### CHAPTER

# 1

# Introduction to Descriptive Physical Oceanography

Oceanography is the general name given to the scientific study of the oceans. It is historically divided into physical, biological, chemical, and geological oceanography. This book is concerned primarily with the physics of the ocean, approached mainly, but not exclusively, from observations, and focusing mainly, but not exclusively, on the larger space and timescales of the open ocean rather than on the near-coastal and shoreline regions.

*Descriptive* physical oceanography approaches the ocean through both observations and complex numerical model output used to describe the fluid motions as quantitatively as possible. *Dynamical* physical oceanography seeks to understand the processes that govern the fluid motions in the ocean mainly through theoretical studies and process-based numerical model experiments. This book is mainly concerned with description based in observations (similar to previous editions of this text); however, in this edition we include some of the concepts of dynamical physical oceanography as an important context for the description. A full treatment of dynamical oceanography is contained in other texts. Thermodynamics also clearly enters into our discussion of the ocean through the processes that govern its heat and salt content, and therefore its density distribution.

Chapter 2 describes the ocean basins and their topography. The next three chapters introduce the physical (and some chemical) properties of freshwater and seawater (Chapter 3), an overview of the distribution of water characteristics (Chapter 4), and the sources and sinks of heat and freshwater (Chapter 5). The next three chapters cover data collection and analysis techniques (Chapter 6 and supplemental material listed as Chapter S6 on the textbook Web site http://booksite. academicpress.com/DPO/; "S" denotes supplemental material.), an introduction to geophysical fluid dynamics for graduate students who have varying mathematics backgrounds (Chapter 7), and then basic waves and tides with an introduction to coastal oceanography (Chapter 8). The last six chapters of the book introduce the circulation and water properties of each of the individual oceans (Chapters 9 through 13) ending with a summary of the global ocean in Chapter 14.

Accompanying the text is the Web site mentioned in the previous paragraph. It has four aspects:

- **1.** Textbook chapters on climate variability and oceanographic instrumentation that do not appear in the print version
- Expanded material and additional figures for many other chapters

- **3.** A full set of tutorials for descriptive oceanography with data and sample scripts provided based on the Java Ocean Atlas (Osborne & Swift, 2009)
- **4.** All figures from the text, with many more in color than in the text, for lectures and presentations.

### 1.1. OVERVIEW

There are many reasons for developing our knowledge of the ocean. Near-shore currents and waves affect navigation and construction of piers, breakwaters, and other coastal structures. The large heat capacity of the oceans exerts a significant and in some cases a controlling effect on the earth's climate. The ocean and atmosphere interact on short to long timescales; for example, the El Niño-Southern Oscillation (ENSO) phenomenon that, although driven locally in the tropical Pacific, affects climate on timescales of several years over much of the world. To understand these interactions it is necessary to understand the coupled ocean-atmosphere system. To understand the coupled system, it is first necessary to have a solid base of knowledge about both the ocean and atmosphere separately.

In these and many other applications, knowledge of the ocean's motion and water properties is essential. This includes the major ocean currents that circulate continuously but with fluctuating velocity and position, the variable coastal currents, the rise and fall of the tides, and the waves generated by winds or earthquakes. Temperature and salt content determine density and hence vertical movement. They also affect horizontal movement as the density affects the horizontal pressure distribution. Sea ice has its own full set of processes and is important for navigation, ocean circulation, and climate. Other dissolved substances such as oxygen, nutrients, and other chemical species, and even some of the biological aspects such as chlorophyll content, are used in the study of ocean physics.

Our present knowledge in physical oceanography rests on an accumulation of data, most of which were gathered during the past 150 years, with a large increase of in situ data collection (within the actual water) starting in the 1950s and an order of magnitude growth in available data as satellites began making ocean measurements (starting in the 1970s).

A brief history of physical oceanography with illustrations is provided as supplemental material on the textbook Web site (Chapter 1 supplement is listed as Chapter S1 on the Web site). Historically, sailors have always been concerned with how ocean currents affect their ships' courses as well as changes in ocean temperature and surface condition. Many of the earlier navigators, such as Cook and Vancouver, made valuable scientific observations during their voyages in the late 1700s, but it is generally considered that Mathew Fontaine Maury (1855) started the systematic large-scale collection of ocean current data using ship's navigation logs as his source of information. The first major expedition designed expressly to study all scientific aspects of the oceans was carried out on the British H.M.S. *Challenger* that circumnavigated the globe from 1872 to 1876. The first large-scale expedition organized primarily to gather physical oceanographic data was carried out on the German FS Meteor, which studied the Atlantic Ocean from 1925 to 1927 (Spiess, 1928). A number of photos from that expedition are reproduced on the accompanying Web site. Some of the earliest theoretical studies of the sea included the surface tides by Newton, Laplace, and Legendre (e.g., Wilson, 2002) and waves by Gerstner and Stokes (e.g., Craik, 2005). Around 1896, some of the Scandinavian meteorologists started to turn their attention to the ocean, because dynamical meteorology and dynamical oceanography have many common characteristics. Current knowledge of dynamical oceanography owes its progress to the work of Bjerknes, Bjerknes, Solberg, and Bergeron (1933), Ekman (1905, 1923), Helland-Hansen (1934), and others.

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#### SPACE AND TIMESCALES OF PHYSICAL OCEANOGRAPHIC PHENOMENA

The post-war 1940s through 1960s began to produce much of the data and especially theoretical understanding for large-scale ocean circulation. With the advent of moored and satellite instrumentation in the 1960s and 1970s, the smaller scale, energetically varying part of the ocean circulation — the mesoscale — began to be studied in earnest. Platforms expanded from research and merchant ships to global satellite and autonomous instrument sampling. Future decades should take global description and modeling to even smaller scales (submesoscale) as satellite observations and numerical modeling resolution continue to evolve, and different types of autonomous sampling within the water column become routine. Physical oceanography has retained an aspect of individual exploration but large, multi-investigator, multinational programs have increasingly provided many of the new data sets and understanding. Current research efforts in physical oceanography are focused on developing an understanding of the variability of the ocean and its relation to the atmosphere and climate as well as continuing to describe its steady-state conditions.

### 1.2. SPACE AND TIMESCALES OF PHYSICAL OCEANOGRAPHIC PHENOMENA

The ocean is a fluid in constant motion with a very large range of spatial and temporal scales. The complexity of this fluid is nicely represented in the sea surface temperature image of the Gulf Stream captured from a satellite shown in Figure 1.1a. The Gulf Stream is the western boundary current of the permanent, large-scale clockwise *gyre* circulation of the subtropical North Atlantic. The Gulf Stream has a width of 100 km, and its gyre has a spatial scale of thousands of kilometers. The narrow, warm core of the Gulf Stream (red in Figure 1.1a) carries warm subtropical water northward from the Caribbean, loops through the Gulf of Mexico around Florida and northward along the east coast of North America, leaves the coast at Cape Hatteras, and moves out to sea. Its strength and temperature contrast decay eastward. Its large meanders and rings, with spatial scales of approximately 100 km, are considered mesoscale (eddy) variability evolving on timescales of weeks. The satellite image also shows the general decrease in surface temperature toward the north and a large amount of smallscale eddy variability. The permanence of the Gulf Stream is apparent when currents and temperatures are averaged in time. Averaging makes the Gulf Stream appear wider, especially after the separation at Cape Hatteras where the wide envelope of meanders creates a wide, weak average eastward flow.

The Gulf Stream has been known and charted for centuries, beginning with the Spanish expeditions in the sixteenth century (e.g., Peterson, Stramma, & Kortum 1996). It was first mapped accurately in 1769 by Benjamin Franklin



**FIGURE 1.1** (a) Sea surface temperature from a satellite advanced very high resolution radiometer (AVHRR) instrument (Otis Brown, personal communication, 2009). This figure can also be found in the color insert.



FIGURE 1.1 (b) Franklin-Folger map of the Gulf Stream. Source: From Richardson (1980a).

working together with whaling captain Timothy Folger (Figure 1.1b; from Richardson, 1980a).<sup>1</sup> The narrow current along the coast of the United States is remarkably accurate. The widening envelope of the Franklin/Folger current after separation from Cape Hatteras is an accurate depiction of the envelope of meandering apparent in the satellite image. When time-mean averages of the Gulf Stream based on modern measurements are constructed, they look remarkably similar to this Franklin/Folger map.

The space and timescales of many of the important physical oceanography processes are shown schematically in Figure 1.2. At the smallest scale is molecular mixing. At small,

<sup>&</sup>lt;sup>1</sup> Franklin noted on his frequent trips between the United States and Europe that some trips were considerably quicker than others. He decided that this was due to a strong ocean current flowing from the west to the east. He observed marked changes in surface conditions and reasoned that this ocean current might be marked by a change in sea surface temperature. Franklin began making measurements of the ocean surface temperature during his travels. Using a simple mercury-in-glass thermometer, he was able to determine the position of the current.



FIGURE 1.2 Time and space scales of physical oceanographic phenomena from bubbles and capillary waves to changes in ocean circulation associated with Earth's orbit variations.

macroscopic scales of centimeters, microstructure (vertical layering at the centimeter level) and capillary waves occur. At the slightly larger scale of meters surface waves are found, which have rapid timescales and somewhat long-lived vertical layers (fine structure). At scales of tens of meters are the internal waves with timescales of up to a day. Tides have the same timescales as internal waves, but much larger spatial scales of hundreds to thousands of kilometers. Surface waves, internal waves, and tides are described in Chapter 8.

Mesoscale eddies and strong ocean currents such as the Gulf Stream are found at spatial scales of tens of kilometers to several hundred kilometers and timescales of weeks to years (Figure 1.1a,b). The large-scale ocean circulation has a spatial scale the size of ocean basins up to the global ocean and a timescale ranging from seasonal to permanent, which is the timescale of plate tectonics that rearranges the ocean boundaries (Chapter 2). The timescales for wind-driven and thermohaline circulation in Figure 1.2 are actually the same for circulation of the flow through those systems (ten years around the gyre, hundreds of years through the full ocean); these are time-mean features of the ocean and have much longer timescales. Climate variability affects the ocean, represented in Figure 1.2 by the El Niño, which has an interannual timescale (several years; Chapter 10); decadal and longer timescales of variability of the ocean circulation and properties are also important and described in each of the ocean basin and global circulation chapters.

We see in Figure 1.2 that short spatial scales generally have short timescales, and long spatial scales generally have long timescales. There are some exceptions to this, most notably in the tides and tsunamis as well as in some fine-structure phenomena with longer timescales than might be expected from their short spatial scales.

In Chapter 7, where ocean dynamics are discussed, some formal *non-dimensional parameters* incorporating the approximate space and timescales for these different types of phenomena are introduced (see also Pedlosky, 1987). A non-dimensional parameter is the ratio of dimensional parameters with identical dimensional scales, such as time, length, mass, etc., which are intrinsic properties of the flow phenomenon being described or modeled. Of special importance is whether the timescale of an ocean motion is greater than or less than about a day, which is the timescale for the earth's rotation. Earth's rotation has an enormous effect on how the ocean water moves in response to a force; if the force and motion are sustained for days or longer, then the motion is strongly influenced by the rotation. Therefore, an especially useful parameter is the ratio of the timescale of Earth's rotation to the timescale of the motion. This ratio is called the Rossby number. For the very small, fast motions in Figure 1.2, this ratio is large and rotation is not important. For the slow, large-scale part of the spectrum, the Rossby number is small and Earth's rotation is fundamental. A second very important nondimensional parameter is the ratio of the vertical length scale (height) to the horizontal length scale; this is called the *aspect ratio*. For large-scale flows, this ratio is very small since the vertical scale can be no larger than the ocean depth. For surface and internal gravity waves, the aspect ratio is order 1. We will also see that dissipation is very weak in the sense that the timescale for dissipation to act is long compared with both the timescale of Earth's rotation and the timescale for the circulation to move water from one place to another; the relevant non-dimensional parameters are the Ekman number and Reynolds number, respectively. Understanding how the small Rossby number, small aspect ratio, and nearly frictionless fluid ocean behaves has depended on observations of the circulation and water properties made over the past century. These are the principal subjects of this text.

### CHAPTER

# 2

## Ocean Dimensions, Shapes, and Bottom Materials

### 2.1. DIMENSIONS

The oceans are basins in the surface of the solid earth containing salt water. This chapter introduces some nomenclature and directs attention to features of the basins that have a close connection with the ocean's circulation and dynamical processes that are of importance to the physical oceanographer. More detailed descriptions of the geology and geophysics of the ocean basins are given in Seibold and Berger (1982), Kennett (1982), Garrison (2001), and Thurman and Trujillo (2002), among others. Updated data sets, maps, and information are available from Web sites of the National Geophysical Data Center (NGDC) of the National Oceanic and Atmospheric Administration (NOAA) and from the U.S. Geological Survey (USGS).

The major ocean areas are the Atlantic Ocean, the Pacific Ocean, the Indian Ocean, the Arctic Ocean, and the Southern Ocean (Figure 2.1). The first four are clearly divided from each other by land masses, but the divisions between the Southern Ocean and oceans to its north are determined only by the characteristics of the ocean waters and their circulations. The geographical peculiarities of each ocean are described in Section 2.11.

The shape, depth, and geographic location of an ocean affect the general characteristics of its circulation. Smaller scale features, such as locations of deep sills and fracture zones, seamounts, and bottom roughness, affect often important details of the circulation and of mixing processes that are essential to forcing and water properties. The Atlantic has a very marked "S" shape while the Pacific has a more oval shape. The Atlantic and Indian Oceans are roughly half the eastwest width of the Pacific Ocean, which impacts the way that each ocean's circulation adjusts to changes in forcing. The Indian Ocean has no high northern latitudes, and therefore no possibility of cold, dense water formation. The edges of the Pacific are ringed with trenches, volcanoes, and earthquakes that signal the gradual descent of the ocean bottom crustal "plates" under the surrounding continental plates. In contrast, the Atlantic is the site of dynamic seafloor spreading as material added in the center of the Mid-Atlantic Ridge (MAR) pushes the plates apart, enlarging the Atlantic Ocean by a few centimeters each year.

*Marginal seas* are fairly large basins of salt water that are connected to the open ocean by one or more fairly narrow channels. Those that are connected by very few channels are sometimes called *mediterranean seas* after the



FIGURE 2.1 Map of the world based on ship soundings and satellite altimeter derived gravity at 30 arc-second resolution. Data from Smith & Sandwell (1997); Becker et al. (2009); and SIO (2008).

prototype, the (European) Mediterranean Sea. The Mediterranean provides an example of a negative water balance in a sea with less inflow (river runoff and precipitation) than evaporation. An excellent example of a positive water balance marginal sea (with net precipitation) is found in the Black Sea, which connects with the Mediterranean Sea. Both of these seas are discussed further in Chapters 5 and 9. Other examples of marginal seas that are separated from the open ocean by multiple straits or island chains are the Caribbean Sea, the Sea of Japan, the Bering Sea, the North Sea, the Baltic Sea, and so forth.

The term *sea* is also used for a portion of an ocean that is not divided off by land but has local distinguishing oceanographic characteristics; for example the Norwegian Sea, the Labrador Sea, the Sargasso Sea, and the Tasman Sea.

More of the earth's surface is covered by sea than by land, about 71% sea to 29% land. (The most recent elevation data for the earth's surface, used to construct Figure 2.2, show that 70.96% of the earth is ocean; see Becker et al., 2009.) Furthermore, the proportion of water to land in the Southern Hemisphere is much greater (4:1) than in the Northern Hemisphere (1.5:1). In area, the Pacific Ocean is about as large as the Atlantic and Indian Oceans combined. If the neighboring sectors of the Southern Ocean are included with the three main oceans north of it, the Pacific Ocean occupies about 46% of the total world ocean area, the Atlantic Ocean about 23%, the Indian Ocean about 20%, and the rest, combined, about 11%.

The average depth of the oceans is close to 4000 m while the marginal seas are generally about 1200 m deep or less. Relative to sea level,



FIGURE 2.2 Areas of Earth's surface above and below sea level as a percentage of the total area of Earth (in 100 m intervals). Data from Becker et al. (2009).

the oceans are much deeper than the land is high. While only 11% of the land surface of the earth is more than 2000 m above sea level, 84% of the sea bottom is more than 2000 m deep. However, the maxima are similar: the height of Mt. Everest is about 8848 m, while the maximum depth in the oceans is 11,034 m in the Mariana Trench in the western North Pacific. Figure 2.2 shows the distributions of land elevations and of sea depths relative to sea level in 100 m intervals as the percentage of the total area of the earth's surface. This figure is based on the most recent global elevation and ocean bathymetry data from D. Sandwell (Becker et al., 2009). (It is similar to Figure 2.2 using 1000 m bins that appeared in previous editions of this text, based on data from Kossina, 1921 and Menard & Smith, 1966, but the 100 m bins allow much more differentiation of topographic forms.)

Although the average depth of the oceans, 4 km, is a considerable distance, it is small compared with the horizontal dimensions of the oceans, which are 5000 to 15,000 km. Relative to the major dimensions of the earth, the oceans are a thin skin, but between the sea surface and the bottom of the ocean there is a great deal of detail and structure.

### 2.2. PLATE TECTONICS AND DEEP-SEA TOPOGRAPHY

The most important geophysical process affecting the shape and topography of the ocean

basins is the movement of the earth's tectonic plates, described thoroughly in texts such as Thurman and Trujillo (2002). The plate boundaries are shown in Figure 2.3. Seafloor spreading creates new seafloor as the earth's plates spread apart. This creates the mid-ocean ridge system; the mid-ocean ridges of Figure 2.1 correspond to plate boundaries. The ocean plates spread apart at rates of about 2 cm/year (Atlantic) to 16 cm/year (Pacific), causing extrusion of magma into the surface at the centers of the ridges. Over geologic time the orientation of the earth's magnetic field has reversed, causing the ferromagnetic components in the molten new surface material to reverse. Spreading at the mid-ocean ridge was proven by observing the reversals in the magnetic orientations in the surface material. These reversals permit dating of the seafloor (Figure 2.3). The recurrence interval for magnetic reversals is approximately 500,000 to 1,000,000 years.

The 14,000 km long MAR is a tectonic *spreading center*. It is connected to the global mid-ocean ridge, which at more than 40,000 km long, is the most extensive feature of the earth's topography. Starting in the Arctic Ocean, the mid-ocean ridge extends through Iceland down the middle of the Atlantic, wraps around the tip of Africa, and then winds through the Indian and Pacific Oceans, ending in the Gulf of California. In all oceans, the mid-ocean ridge and other deep ridges separate the bottom waters, as can be seen from different water properties east and west of the ridge.

Deep and bottom waters can leak across the ridges through narrow gaps, called *fracture zones*, which are lateral jogs in the spreading center. The fracture zones are roughly vertical planes, perpendicular to the ridge, on either side of which the crust has moved in opposite directions perpendicular to the ridge. There are many fracture zones in the mid-ocean



FIGURE 2.3 Sea floor age (millions of years). Black lines indicate tectonic plate boundaries. Source: From Müller, Sdrolias, Gaina, and Roest (2008).

ridges. One example that is important as a pathway for abyssal circulation in the Atlantic is the Romanche Fracture Zone through the MAR close to the equator. Another example is the pair of fracture zones in the South Pacific (Eltanin and Udintsev Fracture Zones, Figure 2.12) that steer the Antarctic Circumpolar Current (ACC).

At some edges of the tectonic plates, one plate subducts (moves under) another. Subduction is accompanied on its landward side by volcanoes and earthquakes. Subduction creates deep *trenches* that are narrow relative to their length and have depths to 11,000 m. The deepest parts of the oceans are in these trenches. The majority of the deep trenches are in the Pacific: the Aleutian, Kurile, Tonga, Philippine, and Mariana. There are a few in other oceans such as the Puerto Rico and the South Sandwich Trenches in the Atlantic and the Sunda Trench in the Indian Ocean. Trenches are often shaped like an arc of a circle with an *island arc* on one side. Examples of island arcs are the Aleutian Islands (Pacific), the Lesser Antilles (Atlantic), and the Sunda Arc (Indian). The landward side of a trench extends as much as 10,000 m from the trench bottom to the sea surface, while the other side is only half as high, terminating at the ocean depth of about 5000 m.

Trenches can steer or impact boundary currents that are in deep water (Deep Western Boundary Currents) or upper ocean boundary currents that are energetic enough to extend to the ocean bottom, such as western boundary currents of the wind-driven circulation. Examples of trenches that impact ocean circulation are the deep trench system along the western and northern boundary of the Pacific and the deep trench east of the Caribbean Sea in the Atlantic.

Younger parts of the ocean bottom are shallower than older parts. As the new seafloor created at seafloor spreading centers ages, it cools by losing heat into the seawater above and becomes denser and contracts, which causes it to be deeper (Sclater, Parsons, & Jaupart, 1981). Ocean bottom depths range from 2 to 3 km for the newest parts of the mid-ocean ridges to greater than 5 km for the oldest, as can be seen by comparing the maps of seafloor age and bathymetry (Figures 2.1 and 2.3).

The rate of seafloor spreading is so slow that it has no impact on the climate variability that we experience over decades to millennia, nor does it affect anthropogenic climate change. However, over many millions of years, the geographic layout of Earth has changed. The paleocirculation patterns of "deep time," when the continents were at different locations, differed from the present patterns; reconstruction of these patterns is an aspect of paleoclimate modeling. By studying and understanding present-day circulation, we can begin to credibly model the paleocirculation, which had the same physical processes (such as those associated with the earth's rotation, wind and thermohaline forcing, boundaries, open east-west channels, equatorial regions, etc.), but with different ocean basin shapes and bottom topography.

Ocean bottom roughness affects ocean mixing rates (Sections 7.2.4 and 7.3.2). The overall roughness varies by a factor of 10. Roughness is a function of spreading rates and sedimentation rates. New seafloor is rougher than old Slow-spreading centers seafloor. produce rougher topography than fast spreading centers. Thus the slow-spreading MAR is rougher than the faster spreading East Pacific Rise (EPR; Figure 2.4). Slow-spreading ridges also have rift valleys at the spreading center, whereas fast-spreading ridges have an elevated ridge at the spreading center. Much of the roughness on the ridges can be categorized as abyssal hills, which are the most common landform on Earth. Abyssal hills are evident in Figures 2.4 and 2.5b, all along the wide flanks of the mid-ocean ridge.

Individual mountains (*seamounts*) are widely distributed in the oceans. Seamounts stand out clearly above the background bathymetry. In



**FIGURE 2.4** Seafloor topography for a portion of (a) the fast-spreading EPR and (b) the slow-spreading MAR. Note the ridge at the EPR spreading center and rift valley at the MAR spreading center. This figure can also be found in the color insert. (Sandwell, personal communication, 2009.)

the maps in Figure 2.4b there are some seamounts on the upper right side of the panel. In the vertical cross-section in Figure 2.5b, seamounts are distinguished by their greater height compared with the abyssal hills. The average height of seamounts is 2 km. Seamounts that reach the sea surface form islands. A *guyot* is a seamount that reached the surface, was worn flat, and then sank again below the surface. Many seamounts and islands were created by volcanic *hotspots* beneath the tectonic plates. The hotspots are relatively stationary in contrast to the plates and as the plates move across the hotspots, chains of seamounts are formed. Examples

include the Hawaiian Islands/Emperor Seamounts chain, Polynesian island chains, the Walvis Ridge, and the Ninetyeast Ridge in the Indian Ocean.

Seamounts affect the circulation, especially when they appear in groups as they do in many regions; for instance, the Gulf Stream passes through the New England Seamounts, which affect the Gulf Stream's position and variability (Section 9.3). Seamount chains also refract tsunamis, which are ocean waves generated by submarine earthquakes that react to the ocean bottom as they propagate long distances from the earthquake source (Section 8.3.5). PLATE TECTONICS AND DEEP-SEA TOPOGRAPHY



FIGURE 2.5 (a) Schematic section through ocean floor to show principal features. (b) Sample of bathymetry, measured along the South Pacific ship track shown in (c).

### 2.3. SEAFLOOR FEATURES

The continents form the major lateral boundaries to the oceans. The detailed features of the shoreline and of the sea bottom are important because of the way they affect circulation. Starting from the land, the main divisions recognized are the shore, the continental shelf, the continental slope and rise, and the deep-sea bottom, part of which is the abyssal plain (Figure 2.5a, b). Some of the major features of the seafloor, including mid-ocean ridges, trenches, island arcs, and seamounts, are the result of plate tectonics and undersea volcanism (Section 2.2 and Figure 2.3).

In some of the large basins the seafloor is very smooth, possibly more so than the plains areas on land. Sedimentation, which is mostly due to the incessant rain of organic matter from the upper ocean, covers the rough bottom and produces large regions of very smooth topography. Stretches of the abyssal plain in the western North Atlantic have been measured to be smooth within 2 m over distances of 100 km. The ocean bottom in the northeast Indian Ocean/Bay of Bengal slopes very smoothly from 2000 m down to more than 5000 m over 3000 km. This smoothness is due to sedimentation from the Ganges and Brahmaputra Rivers that drain the Himalayas. Bottom sediments can be moved around by deep currents; formation of undersea dunes and canyons is common. Erosional features in deep sediments have sometimes alerted scientists to the presence of deep currents.

Bottom topography often plays an important role in the distribution of water masses and the location of currents. For instance, bottom water coming from the Weddell Sea (Antarctica) is unable to fill the eastern part of the Atlantic basin directly due to the height of the Walvis Ridge (South Atlantic Ocean). Instead, the bottom water travels to the north along the western boundary of the South Atlantic, finds a deep passage in the MAR, and then flows south to fill the basin east of the ridge. At shallower depths the *sills* (shallowest part of a channel) defining the marginal seas strongly influences both the mid-level currents and the distribution of water masses associated with the sea. Coastal upwelling is a direct consequence of the shape of the coast and its related bottom topography. Alongshore currents are often determined by the coastal bottom topography and the instabilities in this system can depend on the horizontal scales of the bottom topography. Near the shore bottom topography dictates the breaking of surface gravity waves and also directly influences the local tidal expressions.

Much of the mixing in the ocean occurs near the boundaries (including the bottom). Microstructure observations in numerous regions, and intensive experiments focused on detection of mixing and its genesis, suggest that flow of internal tides over steep bottom slopes in the deep ocean is a major mechanism for dissipating the ocean's energy. Ocean bottom slopes computed from bathymetry collected along ship tracks show that the largest slopes tend to occur on the flanks of the fastest spreading mid-ocean ridges. With bathymetric slopes computed from the most recent bathymetric data (Figure 2.6) and information about the ocean's deep stratification, it appears the flanks of the mid-ocean ridges of the Atlantic, Southern Ocean, and Indian Ocean could be the most vigorous dissipation sites of ocean energy (Becker & Sandwell, 2008).

### 2.4. SPATIAL SCALES

Very often some of the characteristics of the ocean are presented by drawing a vertical cross-section of a part of the oceans, such as the schematic depiction of ocean floor features in Figure 2.5a. An illustration to true scale would have the relative dimensions of the edge of a sheet of paper and would be either



FIGURE 2.6 Mean slope of the ocean bottom, calculated from shipboard bathymetry and interpolated to a 0.5 degree grid. *Source: From Becker and Sandwell (2008)*.

too thin to show details or too long to be convenient. Therefore, we usually distort our crosssection by making the vertical scale much larger than the horizontal one. For instance, we might use a scale of 1 cm to represent 100 km horizontally while depths might be on a scale of 1 cm to represent 100 m (i.e., 0.1 km). In this case the vertical dimensions on our drawing would be magnified 1000 times compared with the horizontal ones (a *vertical exaggeration* of 1000:1). This gives us room to show the detail, but it also exaggerates the slope of the sea bottom or of contours of constant water properties (*isopleths*) drawn on the cross-section (Figure 2.5b). In reality, such slopes are far less than they appear on the cross-section drawings; for instance, a line of constant temperature (*isotherm*) with a real slope of 1 in 10,000 would appear as a slope of 1 in 10 on the plot.

### 2.5. SHORE, COAST, AND BEACH

The *shore* is defined as that part of the landmass, close to the sea, that has been modified by the action of the sea. The words shore and *coast* have the same meaning. Shorelines (coasts) shift over time because of motion of the land over geologic time, changes in sea level, and erosion and deposition. The sedimentary record shows a series of marine intrusions and retreats corresponding to layers that reflect periods when the surface was above and below sea level. Variations in sea level between glacial and interglacial periods have been as much as 120 m. The ability of the coast to resist the erosional forces of the ocean depends directly on the type of material that makes up the coast. Sands are easily redistributed by the ocean currents whereas granitic coasts are slow to erode. Often sea level changes are combined with the hydrologic forces of an estuary, which dramatically change the dynamical relationship between the ocean and the solid surface.

The *beach* is a zone of unconsolidated particles at the seaward limit of the shore and extends roughly from the highest to the lowest tide levels. The landward limit of the beach might be vegetation, permanent sand dunes, or human construction. The seaward limit of a beach, where the sediment movement on- and offshore ceases, is about 10 m deep at low tide.

Coasts can be classified in many different ways. In terms of long timescales (such as those of plate tectonics; Section 2.2), coasts and continental margins can be classified as active or passive. Active margins, with active volcanism, faulting, and folding, are like those in much of the Pacific and are rising. Passive margins, like those of the Atlantic, are being pushed in front of spreading seafloor, are accumulating thick wedges of sediment, and are generally falling. Coasts can be referred to as erosional or depositional depending on whether materials are removed or added. At shorter timescales, waves and tides cause erosion and deposition. At millennial timescales, changes in mean sea level cause materials to be removed or added. Erosional coasts are attacked by waves and currents, both of which carry fine material that abrades the coast. The waves create alongshore and rip currents (Section 8.3) that carry the abraded material from the coastline along and out to sea. This eroded material can be joined by sediments discharged from rivers and form deltas. This type of erosion is fastest on *highenergy* coasts with large waves, and slowest on *low-energy* coasts with generally weak wave fields. Erosion occurs more rapidly in weaker materials than in harder components. These variations in materials allow erosive forces to carve characteristic features on coastlines such as sea cliffs and sea caves, and to create an alternation between bays and headlands.

Beaches result when sediment, usually sand, is transported to places suitable for continued deposition. Again these are often the quiet bays between headlands and other areas of low surf activity. Often a beach is in equilibrium; new sand is deposited to replace sand that is scoured away. Evidence for this process can be seen by how sand accumulates against new structures built on the shore, or by how it is removed from a beach when a breakwater is built that cuts off the supply of sand beyond it. On some beaches, the sand may be removed by currents associated with high waves at one season of the year and replaced by different currents associated with lower waves at another season. These currents are influenced by seasonal and interannual wind variations.

Sea level, which strongly affects coasts, is affected by the total amount of water in the ocean, changes in the containment volume of the world's ocean, and changes in the temperature/salinity characteristics of the ocean that alter its density and hence cause the water to expand or contract. Changes in the total amount of water are due primarily to changes in the volume of landfast ice, which is contained in ice sheets and glaciers. (Because sea ice floats in water, changes in sea ice volume, such as that in the Arctic or Antarctic, do not affect sea level.) Changes in containment volume are due to tectonics, the slow rebound of continents (continuing into the present) after the melt of landfast ice after the last deglaciation, and rebound due to the continuing melt of glacial ice. Changes in heat content cause seawater to expand (heating) or contract (cooling).

DEEP OCEAN

Sea level rose 20 cm from 1870 to 2003, including 3 cm in just the last 10 years (1993–2003). Because good global observations are available for that last 10 years, it is possible to ascribe 1.6 cm to thermal expansion, 0.4 cm to Greenland and Antarctic ice sheet melt, and 0.8 cm to other glacial melt with a residual of 0.3 cm. Sea level is projected to rise  $30 \pm 10$  cm in the next 100 years mainly due to warming of the oceans, which absorb most of the anthropogenic heat increase in the earth's climate system. (See Bindoff et al., 2007 in the 4th assessment report of the Intergovernmental Panel on Climate Change.)

### 2.6. CONTINENTAL SHELF, SLOPE, AND RISE

The continental shelf extends seaward from the shore with an average gradient of 1 in 500. Its outer limit (the shelf break) is set where the gradient increases to about 1 in 20 (on average) to form the continental slope down to the deep sea bottom. The shelf has an average width of 65 km. In places it is much narrower than this, while in others, as in the northeastern Bering Sea or the Arctic shelf off Siberia, it is as much as ten times this width. The bottom material is dominantly sand with less common rock or mud. The shelf break is usually clearly evident in a vertical cross-section of the sea bottom from the shore outward. The average depth at the shelf break is about 130 m. Most of the world's fisheries are located on the continental shelves for a multitude of reasons including proximity of estuaries, depth of penetration of sunlight compared with bottom depth, and upwelling of nutrient-rich waters onto some shelves, particularly those off western coasts.

The *continental slope* averages about 4000 m vertically from the shelf to the deep-sea bottom, but in places extends as much as 9000 m vertically in a relatively short horizontal distance. In general, the continental slope is considerably

steeper than the slopes from lowland to highland on land. The material of the slope is predominantly mud with some rock outcrops. The shelf and slope typically include submarine canyons, which are of worldwide occurrence. These are valleys in the slope, either V-shaped or with vertical sides, and are usually found off coasts with rivers. Some, usually in hard granitic rock, were originally carved as rivers and then submerged, such as around the Mediterranean and southern Baja, California. Others, commonly in softer sedimentary rock, are formed by turbidity currents described in the next paragraph. The lower part of the slope, where it grades into the deep-sea bottom, is referred to as the continental rise.

*Turbidity currents* (Figure 2.7) are common on continental slopes. These episodic events carry a mixture of water and sediment and are driven by the unstable sediments rather than by forces within the water. In these events, material builds up on the slope until it is no longer stable and the force of gravity wins out. Large amounts of sediment and bottom material crash down the slope at speeds up to 100 km/h. These events can snap underwater cables. The precise conditions that dictate when a turbidity current occurs vary with the slope of the valley and the nature of the material in the valley. Turbidity currents carve many of the submarine canyons found on the slopes. Some giant rivers, such as the Congo, carry such a dense load of suspended material that they form continuous density flows of turbid water down their canyons.

### 2.7. DEEP OCEAN

From the bottom of the continental slope, the bathymetric gradient decreases down the continental rise to the *deep-sea bottom*, the last and most extensive area. Depths of 3000–6000 m are found over 74% of the ocean basins with 1% deeper. Perhaps the most characteristic



FIGURE 2.7 Turbidity current evidence south of Newfoundland resulting from an earthquake in 1929. Source: From Heezen, Ericson, and Ewing (1954).

aspect of the deep-sea bottom is the variety of its topography. Before any significant deep ocean soundings were available, the sea bottom was regarded as uniformly smooth. When detailed sounding started in connection with cable laying, it became clear that this was not the case and there was a swing to regarding the sea bottom as predominantly rugged. Neither view is exclusively correct, for we now know that there are mountains, valleys, and plains on the ocean bottom, just as on land. With the advent of satellite altimetry for mapping ocean topography, we now have an excellent global view of the distribution of all of these features (e.g., Figure 2.1; Smith & Sandwell, 1997), and can relate many of the features to plate tectonic processes (Section 2.2) and sedimentation sources and processes.

### 2.8. SILLS, STRAITS, AND PASSAGES

Sills, straits, and passages connect separate ocean regions. A *sill* is a ridge, above the average bottom level in a region, which separates one basin from another or, in the case of a fjord (Section 5.1), separates the landward basin from the sea outside. The *sill depth* is the depth from the sea surface to the deepest part of the ridge; that is, the maximum depth at which direct flow across the sill is possible. An oceanic sill is like a topographic saddle with the sill depth analogous to the saddlepoint. In the deep ocean, sills connect deep basins. The sill depth controls the density of waters that can flow over the ridge.

Straits, passages, and channels are horizontal constrictions. It is most common to refer to a strait when considering landforms, such as the Strait of Gibraltar that connects the Mediterranean Sea and the Atlantic Ocean, or the Bering Strait that connects the Bering Sea and the Arctic Ocean. Passages and channels can also refer to submarine topography, such as in fracture zones that connect deep basins. Straits and sills can occur together, as in both of these examples. The minimum width of the strait and the maximum depth of the sill can hydraulically control the flow passing through the constriction.

### 2.9. METHODS FOR MAPPING BOTTOM TOPOGRAPHY

Our present knowledge of the shape of the ocean floor results from an accumulation of sounding measurements (most of which have been made within the last century) and, more recently, using the gravity field measured by satellites (Smith & Sandwell, 1997). The early measurements were made by lowering a weight

on a measured line until the weight touched bottom, as discussed in Chapter S1, Section S1.1 located on the textbook Web site http:// booksite.academicpress.com/DPO/; "S" denotes supplemental material. This method was slow; in deep water it was uncertain because it was difficult to tell when the weight touched the bottom and if the line was vertical.

Since 1920 most depth measurements have been made with *echo sounders*, which measure the time taken for a pulse of sound to travel from the ship to the bottom and reflect back to the ship. One half of this time is multiplied by the average speed of sound in the seawater under the ship to give the depth. With presentday equipment, the time can be measured very accurately and the main uncertainty over a flat bottom is in the value used for the speed of sound. This varies with water temperature and salinity (see Section 3.7), and if these are not measured at the time of sounding an average value must be used. Research and military ships are generally outfitted with echo sounders and routinely report their bathymetric data to data centers that compile the information for bathymetric mapping. The bathymetry along the research ship track in Figure 2.5b was measured using this acoustic method.

The modern extension of these single echo sounders is a multi-beam array, in which many sounders are mounted along the bottom of the ship; these provide two-dimensional "swath" mapping of the seafloor beneath the ship.

Great detail has been added to our knowledge of the seafloor topography by satellite measurements. These satellites measure the earth's gravity field, which depends on the local mass of material. These measurements allow mapping of many hitherto unknown features, such as fracture zones and seamounts in regions remote from intensive echo sounder measurements, and provide much more information about these features even where they had been mapped (Smith & Sandwell, 1997). Echo sounder measurements are still needed to verify the gravity-based measurements since the material on the ocean bottom is not uniform; for instance, a bottom shaped by extensive sediment cover might not be detected from the gravity field. The bathymetry shown in Figure 2.5c, as well as in the global and basin maps of Figures 2.1 and Figures 2.8–2.12, is a blended product of all available shipboard measurements and satellite-based measurements.

### 2.10. BOTTOM MATERIAL

On the continental shelf and slope most of the bottom material comes directly from the land, either brought down by rivers or blown by the wind. The material of the deep-sea bottom is often more fine-grained than that on the shelf or slope. Much of it is pelagic in character, that is, it has been formed in the open ocean. The two major deep ocean sediments are "red" clay and the biogenic "oozes." The former has less than 30% biogenic material and is mainly mineral in content. It consists of fine material from the land (which may have traveled great distances through the air before finally settling into the ocean), volcanic material, and meteoric remains. The oozes are over 30% biogenic and originate from the remains of living organisms (plankton). The calcareous oozes have a high percentage of calcium carbonate from the shells of animal plankton, while the siliceous oozes have a high proportion of silica from the shells of silica-secreting planktonic plants and animals. The siliceous oozes are found mainly in the Southern Ocean and in the equatorial Pacific. The relative distribution of calcareous and siliceous oozes is clearly related to the nutrient content of the surface waters, with calcareous oozes common in low nutrient regions and siliceous oozes in high nutrient regions.

Except when turbidity currents deposit their loads on the ocean bed, the average rate of deposition of the sediments is from 0.1 to 10 mm per

1000 years, and a large amount of information on the past history of the oceans is stored in them. Samples of bottom material are obtained with a "corer," which is a 2-30 m long steel pipe that is lowered vertically and forced to penetrate into the sediments by the heavy weight at its upper end. The "core" of sediment retained in the pipe may represent material deposited from 1000 to 10 million years per meter of length. Sometimes the material is layered, indicating stages of sedimentation of different materials. In some places, layers of volcanic ash are related to historical records of eruptions; in others, organisms characteristic of cold or warm waters are found in different layers and suggest changes in temperature of the overlying water during the period represented by the core. In some places gradations from coarse to fine sediments in the upward direction suggest the occurrence of turbidity currents bringing material to the region with the coarser material settling out first and the finer later.

Large sediment depositions from rivers create a sloping, smooth ocean bottom for thousands of miles from the mouths of the rivers. This is called a deep-sea sediment *fan*. The largest, the Bengal Fan, is in the northeastern Indian Ocean and is created by the outflow from many rivers including the Ganges and Brahmaputra. Other examples of fans are at the mouths of the Yangtze, Amazon, and Columbia Rivers.

Physical oceanographers use sediments to help trace movement of the water at the ocean floor. Some photographs of the deep-sea bottom show ripples similar to a sand beach after the tide has gone out. Such ripples are only found on the beach where the water speed is high, such as in the backwash from waves. We conclude from the ripples on the deep-ocean bottom that currents of similar speed occur there. This discovery helped to dispel the earlier notion that all deep-sea currents are very slow.

Sediments can affect the properties of seawater in contact with them; for instance,

silicate and carbonate are dissolved from sediments into the overlying seawater. Organic carbon, mainly from fecal pellets, is biologically decomposed (remineralized) into inorganic carbon dioxide in the sediments with oxygen consumed in the process. The carbon dioxiderich, oxygen-poor pore waters in the sediments are released back into the seawater, affecting its composition. Organic nitrogen and phosphorus are also remineralized in the sediments, providing an important source of inorganic nutrients for seawater. In regions where all oxygen is consumed, methane forms from bacterial action. This methane is often stored in solid form called a methane hydrate. Vast quantities (about 10<sup>19</sup> g) of methane hydrate have accumulated in marine sediments over the earth's history. They can spontaneously turn from solid to gaseous form, causing submarine landslides and releasing methane into the water, affecting its chemistry.

### 2.11. OCEAN BASINS

The *Pacific Ocean* (Figure 2.8) is the world's largest ocean basin. To the north there is a physical boundary broken only by the Bering Strait, which is quite shallow (about 50 m) and 82 km wide. There is a small net northward flow from the Pacific to the Arctic through this strait. At the equator, the Pacific is very wide so that tropical phenomena that propagate east-west take much longer to cross the Pacific than across the other oceans. The Pacific is rimmed in the west and north with trenches and ridges. This area, because of the associated volcanoes, is called the "ring of fire." The EPR, a major topographic feature of the tropical and South Pacific, is a spreading ridge that separates the deep waters of the southeast from the rest of the Pacific; it is part of the global mid-ocean ridge (Section 2.2). Fracture zones allow some communication of deep waters across the ridge. Where the major eastward current of the Southern Ocean, the ACC (Chapter 13), encounters the ridge, the current is deflected.

The Pacific has more islands than any other ocean. Most of them are located in the western tropical regions. The Hawaiian Islands and their extension northwestward into the Emperor Seamounts were created by motion of the Pacific oceanic plate across the hotspot that is now located just east of the big island of Hawaii.

The Pacific Ocean has numerous marginal seas, mostly along its western side. In the North Pacific these are the Bering, Okhotsk, Japan, Yellow, East China, and South China Seas in the west and the Gulf of California in the east. In the South Pacific the marginal seas are the Coral and Tasman Seas and many smaller distinct regions that are named, such as the Solomon Sea (not shown). In the southern South Pacific is the Ross Sea, which contributes to the bottom waters of the world ocean.

The Atlantic Ocean has an "S" shape (Figure 2.9). The MAR, a spreading ridge down the center of the ocean, dominates its topography. Deep trenches are found just east of the Lesser Antilles in the eastern Caribbean and east of the South Sandwich Islands. The Atlantic is open both at the north and the south connecting to the Arctic and Southern Oceans. The northern North Atlantic is one of the two sources of the world's deep water (Chapter 9). One of the Atlantic's marginal seas, the Mediterranean, is evaporative and contributes high salinity, warm water to the mid-depth ocean. At the southern boundary, the Weddell Sea is a major formation site for the bottom water found in the oceans (Chapter 13). Other marginal seas connecting to the Atlantic are the Norwegian, Greenland, and Iceland Seas (sometimes known collectively as the Nordic Seas), the North Sea, the Baltic Sea, the Black Sea, and the Caribbean. The Irminger Sea is the region southeast of Greenland, the Labrador Sea is the region between Labrador and Greenland, and the Sargasso Sea is the open ocean region surrounding Bermuda. Fresh outflow 2. OCEAN DIMENSIONS, SHAPES, AND BOTTOM MATERIALS



from large rivers such as the Amazon, Congo, and Orinoco Rivers form marked low-salinity tongues at the sea surface.

The *Indian Ocean* (Figure 2.10) is closed off by land just north of the tropics. The topography of the Indian Ocean is very rough because of the ridges left behind as the Indian plate moved northward into the Asian continent creating the Himalayas. The Central Indian Ridge and Southwest Indian Ridge are two of the slowest spreading ridges on Earth. (As discussed previously, seafloor roughness from abyssal hills and fracture zones is highest at slower spreading rates, which is necessary in understanding the spatial distribution of deep mixing in the global ocean.) The only trench is the Sunda Trench where the Indian plate subducts beneath Indonesia. The eastern boundary of the Indian Ocean is porous and connected to the Pacific Ocean through the Indonesian archipelago. Marginal seas for the Indian Ocean include the Andaman Sea, the Red Sea, and the Persian Gulf. The open ocean region west of India is called the Arabian Sea and the region east of India is called the Bay of Bengal.

The differential heating of land and ocean in the tropics results in the creation of the monsoon weather system. Monsoons occur in many places, but the most dramatic and best described monsoon is in the northern Indian Ocean (Chapter 11). From October to May the Northeast Monsoon sends cool, dry winds from the OCEAN BASINS



FIGURE 2.9 Map of the Atlantic Ocean. Etopo2 bathymetry data from NOAA NGDC (2008).

1000 2000 3000

4000 5000 6000

7000

continental land masses in the northeast over the Indian subcontinent to the ocean. Starting in June and lasting until September, the system shifts to the southwest monsoon, which brings warm, wet rains from the western tropical ocean to the Indian subcontinent. While these monsoon conditions are best known in India, they also dominate the climate in the western tropical and South Pacific.

Most of the rivers that drain southward from the Himalayas — including the Ganges, Brahmaputra, and Irawaddy — flow out into the

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FIGURE 2.10 Map of the Indian Ocean. Etopo2 bathymetry data from NOAA NGDC (2008).

Bay of Bengal, east of India rather than into the Arabian Sea, west of India. This causes the surface water of the Bay of Bengal to be quite fresh. The enormous amount of silt carried by these rivers from the eroding Himalayan Mountains into the Bay of Bengal creates the subsurface geological feature, the Bengal Fan, which slopes smoothly downward for thousands of kilometers. West of India, the Arabian Sea, Red Sea, and Persian Gulf are very salty due to the dry climate and subsequent high evaporation. Similar to the Mediterranean, the saline Red Sea water is sufficiently dense to sink to middepth in the Indian Ocean and affects water properties over a large part of the Arabian Sea and western Indian Ocean. OCEAN BASINS



FIGURE 2.11 The Arctic Ocean. Etopo2 bathymetry data from NOAA NGDC (2008).

The *Arctic Ocean* (Figure 2.11) is sometimes not regarded as an ocean, but rather as a mediterranean sea connected to the Atlantic Ocean. It is characterized by very broad continental shelves surrounding a deeper region, which is split down the center by the Lomonosov Ridge. These shelf areas around the Arctic are called the Beaufort, Chukchi, East Siberian, Laptev, Kara, and Barents Seas. The Arctic is connected to the North Pacific through the shallow Bering Strait. It is connected to the Nordic Seas (Norwegian and Greenland) through passages on either side of Svalbard, including Fram Strait between Svalbard and Greenland. The Nordic Seas are separated from the Atlantic Ocean by the submarine ridge between Greenland, Iceland,



FIGURE 2.12 The Southern Ocean around Antarctica. Etopo2 bathymetry data from NOAA NGDC (2008).

and the UK, with a maximum sill depth of about 620 m in the Denmark Strait, between Greenland and Iceland. Dense water formed in the Nordic Seas spills into the Atlantic over this ridge. The central area of the Arctic Ocean is perennially covered with sea ice.

The *Southern Ocean* (Figure 2.12) is not geographically distinct from the Atlantic,

Indian, and Pacific Oceans, but is often considered separately since it is the only region outside the Arctic where there is a path for eastward flow all the way around the globe. This occurs at the latitude of Drake Passage between South America and Antarctica and allows the three major oceans to be connected. The absence of a meridional (north-south) boundary in Drake Passage changes the dynamics of the flow at these latitudes completely in comparison with the rest of the ocean, which has meridional boundaries. Drake Passage also serves to constrict the width of the flow of the ACC, which must pass in its entirety through the passage. The South Sandwich Islands and trench east of Drake Passage partially block the open circumpolar flow. Another major constriction is the broad Pacific-Antarctic rise, which is the seafloor spreading ridge between the Pacific and Antarctic plates. This fast-spreading ridge has few deep fracture zones, so the ACC must deflect northward before finding the only two deep channels, the Udintsev and Eltanin Fracture Zones.

The ocean around Antarctica includes permanent ice shelves as well as seasonal sea ice (Figures 13.11 and 13.19). Unlike the Arctic there is no perennial long-term pack ice; except for some limited ice shelves and all of the firstyear ice melts and forms each year. The densest bottom waters of the world ocean are formed in the Southern Ocean, primarily in the Weddell and Ross Seas as well as in other areas distributed along the Antarctic coast between the Ross Sea and Prydz Bay.