Deep-Sea Bottom Currents: Their Nature and Distribution

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Introduction

The oceans are stratified into distinct layers with different physical and chemical properties. One of the principal divisions is between the upper warm-water sphere and the deep cold-water sphere, separated by the thermocline. The thermocline is the zone of marked temperature change from about 200 to 1000 m water depth. For physical oceanographers, deep-sea bottom currents are generally defined as the flow of water masses in the cold-water sphere beneath the base of the thermocline (Zenk, 2008).

There are at least three different bottom current types that can be recognized as operating in deep water settings (Shanmugam, 2008; Rebesco et al., 2014; Esentia et al., 2018) including: (a) wind-driven bottom currents, (b) thermohaline bottom currents, and (c) deepwater tidal bottom currents, both barotropic and baroclinic. Bottom currents are also affected by intermittent processes, such as giant eddies, benthic storms, flow cascading, and tsunamis. All of these currents and processes are capable of affecting seafloor sediment through their erosion, transport and deposition.

Early work by the German physical oceanographer Wüst (1933) initially proposed that bottom currents driven by thermohaline circulation might be sufficiently strong to influence sediment flux in the deep ocean basins. But his work was largely ignored until the early 1960s when Bruce Heezen of Woods Hole Oceanographic Institute took up the challenge from a marine geological perspective. In their now seminal paper of 1966, Heezen, et al. demonstrated the very significant effects of contour following bottom currents or contour currents in shaping sedimentation on the deep continental rise off eastern North America. The deposits of these semipermanent alongslope currents soon became known as contourites, clearly distinguishing them from the deposits of downslope event processes known as turbidites. The ensuing decade saw a profusion of research on contourites and bottom currents in and beneath the present-day oceans, and the demarcation of slope-parallel, elongate, mounded sediment bodies made up largely of contourites that became known as contourite drifts (Hollister and Heezen, 1972; McCave and Tucholke, 1986).

For the most part, physical oceanographers have worked independently of geologists on the nature and variability of bottom currents, so that much integration is still required between these disciplines. Important contributions that to some extent bridge this divide have come from the HEBBLE project on the Nova Scotian Rise (Hollister and McCave, 1984), work along the Brazilian continental margin (Viana et al., 1998), and an extensive program of research in the Gulf of Cadiz, culminating in IODP Expedition 339 (Stow et al., 2013; Hernandez-Molina et al. 2014). Prior to this latest mission, the international deep-sea drilling program in its various guises (DSDP, IPOD, ODP, IODP) has contributed enormously to contourite research—the paleoceanographic context and study of oceanic gateways remain primary targets at present. A topical synthesis of ocean currents can be found in Stow (2017).

This contribution in the *Encyclopedia of Ocean Sciences* is one of three on deep-sea bottom currents and their deposits. The focus here is on the nature and variability of bottom currents, based largely on physical oceanographers' observations of modern oceans, their water mass structure and patterns of circulation. The other two contributions outline the contourite drifts, erosion surfaces and bedform morphology caused by bottom current interaction with the seafloor, and the nature of bottom current deposits, known as contourites.

Ocean Stratification

From top to bottom, the ocean is organized into layers, in which the physical and chemical properties of the ocean—temperature, salinity, density, and light penetration—show strong vertical segregation. This layer-cake structure is always present but variable in its stability and subject to periodic breakdown, disruption and change.

 https://doi.org/10.1016/B978-0-12-409548-9.10878-4

Almost all properties of the ocean vary in some way with depth. Perhaps the most important of these, from the perspective of ocean circulation, is temperature. Absorption of incoming solar energy preferentially heats the surface waters, more so at low latitudes and during summer months. This results in a warm surface layer, a broad transition layer (the thermocline) through which the temperature decreases rapidly with depth, and a cold deep homogeneous zone reaching to the ocean floor. Light penetration is attenuated by absorption and scattering, giving an upper photic and lower aphotic zone, with a more or less well defined twilight region in between. Exactly the same broad three-fold layering is true for salinity, which generally shows an increase with depth—through the halocline.

The density of seawater is controlled by its temperature, salinity and pressure, such that colder, saltier and deeper waters are all denser. A rapid density change, known as the pycnocline, is therefore found at approximately the same depth as the thermocline and halocline. This varies from about 10 to 500 m, and is often completely absent at the highest latitudes. Although, winds and waves thoroughly stir and mix the upper layers of the ocean, and commonly destroy the layered structure during major storms, the energy from these forces is rapidly dissipated downwards so that deep ocean waters remain more stable.

These deep waters constitute 80% of the ocean and are themselves stratified into a series of different water masses, most of which are generated at high latitudes by the cooling and sinking of polar waters, which then spread out at depth towards the equator. Depending on just where they are formed (Arctic, Antarctic, Labrador or Greenland Seas, for example) and how much subsequent mixing has taken place, their properties differ. The coldest, most saline waters, formed around Antarctica, are the densest and hence form the deepest layer in the oceans.

However, despite this apparently stable layer-cake structure, the water column is, in fact, complex and variable. The different layers and water masses are constantly on the move as a result of several interacting forces—winds, waves, tides, thermohaline (gravitational) overturn, and the Earth's Coriolis force. Most of this movement is exceptionally slow, but in some parts of each layer the flow of water is more rapid—these are the ocean currents. They occur at the surface, through the mid-ocean layers, and in the bottom waters.

Principal Types of Bottom Current

Wind-Driven Circulation

Wind is the principal force that drives surface currents, but the main surface currents affect flow through several kilometers of the water column, so that their effects are felt even in the deep oceans. The pattern of surface circulation results from a complex interaction of wind drag, pressure gradients and Coriolis deflection. Wind drag is, in fact, a very inefficient process by which the momentum of moving air molecules is transmitted to water molecules at the ocean surface. The speed of water molecules (the current), initially in the direction of the wind, is only about 3%–4% of the wind speed. This means that a wind blowing constantly over a period of time at 50 kph will produce a water current of about 2 kph. The winds blow in a very consistent pattern on a planetary scale.

The second principal force is that caused by seawater being piled up into broad mounds, hundreds of kilometers across, and drawn down into adjacent depressions. Converging currents and persistent onshore winds tend to pile water up faster than it can flow away, while diverging currents result in a drawdown of water and the creation of lows. Water tries to flow back "downhill"—that is, down the pressure gradient so created.

The third factor involved is Coriolis force. This force is the result of planetary rotation and is felt most keenly by all moving objects (water, wind, aeroplanes, etc.) that are not rigidly attached to the Earth's surface. Moving currents of water and of air are affected in exactly the same way by Coriolis force, veering to the right in the northern hemisphere and to the left in the southern hemisphere.

These forces acting on the oceans cause water to flow, be deflected and develop into stronger currents. These, in turn, interact with the topographic barriers caused by the continental land masses and island archipelagos. The surface circulation pattern that results is one of very large oval-shaped gyres north and south of the equator in each ocean, together with a strong and continuous Antarctic Circumpolar Current (Fig. 1). The strongest currents extend their influence throughout much of the water column, still registering significant flow at 4000 m depth. This is especially true of those currents that flow along western margins of ocean basins and are intensified by the Coriolis force, such as the Gulf Stream and Kuroshio Current. The Circumpolar Antarctic Current is also well known for its effect on the deep slope and rise around the Antarctic continent.

Thermohaline Circulation

The pattern of surface circulation is closely linked with a vast deep ocean network of circulating water that transfers energy, nutrients and sediments around the world. This global thermohaline circulation system is driven by density differences linked to water temperature and salinity, rather than by atmospheric winds (Figs. 2 and 3).

The pattern of deep thermohaline circulation in the oceans starts at the surface, mainly at very high latitudes. These are the cold-water kitchens of bottom-water production. During the long polar winters when sunlight is absent or minimal for 24 h a day, seawater becomes extremely cold and begins to freeze. As sea ice forms and expands out from the continents, it incorporates mostly freshwater, making the underlying water more saline. Cold, saline water is denser and sinks rapidly before spreading out across the ocean floors, moving very slowly towards lower latitudes. At the ocean surface, currents flowing from the equator to the poles

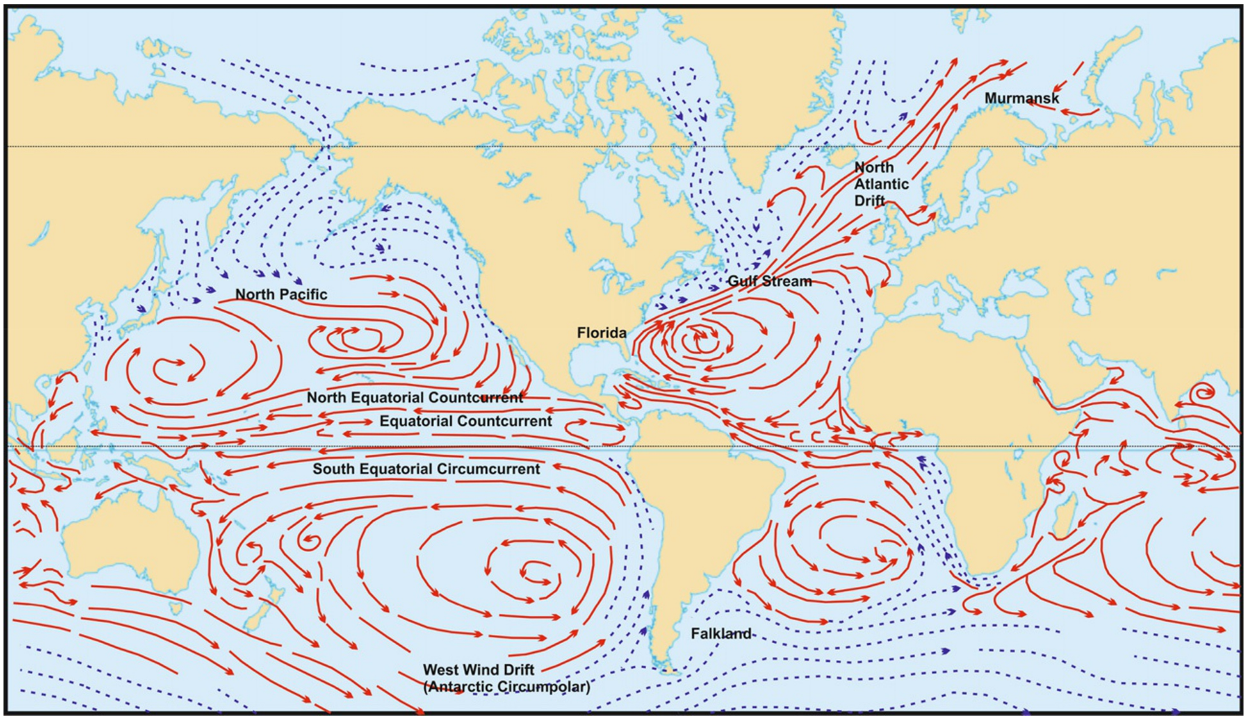

Fig. 1 Wind-driven surface circulation of the global ocean. Note that the principal currents, especially on the western margin of oceans, extend through the ocean and are felt on the deep seafloor.

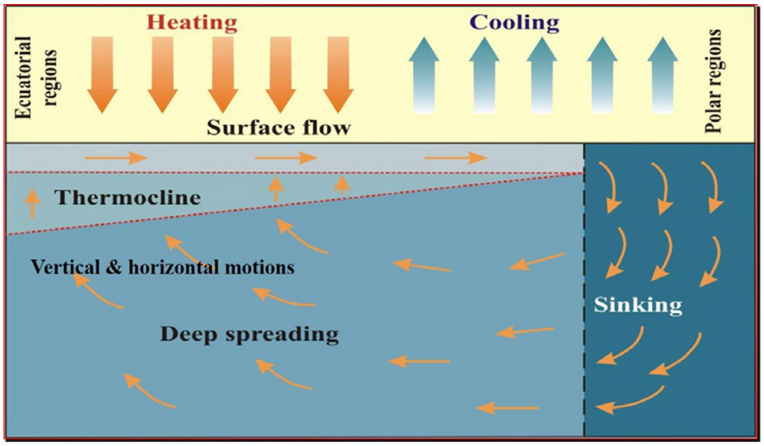

Fig. 2 The mechanism of thermohaline circulation in the ocean basins. The high latitude regions are the principal source areas for the generation of deep ocean water.

replace the waters that have sunk with warmer water, which in its turn will cool and sink. The cycle is apparently without end and has become known as the global ocean conveyer belt (Broecker, 1991).

The principal cold-water kitchens are off Antarctica in the Weddell Sea and the Ross Sea, where the coldest and densest of all water masses is produced. Antarctic Bottom Water, as it is called, cascades down the steep continental slope off Antarctica and spreads out across the floor of the ocean. It flows north into the Atlantic, Pacific and Indian Oceans, crosses the equator and finally mixes upwards far to the northern latitudes. It also becomes partially trapped and compartmentalized by topographic barriers, such as the mid-ocean ridges and aseismic ridge systems, either circulating within the subbasin or escaping to an adjacent sub-basin where the gateway sill is sufficiently deep. Slightly less dense waters are formed in a broad region of the Southern Ocean, before they too plunge beneath the warmer subpolar waters at the Antarctic Convergence zone around 60°S. These give rise to Antarctic Intermediate Waters and associated deep currents.

Cold bottom waters that form beneath sea ice in the Arctic Ocean are mainly trapped there by the high sills that surround the basin. The principal cold-water kitchen in the northern hemisphere, therefore, is the winter cooling of surface waters in the

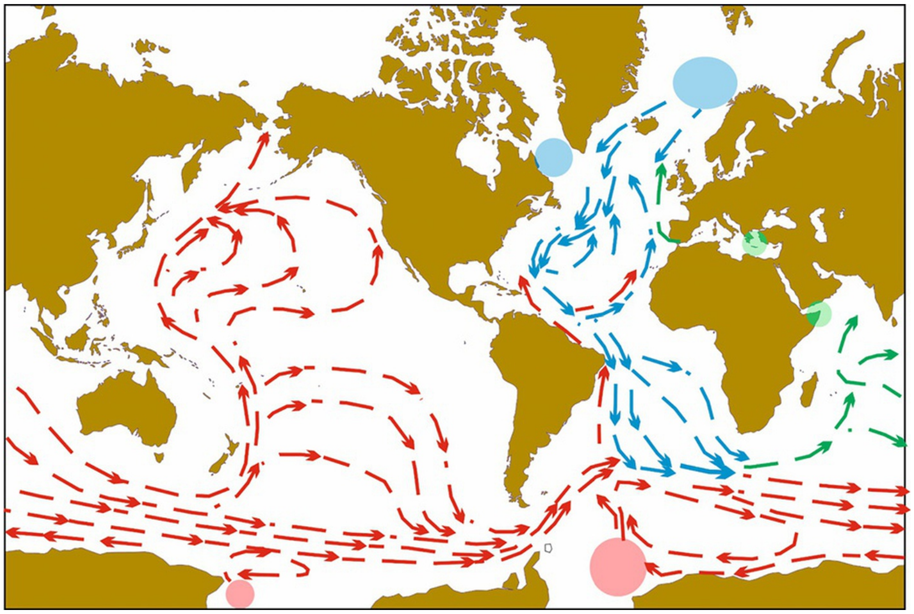

Fig. 3 Thermohaline-driven deep-ocean circulation of the global ocean.

Norwegian and Greenland Seas. As they spill across the Denmark Strait and through the Faeroe-Shetland Channel, they mix with a small amount generated in the Labrador Sea to form North Atlantic Deep Water. This flows southwards until it ultimately blends with Antarctic Bottom Water to form a water mass known simply as Common Water. It is this mixture that dominates the deep Indian Ocean and much of the Pacific as well.

Not all bottom currents are cold. Deep and intermediate depth water is also formed from relatively warm surface waters that are subject to excessive evaporation at low latitudes, and hence to an increase in salinity and density. This process is most effective in semi-enclosed marginal seas and basins. The Mediterranean Sea is currently the principal warm-water kitchen for warm, highly saline water, formed principally in the eastern Mediterranean or Levantine Sea and, to a lesser extent, in the north Aegean Sea. The bottom water so formed flows through the Sicily gateway (between Sicily and North Africa), circulates the western Mediterranean, out through the Straits of Gibraltar and then northwards along the Iberian and north European margin as an intermediate water mass. The Red Sea is a further minor source of warm, saline, intermediate water, flowing out into the Indian Ocean through the Hormuz Strait. At different periods of Earth history, warm saline bottom waters will have been equally or more important than cold water masses in the global thermohaline circulation.

Deep-Water Tidal Bottom Currents

Tides (also known as barotropic tides) are created by the gravitational pull of the Moon and the Sun and by a centrifugal force due to rotation of the Earth. Because the ocean waters are mobile with respect to the solid Earth, lunar gravity is able to pull the water towards it, creating a very slight bulge on the side of the Earth nearest the moon. On the opposite side, centrifugal force pulls water away from the moon, creating a second bulge. These opposing bulges are the high tides. In between, where the water has pulled away, are the areas of low tide. The Sun's gravitational pull is about 40% as strong as the lunar influence, and produces tidal bulges with a quite different periodicity—a solar year rather than the lunar month. When the Earth, moon and sun are aligned, at times of full moon or new moon, then the tidal bulges are in phase, leading to extra high and extra low tides. When the moon and the sun are at right angles to each other with respect to the Earth, then the gravitational attractions are opposed and the tidal range is least.

Much of the global ocean experiences semidiurnal tides that are approximately equal in height. Some of the western Pacific, however, experiences only one high and one low tide per day (diurnal). Tidal ranges across most of the world's coastline are between 1 and 3 m, although in semi-enclosed seas, such as the Mediterranean, Black and Red Seas, they are almost imperceptible. Semi-restricted bays or funnel-shaped estuaries, on the other hand, serve to greatly exaggerate tidal range—the maximum recorded being 16 m in the Bay of Fundy, eastern Canada.

The movement of water by these barotropic tides is generally sluggish throughout the water column, but becomes accentuated where the oceans shallow across the continental shelves and coastal regions, especially so in restricted estuaries and bays. The same increase in tidal current velocity is felt on the deep-sea floor in restricted gateways and channels. Submarine canyons and channels across the continental slope, carved out by turbidity currents and other downslope processes, are affected by diurnal or semidiurnal up-canyon and down-canyon movement of deep tidal bottom currents (Fig. 3). Alongslope contourite channels and the narrow gateways between oceanic basins also act to funnel and increase tidal bottom current velocities.

Tidal energy can also be transferred at depth into an internal wave with tidal periodicity that oscillates along a density interface between the different ocean layers. These are known as internal tides or baroclinic tides. Their interaction with the seafloor and with barotropic tides is still not fully understood.

Bottom Current Characteristics

Thermohaline and Wind-Driven Bottom Currents

We summarize below the principal characteristics of bottom currents, particularly with regard to how they most affect seafloor erosion and contourite deposition (Shanmugam, 2008; Stow et al., 2002, 2008; McCave, 2008; Zenk, 2008; Rebesco et al., 2014) (Fig. 4):

1. They are generally semi-permanent features in the ocean basins, often long-lived through geological time. Therefore, they can be considered to act continuously in affecting sedimentation, rather than as episodic events.
2. They have a net flow alongslope, but can also flow upslope, downslope, around and over topographic obstacles or irregularities. After generation near the surface in the source area, they cascade downslope until they find the appropriate density layer for their salinity-temperature properties. At this level, they turn, under the influence of Coriolis force, to flow alongslope.
3. They typically act as a broad sluggish movement of water (mean velocity < 10 cm s^{-1}) over low gradient slopes and in ocean basins, as more constricted intermediate velocity flows (10–30 cm s^{-1}) over steeper slopes and around topographic obstacles, and as highly constricted high velocity flows (>30 cm s^{-1}) through narrow gateways, passages and over shallow sills. These velocities may exceed 100 cm s^{-1} where the flow is particularly restricted or the slope especially steep.
4. They are highly variable in location, direction and velocity over relatively short timescales (from hours to months). Velocity increase, decrease and flow reversal occur as a result of deep tidal effects. Seasonal changes can result from variation in properties of the water masses generated in the source regions.
5. Mean flow velocity decreases from the core to margins of the current. Flow velocity is directly affected by changes in slope gradient and other topographic irregularities along its course; and also by current meandering and subdivision into two or more strands around obstacles. This kind of flow variability leads to many cycles of deposition, non-deposition and erosion during the course of contourite accumulation.
6. Large eddies develop at the flow margins, where they peel off and move at high angles or in a reverse direction to the main flow. Eddy kinetic energy, sea-surface topographic variations and surface current instabilities can all be transmitted through the water column and so result in marked variation in kinetic energy at the seafloor.

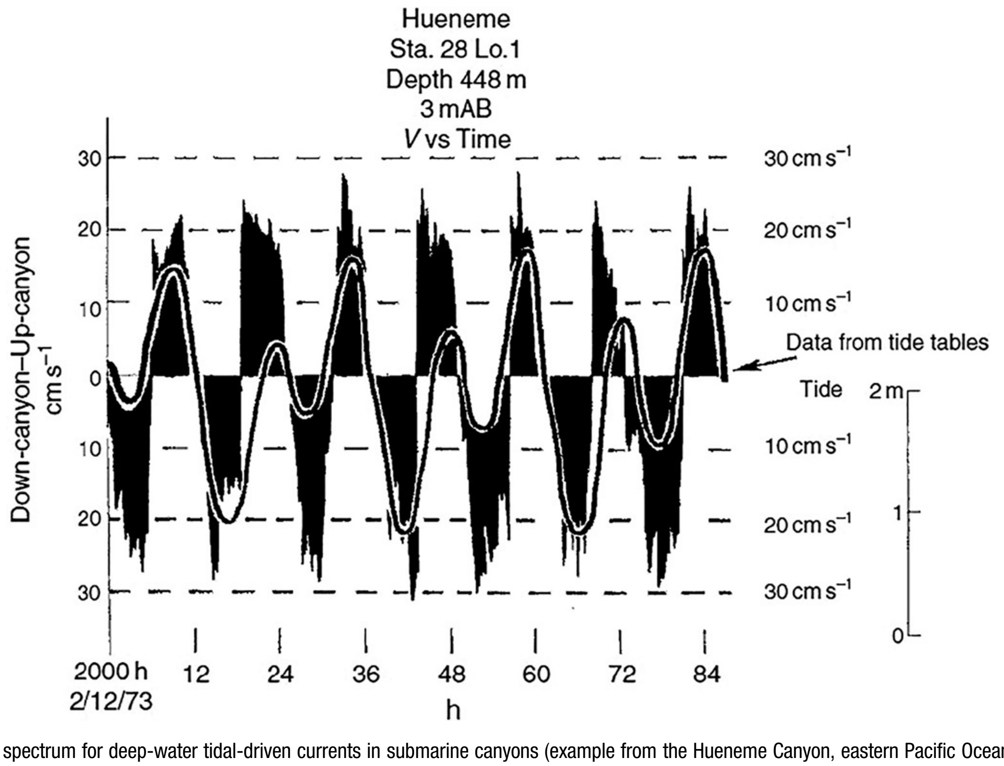

Fig. 4 Velocity spectrum for deep-water tidal-driven currents in submarine canyons (example from the Hueneme Canyon, eastern Pacific Ocean.

7. In places, this induced kinetic energy leads to an alternation of short (days to weeks) episodes of higher velocity benthic storms, and longer periods (weeks to months) of lower velocity. Benthic storms can result in further erosion and resuspension of large volumes of sediment, its incorporation into the bottom nepheloid layer (i.e., suspended sediment load) and transport downstream. Deposition occurs during the quieter low-velocity periods.
8. Bottom currents also show longer-period variability (from decadal to millennial). Some of this can be directly related to changes in climate and sea-level change, for example at the scale of Milankovitch cyclicity. Controls on other period changes are less well understood at present, although their effects on contourite (cyclic) deposition can be observed in the sediment record.

Deep-Water Tidal Bottom Currents

The specific characteristics of tidal currents in deep water are less well known, but we can make a few general observations.

1. They are and have been a continuous process throughout geological time, with a distinctive tidal periodicity. This can lead to alternating normal and reverse current directions, and to periods of higher and lower velocity.
2. The focusing effect of deep-ocean channels leads to maximum current velocities of 25–50 cm s^{-1} in many slope canyons, down to water depths of at least 4600 m, and as high as 75 cm s^{-1} in some cases. These flows may operate at right angles to alongslope bottom currents.
3. Where the tidal bottom current is directed in parallel with alongslope bottom currents, for example through contourite channels or gateways, then the tidal component is added to the alongslope bottom current. This may serve to alternately increase and decrease mean bottom current velocity.

Gateway Currents

The oceans are compartmentalized into abyssal basins separated by submarine mountain ranges and plateaus. Along the length of the global conveyor belt, bottom waters pile up behind such topographic barriers until they reach the spill point. Typically, this occurs as a narrow valley that cuts across the barrier—known as an ocean gateway. As the huge weight of dense water funnels through the narrow gateway, it is severely restricted in width and so accelerates. The bottom currents through such deep ocean passageways can be highly abrasive, scouring away loose sediment, and eroding bare rock. Bottom current velocities of 1–2 m s^{-1} (4–8 kph) are quite common, with a maximum recorded at nearly 3 m s^{-1}.

As the narrow high-velocity bottom current enters the adjacent basin, the dense water spreads out and cascades downslope, as a giant submarine waterfall or cataract. For example, Antarctic Bottom Water piles up behind the Rio Grande Rise in the South Atlantic and then cascades into the Brazilian Basin—a drop of over 1000 m. Even larger in scale is the submarine waterfall located beneath the Denmark Strait in the North Atlantic. Here, 5 million m^3 of water cascades downslope into the North Atlantic basin every second, generating giant eddies and turbulent whirlpools, and drops a vertical distance of over 3.5 km.

Bottom Current Erosion and Deposition

Sediment is eroded by high-energy bottom currents where the current speed exerts sufficient shear stress to overcome the erosion threshold of the substrate sediment. This can occur in gateways and channels, across broad erosional terraces that form in the zone of interaction between different water masses, and wherever localized velocity increase occurs due to benthic storm events or tidal-current reinforcement.

Coarser materials (sands and rarely gravel) are commonly moved as bedload in the current, by processes of rolling and sliding, with finer grain sizes moving by saltation and ultimately by full entrainment into the bottom current. Bedload movement by tractional processes results in the construction of a distinctive seafloor morphology or bottom-current bedforms. Fine to medium sand, silt and clay can all be transported as fully suspended load—also referred to as a nepheloid load or layer. Although typical sediment concentrations are relatively low in nepheloid layers associated with bottom currents (0.01–0.1 ppm, or 0.02–0.2 mg L^{-1}) they are episodically increased up to tenfold as a result of benthic storm erosion and resuspension. This increase also applies to other temporal increases in sediment supply—for example, from localized fine-grained turbidity current input.

Sediment supply to the bottom-current nepheloid layer comes from a range of sources, not only from erosion by the current itself. These include: (a) vertical flux from windblown particles, river suspension plumes, glaciomarine suspension and volcanic dust delivered to the sea surface; (b) vertical flux from sea surface primary productivity, including organic material, calcareous and siliceous bioclastic debris; (c) vertical to slow horizontal advection by a combination of hemipelagic processes, including suspension cascading; (d) direct downslope flux from low-density turbidity currents and hyperpycnal plumes; (e) intermittent downslope flux via spillover processes, including bioturbational and shelf-edge current resuspension; and (f) erosion of the seafloor and resuspension by bottom currents immediately adjacent to and upstream from the site of deposition.

The relative importance of sediment flux from these different sources is poorly known but an estimate has been attempted for the Eirik Drift in the NW Atlantic off the southern tip of Greenland. Eirik Drift measures some 300 km in length, 70 km in width and up to 0.7 km in thickness. The total flow of the Deepwater Boundary Current (DWBC) into the northern end of the drift is measured at approximately 6 Sverdrups (6.10^6 m^3 s^{-1}). If the mean flow concentration is 10^{-4} kg m^{-3}, then the mass sediment flux is around

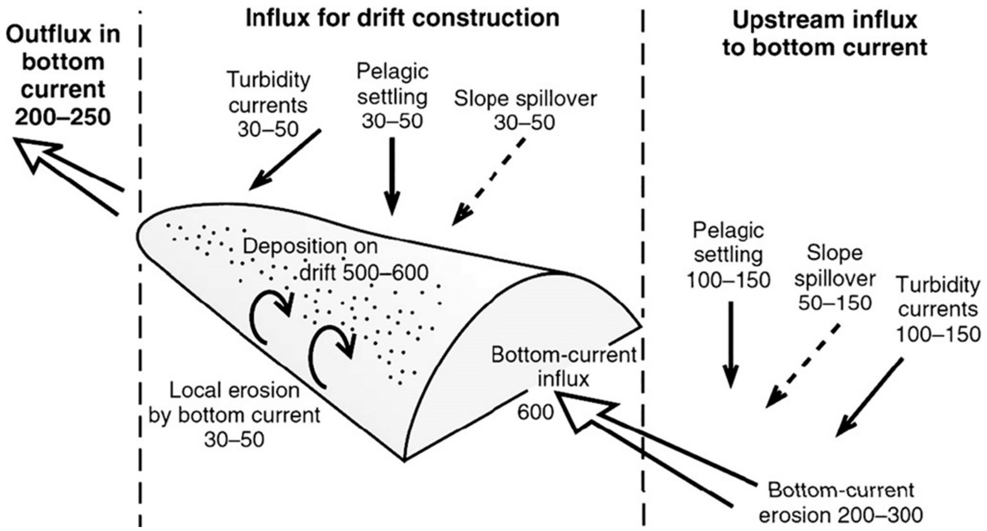

Fig. 5 Contourite drift sediment budget. Estimated sediment flux (in kg/s) from different inputs into the Deepwater Boundary Current that constructs the Eirik Drift, off South Greenland. From Stow, D.A.V., Hunter, S., Wilkinson, D., Hernandez-Molina, J. (2008). The nature of contourite deposition. In: Rebesco, M., Camerlenghi, A. (Eds.), *Contourites. Developments in Sedimentology*, vol. 60, pp. 143–156.

600 kg s^{-1}, yielding annual sediment flux of around 2.10^{10} kg y^{-1} (or 2.10^{7} t per year). The estimated distribution of this flux between the different inputs is shown in Fig. 5.

Where the carrying capacity (or speed) of the bottom current can no longer hold sediment in suspension, then deposition occurs. The average sedimentation rate of contourite deposits varies significantly depending on the location with respect to long-term nature and velocity of the bottom current system and volume and source of sediment supply. Regions of long-term erosion and non-deposition will record zero rates of accumulation, whereas typical sedimentation rates on sheeted drifts range from 3 to 10 cm ky^{-1}, and on mounded drifts from 5 to 30 cm ky^{-1}. Rates in excess of 100 cm ky^{-1} have been recorded on some drifts.

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