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Review article

Contourites and associated sediments controlled by deep-water circulation processes: State-of-the-art and future considerations

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ABSTRACT

The contourite paradigm was conceived a few decades ago, yet there remains a need to establish a sound connection between contourite deposits, basin evolution and oceanographic processes. Significant recent advances have been enabled by various factors, including the establishment of two IGCP projects and the realisation of several IODP expeditions. Contourites were first described in the Northern and Southern Atlantic Ocean, and since then, have been discovered in every major ocean basin and even in lakes. The 120 major contourite areas presently known are associated to myriad oceanographic processes in surface, intermediate and deep-water masses. The increasing recognition of these deposits is influencing palaeoclimatology & palaeoceanography, slope-stability/geological hazard assessment, and hydrocarbon exploration. Nevertheless, there is a pressing need for a better understanding of the sedimentological and oceanographic processes governing contourites, which involve dense bottom currents, tides, eddies, deep-sea storms, internal waves and tsunamis. Furthermore, in light of the latest knowledge on oceanographic processes and other governing factors (e.g. sediment supply and sea-level), existing facies models must now be revised. Persistent oceanographic processes significantly affect the seafloor, resulting in large-scale depositional and erosional features. Various classifications have been proposed to subdivide a continuous spectrum of partly overlapping features. Although much progress has been made in the large-scale, geophysically based recognition of these deposits, there remains a lack of unambiguous and commonly accepted diagnostic criteria for deciphering the small-scaled contourite facies and for distinguishing them from turbidite ones. Similarly, the study of sandy deposits generated or affected by bottom currents, which is still in its infancy, offers great research potential: these deposits might prove invaluable as future reservoir targets. Expectations for the forthcoming analysis of data from the IODP Expedition 339 are high, as this work promises to tackle much of the aforementioned lack of knowledge. In the near future, geologists, oceanographers and benthic biologists will have to work in concert to achieve synergy in contourite research to demonstrate the importance of bottom currents in continental margin sedimentation and evolution.

1. Introduction

The influence of bottom-water circulation in deep-sea sedimentation (i.e. contourites) remains poorly understood. To address this issue, researchers must establish a sound connection between contourite deposits, basin evolution and oceanographic processes. The research on contourites is presently maturing (Rebesco and Camerlenghi, 2008). In fact, contourites were first identified nearly 50 years ago, on the basis of deep-sea bottom photographs of current ripples (Heezen and Hollister, 1964; Hollister, 1967). Over the past few years, this research topic has crystallised, as reflected by number of the publications and international programmes dealing with it (see Section 2). However, many uncertainties remain, such as lack of indisputable diagnostic criteria for identifying contourites. This field is now advancing similarly to turbidite research, which is now mature, progressed in the 1960s. Indeed, there is a glaring disparity in knowledge between the former and the latter: a recent (February 2014) online search for the term *contourites* yielded 256 results on Scopus and 17,300 on Google, whereas a similar search for *turbidites* gave 3841 and 295,000 results, respectively—more than 15 times more in each case.

We are aware of the fact that *contourites* (sediments affected by alongslope bottom currents), *turbidites* (sediments deposited by

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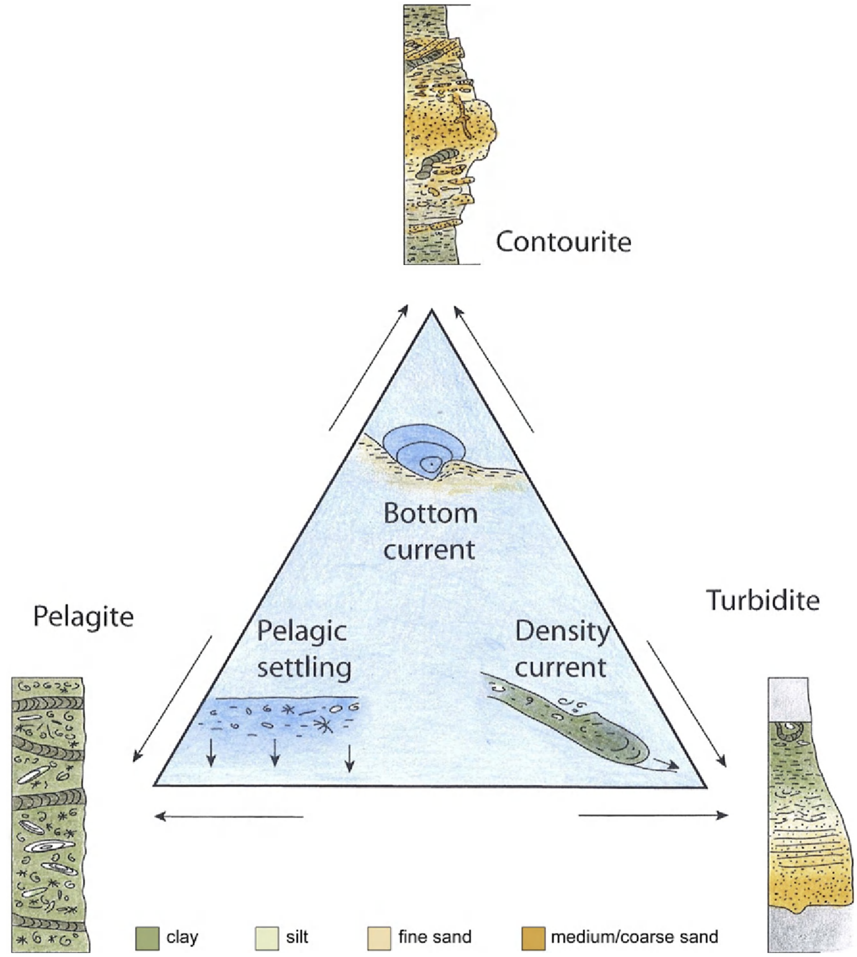


Fig. 1. Conceptual diagram showing the three main types of sedimentary processes operating in the deep sea (within the triangle) and the facies model of the respective depositional products.

downslope density currents) and *pelagites* (sediments deposited by vertical pelagic settling) simply represent extremes in a continuum of deep-sea sedimentary facies (Fig. 1). In fact, putting aside mass-transport deposits, the three main sedimentary processes that occur in the deep sea comprise *the settling of pelagic particles* through the water column, the *predominantly alongslope flow of bottom currents* (relatively clean bottom-water masses), and *downslope density currents* (turbid flows of predominantly terrigenous sediments). The first of these represents a background process that becomes dominant only in very remote abyssal areas. In contrast, episodic, high-energy density flows are commonly superimposed over or interact with relatively permanent flows of bottom currents on many continental margins. Whilst the application of turbidite models and concepts to the interpretation of deep-sea facies remains challenging (e.g. the dominance of the turbidite paradigm; criticised by Shanmugam, 2000), there is increasing recognition that contour currents are important transport and sedimentary phenomena that control much of deep-sea sedimentation.

The study of contourites is now considered crucial for at least three fields of fundamental and applied research (Rebesco, 2005, 2014): *palaeoclimatology & palaeoceanography*; *slope-stability/geological hazard assessment*; and *hydrocarbon exploration* (see Section 3). However, despite the significance of contourites, the fact that they cover large parts of present ocean floors and all of the Earth's continental margins (see Section 4), and the fact that an increasing number of fossil occurrences are being documented (Huneke and Stow, 2008), they remain poorly known amongst non-specialists. The widespread distribution of contourites is connected to the pervasiveness of the transport vehicles that control their deposition: bottom currents and associated oceanographic processes (Fig. 2). In fact, water masses move throughout the ocean basins and, as a general simplification, any "persistent" water current near the seafloor can be considered to be a "bottom current" (see Section 5).

Contourites are defined as "sediments deposited or substantially reworked by the persistent action of bottom currents" (e.g. Stow et al., 2002a; Rebesco, 2005, 2014). The term *contourites*, originally introduced to specifically define the sediments deposited in the deep sea by contour-parallel thermohaline currents, has subsequently been widened to embrace a larger spectrum of sediments that are affected to various extents by different types of current. Moreover, contourites may occur interbedded with other sediment types and interaction of processes is the norm rather than the exception.

On the basis of these considerations, we suggest to use the well-established term *contourite* as a generic such as *mass-wasting deposits* or *gravity-flow deposits*. Such family names include several kinds of sediment that have more specific names (e.g. *turbidites*, *debris-flow deposits*, or *mudflow deposits*). In the case of the deposits affected by bottom currents, such specific names have not been formalised to the same extent (apart from sensu stricto *contourites*, which are produced by thermohaline-induced geostrophic bottom currents). Additionally, numerous associated processes are related to the circulation of deep-water masses and bottom currents (see Section 5), such as benthic storms; overflows; interfaces between water masses; vertical eddies; horizontal vortices; tides and internal tides; internal waves and solitons; tsunami-related traction currents; and rogue or cyclonic waves. Some of these processes are not well known and/or their consequences have not been researched, although their associated energy

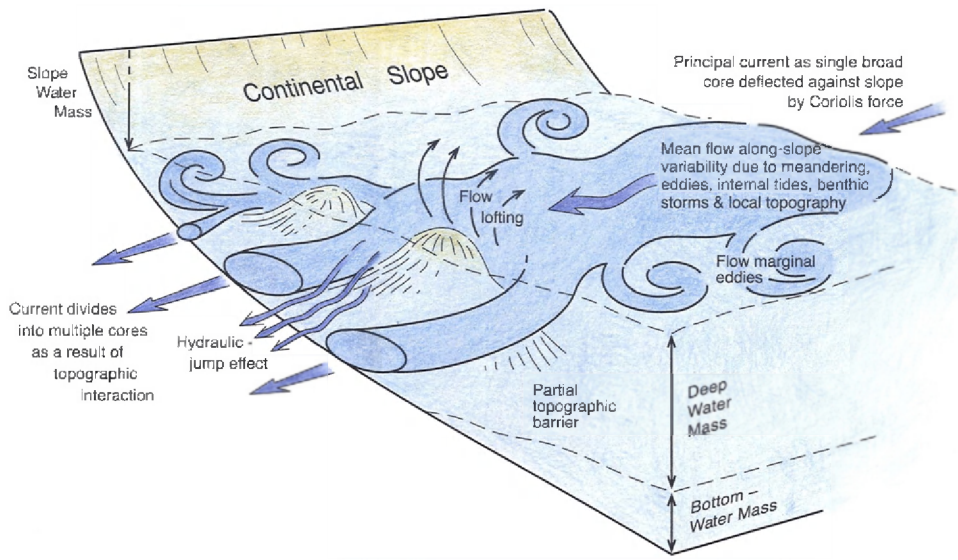


Fig. 2. Schematic diagram summarising the principal bottom-current features.
Modified from work by Stow et al. (2008); with permission from Elsevier.

and influence in shaping the seafloor are very important (e.g. as is known for internal waves). Thus further research aimed at distinguishing between the various types of transport and depositional processes is required and must incorporate the contributions of oceanographers and geologists.

Bottom currents are capable of building thick and extensive accumulations of sediments. Although these sediment bodies have received various names in the past, the term *contourite drifts* should be preferred. Similarly to channel-levee systems generated by turbidity currents, such large bodies normally have a noticeable mounded geometry,

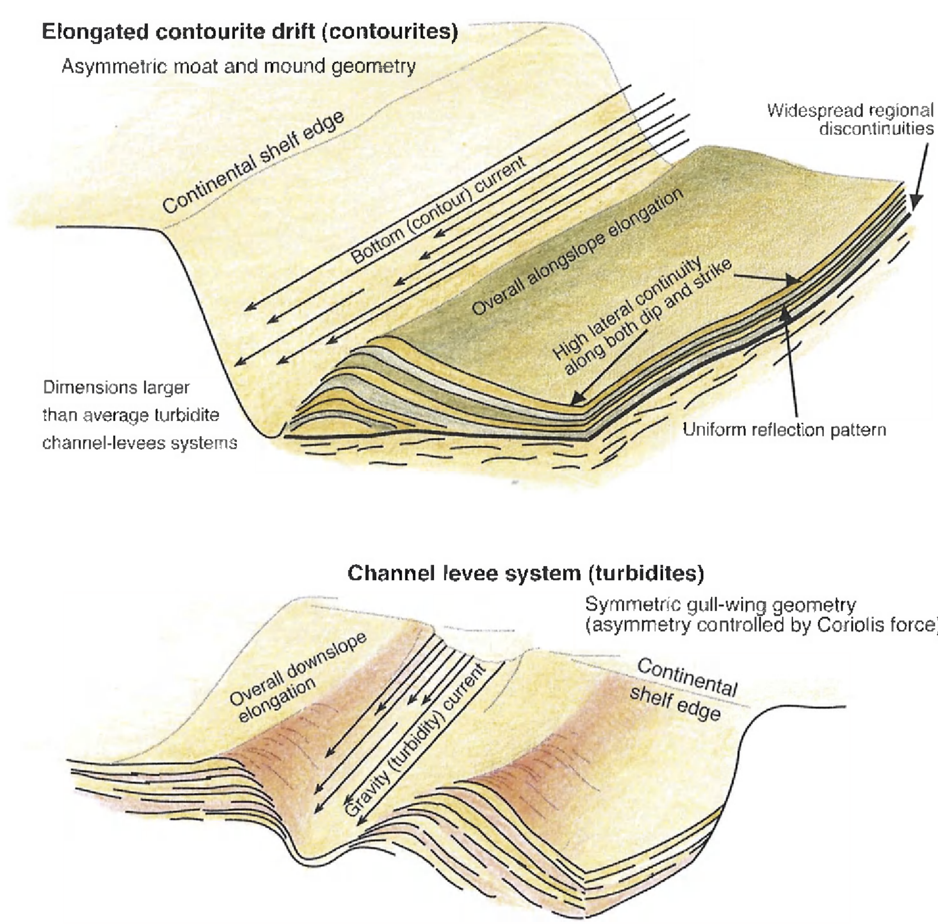
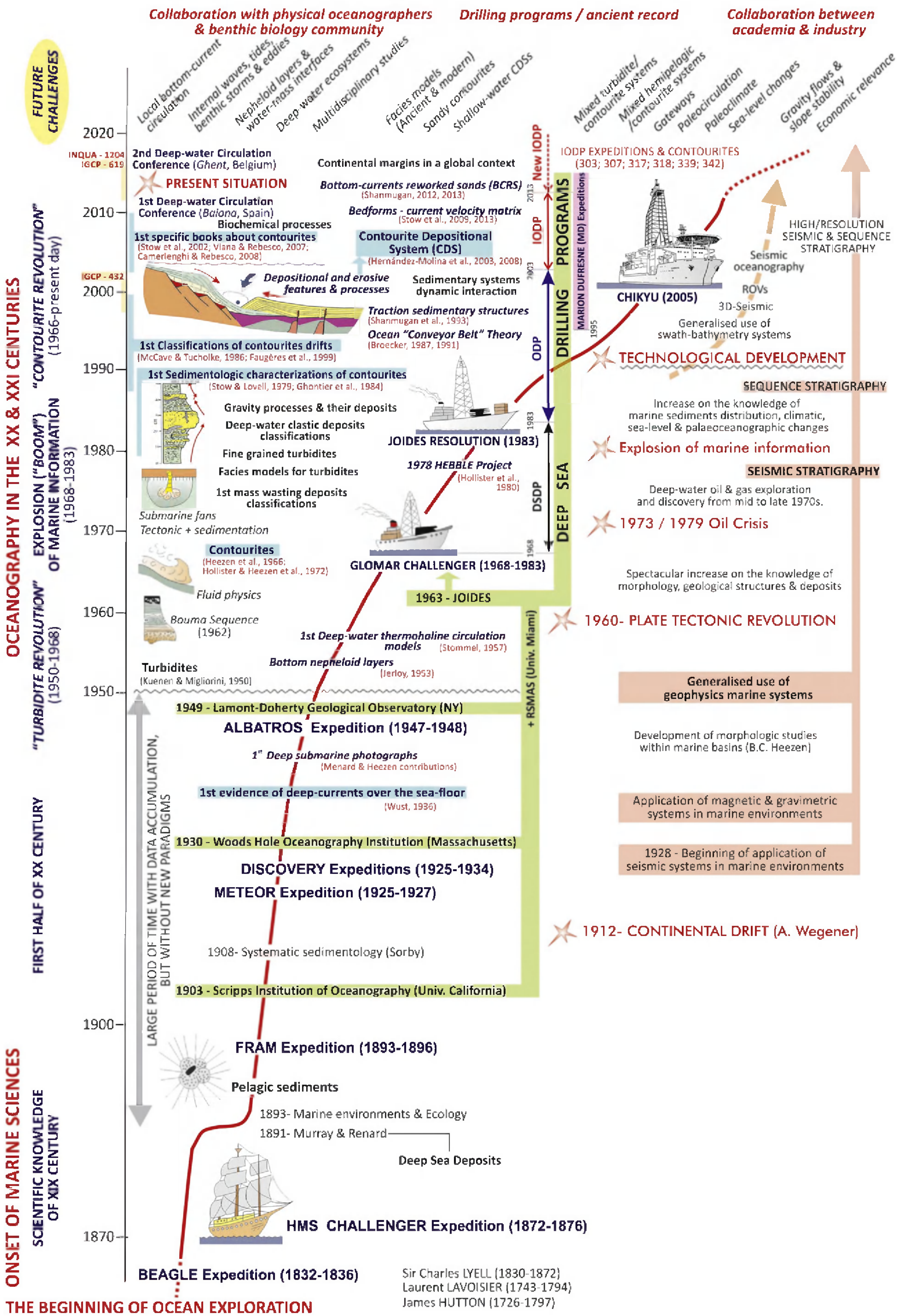


Fig. 3. Schematic model showing ideal, large-scale differences between contourite drifts and channel-levee systems.
Modified from work by Rebesco (2005); with permission from Elsevier.


Collaboration with physical oceanographers & benthic biology community
Drilling programs / ancient record
Collaboration between academia & industry
FUTURE CHALLENGES
Local bottom-current circulation
Internal waves, tides, benthic storms & eddies
Nepheloid layers & water-mass interfaces
Deep-water ecosystems
Multidisciplinary studies
Facies models (Ancient & modern)
Sandy contourites
Shallow-water CDSs
Mixed turbidite/contourite systems
Mixed hemipelagic /contourite systems
Gateways
Paleocirculation
Paleoclimate
Sea-level changes
Gravity flows & slope stability
Economic relevance
INQUA - 1204
IGCP - 619
2nd Deep-water Circulation Conference (Ghent, Belgium)
Continental margins in a global context
PRESENT SITUATION
Bottom-currents reworked sands (BCRS) (Shanmugan, 2012, 2013)
1st Deep-water Circulation Conference (Baiona, Spain)
Bedforms - current velocity matrix (Stow et al., 2009, 2013)
Biochemical processes
1st specific books about contourites (Stow et al., 2002; Viana & Rebesco, 2007; Camerlenghi & Rebesco, 2008)
Contourite Depositional System (CDS) (Hernández-Molina et al., 2003, 2008)
IGCP - 432
Depositional and erosive features & processes
Sedimentary systems dynamic interaction
Traction sedimentary structures (Shanmugan et al., 1993)
Ocean "Conveyor Belt" Theory (Broecker, 1987, 1991)
1st Classifications of contourites drifts (McCave & Tucholke, 1986; Faugères et al., 1999)
1st Sedimentologic characterizations of contourites (Stow & Lovell, 1979; Ghontier et al., 1984)
Gravity processes & their deposits
Deep-water clastic deposits classifications
Fine grained turbidites
Facies models for turbidites
1st mass wasting deposits classifications
Submarine fans
Tectonic + sedimentation
Contourites (Heezen et al., 1966; Hollister & Heezen et al., 1972)
Fluid physics
Bouma Sequence (1962)
Turbidites (Kuenen & Migliorini, 1950)
1st Deep-water thermohaline circulation models (Stommel, 1957)
Bottom nepheloid layers (Jerlov, 1953)
IODP EXPEDITIONS & CONTOURITES (303; 307; 317; 318; 339; 342)
New IODP
IODP
ODP
DSDP
DRILLING PROGRAMS
DEEP SEA
MARION DUFRESNE (MD) Expeditions
CHIKYU (2005)
JOIDES RESOLUTION (1983)
1978 HEBBLE Project (Hollister et al., 1980)
GLOMAR CHALLENGER (1968-1983)
1963 - JOIDES
+ RSMAS (Univ. Miami)
HIGH/RESOLUTION SEISMIC & SEQUENCE STRATIGRAPHY
Seismic oceanography
ROVs
3D-Seismic
Generalised use of swath-bathymetry systems
TECHNOLOGICAL DEVELOPMENT
SEQUENCE STRATIGRAPHY
Increase on the knowledge of marine sediments distribution, climatic, sea-level & palaeoceanographic changes
Explosion of marine information
SEISMIC STRATIGRAPHY
Deep-water oil & gas exploration and discovery from mid to late 1970s.
1973 / 1979 Oil Crisis
Spectacular increase on the knowledge of morphology, geological structures & deposits
1960- PLATE TECTONIC REVOLUTION
Generalised use of geophysics marine systems
Development of morphologic studies within marine basins (B.C. Heezen)
Application of magnetic & gravimetric systems in marine environments
1928 - Beginning of application of seismic systems in marine environments
1912- CONTINENTAL DRIFT (A. Wegener)
1949 - Lamont-Doherty Geological Observatory (NY)
ALBATROS Expedition (1947-1948)
1st Deep submarine photographs (Menard & Heezen contributions)
1st evidence of deep-currents over the sea-floor (Wust, 1936)
1930 - Woods Hole Oceanography Institution (Massachusetts)
DISCOVERY Expeditions (1925-1934)
METEOR Expedition (1925-1927)
1908- Systematic sedimentology (Sorby)
1903 - Scripps Institution of Oceanography (Univ. California)
LARGE PERIOD OF TIME WITH DATA ACCUMULATION, BUT WITHOUT NEW PARADIGMS
FRAM Expedition (1893-1896)
Pelagic sediments
1893- Marine environments & Ecology
1891- Murray & Renard
Deep Sea Deposits
HMS CHALLENGER Expedition (1872-1876)
BEAGLE Expedition (1832-1836)
Sir Charles LYELL (1830-1872)
Laurent LAVOISIER (1743-1794)
James HUTTON (1726-1797)
THE BEGINNING OF OCEAN EXPLORATION
OCEANOGRAPHY IN THE XX & XXI CENTURIES
"CONTOURITE REVOLUTION" (1966-present day)
EXPLOSION ("BOOM") OF MARINE INFORMATION (1968-1983)
"TURBIDITE REVOLUTION" (1950-1968)
FIRST HALF OF XX CENTURY
ONSET OF MARINE SCIENCES
SCIENTIFIC KNOWLEDGE OF XIX CENTURY
2020
2010
2000
1990
1980
1970
1960
1950
1900
1870

which is generally elongated parallel, or lightly oblique to the margin (Fig. 3). Bottom currents and associated processes also generate various other depositional and erosional or non-depositional structures at different scales (see Section 6).

A facies model for contourites (see Section 7) has been proposed for some time, mainly for fine-grained contourites (Gonthier et al., 1984; Stow and Faugères, 2008). However, unambiguous and commonly accepted diagnostic criteria for contourites are still lacking. According to most contourite researchers and based on countless core samples (see, for example: Stow and Faugères, 2008), extensive bioturbation is generally the dominant feature. However, a minority of researchers (see, for example: Huneke and Stow, 2008; Martín-Chivelet et al., 2008; Shanmugam, 2008, 2012a; Mutti and Carminatti, 2012) interpret some distinctly laminated sandy deposits as contourites. They observe that traction structures are abundantly produced by bottom currents on modern ocean floors and wonder whether the general absence of these structures in sediment core studies might be explained by a bias imposed by the limited scale of observation. Furthermore, these authors highlight that extensive bioturbation is also abundant in areas unaffected by bottom currents and in turbidites. This controversy might be down to different researchers having worked in different settings (e.g. muddy contourites vs. sandy contourites or reworked turbidites). Interestingly, there is still some debate over the facies model and over the diagnostic sedimentological criteria for contourites. The forthcoming results from the recent IODP Expedition 339, which specifically targeted contourite deposits (for a preliminary preview, see Section 8), might provide crucial clues to these questions. Further research is needed to define a universally acceptable set of diagnostic criteria for contourites. Such future research (see Section 9) should be aimed mainly at elucidating the processes involved in contourite formation.

2. Brief history

Since the seminal paper of Heezen and Hollister (1964) was published in the first issue of this journal, the contourite paradigm has progressed gradually, although much more slowly (Fig. 4) than has the well-funded area of turbidite research (Rebesco and Camerlenghi, 2008). However, a comparison of the initial ideas on contourites with their presently accepted definition and recorded occurrences reveals a marked shift in researchers' mind-sets: the definition of contourites has shifted from meaning highly localised deposits created by thermohaline circulation in a deep-sea basin (Heezen et al., 1966), to meaning multi-facetted deposits (Lovell and Stow, 1981; Faugères et al., 1999) originating from many possible physical drivers and several types of currents (Rebesco and Camerlenghi, 2008) in the deep ocean (Campbell and Deptuck, 2012; Uenzelmann-Neben and Gohl, 2012), on continental slopes (Preu et al., 2012; Roque et al., 2012; Li et al., 2013), in shallow margins (Verdicchio and Trincardi, 2008; Vandorpe et al., 2011) and in lakes (Ceramicola et al., 2001; Gilli et al., 2005; Heirman et al., 2012).

Although an initial phase of steady research growth witnessed a few milestone papers (Stow, 1982; Stow and Holbrook, 1984; Stow and Piper, 1984; Faugères and Stow, 1993), it was not until 2002 that the first dedicated book on contourites was published (Stow et al., 2002a), as the major outcome of IGCP 432 on Contourites, Bottom Currents and Palaeocirculation (1998 to 2002). This international geoscience correlation programme firmly confirmed the contemporary contourite paradigm and was successful in creating a "contourite community", some of whose members published major books (see Viana and Rebesco (2007) and Rebesco and Camerlenghi (2008)) and brought contourites into the *Encyclopaedia of Geology* (Rebesco, 2005). Thus, the success of IGCP 432 led to improved documentation and classification of the plethora of contourite deposits, with a focus on the link between processes and products.

Since IGCP 432, contourite research has benefitted from a boost in technological and methodological advances in geophysics, palaeoceanography and physical oceanography (Fig. 4) that have enabled numerous discoveries of contourite drifts and led to increasingly detailed studies on possible driving forces behind contourite formation. These advances also yielded better insight into the lateral and temporal variability and connectivity of contourite processes, which in turn led to the definition of the terms *Contourite Depositional Systems* (*CDS*) and *Contourite Depositional Complexes* (Hernández-Molina et al., 2004, 2008a; Rebesco and Camerlenghi, 2008). Much of this insight has emerged from the study of contourite deposits influenced by the Mediterranean Outflow Water (MOW), which has served as a natural laboratory for more than 20 years (Gonthier et al., 1984; Nelson et al., 1993; Llave et al., 2001; Mulder et al., 2003; Hernandez-Molina et al., 2006a; Marchès et al., 2007; Llave et al., 2011; Roque et al., 2012; Brackenridge et al., 2013 amongst others). Moreover, this case not only applies for the Gulf of Cádiz, but also applies further away, in the Bay of Biscay (Ercilla et al., 2008; Van Rooij et al., 2010; Hernandez-Molina et al., 2011a) and in the Porcupine Seabight (Van Rooij et al., 2003, 2009). From 2000 onwards, significant research efforts were made in relation to the MOW and contourites adjacent to the Iberian Peninsula (Hernandez-Molina et al., 2011a). Major contributors included the Spanish projects CONTOURIBER (2009 to 2012) and MOWER (2013 to 2015), and the European project EC FP5 HERMES (2005 to 2009), which enabled better comprehension of the delicate interplay between bottom currents and deep-water ecosystems, such as the association between the giant cold-water coral mounds and the Porcupine CDS, off Ireland (Van Rooij et al., 2007a,b; Huvenne et al., 2009). In parallel, other continental margins have been studied by researchers from various countries, seeking to ascertain the Base of Slope (BOS) for determining the outer limits of the juridical Continental Shelf, per the recommendations of the United Nations Convention on the Law of the Sea (UNCLOS; ABLOS, 2006). These studies have revealed the abundance of contourites and associated sediments in deep-water settings and illustrated the importance of these features in controlling the continental slope and rise morphology (see, for example: Hernández-Molina et al., 2009, 2010).

Momentum following IGCP 432 led to the first International Conference on Deep-water Circulation: Processes and Products, held in Baiona, Spain (16 to 18 June 2010), resulting in a special issue of *Geo-Marine Letters* (Hernandez-Molina et al., 2011b) and laying the groundwork for the second edition, to be held in, Ghent, Belgium (September 2014). It also led to proposal of a new IGCP, which began in 2012: IGCP 619 "Contourites: processes and products". Some authors (Shanmugam, 2006; Mutti and Carminatti, 2012; Shanmugam, 2012a; Gong et al., 2013; Shanmugam, 2013a,b) have proposed new concepts about Bottom-Current Reworked Sands (BCRS), thus providing new perspectives on deep-water research in terms of ancient records as well as recent and present marine basins.

The primary aim in this new phase of improved comprehension is to better correlate contourite processes to their products, through greater cooperation amongst researchers from diverse disciplines. However, the ultimate success of the Cádiz CDS natural laboratory is still to come, through the exploitation of the data from IODP Expedition 339, which acquired more than 5 km of core sample (Expedition 339 Scientists, 2012; Hernández-Molina et al., 2013; Stow et al., 2013a). Moreover, over the past decade, several IODP expeditions were dedicated to drilling sediment drifts, focusing predominantly on the palaeoceanographic potential of these drifts at strategic sites in areas

Fig. 4. Sketch showing the recent evolution of knowledge on contourite processes and products in the very general context of oceanography and marine geology development, including the potential future challenges. Only those four American institutions that joined together to develop the JOIDES programme and the DSDP are shown; however, many other institutions from around the world have contributed to this field of research.
Modified from work by Hernandez-Molina et al. (2011b); with permission from Springer.

including the North Atlantic Ocean (*IODP 303*: Eirik and Gardar Drift; *IODP 307*: Porcupine CDS; and *IODP 342*: Newfoundland Drifts) and the Pacific and Southern oceans (*IODP 317*: Canterbury Drift; and *IODP 318*: Wilkes Land Drift).

3. Implications of contourites for palaeoclimate, slope stability, and hydrocarbon exploration

For many years the research on contourites was the realm of a few specialists. However, it has recently been garnering interest amongst more and more scientists, in parallel to increasing awareness about the effects of bottom currents and associated oceanographic processes (see Section 5). Such processes occur almost everywhere in the oceans and are crucial for shaping the seafloor. Many scientists, even those who are not specialised in contourites, must deal with sediments affected by bottom currents in their own field of research. Contourites are paramount in three areas: *palaeoclimatology* & *palaeoceanography*; *slope stability/geological hazard assessment*; and *hydrocarbon exploration*.

3.1. Palaeoclimatology & palaeoceanography

Contouritic sediments are usually fairly continuous and yield temporal records of relatively high resolution, as the accumulation rates in contourite drifts are higher than those in adjacent, condensed pelagic sequences. Therefore, these sediments yield valuable information on the variability of ocean circulation patterns, current velocities, oceanographic history and basin interconnectivity. In polar areas these sediments provide information on ice-sheet stability and sea-ice coverage (Camerlenghi et al., 1997a; Rebesco et al., 1998; Sagnotti et al., 2001; Lucchi et al., 2002a,b; Grützner et al., 2003; Rebesco, 2003; Villa et al., 2003; Grützner et al., 2005; Amblas et al., 2006). Seeking a better understanding of how the ocean affects the Earth's climate, researchers have been extracting detailed palaeoclimate information from sediments formed by persistent oceanic currents.

The history of ocean circulation and climate can be extracted from contourite deposits, using discrete sampling analyses (with geochemical, faunal, sedimentological techniques), continuous geophysical–chemical logging and seismic imaging. The latter enables visualisation of drift geometry, internal reflections configuration and seismic facies, thereby providing palaeoceanographic information on palaeo-current pathways and on changes in current energy and direction, on timescales from tens of thousands to millions of years. In particular, high-quality 3D seismic data collected by the petroleum industry and made available to the academic community over the past few have enabled detailed morphological reconstructions of contourite drifts (Fig. 5).

Contourite research addresses a broad range of time scales and Earth-system processes that range from millions of years (*tectonic*; e.g. the opening of the oceanic gateways) to tens of years (*human*;

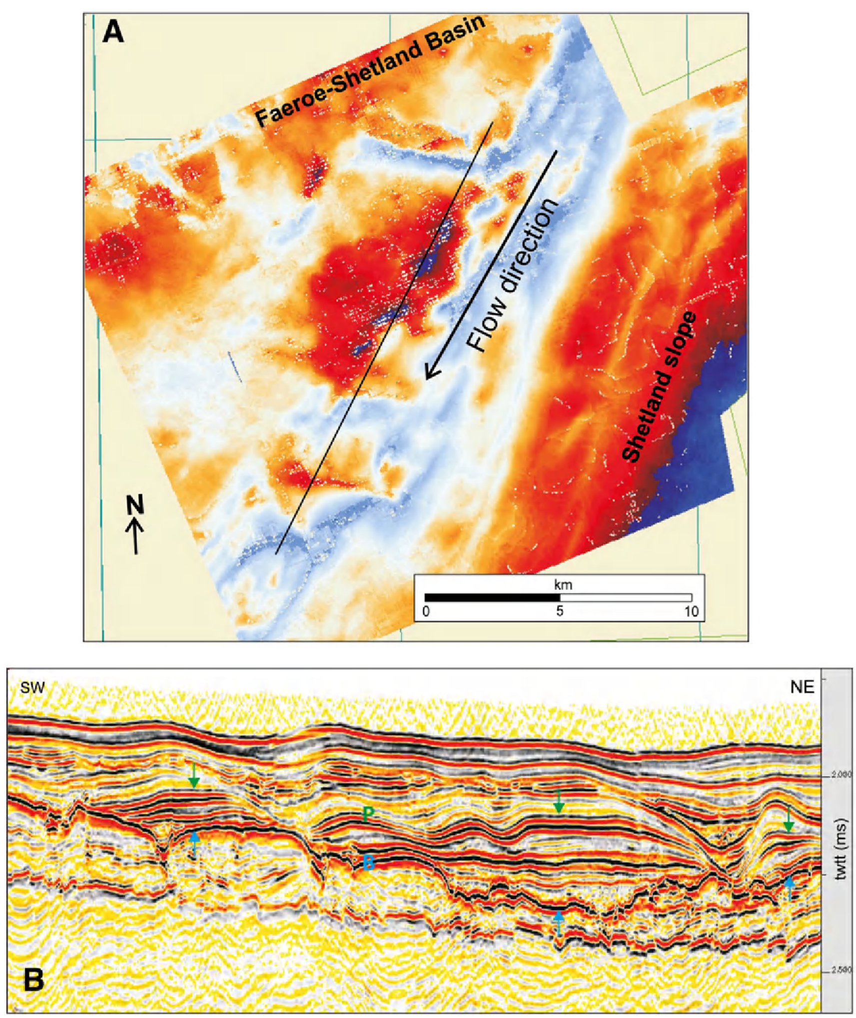


Fig. 5. Reconstruction of palaeocurrent patterns across the West Shetland Drift. (A) 3D seismic data isochrone map between an internal Pliocene reflector and the basal unconformity (P and B, respectively, in the seismic profile below). The dark-blue area corresponds to sediment thicknesses > 150 m, whereas the light blue area represents non-deposition. (B) Seismic profile (the thin line in A) showing along-slope SW progradation of internal reflections and downlap/onlap onto the erosional basal unconformity (Late Miocene to Early Pliocene). Modified from work by Knutz (2008); with permission from Elsevier.

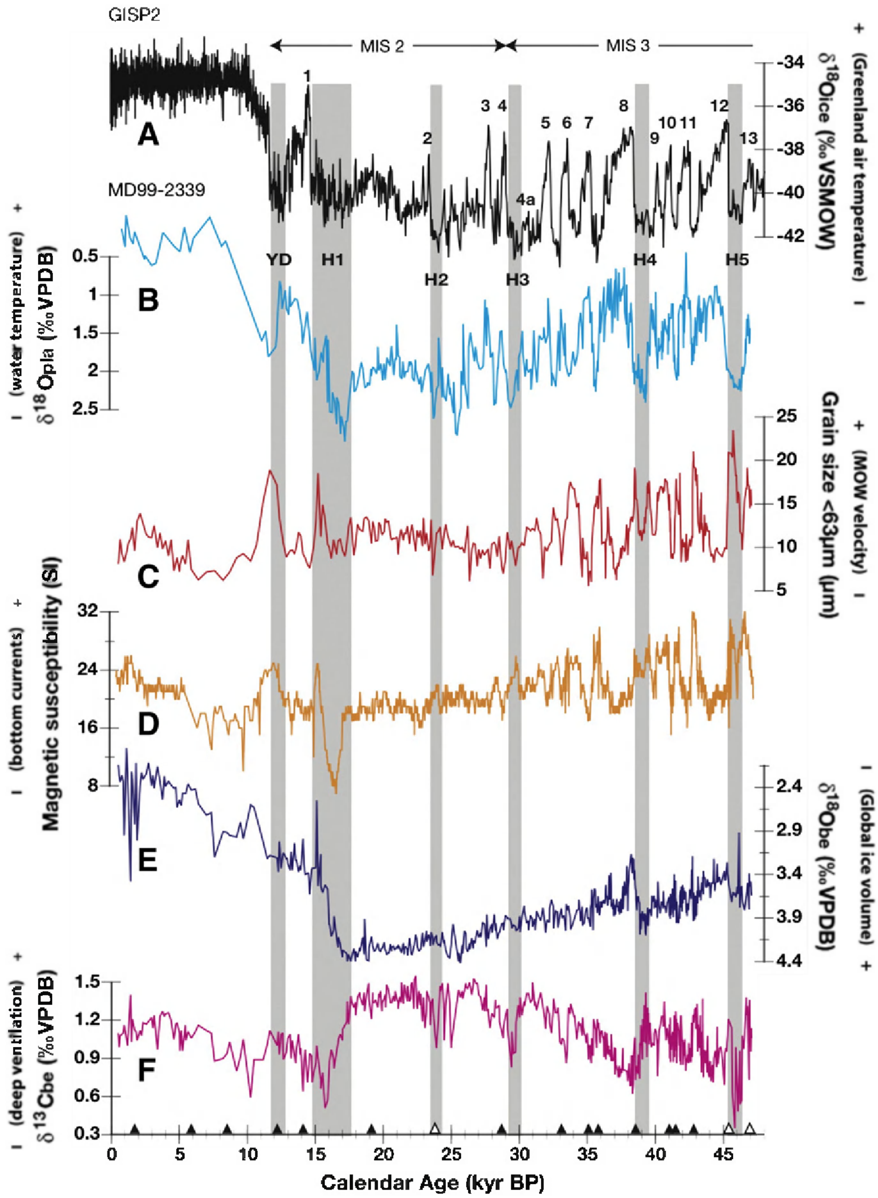


Fig. 6. Comparison of palaeoclimate records from Greenland Ice Sheet Project Two (GISP2) and core MD99-2339 from the Gulf of Cádiz over the last 48 calendar kyr. The MOW variation correlates exactly with Heinrich events (H1 to 5) and D/O cycles during MIS 3. (A) GISP2 δ^{18}O temperature record (δ^{18}Oice; Stuiver and Grootes, 2000); (B) MD99-2339 planktonic δ^{18}O record (δ^{18}Opla), indicative of water temperature. (C) MD99-2339 mean grain size (gs) < 63 μm, which reveal changes in the strength of the MOW; (D) MD99-2339 magnetic susceptibility, which reveals changes in the concentration of ferromagnetic grains and therefore, depict hydrodynamic changes (stronger bottom currents led to increased deposition of lithic and consequently, ferromagnetic grains); (E) MD99-2339 benthonic δ18O record (δ^{18}Obe), which is a proxy of sea-level/Ice volume; (F) MD99-2339 epibenthic δ^{13}C data (δ^{18}Cbe), which is a function of decay of organic matter and enables reconstruction of the deep-water circulation (ventilation). Numbers above the GISP2 δ^{18}O record indicate Dansgaard-Oeschger interstadials 1 to 13, H1 to H5 below Heinrich events 1 to 5 (also marked by grey squares). The triangles at the bottom indicate age-control points based on ^{14}C dates (black) with tuning points to the GISP2 chronology (grey). The arrows on the top show periods of Marine Isotope Stages (MIS) 2 and 3. VPDB: Vienna PeeDee Belemnite; VSMOW: Vienna Standard Mean Ocean Water. Modified from work by Voelker et al. (2006); with permission from Elsevier.

e.g. rapid ocean-climate variability in the North Atlantic) (Knutz, 2008). The reconstruction of leads and lags between various parameters of ocean-climate changes at multi-decadal time scales can be measured in the records from rapidly accumulating muddy contourite deposits. This information, whose resolution approaches that of ice-core archives (e.g. Llave et al., 2006; Voelker et al., 2006; Knutz et al., 2007; Toucanne et al., 2007, Fig. 6), is crucial for elucidating the global teleconnections, feedback thresholds and forcing mechanisms that determined the past climate systems and are dictating the present one.

3.2. Slope-stability/geological hazard assessment

The stability of submarine slopes commonly relates to the distribution, composition and physical properties of contourites (Solheim et al., 2005, Fig. 7), including some of the largest known ones (Bryn et al., 2005). The spatial and temporal variations in sediment erosion, transport and deposition generate sedimentary successions that are prone to becoming gravitationally unstable (Laberg et al., 2005). Fine-grained, low-permeability, high pore-water content contourites favour the formation of over-pressurised gliding planes (Rebesco, 2005).

According to Laberg and Camerlenghi (2008) contouritic sediments tend to fail because of five main factors:

(a) *Geometry and location:* Contourites (as opposed to *sheeted turbidites*) form large sediment mounds on inclined continental slopes that are prone to mass wasting (often large due to broad areal distribution of contourites resulting from ample extent of geostrophic currents, Rebesco et al., 2002, 2007).

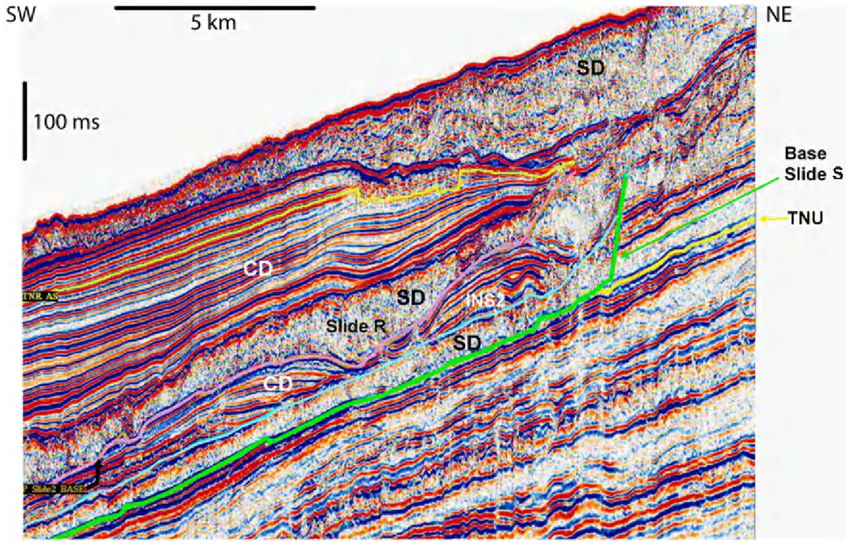


Fig. 7. Seismic profile from Storegga Slide, offshore Norway showing palaeo-slide S, infilling contourite drift and palaeo-slide R. The glide plane of palaeoslide S is indicated by the green horizon, and that of palaeo-slide R, by the pink horizon labelled P Slide 2 Base. The latter follows the top of the drift deposits and is parallel with the internal reflections of the drift. CD: contouritic drift deposits; SD: slide debris; TNU: top Naust unit U reflection; INS2: intra Naust unit S reflection 2; TNR: top Naust unit R reflection. For further discussion, see Solheim et al. (2005).
Modified from work by Laberg and Camerlenghi (2008); with permission from Elsevier.

(b) *Low shear-strength:* This results from relatively high sedimentation rates (Mulder et al., 2003; Laberg and Vorren, 2004) and well-sorted grain-size (Wilson et al., 2004), both of which imply high water content (Kvalstad et al., 2005).
(c) *Under-consolidation and excess pore pressure:* These are especially generated in low-permeability, fine-grained contourites, and high-porosity siliceous ooze layers (Volpi et al., 2003).
(d) *Loading:* For contourites on continental slopes this can be rapid (e.g. glacigenic sediments on high-latitude margins; see: Laberg and Vorren, 2004) and cyclic (earthquakes), causing liquefaction (see: Sultan et al., 2004).
(e) *Gas charging:* This includes dissociation of gas hydrates after ocean-warming by thermohaline currents (Mienert et al., 2005) and migration of gases from relatively high organic-carbon content produced by water masses.

3.3. Hydrocarbon exploration

Contour currents affect petroleum systems in many ways, including reservoir geometry and quality, and the distribution of sealing rocks (Shanmugam, 2006; Viana et al., 2007; Viana, 2008; Mutti and Carminatti, 2012; Shanmugam, 2012a; Brackenridge et al., 2013; Shanmugam, 2013a,b; Stow et al., 2013a,b). Changes in the seafloor topography produced by erosion or deposition induced by bottom currents can lead to the re-adjustment of the sediment accommodation space and to the creation of sub-basins, which act as sediment traps or as gateways for sediment transfer (Viana, 2008; Campbell and Deptuck, 2012). Furthermore, *coarse-grained contourites* deposited by robust flows can represent hydrocarbon reservoirs, whereas *fine-grained contourites* accumulated by weak bottom currents can provide sealing (and source) rocks (Rebesco, 2005).

3.3.1. Coarse-grained contourite reservoirs

Rippled, top-truncated, fine-to-medium-grained, sand-rich oil-bearing beds of Eocene age in the Campos Basin comprise contourite sands (Souza Cruz, 1995, 1998; Viana and Faugères, 1998; Viana, 2008; Mutti and Carminatti, 2012). Contourite reservoirs have also been described in the North Sea (Enjorlas et al., 1986). Traction-dominated deep-water sand deposits in the Gulf of Mexico that show very good potential as reservoirs (high porosities and high permeability values) have been interpreted as sandy contourites by Shanmugam et al. (1993b). However, they were later interpreted as fine-grained turbidites by Stow et al. (1998a), since they were not considered to unequivocally exhibit characteristics of contourites. Large volumes of sand transported and re-deposited by persistent and efficient hydrodynamic regimes have been reported for the Carnegie Ridge (Lonsdale and Malfait, 1974), the upper slope of the Campos Basin (Viana and Faugères, 1998), the lower Mississippi Fan (Kenyon et al., 2002), the Gulf of Cádiz (Habgood et al., 2003; Llave et al., 2005; Hernandez-Molina et al., 2006a; Brackenridge et al., 2013; Stow et al., 2013a,b; Hernández-Molina et al., 2014), and the Faroe–Shetland Channel (Wynn et al., 2002; Masson et al., 2004; Akhmetzhanov et al., 2007). In most cases, the occurrence of sandy contourites with potential for hydrocarbon exploration depends on the vicinity of a sand-rich area prone to sweeping by contour currents (Fig. 8).

3.3.2. Fine-grained contourite sealing rocks

Fine-grained contourites, which are related to sealing facies/permeability barriers and to source-rock accumulation, play an important role in the characterisation of deep-water petroleum systems. For instance, in the Santos Basin excellent sealing rocks of bottom-current origin have been found overlying the oil-bearing sandstones (Duarte and Viana, 2007); and in the Campos Basin fine-grained, extremely bioturbated sediments related to the action of bottom currents act as important permeability barriers and internal heterogeneities within thick packages of oil-rich sandstones (Moraes et al., 2007). Understanding the distribution and thickness of these rocks is fundamental for water-injection and recovery projects. Although the potential of contourites as source rocks is generally low, due to the ventilation induced by contour currents, contourite deposits have frequently been associated with accumulations of gas hydrates and of free gas (Viana, 2008; Mosher, 2011).

Despite the economic significance of contourites, the number of publications dealing with this topic is rather low. According to Viana et al. (2007), this might be explained by various factors: the early establishment of gravity-flow accumulation models; the volumetric predominance of sandy gravity-flow deposits; a lack of unequivocal sedimentological criteria; and poorly investigated oceanography aspects for contourite identification. However, the greater access to high-resolution seafloor imaging techniques and industrial 3D seismic

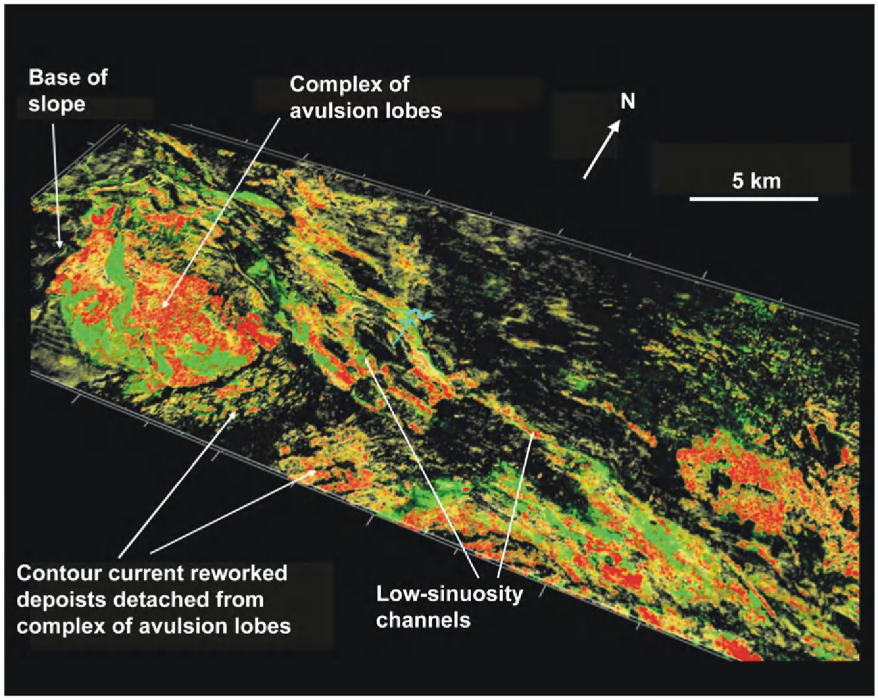


Fig. 8. Seismic-amplitude map of a subcropping Late Pleistocene turbidite/contourite complex. The warm colours (yellow and red) correspond to coarse-grained sediments, and the green, to fine-grained sediments. Low-sinuosity channels flow diagonally across the image, from NW to SE. Laterally to this system, to the South, occurs a fan-shaped body corresponding to a complex of avulsion lobes, distally reworked by contour currents developing a fringe of irregular patch-like small sedimentary bodies.
Modified from work by Viana (2008); with permission from Elsevier.

data that academic researchers have been enjoying is providing them with a deeper understanding of contourite deposits and confirming that these deposits are important constituents of petroleum systems (Shanmugam, 2006; Viana et al., 2007; Viana, 2008; Brackenridge et al., 2011; Mutti and Carminatti, 2012; Shanmugam, 2012a; Brackenridge et al., 2013; Hernández-Molina et al., 2013; Shanmugam, 2013a,b; Stow et al., 2013a,b). This new perspective is crucial for understanding the origin of deep-water sandstones and for predicting the distribution of

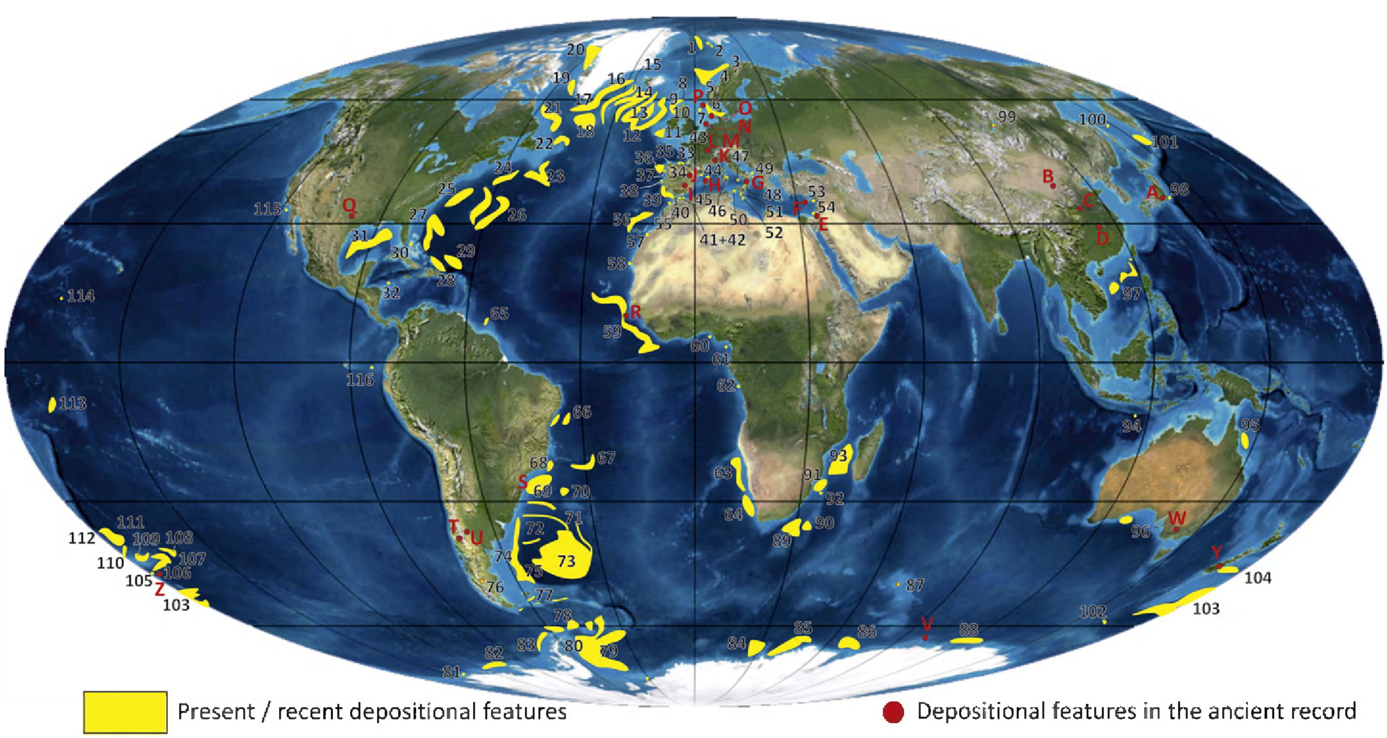


Fig. 9. Occurrence of large contourite deposits in the present (recent) ocean basins (yellow areas) and in the ancient sedimentary record (black points). This compilation was done specifically for the present review, but was subsequently archived and visualised by Claus et al. (in press) on the Marine Regions website (http://www.marineregions.org). The numbers and letters in each area of the Figure refer to the key references in Appendix A.

these sediments in future petroleum exploration and production endeavours worldwide (Shanmugam, 2013b).

The most recent findings on contourites in the context of hydrocarbon exploitation have been provided by IODP Expedition 339, which verified the presence of clean, well-sorted contourite sands (thickness: up to 10 m) from high-quality potential hydrocarbon reservoirs, that are completely different from the turbidite sands from deep-water oil and gas plays (see Section 8). These sands are associated with very thick contourite muds that are moderately rich in organic carbon and could provide potential suitable seals and/or source rocks. These new findings could herald a paradigm shift in deep-water targets for hydrocarbon exploration (Expedition 339 Scientists, 2012; Hernández-Molina et al., 2013; Stow et al., 2013b).

4. The occurrence of contourites

Since the pioneering work of McCave and Tucholke (1986) and of Faugères et al. (1993), contourites have been described mainly in the North and South Atlantic basins, predominantly associated to the deeper water masses (e.g. NADW and AABW). However, over the past 2 decades, contourite features have been described in other areas in the Atlantic Ocean as well as in the Mediterranean, Indian, Pacific, Arctic and Antarctic realms. Fig. 9 shows an updated compilation demonstrating that contourite features are ubiquitous within the oceanic basins. The figure shows some 116 identified areas, including the major, published contourite features from the present or recent past (as in Fig. 10), but excluding very minor features and unpublished data. New findings have been observed in different settings and associated to either *deep* (e.g. Borisov et al., 2013; Martos et al., 2013), *intermediate* (e.g. Van Rooij et al., 2010; Rebesco et al., 2013) or shallow water masses (e.g. Vandorpe et al., 2011). Most of the described large contourite depositional and erosional features are located in the western side of the largest oceanic basins, and around the Antarctic and Arctic oceans (Fig. 9). They extend from the upper slope/outer shelf to the abyssal plains (Fig. 10). However, the absence of contourites does not necessarily imply an absence of depositing density currents, since many years of accumulation can be erased by highly intermittent or episodic oceanographic processes (see Section 5).

Faugères et al. (1993) reported significant differences in the distribution of different types of drift between the North Atlantic and the South Atlantic, suggesting that giant marginal elongate drifts were the prominent drift type in the former, whereas contourite sheets and channel-related drifts were more typical in the latter. They considered the morphological pattern, and its interaction with bottom water, as the major factors that had controlled the style of contourite accumulation. However, according to current knowledge, their findings probably stemmed from the fact that in the North Atlantic researchers had extensively studied the continental slopes, whereas in the Southern Atlantic

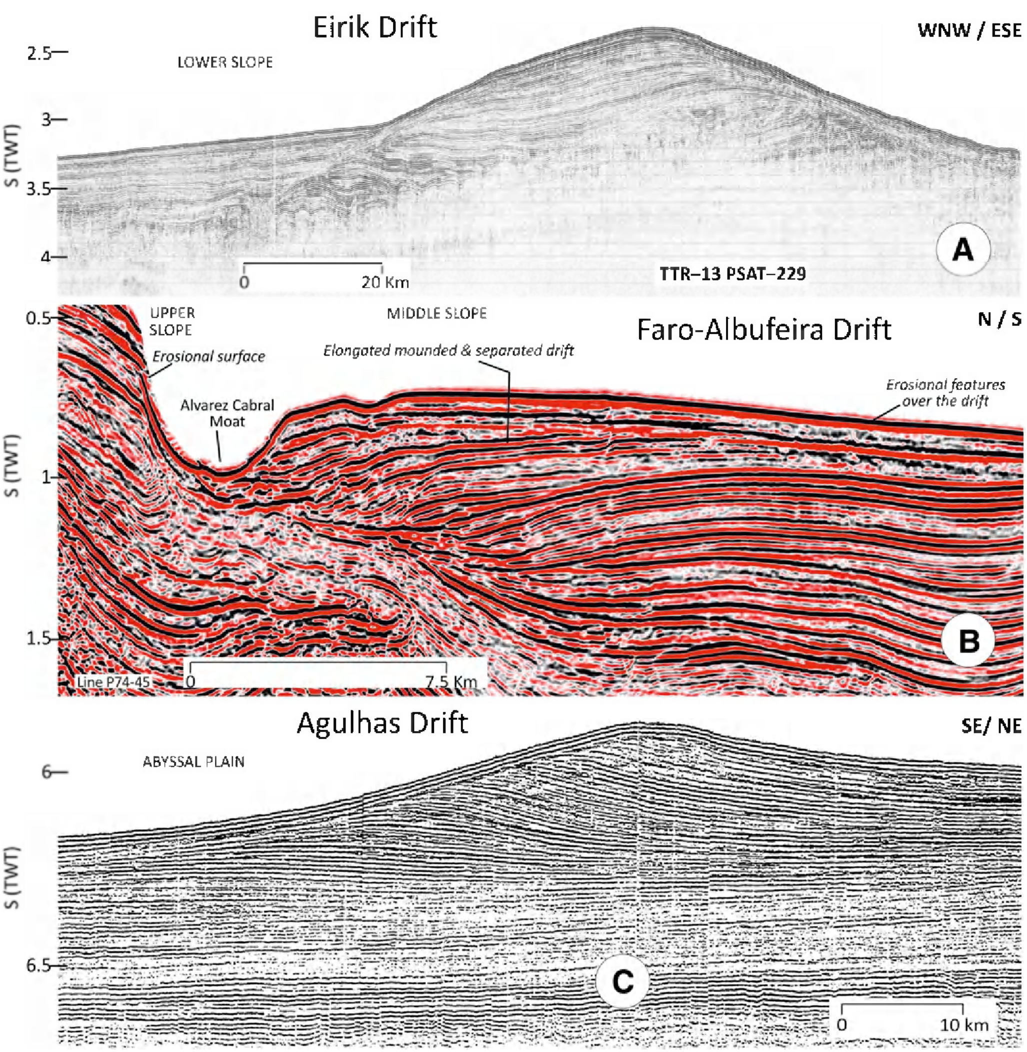


Fig. 10. Examples of three large contourite drifts: A) Eirik Drift, Greenland margin, northern hemisphere; B) Faro-Albufeira Drift, Gulf of Cádiz margin, northern hemisphere (Llave et al., 2001; Brackenridge et al., 2013; Stow et al., 2013a, Seismic line provided by REPSOL Oil for this work); and C) Agulhas Drift, Transkei Basin, southern hemisphere.
Panel A: from Hunter et al. (2007b) and Hernández-Molina et al. (2008b); with permission from Elsevier; Panel C: Niemi et al. (2000), with permission from Elsevier.

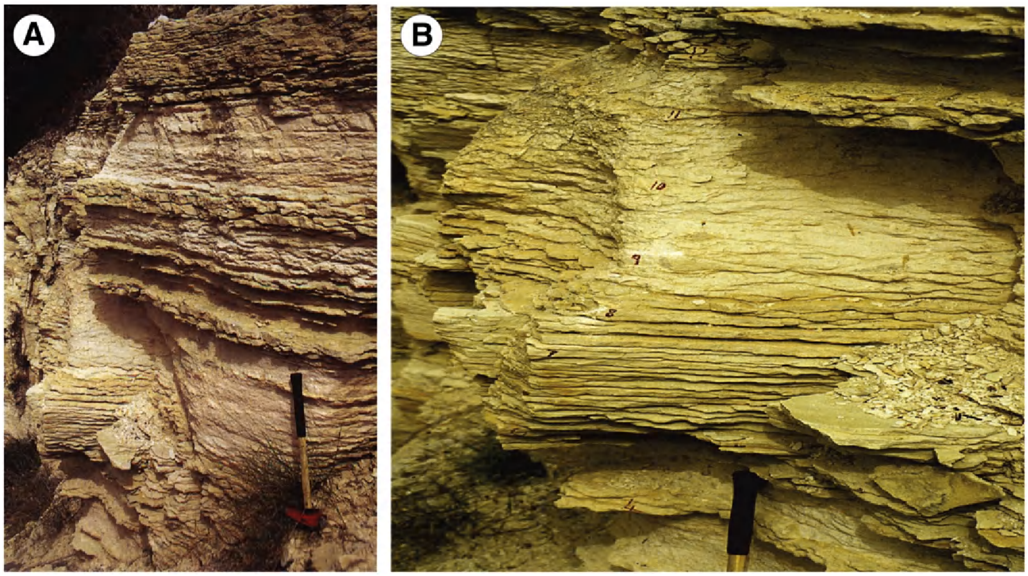


Fig. 11. Ancient contourite outcrops (A & B) of the Lefkara Formation, Cyprus (courtesy of D.A.V. Stow, Heriot-Watt University, UK).

researchers had studied mainly the abyssal plains. New discoveries have confirmed that contourite features are very common on many slopes, continental rises, abyssal plains and around banks, seamounts, etc. (see the compilations in Hernández-Molina et al., 2008a,b). Drift types (see Section 6) are significantly controlled by the physiographic and geological setting in which they develop and by different water masses that flow at different depths, different velocities and in the same or opposite directions.

Recognising contourite deposits in ancient sedimentary series (Fig. 11) that are presently exposed on land is not trivial (Stow et al., 1998a,b; Huneke and Stow, 2008). Moreover, there is some controversy surrounding the recognition and interpretation of bottom-current deposits and of (contour-current) reworked turbidites in ancient series (Stanley, 1988; Pickering et al., 1989, 1995; Stow et al., 1998a,b; Shanmugam, 2000; Stow et al., 2002a; Martín-Chivelet et al., 2003; Huneke and Stow, 2008; Martín-Chivelet et al., 2008; Stow et al., 2008; Mutti and Carminatti, 2012; Shanmugam, 2012a, 2013b). The diagnostic criteria for these features comprise their facies, structures, ichnofacies, texture, sequences, microfacies, and composition (Huneke and Stow, 2008). Furthermore, sedimentary structures of ancient contourites are also "diagnostic indicators" of the currents from which they were derived; however, interpretation of these contourites requires consideration of their full context—particularly, the genetic association between them and other structures, and the palaeogeographic framework (Martín-Chivelet et al., 2008, 2010). Fieldwork and observation (where permitted by exposure) of medium-scale criteria, such as the recognition of hiatuses and condensed deposits, or of variation in the thickness of depositional units, geometry, palaeowater depth or geological context, can be definitive for contourite recognition. Large-scale diagnostic criteria, including palaeoceanographic features and continental margin reconstructions, are essential, although their application to outcrops is generally problematic (Huneke and Stow, 2008). Thus any gain in knowledge on ancient contourite deposits will require a better understanding of the present and recent contourite processes and products, as well as better comprehension of the palaeogeography of ancient oceanic basins and of the past water-circulation model. Over the past few years examples of contourites in ancient rock series in different areas were published (see tentative compilation in Fig. 9). These range in age from Cambro-Ordovician to Neogene and their distribution denotes that most of them are located following an east-to-west trend, which is probably associated to evolution of the Palaeo Tethys Ocean.

5. Oceanographic processes that affect contourite formation

The deep waters of the oceans are formed primarily in marginal seas or shallow shelf regions where the water is made cold and dense by cooling and/or ice formation (Fig. 12), or highly saline upon strong evaporation (e.g. Ambar and Howe, 1979; Dickson and Browne, 1994; Price and Baringer, 1994; Girton and Sanford, 2003; Rahmstorf, 2006; Kuhlbrodt et al., 2007). The relatively dense water formed there flows into the ocean via narrow or shallow straits, or via the continental margin; in the Northern Hemisphere, it is steered to the right by the Earth's rotation (Fig. 12). Once the water is no longer constricted by the topography, it reshapes into a wider structure that adjusts to the forces of gravity, Earth's rotation, and bottom friction. The distribution of salt and heat in the deep ocean is strongly related to these dense currents and to the rates at which they descend to greater depths and subsequently mix with ambient fluid (Munk and Wunsch, 1998; Wells and Wettlaufer, 2005; Wåhlin and Cenedese, 2006; Legg et al., 2009; Akimova et al., 2011). Since the deep waters of the oceans are formed by atmospheric forcing, their existence and properties give information on the regional climate of the areas in which they are formed (Bartoli et al., 2005; Rogerson et al., 2012) and in some cases, also on the global climate (Broecker, 1991; Rahmstorf, 2006; Kuhlbrodt et al., 2007; Legg et al., 2009).

The Earth's rotation tends to steer bottom currents to flow parallel to large-scale bathymetry (e.g. along the continental margins). Small-scale topographic features such as seamounts, ridges, straits, mounds, banks and canyons can disrupt and accelerate the flow. Once the current velocity becomes sufficiently high, the sediment erodes, and as the velocity later decreases, the sediment is deposited (Stow et al., 2002a; Rebesco and Camerlenghi, 2008). Fig. 13 is a sketch of some of the processes that affect erosion and deposition in continental margins and deep basins.

Currents can be either barotropic or baroclinic. Barotropic flows are currents with constant pressure surfaces parallel to surfaces of constant density (e.g. the well-mixed shallower [shelf] part of the ocean; see Shanmugam, 2013a). In general, these flows are driven by pressure gradients caused by sloping sea surface. Examples of barotropic flows include tides, wind-driven currents and surface waves. In contrast, baroclinic flows are currents with constant pressure surfaces that are not parallel to surfaces of constant density (e.g. a cold dense plume flowing into a warmer, lighter ocean environment). These flows are usually driven by pressure gradients caused by horizontal density

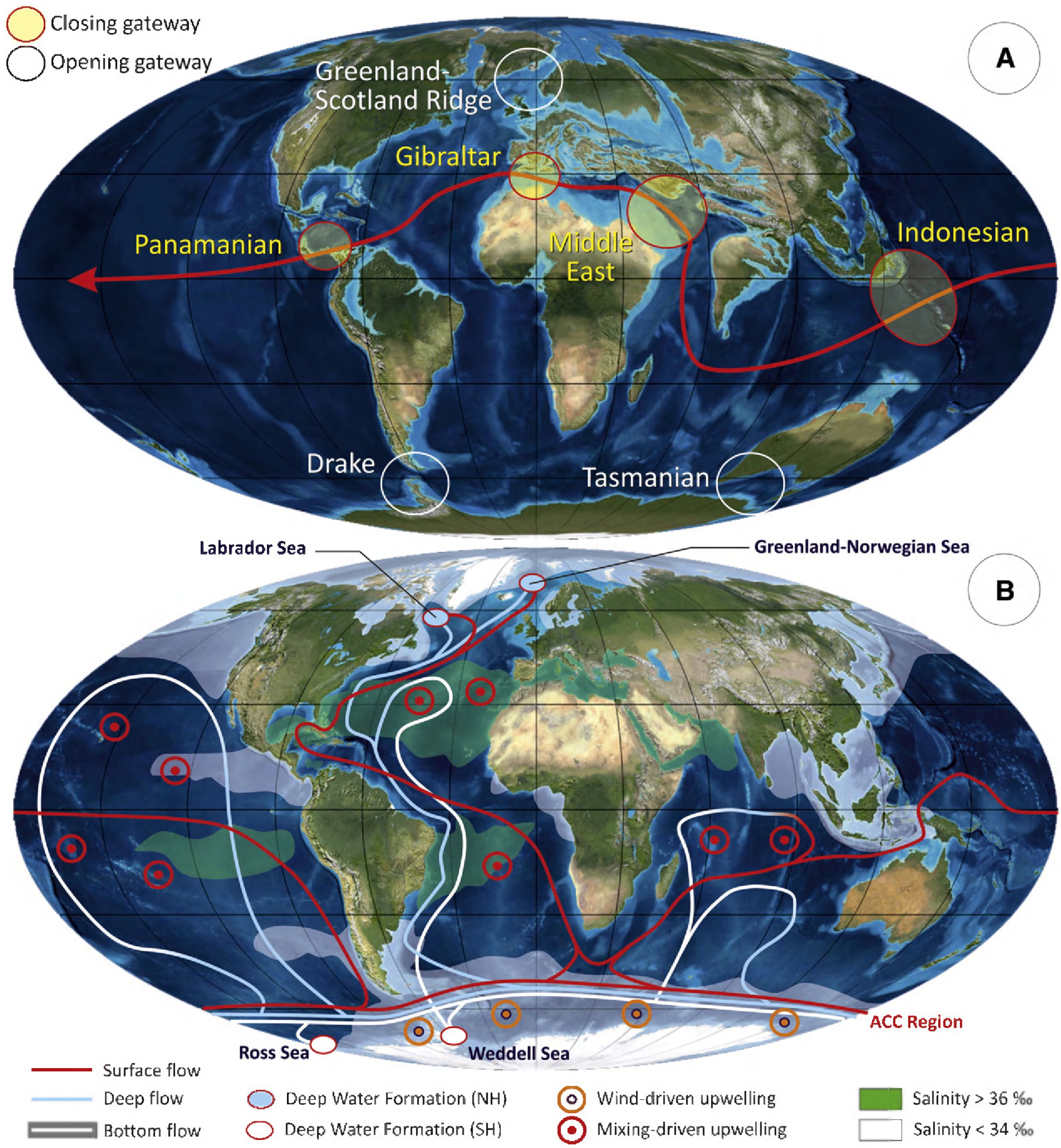


Fig. 12. A) Global distribution of the marine basin during the Eocene (56–33.9 Ma.), illustrating the closing and opening of the main gateways, which have determined drastic changes in the deep-water circulation. B) Global thermohaline circulation. Red: surface currents; Light blue: deep water; White: bottom water; Orange: main sites of deep-water formation. In the Atlantic, warm and saline waters flow northward from the Southern Ocean into the Labrador and Nordic seas. In contrast, there is no deep-water formation in the North Pacific, whose surface waters are consequently fresher. Deep waters formed in the Southern Ocean become denser than those from the North Atlantic and therefore spread at deeper levels. Note the small, localised deep-water formation areas in comparison to the widespread zones of mixing-driven upwelling. Wind-driven upwelling occurs along the Antarctic Circumpolar Current.
Panel A: Adapted from Seibold and Berger (1993); Panel B: Adapted from Rahmstorf (2006) and Kuhlbrodt et al. (2007); reproduced with permission of Elsevier. The base maps from A & B are from Ron Blakey, Colorado Plateau Geosystems (http://cpgeosystems.com/mollglobe.html).

differences and fronts. Baroclinic currents are found along the ocean floor of continental slopes and in submarine canyons (Fig. 13) (Robertson and Ffield, 2005; Allen and Durrieu de Madron, 2009; Shanmugam, 2012a, 2013a), amongst other locations. Examples of baroclinic flows include internal waves and internal tides that propagate on density surfaces (pycnoclines) in stratified waters.

Bottom currents are typically baroclinic: their velocity typically correlates to the strength of their density gradient. The water velocity at the seafloor can also be affected by *barotropic currents* (Stow et al., 2002a; Rebesco, 2005; He et al., 2008), *tides* (Gao et al., 1998; Stow et al., 2013a) or *intermittent processes* such as giant *eddies* (Serra et al., 2010), *deep sea storms* (Hollister et al., 1974; Hollister, 1993), *vortices* (Preu et al., 2013), *internal waves* and *tsunamis* (Shanmugam, 2012b, 2013a,b). Most of these phenomena are associated with stronger currents and therefore, are likely to induce erosion. However, they can also be associated with transport of sediment-laden water and consequently, with deposition of sediments (as in the case of tsunamis). Below, some of these processes are overviewed and their possible influence in the formation of erosional and depositional seafloor features is discussed.

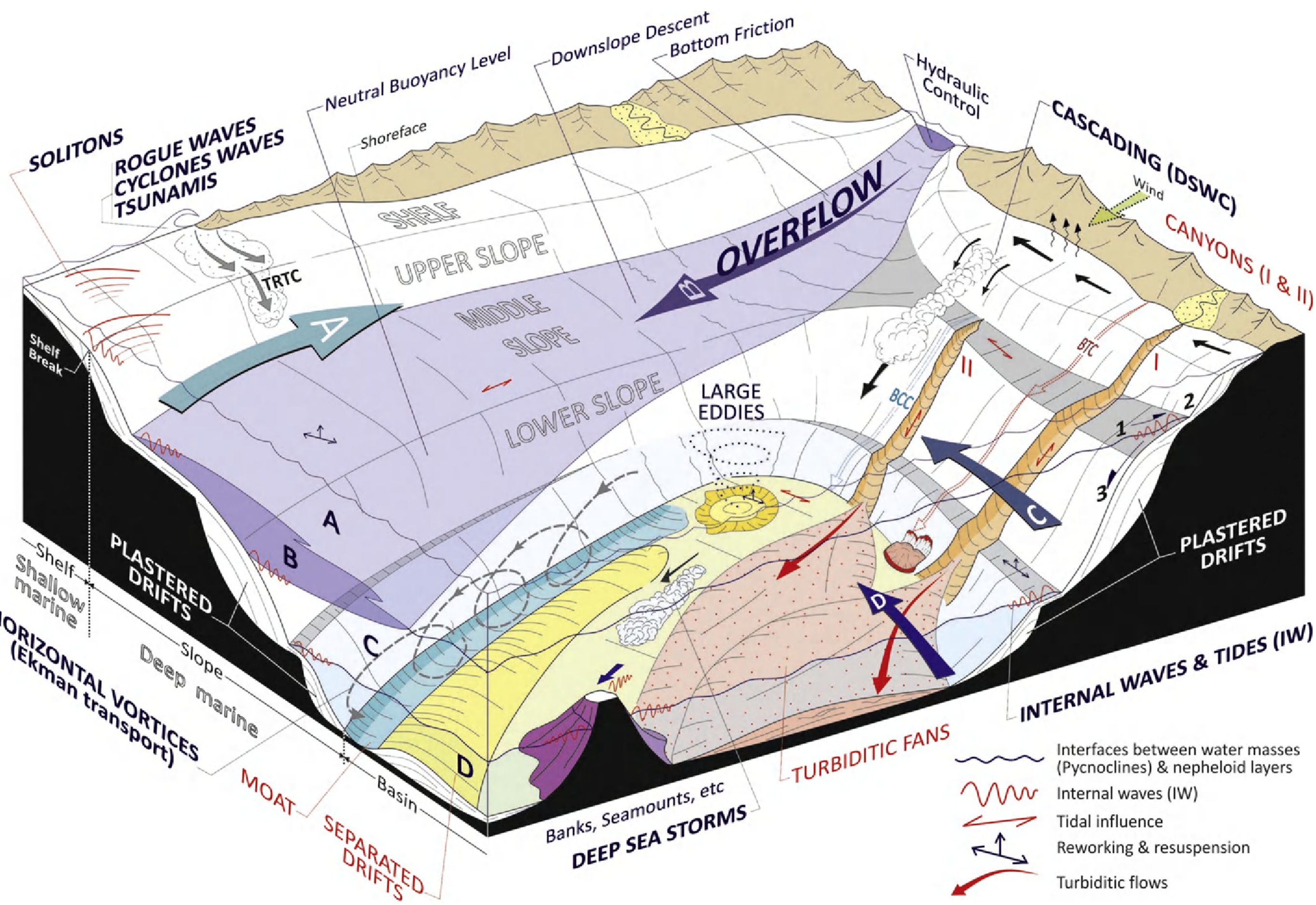


Fig. 13. 3D sketch depicting the possible oceanographic processes in deep-water environments. The velocity at the seafloor can be affected by density currents and overflows, as well as by barotropic currents and by intermittent processes such as cascading, giant eddies, deep sea storms, vortices, internal waves, internal tides, tsunamis, cyclone waves, and rogue waves (for further explanations, see text).

5.1. Density-driven currents

Density-driven currents that are wider than a few tens of km tend to flow in geostrophic balance parallel to depth contours (see, for example: Wåhlin and Walin, 2001; Legg et al., 2009). The density forcing can be maintained by cooling, evaporation or a combination of both, in which case the current is said to be *thermohaline*. These currents are responsible for the deposition of sensu stricto contourites. Density-driven bottom currents (Figs. 12 and 14) can be found in various areas, such as the *Denmark Strait* (e.g. Smith, 1975; Girton and Sanford, 2003; Käse et al., 2003; Karcher et al., 2011); *Faeroe-Bank channel* (Borenäs and Lundberg, 2004; Mauritzen et al., 2005; Riemenschneider and Legg, 2007; Darelius et al., 2011); *Gulf of Cádiz* (Smith, 1975; Borenäs et al., 2002; Johnson et al., 2002); *Red Sea* (Peters et al., 2005; Matt and Johns, 2007), the *South China Sea* (Gong et al., 2013); *Weddell Sea* (Orsi et al., 1999; Naveira Garabato et al., 2002a,b, 2003; Foldvik et al., 2004; Nicholls et al., 2009); and *Ross Sea* (Carter et al., 2008; Capello et al., 2009; Muench et al., 2009a,b).

The bottom currents in the North Atlantic are the sources for one of the major deep-water masses on Earth (Fig. 12): the North Atlantic Deep Water (NADW) (Dickson and Browne, 1994). The combined overflows in the Southern Ocean form the Antarctic Bottom Water (AABW) (Nicholls et al., 2009; Kida, 2011), the other major and the densest deep-water mass on Earth in the present-day climate. The Mediterranean Outflow Water (Fig. 12) contributes to a relatively warm and salty water mass found at intermediate depths (ca. 1000 to 1500 m) in the Atlantic (Ambar and Howe, 1979; Baringer and Price, 1997). Most density-driven bottom currents are ultimately created by atmospheric processes (e.g. Kida et al., 2009). For example, the Mediterranean and the Red Sea Outflows are caused by high evaporation in the warm regional climates of the Mediterranean and Red Seas (Ambar and Howe, 1979; Arnone et al., 1990; Baringer and Price, 1997; Peters et al., 2005), whereas the North Atlantic and Southern Ocean overflows are characterised by cold and salty water produced by surface cooling and/or freezing (Dickson and Browne, 1994; Rahmstorf, 2006; Kuhlbrodt et al., 2007; Carter et al., 2008; Nicholls et al., 2009). Regardless of their mechanisms of formation, all of these currents exit, through a narrow or shallow strait, into a larger ocean. Upon exit, the dense water is steered to the right (in the Northern Hemisphere) by the Earth's rotation. Once the water is no longer constricted by the topography (Figs. 13 and 14), it reshapes into a wider structure in which the scale speed is proportional to the slope of the bottom and to the density difference between the density current and the overlying water mass (e.g. Price and Baringer, 1994; Borenäs and Wåhlin, 2000; Cenedese et al., 2004; Kida et al., 2009; Legg et al., 2009; Akimova et al., 2011), according to:

$$U_N = \frac{g'a}{f} \tag{1}$$

whereby $g' = g\frac{\Delta\rho}{\rho_0}$ is the reduced gravity (in which g is the gravitational acceleration; and $\Delta\rho = \rho - \rho_0$ is the difference in density between the

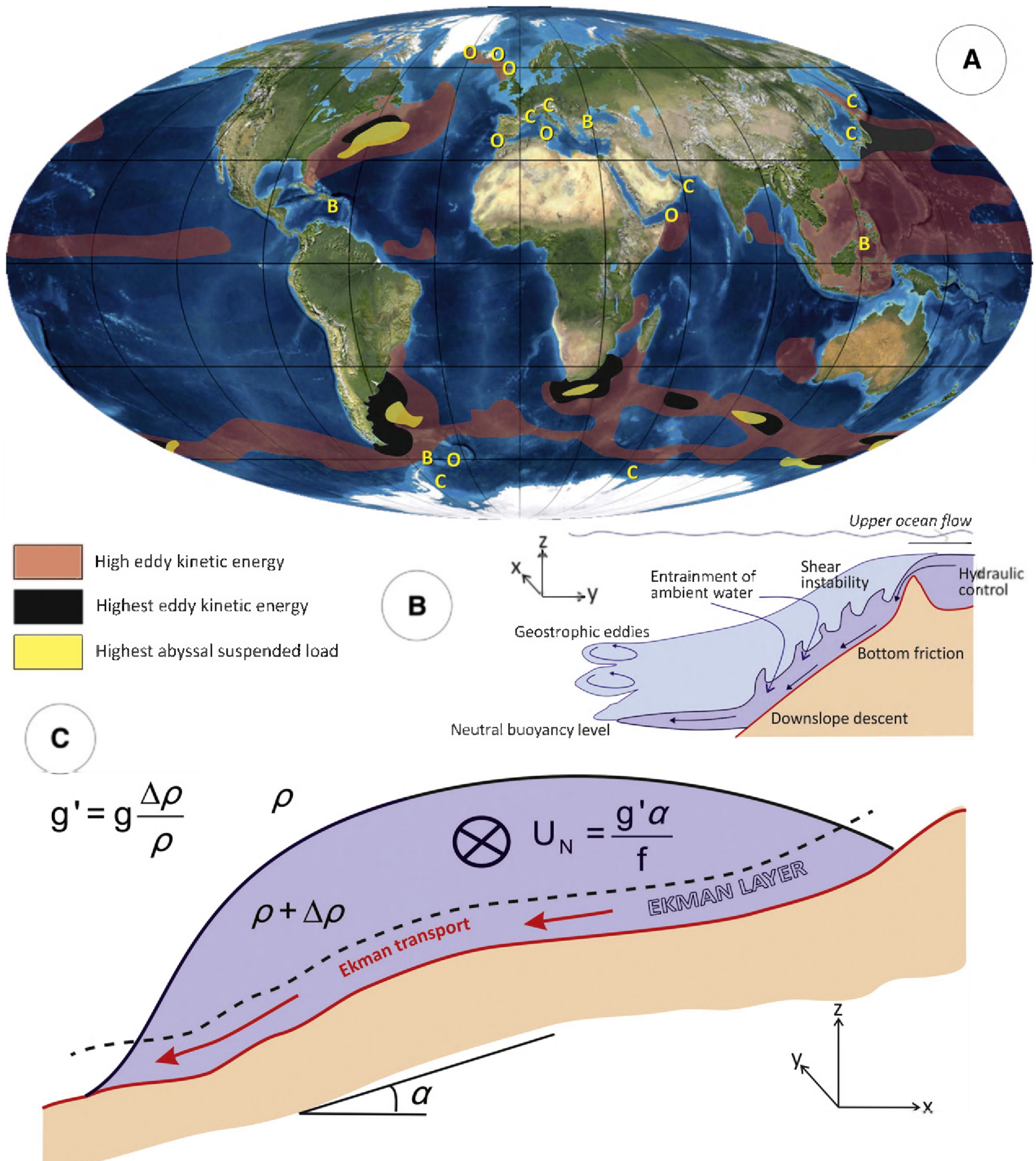


Fig. 14. A) Map showing the relationship between kinetic energy and suspended load, indicating the areas with higher suspended load in deep basins (Bearmon, 1989; Pickering et al., 1989). The position of main gravity currents by type is also indicated (O: Overflow across a topographic barrier from a regional basin into the open ocean; B: Open-ocean overflow into an isolated regional basin; C: Cascade of dense water from a continental shelf). Not shown: numerous overflows across multiple sills of the mid-ocean ridge system, within the series of basins of the western South Pacific; and the cascades of shelf water over the slope of the Arctic Sea (adapted from Legg et al., 2009). The base map is from Ron Blakey, Colorado Plateau Geosystems by (http://cpgeosystems.com/mollglobe.html). B) Physical processes acting in overflows (adapted from Legg et al., 2009). C) Sketch of a dense overflow showing the coordinate system and some of the notations used (ambient density: ρ; plume density: $\rho + \Delta\rho$; reduced gravity: g'; bottom slope: α; Coriolis parameter: f; and Nof velocity: U_N). Also shown are the Ekman layer and the benthic Ekman transport (see, for example: Pedlosky, 1996; Wåhlin and Walin, 2001).

density current [ρ] and the ambient water [ρ_0]); α is the slope of the bottom; and f is the Coriolis frequency (Fig. 14).

Table 1 lists the properties of some of the major dense outflows after they enter onto the continental slope. As observed in the table, the flows are wide (as compared to the Rossby radius of deformation) but thick (as compared to the frictional boundary layer). Consequently, they are expected to be geostrophically balanced to lowest order (i.e. to have the average velocity [1] [also called the *Nof velocity*]). The velocity (1) is also the speed at which cold eddies translate along the slope (Nof, 1984) and at which the leading front of a dense water mass propagates (Wåhlin, 2004).

Entrainment of bottom water, bottom friction, and inertial accelerations modify the bottom current on the slope (Fig. 14). Bottom friction induces an Ekman transport in the bottom boundary layer (e.g. Pedlosky, 1996; Wåhlin and Walin, 2001), which in the Northern Hemisphere is directed to the left of the flow velocity. Although the frictional transport is confined to a thin layer next to the bottom, *the Ekman boundary layer*, it affects the entire water column. The dense water adjusts to the

Table 1
Key properties of some of the large dense outflows. The parameters at the bottom of the table show that whilst rotation has varying degrees of influence, in all the overflows the Froude number can exceed 1 in regions associated with entrainment.
Adapted from Legg et al. (2009).

KEY PARAMETERS OF LARGE DENSE OVERFLOWS

	Parameters	Faroe Bank	Denmark Strait	Ross Sea	Weddell Sea	Red Sea	Mediterranean Sea
SOURCE WATER	Potencial temperature °C	0	0.25	−1.9	−1.9	22.8	14
	Salinity	34.92	34.81	34.8	34.67	39.8	38.4
	Density (σ_0)	28.07	27.94	27.9	27.8	27.7	28.95
	Sill depth (m)	800	500	600	500	200	300
PRODUCT WATER	Potencial temperature °C	3.3	2.1	−1.0	−1.0	21.7	11.8
	Salinity	35.1	34.84	34.72	34.67	39.2	36.4
	Density (σ_0)	27.9	27.85	27.85	27.75	27.48	27.6
	Depth (m)	3000	1600	>3000	2000	750	850
TRANSPORT (Sv)	Source	1.8	2.9	0.6	1.0	0.3	0.8
	Product	3.3	5.2	2	5	0.55	2.3
VELOCITY *U* (m / s)	Source	1	0.7	1	1	0.55	1
	Product	0.5	0.4	0.5	0.4	Small	1
THICKNESS *H* (m)	Source	200	200	150	100	100	200
	Product	150	200	250	300	140	200
TRANSIT TIME SOURCE TO PRODUCTS (Days)		3 – 5	4 – 7	1	1	3.5	3 - 5
TIDAL CURRENT (m / s)		0.2	0.1	0.3	0.2	0.8	0.1
REDUCED GRAVITY (g') $g' = g\,\Delta\rho / \rho_0$			2×10^{-3}	2×10^{-3}	2×10^{-3}	1.4×10^{-2}	1.6×10^{-3}
FROUDE NUMBER (Fr) $Fr = U / \sqrt{g'H/f}$		1	0.3 – 1.2	0.9 – 1.1	1	0.6 – 1.3	1
ENTRAINMENT RATE (W_e / U)		5×10^{-4}	1×10^{-3}	6×10^{-3}		2×10^{-4}	$5 - 20 \times 10^{-4}$
CORIOLIS PARAMETER f (s^{-1})		1.3×10^{-4}	1×10^{-4}	1.3×10^{-4}	1.3×10^{-4}	1.3×10^{-5}	3.1×10^{-5}
DEFORMATION RADIUS (km) $L_R = \sqrt{g'H/f}$		30	5	7	7	40	100

divergence of the frictional transport, which acts as a horizontal diffusive process, minimising the curvature of the dense interface. The lower (seaward) edge moves downhill as the Ekman transport from the interior is expelled from the Ekman layer (e.g. Condie, 1995; Wåhlin and Walin, 2001). At the upper (landward) edge, the dense interface instead becomes nearly horizontal, with low geostrophic velocities and minor frictional transport (e.g. Jungclaus and Backhaus, 1994; MacCready, 1994; Wåhlin and Walin, 2001). The combined frictional effect over the entire outflow causes it to gradually widen (Fig. 14), keeping the upper horizontal boundary at a nearly constant depth (e.g. Jungclaus and Backhaus, 1994; Price and Baringer, 1994; Borenäs and Wåhlin, 2000). If these processes occur over a laterally varying bottom slope, they can split the flow into two or more cores and branches, as has been suggested for the Mediterranean Outflow (Borenäs et al., 2002; Serra et al., 2010), amongst other cases. The inertial accelerations induce waves and eddies in the overflow (Fig. 14). The stability of a dense current flowing along a sloping bottom has been analysed by Griffiths and Linden (1981) and by Sutherland et al. (2004). Their results suggest that, if friction is disregarded, these flows are nearly always unstable, and that this instability breaks the dense current up into a train of dense eddies. This effect has also been studied in the outflow from the Red Sea (Nof et al., 2002).

As dense water passes from its site of formation, through an overflow and into the open ocean, it mixes with overlying water (Figs. 13 and 14). This mixing determines the hydrographic properties and volume of the produced water mass (Dickson and Browne, 1994; Kida et al., 2009). Laboratory experiments (see, for example: Ellison and Turner, 1959; Turner, 1973; Cenedese et al., 2004) have shown that if the speed of the dense water exceeds the phase speed of a long internal wave, then the flow becomes unstable and starts to overturn. This induces mixing between the dense water and the overlying water mass, which is rarely resolved in models and nearly always needs to be parameterised. Most of the presently used entrainment parameterisations have forms in which the entrainment rapidly transitions into a high-entrainment flow, when the velocity exceeds that of the speed of a long internal wave. Cenedese et al. (2004) found that in overflows and in dense currents the presence of eddies and waves can further enhance the vertical mixing.

Expression (1) is a rule-of-thumb for the large-scale average velocity. Locally, the velocity can reach much higher values: for instance, when it encounters small-scale topographical features such as *submarine canyons* (e.g. Wåhlin, 2004; Allen and Durrieu de Madron, 2009; Muench et al., 2009a,b), ridges (Darelius and Wåhlin, 2007) or *seamounts* (Kennett,

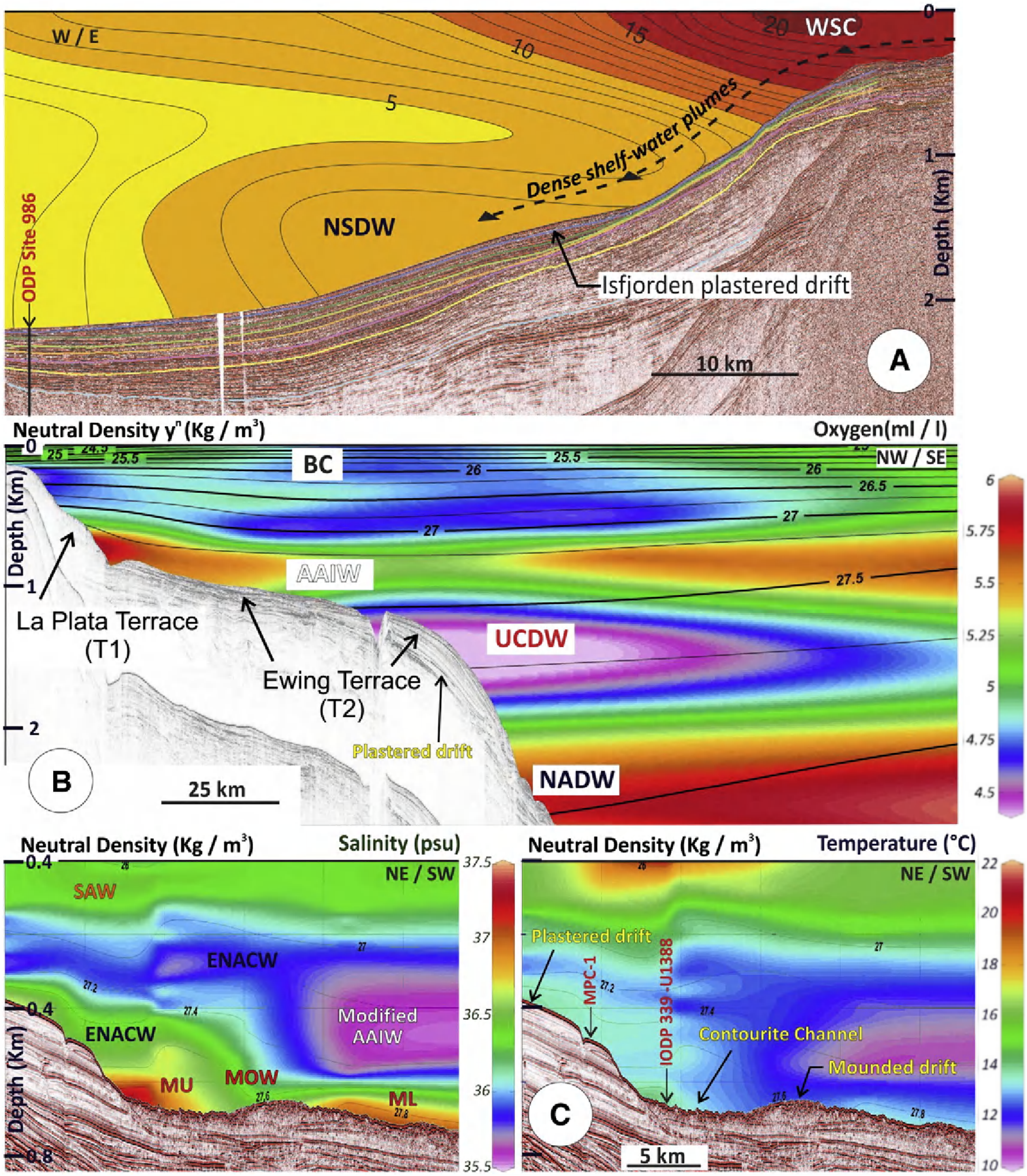


Fig. 15. A. Examples of combining physical oceanographic data with geologic/geophysical data, showing the relationship amongst the long-term current regime, the seafloor morphology and the sub-bottom sediment geometry. A) Western Spitsbergen margin; B) Argentine margin, North of the Mar del Plata Canyon; and C) Gulf of Cádiz, from the exit of the Strait of Gibraltar. The black numbers and lines in (A) refer to current velocity (cm/s), but in (B) and (C) they refer to isopycnals and neutral density (kg/m^3). Legend for water masses, in alphabetical order (AAIW: Antarctic Intermediate Water; BC: Brazil Current; ENCW: Eastern North Atlantic Central Water; MC: Malvinas Current; Modified AAIW: Modified Antarctic Intermediate Water; MOW: Mediterranean Outflow Water [MU: Upper Core, and ML: Lower Core]; NADW: North Atlantic Deep Water; NSDW: Norwegian Sea Deep Water; SAW: Surface Atlantic Water; UCDW: Upper Circumpolar Deep Water; WSC: West Spitsbergen Current).
Panel A: Rebesco et al. (2013), with permission from Elsevier; Panel B: Preu et al. (2013), with permission from Elsevier; Panel C: Hernández-Molina et al. (2014), with permission from Geological Society of America.

1982; McCave and Tucholke, 1986; Hernández-Molina et al., 2004, 2006b; Stow et al., 2009; Hernandez-Molina et al., 2011a). Such regional variations are important to the formation of sediment deposits (Fig. 15A), since they can locally erode the seafloor and keep particles in suspension for longer periods. Thus, the velocity of deep currents, including any small-scale variations, affects the lateral transport of sediment. An example of this is the deep western boundary currents, which on average, flow with the large-scale velocity (1), but locally can reach velocities that are many times higher (Table 1). The amount of suspended particulate matter can be ten times greater in the deep western boundary currents than in overlying water masses (Ewing et al., 1971; Kennett, 1982; Gao et al., 1998; Tucholke, 2002). Other examples of sedimentary structures induced by bottom currents are erosional features (see Section 6). The thickness of the nepheloid layer is generally 150 to 1500 m, and the average concentration of suspended matter is ca. 0.01 to 0.5 mg L^{-1} (McCave, 2008). The residence time for particulate material in deep nepheloid layers is estimated to last several days to weeks for the first 15 m above the seafloor, and weeks to months for the first 100 m above the seafloor (Kennett, 1982; Gao et al., 1998).

An active bottom current, acting for a prolonged period of time, will affect the seafloor, leading to processes generated from winnowing of fine-grained sediments up to large-scale erosion and deposition (Heezen, 1959; Stow and Lovell, 1979; Kennett, 1982; Pickering et al., 1989; Seibold and Berger, 1993; Stow, 1994; Einsele, 2000; Shanmugam, 2006; Rebesco and Camerlenghi, 2008; Stow et al., 2009). The generation of large depositional and/or erosional features requires a bottom current that is persistent on geological time scales (i.e. millions of years; Hernández-Molina et al., 2008a,b). An example of this is the main CDS that, in many cases, began to grow around the Eocene/Oligocene boundary (ca. 32 Ma) and were later reactivated in the Middle Miocene (Kennett, 1982; Sykes et al., 1998; Niemi et al., 2000; Uenzelmann-Neben, 2001; Flood and Giosan, 2002; Pfuhl and McCave, 2005).

5.2. Processes in the interface between water masses

The interface between two overlying water masses of different density is called a *pycnocline* (Fig. 13). It can be sharp and well-defined, or diffuse with a gradual transition from one water mass to the other. The pycnocline is maintained by stratifying processes that are caused by a regional positive buoyancy flux at the surface, and it becomes eroded by turbulent mixing between the water masses, such as that caused by tides. The relative strength of these two effects determines how well defined the pycnocline is in different regions and at different times of year. The interface often tilts in one direction (e.g. Reid et al., 1977) but can be locally and temporarily displaced by eddies (e.g. Piola and Matano, 2001; Arhan et al., 2002, 2003) and internal waves (i.e. waves that travel on the interface). Pycnoclines are characterised by energetic current patterns associated with these waves and eddies (Reid et al., 1977), which can shape the seafloor (e.g. Hernández-Molina et al., 2009, 2011a; Preu et al., 2013. Fig. 15B) and result in erosion and re-suspension of sediments (Dickson and McCave, 1986; van Raaphorst et al., 2001; Bonnin et al., 2002; Cacchione et al., 2002; Hosegood and Van Haren, 2003; Shanmugam, 2013a). Hence, the detection and characterisation of pycnoclines and of their historical records are important for multidisciplinary research in modern oceans (Hernandez-Molina et al., 2011a; Preu et al., 2011, 2013).

5.3. Deep-water tidal currents

Tides are generated at the surface and therefore, are barotropic (Fig. 13). However, tidal energy can transfer to a baroclinic wave, in what are known as *baroclinic tides* (Fig. 13). Baroclinic tides can move tidal energy away from its region of origin such that it dissipates, for example, on abrupt topography miles away. This mechanism provides a substantial part of the ocean mixing that is required for sustaining the global meridional overturning circulation (Munk and Wunsch, 1998). Barotropic and baroclinic tides both influence the bottom-water circulation in deep-water environments (Dykstra, 2012). Tidal currents change direction with phase, describing an elliptical flow path often aligned along bathymetric features. Tidal energy tends to be elevated within submarine canyons and adjacent areas (e.g. Shepard, 1976; Shepard et al., 1979; Petruncio et al., 1998; Viana et al., 1998a; Kunze et al., 2002; Garrett, 2003; Shanmugam, 2012b; Gong et al., 2013; Shanmugam, 2013b) and in some contourite channels (Stow et al., 2013a). Shanmugam (2012a) has proposed that barotropic tide currents affect land- or shelf-incising canyons whose heads are connected to rivers or estuaries, but that baroclinic tide currents affect slope-incising canyons with no clear connection to a major river or estuarine system. Inversion of the bottom current direction by tidal influence has been reported outside these canyons (Kennett, 1982; Stow et al., 2013a). Deep-marine tidal bottom currents have velocities that commonly range from 25 to 50 cm/s (but that can reach 70 to 75 cm/s) and periods of up 1 to 20 h (Shanmugam, 2012b).

5.4. Deep-sea storms

One intermittent deep-water process that is closely related to eddy formation is the generation of *deep-sea storms* (also called *benthic storms* or *abyssal storms*), which remains poorly understood. These storms involve the periodic intensification of normal bottom-current flow alongslope or following the isobaths (Fig. 13), where their mean flow velocity typically increases by two to five times, especially close to boundaries of strong surface currents. The HEBBLE project was the first to document the occurrence of benthic storm events and demonstrated their importance in the winnowing, transport and redistribution of sediments (Hollister et al., 1974; Nowell and Hollister, 1985; Hollister, 1993). Once ripped up by the erosional effects of increased bottom shear, sediments can be transported by bottom currents and deposited in quiet regions downstream (Hollister and McCave, 1984; Flood and Shor, 1988; Von Lom-Keil et al., 2002). In some cases, the flow has a velocity exceeding 20 cm/s, a very high concentration of suspended matter (up to 5 g L^{-1}), and strong erosional capability. Although benthic storms typically last from 2 to 20 days (most often, 3 to 5 days), they can have much longer-lasting effects on the suspension of bottom sediment, production of plankton blooms, and supply of considerable amounts of organic matter to the drifts (Richardson et al., 1993; Von Lom-Keil et al., 2002). The regions subjected to particularly intense deep-sea storms can also exhibit significant erosion in their continental slopes and produce large submarine slides (Pickering et al., 1989; Stow et al., 1996; Gao et al., 1998; Einsele, 2000).

5.5. Eddies

The generation of vortices is commonly associated to the lateral distribution of water masses (Serra et al., 2010), and appears to be a significant mechanism for both the formation of nepheloid layers and the long-distance transport of sediment (Fig. 13). Eddies can arise when a water mass interleaves into a stratified environment, or when a current flows along a seafloor irregularity such as a canyon, seamount, or cape (Roden, 1987; Rogers, 1994; Arhan et al., 2002; Serra et al., 2010). In some regions, large eddies on the seafloor span thousands of kilometres, such as in the Argentine Basin (Cheney et al., 1983; Flood and Shor, 1988; Arhan et al., 2002; Hernández-Molina et al., 2009), the Weddell and Scotia Seas (Hernández-Molina et al., 2008a), Mozambique slope (Preu et al., 2011), and the Gulf of Cádiz and the margins west off Portugal (Serra et al., 2010).

5.6. Secondary circulation

The main current-cores of dense water masses are usually parallel to the isobaths (Fig. 13 and 15). These cores have been associated with certain contourite erosional elements such as moats and channels (e.g. McCave and Tucholke, 1986; Faugères et al., 1993; Faugères et al., 1999; Rebesco and Stow, 2001; Stow et al., 2002a; Rebesco and Camerlenghi, 2008; Stow et al., 2009; Faugères and Mulder, 2011). They are often associated with deposition on the downslope side, and erosion on the upslope side, and their origin has been linked to a helicoidal flow path (i.e. in the bottom-current there is the core flow plus a clockwise circulation). In geophysics, this type of flow structure is called a *horizontal eddy* (Davies and Laughton, 1972; Roberts et al., 1974; Roden, 1987; Rogers, 1994; McCave and Carter, 1997; Hernández-Molina et al., 2008a; Serra, 2004; Zenk, 2008). Although researchers have inferred a helicoidal flow from the morphology of the seafloor features, they have yet to explain these features based on quantification of the relevant oceanographic processes and morphological results—work that should pose an interesting research challenge for oceanographers and sedimentologists. Nonetheless, these features likely result from the Coriolis forces that focus the vortex against the adjacent seafloor of the slope, erode the right flank (in the Northern Hemisphere) of the channel and deposit sediment (drift) on the left side, where the current velocity is lower (Faugères et al., 1999; Llave et al., 2007). Due to the combined effects of Coriolis force and bottom friction (Fig. 14C) a secondary circulation is nearly always created in large-scale bottom currents. The secondary flow has fluid moving downhill comparatively fast in a thin bottom layer and advected back uphill at slower speeds throughout the full layer (for selected examples, see Wåhlin and Walin, 2001; Muench et al., 2009a,b; Cossu et al., 2010). Together with the main flow this creates an asymmetric helical flow path which has been visualized in the lab (see e.g. Fig. 11 in Darelius, 2008) and observed e.g. in the Baltic Sea (see Fig. 8 in Umlauf and Arneborg, 2009). This is coherent with erosion at the upper side and deposition/drift formation at the lower side as observed (Faguéres et al, 1999; Llave et al, 2007).

5.7. Dense shelf water cascading

Cascades are a type of buoyancy-driven current in which dense water formed by cooling, evaporation or freezing in the surface layer over the continental shelf descends the continental slope, down to a greater depth (Fig. 13). *Dense shelf water cascading (DSWC)* is an intermittent process and a component of ventilation of intermediate and deep waters; hence, it affects the generation of turbidity currents and bio-geochemical cycles (Huthnance, 1995; Shapiro et al., 2003; Legg et al., 2009). Cascades propagate along-slope and across-slope under the influence of gravity, the Earth's rotation, friction and mixing (Shapiro et al., 2003). Cascades can be found in the Southern Ocean (Baines and Condie, 1998), the Arctic shelves (Aagaard et al., 1981; Turrell et al., 1999) and the Gulf of Lion (Canals et al., 2006; Gaudin et al., 2006; Palanques et al., 2006, 2008). Dense shelf water cascading varies spatially and temporally. The dense water masses that result from it are likely to be comprise of *mesoscale individual cold/salty lenses perched on the slope* (Saunders, 1990); *dense plumes with significant 3D structure* (Shapiro and Hill, 2003), which can generated massive sand beds within submarine canyon (Gaudin et al., 2006); *downslope currents*, that pass mainly through canyons until reaching their hydrostatic equilibrium level (Canals et al., 2006; Palanques et al., 2008; Ribó et al., 2011); and *along-slope currents*, by the formation of spiral waves, meanders and eddies (Shapiro et al., 2003).

5.8. Internal waves and solitons

Internal waves are gravity waves that oscillate along the interface between two water masses of different densities (Farmer and Armi, 1999; Apel, 2000, 2004). They can be generated by any disturbance that penetrates the pycnocline (Fig. 13). Internal waves have been described off California (Emery, 1956), in the Sea of Japan (Navrotsky et al., 2004), in the Indian Ocean (Santek and Winguth, 2005) and on the Iberian Margin (Hernandez-Molina et al., 2011a). They typically have much lower frequencies (periods from several hours to days) and higher amplitudes (up to hundreds of metres) than do surface waves, because the density difference between the two water layers is lesser than that between water and air. The energy associated with internal waves is particularly high close to the continental margins (maximum horizontal velocities up to 200 cm/s and vertical velocities of 20 cm/s, Shanmugam, 2012a, 2013a), and they have been postulated as a major mechanism in the production of along-slope and across-slope processes and in the maintenance of intermediate and bottom nepheloid layers (McCave, 1986; Dickson and McCave, 1986; Cacchione et al., 2002; Puig et al., 2004), as well as in the erosion of contourite terraces (Hernández-Molina et al., 2009; Preu et al., 2013, Fig. 15).

Internal waves that correspond to periods of tides are called *internal tides* (Shepard, 1976; Garrett and Kunze, 2007; Shanmugam, 2012a,b, 2013a). Baroclinic tidal currents are commonly generated above areas of steep bathymetric variation such as the shelf break, as in the Bay of Biscay (Baines, 1982) and the Sea of Japan (Matsuyama et al., 1993). Internal *solitary waves* (or *solitons*) comprise large-amplitude, non-linear internal waves, whether single or in groups (Hyder et al., 2005), as in the equatorial Atlantic (Brandt et al., 2002) and the Bay of Bengal (Hyder et al., 2005). In the Strait of Gibraltar, the Camarinal and Spartel sills produce solitons with amplitudes of 50 to 100 m and wavelengths of 2 to 4 km (Armi and Farmer, 1988; Farmer and Armi, 1988; Brandt et al., 1996; Jackson, 2004). These solitons extend at least 200 km into the Western Mediterranean and last for more than 2 days, before decaying to background levels (Apel, 2000; Jackson, 2004). Similar observations have been made along the north-western coast of the Iberian Peninsula and around the Galicia Bank (Correia, 2003; Jackson, 2004).

5.9. Tsunami-related traction currents, rogue waves and cyclonic waves

Tsunamis comprise a wave or series of waves that have long wavelengths and long periods (Fig. 13), caused by an impulsive vertical displacement of the water by earthquakes, landslides, volcanic explosions or extra-terrestrial (meteorite) impacts (Shanmugam, 2006, 2011, 2012b). Tsunami waves carry energy through the water, but do not move the water itself, nor do they transport sediment. However, during the transformation stage, the tsunami waves erode and incorporate sediment into the incoming wave. Therefore, tsunami-related traction currents can transport large concentrations of sediment in suspension. In addition, they are important triggering mechanisms of sediment failures (Wright and Rathje, 2003).

Rogue waves and cyclonic waves have been proposed as intermittent processes similar to tsunami waves and therefore, should also be considered in the context of contourites (Shanmugam, 2012b). They can trigger bottom currents as well as submarine mudflows and slope instabilities, thereby accelerating deep-water sedimentation. However, it is not possible to differentiate between deposits generated by tsunamis and those generated by cyclonic waves (Shanmugam, 2011).

6. Depositional and erosional features

Persistent bottom-current systems and associated oceanographic processes (see Section 5) strongly affect the seafloor, ultimately conferring it with pervasive erosional and depositional features (Fig. 15). These features can be isolated, but when alongslope processes dominate, are more likely to be form part of a *Contourite Depositional System* (*CDS*), which is an association of various drifts and related erosional features (Hernández-Molina et al., 2003, 2008a, 2009). Similarly, distinct but connected CDS within the same water mass can be considered to be a *Contourite Depositional Complex* (*CDC*) (Hernández-Molina et al., 2008a). However, contourites also occur interbedded with other deep-water facies types, and do not necessarily form individual sedimentary bodies. The erosional and depositional features produced by bottom currents are found at various scales: they range from small bedforms to large sediment drifts.

6.1. Large depositional and erosional features

Bottom currents are known to construct large accumulations of sediments, known as *contourite drifts*. Over the past 50 years, researchers have studied numerous drifts (Fig. 9) using a combination of techniques, finding that drifts vary greatly in location, morphology, size, sediment patterns, construction mechanisms and controls (Faugères and Stow, 2008).

Contourite drifts are most easily recognised when they have an along-slope, elongated mounded shape, and an adjacent concave moat (see Erosional Features, below). They can be more than 100 km wide, several (up to tens) hundreds of kilometres long, up to 2 km thick and have a relief of up to 1.5 km. Similarly to deep-water down-slope turbidity deposits, they range in dimensions from ca. 100 km^2 (small patch drifts, equivalent in size to isolated turbidite lobes) to >100,000,000 km^2 (giant elongated drifts matching the size of the largest deep-sea fans).

A three-tiered classification of contourite drifts, based on drift depth, was proposed over a decade ago by Viana et al. (1998a) and by Stow

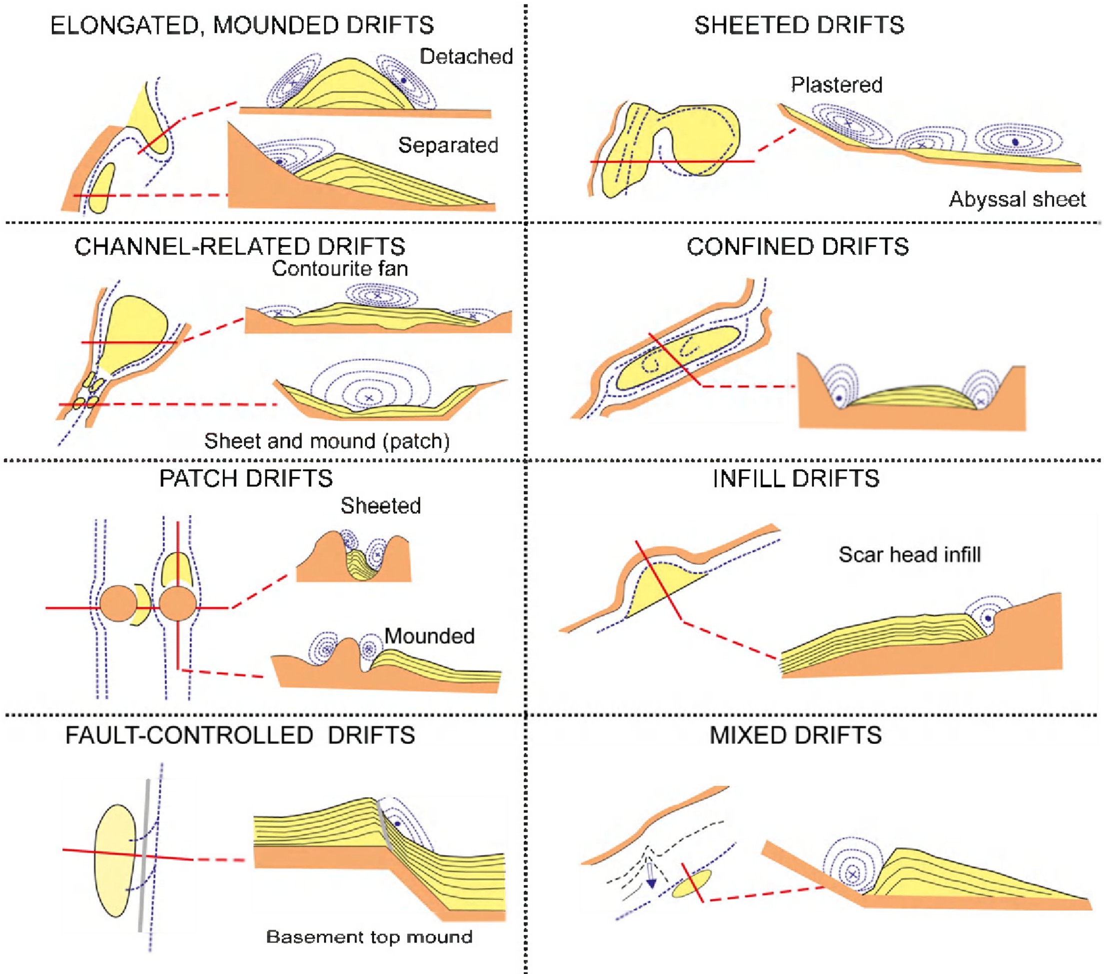


Fig. 16. Sediment drift types and inferred bottom-current paths.
Modified from work by Rebesco (2005), and by Hernández-Molina et al. (2008b), with permission from Elsevier. The original classification was adapted from Rebesco and Stow (2001) and Stow et al. (2002a).

et al. (2002a). However, this system now appears irrelevant, especially for non-recent drifts, whose palaeodepth is often unknown (Rebesco, 2005). Other classification systems, based on drift morphology or location, have also been conceived (McCave and Tucholke, 1986; Faugères et al., 1993, 1999; Rebesco and Stow, 2001; Rebesco, 2005; Faugères and Stow, 2008). Pragmatically speaking, there is some overlap amongst different drift types, such that they are actually within a continuous spectrum of deposits.

All contourite drifts are characterised by a variable degree of mounding and somewhat evident elongation (Fig. 16). The largest elongated mounded drifts (*giant elongated mounded drifts*) are generally found on the lower slope and can be divided into two types: *separated* and *detached*. *Separated drifts* are most associated with steeper slopes, from which they are separated by a distinct erosional/non-depositional moat (Fig. 10B) along which the flow is focused (e.g. North Iberian margin, Van Rooij et al., 2010). *Detached drifts* typically present an elongation that deviates from the adjacent slope against which it first began to form. Such a drift development can result from a change in the margin's trend (e.g. Eirik Drift, Fig. 10A, Hunter et al., 2007a,b). *Sheeted drifts*, most commonly found on abyssal plains, are characterised by a broad, faintly mounded geometry, with very slight thinning towards the margins (e.g. Gulf of Cádiz; see Llave et al., 2001, 2007; Hernández-Molina et al., 2008b). They show a fairly uniform thickness and a predominantly aggradational stacking pattern. *Plastered drifts* are generally more subdued and smaller than giant, elongated mounded drifts (Fig. 15A and B), but are more mounded and located in shallower positions than are sheeted drifts (e.g. Preu et al., 2013; Rebesco et al., 2013). Given their location along a gentle slope swept by relatively low-velocity currents, in the classification of Fig. 16 they are included along with the sheeted drifts, but in other classifications (e.g. Faugères and Stow, 2008) they are considered along with the giant drifts. Some plastered drifts can actually be considered as sheeted drifts, whereas others must be considered as mounded, elongated drifts; regardless, there is a continuity of examples in between these two end members. *Channel-related drifts* lie in gateways in which currents are constrained and flow velocities are higher (e.g. Antarctica, Maldonado et al., 2005 or Vena Channel, Brazil, Mézerais et al., 1993). *Confined drifts* are mounded, with distinct moats along both flanks, and elongated parallel to the axis of a relatively small confining basin (e.g. Lake Baikal, Ceramicola et al., 2001). *Patch drifts* are small, elongated-to-irregular drifts characterised by a random (patchy) distribution controlled by the interaction between bottom currents and irregular seafloor morphology (Hernández-Molina et al., 2006b). *Infill drifts* typically form at the head of a scar and are characterised by a moderate relief and extension, as well as a mounded geometry that progressively infills the topographic depression (Laberg et al., 2001). *Fault-controlled drifts*, characterised by the influence of faulting in their development, develop either at the base or at the top of a fault-generated basement relief in response to perturbations in the bottom-current flow pattern (Rebesco, 2005). *Mixed drifts* are those that involve the significant interaction of

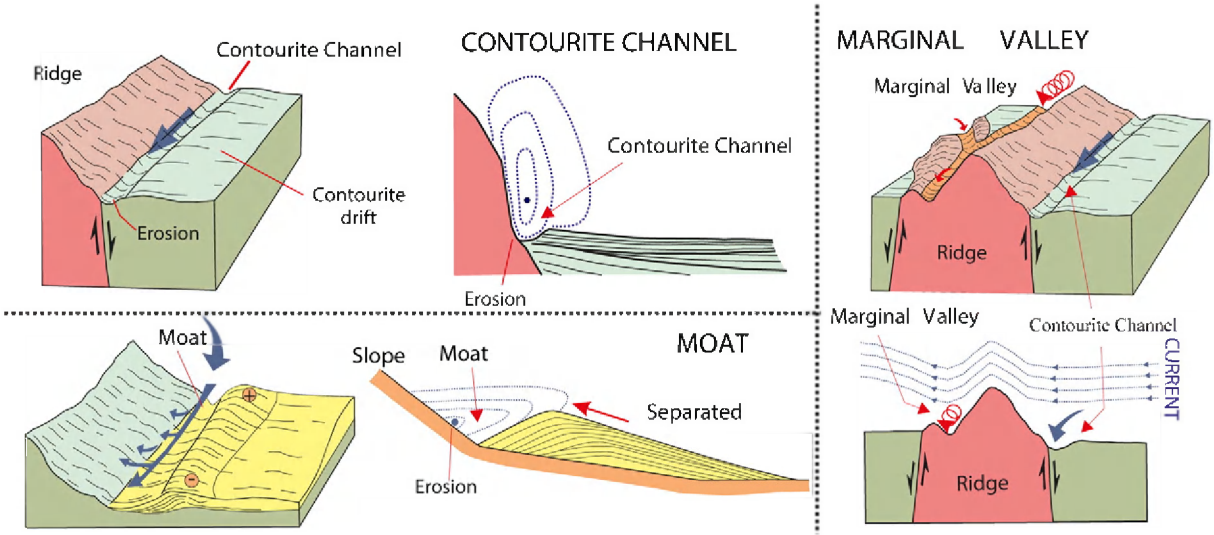


Fig. 17. Main characteristics of areal and linear, large-scale contourite erosional features.
Modified from work by Hernández-Molina et al. (2008b) and by García et al. (2009), with permission from Elsevier.

along-slope contour currents with other depositional processes in the building of the drift body, as in the Antarctic Peninsula (Camerlenghi et al., 1997b; Giorgetti et al., 2003; Hillenbrand et al., 2008), the Argentine slope (Hernández-Molina et al., 2009) and the Gulf of Cádiz (Llave et al., 2007). The many interrelated and overlapping factors that control drift morphology (Faugères et al., 1993; Rebesco, 2005; Ercilla et al., 2008; Faugères and Stow, 2008; Ercilla et al., 2011) include physiographic and tectonic settings, current regime, sediment input, interacting processes, and changes in climate and in sea level, as well as the length of time that these processes have operated.

Large-scale erosional features are also common in CDS, although they have not been as well studied as have depositional ones. They typically occur in association with contourite drifts (Faugères et al., 1999; Stow and Mayall, 2000), but can also be found in a broad area of continental slopes (Viana, 2001; Hernández-Molina et al., 2003, 2006a, 2009; Ercilla et al., 2011). The authors of this present review propose a preliminary reconsideration (Fig. 17) of the only systematic classification of large-scale erosional features that has been attempted to date (Hernández-Molina et al., 2008b; García et al., 2009). Two main types of large-scale erosions are considered: *areal* and *linear*.

Areal erosional features are further subdivided into two types: *terraces* and *abraded surfaces*. *Terraces* are broad, low-gradient, slightly seaward-dipping, along-slope surfaces produced by erosional as well as depositional processes (Fig. 15B). They are generally found on the upper and middle slopes relative to the position of the interfaces between different water masses, but could be identified at any depth over continental slopes (e.g. Viana, 2001; Viana et al., 2002a,b; Hernández-Molina et al., 2009, 2014; Brackenridge et al., 2011; Preu et al., 2013). *Abraded surfaces* are localised areas eroded by strong tabular currents. They are often found in association with scours, sediment waves, dunes and sand banks (Hernandez-Molina et al., 2011a; Ercilla et al., 2011; Sweeney et al., 2012).

Large, linear erosional features have been further subdivided by Hernández-Molina et al. (2008b) and García et al. (2009) into three types: *contourite channels*, *moats*, and *marginal valleys*. *Contourite channels* are elongate erosional depressions formed mainly by the action of bottom currents. They are characterised by the presence of truncated reflections and can be along-slope trending, or sinuous and oblique relative to the slope. Moats are channels parallel to the slope and originated by non-deposition and localised erosion beneath the core of the bottom current, accentuated by the Coriolis force. Hernández-Molina et al. (2008b) suggest that the term *moat* be used only for those features that have a genetic relationship with giant, elongated, mounded contourite drifts of separated type (Fig. 10B and 15B and C); *Marginal valleys* (or *scours*) are, according to the aforementioned authors, those elongated erosional channels that are generated by the effects of a bottom current impinging against and around topographic obstacles (e.g. seamounts, diapiric ridges, and mud volcanoes). Furrows are set apart in this contribution and included within bedforms (see Section 6.2), as they are much narrower and less incised than are contourite channels. However, since in exceptional cases they can reach lengths of up to a few tens of kilometres, they should be mentioned together with the large-scale erosional features. Their origin has been associated to small, detached filaments of flow separated from the main bottom current (possibly as a result of topographic effects).

These distinctions, developed mainly from observations in the Gulf of Cádiz, Antarctica and Argentine basins, can likely be identified in many other margins. Nonetheless, more detailed knowledge on erosional features and associated oceanographic processes is required. Many other areas have yet to be analysed to improve this preliminary classification, as well as to clarify the genetic spatial and vertical relationships between the erosional features and the adjacent depositional features within a CDS.

6.2. Bedforms

Various depositional and erosional bedforms are generated by bottom currents. These bedforms can occur in a wide range of deep-water environments, but are often found in association with contourite drifts or with large scale erosions in gateways, channels or adjacent to seafloor obstacles (Stow et al., 2013a). They are highly variable in terms of sediment composition, morphology and dimension (Wynn and Masson, 2008), with the latter ranging from decimetres (detected with bottom photographs) to kilometres (detected with seafloor imaging and other high-resolution geophysical tools). The detection of bedforms can be important for the reconstruction of bottom-current velocity (Stow et al., 2009, Fig. 18) and for geohazard assessment (where bedforms are indicative of velocities higher than 1 m/s, which can damage seafloor infrastructure, including pipelines and telecommunication cables).

Bottom-current bedforms can be divided into two types, based on their spatial relationship to the flow: *longitudinal*, which are elongated

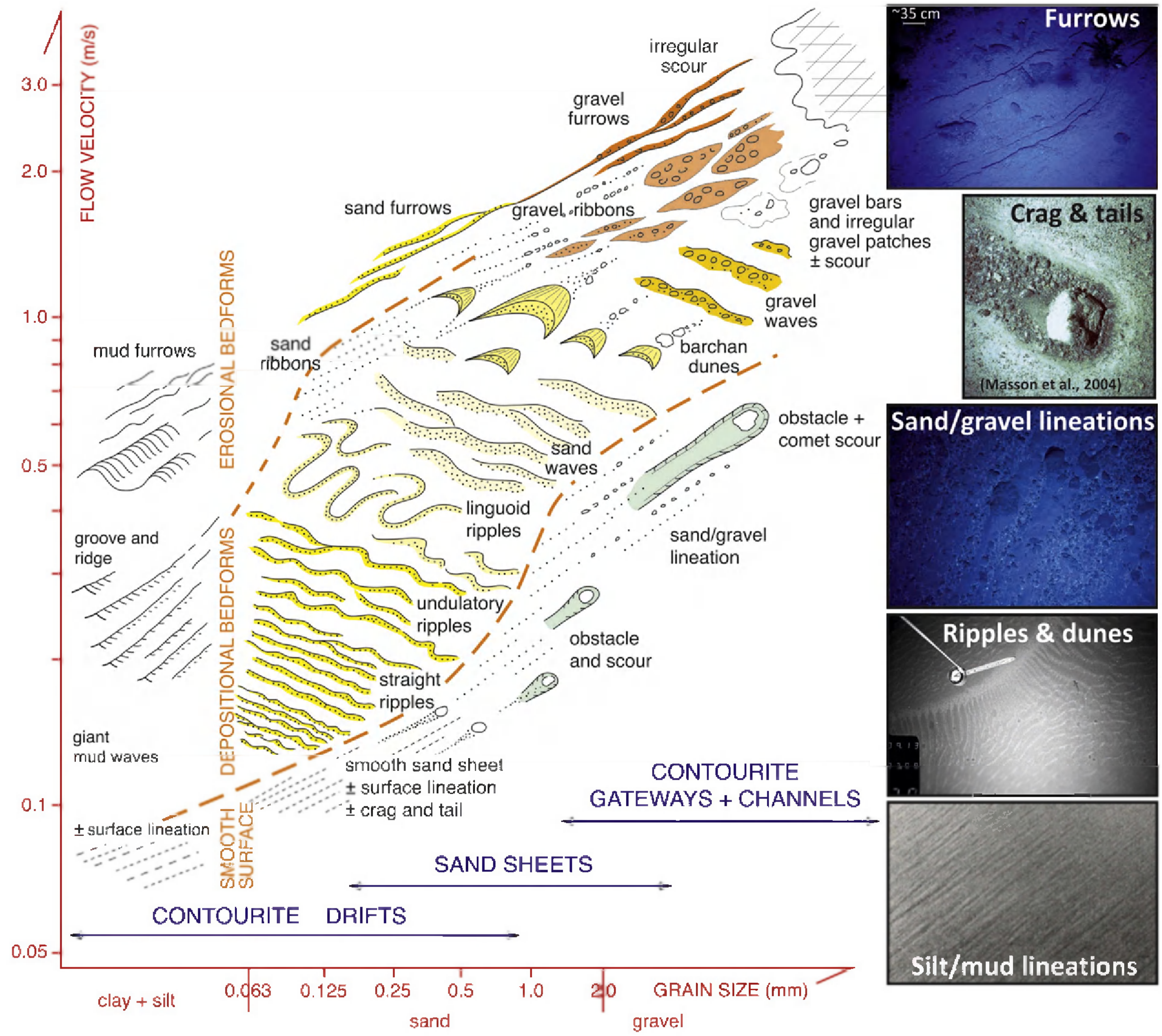


Fig. 18. Bedform-velocity matrix for deep-water bottom current systems, showing a schematic representation of bedforms in function of mean grain size of sediment versus flow velocity at or near the seafloor.
Modified from work by Stow et al. (2009), with permission from the Geological Society of America.

parallel to the flow and are essentially erosional; and *transverse*, which are mostly depositional. However, these longitudinal and transverse bedforms can both be related to the velocity range of the bottom current, in function of the mean grain size of sediments (Stow et al., 2009, 2013a, Fig. 18).

6.2.1. Longitudinal bedforms

Low-relief, sub-parallel surface lineations range from millimetre-spaced silt streaks to decimetre-spaced gravel stringers (e.g. Hollister and McCave, 1984). *Groove and ridge* is used to describe a cohesive, muddy substrate that shows decimetre-to-metre-spaced, distinctly erosive grooves between remnant or depositional ridges (also called *longitudinal triangular ripples*; e.g. Flood, 1981). *Crag and tail* refers to the centimetre-to-decimetre-sized, elongated depositional mound (the *tail*) downstream of an obstacle (the *crag*) (Heezen and Hollister, 1964). At higher flow velocities, these can be substituted with *comet scours*, metre-to-hectometre-long, crescentic-to-elongate, erosional scour marks that occur around, and extend downstream from, an obstacle (Masson et al., 2004). Without an associated obstacle, erosional scour crescents, irregular pluck marks, and/or or tool marks can be found. *Ribbon marks* are elongated mounded filaments of sand that are up to 500 m wide and several kilometres long, often merge into or diverge from broad sand sheets, and are produced by deposition following erosion and winnowing (Viana et al., 2007). *Erosional furrows* are elongate sub-parallel lineations that are somewhat regularly spaced, typically a few kilometres in length, a few tens of metres wide, and a few tens of centimetres deep. In some cases they are cut into coarse gravel and sand substrates (Masson et al., 2004; Stow et al., 2013a) or into fine-grained cohesive sediments (Flood, 1983). Kilometre-scale sub-circular-to-oval scour hollows, or elongate-to-irregular erosional scours, are attributed to vertical spouts of water and to catastrophic, high-energy, bottom-current flow (Bulat and Long, 2001; Holmes et al., 2003; Stoker et al., 2003).

6.2.2. Transverse bedforms

Transverse bedforms exist in different sizes and shapes. The smallest ones are *ripples*, which have wavelengths of a few decimetres, heights of a few centimetres, and exhibit straight, sinuous, and linguoid trends (Stow, 2005). Larger transverse bedforms are *dunes*, which can be sinuous-crested and barchanoid in planform

(Wynn et al., 2002), and *sand waves*, generally flatter and sinuous-crested (Kuijpers et al., 1993). They range from tens to hundreds of metres in wavelength, and from a few decimetres to a few metres in height. *Sediment waves* (or *mud waves*) belong to a continuum of bottom-current features that fall somewhere between ripples and contourite drifts. They can have wavelengths ranging from 0.5 to 10 km in length, heights usually up to 50 m (occasionally, up to 150 m), and wave crests that are often longer than 10 km (Wynn and Stow, 2002; Stow et al., 2013a).

7. Contourite types and facies models

Specific bottom-current facies have been described by many authors (e.g. Stow and Lovell, 1979; Stow and Holbrook, 1984; Stow and Piper, 1984; Pickering et al., 1989; Faugères and Stow, 1993; Gao et al., 1998; Faugères et al., 1999; Stow et al., 2002a; Rebesco, 2005; Llave et al., 2006; Øvrebø et al., 2006; Shanmugam, 2006; Stow and Faugères, 2008; Shanmugam, 2012a, 2013b; Stow et al., 2013b). Some of them are overviewed below.

7.1. Small-scale characteristics

The characterisation of contourite facies is based primarily on their small-scale characteristics, as identified on descriptions of cores and/or outcrops.

7.1.1. Lithology

Contourites vary widely in their lithology (Stow et al., 1996, 1998a,b, 2002b; Shanmugam, 2006; Stow and Faugères, 2008; Shanmugam, 2013b), exhibiting different grain sizes (from clay to gravel) and

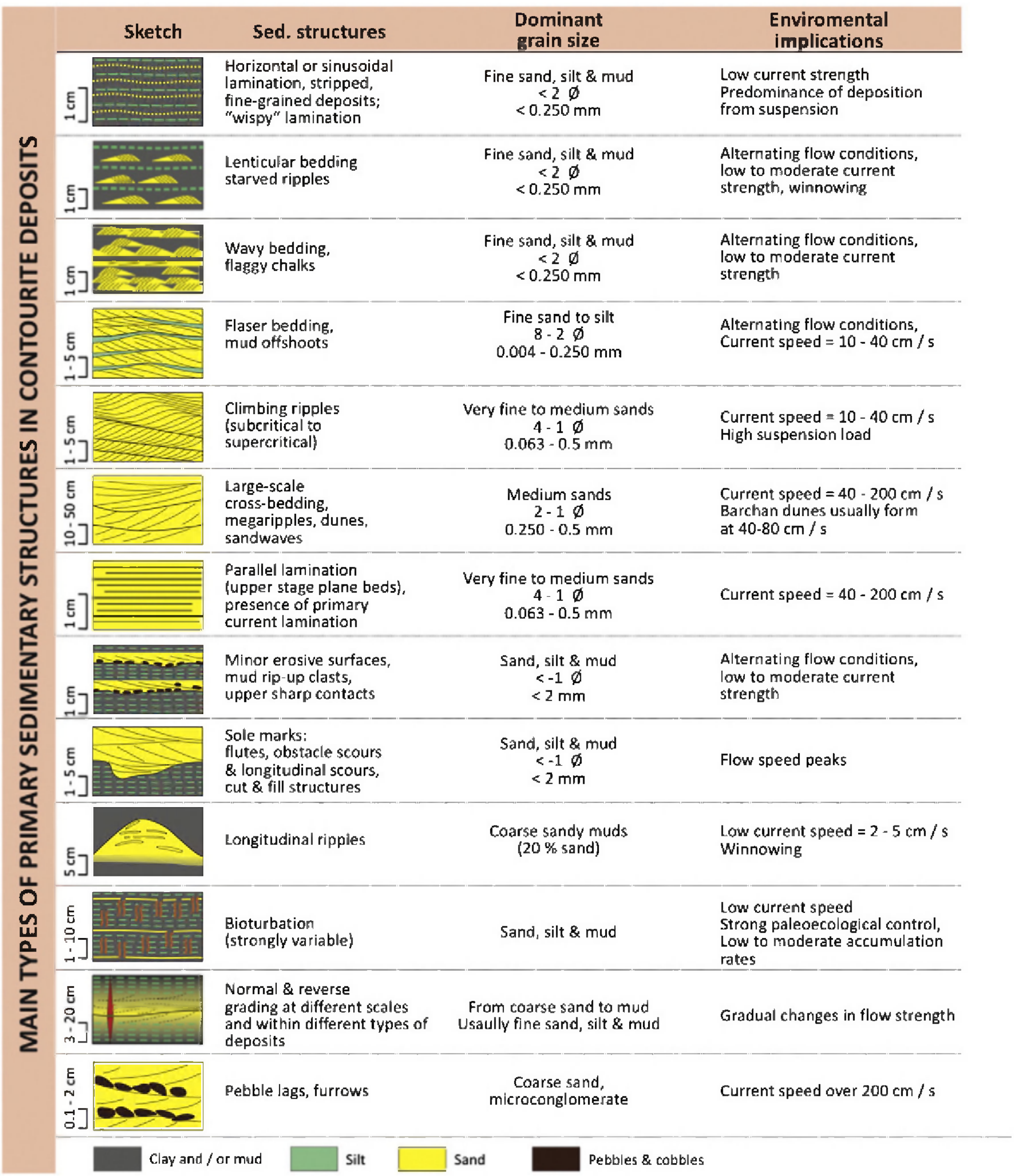


Fig. 19. Main types of sedimentary structures in contourite deposits. From Martín-Chivelet et al. (2008); with permission from Elsevier.

composition (terrigenous, biogenic, volcanic and/or mixed). Although some contourites exhibit a rather homogeneous composition (e.g. *mid-ocean drifts*, which are more than 90% pelagic biogenic material; and *high-latitude drifts*, which are more than 90% glaciomarine hemipelagic material), most bottom-current deposits show a characteristically mixed composition (Stow et al., 2008). Within the present-day ocean basins, they usually contain a mixture of biogenic, terrigenous, volcanoclastic and authigenic components (Gao et al., 1998). Compared to adjacent pelagic or hemipelagic sediments, they can have identical composition but differ in their texture and fabric. The most common contouritic deposit is a rather poorly sorted, mud-rich (between 5 and 40 μm) facies, which is intensively bioturbated, intercalated by thinner horizons of fine-grained sands and silt, and typically shows a somewhat rhythmic bedding (Stow et al., 2002b). Sandy contourite deposits are less common, albeit more commonly than initially thought, and have been reported on the Brazilian margin (Viana et al., 2002a), in the Gulf of Mexico (Shanmugam, 2006, 2012a, 2013b) and in the Gulf of Cádiz (Hanquiez et al., 2007; Hernández-Molina et al., 2012; Brackenridge et al., 2013; Stow et al., 2013a; Hernández-Molina et al., 2014). Also, reworked sandy turbidites have been reported in different areas (Stanley, 1993; Shanmugam, 2006; Gong et al., 2012; Shanmugam, 2012a, 2013b). At high latitudes, contourites can contain much gravel-sized material, brought in as ice-rafted debris. Moreover, gravel-lag contourites are found near oceanic gateways, shallow straits and moats in all latitudes.

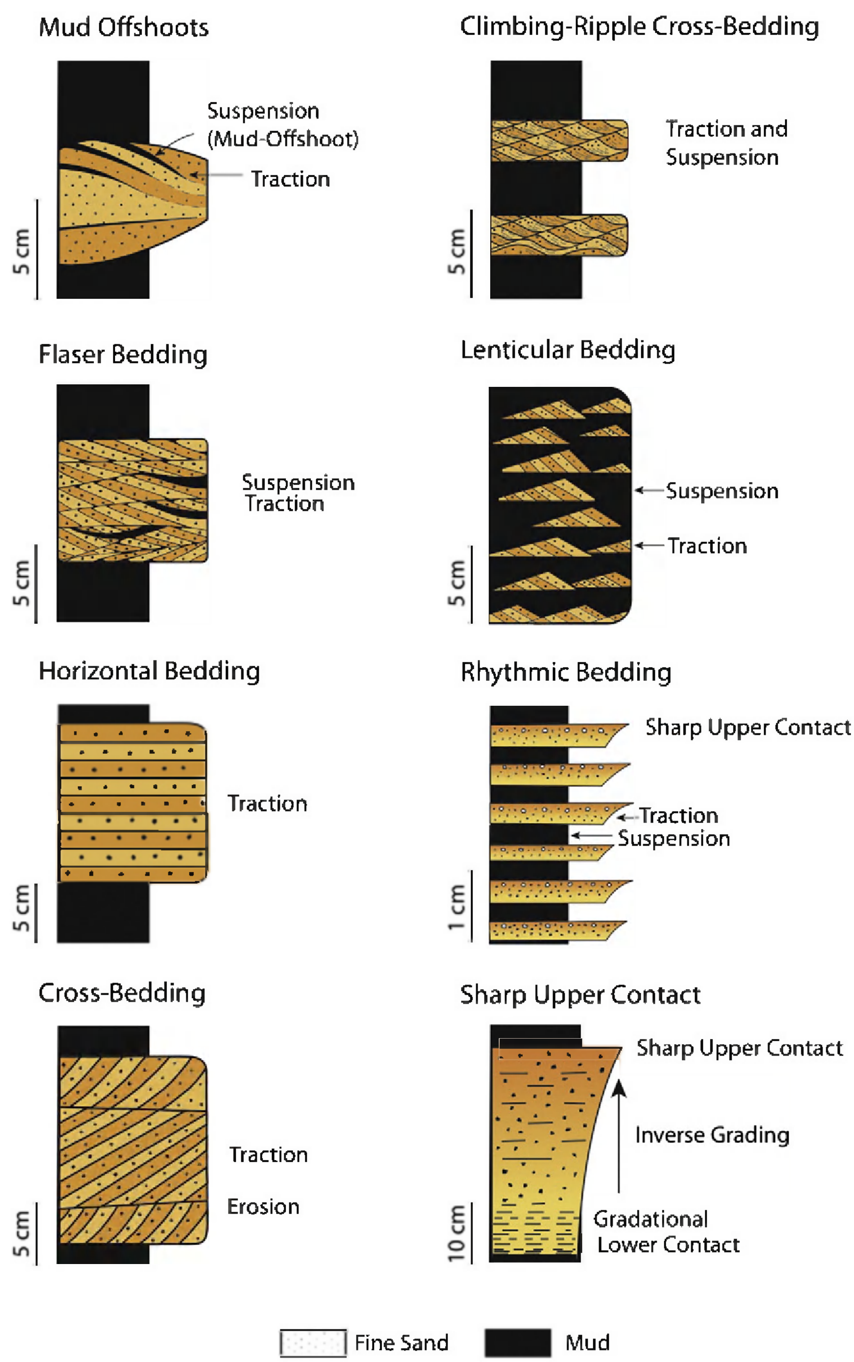


Fig. 20. Traction features interpreted as an indication of bottom-current reworked sands (BCRS).
Modified from work by Shanmugam et al. (1993b) and by Shanmugam (2008), with permission from Elsevier.

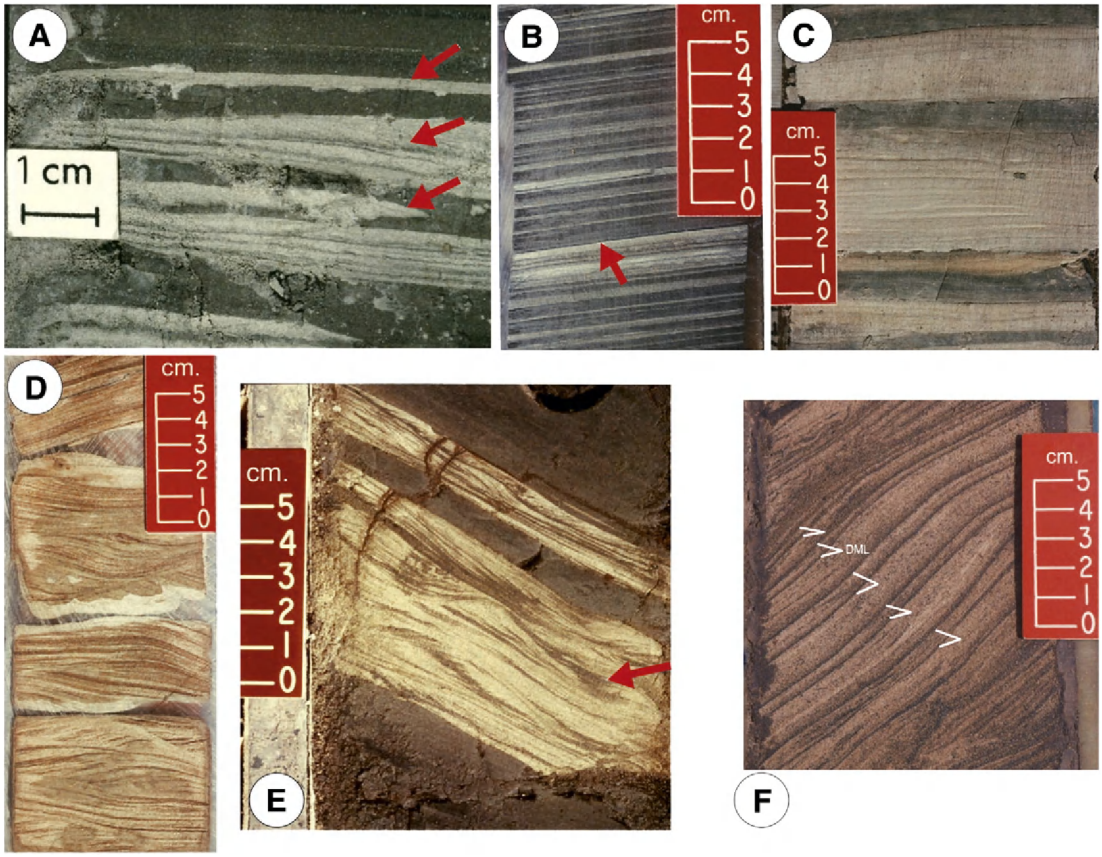


Fig. 21. Core photographs of representative sedimentary structures related to bottom-current reworked sands (BCRS): A) Discrete thin sand layers with sharp upper contacts (top red arrow). Traction structures include horizontal laminae, low-angle cross-laminae, and starved ripples (Middle Pleistocene, Gulf of Mexico) (according to Shanmugam et al., 1993b; reprinted with permission of the AAPG, whose permission would be required for further use); B) Rhythmic layers of sand and mud, inverse grading, and sharp upper contacts of sand layers (red arrow) (Palaeocene, North Sea) (Shanmugam, 2008; with permission from Elsevier); C) Horizontal lamination with gradational upper contact. Note the faint normal grading (Pliocene sand. Edop Field, offshore Nigeria) (Shanmugam, 2012a; with permission from Elsevier); D) Convex-up laminae (hummocky cross-stratification [HCS]?) and concave-up laminae (wave ripples) in fine-grained sand (Eocene, North Sea) (Shanmugam, 2012a; with permission from Elsevier); E) Flaser bedding. Note the presence of mud in ripple troughs. Core photograph showing double mud layers (arrow) (Pliocene sand. Edop Field) (Shanmugam et al., 1993b; with permission from AAPG); and F) Double mud layers (white arrow) (Pliocene sand. Edop Field) (Shanmugam, 2003, 2012a; with permission from Elsevier).

7.1.2. Sedimentary structures

Various sedimentary structures have been described for contourites in present and ancient deposits (Martín-Chivelet et al., 2003; Shanmugam, 2006; Martín-Chivelet et al., 2008; Stow et al., 2008; Shanmugam, 2012a). However, in areas of intense bioturbation from benthic activity, the preservation potential of some of these structures can be low. The small-to-medium-scale sedimentary structures described for contourite deposits comprise (Figs. 19–21) *ripples and cross-laminations* (mainly in fine sandy deposits); *flaser and lenticular beddings* (fine sands, silt and clays); *horizontal, sub-horizontal and sinusoidal laminations*; *parallel laminations*; *large scale cross-stratification*; *erosive scarps* (and associated structures); *gravel lags*; *grading*; *symmetric ripples*; and *longitudinal triangular ripples* (Martín-Chivelet et al., 2003, 2008; Stow and Faugères, 2008; Stow et al., 2009). Most of these structures are also present in other deep-water deposits (e.g. turbidites), but some have been suggested to be a clear diagnostic feature for bottom-current deposits, such as (Figs. 19 and 21): *negative grading* (Shanmugam et al., 1993a,b; Shanmugam, 2012a, 2013b); *longitudinal triangular ripples* (Heezen and Hollister, 1964; Flood, 1981; Tucholke, 1982; McCave et al., 1984); and *double mud layers and sigmoidal cross-bedding*, which are unique to deep-water tidal deposits in submarine canyons (Shanmugam, 2006, 2012a). Sedimentary structures and grain size can facilitate decoding of the bottom current intensity (e.g. Wynn et al., 2002; Stow et al., 2009). A bedform-velocity matrix for deep-water bottom-currents has been proposed by Stow et al. (2009). It might be practical for local studies, although if it is applied, then certain other aspects must also be considered (Shanmugam, 2012a).

7.1.3. Biogenic structures

The activity of benthic organisms is common in contourite deposits, and effective for destroying the original sediment fabric and structures. The dominant processes and structures are bioturbation (mottling) and organic traces, respectively (Wetzel et al., 2008). Contourites have been with very low (Shanmugam et al., 1993b), medium (Martín-Chivelet et al., 2003) and very high (Faugères and Stow, 1993) bioturbation that has been described. Ichnofacies associations for the *fodinichnia* (feeding traces), *pascichnia* (shepherding traces) and *dominichnia* (living traces) types have been described, defining a continuum within sub-to-well-oxygenated conditions (Ekdale and Mason, 1988).

7.1.4. Palaeontological content

Frequently, the palaeontological content of contourite deposits is rather similar to that of the surrounding pelagic and/or hemipelagic deposits. It usually comprises planktonic and benthic foraminifera, ostracods, nannoplankton, etc. and sometimes contains parts of molluscs, echinoderms and brachiopods, although reworked contourites also contain shallow-water fauna and flora (Martín-Chivelet et al., 2003, 2008).

7.2. Classifications for contourite deposits

Initially, contourite deposits were classified exclusively according to their lithology and texture (Stow and Lovell, 1979; Gonthier et al., 1984; Stow and Holbrook, 1984), which led to the definition of four types of facies: *clays*, *mottled silts*, *sand*, and *gravel lags*. Other, complementary classification systems were later proposed (Stow and Lovell, 1979;

CLASSIFICATION FOR CONTOURITE DEPOSITS	Type of deposits	Grain size	Characteristics	Sed. Structures	Examples
CLASTIC CONTOURITES	Muddy contourites	5 - 11 Ø < 0.31 mm	50 % Clays < 15 % of sands ≤20 -30 % of bioclastic and / or carbonate components Homogeneous & highly bioturbated Poor sorting	Rare laminations Bioturbatiion Indistinct mottled appearance	Rockall Trough (Stow & Faugères, 2008)
CLASTIC CONTOURITES	Silty contourites	4 - 8 Ø 0.063-0.004 mm	40 - 60 % Silts Interbedded between muddy & sandy contourites Poor sorting Sharp to irregular tops and bases	Tractive structures (ripples, sed.waves) Bioturbational mottling Ichonofacies	Gulf of Cadiz (Stow & Faugueres, 2008)
CLASTIC CONTOURITES	Sandy contourites	-1 - 4 Ø 2 - 0.063 mm Normally does not exceed fine sands	Sheeted to wedge beeding Well-sorted deposits, but can be poor to moderate Mixed siliciclastic / biogenic composition Heavy mineral concentration Both positive & negative grading Gradational or erosive contacts	Tractive structures (e.g.: horizontal lamination, cross-lamination, ripples, etc) Bioturbation (sub.vertical burrows) Massive layers (structureless) Erosional or gradational contacts	Gulf of Cadiz (Stow & Faugueres, 2008)
CLASTIC CONTOURITES	*Bottom current reworked sands (BCRS)*		From previous turbiditic deposits Rhytmic layers Lenticular bedding Well sorted Coarsening upward sequences Sharp to gradational bottom contact and sharp (nonerosiona) upper contactl	Tractive sturctures (e.g.: horizontal lamination, low-angle cross lamination; mud-offshoots in ripples, mud-drapes, flaser, etc)	Gulf of Mexico (Shanmugam, 2006; 2012)
CLASTIC CONTOURITES	Gravel contourites	< -1 Ø > 2 mm	Winnowing & erosion (channels, moats, etc) Irregular layer and lenses Poorly to very poorly sorted	Sandy gravel lag	Faeroe-Shetland Channel (Akhurst et al., 2002)
VOLCANOCLASTIC CONTOURITES		Mud, silt or sands	Similar to the siliciclastic facies Composition is dominated by volcaniclastic material	Similar to the clastic contourites	Hawaii (Puga-bernabéu, A., Pers. Comm.)
SHALE-CLASTS OR SHALE-CHIP LAYERS		Shale clasts generally mm in size	Developed in muddy & sandy contourite facies From substrate erosion by strong bottom currents Burrowing on the nondeposition surface	Clats axes sub-parallel both to beeding and to the current direction	Brazilian Basin (Faugueres et al., 2002)
CALCAREOUS CONTOURITES	Calcareous muddy & silty contourites	> 4 Ø < 0.063 mm Silty clay to clavey silts	>70 % of bioclastic and / or carbonate components Dominant biogenic input Poorly sorted Distint sand-size fractions (biogenic particles) Composition: pelagic to hemipelagic, including nannofossils & foraminifers as dominat elements Admixture of siliciclastic or volcanicclastic material	Bedding is indistint, but may be enhanced by cyclic variations in composition / grain size. Bioturbation	Ortegal Spur (Belgica GENESIS Cruise)
CALCAREOUS CONTOURITES	Calcareous sandy contourites	-1 - 4 Ø 2 - 0.063 mm Sands	Equivalent of sandy contourites Both well-sorted to poorly sorted Particles from pelagic, benthic, off-shelf & off-reef sources Admixture of siliciclastic, volcanic & silicieous material	Thin-bedded cross-laminated foraminera contourite. Lenticularity Hardgrounds.non de`positional surfaces Bioturbation & burrowing	
CALCAREOUS CONTOURITES	Calcareous gravel lag contourites	< -1 Ø > 2 mm Gravel	Clasts or chips derived from erosion of the subtrate		Gulf of Cadiz (F.J. Sierro, Pers. Comm.)
SILICEOUS BIOCLASTIC CONTOURITES		Mud, silt or sands	Rich in diatomaceous & radiolarian material	Laminated and / or cross-laminated sands	Blake-Bahama Basin (DSDP, Site 534, Sheridan et al 1983)
CHEMOGENIC CONTOURITES	Manganiferous contourites		Manganiferous or ferro-manganiferous horizonts Areas with ferro-manganese nodules & pavements	Bioturbation & burrowing	Brazilian Basin (Gonthier et al., 2003)
CHEMOGENIC CONTOURITES	Chemogenic gravel-lag contourites		Deep-water chenoherms (chemical - biogenic precipitates) of metal - carbonate chimneys, mounds & encrustations Winnowed and aligned into chemogenic gravel-lags	Strewn od debris Alineation of gravel-lag	Bellingshausen Basin (Heezen & Hollister, 1971)

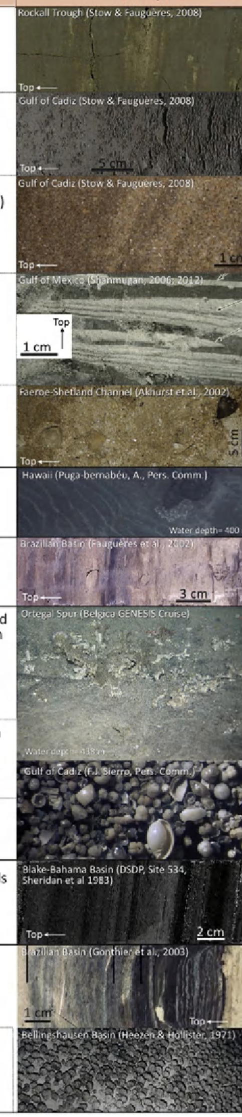


Fig. 22. Classification of contourite deposits. Types of deposits, grain size, general characteristics, common sedimentary structures, and a representative example of each type are included. Bottom current reworked sands (BCRS), as defined by Shanmugam (2006, 2012a), within sandy contourites are also considered here.
Adapted from Stow and Faugères (2008).

Stow, 1982; Faugères and Stow, 1993; Stow et al., 1996; Gao et al., 1998; Stow and Faugères, 2008). The most recent integrated classification for contourites, proposed by several authors and summarised in Stow and Faugères (2008), is briefly compiled and described here (Fig. 22), taking into consideration certain ideas from other authors (e.g. Shanmugam, 2006, 2012a,b, 2013b).

7.2.1. Clastic contourites

7.2.1.1. Muddy contourites. These are homogeneous and highly bioturbated deposits, often with an indistinct mottled appearance, and can also show distinct burrows of varied ichnofacies. They are typically >50% clay, have an average grain size ranging from 5 to 11 Φ, poor sorting (typically 1.4 to 2 Φ) and <15% of sand. Bioclastic and/or carbonate components can be present, generally at a maximum of 20 to 30%; these include planktonic and benthic calcareous and siliceous organisms that are frequently broken and are impregnated with iron oxides. On rare occasions, muddy contourites can exhibit a primary lamination, which is often characterised by a colour change or by irregular winnowed concentrations of coarser material. The composition is dominantly siliciclastic. These deposits occur in thick units and are rather difficult to differentiate from hemipelagic deposits. The components are partly local (they include a pelagic contribution) and are partly far-travelled.

7.2.1.2. Silty contourites. Although these deposits are very similar to muddy contourites, they represent a transitional point between the latter and sandy contourites, and are commonly interbedded between these types. Silt is the dominant grain size (between 40 and 60%) and certain traction sedimentary structures, as well as bioturbational mottling and ichnofacies, are frequently present. The range of grain sizes (3 to 11 Φ) can be still wider than for muddy contourites; thus, the sorting can be very poor (>2 Φ) and there can be some evidence of indistinct discontinuous lamination (partly destroyed by bioturbation). This typically comprises silty layers with sharp to irregular tops and bases, together with thin lenses of coarser material.

7.2.1.3. Sandy contourites. These contourites, in which sand-sized components predominate, are characterised by well-sorted deposits with lamination and/or tabular-to-wedge bedding, from a few centimetres to several metres in thickness. The mean grain size does not normally

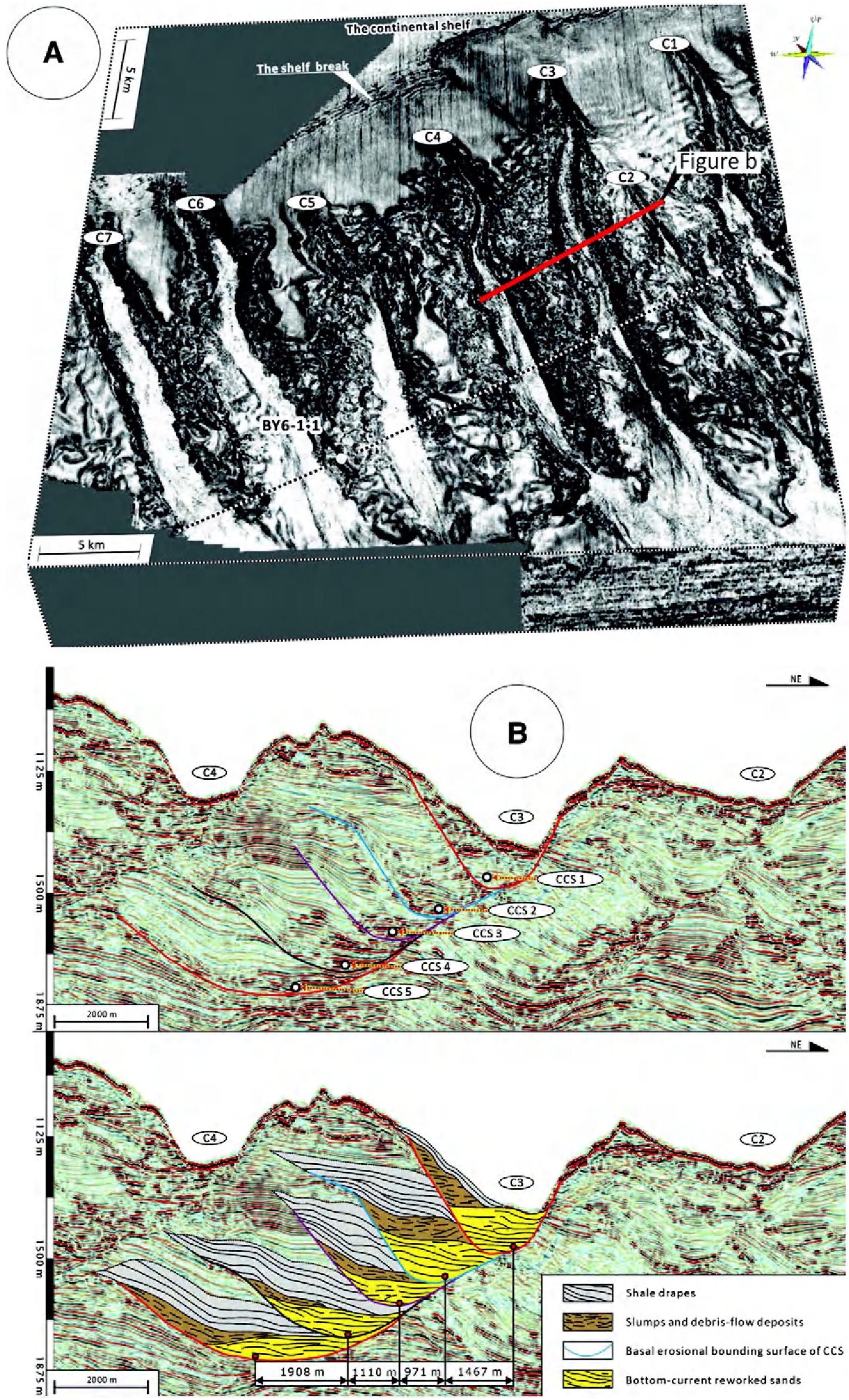


Fig. 23. A) Three-dimensional (3D) coherence volume showing unidirectionally migrating deep-water channels (C1 to C7). B) Facies and architecture within unidirectionally migrating deep-water channel 3 (C3). Five channel-complex sets (CCS1 to CCS5), are identified, each of which comprises bottom-current reworked sands (BCRS) in the lower part, grading upward into slumps and debris-flow deposits and, finally, into shale drapes. The BCRS are represented by subparallel and high-amplitude reflections with external lens shapes and are systematically nested in the direction of channel migration (Gong et al., 2013; with permission from the AAPG).

exceed that of fine sand (apart from coarser-grained horizons and lags), and sorting is mostly poor to moderate (0.8 to 2 Φ), which is partly due to bioturbational mixing. Both positive grading and negative grading can be present. The sediment has a mixed siliciclastic/biogenic composition, with evidence of abrasion, fragmented bioclasts and iron-oxide staining. In sandy contourites, traction sedimentary structures (e.g. horizontal structures and cross-laminations) become dominant. Nevertheless, they can be bioturbated throughout, with common sub-vertical burrows, and can appear as massive (structureless) at first sight. The layering can have gradational or erosive contacts.

7.2.1.4. Bottom-current reworked sands. Shanmugam (2006, 2012a, 2013b) emphasised the consideration of bottom-current reworked sands (BCRS) from previous turbiditic deposits as a pragmatic alternative to the application of conventional turbidite concepts, and as a new concept for understanding the origin and predicting the distribution of deep-water sandstones. Several authors have determined the general characteristics of BCRS: rhythmic layers of fine-grained sands and silts, in which the sand is well sorted (Hubert, 1964; Hollister, 1967; Hollister and Heezen, 1972; Bouma and Hollister, 1973; Unrug, 1977; Stow and Lovell, 1979; Lovell and Stow, 1981; Shanmugam, 1990, 2000; Ito, 2002; Gong et al., 2013; Stow et al., 2013b). BCRS frequently contain different seismic facies (Fig. 23; see Gong et al., 2013) and sedimentary structures (Fig. 21; see Shanmugam, 2012a, 2013b), such as horizontal laminations, low-angle cross laminations, lenticular beddings, mud-offshoots in ripples, mud drapes, flaser beddings and various traction structures. The sand usually coarsens upwards at different scales and frequently exhibits a sharp-to-gradational bottom contact and a sharp (non-erosional) upper contact. Shanmugam (2006, 2012a, 2013b) also distinguished between sands reworked by three types of bottom current: *thermohaline, wind-driven* (in which traction structures are dominant), and *tidal.* Deep-marine BCRS deposits are characterised by sand–mud rhythmites, double mud layers, climbing ripples, mud-draped ripples, alternation of parallel and cross-lamination, sigmoidal cross-bedding with mud drapes, internal erosional surfaces, lenticular bedding, and flaser bedding. These features represent alternating events of traction and suspension deposition.

7.2.1.5. Gravel contourites. Gravel-rich and gravel-bearing contourites are commonly found in drift deposits, as the result of winnowing and seafloor erosion under strong bottom-currents that yield irregular layers and lenses of poorly-to-very-poorly-sorted (1 to >2 Φ), sandy gravel-lag. At high latitudes, the gravel and coarse sandy material from ice rafting remains as a passive input into muddy, silty or sandy contourite deposits and is not subsequently reworked to any great extent by bottom currents. Similar coarse-grained concentrations and gravel pavements develop locally in response to high-velocity bottom-current activity in shallow straits, narrow contourite moats, and passageways.

7.2.1.6. Volcaniclastic contourites. These contourites are similar to the siliciclastic facies described above, except that they are made up mostly of volcaniclastic material.

7.2.1.7. Shale-clast or shale-chip layers. These layers can develop in muddy and sandy contourites, as the result of substrate erosion by strong bottom currents, under conditions in which erosion has reached a firmer substrate and, in some cases, in which burrowing on the non-depositional surface has helped break up the semi-firm mud. The shale clasts are generally millimetre-sized, and occur with their long axes sub-parallel to the bedding and, presumably, to the current direction as well.

7.2.2. Calcareous contourites

7.2.2.1. Calcareous muddy and silty contourites. Calcareous contourites commonly occur in regions with a dominant biogenic input (including open-ocean sites), beneath areas of upwelling, or down-current from a source of biogenic/bioclastic material. The bedding is usually indistinct, but can be enhanced by cyclic variations in composition and/or grain size. Primary sedimentary structures are poorly developed or absent, partly due to bioturbation, but some parallel-to-sub-parallel, indistinct primary lamination may be preserved. The mean grain size usually ranges from silty clay to clayey (and/or sandy) silt. The grain is poorly sorted and in some cases, exhibits a distinct sand-sized fraction that comprises coarser pelagic biogenic particles. The typical composition is pelagic-to-hemipelagic, and includes nannofossils and foraminifers as dominant elements, but in some cases the deposits can contain large amounts of reworked shallow-water carbonate debris from off-shelf or off-reef supplies. There is a variable admixture of siliciclastic or volcaniclastic material.

7.2.2.2. Calcareous sandy contourites. Calcareous sandy contourites are the calcareous equivalent of sandy contourites. In thinner beds, primary sedimentary structures can be affected by bioturbation, but thin-bedded, cross-laminated foraminiferal contourites are also described. Thicker beds can preserve more structures, although lenticularity, non-depositional surfaces, hardgrounds and burrowing also commonly appear in this facies. The mean grain size is sand, and both poorly sorted and well-sorted examples are recognised. These coarse-grained biogenic particles can derive from pelagic, benthic, off-shelf and off-reef sources, and may comprise a variable admixture of siliciclastic, volcaniclastic and siliceous biogenic material. The bioclasts are often fragmented due to transport and iron-stained due to oxidation.

7.2.2.3. Calcareous gravel-lag contourites. Calcareous gravel-lag contourites, including those comprising calcilutite microclasts or chips derived from the erosion of the substrate are not well known from the modern record. Thus, they have been inferred and described from ancient contourite successions.

7.2.2.4. Siliceous bioclastic contourites. Siliceous bioclastic contourites of mud, silt or sand are also described from modern systems. Both muddy (siliciclastic) and calcareous (bioclastic) contourites can be relatively rich in diatomaceous and radiolarian material, particularly at higher latitudes, but are rarely dominated by siliceous bioclastic material. Cross-laminated radiolarian-rich contourite sands are only known from ancient contourite successions.

7.2.3. Chemogenic contourites

Chemogenic contourites are those in which chemical precipitation directly from seawater occurs in association with contourite deposition and/or erosion (hardgrounds) and hiatus surfaces.

7.2.3.1. Manganiferous contourites. These contourites contain manganiferous or ferro-manganiferous horizons, expressed as very fine dispersed particles; a coating on individual particles of the background sediment; fine encrusted horizons or laminae; and micronodules. Bioturbation and burrowing are particularly evident in such cases, where they form a tiered ichnofacies assemblage in the sediment below the non-depositional surface. Extensive areas with larger ferro-manganese nodules on the seafloor as well as pavements are well known.

7.2.3.2. Chemogenic gravel-lag contourites. Where deep-water chemoherms (chemical–biogenic precipitates) of metal–carbonate chimneys, mounds, and encrustations occur in the path of bottom currents, the seafloor is strewn with the fallen and/or eroded debris of the chemoherm material. Locally, this has clearly been winnowed and aligned into chemogenic gravel-lag contourites.

7.3. Contourite facies model

The creation of a definitive facies model for contourites poses major challenges. Obviously, the knowledge on the regional oceanographic/physiographic setting is crucial, but this information is very difficult to obtain for ancient systems. A better understanding of the oceanographic processes related to CDS is needed. In particular, reconciliation between theory and observations will require greater collaboration between physical oceanographers and geologists (see Section 9).

The standard contourite facies model sequence was first proposed by Gonthier et al. (1984) and by Faugères et al. (1984), and was derived from the Faro Drift within the middle slope of the Gulf of Cádiz. This model implies a cyclic trend, encompassing three main facies:

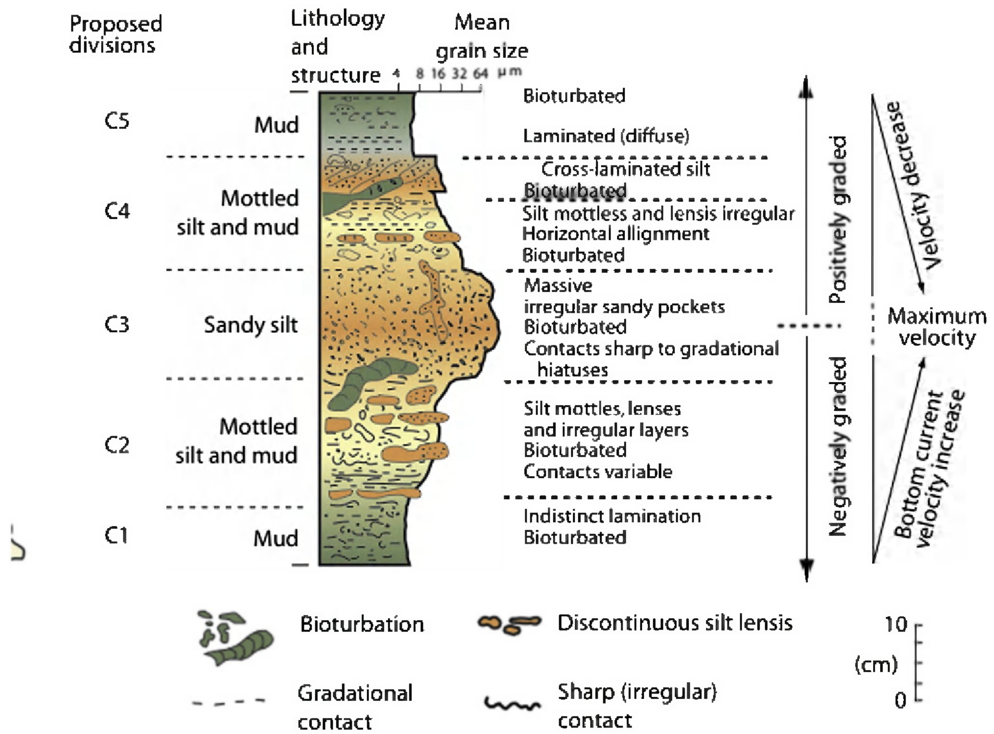


Fig. 24. Standard facies model of contourite sequence, linked to variation in contour-current velocity.
From Stow and Faugères (2008), based on the original figure from Gonthier et al. (1984); with permission from Elsevier.

homogeneous mud; *mottled silt & mud*; and *sand & silt*. These facies are typically arranged in a coarsening-up/fining-up cycle that defines the standard bi-gradational sequence for contourites (Fig. 24). Similar sequences have been illustrated in other margins, in recent and present-day contourite deposits, as on the Brazilian margin (Viana and Faugères, 1998) and the Irish margin (Øvrebø et al., 2006), as well as in the ancient record (e.g. China outcrops, Gao et al., 1998). Other authors have also reported that partial or incomplete sequences are common (Howe et al., 1994; Stoker et al., 1998; Shanmugam, 2000; Howe et al., 2002; Stow et al., 2002a, 2013a; Mulder et al., 2013). A few subsequent modifications (Stow et al., 2002a; Stow, 2005) have demonstrated that the facies and facies sequences associated to contourites vary greatly, making any singular, systematic characterisation of facies rather difficult for the moment. Stow et al. (2002b) slightly modified the standard sequence by using five principal divisions (C1–C5, Fig. 24), and Stow and Faugères (2008) later proposed a model for

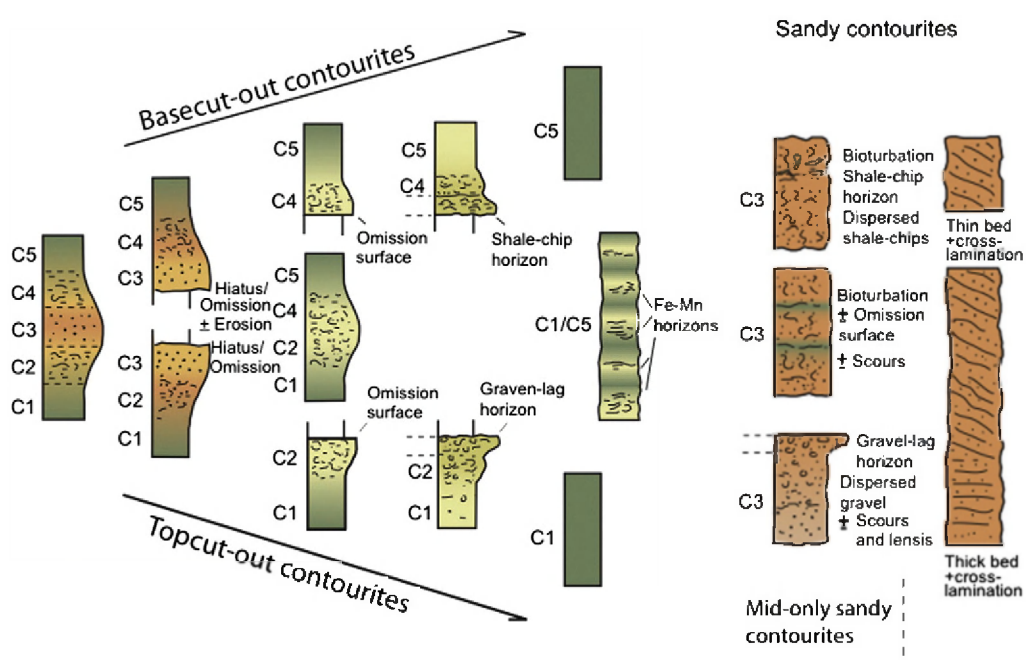


Fig. 25. Modifications and variations on the standard contourite facies models, showing the range of contourite facies, sequences, and partial sequences commonly encountered in contourite successions.
From Stow and Faugères (2008); with permission from Elsevier.

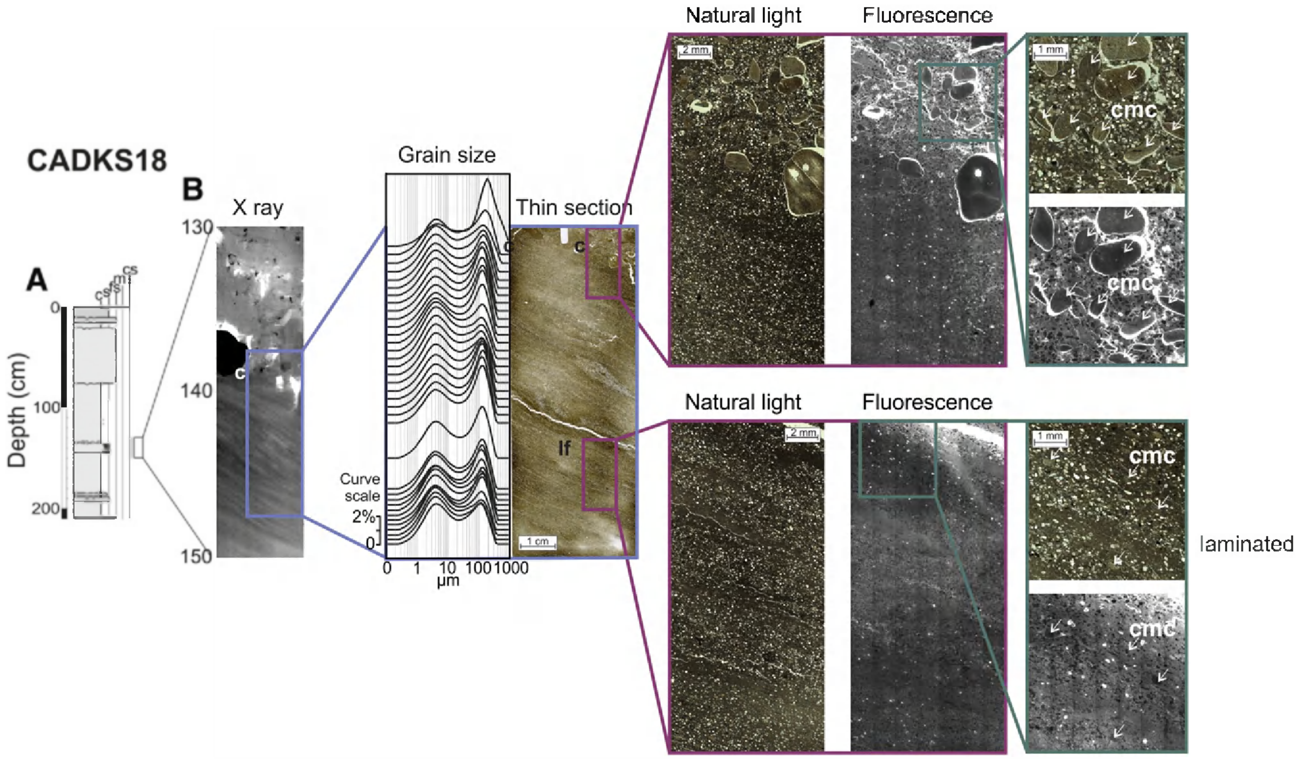


Fig. 26. Example of a gravely contourite in the Gulf of Cádiz (Core CADKS18). A) Core log; B) Laminated and gravely contourite, X-ray, grain size, and indurated thin sections under natural light and fluorescence. c: contact; lf: laminated facies; cmc: consolidated mud clasts.
From Mulder et al. (2013); with permission from Springer.

partial sequences (Fig. 25), which are equally or more common than the full bi-gradational sequence.

In theory, the general model for contourites includes two shifts in the strength of the bottom-current flow: from *weak to strong*, and then *back to weak* (Stow and Holbrook, 1984; Stow et al., 2002a; Huneke and Stow, 2008). Very recently, Mulder et al. (2013) demonstrated that this sequence of facies is only partly related to changes in bottom-current velocity and flow competency, and that it might also be related to the supply of a coarser, terrigenous particle stock. They suggest that the stock could be provided by either increased erosion of indurated mud along the flanks of confined contourite channels (mud clasts), or by increased sediment supply through rivers (quartz grains) and downslope mass transport on the continental shelf and upper slope. These results are consistent with the hypothesis of Masson et al. (2010), who determined that the classical contourite depositional sequence proposed by Gonthier et al. (1984) had to be interpreted with greater care and in the context of the regional sedimentological background.

The main criticism for considering the Faro Drift deposits as the standard contourites facies sequence relates to two facts: this drift is predominantly muddy, and it is located in the distal part of a huge CDS. Moreover, other facies in other parts of the same depositional system have recently been reported (e.g. Hernández-Molina et al., 2012, 2013; Mulder et al., 2013; Stow et al., 2013a; Hernández-Molina et al., 2014), but it is difficult to apply the conceptual model to them. In fact, Mulder et al. (2013) showed that most of the contacts between the classical contourite facies (mottled, fine sand, and coarse sand) are sharp rather than transitional (Fig. 26), which is agreement with the ideas of Shanmugam (2006, 2012a, 2013b).

Given the aforementioned findings, the proposed facies sequence for the Faro Drift could be considered a good model for fine-grained contourite deposits and pervasive bioturbation would be a diagnostic feature of muddy/silty contourites. However, this sequence is not representative for other types of contourite deposits (Martín-Chivelet et al., 2008; Shanmugam, 2012a; Mulder et al., 2013; Shanmugam, 2013b). Authors working in contourite settings in which sandy deposits are more common (Shanmugam et al., 1993a,b, 1995; Shanmugam, 2000, 2012a, 2013b) have reported that bottom currents prevail over burrowing. They emphasise the importance of traction sedimentary structures as diagnostic indicators of the bottom currents from which they derive. Nevertheless, until the full context of these structures is understood (particularly, the genetic association with other structures, and the palaeogeographic and palaeoceanographic frameworks), interpretation of them should be based only on the processes, rather than on the type of sedimentary events or environments (Martín-Chivelet et al., 2008). Laminated, barren, glacigenic muddy contourites observed on Polar margins are often non-bioturbated (Anderson et al., 1979; Pudsey et al., 1988; Mackensen et al., 1989; Grobe and Mackensen, 1992; Pudsey, 1992; Gilbert et al., 1998; Anderson, 1999; Yoon et al., 2000; Lucchi et al., 2002a,b). This particular type of glacigenic contourite facies appears associated to glacial times only, and has been interpreted as resulting from unusual, climate-related, environmental conditions of suppressed primary productivity and oxygen-poor deep waters (Lucchi and Rebesco, 2007).

The controversy over which feature—primary traction structures, or bioturbation—should be the basic diagnostic criterion for the recognition of contourite deposits, might have limited significance, since different authors have worked in different settings. Regardless, the previous research on this issue holds two important lessons: firstly, that there is no unique facies sequence for contourites; and secondly, that traction sedimentary structures are also common within contourites (Carter et al., 1996; Masson et al., 2002; Wynn et al., 2002; Shanmugam, 2006; Martín-Chivelet et al., 2008; Shanmugam, 2012a, 2013b). In recent work, Mutti and Carminatti (2012) have addressed this variability and proposed a preliminary contourite facies tract inferred from core observation in deep-water sands of the Brazilian offshore basins. They have proposed the following six types of facies: muddy fine sand with abundant mudstone clasts (CFA); metre-thick, well-sorted, horizontallylaminated fine and very fine sand (CFB); metre-thick, well-sorted fine and very fine sand with large ripples containing internal sigmoidal laminae (CFC); alternating centimetre-thick packages of ripple-laminated fine-grained sand and bioturbated muddier units with sand streaks (CFD); centimetre-thick packages of lenticular rippled sand and sand streaks alternating with mudstones, in which bioturbation is very common; (CFE); and highly bioturbated, terrigenous, mixed and biogenic (calcareous) mudstones (CFF). Of these facies types, only CFE corresponds to the classic contourite model from Gonthier et al. (1984), Stow et al. (2002a) and Stow and Faugères (2008).

The facies sequence reported by Mutti and Carminatti (2012) demonstrates a greater spectrum of contourite facies than that which had previously been reported, especially for cases in which bottom currents strongly contributed to the reworking and redistribution of turbidite fine sands derived from basin margins (thereby generating mixed turbidite/contourite depositional systems). A better understanding of the CDS and related oceanographic processes is needed, that would describe the small-scale and the large-scale features, as well as their associated facies. Given the economic importance of mixed turbidite/contourite systems, considerable research efforts will have to be made to elucidate their geometry and facies distribution patterns. Indeed, completion of this work will provide the optimal conditions required for proposing standard facies sequences for the variety of contourite deposits.

7.4. Contourites versus turbidites: differentiation and diagnostic criteria

The differentiation between contourites and turbidites has been a controversial issue in sedimentology research since the 1970s (Hollister, 1967; Bouma, 1972; Hollister and Heezen, 1972; Piper, 1972; Bouma, 1973; Bouma and Hollister, 1973). There is no general agreement on which structures can be used to distinguish contourites from other deep-sea deposits such as fine-grained turbidites. Moreover, contourite processes can trigger certain gravity processes and formation of submarine lobes, as occurs in the Gulf of Cádiz (Habgood et al., 2003; Hanquiez et al., 2010), or rework previous turbiditic deposits (Shanmugam et al., 1993a,b; Shanmugam, 2012a, 2013b). Therefore, contourite and turbiditic processes can be linked both vertically and laterally. Consequently, distinction of their products (i.e. mixed deposits) will pose a challenge in future research.

Presently, there is a lack of commonly accepted criteria to distinguish between the contourite components and the turbidite components in mixed deposits at a small scale (cores and outcrops on the basis of lithologic facies), although the difference between them is very clear at a large scale (depositional systems, on the basis of seismic facies; see Figs. 3 and 10). Some authors (Lovell and Stow, 1981; Stow and Piper, 1984) have already considered that contourite deposits can be differentiated from fine turbiditic deposits based on three factors: certain characteristics of the former (widespread burrowing; bioturbation; a lack of a vertical sequence of structures; and a low likelihood of preservation of primary sedimentary structures); and the fact that palaeocurrents are a good criterion for distinguishing between along-slope (i.e. contourite) and down-slope (i.e. turbidite) systems. However, as mentioned above, traction sedimentary structures are present in contourites and are considered by some authors to be viable diagnostic criteria (Carter et al., 1996; Wynn et al., 2002; Martín-Chivelet et al., 2003; Shanmugam, 2006; Martín-Chivelet et al., 2008; Shanmugam, 2012a, 2013b). Also the presence of laminations, enhanced by shell fragments and the concentration of quartz grains, has been reported as resulting from a discrete grain (sortable silt) input as bed load at the base of the contour current (McCave, 2008; Masson et al., 2010; Mulder et al., 2013). This confirms that bed-load transport is a major characteristic of contour current deposition (Shanmugam, 2012a), and explains the superior sorting of sandy contourites relative to sandy turbidites, as has been suggested by Shanmugam (2012a, 2013b).

Presently, the scale of observation is essential for differentiating contourites from turbidites, since the overall geometry, the stratigraphic stacking pattern and facies association differ greatly in each case. Further information could come from more detailed studies in clearly recognised modern CDS, based not only on sediment core analysis, but also on visual (ROV-assisted) analysis of local outcrops on the seafloor (e.g. channel flanks and slide scarps), and on comparison with the ancient record facies, through coring (IODP, etc.) of their facies and facies sequences.

8. Most recent understandings from IODP Expedition 339

Of the 5.5 km of core recovered during IODP Expedition 339 in the Gulf of Cádiz and west off Portugal (http://iodp.tamu.edu/scienceops/expeditions/mediterranean_outflow.html), at least 4.5 km belongs to a CDS (Expedition 339 Scientists, 2012; Hernández-Molina et al., 2013; Stow et al., 2013b). The predominant sedimentary facies includes pelagites, hemipelagites, contourites, turbidites, debrites and slump deposits (Fig. 27). Contourites are the dominant sediment type, comprising 95% of the Quaternary and ca. 50% of the recovered Pliocene succession. This facies group includes sand-rich, muddy sand, silty-mud and mud-rich contourites, all of which were deposited at moderate (20 to 30 cm/ky) to very high (>100 cm/ky) rates of sedimentation. The recovered contourites are remarkably uniform in composition and textural attributes. These contourites feature intense continuous bioturbation throughout, and the muddy and silty contourite deposits are distinguished by a conspicuous absence of primary sedimentary structures. They are characterised by bi-gradational sequences from inverse to normal grading with numerous partial sequence types (Fig. 27). These preliminary results are in agreement with the previously proposed idea that there is a greater variety of facies sequences for bottom-current deposits than what is presently represented in the most commonly accepted contourite facies model (Shanmugam et al., 1993a; Martín-Chivelet et al., 2008; Shanmugam, 2012a; Hernández-Molina et al., 2013; Mulder et al., 2013; Shanmugam, 2013b; Stow et al., 2013b). Moreover, this illustrates that there have been massive spatial and temporal changes in the facies of the same CDS. Therefore, the contourite facies model must be refined: for example, the sand–silt contributions and the role of sediment supply in it must be incorporated. An enormous quantity and extensive distribution of contourite sands (and bottom-current-modified turbidite sands) have been reported (Expedition 339 Scientists, 2012; Hernández-Molina et al., 2013; Stow et al., 2013b) (Fig. 27), especially in the proximal part of the CDS close to the Strait of Gibraltar, where traction sedimentary structures have been found in very thick, sandy contourite layers (>10 m) that had been drilled (Hernández-Molina et al., 2013, 2014).

Additionally, remarkable interactions between contourite and turbidite processes have been reported that are completely new and different from the current facies models. Therefore, the results and forthcoming data analysis from IODP Expedition 339 Scientists will be very important for the future use of contourite systems in palaeoceanographic studies. Additionally, the drilled sandy contourites in the aforementioned proximal part of the CDS are completely different deep-water sands than the turbidite sands that are currently dominant as deep-water oil and gas plays. They seem to be formed in different depositional settings, have different depositional architectures, and are clean and well sorted. Deeply buried sediments with these characteristics would be high quality potential reservoirs. Additionally, the associated contourite muds are very thick, rapidly deposited, and moderately rich in organic carbon (up to 2 wt.%). They could provide potential source rocks in the subsurface, as well as suitable seals in stratigraphic traps. These new findings could herald a paradigm shift for exploration targets in deep-water settings (Hernández-Molina et al., 2013; Stow et al., 2013b).

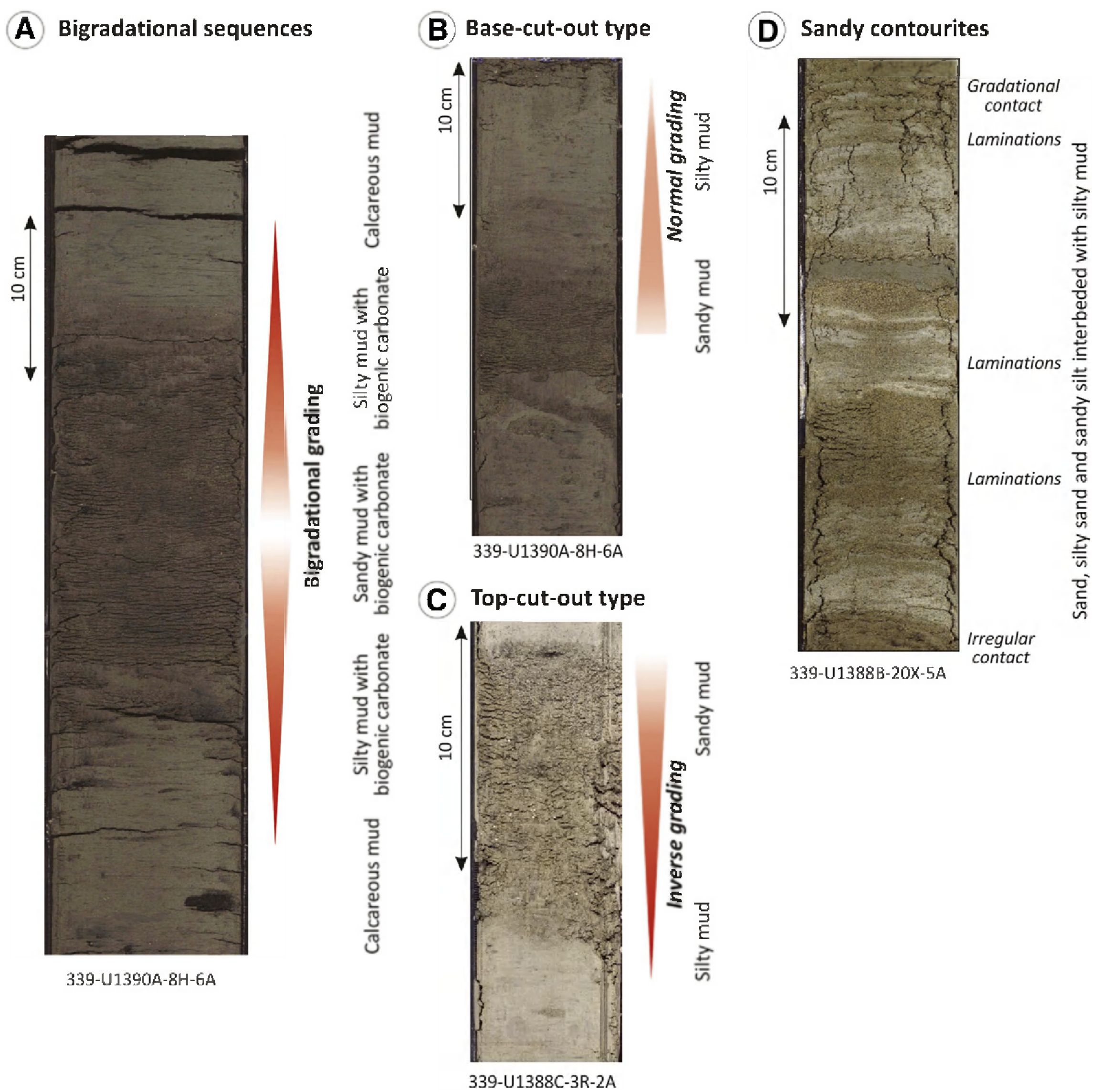


Fig. 27. Examples of the principal sedimentary facies for the contourites recovered during IODP Expedition 339. From Hernández-Molina et al. (2013).

9. Insight and perspectives for the future

As in many marine sciences, most of the future perspectives in contourite research strongly depend on continuous technological advances, which are enabling more and more accurate and detailed studies, both on the seafloor, and on samples retrieved from contourite deposits. More than ever, synergy amongst various datasets (Palomino et al., 2011; Preu et al., 2013; Hernández-Molina et al., 2014) will be needed to enable better recognition of the interactive roles of the individual drivers. One essential task is to establish connections between contourite features, their temporal and spatial evolution, and the oceanographic processes that form them. Scale will be a especially important factor in this quest to better discriminate amongst products and to understand their different processes. Firstly, albeit most of the large CDS have already been extensively mapped, the use of advanced "predictive" and diagnostic knowledge, and of high-resolution geophysical surveying, will enable identification of smaller contourite deposits (in all domains, both marine and lacustrine), which in turn will elucidate the interplay amongst ambient processes and will enable high-resolution palaeoclimatological and palaeoceanographic studies. Secondly, increased resolution will enable improved documentation of the detailed spatial and temporal variability within a single contourite deposit, thereby providing a more realistic view on the nature and variability of the responsible formation processes. Finally, from a larger-scale perspective, it may be necessary to take a step back and to provide an overarching view on contourite drifts or CDS in the same basin, created by the same water masses or at the same time scale. One possible benefit of such an approach would be to elucidate the growth mechanisms of the North and/or South Atlantic CDS.

More specifically, with respect to the driving processes, a more intensive collaboration between physical oceanographers and geologists is highly encouraged. Only such concerted effort will enable reconciliation between theory and experimental observations of contourite deposition and erosion. This work will help to provide a new understanding of contourite development from the combined perspective of sedimentology and fluid dynamics. Here, the use of numerical or sand-box (analogue) modelling could yield significant advances to understand CDS evolution. In fact, such an approach has already been employed to model turbidite processes (Salles et al., 2008; McHargue et al., 2011) and tectonics (Morley et al., 2011), but remains underrepresented in the study of alongslope depositional and erosional processes. Nevertheless, improved modelling of contourite deposits will also require more-detailed, high-resolution in situ observations (van Haren et al., 2013). Special attention should be paid to the future development of seismic oceanography (Pinheiro et al., 2010; Carniel et al., 2012), which will enable combined use of these in situ observations to 2D (or even 3D) insight into the oceanographic processes. An integrated acoustic approach coupled with oceanographic data (e.g. CTD or (L)ADCP; see: Preu et al., 2011, 2012; Hernández-Molina et al., 2014) and moorings (Rebesco et al., 2013) will be essential for identifying oceanographic processes and for correlating these processes with seafloor morphologic features (Fig. 15). Advances in palaeoceanography will likely assist in deconvolution of the contourite record, which is not a straightforward task. For example, the fingerprinting of water masses using isotopic tools such as Nd isotope ratios from the authigenic ferromanganese oxide component, available in bulk sediment and foraminifers (Frank, 2002; Khelifi et al., 2009; Piotrowski et al., 2012), and in cold-water corals (Copard et al., 2011), is a steadily expanding technique that will likely become standard in the future. Similarly, ferromanganese nodules could provide interesting palaeoceanographic data (e.g. González et al., 2010). Lastly, better interpretation and evaluation of grain-size techniques will be necessary (Mulder et al., 2013).

Better characterisation of the contourite products will be fully linked to advances in marine geophysics, making high-resolution 3D seismics (Campbell and Deptuck, 2012), multibeam and backscatter data (Palomino et al., 2011; Sweeney et al., 2012) more readily available for academic purposes, and enabling even more-detailed observations from Automated Underwater Vehicles (AUV). This in turn will improve the further refining of depositional models for CDS and contourite erosive features that involve the initiation, processes, seismic facies and architectural elements of these structures. Interestingly, Brackenridge et al. (2011) have already made a first attempt to fit contourites into a sequence stratigraphic framework, which could ultimately benefit from elucidation of the different orders and types of vertical contourite sequence, especially in high-resolution seismic profiles. Furthermore, the interpretation of these geophysical datasets will require more-accurate "hard" geological data and direct access to these data. Such data include those documented by gradual rollout of the IODP Expedition 339 results (Hernández-Molina et al., 2013) or by its associated MOWER Project (Hernández-Molina et al., 2014). This does not only concerns the ground-truthing of seismic geometries or amplitude anomalies, but also applies to backscatter anomalies on multibeam data that are often used as a proxy for sediment texture. Moreover, the sedimentary analyses of facies and structures are now providing better insight, thanks to state-of-the-art imaging techniques such as indurated thin sections (Mulder et al., 2013), Ichnological Digital Analysis Images Package (IDIAP, Dorador et al., 2013, 2014), and Computerised Tomography (CT) scanning (Flisch and Becker, 2003), which already has been successfully applied in oceanic cores for contourite research (Mena et al., 2011; Mena, 2014) and other sedimentological studies (Pirlet et al., 2012). This work might also elucidate the geotechnical properties of contourites, with respect to slope stability on continental margins. Only high-resolution studies, dedicated coring, and geotechnical analyses can provide more insight into the occurrence of weak layers within contourite deposits that could lead to sediment failure.

Another aspect of contourite processes and products that is poised to receive more attention is their economic potential in the context of hydrocarbon exploration: for example, further exploration of the data from IODP Expedition 339, which suggests an extensive distribution of clean and well-sorted sands (Expedition 339 Scientists, 2012; Hernández-Molina et al., 2013; Stow et al., 2013b). This will enable better evaluation of the potential of sandy contourite systems as reservoirs for oil and gas, as well as the potential of muddy contourites both as source rocks for hydrocarbons and as unconventional reservoirs (Viana, 2008; Shanmugam, 2012a; Brackenridge et al., 2013; Shanmugam, 2013a,b; Stow et al., 2013b). Moreover, the frequent association of contourite deposits, both sandy and muddy with cold-water coral mounds (Huvenne et al., 2009; Van Rooij et al., 2011), could also be regarded as a very interesting unconventional reservoir (Henriet et al., 2014). Therefore, the role of deep-water circulation, and of its variability due to climate, on the ecological health status of deep-water ecosystems (e.g. reefs) must be elucidated. An emerging field that requires more insight into deep bottom-water circulation is the renewed exploration phase for manganese nodules (Hoffert, 2008; Rona, 2008). More knowledge on their formation with respect to bottom currents and seafloor morphology will be required for predictive mapping of these marine resources.

Finally, contourite processes and products should be considered for reclassification under international law, since many continental slopes do not coincide with conceptual models and/or the recommendations of the United Nations Convention on the Law of the Sea (UNCLOS; ABLOS, 2006). In fact, important changes in the slope gradient trend are due to the occurrence of large contourite depositional or erosional features, in some case at the base of the slope (e.g. Hernández-Molina et al., 2009, 2010).

The advances expected in contourite research should lead to the establishment of better diagnostic criteria for contourite identification. This will require cooperation and synergy amongst researchers from all involved disciplines, in order to bridge gaps between theory and experimental observations, and between physical oceanography (present, "short time-scale" processes) and geology (past, "long time-scale" products).

10. Conclusions

Fifty years after contourites were discovered and this journal began, research on the recognition, occurrence, palaeoceanography, and sedimentology of contourite deposits and their related processes has advanced considerably. The term *contourite* has since grown to encompass all marine and (some) lacustrine sedimentary basins. Since the 1990s, several specialised articles have been published by researchers seeking to better define diagnostic criteria and to improve facies models. However, this work is far from being finished, as clearly illustrated in the present review. The main points to be addressed in future research are described below.

1. *Contourite processes are not as simple as initially thought.* As reported in Rebesco and Camerlenghi (2008), the main contourite processes are linked to the bottom-current dynamics. However, bottom currents can be driven by myriad oceanographic processes, most of which are not fully understood.
2. *Given the complexity of contourite processes, the contourite nomenclature might need to be reconsidered* (e.g. when dealing with internal waves, and other episodic processes that affect the seafloor).
3. *The contourite model originally proposed by* Gonthier et al. (1984) *remains valid for muddy deposits.* However, new models will need to be considered for other types of bottom-current deposits. This does not mean that the previously proposed models are wrong: there is simply too great a variety of deposits affected by bottom currents to be described using a single model. These deposits must be documented, and new associations of facies and facies models must be established, based on present-day marine data and outcrops.
4. *More work is needed to understand sandy contourites and their differences with bottom-current reworked turbidite sands.* The economic potential of these deep-water deposits should be explored and evaluated.
5. *Integrated studies drawing on specialists from geology, oceanography and benthic biology will be essential for providing a holistic perspective on, and further advancing, contourite research.*
6. *Given that bottom current-controlled depositional and erosional features are very common in marine basins, at different depths and in various settings, the hitherto underestimated pervasiveness of bottom-water circulation and associated processes in shaping the seafloor and*

in controlling the sedimentary stacking pattern on continental margins must be reconsidered.

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Appendix A

This appendix includes the legend for the numbers and letters in each area in Fig. 9, showing large contourite deposits in the present (recent) ocean basins and in the ancient sedimentary record. Though only a selection of critical references (taken from amongst many others) for each area has been included, numerous references were used to compile Fig. 9. The contourite drifts in present basins were specifically complied for the present work, and later archived and visualised under the “Global Contourite Distribution” in the Marine Regions website (http://www.marineregions.org), under the Thematic Gazetteer (Claus et al., in press).

I) *Contourite drifts in present (recent) oceanic basins*:

1: Fram Strait (Eiken and Hinz, 1993; Howe et al., 2008);
2: Spitsbergen (Rebesco et al., 2013);
3: Vesteralen (Laberg et al., 2002, 2005);
4: Lofoten (Laberg et al., 2002; Laberg and Vorren, 2004);
5: Nyk (Laberg et al., 2001, 2002)
6: Skagerrak Strait (Hass, 1993; Kuijpers et al., 1993);
7: North Sea (Knutz, 2010; Kilhams et al., 2011);
8: Faroe–Shetland (Howe et al., 2002; Stoker et al., 2003; Masson et al., 2004);
9: Hebrides slope (Armishaw et al., 1998, 2000; Howe et al., 2002; Knutz et al., 2002a,b; Masson et al., 2002);
10: Rockall Trough (Stoker et al., 1998, 2001; Akhurst et al., 2002; Howe et al., 2002; Masson et al., 2002; Howe et al., 2006);
11: Rockall-Porcupine region (Van Rooij et al., 2003; Øvrebø et al., 2006; Van Rooij et al., 2007a,b);
12: Feni (McCave and Tucholke, 1986; Stoker et al., 2005; Hassold et al., 2006);
13: Hatton (McCave and Tucholke, 1986; Faugères et al., 1993; 1999; Stoker et al., 2005; Sayago-Gil et al., 2010);
14: Gardar (McCave and Tucholke, 1986; Faugères et al., 1993, 1999; Hassold et al., 2006);
15: Bjorn (McCave and Tucholke, 1986; Faugères et al., 1993, 1999; Stoker et al., 2005);
16: Snorri (McCave and Tucholke, 1986; Faugères et al., 1993, 1999);
17: Eirik (McCave and Tucholke, 1986; Hunter et al., 2007a,b; Müller-Michaelis et al., 2013);
18: Gloria (McCave and Tucholke, 1986; Faugères et al., 1993, 1999);
19: David Strait (Nielsen et al., 2011);
20: Baffin Bay (Knutz et al., 2010);
21: Labrador/Orphan (Faugères et al., 1993; Piper and Gould, 2004; Piper, 2005);
22: Flemish Cap (Faugères et al., 1993, 1999);
23: New Foundland (McCave and Tucholke, 1986; Faugères et al., 1993, 1999);
24: Nova Scotia rise/Canada (Hollister and Heezen, 1972; Campbell, 2011; Campbell and Deptuck, 2012);
25: Hatteras (McCave and Tucholke, 1986; Locker and Laine, 1992; Faugères et al., 1993, 1999);
26: Bermuda rise (McCave and Tucholke, 1986; Faugères et al., 1993, 1999);
27: Blake Blake-Bahama (McCave and Tucholke, 1986; Faugères et al., 1993, 1999; Flood and Giosan, 2002);
28: Caicos (McCave and Tucholke, 1986; Tucholke, 2002);
29: Antilles (McCave and Tucholke, 1986; Tucholke, 2002);
30: Florida Strait (Gardner et al., 1989; Faugères et al., 1999);
31: Gulf of Mexico (Shanmugam, 2012a; Shanmugam et al., 1993a);
32: Bawihka Channel (Hine et al., 1992, 1994);
33: Cap Ferret (Faugères et al., 1998);
34: Le Danois (Ercilla et al., 2008; Van Rooij et al., 2010; Hernandez-Molina et al., 2011a);
35: Ortegal (Jané et al., 2010a,b; Hernandez-Molina et al., 2011a; Maestro et al., 2013);
36: Galicia Bank (Ercilla et al., 2011);
37: Galicia margin (Bender et al., 2012);
38: W Portugal (Alves et al., 2003; Hernandez-Molina et al., 2011a);
39: Gulf of Cádiz (Kenyon and Belderson, 1973; Gonthier et al., 1984; Nelson et al., 1993, 1999; Faugères et al., 1999; Habgood et al., 2003; Mulder et al., 2003; Hernandez-Molina et al., 2006a; Mulder et al., 2006; Hanquiez et al., 2007; Llave et al., 2007; Marchès et al., 2007; Hernandez-Molina et al., 2011a; Llave et al., 2011; Roque et al., 2012; Brackenridge et al., 2013; Stow et al., 2013a,b);
40: Strait of Gibraltar (Kelling and Stanley, 1972; Esteras et al., 2000; Hernández-Molina et al., 2012);
41: Ceuta (Ercilla et al., 2002);
42: Alboran Sea (Palomino et al., 2011; Ercilla et al., 2012; Juan et al., 2012);
43: Rosas (Canals, 1985);
44: Menorca (Mauffret, 1979; Velasco et al., 1996; Frigola et al., 2008);

45: Mallorca (Vandorpe et al., 2011);
46: Corsica (Roveri, 2002; Toucanne et al., 2012);
47: Thyrrenian (Falcini et al., 2010);
48: Messina (Viana et al., 1998a);
49: Adriatic (Verdicchio and Trincardi, 2008; Falcini et al., 2010);
50: Silicia Channel (Martorelli et al., 2011);
52: Malta slope (Micallef et al., 2013);
53: Latakia (Tahchi et al., 2010);
54: Israel (Golik, 1993);
55: Moroccan margin (Casas et al., 2010; Antón et al., 2012; Vandorpe et al., 2014);
56: Madeira (Faugères et al., 1999);
57: Cape Yubi (Schwenk et al., 2010);
58: Cape Blanc (Schwenk et al., 2010);
59: Sierra Leone & Gambia basins (Westall et al., 1993; Jones and Okada, 2006);
60: Pliocene, Equatorial Guinea (Shanmugam, 2012a);
61: Pliocene sand. Edop Field offshore Nigeria. (Shanmugam, 2003, 2012a);
62: Gabon (Biscara et al., 2010);
63: Namibia (Keil and Spieß, 2010);
64: Cape Basin (Weigelt and Uenzelmann-Neben, 2004);
65: Demerara Rise (Damuth and Kumar, 1975; Viana et al., 1998a);
66: Brasil (Viana et al., 2002a,b; Viana, 2008; Borisov et al., 2013);
67: Columbia (Massé et al., 1998; Faugères et al., 2002a);
68: Santos (Duarte and Viana, 2007);
69: Campos (Mutti et al., 1980; Viana et al., 1998b; Lima et al., 2007; Moraes et al., 2007; Viana, 2008);
70: Vema (Mézerais et al., 1993; Faugères et al., 2002b);
71: Ewing (Ewing et al., 1971; Flood and Shor, 1988);
72: Argiro (Ewing et al., 1971);
73: Zapiola (Ewing et al., 1971; Flood and Shor, 1988; Flood et al., 1993; Von Lom-Keil et al., 2002);
74: Argentine & Uruguayan margin (Hernández-Molina et al., 2009; Violante et al., 2010; Bozzano et al., 2011; Grützner et al., 2011; Muñoz et al., 2012; Preu et al., 2013);
75: Argentine Basin (Flood and Shor, 1988; Klaus and Ledbetter, 1988; Flood et al., 1993);
76: Lago Cardiel drift (Gilli et al., 2005);
77: Malvinas/Falkland Trough (Cunningham et al., 2002; Koenitz et al., 2008);
78: Scotia basins (Howe and Pudsey, 1999; Pudsey and Howe, 2002; Maldonaldo et al., 2003, 2005; Lobo et al., 2011; Martos et al., 2013);
79: Weddell Sea basin (Michels et al., 2002; Howe et al., 2004; Maldonado et al., 2006; Lindeque et al., 2012);
80: Eastern Weddell Sea margin (Pudsey et al., 1988; Gilbert et al., 1998; Camerlenghi et al., 2001; Pudsey, 2002);
81: Pennell Coast, Antarctica (Rodríguez and Anderson, 2004);
82: Bellingshausen (Scheuer et al., 2006; Uenzelmann-Neben and Gohl, 2010);
83: Antarctica Peninsula (Rebesco et al., 1996, 1997; Pudsey, 2000; Lucchi et al., 2002a,b; Rebesco et al., 2002; Lucchi and Rebesco, 2007; Rebesco et al., 2007);
84: Larsen Sea (Kuvaas et al., 2004; Gandjukhin, 2004);
85: Cosmonaut (Kuvaas et al., 2005);
86: Prydz Bay (Kuvaas and Leitchenkov, 1992);
87: Lac d'Armor drift, Kerguelen (Heirman et al., 2012);
88: Wilkes Land (Escutia et al., 2002);
89: Transkei Basin/Agulhas (Niemi et al., 2000; Uenzelmann-Neben, 2002; Schlüter and Uenzelmann-Neben, 2007, 2008)
90: Mozambique ridge (Uenzelmann-Neben et al., 2010);
91: Mozambique margin (Preu et al., 2011);
92: Sodwana Bay (Flemming, 1981; Ramsay, 1994);
93: Mozambique Channel (Faugères et al., 1999);
94: Sumba (Reed et al., 1987; Faugères et al., 1999);
95: Marion drift, NE Australian margin (Davies et al., 1991; Faugères et al., 1999);
96: Great Australian Bight (Anderskouv et al., 2010a,b);
97: South China Sea (Lüdmann et al., 2005; Luo et al., 2010; Wang et al., 2010; Gong et al., 2012, 2013; Chen et al., 2013; Li et al., 2013);
98: Miyako Island (Tsuji, 1993; Viana et al., 1998a);
99: Baikal lake (Ceramicola et al., 2001);
100: Okhotsk Sea (Wong et al., 2003);
101: Kurile (Karp et al., 2006);
102: Esmeralda (Carter et al., 2004a,b,c);
103: Campbel skin (Carter et al., 2004a,b,c);
104: Canterbury (Lu et al., 2003; Lu and Fulthorpe, 2004; Fulthorpe et al., 2011);
105: N Bounty (Carter et al., 2004a,b,c);
106: Chatham terrace (Carter et al., 2004a,b,c);
107: Chatham (McCave and Carter, 1997; Carter et al., 2004a,b,c);
108: Louisville (Carter and McCave, 1994; Carter et al., 2004a,b,c);
109: N. Chatham (Wood and Davy, 1994; Carter et al., 2004a,b,c);
110: Rekohu (Carter et al., 2004a,b,c; Joseph et al., 2004);
111: Hikurangi Fan-drift (Carter et al., 2004a,b,c; Venuti et al., 2007);
112: Subducting drift (Carter et al., 2004a,b,c; Joseph et al., 2004);
113: Samoa (Lonsdale, 1981; Faugères et al., 1999);
114: Mid Pacific Seamounts (Lonsdale et al., 1972; Cacchione et al., 1988; Viana et al., 1998a);
115: California Bordeland (Robinson et al., 2007);
116: Carnegie Ridge (Lonsdale et al., 1972; Lonsdale and Malfait, 1974).

II) *Contourite deposits in the ancient record*:
A: Miura-Boso, Japan (Stow and Faugères, 1990; Ito, 1996; Stow et al., 1998b, 2002b);
B: Ningxia (Hua et al., 2010; He et al., 2011; Shanmugam, 2011, 2012a);
C: Pigliang Drif, Gansu, China (Gao et al., 1998; Luo et al., 2002);
D: Yangjiaping, Jiuxi Drift, Hunan, China (Duan et al., 1993; Gao et al., 1998; Stow et al., 1998a; Luo et al., 2002; He et al., 2010);
E: Talme. Israel (Bein and Weiler, 1976);
F: Cyprus (Kähler, 1994; Kähler and Stow, 1998; Stow et al., 2002c; Turnbull, 2004);
G: Calabria (Collela and d'Alessandro, 1988);
H: Mallorca (Barnolas, 2010; Barnolas and Llave, 2012);
I: Caravaca, Spain (Martín-Chivelet et al., 2003, 2008);
J: Ricla, NE Spain. Upper Jurassic (Bádenas et al., 2012; Pomar et al., 2012; Shanmugam, 2012a).
K: Annot Sandstone. Eocene–Oligocene, SE France (Shanmugam, 2003, 2006, 2012a);
L: French–Swiss Alps, Cretaceous (Villars, 1991);
M: Europe and North Africa (Oczlon, 1991, 1994; Hüneke, 2001, 2006, 2007);
N: North Sea (Shanmugam, 2008, 2012a);
O: Danish Basin (Lykke-Andersen and Surlyk, 2004; Esmerode et al., 2007; Surlyk and Lykke-Andersen, 2007; Esmerode et al., 2008);
P: Cretaceous, offshore Norway (Shanmugam, 2012a);
Q: Delaware Basin (Mutti, 1992);
R: Equatorial East Atlantic, Eocene (Sarnthein and Faugères, 1993).

S: Campos (Paleogene) (Mutti et al., 1980; Mutti, 1992; Moraes et al., 2007);

T: Los Molles, Chile (Suarez et al., unpublished in Stow et al., 1998a);

U: Neuquen Basin, Central Andes, Argentina (Martín-Chivelet et al., 2008);

V: DSDP Leg 28, Site 268. Antarctica (Oligocene). (Piper and Brisco, 1975);

W: Lachlan, eastern Australia (Jones et al., 1993);

Y: Waihao Forks, South Island, New Zealand (Carter et al., 2004a).

Z: South of the Chatham Islands, New Zeland (Party, 1999).

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