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NON-ROTATING GRAVITY CURRENTS

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Gravity currents (also known as 'density currents') in the ocean are flows of water that are principally due to differences in density (due to differing temperature and/or salinity) between neighboring water bodies. These density differences cause lateral pressure gradients that produce the horizontal motion. They may occur on length scales ranging from centimeters to hundreds of kilometers, and they have a characteristic structure that is described below. In the oceanic environment there are three main types, as follows. Firstly, gravity currents may move along the ocean bottom, with denser (colder or saltier) water moving under a lighter water mass. Secondly, a body of lighter water may move along the ocean surface, above denser water, and thirdly, a homogeneous body of water of intermediate density may penetrate a larger stratified water body of varying density, with lighter fluid above and denser fluid below. The latter process is termed an intrusion. If these flows last for more than a significant fraction of the inertial period, the Earth's rotation (via the Coriolis force) has an important effect on the flow, and these rotational effects are discussed elsewhere (*see* **Rotating Gravity Currents**). This article gives some examples of these flows, and then describes their basic dynamical properties for flows that move horizontally in one direction, and that move radially outward in two dimensions. It then proceeds to discuss gravity currents that flow down slopes, taking into account the effects of environmental stratification.

Examples

Gravity currents exist in the ocean in a wide variety of forms. On the smallest scale of interest here, they are man-made. Prominent examples are sewage outfalls, in which effluent is piped to some distance offshore and is then released into the ocean. If the effluent is denser than the environment it may spread laterally over the local bottom, but if lighter, it may rise in a plume to the surface and spread there. Another example is effluent from power stations; this may be released on the surface, but it then behaves in a similar fashion to the sewage outfall. Oil released on the surface (as from a shipwrecked tanker) will initially spread as a buoyant gravity current, although being immiscible with water, its mixing properties will be quite different from the previous examples.

Moving to larger scale, gravity currents may be found in rivers, lakes, and estuaries. A sudden surge of water down a river into a lake or estuary can form a gravity current on the bottom or at the surface. If the lake is deep and density-stratified, and the flow into the lake is denser (e.g., colder) than the surface water, most of this inflow may begin as a downslope current, but end up as an intrusion.

Several examples in the coastal environment are due to tides, which can cause the shoreward flow of buoyant fluid into a region on a broad front, or alternatively, a salt wedge of dense fluid moving inward on the bottom. In the open ocean, fronts (*see* **Shelf-sea and Slope Fronts**) that separate water of different densities can exist for periods of days, weeks, and months in much the same location. These fronts are maintained in dynamical balance by the Coriolis force, but they may exhibit gravity current character locally at times when the flow is disturbed and the geostrophic balance is broken.

Sometimes turbulence within a current will cause the sediment on the bottom to become suspended in the water. When this occurs, the sediment can add to the fluid density, and the result constitutes a turbidity current. This may cause a redistribution of sediment, often to deeper water. Earthquakes may trigger submarine landslides that transport and redistribute sediment over large distances. Over timescales of millions of years, these may become conspicuous layers of sedimentary rock (turbidites) that leave a notable signature in the geological record.

Unidirectional Density Currents

The dynamical essentials of the above phenomena can be encapsulated by looking at simple idealized flows that represent them. These can best be seen from simple laboratory experiments. A density current can be created in the laboratory by releasing cold water into relatively warmer water, but it is usually simpler to work instead with isothermal fresh water, and use dissolved salt to make the water denser. In **Figure 1** a body of dense, dyed salty water has been released into a tank of fresh water, producing a density current moving from left to right over a horizontal surface, viewed from the side. The leading part of the current consists of a 'head' with an overhanging leading nose. This head may be regarded as a limiting form of hydraulic jump in a cold layer of dense fluid (*see* **Overflows and Cascades**), in which the depth upstream of the jump is zero. Behind the head the fluid has a three-layer structure, of which the first layer of dense fluid constitutes the main part of the current, at the bottom. This fluid moves faster than the head, and catches up with it. Inside the head it rises and is mixed with the surrounding lighter environmental fluid, and forms a density-stratified layer, spread out above the bottom current. Above this is the third layer of unmixed environmental fluid. Vorticity is produced in the upper part of the head by shear between the head and the ambient fluid, and this is also deposited in the stratified layer. This mixed stratified layer moves slowly in the direction of the current. It is highly turbulent immediately behind the head, where most of the mixing takes place, and this turbulence decays with distance from the head. Here the interfaces between the mixed layer and the dense fluid below and the environmental fluid above are generally stable, so that apart from the decaying turbulence little further significant mixing occurs.

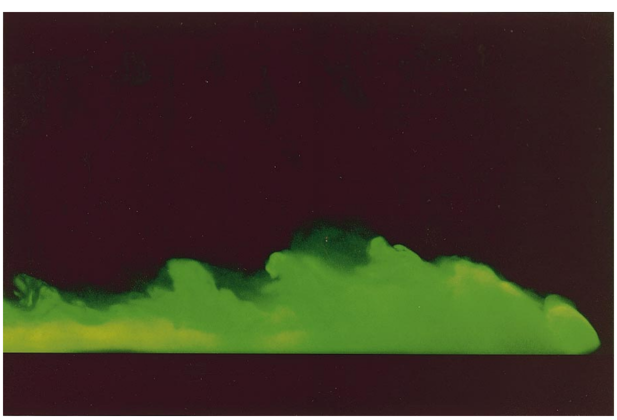

Figure 1 A vertical cross-section showing the structure of the flow in and behind the head of a gravity current advancing over a rigid bottom in a laboratory experiment, from left to right. The dense fluid has been marked by fluorescein dye. Note the dense fluid layer on bottom at the left, and the mixed layer above it. (Dyed salt water inflow density 1.05 g cm^{-3} into fresh water, current depth of order 1–2 cm.)

Density current heads have three-dimensional structure containing many lumps and bumps, and these shapes change continually as the head propagates. Some small parts of the head are more advanced than others, but these then disappear and are overtaken by others. This lobe and cleft structure is due to the drag on the current by the rigid lower surface over which it propagates. This retards the lowest levels of dense fluid, causing the overhanging noses. The lighter fluid beneath these noses then causes the unstable three-dimensional lobe and cleft structure.

In order to quantify this process, an idealized two-dimensional model of a density current can be considered, in which the volume flux Q and density ρ_1 of the dense fluid are specified at a given source location. The Reynolds number ($= Q/\nu$ where ν is kinematic viscosity) is assumed to be large. If the density of the environmental fluid is ρ_0, the buoyancy of the dense fluid is then $g' = (\rho_1 - \rho_0)g/\rho_0$. Behind the head, the velocity v_1 and thickness d_1 of the dense layer are approximately constant. We then have $Q = \mathrm{v}_1 d_1$, and from dimensional analysis alone we have:

$$d_1 \sim (Q^2/g')^{1/3} \quad \mathrm{v}_1 \sim (Qg')^{1/3} \sim (g'd_1)^{1/2} \qquad [1]$$

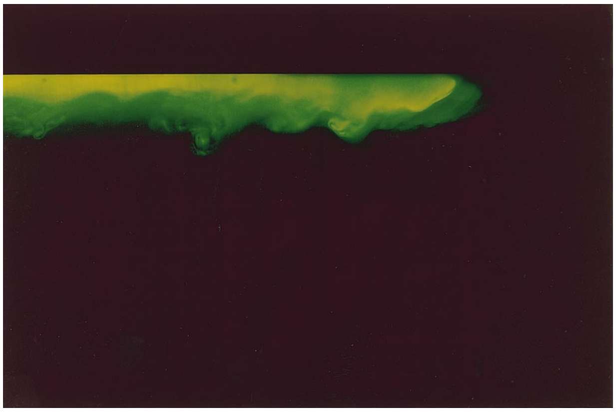

Figure 2 As for **Figure 1**, but showing a section along a gravity current advancing to the left beneath a free surface. Note the (relative) absence of an overhanging nose. (Dyed freshwater inflow into salt water of density 1.1 g cm^{-3}, current depth of order 1–2 cm.)

The speed v_H of the head also scales with v_1, with $v_H < v_1$. Laboratory observations of flow into a deep homogeneous environment at large Reynolds numbers give:

$$v_H = 1.2(g'd_1)^{1/2} \quad v_1 = 1.4(g'd_1)^{1/2} \qquad [2]$$

which are approximately constant with time for a sustained source of fluid. The same observations also show that the total height of the head is typically $3d_1$, and the height of the nose is about $0.15d_1$. These expressions apply to density current heads in general, when the buoyancy and thickness of the dense fluid approaching the head are given by g' and d_1. This steady pattern is maintained because the driving pressure gradient is retarded by loss of momentum through mixing in the head, and the following current is restrained by friction with the bottom surface and the fluid above. When the current becomes long, the thickness of the dense layer decreases gradually away from the source, providing a small pressure gradient to overcome friction.

If buoyant fluid is released into denser fluid causing a gravity current that flows along the surface, the lack of stress on the surface implies that the leading nose is smooth rather than overhanging, and the lobe and cleft structure is largely absent (*see* **Figure 2**). Experiments show that these flows may be significantly affected by surface tension unless the latter is very small compared with buoyancy, but where buoyancy dominates (as in most cases in the ocean) the above dynamical relationships in eqn [2] apply with slightly different constants.

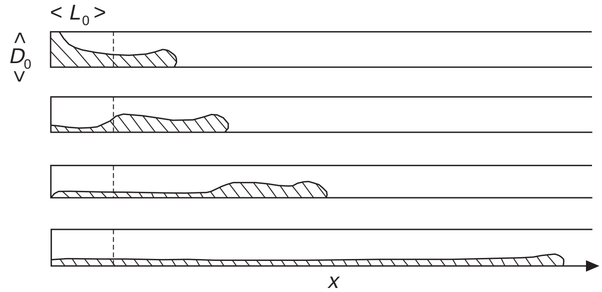

Figure 3 Schematic diagrams showing various stages at successive times in the initial slumping phase of the collapse of a rectangular volume (shown dashed) of dense fluid with depth d_0 equal to the total depth D_0. The mixed region has been omitted from this sketch.

Unidirectional Density Currents due to Collapse of a Dense Body of Fluid

Density currents may also be created by suddenly releasing a large body of dense fluid within a stationary environment of lighter fluid. This models a finite-sized source, and is readily simulated in a laboratory tank by raising a vertical barrier that separates the dense and lighter fluid. In a two-dimensional situation where the fluid spreads in the x-direction only, the flow passes successively through three stages: the slumping, inertia–buoyancy, and viscous stages. In the first stage of collapse of the dense fluid (of initial depth d_0, length L_0, and area $A = A_0 = d_0L_0$ in fluid of total depth D_0), termed the slumping stage, the front travels as a density current of constant speed and depth. However, the surface of the dense fluid behind this front is not horizontal, as large amplitude internal waves propagate on it (see **Figure 3**). These waves reflect from the left-hand end (or center, for a symmetric collapsing body), and the character of the flow changes when they catch up to the front, or density current head.

If the total depth is sufficiently small, or if the upper level fluid is sufficiently strongly stratified to (partially) simulate a rigid lid, the resulting motion of the ambient fluid may affect the waves on the interface with the dense fluid, and hence affect the details of the slumping behavior. This effect may be represented in experiments by a finite total depth, as shown schematically in **Figure 3.** In particular, the presence of a reversed flow in the upper layer causes a corresponding reduction in the speed over the ground of the collapsing front, relative to eqn [2]. The distance x_s at which the slumping stage ends and the waves reach the head, measured from the end of the tank, is then observed to be:

$$x_s/L_0 = 3 + 7.4d_0/D_0 \qquad [3]$$

After the reflected waves have caught up with the density current head the flow enters the second self-similar stage, which is dominated by inertia, buoyancy, and mixing. Here the dense fluid collapses in the form of a rectangle of approximately uniform area, with increasing length and uniformly decreasing height d_1. This rectangular uniformity is maintained by internal waves propagating outwards on the dense fluid interface. Mixing in the main body of the collapsing rectangular current is generally very small. This is because the mean gradient Richardson number at its upper boundary (on which mixing depends) is generally > 0.25, implying that the mean flow is stable and does not generate local mixing (*see* **Upper Ocean Mixing Processes**). Hence the density within the collapsing rectangular body of dense fluid remains largely unaltered until the fluid enters the head, where most of the mixing takes place. The total rectangular area $A(t)$ of unmixed dense fluid continually decreases because $v_1 > v_H$ in eqn [2]. During this stage, the length of the dense fluid increases as $(Ag't^2)^{1/3}$, so that the velocity of the density current head decreases with time as $(Ag'/t)^{1/3}$ because of the steady decrease of d_1. This continues until the third stage is reached, where the dense layer becomes sufficiently thin and slow for viscous effects to become important, a regime that is not relevant here.

In the second stage, dense fluid enters the head from behind where it is mixed with environmental fluid. The resulting mixture is spread out behind, in a layer over the dense layer. The rate of mixing within the head is roughly proportional to its mean height, and hence decreases with it. At early times in this stage ($x > x_s$), the detrained mixed fluid consists mostly of the dense fluid with a small part of environmental fluid. However, as the current proceeds, this proportion reverses and near the end ($x \gg x_s$), nearly all the mixed fluid is environmental. When the end of the second stage has been reached (at $x = x_s + 29A_0^{1/2}$), and mixing has effectively ceased, the total volume of mixed fluid that has been produced is slightly more than twice the initial volume of dense fluid. The volume of the remaining unmixed dense fluid is quite small, so that overall, the dense fluid and environmental fluid have been mixed in approximately equal proportions.

Radial Density Currents

Collapsing localized bodies of dense fluid may be constrained to spread in one direction if they are spatially confined, such as in a submarine valley, but often they are free to spread horizontally without confinement. These have a curved front or head, expanding radially away from a central source. Provided that the curvature of the front is not too large, the speed of the front and of the fluid behind it are given approximately by eqn [2].

Such flows may be modeled by axisymmetric collapsing bodies of dense fluid, which have some properties that resemble those of one-dimensional spreading. Again, there are three stages of spreading: the slumping, inertia–buoyancy balance, and viscous stages. For a body of dense fluid of initial height d_0, radius R_0, and volume V_0, in fluid of overall depth D_0, the initial collapse occurs in the 'slumping' stage, governed by wave propagation and adverse flow of the ambient fluid. Here the radial position R of the front increases at a constant speed in the range $R_0 < R < R_s$, where R_s corresponds to x_s and denotes the limit of the slumping stage. The shear between the front of dense fluid and the ambient fluid produces vorticity that is initially contained within the mixing head of the current, and as the front expands this vorticity increases by vortex stretching. This initial vortex intensifies as it expands, and may be much deeper than the following fluid. For a small initial volume of dense fluid, this expanding vortex ring of dense mixed fluid is all that is produced. But, for a larger initial volume, as the dense fluid expands and the head moves forward, the initial vortex progressively breaks up and is subsumed into the mixed layer. New vorticity is continually created at the head, as for the unidirectional currents, but this is associated with newly mixed fluid and the stretching is much less than that in the initial vortex. In this second inertial–buoyancy phase (where $R > R_s$), gravity waves maintain approximately uniform depth of the body of dense fluid, and it collapses in the form of an axisymmetric pillbox, with the volume slowly decreasing due to mixing and detrainment behind the circular head. If the volume of dense fluid at any given time is $V = \pi R^2 d_1$, and the properties of the front are given by eqn [2], the radius R increases as:

$$R \sim (g'V)^{1/4} t^{1/2} \qquad [4]$$

V slowly decreases as dense fluid entering the head is mixed, and this dynamical regime continues until (in the laboratory) the viscous regime is reached.

If the source of dense fluid is maintained with constant volume flux Q_1, the initial vortex and head form as above, but the flow behind it evolves differently, as the depth d_1 and velocity v_1 of the following radial outflow are not uniform and decrease with radial distance r. The only significant mixing occurs behind the head, and for steady flow, dimensional analysis gives:

$$d_1 \sim \left(\frac{Q_1^2}{g' r^2}\right)^{1/3} \quad v_1 \sim \left(\frac{Q_1 g'}{r}\right)^{1/3} \qquad [5]$$

In the inertio-gravity range, the radial speed of the head is then given by:

$$v_H = 0.63\left(\frac{Q_1 g'}{t}\right)^{1/4} \qquad [6]$$

Density Currents Down Slopes

When the terrain is not horizontal but slopes downwards at angles greater than about 5°, buoyancy acts directly to drive the current downward, and this provides stronger forcing than the indirect effect of establishing a horizontal pressure gradient as described above. The downslope buoyancy force is now $g' \sin\theta$ per unit mass of dense fluid, where θ is the angle between the slope of the terrain and the horizontal. The onset of a steady source of dense fluid at the top of the slope leads to the formation of a density current head similar to that described above. In the current following the head the physics is different because the flow is now unstable, unless the slope angle is very small, and mixing now occurs along the whole length of the density current, not just behind the head. After the passage of the head an approximately steady flow is established, with the buoyancy force being balanced primarily by the entrainment of environmental fluid into the current from above and, to a much lesser extent, by the drag on the bottom surface. For a homogeneous environment, both the size of the head and the thickness of the following current increase with downslope distance.

Behavior of the Head

For a constant supply of dense fluid of buoyancy g'_0 and flow rate Q_0 at the top of the slope, the speed of the head v_H is almost independent of slope angle and is given approximately by:

$$v_H = (1.5 \pm 0.2)\,(g'_0 Q_0)^{1/3} \quad 5^\circ \le \theta \le 90^\circ \qquad [7]$$

However, in spite of this uniform speed, the size of the head increases as it moves downslope, at the rate:

$$\mathrm{d}H/\mathrm{d}x = 0.72\theta/\pi \qquad [8]$$

where H is the height (or thickness) of the head, x is downslope distance, and θ is in radians. The increased buoyancy at steeper slopes is balanced by increased entrainment that acts to keep the head speed constant, regardless of downslope distance and slope angle.

Entrainment in Downslope Flows

Behind the head the flow is unstable, and mixing with the overlying fluid results. This process becomes stronger with increasing slope angle θ. The mixed fluid above the dense layer is also denser than the environment, and hence it also moves downslope under gravity but at a reduced speed. The thickness and volume of this mixed layer both increase with downslope distance. This combined flow may now be regarded as a single entity, and the net downslope buoyancy flux is constant with x and t, and equal to $g'_0 Q_0$. The net entrainment of environmental fluid into this overall downslope flow may be described by an entrainment coefficient E which is a function of the bulk Richardson number R_i, defined by:

$$R_i = g' d \cos\theta / U^2 \qquad [9]$$

where U is the mean velocity and d the mean thickness of the total flow (*see* **Upper Ocean Mixing Processes**). At each level, the velocity of inflow w_e of the environmental fluid into the downflow is given by:

$$w_e = E(R_i)U \qquad [10]$$

where U is the mean velocity of the downflow. E decreases monotonically with increasing R_i, from $E = 0.075$ at $R_i = 0$ to very small values for $R_i > 0.8$. In these flows, R_i is approximately constant with downslope distance, and decreases with increasing slope angle. Experiments show that U, and hence w_e, are also constant, so that the downslope flux Q and the lateral spreading increase linearly with distance.

Downslope Flows into Stratified Environments

If the environmental fluid is density stratified, the effects of this stratification on density currents flowing over a horizontal surface are mostly limited to its effects on the surrounding flow, including the generation of internal waves (*see* **Internal Waves**). However, for flows down slopes, if ambient stratification is present it is a major parameter. If N is the buoyancy frequency (*see* **Upper Ocean Vertical Structure**) of the ambient stratification, a depth D may be identified below the source, defined by $N^2 = g'_0/D$, where the ambient density equals the

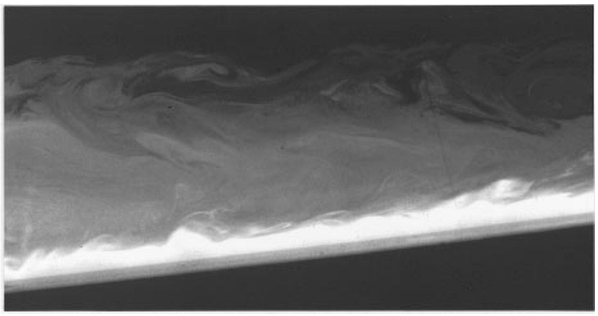

Figure 4 An example of flow of a layer of dense fluid down a slope at 6° to the horizontal, in a density-stratified environment, showing the structure of the interface. The dense fluid is the lightly colored dyed layer close to the boundary. Note the filaments extending from it, and the mixed fluid above. (M = 0.015, with current depth of order 1 cm, buoyancy frequency of stratification $N = 1.24\ \mathrm{rad\,s^{-1}}$, $Q = 2.9\ \mathrm{cm^2\,s^{-1}}$, $g' = 19.4\ \mathrm{cm\,s^{-2}}$.)

inflowing density. If there were no mixing, all of the inflowing fluid would be expected to reach and spread horizontally at this level. The speed of the head of a downslope gravity current into a stratified environment is again approximately constant over most of its distance travelled, and scales with eqn [7] above. The main differences from the homogeneous case concern the following current, which is discussed next.

For slope angles less than about 20°, which covers most cases of interest in the ocean, it is appropriate to regard the main dense current and the mixed layer above it as separate entities. In laboratory experiments a clear interface is visible between them, as seen in **Figure 4**, although there are turbulent fluxes across it. These flows are governed by two dimensionless parameters – a bulk Richardson number R_i defined as in eqn [9] but based on the dense layer only, and a parameter M, defined by:

$$M = QN^3/g'^2 \qquad [11]$$

which is a measure of the effect of the ambient stratification. The dense layer is observed to have approximately uniform thickness over most of its length, but its velocity mostly decreases with downslope distance. It loses fluid to the mixed layer by detrainment, but also entrains fluid from it so that its density progressively decreases, and the fluid that remains in the dense current reaches its ambient level where it spreads out, at a height somewhere above the level of D below the source. Entrainment into this dense layer may be expressed in terms of an entrainment coefficient that depends on R_i, in a manner similar to eqn [10]. There is also a loss of dense fluid to the mixed layer, and an exchange between the mixed layer and the environment. Fluid may leave the mixed layer to find its own neutral level, and this occurs continuously along the path of the current. Hence dense fluid from the source may be distributed over a range of depths, and not all at one level. This detrainment into the environment depends on both R_i and M – larger M implies larger detrainment, to the extent that all the dense fluid may be detrained before it reaches its ambient level.

The above sections describe the 'pure' dynamical form of a gravity current, in various different situations. But when gravity currents occur in the ocean, they are often strongly influenced by other factors over most of their life cycles. In particular, the flow of dense fluid down continental slopes is usually affected by the Earth's rotation, which means that the flow is at least partly in geostrophic balance, with an Ekman layer next to the bottom boundary (*see* **Rotating Gravity Currents**). This applies in particular to the deep flows through Denmark Strait and around Antarctica, that drive the overturning thermohaline circulation. Also, sediment suspension is a common cause of submarine gravity currents, and it may also be a factor in these deep flows.

See also

Internal Waves. Rotating Gravity Currents. Shelf-sea and Slope Fronts. Upper Ocean Mixing Processes. Upper Ocean Vertical Structure.

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