low-temperature condensation compounds such as the chlorides, sulfates, and carbonates were present. Indeed, salts of these anions have been leached from primitive meteorites such as the carbonaceous chondrites as well as some other meteorites that have undergone some reheating. On the basis of theoretical calculations, similar compounds should have been present in the accreting Earth. With leaching of these compounds by the original water released from the Earth early in its history, it can be assumed that the oceans were always salty.

Indeed, the argument that oceans over several hundred million years ago might actually have been twice as saline as the contemporary ocean is based on the fact that there are many geological salt deposits and deep brines, often associated with oil fields. The assertion is based on a reasonable premise that salt deposits became important about 400 million years ago. The salt prior to that time had to be stored in the oceans, thus increasing its salinity by a factor of about two higher than the contemporary ocean.

The Future of the Oceans

If most of the water found on Earth was primarily in the oceans, with some dissolved in a molten mantle existing in the early days of the history of the Earth, then as we have seen, there was a decrease in the size of the original oceans. This decrease occurred because of photolysis of water vapor and subsequent loss of hydrogen from the atmosphere or the entrapment of water in hydrated minerals that were then subducted into the mantle.

The rate of photolytic loss must be considerably smaller at present compared to that on the early Earth because of the decrease in the extreme ultraviolet flux from the Sun. Also, the rate of subduction of the hydrated crust must be less now than it was early in the Earth's history because the driving forces for mantle convection and thus plate tectonics are gravitational heat from accumulation and fractionation and heat production by radioactive nuclides, both of which are waning with time. Therefore, the rate of supply of water to the mantle is now diminishing and there may actually be a release of the water stored in the mantle from previous times. As in the case of carbon dioxide, we may be in a steady-state of water supply from the mantle and return of water to the mantle, thereby maintaining the size of the oceans. At any rate, changes in the volume of water will probably not be large in future.

See also

Conservative Elements. Elemental Distribution: Overview. Hydrothermal Vent Fluids, Chemistry of. Mid-ocean Ridge Geochemistry and Petrology. Volcanic Helium.

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OVERFLOWS AND CASCADES

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Introduction

When dense water enters an ocean basin it often does so at a shallow depth passing through a narrow strait or over a sill from a neighboring basin or marginal sea, or flowing down the continental slope from a shelf sea. The dense water flow is affected by the Earth's rotation and so does not flow directly down slope but instead turns to flow partly along the slope. Large-scale, continuous flows from one ocean basin or large sea down into another are referred to as 'overflows,' while smaller scale, intermittent flows of dense water from shelf seas down continental slopes are referred to as 'cascades.' In both cases, the flows can be summarized as gravity currents on slopes in a rotating system (*see* **Rotating Gravity Currents**).

The deep ocean is divided into many sub-basins by ridges, so that overflows play an important role in the global thermohaline circulation, providing the mechanism by which dense waters created in the polar regions flow from one ocean basin down into another. The flow speeds in overflows can be 10–100 times faster than most other deep ocean flows, generating turbulence and entraining ambient sea water into the overflow. Thus the mixing of deep waters is often dominated by the mixing that happens at overflows.

The combination of density contrast, slope, and rotation produces complicated dynamics that are still not fully understood. The effects of the overflow are not limited to the waters immediately around the overflow, but can extend up through the water column producing, in some cases, eddies that can be observed at the sea surface, hundreds of meters above the overflow. These flows can be characterized in terms of the density contrast, flow rate, slope, water depth, and effect of the Earth's rotation.

Observations

Waters of increased density are formed at a wide range of locations throughout the globe. In low latitudes, waters of increased density are formed through evaporation increasing the salinity, such as in the Mediterranean Sea and the Red Sea. At mid-latitudes, wintertime cooling on continental shelves leads to the formation of cold, dense waters that flow down the continental slope into the deep ocean. In polar regions cooling again increases the density, but this can be further enhanced by brine rejection (and thus increased salinity) during ice formation.

The salinity of the Mediterranean is approximately two parts per thousand more than the Atlantic. The resulting density difference drives an exchange flow through the Strait of Gibraltar, with fresher Atlantic water entering the Mediterranean at the surface while the denser, more saline Mediterranean water flows into the Atlantic beneath. The flow of Mediterranean water through the Strait is approximately 0.7 Sv ($1\,\mathrm{Sv} = 10^6\,\mathrm{m^3\,s^{-1}}$). The dense water descends the slope in the eastern Gulf of Cadiz into the Atlantic, with the flow deflected to the right by Coriolis forces and ambient Atlantic water entrained into this fast moving flow (**Figure 1**). The path of the Mediterranean water is strongly influenced by the local bathymetry, especially by canyons that cut the continental slope and provide 'short-cuts' into deeper water. With the strong entrainment of Atlantic water the overflow reaches a neutral level at a depth of approximately 1000 m, where the density of the overflow matches that of the ambient sea water. The Mediterranean waters then flow along the continental slope at this level, although rotating lenses of water are shed from the flow, perhaps through the strong changes in direction of the flow caused by the capes along the flow

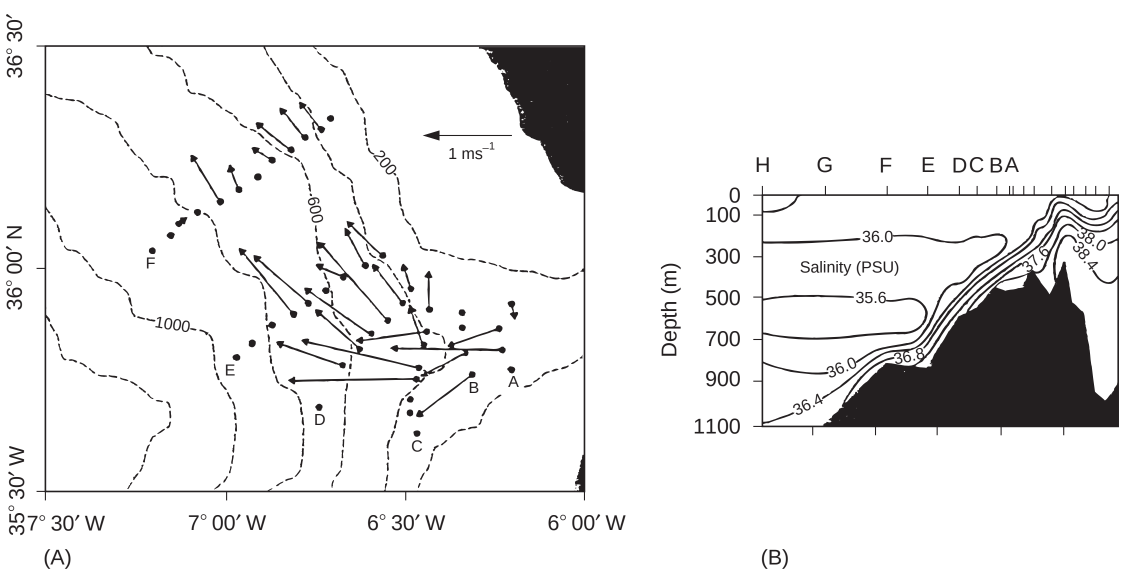

Figure 1 The Mediterranean outflow. (A) The peak outflow velocities in the region just to the west of the Strait of Gibraltar, with depths given in meters. Note how the flow is initially downslope before turning and flowing along the slope. (B) Salinity section along the axis of the outflow, with the letters A–F corresponding to the lines of stations in (A). (Reproduced with permission from Price *et al.*, 1993.)

path. These relatively well-mixed lenses of high salinity water (known as 'Meddies') have been observed to propagate around the North Atlantic as coherent features for considerable periods of time.

On the North-West European Shelf, wintertime cooling results in colder waters on the shelf than in the neighboring deeper waters. In the deeper waters, the cooling produces a surface mixed layer that is deeper than the waters on the shelf. For the same heat loss, the shallower layer of water on the shelf becomes colder than the deeper surface layer off-shelf. Furthermore, during the spring the off-shelf, deep waters are replaced more quickly by warmer water from the south while the flows of waters onto and off the shelf are relatively slow. As the surface waters on the shelf begin to warm and stratify, the deeper waters on the shelf retain their lower temperatures, becoming significantly colder than any nearby waters. These waters tend to flow off-shelf in intermittent cascades. While cascades are smaller in magnitude and shorter in duration than overflows, they obey the same dynamics, turning under the influence of the Earth's rotation to flow along rather than directly downslope. Cascades are also observed flowing off the shelf to the east of Bass Strait, between Tasmania and the mainland of Australia. These cascades reach a flow rate of up to 3 Sv, and have a typical duration of 2–3 days.

Dense waters formed in the Arctic oceans flow south through the Denmark Strait, which lies between Iceland and Greenland, forming one of the largest overflows with a flux of 2.7 Sv. The sill depth at Denmark Strait is approximately 700 m and the overflow descends down into the North Atlantic reaching a depth of between 1000 m and 1500 m before flowing along the slope. The overflow remains at approximately this depth for at least 500 km. It forms cyclonic eddies in the overlying water which have been tracked by deep-drogued buoys (**Figure 2**). Measurements from these, and from moored current meters, show that these eddies have a regular frequency and a strong vorticity.

Similar overflows are observed in the southern hemisphere. Dense ice shelf water (with a temperature below the surface freezing point) flows out of the Filchner Depression in the south-east corner of the Weddell Sea. This flow (with a flux of 0.5 Sv) mixes as it flows down into the Weddell Sea, eventually contributing to the formation of Antarctic Deep and Bottom Waters. A flow in the opposite direction can be found in the south-west corner of the Weddell Sea, where high salinity shelf water, formed by brine rejection during sea ice formation in open water, flows southward underneath the Ronne Ice Shelf. Both these overflows are thought to form eddies, but further measurements are required to establish the details of the flow.

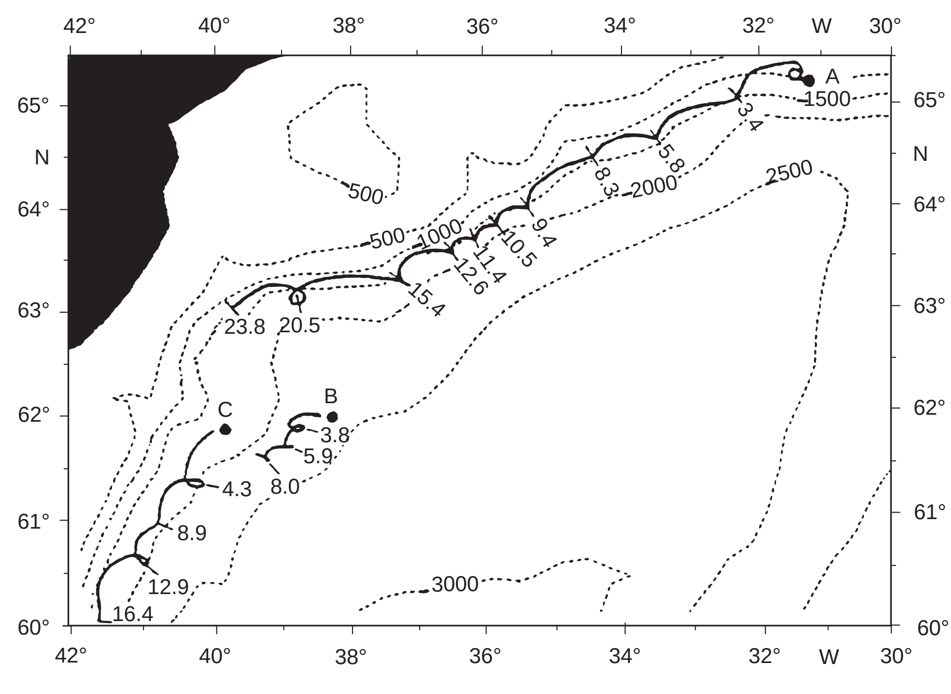

Figure 2 Trajectories of satellite-tracked buoys trapped in cyclonic eddies on the East Greenland continental slope, downstream of the Denmark Strait overflow. Tick marks give time in days, depths are given in meters. The flow here is mainly along the slope, with a superimposed cyclonic (anticlockwise) motion. (Reproduced with permission from Krauss (1996) *Deep-Sea Research I* 43(10): 1661–1667.)

Flow Characteristics

Parameters

Overflows and cascades are formed by sources of dense water on slopes in a rotating system. The source and the system into which the source is flowing can be characterized by five main parameters (**Figure 3**):

1. The source flux (Q), volume per unit time.
2. The density contrast between the source water and the ambient sea water ($\Delta\rho$).
3. The tangent of the slope angle (β).
4. The local value of the Coriolis parameter (f).
5. The depth of ambient sea water above the source (H).

Although the overflow may initially be constrained by flowing through a narrow strait or channel, the width and depth of the overflow are not generally independent parameters. The shape of the overflow (and thus its width and height) adjusts in response to the effects of gravity and rotation as it flows over the slope. Where the ambient sea water is stratified, the last parameter (the depth of ambient water above the source, H) may be replaced by a vertical length scale dependent on the strength and nature of the stratification (see below).

Strong currents in the ambient sea water into which the overflow is flowing have an effect on the overflow, as do irregularities in the slope. In particular, channels and canyons can direct the dense fluid downslope much more rapidly than a similar flow on a smooth slope.

Basic Behavior

A basic description of the behavior of overflows can be given based on the results of laboratory experiments, field observations, and numerical models. As the dense water leaves the channel or flows over the sill that marks the source at the top of the slope, it initially flows directly down the slope under the influence of gravity. The effects of the Earth's rotation deflect the current (to the left in the southern hemisphere, to the right in the north) so that the flow curves to eventually flow mainly along the slope, maintaining its depth. The distance over which the adjustment from downslope to along-slope flow takes place scales with the Rossby radius of deformation (discussed below). The along-slope flow maintains an inviscid geostrophic balance, but is continuously drained by a viscous Ekman layer at its base. The viscous draining flow takes fluid from the base of the current at an angle down the slope.

The inviscid along-slope flow is not always steady. The overflow carries columns of overlying ambient sea water out into deeper water so that these columns of ambient sea water will be stretched (**Figure 4**). This stretching produces cyclonic vorticity in the columns as they conserve their angular momentum. The vorticity in the overlying water breaks up the dense overflow into a series of domes. The cyclonic eddies in the ambient water lie above the domes of

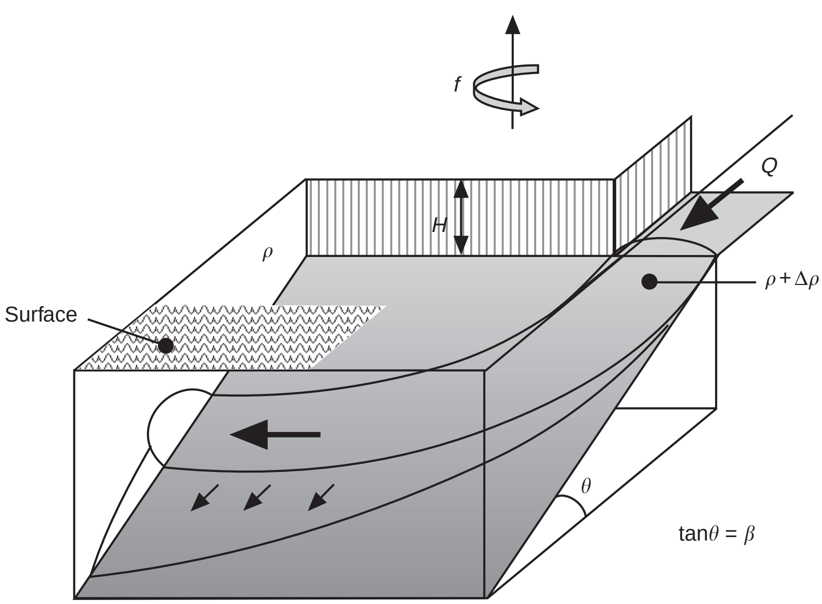

Figure 3 The basic parameters and behavior of overflows. A source (flow rate Q) of relatively dense fluid (density $\rho + \Delta\rho$, where the ambient sea water has density ρ) flows onto a slope (of tangent β) in water of depth H. The local value of the Coriolis parameter (which varies with latitude) is f. Initially the fluid flows down the slope, turning to the right (in the northern hemisphere) under the influence of the Earth's rotation until the main flow is flowing along the slope. A thin viscous sub-layer continuously drains the main flow, taking fluid at an angle down the slope.

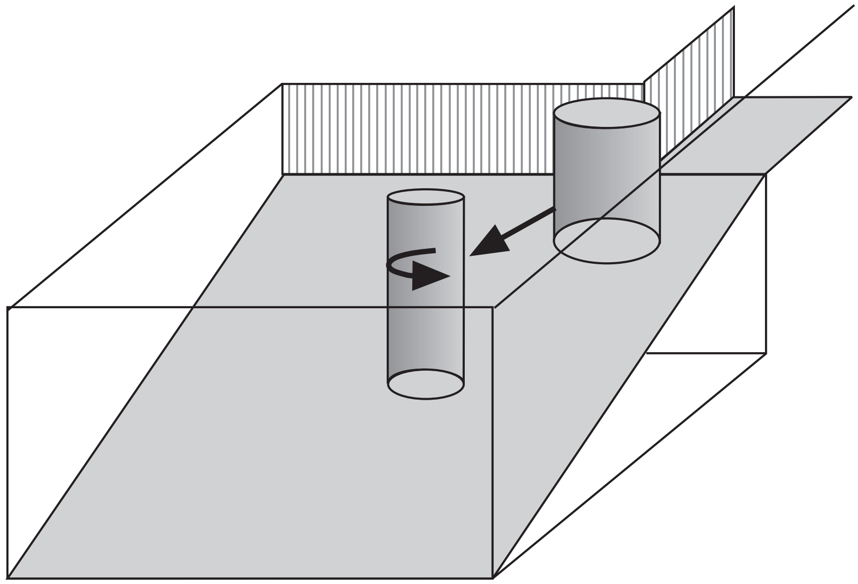

Figure 4 Columns of ambient fluid taken into deeper water stretch and become thinner. Viewed from a stationary observer outside the rotating Earth, columns that appear stationary on the Earth are actually rotating (with vertical vorticity f). When the column becomes thinner it must spin even faster to conserve angular momentum (as ice skaters spin faster when they draw their arms in), giving it cyclonic vorticity relative to the Earth. (Cyclonic is anticlockwise in the northern hemisphere, clockwise in the southern hemisphere.)

dense overflow water and the eddy-dome structures propagate along the slope (but the dense fluid is still continuously drained by the viscous sub-layer). The downslope motion of columns of ambient fluid may occur in the initial downslope flow and adjustment near the source, or as a result of instabilities in the along-slope flow further from the source (see the section on instabilities and mixing below).

Scalings

We make use of the parameters introduced above, but it is useful to express density contrast in terms of the reduced gravity $g' = (\Delta\rho/\rho)g$, where g is the acceleration due to gravity and ρ is the density of the ambient sea water. Two different horizontal length scales have been proposed as the appropriate Rossby radius, the first based on the dynamics of a rotating gravity current and the second based on the injection of fluid of the same density as the ambient sea water. These depend on the source flux, the local value of Coriolis parameter (which varies with latitude), and (for the first type) the density contrast between the dense current and the ambient sea water, with:

$$R_1 = (2Qg')^{1/4} f^{-3/4} \text{ and } R_2 = (Q/f)^{1/3}$$

Typically both these scalings give a Rossby radius of order 10 km for the cases described above. If a column is taken into deeper water by a distance R over a slope of tangent β, then the column will increase in height by an amount βR. This increase in height divided by the original height, H, gives the relative stretching of the water columns and thus give useful nondimensional parameters (depending on the version of R used):

$$\Gamma = \beta R_1/H \text{ or } \mathrm{G} = \beta R_2/H$$

In practice, these parameters are very similar in magnitude and it has yet to be established which is the most useful in describing the behavior of overflows. Comparisons with experiments show that an increase in these parameters (i.e. increased stretching) gives an increase in the frequency of eddy production. Where the ambient sea water is stratified (for example, by a strong thermocline above the source) this can put an effective 'lid' on the vertical influence of the flow and the total depth H is replaced by a height scale derived from the stratification (this would be the height of the thermocline above the source for the simple example mentioned earlier, but a height scale can be obtained for more complicated stratification too).

The appropriate scales for the speed and thickness of the viscous draining layer can also be estimated. This makes it possible to estimate the along-slope distance over which the initial flux is entirely drained into the viscous Ekman layer:

$$Y = \frac{Qf^{3/2}}{g'\beta\sqrt{2v}}$$

where v is the molecular viscosity for laboratory experiments (which have smooth, laminar boundary layers) or a vertical eddy viscosity for oceanographic flows (which have turbulent boundary layers).

Dividing Y by a Rossby radius gives a nondimensional parameter that can be used to describe the importance of the draining flow, for example:

$$Y/R_1 = \frac{Q^{3/4} f^{9/4}}{2^{3/4} g'^{5/4} v^{1/2} \beta}$$

The draining flow is expected to dominate when this parameter is small. In laboratory experiments it has been found that the speed at which the eddies propagate along the slope is determined by Y/R, with slower propagation speeds for low values of Y/R and no eddies at all for sufficiently small values ($Y/R < 1$, approximately).

Modelling and Interpretation

Streamtube Models

In developing a mathematical description of overflows there has been considerable use of 'streamtube' models. These are 'integral' models (meaning that the results can be obtained by integrating the equations of motion over planes perpendicular to the flow) that describe the flow in terms of mean properties as a function of the distance (e.g. s) along the flow center-line (**Figure 5**). Thus the overflow is described by a mean center-line speed, $U(s)$, density contrast, $\Delta\rho(s)$, and shape parameters (e.g. width and height $w(s)$ and $h(s)$). With prescribed parameterizations for the entrainment of ambient sea water into the flow and for the frictional drag at the base of the flow, the path of the overflow centerline $(x(s), y(s))$ can be calculated.

Streamtube models give a simple mathematical formulation that can be used to analyze some of the aspects of overflows. However, the averaging of properties throughout the flow (especially momentum) leads to a somewhat misleading picture of the flow. While the continuously drained along-slope flow (described earlier) does have a mean center position that moves gradually downslope, thinking of the flow in that way does not help comparisons with observations. In many overflows the effect of the bottom friction is not distributed throughout the overflow, but confined to a lower boundary layer, giving the flow sketched in **Figure 3**. However, there are some strong, turbulent overflows (e.g. Mediterranean outflow) in which properties are well mixed and the streamtube approach provides a useful model for direct comparison with oceanographic observations and numerical models.

Laboratory models

Laboratory experiments allow a wide range of parameters and flows to be examined in detail and much of the insight into the behavior of overflows has come from laboratory modeling. Experiments are conducted in tanks mounted on rotating tables to simulate the effects of the Earth's rotation (see **Figure 6** for an example). Typically the scale of the laboratory tanks is of the order of ≤ 1 m, but rotat-

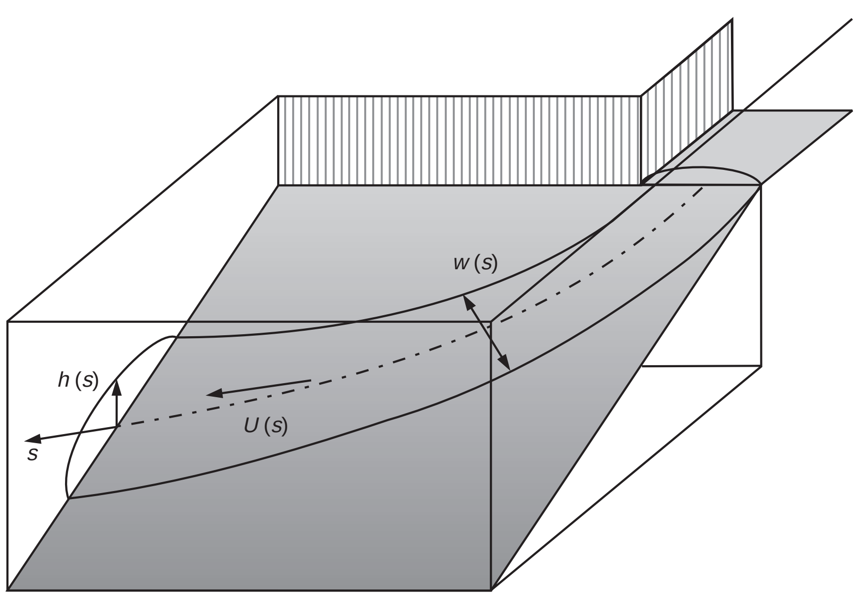

Figure 5 Streamtube models represent the overflow in terms of a mean center-line position with mean overflow properties described as functions of the distance (s) along the center-line.

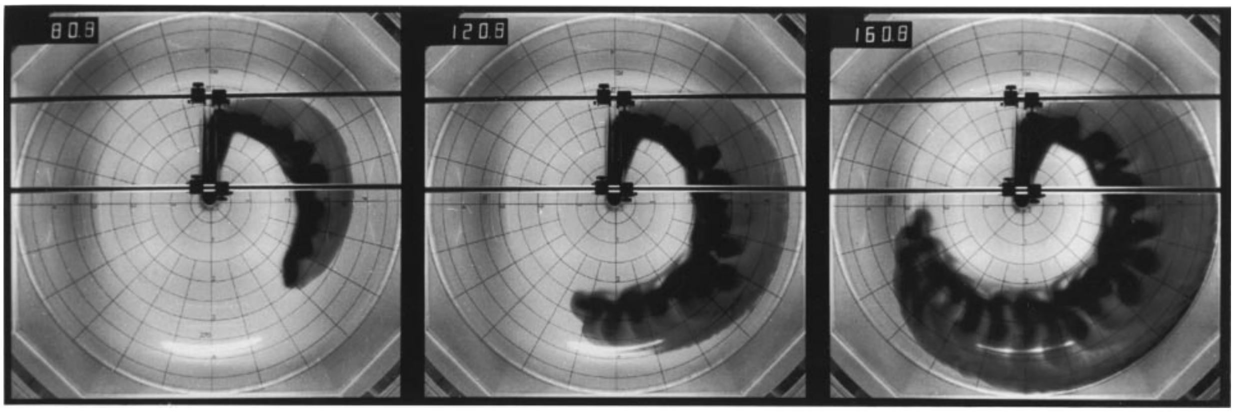

Figure 6 A sequence of photographs of a laboratory experiment (plan view). Dense fluid (dyed) is released on an axisymmetric hill. The inviscid core flows around the hill at constant depth (darkest fluid) while a thin layer of fluid drains down the slope (towards the edge of the image). The dense fluid breaks up into a series of domes with 'sub-plumes' extending downslope from the domes. (Reproduced with permission from Baines and Condie, 1998.)

ing tables with diameters of up to 13 m are used. In order to apply the results from the laboratory experiments to oceanographic flows it is necessary to scale the parameters carefully. The inviscid aspects of the flow can be simulated with confidence in the laboratory (e.g. by matching the values of the nondimensional stretching parameters) so that the along-slope flow and eddy formation processes are likely to be well represented. However, in the laboratory experiments the viscous draining layer is smooth and usually laminar and driven by molecular viscosity, while in the corresponding oceanographic flows the Ekman layer is a turbulent boundary layer. It is therefore necessary to identify an effective vertical eddy viscosity in the oceanographic flows to allow comparison of the draining flow with the laboratory experiments. Reasonable choices of the turbulent eddy viscosity have allowed successful comparisons between the laboratory experiments and oceanographic measurements, but a better understanding of the turbulent oceanic bottom boundary layer would allow more reliable predictions.

In addition to flows where the ambient fluid is homogeneous, experiments have also been conducted where the flow is directed into a stratified ambient fluid. As mentioned earlier, the effect of stratification is to reduce the effective vertical height scale, thus replacing the fluid depth H with a smaller height. This increases the stretching parameters and enhances eddy production. There is some evidence that the propagation speed of the eddies along the slope is reduced by the presence of stratification, perhaps because of internal wave generation. In experiments with a homogeneous ambient fluid, any mixed fluid created at the interface between the overflow and the ambient fluid is still denser than the ambient fluid and eventually rejoins the current. Where the ambient fluid is stratified, mixed fluid may have the same density as ambient fluid above the final depth reached by the main overflow and thus may detrain from the overflow at shallower depths. This distributed injection of overflow water into a stratified ambient has been observed in detail in nonrotating experiments but has yet to be quantified in rotating experiments (*see* **Non-rotating Gravity Currents**).

Numerical Models

In numerical ocean models the ocean is generally divided into a series of boxes, the horizontal dimensions of which are at least 10 km and often much larger (100 km or so for many climate simulations). These boxes are too large to represent overflows accurately since typical overflow widths are of the order of 10 km or less. There are also problems with the accurate representation of mixing at overflows, which often has a dramatic effect on the wider ocean circulation. For example, in a recent project to compare three different models of the North Atlantic, the different models showed very different behavior for meridional overturning and water mass formation. These differences were found to be strongly dependent on the way the mixing of the Denmark Strait overflow was treated. Of the numerical models in that comparison, the isopycnic model (which treats the ocean as made up of a series of layers of uniform density) seems the most promising, since it had too little mixing (when compared with oceanographic observations), whereas the standard level model (with 'step-like' bottom topography) and sigma-coordinate model (with terrain following coordinates) had too much mixing.

Numerical models of idealized geometries (similar to the laboratory models described above) with high resolution have shown similar results to the laboratory models, with both the formation of eddies and draining downslope flow represented provided that the resolution is high enough. These simpler models give useful insights into how numerical ocean models might be improved. In addition to resolution and mixing, the way the bottom drag is treated is also important. The inclusion of special benthic boundary layer models in future ocean models may address this problem.

Instabilities and Mixing

One of the striking features of overflows is their ability to generate strong cyclonic vortices (or eddies) in the overlying water. These have been observed above the Denmark Strait overflow (**Figure 2**) and even above overflows beneath the Ronne Ice Shelf (Antarctica). These eddies fall into two main groups: strong eddies that form very close to the source (referred to as PV for potential vorticity eddies) and those that form after the along-slope flow has been established (BI for baroclinic instability eddies). In flows with PV eddies, much of the dense fluid is concentrated in the domes moving along the slope (although with a thin viscous draining layer), while the BI eddies are accompanied by more substantial 'sub-plumes' of dense fluid extending downslope from the flow. The BI eddies have a weaker effect on the overlying fluid and appear to have a weaker effect on mixing.

Both types of eddies are observed in laboratory experiments and numerical models, and the instabilities leading to BI eddies have been analyzed mathematically. From the mathematical studies an 'interaction' parameter has been defined:

$$\mu = d/\beta R_{\mathrm{H}}$$

where d is the depth of the overflow current as it flows along the slope and R_{H} is a Rossby radius based on the total depth of the fluid:

$$R_H = \frac{\sqrt{g'H}}{f}$$

Thus βR_{H} is the vertical scale corresponding to horizontal motions of scale R_{H} on a slope of tangent β, and the interaction parameter compares this vertical scale to the depth of the current. The mathematical analysis is formally valid for small values of μ and laboratory experiments suggest that BI eddies occur for small μ, while PV eddies occur for large μ.

While these eddy types have been identified, and the conditions under which they occur in laboratory experiments have been approximately defined, the clear characterization of eddies in oceanographic flows has yet to be attempted. The viscous draining flows and mixing behavior of real overflows have also not been quantified in much detail, although shallow layers of relatively unmixed fluid extending downslope with a more mixed flow remaining at constant depth have been observed (e.g. downstream of the Faeroes–Shetland channel in the north-east Atlantic).

Summary

The flow of dense fluid down slopes occurs at a range of scales in the oceans, from small, temporary cascades from shallow shelf seas to the large, continuous flows into major ocean basins. These flows play an important role in controlling the large-scale thermohaline circulation, but their representation in numerical ocean models (especially those used for climate prediction) is poor because of problems with resolution, mixing, and bottom drag. Overflows display a rich behavior and can have a strong effect on the overlying sea water. This behavior is being illuminated through laboratory experiments, numerical and mathematical models, and through increasingly detailed and sophisticated field observations.

See also

Bottom Water Formation. Flows in Straits and Channels. Fluid Dynamics, Introduction and Laboratory Experiments. Meddies and Subsurface Eddies. Non-rotating Gravity Currents. Rotating Gravity Currents. Thermohaline Circulation. Topographic Eddies. Turbulence in the Benthic Boundary Layer.

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OXYANIONS

See **METALLOIDS AND OXYANIONS**

OXYGEN ISOTOPES IN THE OCEAN

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The oxygen isotope signature of sea water varies as a function of the processing of water in the oceanic cycle. The two chemical parameters, salinity and oxygen isotope ratio, are distinctive for various water types. The oxygen-18 to oxygen-16 ratio is

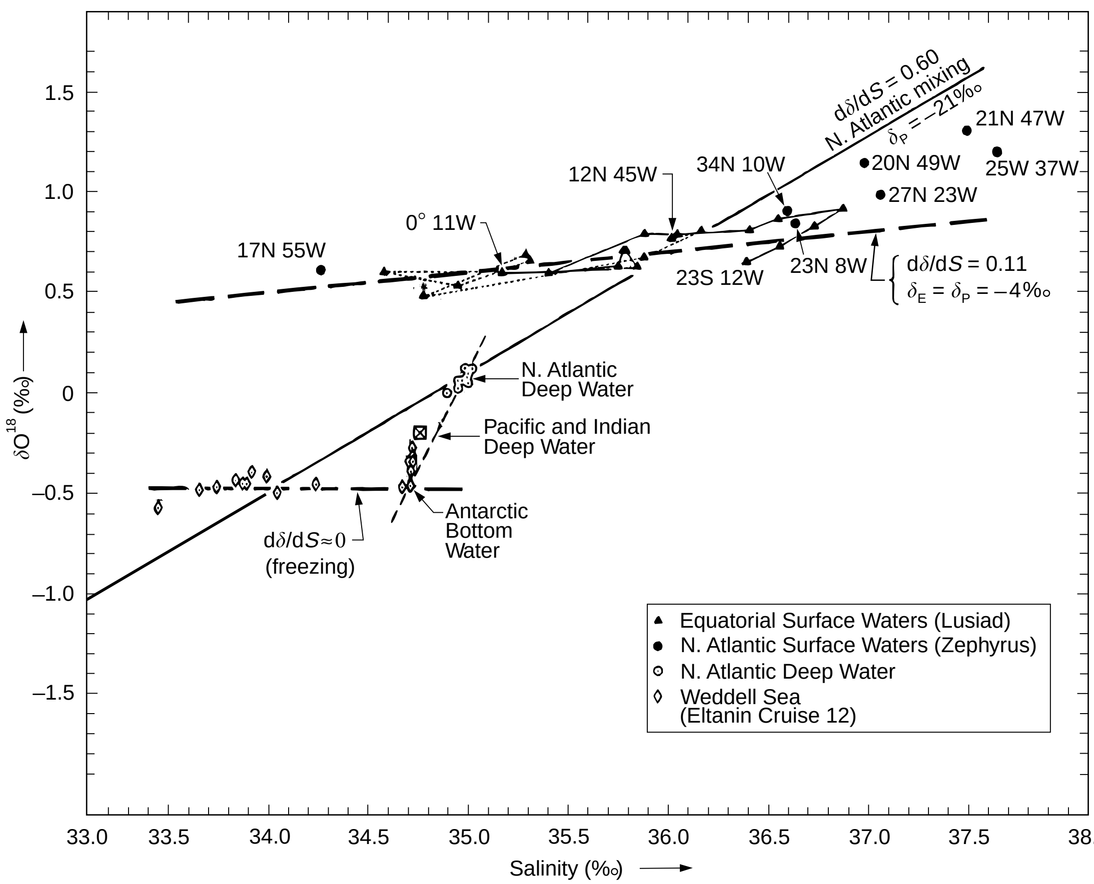

Figure 1 Oxygen-18–salinity relationships in Atlantic surface and deep waters. δ_E and δ_P refer to the isotopic composition of evaporating vapor and precipitation, respectively. (From Craig and Gordon, 1965.)