Introduction

Buoyancy is one of the main forces driving flows on our planet and buoyancy-driven flows encompass a wide spectrum of geophysical flows. In this book, contributions by leading world scientists summarize our present theoretical, observational, experimental, and modeling understanding of buoyancy-driven flows. These flows range from buoyant coastal currents to dense overflows in the ocean, and from avalanches to volcanic pyroclastic flows on the Earth's surface. By design, there is a strong emphasis on the ocean where a wide range of buoyancy-driven flows is observed. Buoyancy-driven currents play a key role in the global ocean circulation and in climate variability through deep-water formation. Formation of dense water usually occurs in marginal seas, which are either cooler (at high latitudes) or saltier (due to greater evaporation rates). These dense waters enter the ocean as a gravity-driven current, entrain surrounding waters as they descend along the continental slope, and modify the ocean's stratification as they become part of the global ocean circulation. Buoyancy-driven currents are also primarily responsible for the redistribution of fresh water throughout the world's oceans. In particular, buoyant coastal currents transport fresh water, heat, nutrients, sediments, biogeochemicals, pollutants, and biological organisms along many continental shelves and thus have significant impacts on ecosystems, fisheries, and coastal circulation.

In our examination of oceanic buoyancy-driven flows, we first provide a broad overview of our current understanding of these flows from observations, laboratory experiments, and idealized model configurations (Chapters 1–5). This is followed by an in-depth discussion on the importance of correctly representing processes in buoyancy-driven flows that are not currently resolved in the ocean component of climate models (Chapter 6) and on the difficulty of properly representing these flows in eddy-resolving ocean models (Chapter 7). Finally, oceanic buoyancy-driven flows are put in the context of a wider range of geophysical problems (atmospheric flows, volcanic flows, and avalanches; Chapters 8–10).

The dynamics regulating buoyancy-driven flows are almost equivalent regardless of whether the phenomena are coastal currents or avalanches. The driving forces are similarly represented in the equations of motion, but the latter may be characterized by terms specific to the kind of motion being investigated. For example, terms representing two-phase flows may be necessary when investigating volcanic flows, whereas terms representing mixing may be fundamental when investigating dense overflows.

The term "buoyancy-driven flow" (or alternatively, "density-driven flows") is used to identify those flows whose motion is forced by a horizontal density difference with the surrounding fluid. A flow is often referred to as *buoyant* when the moving fluid is lighter than the surrounding fluid and as *dense* when the fluid is heavier than the ambient fluid. Both currents, light and dense, may also be called "gravity currents" to indicate that the forcing term in the equations of motion is the gravitational term that arises from the density difference between the fluid in consideration and the surrounding fluid. When only two fluids are considered, the gravitational term contains the so-called reduced gravity, defined as

$$g' = g\frac{\Delta\rho}{\rho},$$

which substitutes the gravitational acceleration g in the equations of motion to take into account that the fluid in consideration (lighter or denser) is embedded in another fluid with a slightly different density. For flows that cannot be simply represented by two fluids with different densities, the "reduced gravity" is replaced by an expression that considers the continuous change of density with depth, in contrast with the sharp change in density characteristic of two fluids. This expression can be written as

$$g' = N^2 H,$$

where $N^2 = \frac{g}{\rho}\frac{\partial\rho}{\partial z}$ is the Brunt-Väisälä frequency, or buoyancy frequency, which is the frequency at which a vertically displaced parcel will oscillate within a statically stable environment, and H is the distance over which the vertical density gradient has been calculated.

Although the Earth's rotation influences many of the flows on our planet, our understanding of buoyancy-driven flows was first developed by ignoring the effects of planetary rotation on the fluid motion. An example of a flow that is density driven, but not strongly influenced by the Earth's rotation, is the sea breeze, a significant feature in coastal meteorology. During the day, the sun heats the land more than it heats the ocean; the air just above the land is warmer, and therefore lighter, than the air over the ocean. Hence, colder, denser air will flow from the ocean to the land. The dynamics of the sea breeze helped inspire Dr. John Simpson to devote his life to the understanding of gravity currents in the absence of rotation using mainly laboratory experiments (see Simpson 1997, for a review). One of the first theories to describe the motion of gravity currents was developed by Benjamin (1968) to predict the frontal velocity of a gravity current generated by the release of harmful gases in coal mines. Subsequent laboratory experiments (Gardner and Crow 1970; Lowe et al. 2005), using the so-called "lock release technique," verified the validity of the early theories by investigating an air cavity in a rectangular horizontal duct. Although these theories neglected the influence of the Earth's rotation, they have proven useful for understanding the behavior of a variety of gravity currents, including the dynamics regulating the natural ventilation in a building and the dynamics of ocean currents. Chapter 1 of this book discusses the fundamental aspects of nonrotating buoyancy-driven currents using theories and laboratory experiments. The limitations of earlier theories are discussed and new approaches are presented in order to examine some of the details in gravity currents that are still not fully understood, such as the generation of internal waves by a propagating gravity current in a stratified environment.

The introduction of the Earth's rotation does complicate the dynamics of buoyancy-driven flows (see Colin de Verdière 1988, for a discussion of large-scale flow patterns driven by a surface buoyancy flux). Chapter 2 illustrates how an ocean basin responds to buoyancy (i.e., cooling or heating) applied at its surface. The mechanism for propagation of information across the basin is a function of stratification and the variation of the Coriolis parameter with latitude (i.e., the β-effect). A full comprehension of buoyancy-driven flows in enclosed ocean basins requires a thorough understanding of the dynamics close to the boundaries, where dissipation effects are important. The buoyancy forcing applied at the free surface generates large vertical motions near the basin boundaries in regions that are not close to the forcing. Chapter 3 further investigates this response by adding more complexity to the problem. Numerical models are used to investigate the effect of buoyancy forcing under a more realistic scenario (i.e., one that includes unsteady flows and eddies). Building on the theoretical results of Chapter 2, Chapter 3 shows that baroclinic eddies can be an effective mechanism for moving the buoyancy response away from the forcing location. As in Chapter 2, the mean downwelling (i.e., vertical velocity) is found to be concentrated near the lateral boundary.

In a rotating environment, like the Earth, buoyancy forcing can be applied on the free surface of an ocean basin in the form of cooling and heating as discussed previously, but it can also be applied on the ocean basin boundary in the form of a buoyancy discharge, as it happens in the presence of a river estuary. The fresh water river outflow exiting at the ocean boundary will have a tendency to spread horizontally because of its positive buoyancy (i.e., the river outflow is lighter than the ocean water). However, in the presence of rotation, the buoyant waters cannot spread uniformly, and after a length scale of a few Rossby deformation radii, the Coriolis force turns the buoyant outflow to the right (left) in the Northern (Southern) Hemisphere. The presence of a coastline, then, is fundamental for the generation of buoyant "coastal" currents. Since there is no flow normal to the coast, there is no Coriolis force parallel to the coast, and the resulting motion is along the coast. The buoyant "coastal" current is forced to hug the coastline with the coast on its right (left) looking downstream in the Northern (Southern) Hemisphere. Buoyant coastal currents can be found in many parts of the world's oceans. Particularly striking examples are the Leeuwin Current (Griffith and Pearce 1985; Pearce and Griffith 1991), the Norwegian Coastal Current (Johannessen and Mork 1979), and the East Greenland Current (Wadhams et al. 1979). These buoyancy-driven currents are complex turbulent current systems, as can be seen in satellite images (see figure 5 of Legeckis 1978 or figure 12 of Wadhams et al. 1979). The fronts of these currents delineate boundaries between different water properties, and the stability of these fronts is of fundamental importance for the exchange of water properties (i.e., mixing) across the front (e.g., Cenedese and Linden 2002). Eddies (with scales of 10–100 km) detaching from an unstable current can transport a large volume of water across fronts and can be a very efficient mechanism for transferring water properties from one region to another, affecting, for example, the local fisheries and ecosystems.

The dynamics of buoyant coastal currents have been studied extensively (e.g., O'Donnell 1990; Yankovsky and Chapman 1997; Garvine 2001; Fong and Geyer 2002; Garcia Berdeal et al. 2002; Hetland 2005). Laboratory experiments have provided numerous insights into the dominant scales present in these flows as well as into their dynamics and stability (Griffiths and Linden 1981a; Lentz and Helfrich 2002). Chapter 4 shows that a buoyant coastal current can flow along a coastline in one of two forms: a surface-trapped current or a slope-controlled current. A surface-trapped current forms a shallow layer that intersects the sloping bottom topography close to shore; bottom topography has virtually no effect on its dynamics. Numerous studies have examined the behavior (formation, propagation, stability, etc.) of a surface-trapped current along a vertical wall (over a flat bottom) in various configurations (e.g., Griffiths and Linden 1981a; 1982; Griffiths and Hopfinger 1983; Chabert d'Hières et al. 1991; Garvine 1999; Fong and Geyer 2002; Geyer et al. 2004). A slope-controlled current is fundamentally different because bottom topography plays a leading dynamical role. The dynamics of a slope-controlled current were first described by Chapman and Lentz (1994) and then further investigated by Yankovsky and Chapman (1997) on the basis of numerical model calculations. Lentz and Helfrich (2002) confirmed the existence of the slope-controlled current using a laboratory model and developed a scaling theory that smoothly links the surface-trapped and slope-controlled currents.

The presence of bottom topography introduces a strong topographic β-effect, which influences the stability of buoyant coastal currents. For example, several studies (e.g., Cenedese and Linden 2002; Lentz and Helfrich 2002; Wolfe and Cenedese 2006) have shown that a sloping bottom tends to stabilize a current's front. Buoyant coastal currents over a flat bottom become unstable to nonaxisymmetric disturbances (Griffiths and Linden 1981a) that can be interpreted as a mix of baroclinic and barotropic instabilities. After the instability grows to large amplitude, dipoles form and propagate radially outwards. Griffiths and Linden (1981b) developed a simplified model for baroclinic instability, following Phillips (1954), which qualitatively agrees with their laboratory experiments. Griffiths and Linden (1981a) further developed the model to include frictional dissipation due to Ekman layers. Buoyant coastal currents often flow over a shelf break, where the flatter continental shelf ends and the steeper continental slope begins. The front in the Middle Atlantic Bight is an example of a buoyant coastal current's front flowing over the shelf break. This front, which originates in Nova Scotia and ends near Cape Hatteras, is the barrier between the warm and saline waters of the North Atlantic Ocean and the cooler, fresher waters of the continental shelf (Wright 1976; Linder and Gawarkiewicz 1998).

Buoyant coastal currents are also influenced dramatically by the wind, which can enhance or inhibit the offshore movement and dispersal of such currents. Upwelling, along shelf, wind forcing tends to flatten the current front, causing the buoyant current to thin and widen, whereas moderate downwelling winds steepen the front, causing the current to thicken and narrow. However, strong downwelling winds force vertical mixing that widens the current front, but leaves the current width almost unaltered. Several studies have investigated the response of a buoyant coastal current to wind events (Fong et al. 1997; Fong and Geyer 2001; Garcia Berdeal et al. 2002; Geyer et al. 2004; Lentz 2004; Hetland and Signell 2005) as discussed in detail in Chapter 4.

The densest waters in the oceans are generated at high latitudes where strong atmospheric cooling and the formation of ice with consequent brine rejection contribute to the formation of salty, cold water in the Nordic seas basins and along the continental shelf of Antarctica (Baines and Condie 1998; Ivanov et al. 2004). Dense warm and salty waters are also formed in marginal seas (e.g., the Mediterranean Sea and the Red Sea) where evaporation overcomes the input of fresh water from river runoff and precipitation. This dense water usually moves over a sill and/or through a constriction before generating a dense downslope current (called an overflow) over the continental slope. Major examples of these overflows can be found in the Denmark Strait (Dickson and Brown 1994; Girton and Sanford 2003; Käse et al. 2003), in the Faroe Bank Channel (Saunders 1990; Mauritzen et al. 2005; Fer et al. 2010), in the Baltic Sea (Arneborg et al. 2007; Umlauf and Arneborg 2009), in the Strait of Gibraltar (Baringer and Price 1997; Price et al. 1993), at the mouth of the Red Sea (Peters and Johns 2005; Peters et al. 2005), at various locations along the Arctic (Aagaard et al. 1981), and on the shelves of the Weddell Sea (Foster and Carmack 1976) and Ross Sea (Muench et al. 2009; Padman et al. 2009) in Antarctica. As the dense water descends over the continental slope, its basic dynamics can be described as a dense current driven by buoyancy and influenced by the Coriolis acceleration and bottom drag. These "stream tube models" have been successful at modeling the trajectory of these dense currents over the slope by balancing the above-mentioned forces (Smith 1975; Killworth 1977; Price and Baringer 1994). The dense current motion is deflected to the right when looking downslope in the Northern Hemisphere (to the left in the Southern Hemisphere) because of the Coriolis force. After a transition period, the dense current moves mainly along the slope, with a small downslope component due to the bottom drag. As the current moves over the slope, it entrains ambient water with a consequent decrease in density and increase in transport. The dense current eventually reaches either the bottom of the ocean or a level of neutral buoyancy where the ambient surrounding waters have the same density as the dense current. The dense current is then observed to spread horizontally to fill the bottom of the ocean, as in the case of the water formed on the Weddell Sea continental slope, which forms Antarctic Bottom Water (AABW). Alternatively, if the difference in density with the surrounding water is zero before the current reaches the bottom of the ocean, the current intrudes into the water column at the level of neutral buoyancy to form an intermediate layer, as observed with the Mediterranean overflow. Another dense current reaching a neutrally buoyant level is the product water of the Denmark Strait and Faroe Bank Channel overflows, which flows down the continental slope to the southern tip of Greenland to form North Atlantic Deep Water (NADW). This current is observed to hug the continental slope on the western side of the North Atlantic.

Chapter 5 reviews the fundamental characteristics of these overflows and the subsequent fate of the dense water as it moves over the slope. The final location and depth of these dense waters will depend on the amount of ambient water entrained by the dense current as it moves across a sill/constriction and descends over the continental shelf and slope. At the sill/constriction, the velocity of the currents is usually large compared to the velocity of the ambient waters surrounding them; hence, the currents can generate small-scale instabilities (i.e., turbulence) that lead to the entrainment of lighter ambient water, resulting in a decrease in density and an increase in transport. Strong entrainment is usually present near the location of the sill/constriction where the dense current maximum velocities are typically observed. For example, the Mediterranean overflow has been observed to entrain mainly within 50 km, or half a day, from the current exiting the Strait of Gibraltar, where the dense current velocity reaches its maximum (Price and Baringer 1994). This entrainment determines the main characteristics of temperature and salinity of the current as well as the depth of the neutrally buoyant layer into which the dense current will intrude (approximately 1000 m for the Mediterranean overflow water). However, entrainment has also been observed to occur over the slope where the current's velocity is much lower (Lauderdale et al. 2008; Cenedese et al. 2004; Cenedese and Adduce 2008; 2010). For example, the moderate entrainment that occurs along the Denmark Strait overflow's 1,000-km-long path on the continental slope between the Denmark Strait and Cape Farewell increases the volume transport of the current by the same amount as the intense entrainment that occurrs in the first 100 km near the actual Denmark Strait (Lauderdale et al. 2008). Hence, it is of fundamental importance to consider not only the supercritical entrainment occurring near the sill/constriction but also the subcritical entrainment occurring over the slope (Hughes and Griffiths 2006; Wåhlin and Cenedese 2006; Lauderdale et al. 2008). A correct prediction of the location, depth, and density of the NADW

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originating from the Denmark Strait overflow can be obtained only by considering the entrainment occurring immediately downstream of the sill as well as over the slope. The dynamics regulating the entrainment in a dense current were first investigated by Ellison and Turner (1959) using laboratory experiments and have recently been revisited in the context of overflows using numerical models (Ezer 2005; 2006; Legg et al. 2006; Özgökmen et al. 2006; 2009; Riemenschneider and Legg 2007; Chang et al. 2008) and laboratory experiments (Baines 2001; 2002; 2005; 2008; Cenedese et al. 2004; Wells and Wettlaufer 2005; Wells 2007; Cenedese and Adduce 2008). An accurate representation of the dynamics of buoyancy-driven flows in ocean numerical models is primarily dependent upon the horizontal and vertical resolution choices and the parameterization of unresolved subgrid scale processes (Chapters 6 and 7). In the case of overflows, the models' resolution at best marginally resolves the topography and is not fine enough to represent the small-scale processes related to the entrainment. The parameterization of the unresolved physics is model dependent and has a strong impact on the downstream evolution of water masses (Wu et al. 2007; Danabasoglu et al. 2010). This is an area of active research (Hallberg 2000; Xu et al. 2006; Jackson et al. 2008; Cenedese and Adduce 2010) that directly impacts our ability to represent the long-term evolution of oceanic water masses associated with climate variability and climate change (Chapter 6). Even when the model is eddy resolving (Chapter 7), the dynamics of overflows remain highly dependent on topographic details and eddy dynamics. Mesoscale eddies generated near the sill modulate the entrainment that takes place in the downslope flow of dense water. The simulation of these downslope flows of dense water is nontrivial and differs strongly among ocean models based on different vertical coordinate schemes (Griffies et al. 2000).

There are other areas in oceanography in which "buoyancy-driven" flows are manifested. In particular, much oceanographic and fluid dynamics research has focused on classical convection (e.g., Rayleigh–Bénard) and its oceanic manifestation. Although we have decided not to dedicate an entire chapter to this particular topic, Chapters 3 and 5 do provide an introduction to, as well as some examples of, convective flows.

Dynamics similar to those regulating oceanic buoyancy-driven flows also describe flows in the atmosphere, volcanoes, and avalanches (Chapters 8, 9, and 10, respectively). However, specific aspects of the medium may be fundamental in the dynamics of these different flows. For example, in the atmosphere (Chapter 8), the air is 1,000 times lighter than the water and is much more compressible; hence, some of the simplifications in the equations of motion made for the oceanic flows no longer hold. The introduction of moisture and phase transition adds complexity in the atmosphere. In volcanic flows (Chapter 9), the different phases of the fluids found in a magma chamber, and in particular the gas content in these phases, are important variables in the description of the flow. The erupting buoyant plumes are similar to oceanic buoyant plumes, and a correct representation necessitates including the effects of the particle content, also necessary for the description of turbidity currents in the ocean.

In avalanches, the fluid (air) is mixed with particles in the form of rocks and snow, and the flow can be dominated either by inertia or friction, as discussed in Chapter 10.

In summary, although we have made much progress in improving our understanding of oceanic buoyancy-driven flows, correctly representing buoyancy-driven processes not currently resolved in the ocean component of climate models remains a challenge. The credibility of these models is strongly limited by their ability to represent the climatologically important processes that occur on scales smaller than the climate models' grid scale (currently typically 100 km), and a correct physical parameterization of these buoyancy-driven processes will be essential for an accurate projection of the long-term water mass evolution in these climate models.

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