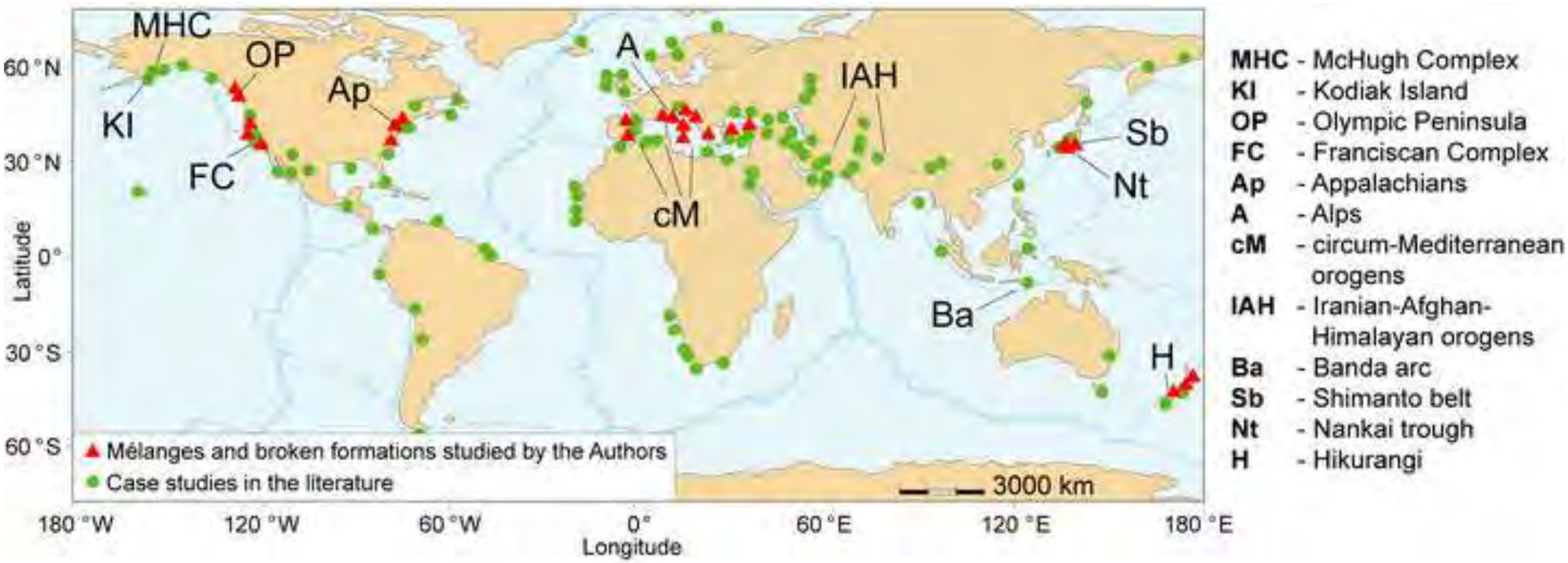

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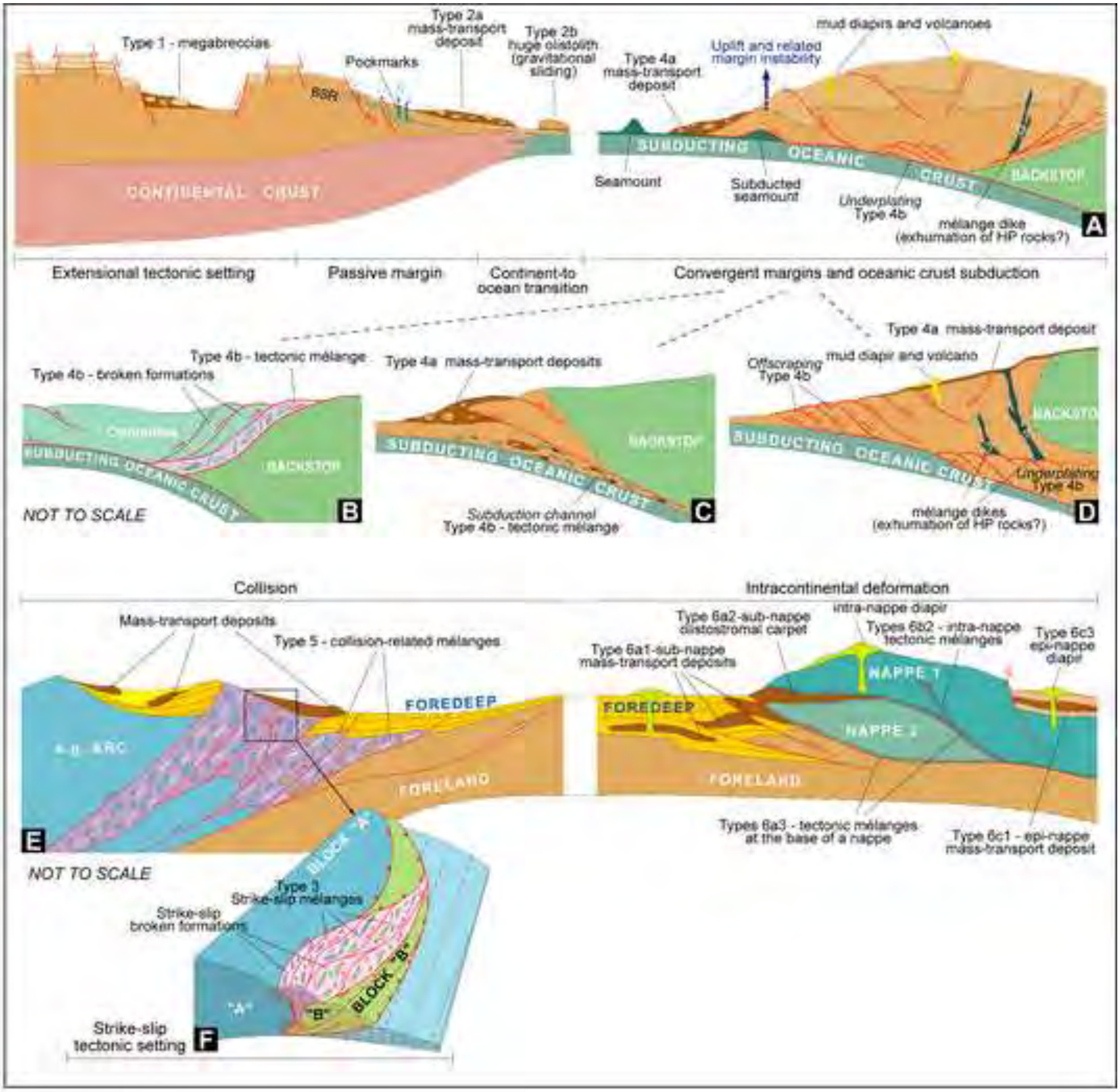

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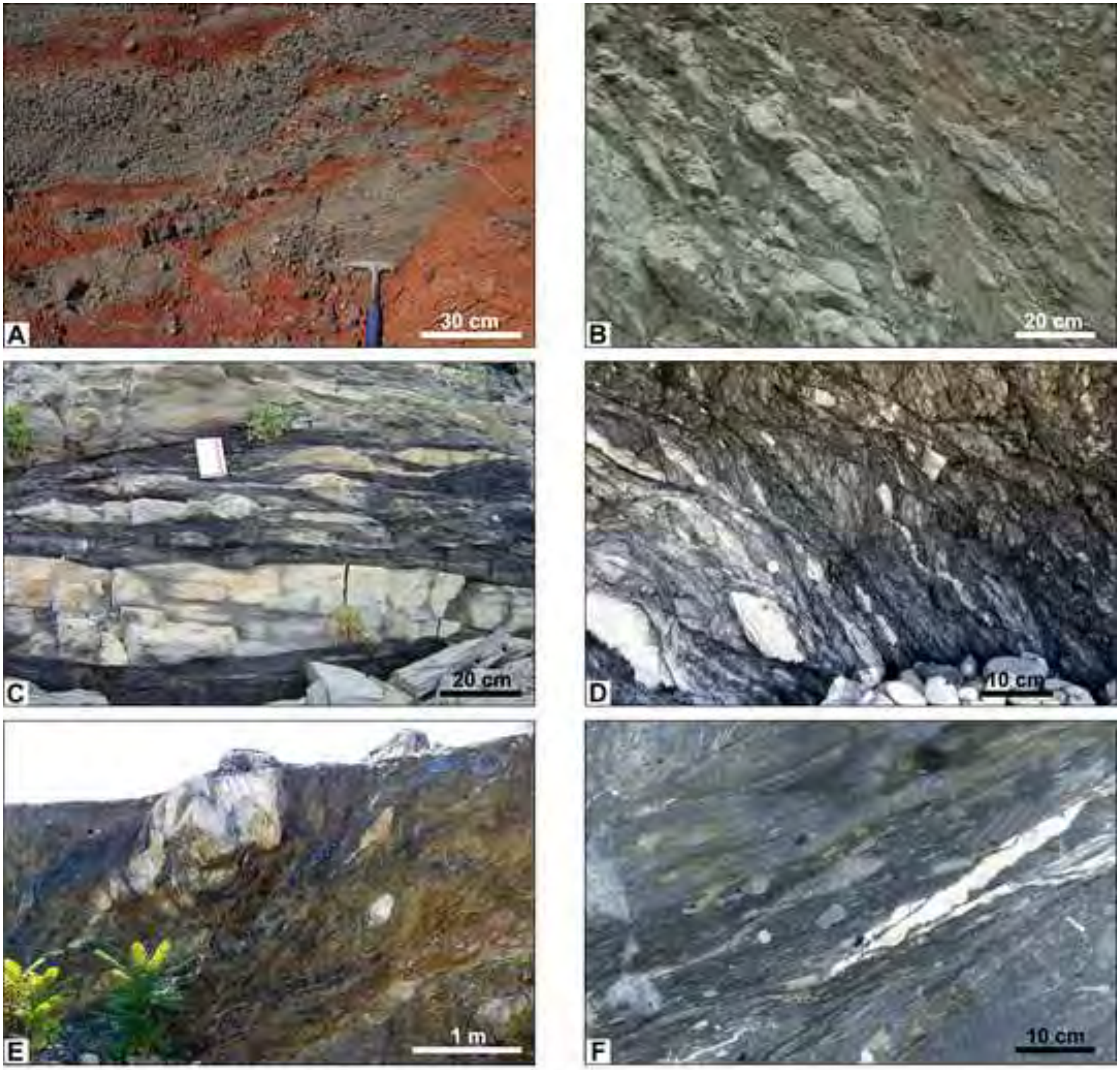

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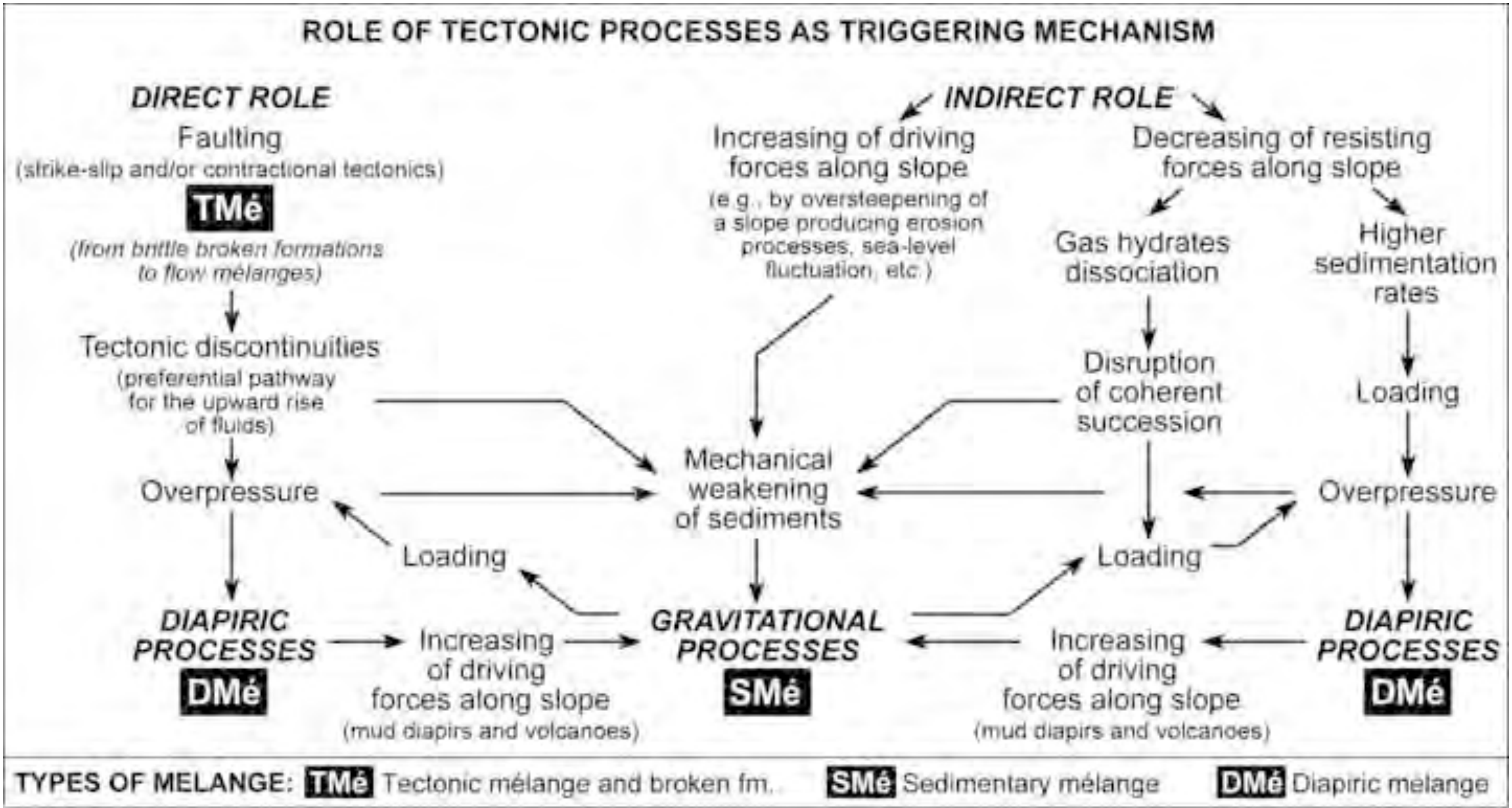

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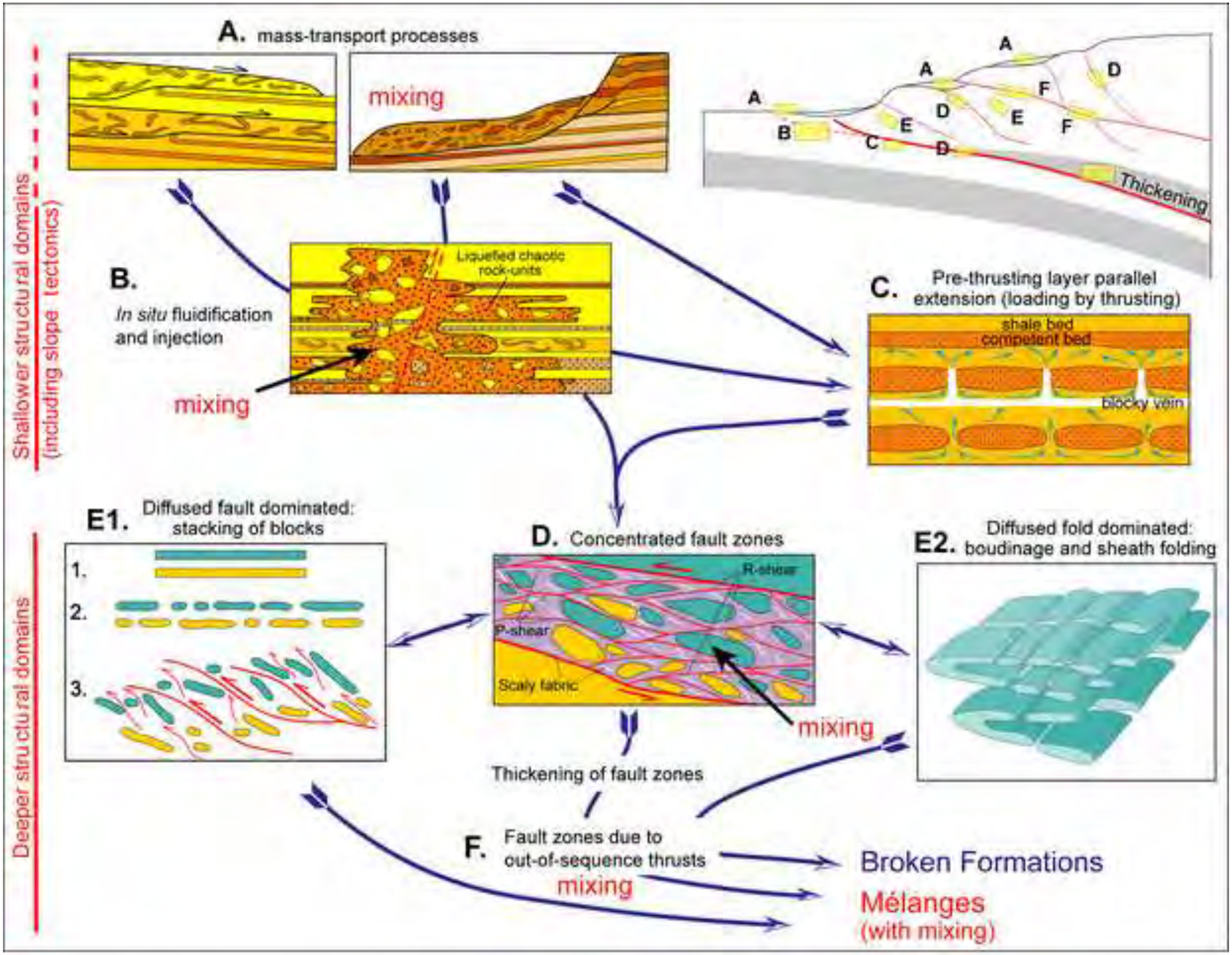

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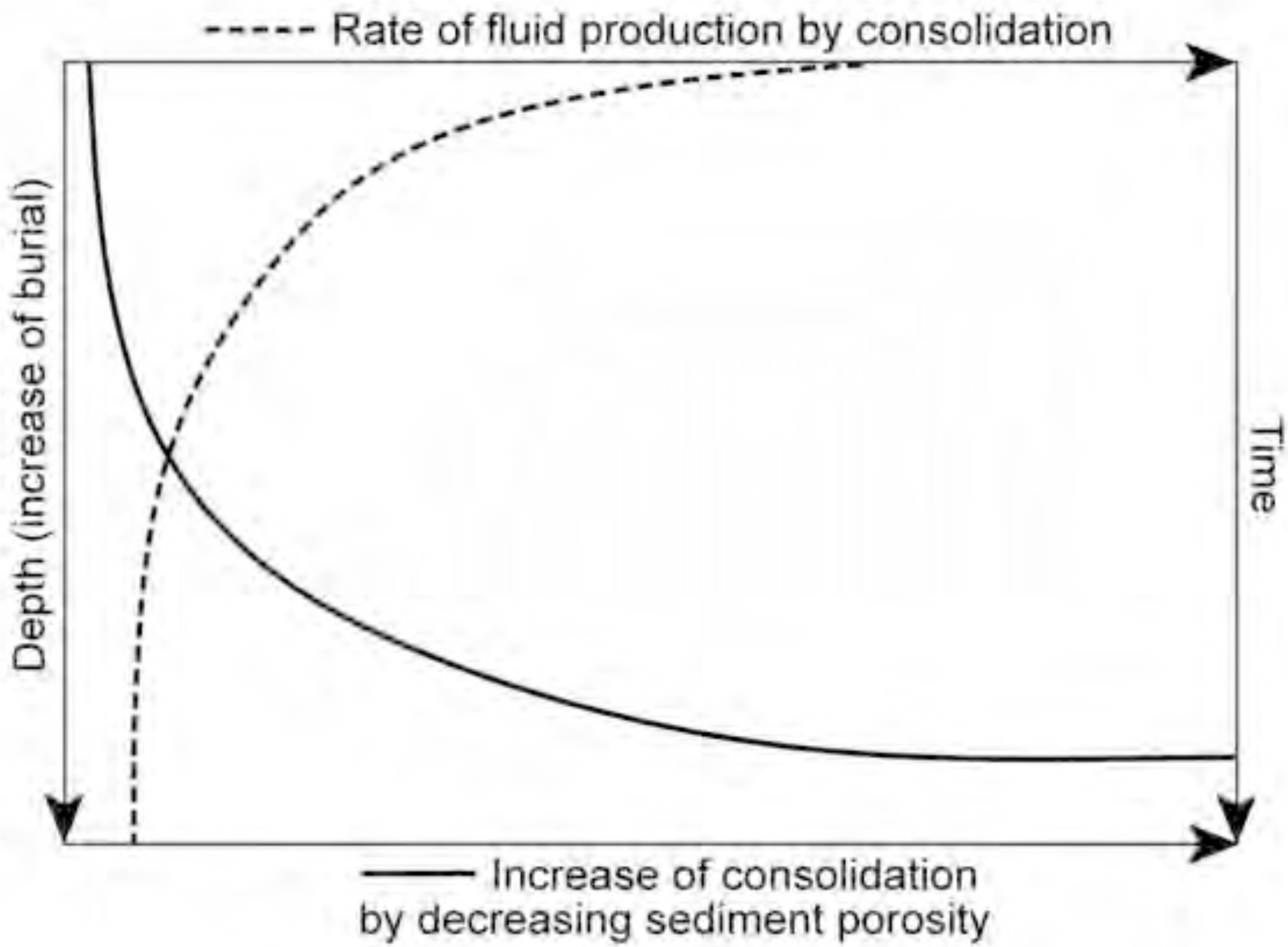

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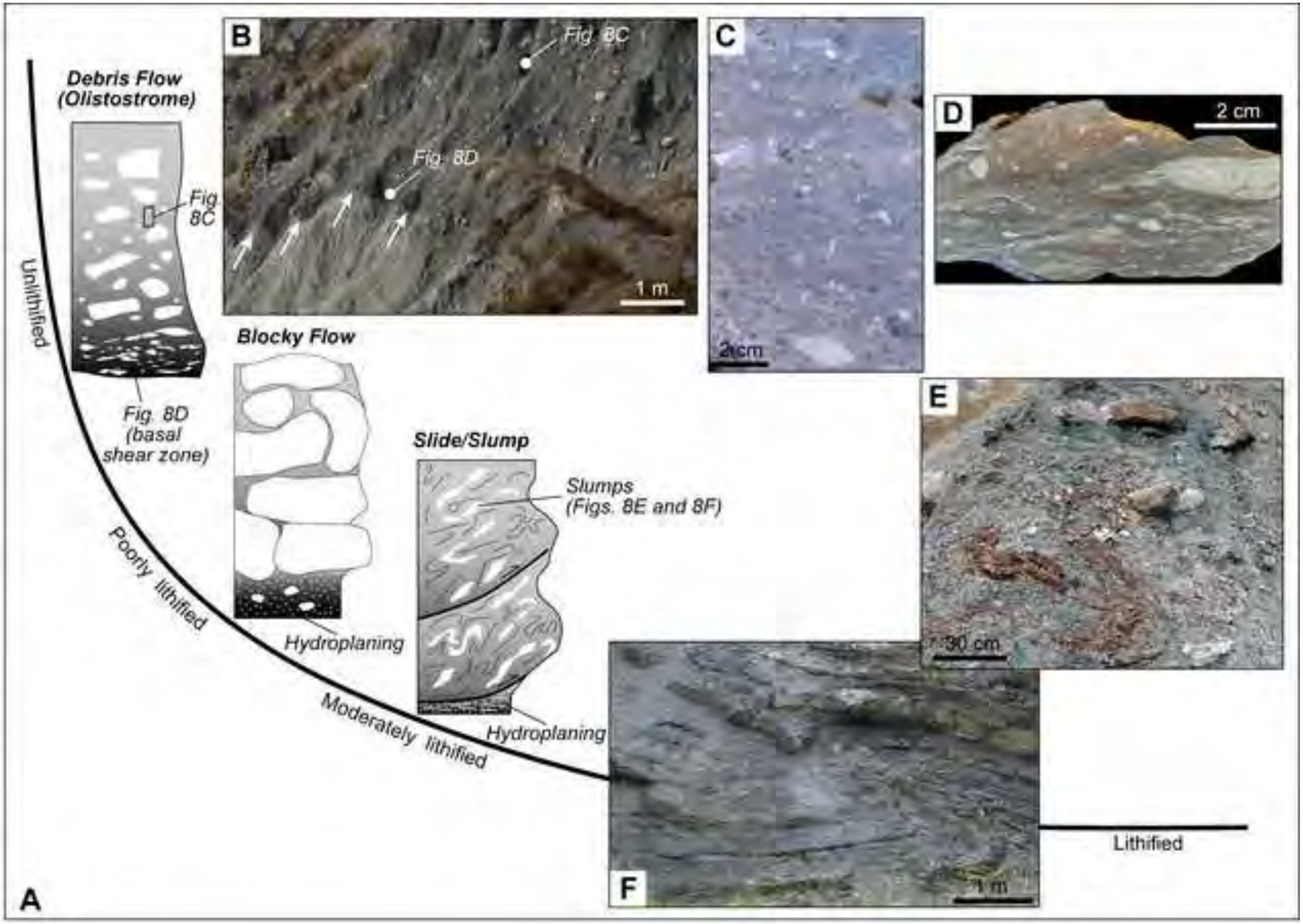

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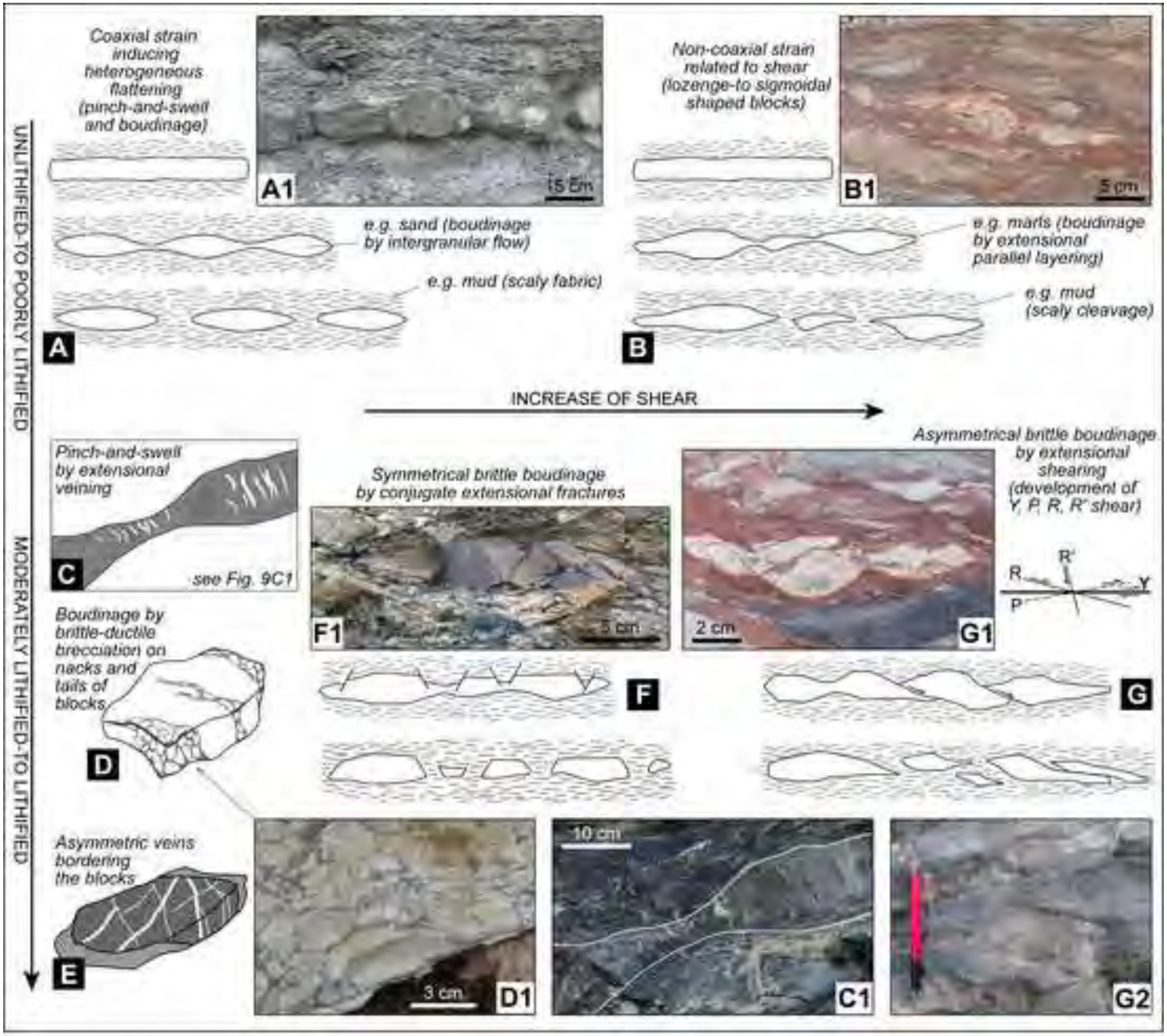

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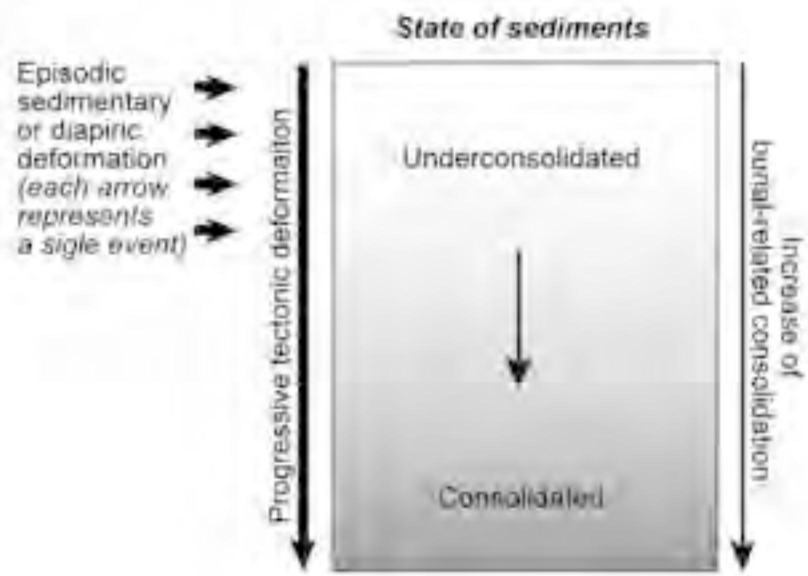

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