

INTRODUCTION TO PHYSICAL OCEANOGRAPHY



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Introduction to Physical Oceanography

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Licensing

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Preface

This book is written for upper-division undergraduates and new graduate students in meteorology, ocean engineering, and oceanography. Because these students have a diverse background, I have emphasized ideas and concepts more than mathematical derivations.

Unlike most books, I am distributing this book for free in digital format via the world-wide web. I am doing this for two reasons:

1. Textbooks are usually out of date by the time they are published, usually a year or two after the author finishes writing the book. Randol Larson, writing in *Syllabus*, states: “In my opinion, technology textbooks are a waste of natural resources. They’re out of date the moment they are published. Because of their short shelf life, students don’t even want to hold on to them”—(Larson, 2002). By publishing in electronic form, I can make revisions every year, keeping the book current.
2. Many students, especially in less-developed countries, cannot afford the high cost of textbooks from the developed world. This then is a gift from the US National Aeronautics and Space Administration (NASA) to the students of the world.

Acknowledgements

I have taught from the book for several years, and I thank the many students in my classes and throughout the world who have pointed out poorly written sections, ambiguous text, conflicting notation, and other errors. I also thank Professor Fred Schlemmer at Texas A&M Galveston who, after using the book for his classes, has provided extensive comments about the material.

I also wish to thank many colleagues for providing figures, comments, and helpful information. I especially wish to thank Aanderaa Instruments, Bill Allison, Kevin Bartlett, James Berger, Gerben de Boer, Daniel Bourgault, Don Chambers, Greg Crawford, Thierry De Mees, Richard Eanes, Peter Etnoyer, Tal Ezer, Gregg Foti, Nevin S. Fučkar, Luiz Alexandre de Araujo Guerra, Hazel Jenkins, Andrew Kiss, Jody Klymak, Judith Lean, Christian LeProvost, Brooks Martner, Nikolai Maximenko, Kevin McKone, Mike McPhaden, Thierry De Mees, Pim van Meurs, Gary Mitchum, Joe Murtagh, Peter Niiler, Nuno Nunes, Ismael Núñez-Riboni, Alex Orsi, Kym Perkin, Mark Powell, Richard Ray, Joachim Ribbe, Will Sager, David Sandwell, Sea-Bird Electronics, Achim Stoessel, David Stooksbury, Tom Whitworth, Carl Wunsch and many others.

Of course, I accept responsibility for all mistakes in the book. Please send me your comments and suggestions for improvement.

Figures in the book came from many sources. I particularly wish to thank Link Ji for many global maps, and colleagues at the University of Texas Center for Space Research. Don Johnson redrew many figures and turned sketches into figures. Trey Morris tagged the words used in the index. I especially thank NASA’s Jet Propulsion Laboratory and the Topex/Poseidon and Jason Projects for their support of the book through contracts 960887 and 1205046. Cover photograph of the resort island of Kurumba in North Male Atoll in the Maldives was taken by Jagdish Agara (copyright Corbis). Cover design is by Don Johnson.

The book was produced in LATEX 2 ϵ using TeXShop 2.14 on an Intel iMac computer running OS-X 10.4.11. I especially wish to thank Gerben Wierda for his very useful i-Installer package that made it all possible, Richard Koch, Dirk Olmes and many others for writing the TeXShop software package, and Andrew Kiss at the University of New South Wales in Canberra Australia for help in using the hyperref package. The LATEX 2 ϵ software is a pleasure to use. All figures were drawn in Adobe Illustrator.

CHAPTER OVERVIEW

1: A Voyage of Discovery

The role of the ocean on weather and climate is often discussed in the news. Who has not heard of El Niño and changing weather patterns, the Atlantic hurricane season and storm surges? Yet, what exactly is the role of the ocean? And, why do we care?

[1.1: Why Study the Physics of the Ocean?](#)

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1.1: Why Study the Physics of the Ocean?

The answer depends on our interests, which devolve from our use of the ocean. Three broad themes are important:

1. We get food from the ocean. Hence we may be interested in processes which influence the sea, just as farmers are interested in the weather and climate. The ocean not only has weather such as temperature changes and currents, but the oceanic weather fertilizes the sea. The atmospheric weather seldom fertilizes fields except for the small amount of nitrogen fixed by lightning.
2. We use the ocean. We build structures on the shore or just offshore. We use the ocean for transport. We obtain oil and gas below the ocean. And, we use the ocean for recreation, swimming, boating, fishing, surfing, and diving. Hence we are interested in processes that influence these activities, especially waves, winds, currents, and temperature.
3. The ocean influences the atmospheric weather and climate. The ocean influences the distribution of rainfall, droughts, floods, regional climate, and the development of storms, hurricanes, and typhoons. Hence we are interested in air-sea interactions, especially the fluxes of heat and water across the sea surface, the transport of heat by the ocean, and the influence of the ocean on climate and weather patterns.

These themes influence our selection of topics to study. The topics then determine what we measure, how the measurements are made, and the geographic areas of interest. Some processes are local, such as the breaking of waves on a beach, some are regional, such as the influence of the North Pacific on Alaskan weather, and some are global, such as the influence of the ocean on changing climate and global warming.

If indeed these reasons for the study of the ocean are important, let's begin a voyage of discovery. Any voyage needs a destination. What is ours?

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1.2: Goals

At the most basic level, I hope you, the students who are reading this text, will become aware of some of the major conceptual schemes (or theories) that form the foundation of physical oceanography, how they were arrived at, and why they are widely accepted, how oceanographers achieve order out of a random ocean, and the role of experiment in oceanography (to paraphrase Shamos, 1995: p. 89).

More particularly, I expect you will be able to describe physical processes influencing the ocean and coastal regions: the interaction of the ocean with the atmosphere, and the distribution of oceanic winds, currents, heat fluxes, and water masses. The text emphasizes ideas rather than mathematical techniques. I will try to answer such questions as:

1. What is the basis of our understanding of physics of the ocean?
 - (a) What are the physical properties of sea water?
 - (b) What are the important thermodynamic and dynamic processes influencing the ocean?
 - (c) What equations describe the processes and how were they derived?
 - (d) What approximations were used in the derivation?
 - (e) Do the equations have useful solutions?
 - (f) How well do the solutions describe the process? That is, what is the experimental basis for the theories?
 - (g) Which processes are poorly understood? Which are well understood?
2. What are the sources of information about physical variables?
 - (a) What instruments are used for measuring each variable?
 - (b) What are their accuracy and limitations?
 - (c) What historic data exist?
 - (d) What platforms are used? Satellites, ships, drifters, moorings?
3. What processes are important? Some important process we will study include:
 - (a) Heat storage and transport in the ocean.
 - (b) The exchange of heat with the atmosphere and the role of the ocean in climate.
 - (c) Wind and thermal forcing of the surface mixed layer.
 - (d) The wind-driven circulation including the Ekman circulation, Ekman pumping of the deeper circulation, and upwelling.
 - (e) The dynamics of ocean currents, including geostrophic currents and the role of vorticity.
 - (f) The formation of water types and masses.
 - (g) The deep circulation of the ocean.
 - (h) Equatorial dynamics, El Niño, and the role of the ocean in weather.
 - (i) Numerical models of the circulation.
 - (j) Waves in the ocean including surface waves, inertial oscillations, tides, and tsunamis.
 - (k) Waves in shallow water, coastal processes, and tide predictions.
4. What are a few of the major currents and water masses in the ocean, and what governs their distribution?

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1.3: Organization

Before beginning a voyage, we usually try to learn about the places we will visit. We look at maps and we consult travel guides. In this book, our guide will be the papers and books published by oceanographers. We begin with a brief overview of what is known about the ocean. We then proceed to a description of the ocean basins, for the shape of the seas influences the physical processes in the water. Next, we study the external forces, wind and heat, acting on the ocean, and the ocean's response. As we proceed, I bring in theory and observations as necessary.

By the time we reach chapter 7, we will need to understand the equations describing dynamic response of the ocean. So we consider the equations of motion, the influence of earth's rotation, and viscosity. This leads to a study of wind-driven ocean currents, the geostrophic approximation, and the usefulness of conservation of vorticity.

Toward the end, we consider some particular examples: the deep circulation, the equatorial ocean and El Niño, and the circulation of particular areas of the ocean. Next we look at the role of numerical models in describing the ocean. At the end, we study coastal processes, waves, tides, wave and tidal forecasting, tsunamis, and storm surges.

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1.4: The Big Picture

The ocean is one part of the earth system. It mediates processes in the atmosphere by the transfers of mass, momentum, and energy through the sea surface. It receives water and dissolved substances from the land. And, it lays down sediments that eventually become rocks on land. Hence an understanding of the ocean is important for understanding the earth as a system, especially for understanding important problems such as global change or global warming. At a lower level, physical oceanography and meteorology are merging. The ocean provides the feedback leading to slow changes in the atmosphere.

As we study the ocean, I hope you will notice that we use theory, observations, and numerical models to describe ocean dynamics. *None is sufficient by itself.*

1. Ocean processes are nonlinear and turbulent. Yet we don't really understand the theory of non-linear, turbulent flow in complex basins. Theories used to describe the ocean are much simplified approximations to reality.
2. Observations are sparse in time and space. They provide a rough description of the time-averaged flow, but many processes in many regions are poorly observed.
3. Numerical models include much more realistic theoretical ideas, they can help interpolate oceanic observations in time and space, and they are used to forecast climate change, currents, and waves. Nonetheless, the numerical equations are approximations to the continuous analytic equations that describe fluid flow, they contain no information about flow between grid points, and they cannot yet be used to describe fully the turbulent flow seen in the ocean.

By combining theory and observations in numerical models we avoid some of the difficulties associated with each approach used separately (figure 1.4.1). Continued refinements of the combined approach are leading to ever-more-precise descriptions of the ocean. The ultimate goal is to know the ocean well enough to predict the future changes in the environment, including climate change or the response of fisheries to overfishing.

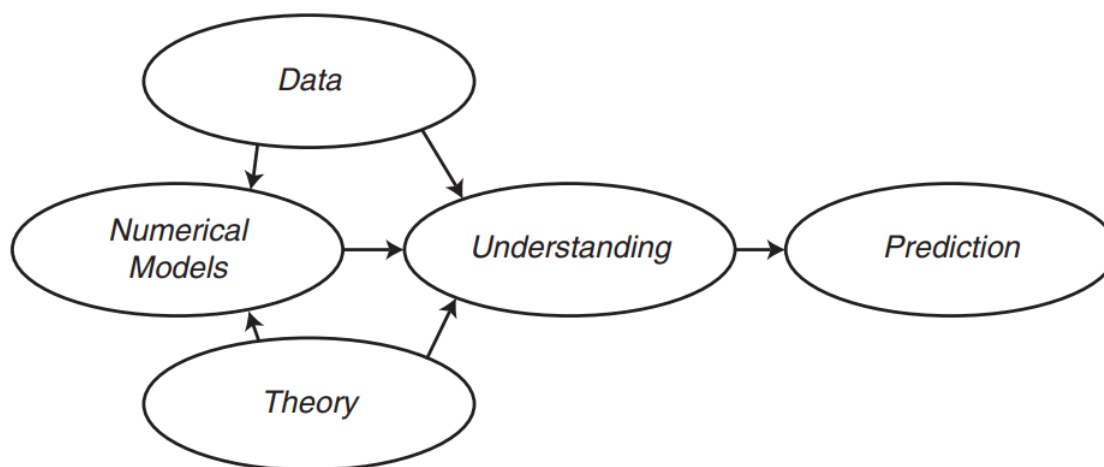


Figure 1.4.1: Data, numerical models, and theory are all necessary to understand the ocean. Eventually, an understanding of the ocean-atmosphere-land system will lead to predictions of future states of the system.

The combination of theory, observations, and computer models is relatively new. Four decades of exponential growth in computing power has made available desktop computers capable of simulating important physical processes and oceanic dynamics.

“All of us who are involved in the sciences know that the computer has become an essential tool for research ...scientific computation has reached the point where it is on a par with laboratory experiment and mathematical theory as a tool for research in science and engineering.”

—Langer (1999)

The combination of theory, observations, and computer models also implies a new way of doing oceanography. In the past, an oceanographer would devise a theory, collect data to test the theory, and publish the results. Now, the tasks have become so specialized that few can do it all. Few excel in theory, collecting data, and numerical simulations. Instead, the work is done more and more by teams of scientists and engineers.

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1.5: Further Reading

If you know little about the ocean and oceanography, I suggest you begin by reading MacLeish's (1989) book *The Gulf Stream: Encounters With the Blue God*, especially his Chapter 4 on "Reading the ocean." In my opinion, it is the best overall, non-technical, description of how oceanographers came to understand the ocean.

You may also benefit from reading pertinent chapters from any introductory oceanographic textbook. Those by Gross, Pinet, or Segar are especially useful. The three texts produced by the Open University provide a slightly more advanced treatment.

Gross, M. Grant and Elizabeth Gross (1996). *Oceanography—A View of Earth*. 7th edition. Prentice Hall.

MacLeish, William (1989). *The Gulf Stream: Encounters With the Blue God*. Houghton Mifflin Company.

Pinet, Paul R. (2006). *Invitation to Oceanography*. 4nd edition. Jones and Bartlett Publishers.

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Open University (1989). *Waves, Tides and Shallow-Water Processes*. Pergamon Press.

Segar, Douglas A. (2007). *Introduction to Ocean Sciences*. 2nd edition. W. W. Norton.

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CHAPTER OVERVIEW

2: The Historical Setting

Our knowledge of oceanic currents, winds, waves, and tides goes back thousands of years. Polynesian navigators traded over long distances in the Pacific as early as 4000 BC (Service, 1996). Pytheas explored the Atlantic from Italy to Norway in 325 BC. Arabic traders used their knowledge of the reversing winds and currents in the Indian Ocean to establish trade routes to China in the Middle Ages and later to Zanzibar on the African coast. And, the connection between tides and the sun and moon was described in the Samaveda of the Indian Vedic period extending from 2000 to 1400 BC (Pugh, 1987). Those oceanographers who tend to accept as true only that which has been measured by instruments, have much to learn from those who earned their living on the ocean.

Modern European knowledge of the ocean began with voyages of discovery by Bartholomew Dias (1487–1488), Christopher Columbus (1492–1494), Vasco da Gama (1497–1499), Ferdinand Magellan (1519–1522), and many others. They laid the foundation for global trade routes stretching from Spain to the Philippines in the early 16th century. The routes were based on a good working knowledge of trade winds, the westerlies, and western boundary currents in the Atlantic and Pacific (Couper, 1983: 192–193).

The early European explorers were soon followed by scientific voyages of discovery led by (among many others) James Cook (1728–1779) on the *Endeavour*, *Resolution*, and *Adventure*, Charles Darwin (1809–1882) on the *Beagle*, Sir James Clark Ross and Sir John Ross who surveyed the Arctic and Antarctic regions from the *Victory*, the *Isabella*, and the *Erebus*, and Edward Forbes (1815–1854) who studied the vertical distribution of life in the ocean. Others collected oceanic observations and produced useful charts, including Edmond Halley who charted the trade winds and monsoons and Benjamin Franklin who charted the Gulf Stream.

Slow ships of the 19th and 20th centuries gave way to satellites, drifters, and autonomous instruments toward the end of the 20th century. Satellites now observe the ocean, air, and land. Thousands of drifters observe the upper two kilometers of the ocean. Data from these systems, when fed into numerical models allows the study of earth as a system. For the first time, we can study how biological, chemical, and physical systems interact to influence our environment.

[2.1: Definitions](#)

[2.2: Eras of Oceanographic Exploration](#)

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2.1: Definitions

The long history of the study of the ocean has led to the development of various, specialized disciplines each with its own interests and vocabulary. The more important disciplines include:

Oceanography is the study of the ocean, with emphasis on its character as an environment. The goal is to obtain a description sufficiently quantitative to be used for predicting the future with some certainty.

Geophysics is the study of the physics of the earth.

Physical Oceanography is the study of physical properties and dynamics of the ocean. The primary interests are the interaction of the ocean with the atmosphere, the oceanic heat budget, water mass formation, currents, and coastal dynamics. Physical Oceanography is considered by many to be a subdiscipline of geophysics.

Geophysical Fluid Dynamics is the study of the dynamics of fluid motion on scales influenced by the rotation of the earth. Meteorology and oceanography use geophysical fluid dynamics to calculate planetary flow fields.

Hydrography is the preparation of nautical charts, including charts of ocean depths, currents, internal density field of the ocean, and tides.

Earth-system Science is the study of earth as a single system comprising many interacting subsystems including the ocean, atmosphere, cryosphere, and biosphere, and changes in these systems due to human activity.

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2.2: Eras of Oceanographic Exploration

The exploration of the sea can be divided, somewhat arbitrarily, into various eras (Wust, 1964). I have extended his divisions through the end of the 20th century.

1. Era of Surface Oceanography: Earliest times to 1873. The era is characterized by systematic collection of mariners' observations of winds, currents, waves, temperature, and other phenomena observable from the deck of sailing ships. Notable examples include Halley's charts of the trade winds, Franklin's map of the Gulf Stream, and Matthew Fontaine Maury's Physical Geography of the Sea.
2. Era of Deep-Sea Exploration: 1873–1914. Characterized by a few, wide-ranging oceanographic expeditions to survey surface and subsurface conditions, especially near colonial claims. The major example is the Challenger Expedition (figure 2.2.1), but also the Gazelle and Fram Expeditions.

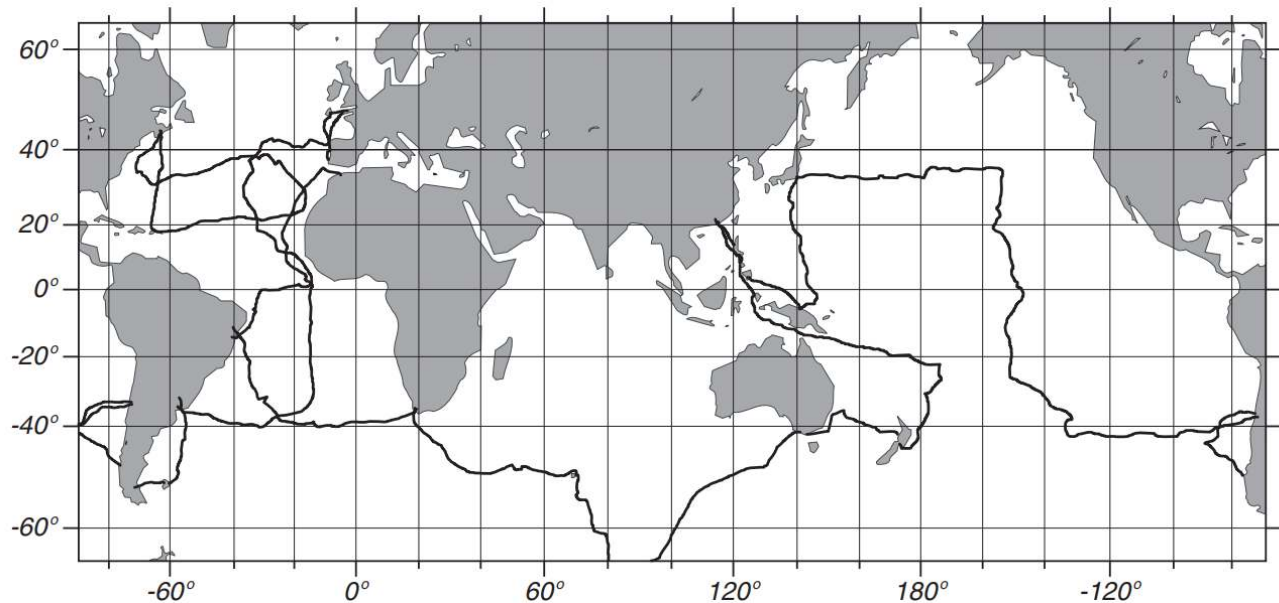


Figure 2.2.1: Example from the era of deep-sea exploration: Track of H.M.S. *Challenger* during the British Challenger Expedition, 1872–1876. After Wust (1964).

3. Era of National Systematic Surveys: 1925–1940. Characterized by detailed surveys of colonial areas. Examples include Meteor surveys of the Atlantic (figure 2.2.2), and the Discovery Expeditions.

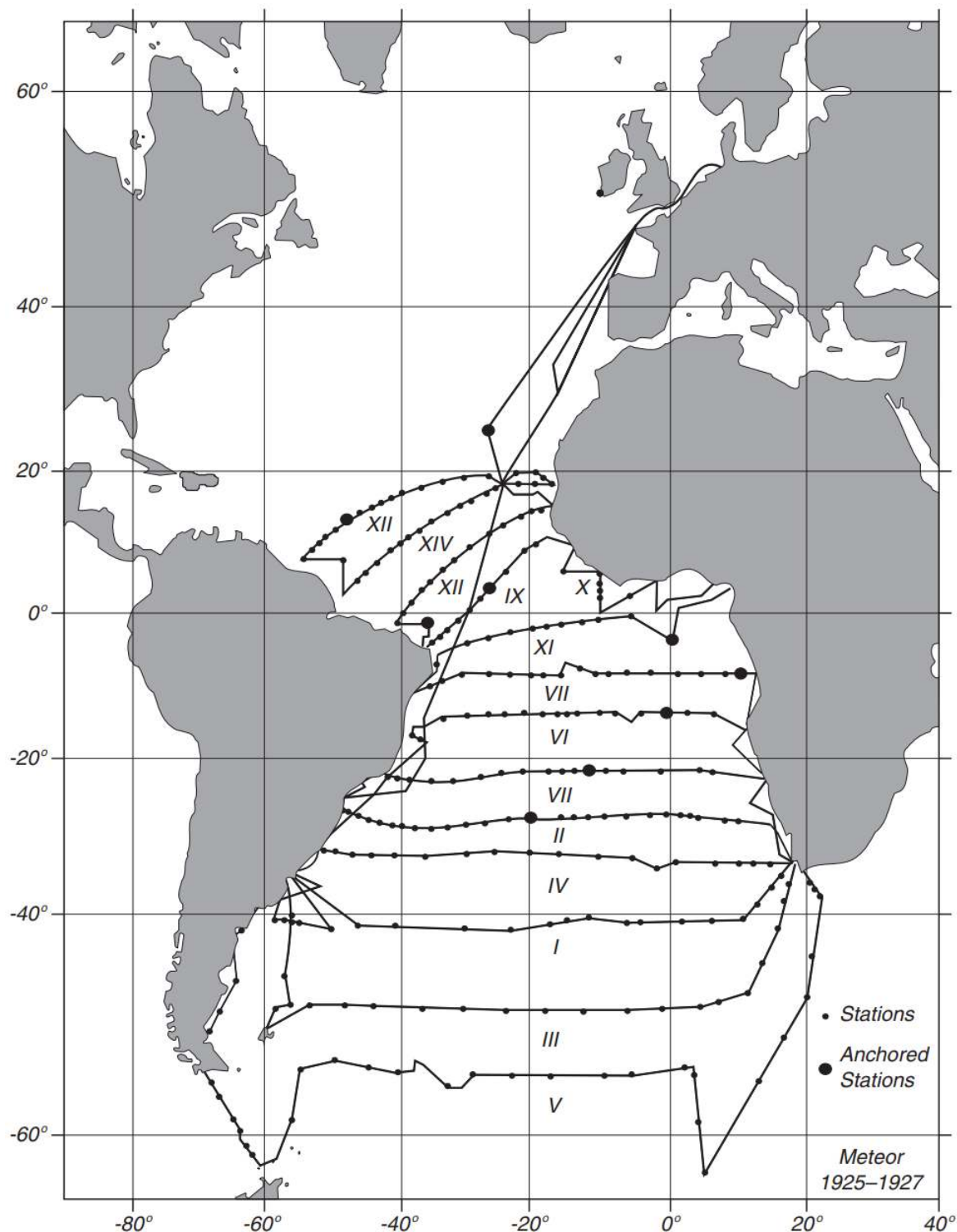


Figure 2.2.2: Example of a survey from the era of national systematic surveys. Track of the R/V *Meteor* during the German *Meteor* Expedition. Redrawn from Wust (1964).

4. Era of New Methods: 1947–1956. Characterized by long surveys using new instruments (figure 2.2.3). Examples include seismic surveys of the Atlantic by *Vema* leading to Heezen’s maps of the sea floor.

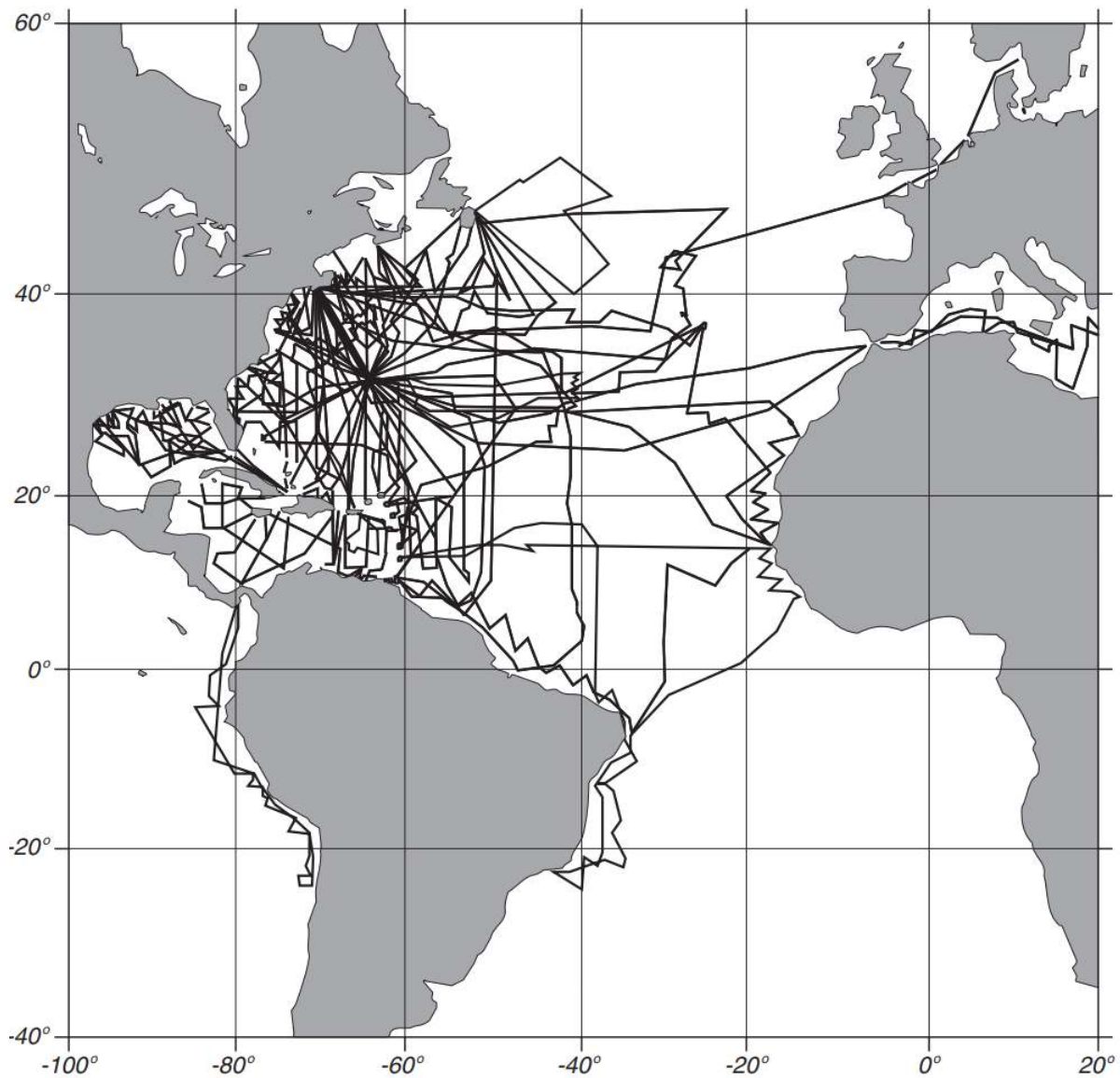


Figure 2.2.3: Example from the era of new methods. The cruises of the R/V *Atlantis* out of Woods Hole Oceanographic Institution. After Wust (1964).

5. Era of International Cooperation: 1957–1978. Characterized by multinational surveys of ocean and studies of oceanic processes. Examples include the Atlantic Polar Front Program, the norpac cruises, the International Geophysical Year cruises, and the International Decade of Ocean Exploration (figure 2.2.4). Multiship studies of oceanic processes include mode, polymode, norpax, and jasin experiments.

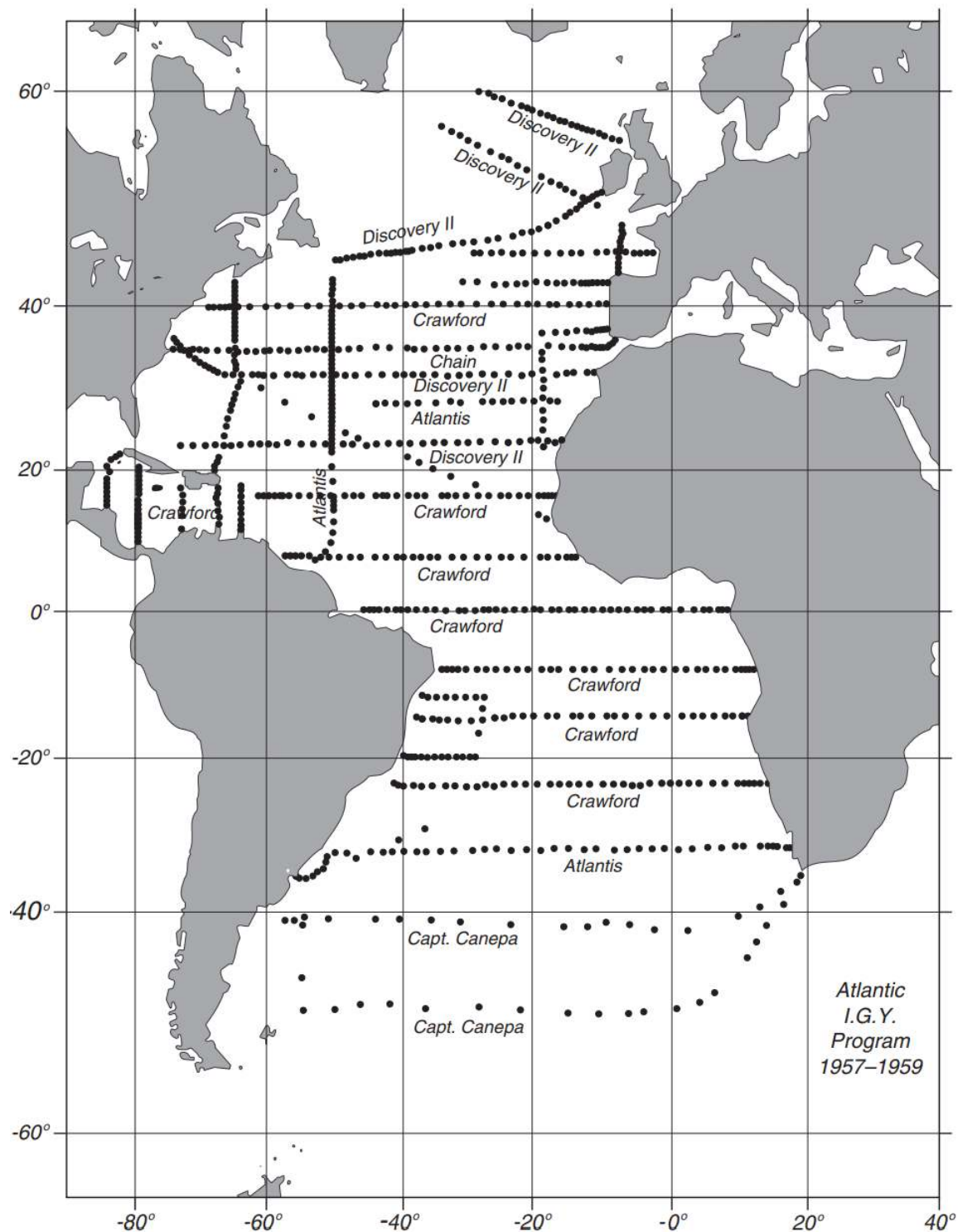


Figure 2.2.4: Example from the era of international cooperation. Sections measured by the International Geophysical Year Atlantic Program, 1957-1959. After Wust (1964).

6. Era of Satellites: 1978–1995. Characterized by global surveys of oceanic processes from space. Examples include Seasat, noaa 6–10, nimbus–7, Geosat, Topex/Poseidon, and ers–1 & 2.
7. Era of Earth System Science: 1995– Characterized by global studies of the interaction of biological, chemical, and physical processes in the ocean and atmosphere and on land using in situ (which means from measurements made in the water) and space data in numerical models. Oceanic examples include the World Ocean Circulation Experiment (woce) (figure 2.2.5) and

Topex/Poseidon (figure 2.2.6), the Joint Global Ocean Flux Study (JGOFS), the Global Ocean Data Assimilation Experiment (GODAE), and the SeaWiFS, Aqua, and Terra satellites.

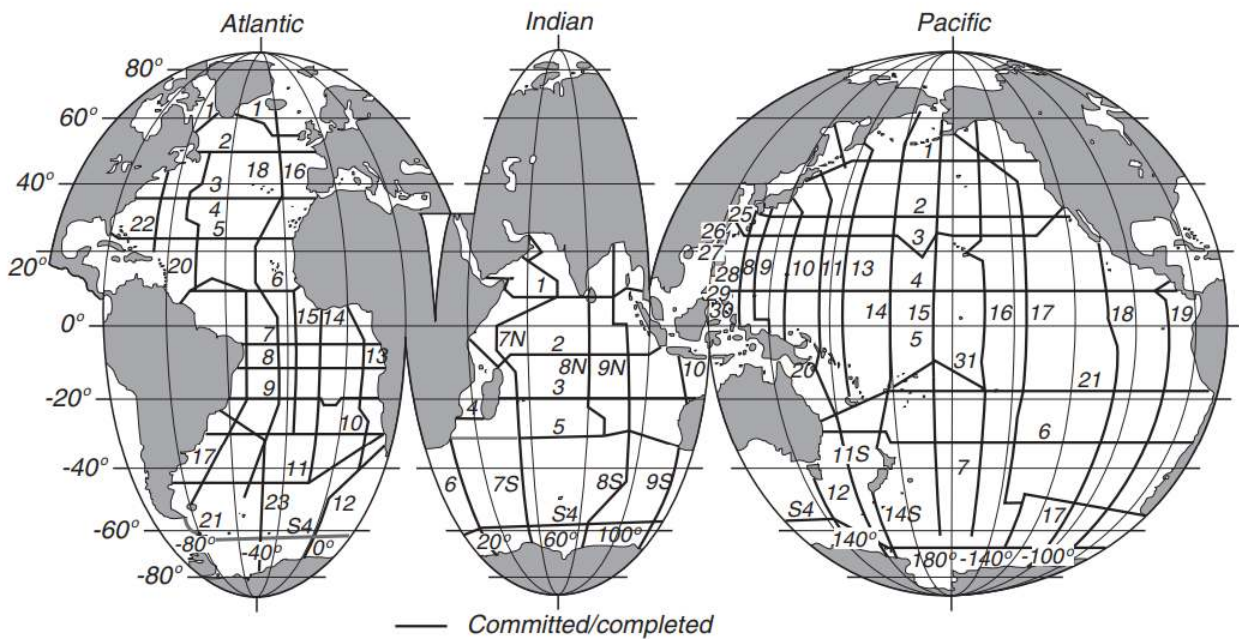


Figure 2.2.5: Tracks of research ships making a one-time global survey of the ocean of the world. From World Ocean Circulation Experiment.

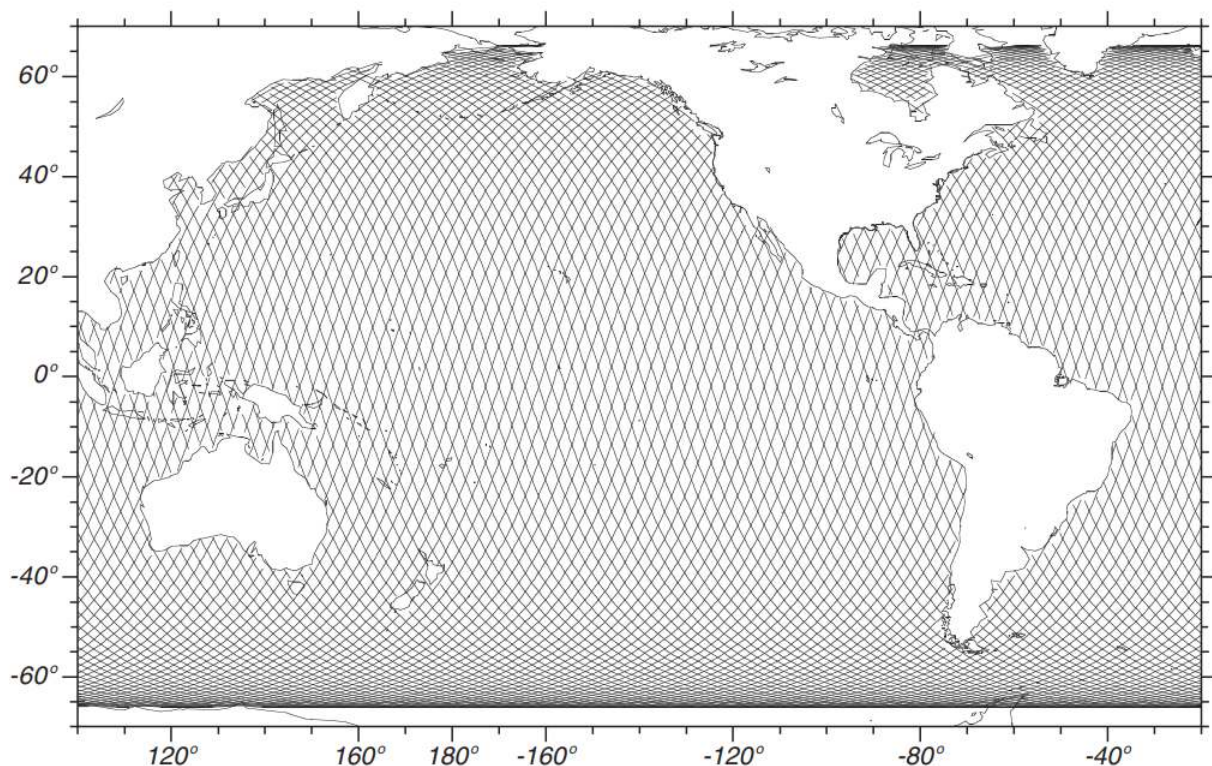


Figure 2.2.6: Example from the era of satellites. Topex/Poseidon tracks in the Pacific Ocean during a 10-day repeat of the orbit. From Topex/Poseidon Project.

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2.3: Milestones in the Understanding of the Ocean

What have all these programs and expeditions taught us about the ocean? Let's look at some milestones in our ever-increasing understanding of the ocean, beginning with the first scientific investigations of the 17th century. Initially progress was slow. First came very simple observations of far-reaching importance by scientists who probably did not consider themselves oceanographers, if the term even existed. Later came more detailed descriptions and oceanographic experiments by scientists who specialized in the study of the ocean.

1685 Edmond Halley, investigating the oceanic wind systems and currents, published "An Historical Account of the Trade Winds, and Monsoons, observable in the Seas between and near the Tropicks, with an attempt to assign the Physical cause of the said Winds" *Philosophical Transactions*.

1735 George Hadley published his theory for the trade winds based on conservation of angular momentum in "Concerning the Cause of the General Trade-Winds" *Philosophical Transactions*, 39: 58-62.

1751 Henri Ellis made the first deep soundings of temperature in the tropics, finding cold water below a warm surface layer, indicating the water came from the polar regions.

1769 Benjamin Franklin, as postmaster, made the first map of the Gulf Stream using information from mail ships sailing between New England and England collected by his cousin Timothy Folger (figure 2.3.1).

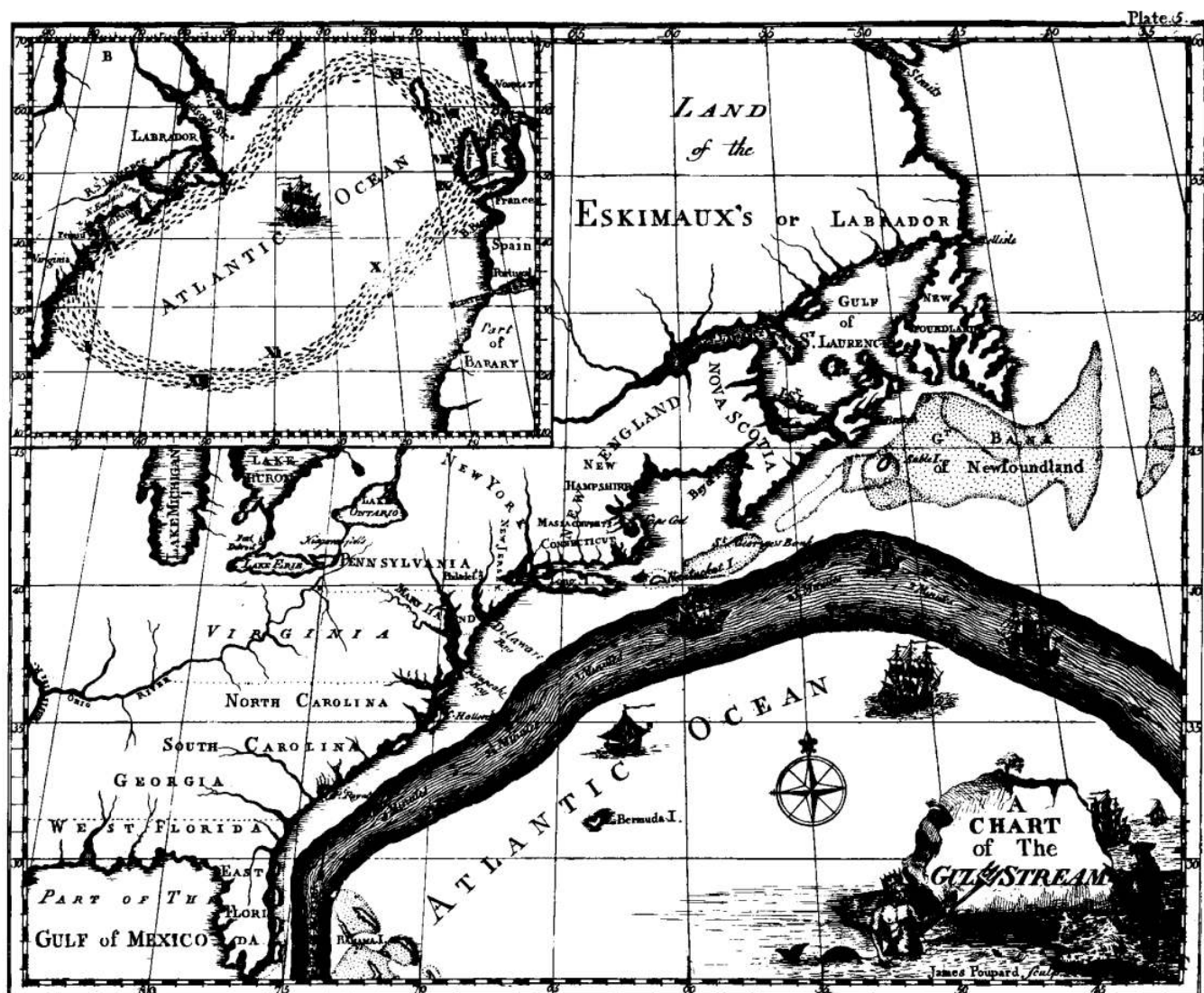


Figure 2.3.1: The 1786 version of Franklin-Folger map of the Gulf Stream.

1775 Laplace published his theory of tides.

1800 Count Rumford proposed a meridional circulation of the ocean with water sinking near the poles and rising near the Equator.

1847 Matthew Fontaine Maury published his first chart of winds and currents based on ships logs. Maury established the practice of international exchange of environmental data, trading logbooks for maps and charts derived from the data.

1872–1876 Challenger Expedition marks the beginning of the systematic study of the biology, chemistry, and physics of the ocean of the world.

1885 Pillsbury made direct measurements of the Florida Current using current meters deployed from a ship moored in the stream.

1903 Founding of the Marine Biological Laboratory of the University of California. It later became the Scripps Institution of Oceanography.

1910–1913 Vilhelm Bjerknes published *Dynamic Meteorology and Hydrography*, which laid the foundation of geophysical fluid dynamics. In it he developed the idea of fronts, the dynamic meter, geostrophic flow, air-sea interaction, and cyclones.

1930 Founding of the Woods Hole Oceanographic Institution.

1942 Publication of *The Ocean* by Sverdrup, Johnson, and Fleming, a comprehensive survey of oceanographic knowledge up to that time.

Post WW2 The need to detect submarines led the navies of the world to greatly expand their studies of the sea. This led to the founding of oceanography departments at state universities, including Oregon State, Texas A&M University, University of Miami, and University of Rhode Island, and the founding of national ocean laboratories such as the various Institutes of Oceanographic Science.

1947–1950 Sverdrup, Stommel, and Munk publish their theories of the wind-driven circulation of the ocean. Together the three papers lay the foundation for our understanding of the ocean's circulation. 1949 Start of California Cooperative Fisheries Investigation of the California Current. The most complete study ever undertaken of a coastal current.

1952 Cromwell and Montgomery rediscover the Equatorial Undercurrent in the Pacific.

1955 Bruce Hamon and Neil Brown develop the CTD for measuring conductivity and temperature as a function of depth in the ocean.

1958 Stommel publishes his theory for the deep circulation of the ocean.

1963 Sippican Corporation (Tim Francis, William Van Allen Clark, Graham Campbell, and Sam Francis) invents the Expendable BathyThermograph (XBT), now perhaps the most widely used oceanographic instrument deployed from ships.

1969 Kirk Bryan and Michael Cox develop the first numerical model of the oceanic circulation.

1978 NASA launches the first oceanographic satellite, Seasat. The project developed techniques used by generations of remotesensing satellites.

1979–1981 Terry Joyce, Rob Pinkel, Lloyd Regier, F. Rowe and J. W. Young develop techniques leading to the acoustic-doppler current profiler for measuring ocean-surface currents from moving ships, an instrument widely used in oceanography.

1988 NASA Earth System Science Committee headed by Francis Bretherton outlines how all earth systems are interconnected, thus breaking down the barriers separating traditional sciences of astrophysics, ecology, geology, meteorology, and oceanography.

1991 Wally Broecker proposes that changes in the deep circulation of the ocean modulate the ice ages, and that the deep circulation in the Atlantic could collapse, plunging the northern hemisphere into a new ice age.

1992 Russ Davis and Doug Webb invent the autonomous, pop-up drifter that continuously measures currents at depths to 2 km.

1992 NASA and CNES develop and launch Topex/Poseidon, a satellite that maps ocean surface currents, waves, and tides every ten days, revolutionizing our understanding of ocean dynamics and tides.

1993 Topex/Poseidon science-team members publish first accurate global maps of the tides.

More information on the history of physical oceanography can be found in Appendix A of W.S. von Arx (1962): *An Introduction to Physical Oceanography*.

Data collected from the centuries of oceanic expeditions have been used to describe the ocean. Most of the work went toward describing the steady state of the ocean, its currents from top to bottom, and its interaction with the atmosphere. The basic description was mostly complete by the early 1970s. Figure 2.3.2 shows an example from that time, the surface circulation of the ocean. More recent work has sought to document the variability of oceanic processes, to provide a description of the ocean sufficient to predict annual and interannual variability, and to understand the role of the ocean in global processes.

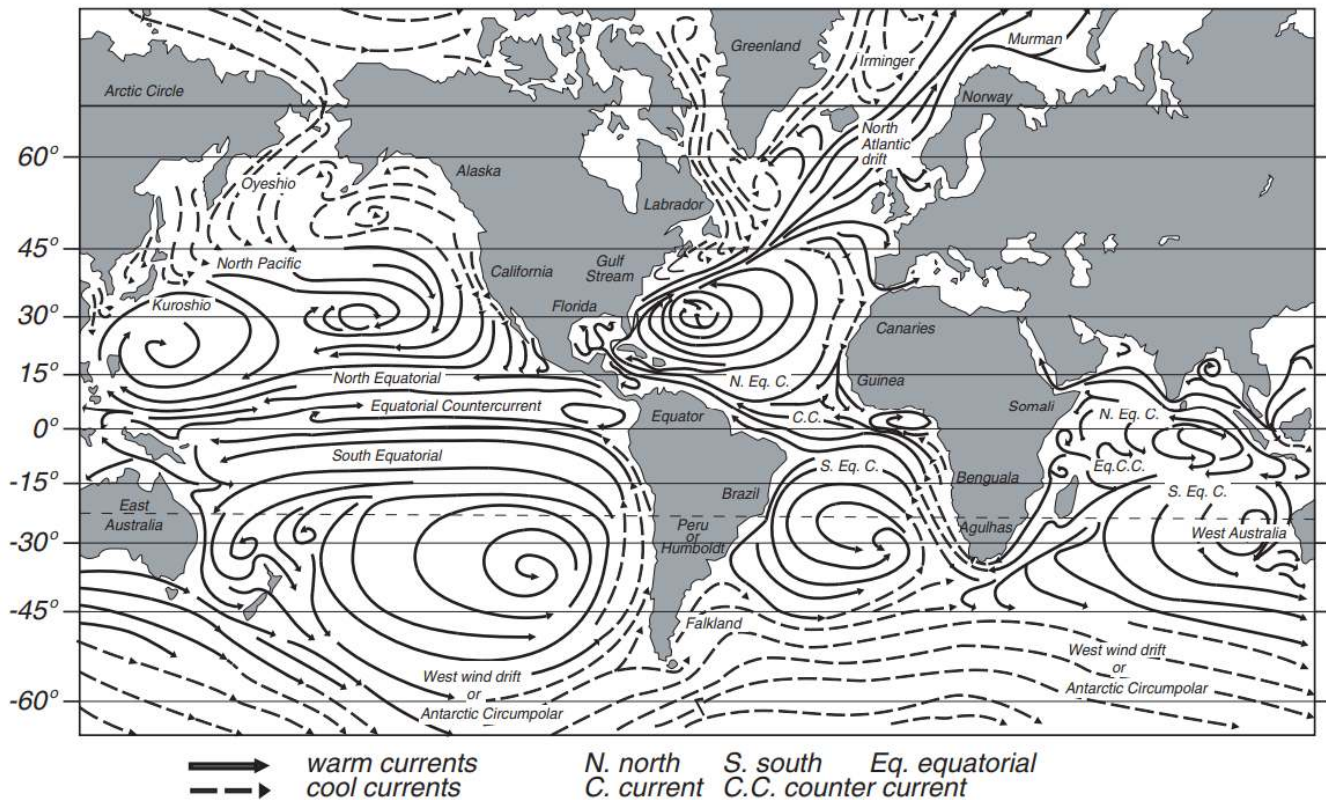


Figure 2.3.2: The time-averaged, surface circulation of the ocean during northern hemisphere winter deduced from a century of oceanographic expeditions. After Tolmazin (1985: 16).

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2.4: Evolution of Some Theoretical Ideas

A theoretical understanding of oceanic processes is based on classical physics coupled with an evolving understanding of chaotic systems in mathematics and the application to the theory of turbulence. The dates given below are approximate.

19th Century Development of analytic hydrodynamics. Lamb's *Hydrodynamics* is the pinnacle of this work. Bjerknes develops geostrophic method widely used in meteorology and oceanography.

1925–40 Development of theories for turbulence based on aerodynamics and mixing-length ideas. Work of Prandtl and von Karman.

1940–1970 Refinement of theories for turbulence based on statistical correlations and the idea of isotropic homogeneous turbulence. Books by Batchelor (1967), Hinze (1975), and others.

1970– Numerical investigations of turbulent geophysical fluid dynamics based on high-speed digital computers.

1985– Mechanics of chaotic processes. The application to hydrodynamics is just beginning. Most motion in the atmosphere and ocean may be inherently unpredictable.

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2.5: The Role of Observation in Oceanography

The brief tour of theoretical ideas suggests that observations are essential for understanding the ocean. The theory describing a convecting, wind-forced, turbulent fluid in a rotating coordinate system has never been sufficiently well known that important features of the oceanic circulation could be predicted before they were observed. In almost all cases, oceanographers resort to observations to understand oceanic processes.

At first glance, we might think that the numerous expeditions mounted since 1873 would give a good description of the ocean. The results are indeed impressive. Hundreds of expeditions have extended into all oceans. Yet, much of the ocean is poorly explored.

By the year 2000, most areas of the ocean will have been sampled from top to bottom only once. Some areas, such as the Atlantic, will have been sparsely sampled three times: during the International Geophysical Year in 1959, during the Geochemical Sections cruises in the early 1970s, and during the World Ocean Circulation Experiment from 1991 to 1996. All areas will be vastly under-sampled. This is the sampling problem (See box below). Our samples of the ocean are insufficient to describe the ocean well enough to predict its variability and its response to changing forcing. *Lack of sufficient samples is the largest source of error in our understanding of the ocean.*

The lack of observations has led to a very important and widespread conceptual error:

“The absence of evidence was taken as evidence of absence.” The great difficulty of observing the ocean meant that when a phenomenon was not observed, it was assumed it was not present. The more one is able to observe the ocean, the more the complexity and subtlety that appears.
— Wunsch (2002a)

As a result, our understanding of the ocean is often too simple to be correct.

Sampling Error

Sampling error is the largest source of error in the geosciences. It is caused by a set of samples not representing the population of the variable being measured. A population is the set of all possible measurements, and a sample is the sampled subset of the population. We assume each measurement is perfectly accurate.

To determine if your measurement has a sampling error, you must first completely specify the problem you wish to study. This defines the population. Then, you must determine if the samples represent the population. Both steps are necessary.

Suppose your problem is to measure the annual-mean sea-surface temperature of the ocean to determine if global warming is occurring. For this problem, the population is the set of all possible measurements of surface temperature, in all regions in all months. If the sample mean is to equal the true mean, the samples must be uniformly distributed throughout the year and over all the area of the ocean, and sufficiently dense to include all important variability in time and space. This is impossible. Ships avoid stormy regions such as high latitudes in winter, so ship samples tend not to represent the population of surface temperatures. Satellites may not sample uniformly throughout the daily cycle, and they may not observe temperature at high latitudes in winter because of persistent clouds, although they tend to sample uniformly in space and throughout the year in most regions. If daily variability is small, the satellite samples will be more representative of the population than the ship samples.

From the above, it should be clear that oceanic samples rarely represent the population we wish to study. We always have sampling errors.

In defining sampling error, we must clearly distinguish between instrument errors and sampling errors. Instrument errors are due to the inaccuracy of the instrument. Sampling errors are due to a failure to make a measurement. Consider the example above: the determination of mean sea-surface temperature. If the measurements are made by thermometers on ships, each measurement has a small error because thermometers are not perfect. This is an instrument error. If the ships avoids high latitudes in winter, the absence of measurements at high latitude in winter is a sampling error.

Meteorologists designing the Tropical Rainfall Mapping Mission have been investigating the sampling error in measurements of rain. Their results are general and may be applied to other variables. For a general description of the problem see North & Nakamoto (1989).

Selecting Oceanic Data Sets

Much of the existing oceanic data have been organized into large data sets. For example, satellite data are processed and distributed by groups working with NASA. Data from ships have been collected and organized by other groups. Oceanographers now rely more and more on such collections of data produced by others.

The use of data produced by others introduces problems: i) How accurate are the data in the set? ii) What are the limitations of the data set? And, iii) How does the set compare with other similar sets? Anyone who uses public or private data sets is wise to obtain answers to such questions.

If you plan to use data from others, here are some guidelines:

1. *Use well documented data sets.* Does the documentation completely describe the sources of the original measurements, all steps used to process the data, and all criteria used to exclude data? Does the data set include version numbers to identify changes to the set?
2. *Use validated data.* Has accuracy of data been well documented? Was accuracy determined by comparing with different measurements of the same variable? Was validation global or regional?
3. *Use sets that have been used by others and referenced in scientific papers.* Some data sets are widely used for good reason. Those who produced the sets used them in their own published work and others trust the data.
4. *Conversely, don't use a data set just because it is handy.* Can you document the source of the set? For example, many versions of the digital, 5-minute maps of the sea floor are widely available. Some date back to the first sets produced by the U.S. Defense Mapping Agency, others are from the etopo-5 set. Don't rely on a colleague's statement about the source. Find the documentation. If it is missing, find another data set.

Designing Oceanic Experiments

Observations are exceedingly important for oceanography, yet observations are expensive because ship time and satellites are expensive. As a result, oceanographic experiments must be carefully planned. While the design of experiments may not fit well within an historical chapter, perhaps the topic merits a few brief comments because it is seldom mentioned in oceanographic textbooks, although it is prominently described in texts for other scientific fields. The design of experiments is particularly important because poorly planned experiments lead to ambiguous results, they may measure the wrong variables, or they may produce completely useless data.

The first and most important aspect of the design of any experiment is to determine why you wish to make a measurement before deciding how you will make the measurement or what you will measure.

1. What is the purpose of the observations? Do you wish to test hypotheses or describe processes?
2. What accuracy is required of the observation?
3. What resolution in time and space is required? What is the duration of measurements?

Consider, for example, how the purpose of the measurement changes how you might measure salinity or temperature as a function of depth:

1. If the purpose is to describe water masses in an ocean basin, then measurements with 20–50 m vertical spacing and 50–300 km horizontal spacing, repeated once per 20–50 years in deep water are required.
2. If the purpose is to describe vertical mixing in the open equatorial Pacific, then 0.5–1.0 mm vertical spacing and 50–1000 km spacing between locations repeated once per hour for many days may be required.

Accuracy, Precision, and Linearity

While we are on the topic of experiments, now is a good time to introduce three concepts needed throughout the book when we discuss experiments: precision, accuracy, and linearity of a measurement.

Accuracy is the difference between the measured value and the true value.

Precision is the difference among repeated measurements.

The distinction between accuracy and precision is usually illustrated by the simple example of firing a rifle at a target. Accuracy is the average distance from the center of the target to the hits on the target. Precision is the average distance between the hits. Thus,

ten rifle shots could be clustered within a circle 10 cm in diameter with the center of the cluster located 20 cm from the center of the target. The accuracy is then 20 cm, and the precision is roughly 5 cm.

Linearity requires that the output of an instrument be a linear function of the input. Nonlinear devices rectify variability to a constant value. So a nonlinear response leads to wrong mean values. Non-linearity can be as important as accuracy. For example, let

$$\begin{aligned} \text{Output} &= \text{Input} + 0.1(\text{Input})^2 \\ \text{Input} &= a \sin \omega t \end{aligned}$$

then

$$\begin{aligned} \text{Output} &= a \sin \omega t + 0.1(a \sin \omega t)^2 \\ \text{Output} &= \text{Input} + \frac{0.1}{2}a^2 - \frac{0.1}{2}a^2 \cos 2\omega t \end{aligned}$$

Note that the mean value of the input is zero, yet the output of this nonlinear instrument has a mean value of $0.05a^2$ plus an equally large term at twice the input frequency. In general, if *input* has frequencies ω_1 and ω_2 , then *output* of a non-linear instrument has frequencies $\omega_1 \pm \omega_2$. Linearity of an instrument is especially important when the instrument must measure the mean value of a turbulent variable. For example, we require linear current meters when measuring currents near the sea surface where wind and waves produce a large variability in the current.

Sensitivity to other variables of interest.

Errors may be correlated with other variables of the problem. For example, measurements of conductivity are sensitive to temperature. So, errors in the measurement of temperature in salinometers leads to errors in the measured values of conductivity or salinity.

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2.6: Important Concepts

From the above, I hope you have learned:

1. The ocean is not well known. What we know is based on data collected from only a little more than a century of oceanographic expeditions supplemented with satellite data collected since 1978.
2. The basic description of the ocean is sufficient for describing the time-averaged mean circulation of the ocean, and recent work is beginning to describe the variability.
3. Observations are essential for understanding the ocean. Few processes have been predicted from theory before they were observed.
4. Lack of observations has led to conceptual pictures of oceanic processes that are often too simplified and often misleading.
5. Oceanographers rely more and more on large data sets produced by others. The sets have errors and limitations which you must understand before using them.
6. The planning of experiments is at least as important as conducting the experiment.
7. Sampling errors arise when the observations, the samples, are not representative of the process being studied. Sampling errors are the largest source of error in oceanography.
8. Almost all our observations of the ocean now come from satellites, drifters, and autonomous instruments. Fewer and fewer observations come from ships at sea.

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CHAPTER OVERVIEW

3: The Physical Setting

Earth is an oblate ellipsoid, an ellipse rotated about its minor axis, with an equatorial radius of $R_e = 6,378.1349$ km (West, 1982) slightly greater than the polar radius of $R_p = 6,356.7497$ km. The small equatorial bulge is due to Earth's rotation.

Distances on Earth are measured in many different units, the most common are degrees of latitude or longitude, meters, miles, and nautical miles. Latitude is the angle between the local vertical and the equatorial plane. A meridian is the intersection at Earth's surface of a plane perpendicular to the equatorial plane and passing through Earth's axis of rotation. Longitude is the angle between the standard meridian and any other meridian, where the standard meridian is the one that passes through a point at the Royal Observatory at Greenwich, England. Thus longitude is measured east or west of Greenwich.

A degree of latitude is not the same length as a degree of longitude except at the equator. Latitude is measured along great circles with radius R , where R is the mean radius of earth. Longitude is measured along circles with radius $R \cos \varphi$, where φ is latitude. Thus 1° latitude = 111 km, and 1° longitude = $111 \cos \varphi$ km.

Because distance in degrees of longitude is not constant, oceanographers measure distance on maps using degrees of latitude.

Nautical miles and meters are connected historically to the size of Earth. Gabriel Mouton proposed in 1670 a decimal system of measurement based on the length of an arc that is one minute of a great circle of Earth. This eventually became the nautical mile. Mouton's decimal system eventually became the metric system based on a different unit of length, the meter, which was originally intended to be one ten-millionth the distance from the Equator to the pole along the Paris meridian. Although the tie between nautical miles, meters, and Earth's radius was soon abandoned because it was not practical, the approximations are very good. For example, Earth's polar circumference is approximately 40,008 km. Therefore one ten-millionth of a quadrant is 1.0002 m. Similarly, a nautical mile should be 1.8522 km, which is very close to the official definition of the international nautical mile: $1 \text{ nm} \equiv 1.8520 \text{ km}$.

[3.1: Ocean and Seas](#)

[3.2: Dimensions of the Ocean](#)

[3.3: Sea-Floor Features](#)

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3.1: Ocean and Seas

There is only one ocean. It is divided into three named parts by international agreement: the Atlantic, Pacific, and Indian Oceans (International Hydrographic Bureau, 1953). Seas, which are part of the ocean, are defined in several ways. I consider two.

The Atlantic Ocean extends northward from Antarctica and includes all of the Arctic Sea, the European Mediterranean, and the American Mediterranean more commonly known as the Caribbean sea (figure 3.1.1). The boundary between the Atlantic and Indian Ocean is the meridian of Cape Agulhas (20°E). The boundary between the Atlantic and Pacific is the line forming the shortest distance from Cape Horn to the South Shetland Islands. In the north, the Arctic Sea is part of the Atlantic Ocean, and the Bering Strait is the boundary between the Atlantic and Pacific.

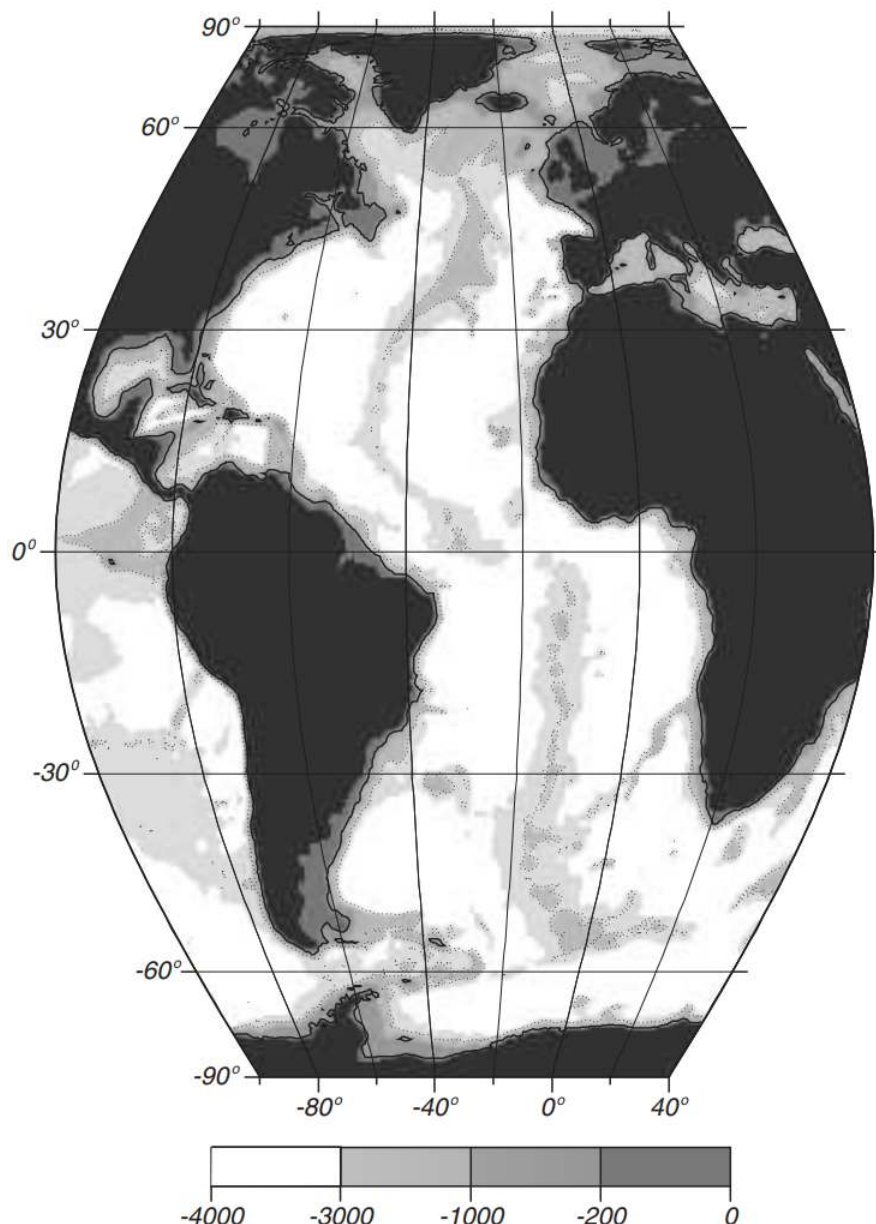


Figure 3.1.1: The Atlantic Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the etopo 30' data set. The 200 m contour outlines continental shelves.

The Pacific Ocean extends northward from Antarctica to the Bering Strait (figure 3.1.2). The boundary between the Pacific and Indian Ocean follows the line from the Malay Peninsula through Sumatra, Java, Timor, Australia at Cape Londonderry, and Tasmania. From Tasmania to Antarctica it is the meridian of South East Cape on Tasmania 147°E.

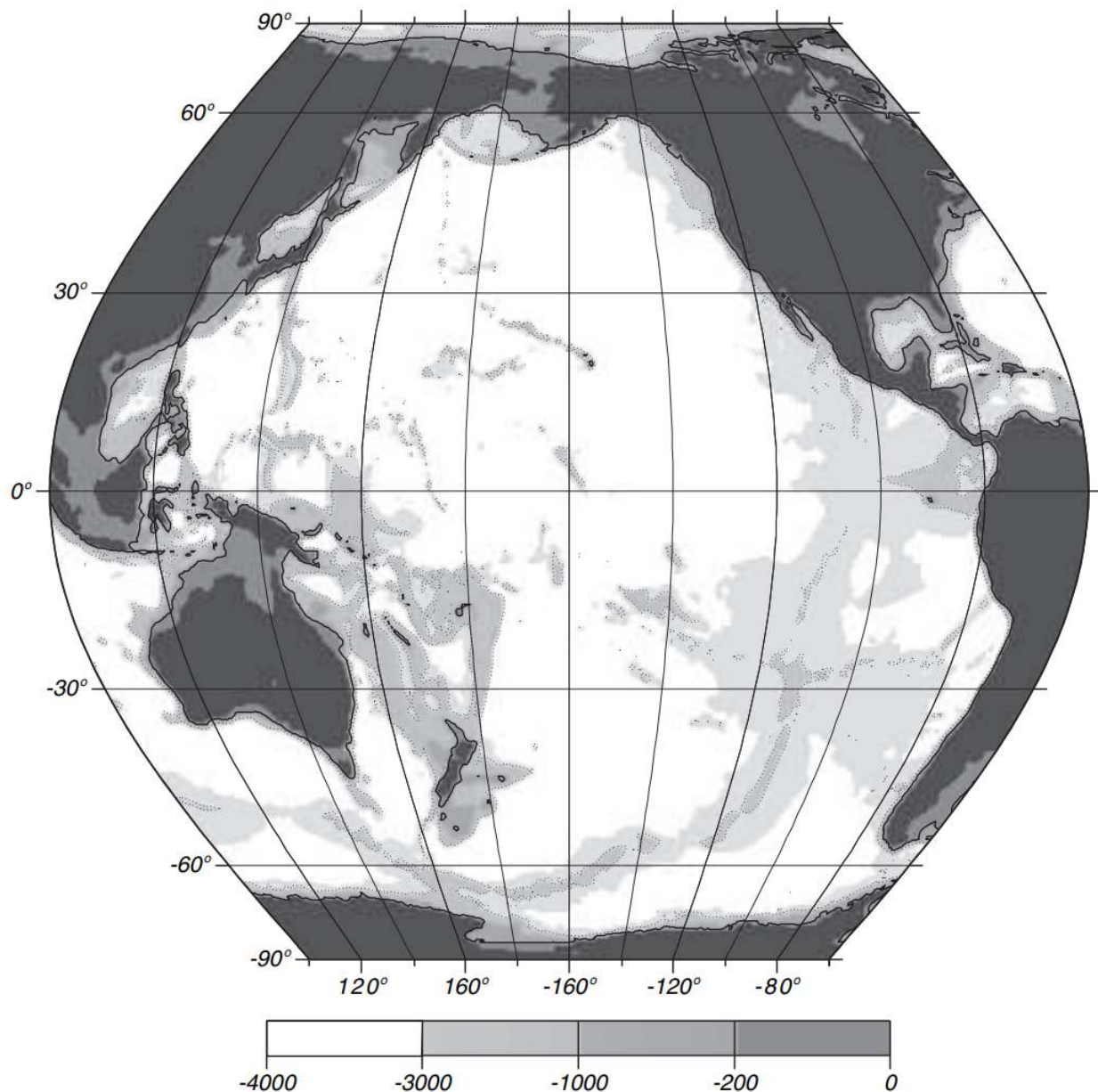


Figure 3.1.2: The Pacific Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the etopo 30' data set. The 200 m contour outlines continental shelves.

The Indian Ocean extends from Antarctica to the continent of Asia including the Red Sea and Persian Gulf (figure 3.1.3). Some authors use the name Southern Ocean to describe the ocean surrounding Antarctica.

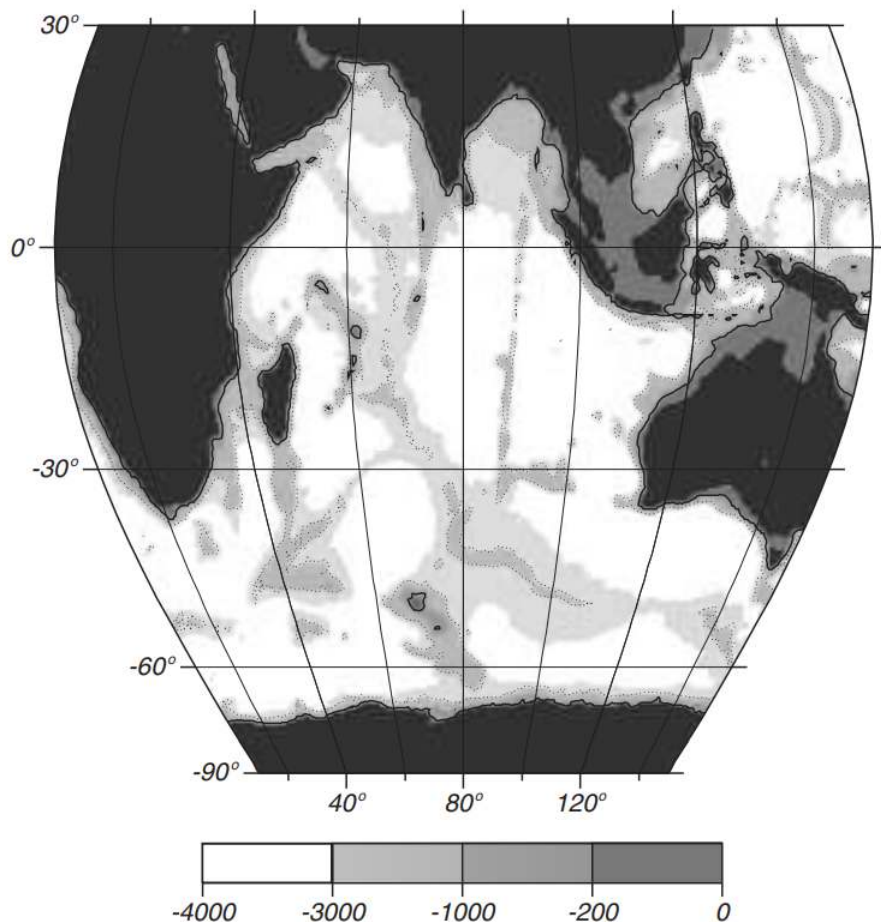


Figure 3.1.3: The Indian Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the etopo 30' data set. The 200 m contour outlines continental shelves.

Mediterranean Seas are mostly surrounded by land. By this definition, the Arctic and Caribbean Seas are both Mediterranean Seas, the Arctic Mediterranean and the Caribbean Mediterranean.

Marginal Seas are defined by only an indentation in the coast. The Arabian Sea and South China Sea are marginal seas.

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3.2: Dimensions of the Ocean

The ocean and seas cover 70.8% of the surface of earth, which amounts to $361,254,000 \text{ km}^2$. The areas of the named parts vary considerably (table 3.2.1).

Table 3.2.1. Surface Area of the ocean[†].

Ocean	Surface Area (km^2)
Pacific Ocean	181.34×10^6
Atlantic Ocean	106.57×10^6
Indian Ocean	74.12×10^6

[†]From Menard and Smith (1966)

Oceanic dimensions range from around 1500 km for the minimum width of the Atlantic to more than 13,000 km for the north-south extent of the Atlantic and the width of the Pacific. Typical depths are only 3–4 km. So horizontal dimensions of ocean basins are 1,000 times greater than the vertical dimension. A scale model of the Pacific, the size of an 8.5×11 in sheet of paper, would have dimensions similar to the paper: a width of 10,000 km scales to 10 in., and a depth of 3 km scales to 0.003 in., the typical thickness of a piece of paper.

Because the ocean is so thin, cross-sectional plots of ocean basins must have a greatly exaggerated vertical scale to be useful. Typical plots have a vertical scale that is 200 times the horizontal scale (figure 3.2.1). This exaggeration distorts our view of the ocean. The edges of the ocean basins, the continental slopes, are not steep cliffs as shown in the figure at 41°W and 12°E . Rather, they are gentle slopes dropping down 1 meter for every 20 meters in the horizontal.

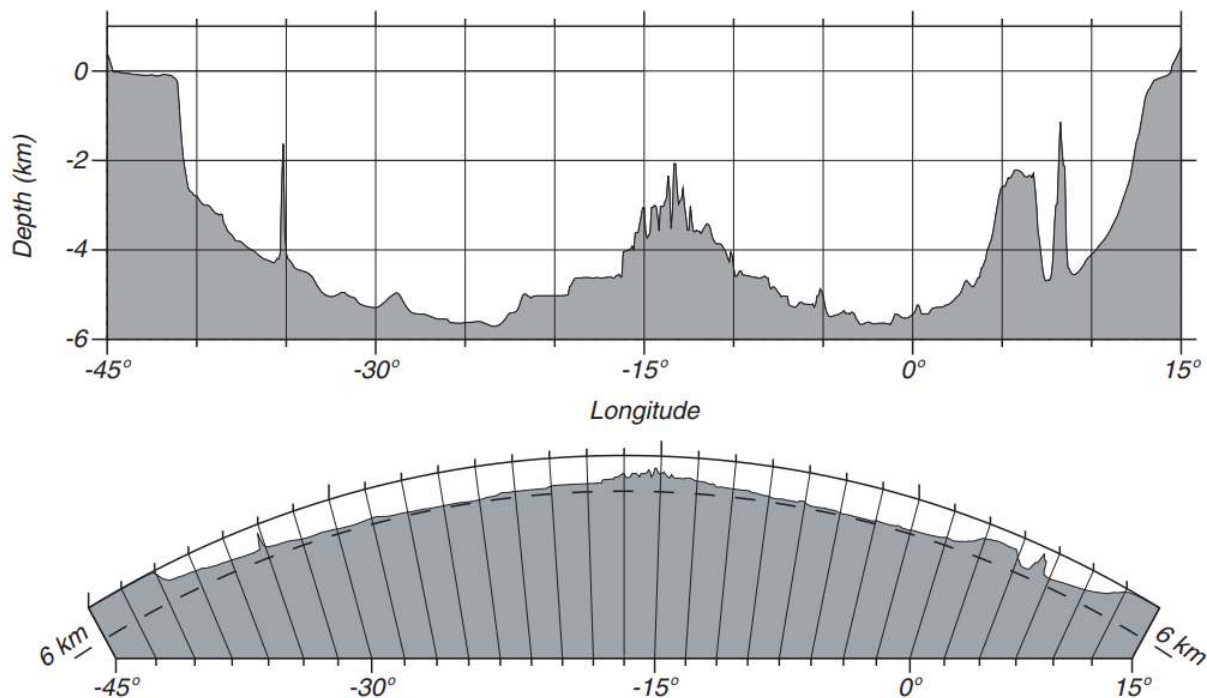


Figure 3.2.1: Cross-section of the south Atlantic along 25°S showing the continental shelf offshore of South America, a seamount near 35°W , the mid-Atlantic Ridge near 14°W , the Walvis Ridge near 6°E , and the narrow continental shelf off South Africa. **Upper:** Vertical exaggeration of 180:1. **Lower:** Vertical exaggeration of 30:1. If shown with true aspect ratio, the plot would be the thickness of the line at the sea surface in the lower plot.

The small ratio of depth to width of the ocean basins is very important for understanding ocean currents. Vertical velocities must be much smaller than horizontal velocities. Even over distances of a few hundred kilometers, the vertical velocity must be less than 1% of the horizontal velocity. I will use this information later to simplify the equations of motion.

The relatively small vertical velocities have great influence on turbulence. Three-dimensional turbulence is fundamentally different than two-dimensional turbulence. In two dimensions, vortex lines must always be vertical, and there can be little vortex stretching. In three dimensions, vortex stretching plays a fundamental role in turbulence.

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3.3: Sea-Floor Features

Earth's rocky surface is divided into two types: oceanic, with a thin dense crust about 10 km thick, and continental, with a thick light crust about 40 km thick. The deep, lighter continental crust floats higher on the denser mantle than does the oceanic crust, and the mean height of the crust relative to sea level has two distinct values: continents have a mean elevation of 1100 m, the ocean has a mean depth of -3400 m (figure 3.3.1).

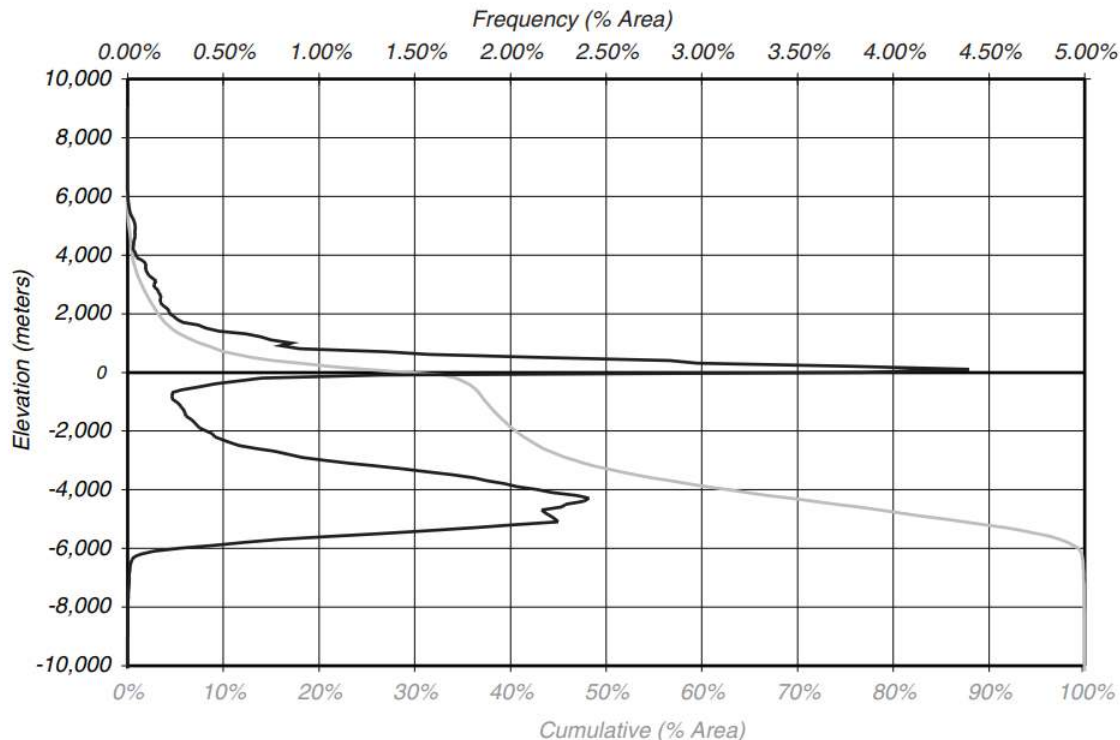


Figure 3.3.1: Histogram of height of land and depth of the sea as percentage of area of earth in 100-m intervals, showing the clear distinction between continents and sea floor. The cumulative frequency curve is the integral of the histogram. The curves are calculated from the etopo 2 data set by George Sharman of the NOAA National Geophysical Data Center.

The volume of the water in the ocean exceeds the volume of the ocean basins, and some water spills over on to the low lying areas of the continents. These shallow seas are the continental shelves. Some, such as the South China Sea, are more than 1100 km wide. Most are relatively shallow, with typical depths of 50–100 m. A few of the more important shelves are: the East China Sea, the Bering Sea, the North Sea, the Grand Banks, the Patagonian Shelf, the Arafura Sea and Gulf of Carpentaria, and the Siberian Shelf. The shallow seas help dissipate tides, they are often areas of high biological productivity, and they are usually included in the exclusive economic zone of adjacent countries.

The crust is broken into large plates that move relative to each other. New crust is created at the mid-ocean ridges, and old crust is lost at trenches. The relative motion of crust, due to plate tectonics, produces the distinctive features of the sea floor sketched in Figure 3.3.2, including mid-ocean ridges, trenches, island arcs, and basins. The names of the sub-sea features have been defined by the International Hydrographic Organization (1953), and the following definitions are taken from Sverdrup, Johnson, and Fleming (1942), Shepard (1963), and Dietrich et al. (1980).

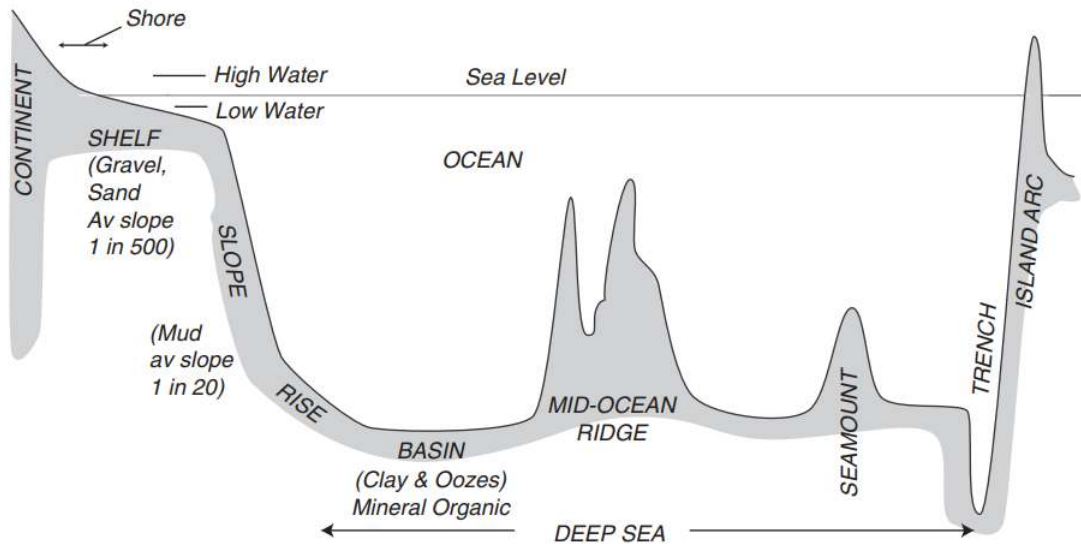


Figure 3.3.2: Schematic section through the ocean showing principal features of the sea floor. Note that the slope of the sea floor is greatly exaggerated in the figure.

Basins are deep depressions of the sea floor of more or less circular or oval form.

Canyons are relatively narrow, deep furrows with steep slopes, cutting across the continental shelf and slope, with bottoms sloping continuously downward.

Continental shelves are zones adjacent to a continent (or around an island) and extending from the low-water line to the depth, usually about 120 m, where there is a marked or rather steep descent toward great depths. (figure 3.3.3)

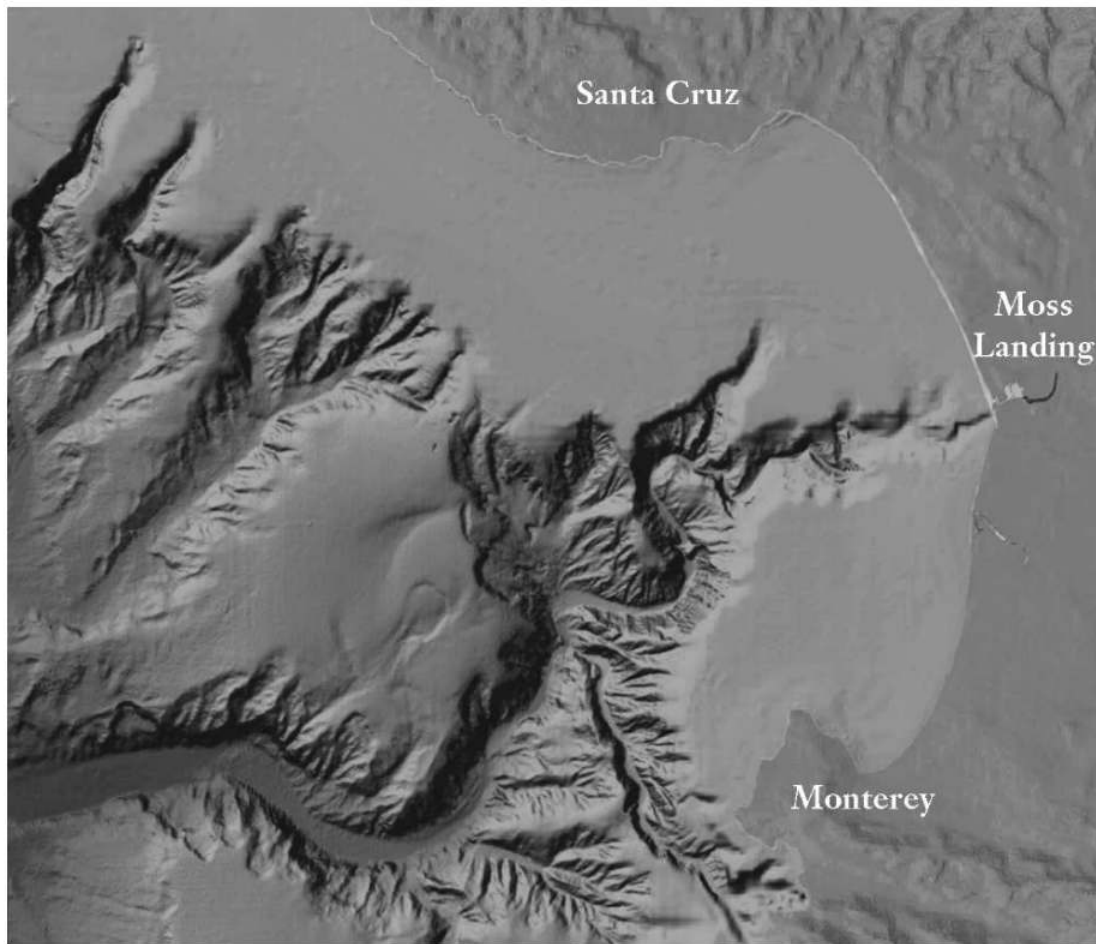


Figure 3.3.3: An example of a continental shelf, the shelf offshore of Monterey California showing the Monterey and other canyons. Canyons are common on shelves, often extending across the shelf and down the continental slope to deep water. Figure copyright Monterey Bay Aquarium Research Institute (MBARI).

Continental slopes are the declivities seaward from the shelf edge into greater depth.

Plains are very flat surfaces found in many deep ocean basins.

Ridges are long, narrow elevations of the sea floor with steep sides and rough topography.

Seamounts are isolated or comparatively isolated elevations rising 1000 m or more from the sea floor and with small summit area (figure 3.3.4).

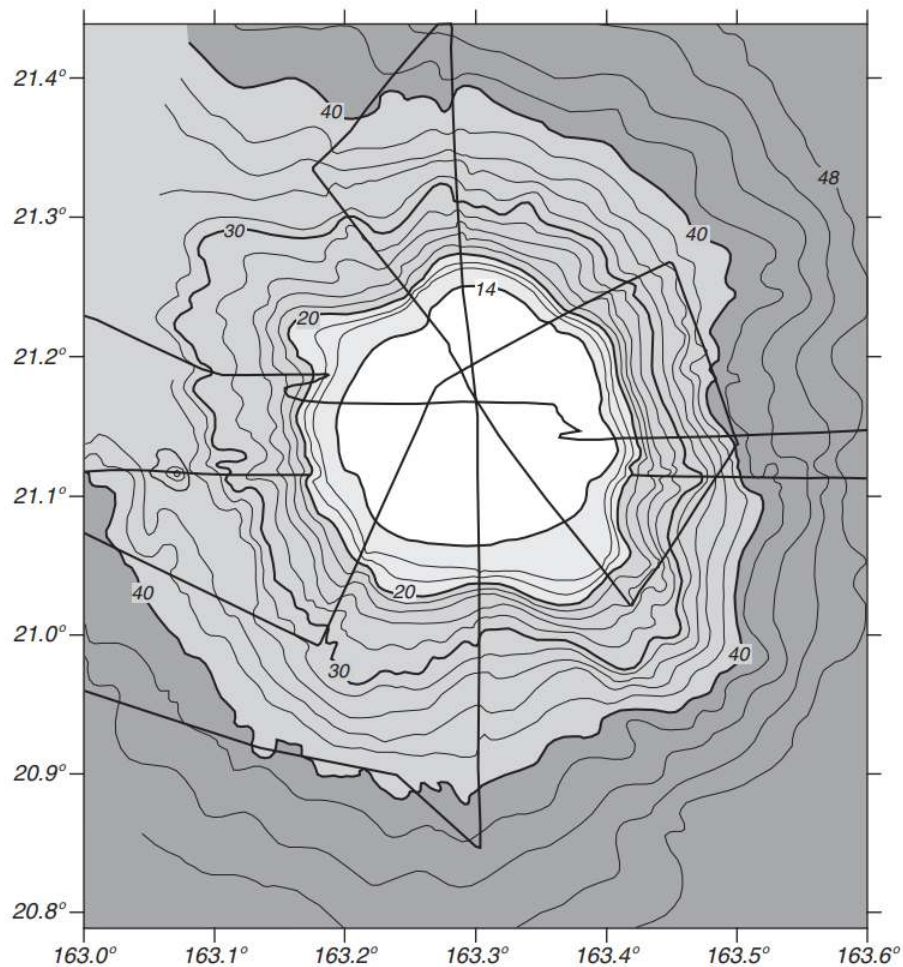


Figure 3.3.4: An example of a seamount, the Wilde Guyot. A guyot is a seamount with a flat top created by wave action when the seamount extended above sea level. As the seamount is carried by plate motion, it gradually sinks deeper below sea level. The depth was contoured from echo sounder data collected along the ship track (thin straight lines) supplemented with side-scan sonar data. Depths are in units of 100 m. From William Sager, Texas A&M University

Sills are the low parts of the ridges separating ocean basins from one another or from the adjacent sea floor.

Trenches are long, narrow, and deep depressions of the sea floor, with relatively steep sides (figure 3.3.5).

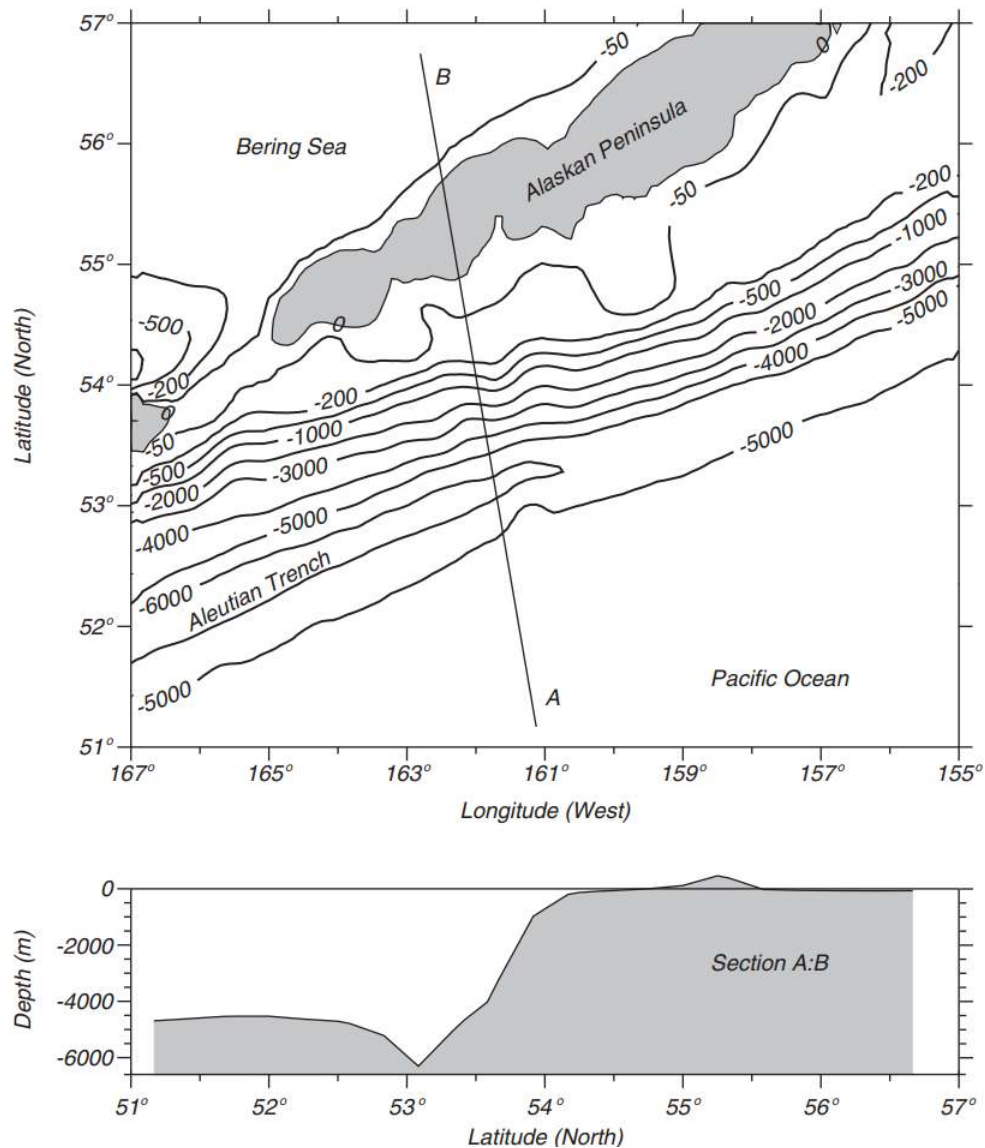


Figure 3.3.5: An example of a trench, the Aleutian Trench; an island arc, the Alaskan Peninsula; and a continental shelf, the Bering Sea. The island arc is composed of volcanos produced when oceanic crust carried deep into a trench melts and rises to the surface.
Top: Map of the Aleutian region of the North Pacific. **Bottom:** Cross-section through the region.

Sub-sea features strongly influence the ocean circulation. Ridges separate deep waters of the ocean into distinct basins. Water deeper than the sill between two basins cannot move from one to the other. Tens of thousands of seamounts are scattered throughout the ocean basins. They interrupt ocean currents and produce turbulence, leading to vertical mixing in the ocean.

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3.4: Measuring the Depth of the Ocean

The depth of the ocean is usually measured two ways: 1) using acoustic echo-sounders on ships, or 2) using data from satellite altimeters.

Echo Sounders

Most maps of the ocean are based on measurements made by echo sounders. The instrument transmits a burst of 10–30 kHz sound and listens for the echo from the sea floor. The time interval between transmission of the pulse and reception of the echo, when multiplied by the velocity of sound, gives twice the depth of the ocean (figure 3.4.1).

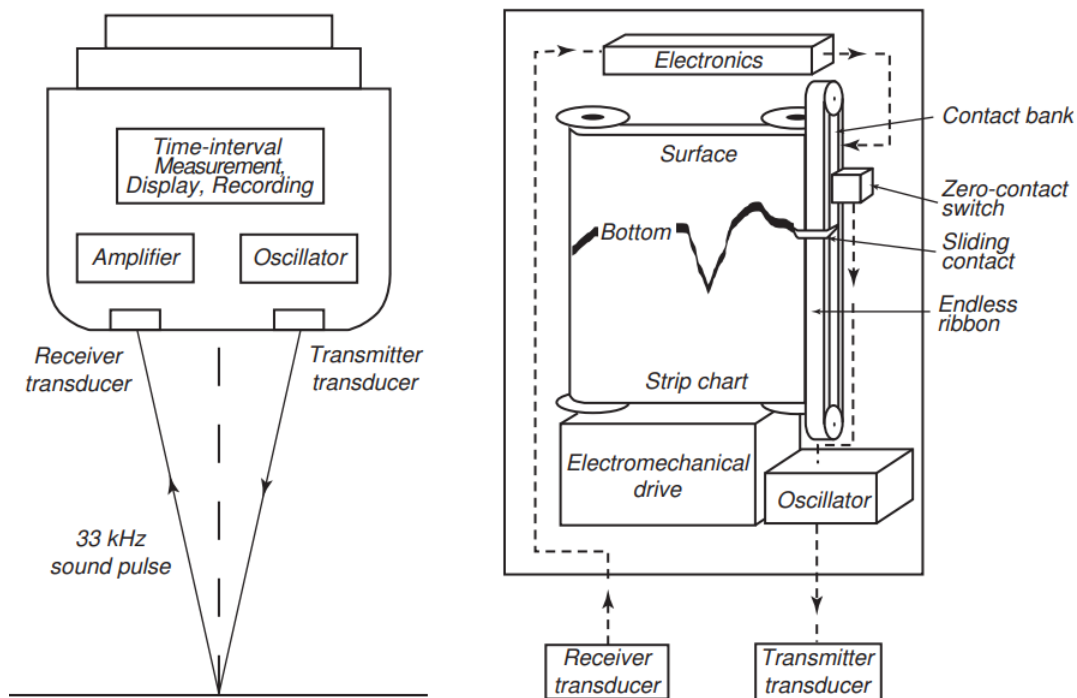


Figure 3.4.1: **Left:** Echo sounders measure depth of the ocean by transmitting pulses of sound and observing the time required to receive the echo from the bottom. **Right:** The time is recorded by a spark burning a mark on a slowly moving roll of paper. After Dietrich et al. (1980: 124).

The first transatlantic echo soundings were made by the U.S. Navy Destroyer *Stewart* in 1922. This was quickly followed by the first systematic survey of an ocean basin, made by the German research and survey ship *Meteor* during its expedition to the south Atlantic from 1925 to 1927. Since then, oceanographic and naval ships have operated echo sounders almost continuously while at sea. Millions of miles of ship-track data recorded on paper have been digitized to produce data bases used to make maps. The tracks are not well distributed.

Tracks tend to be far apart in the southern hemisphere, even near Australia (figure 3.4.2), and closer together in well-mapped areas such as the North Atlantic. Echo sounders make the most accurate measurements of ocean depth. Their accuracy is $\pm 1\%$.

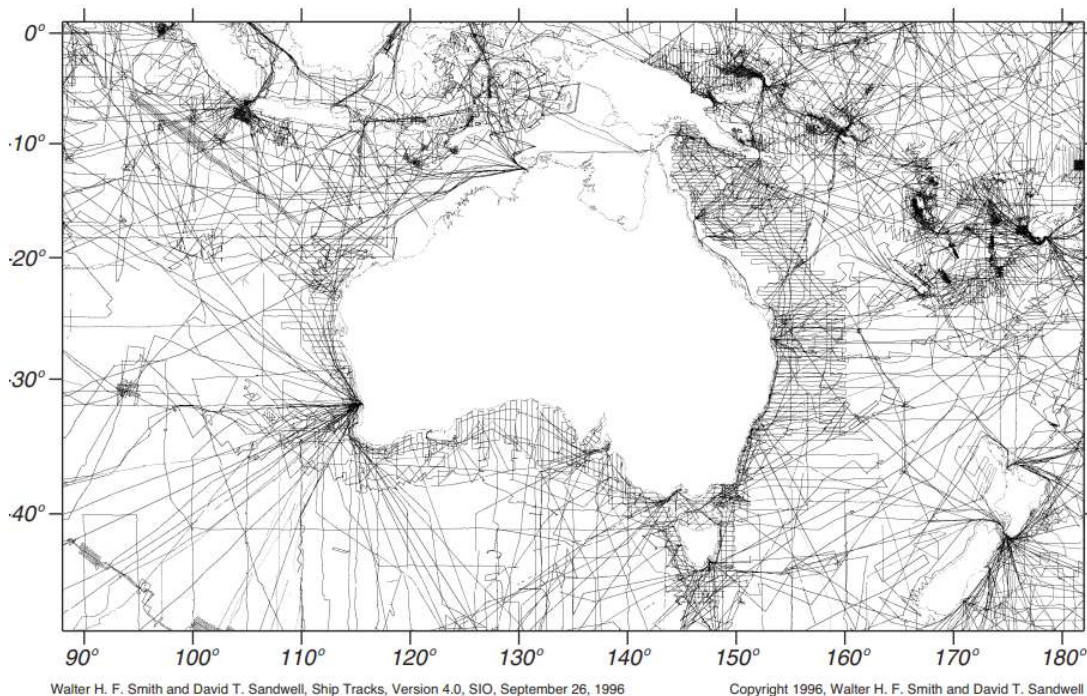


Figure 3.4.2: Locations of echo-sounder data used for mapping the ocean floor near Australia. Note the large areas where depths have not been measured from ships. From David Sandwell, Scripps Institution of Oceanography.

Satellite Altimetry

Gaps in our knowledge of ocean depths between ship tracks have now been filled by satellite-altimeter data. Altimeters profile the shape of the sea surface, and this shape is very similar to that of the sea floor (Tapley and Kim, 2001; Cazenave and Royer, 2001; Sandwell and Smith, 2001). To see this, we must first consider how gravity influences sea level.

The Relationship Between Sea Level and the Ocean's Depth

Excess mass at the sea floor, for example the mass of a seamount, increases local gravity because the mass of the seamount is larger than the mass of water it displaces. Rocks are more than three times denser than water. The excess mass increases local gravity, which attracts water toward the seamount. This changes the shape of the sea surface (figure 3.4.3).

Let's make the concept more exact. To a very good approximation, the sea surface is a particular level surface called the **geoid** (see box). By definition a level surface is a surface of constant gravitational potential, and it is everywhere perpendicular to gravity. In particular, it must be perpendicular to the local vertical determined by a plumb line, which is "a line or cord having at one end a metal weight for determining vertical direction" (Oxford English Dictionary).

The Geoid

The level surface that corresponds to the surface of an ocean at rest is a special surface, the **geoid**. To a first approximation, the geoid is an ellipsoid that corresponds to the surface of a rotating, homogeneous fluid in solid-body rotation, which means that the fluid has no internal flow. To a second approximation, the geoid differs from the ellipsoid because of local variations in gravity. The deviations are called **geoid undulations**. The maximum amplitude of the undulations is roughly ± 60 m. To a third approximation, the geoid deviates from the sea surface because the ocean is not at rest. The deviation of sea level from the geoid is defined to be the **topography**. The definition is identical to the definition for land topography, for example the heights given on a topographic map.

The ocean's topography is caused by tides, heat content of the water, and ocean surface currents. I will return to their influence in chapters 10 and 17. The maximum amplitude of the topography is roughly ± 1 m, so it is small compared to the geoid undulations.

Geoid undulations are caused by local variations in gravity due to the uneven distribution of mass at the sea floor. Seamounts have an excess of mass because they are more dense than water. They produce an upward bulge in the geoid (see below).

Trenches have a deficiency of mass. They produce a downward deflection of the geoid. Thus the geoid is closely related to sea-floor topography. Maps of the oceanic geoid have a remarkable resemblance to the sea-floor topography.

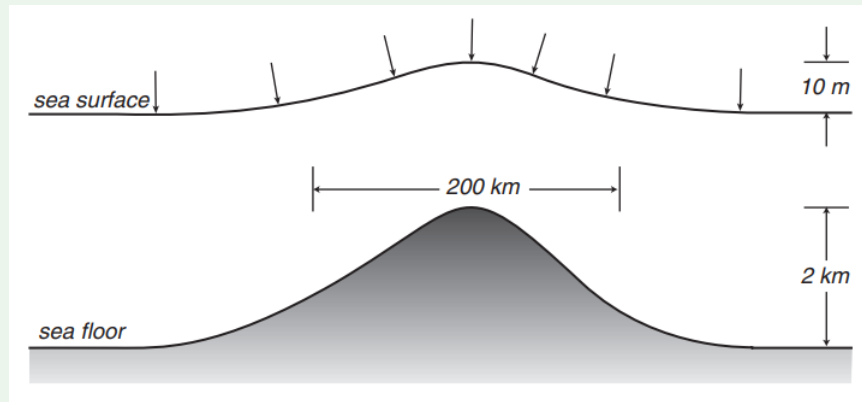


Figure 3.4.3: Seamounts are more dense than sea water. They increase local gravity, causing a plumb line at the sea surface (arrows) to be deflected toward the seamount. Because the surface of an ocean at rest must be perpendicular to gravity, the sea surface and the local geoid must have a slight bulge as shown. Such bulges are easily measured by satellite altimeters. As a result, satellite altimeter data can be used to map the sea floor. Note: the bulge at the sea surface is greatly exaggerated; a two-kilometer high seamount would produce a bulge of approximately 10 m.

The excess mass of the seamount attracts the plumb line's weight, causing the plumb line to point a little toward the seamount instead of toward Earth's center of mass. Because the sea surface must be perpendicular to gravity, it must have a slight bulge above a seamount as shown in figure 3.4.3. If there were no bulge, the sea surface would not be perpendicular to gravity. Typical seamounts produce a bulge that is 1–20 m high over distances of 100–200 kilometers. This bulge is far too small to be seen from a ship, but it is easily measured by satellite altimeters. Oceanic trenches have a deficit of mass, and they produce a depression of the sea surface.

The correspondence between the shape of the sea surface and the depth of the water is not exact. It depends on the strength of the sea floor, the age of the sea-floor feature, and the thickness of sediments. If a seamount floats on the sea floor like ice on water, the gravitational signal is much weaker than it would be if the seamount rested on the sea floor like ice resting on a tabletop. As a result, the relationship between gravity and sea-floor topography varies from region to region.

Depths measured by acoustic echo sounders are used to determine the regional relationships. Hence, altimetry is used to interpolate between acoustic echo sounder measurements (Smith and Sandwell, 1994).

Satellite-altimeter systems

Now let's see how altimeters measure the shape of the sea surface. Satellite altimeter systems include a radar to measure the height of the satellite above the sea surface and a tracking system to determine the height of the satellite in geocentric coordinates. The system measures the height of the sea surface relative to the center of mass of earth (figure 3.4.4). This gives the shape of the sea surface.

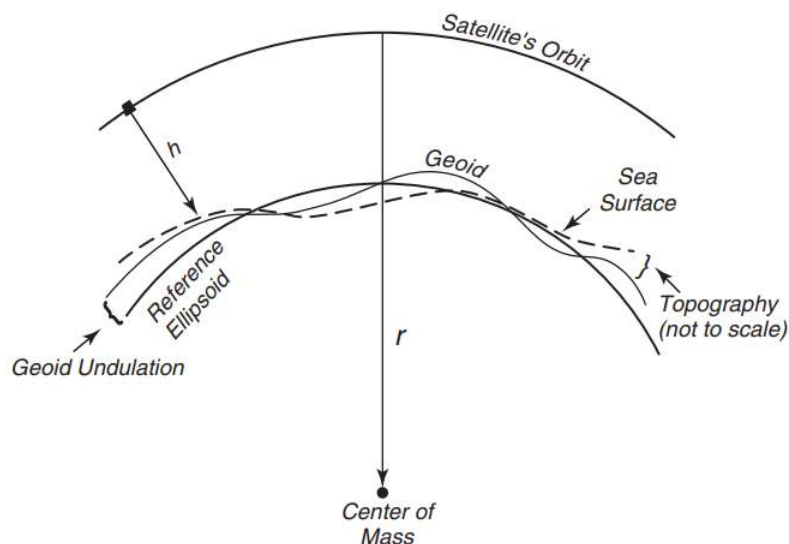


Figure 3.4.4: A satellite altimeter measures the height of the satellite above the sea surface. When this is subtracted from the height r of the satellite's orbit, the difference is sea level relative to the center of Earth. The shape of the surface is due to variations in gravity, which produce the geoid undulations, and to ocean currents which produce the oceanic topography, the departure of the sea surface from the geoid. The reference ellipsoid is the best smooth approximation to the geoid. The variations in the geoid, geoid undulations, and topography are greatly exaggerated in the figure. From Stewart (1985).

Many altimetric satellites have flown in space. All observed the marine geoid and the influence of sea-floor features on the geoid. The altimeters that produced the most useful data include Seasat (1978), GEOSAT (1985–1988), ERS-1 (1991– 1996), ERS-2 (1995–), Topex/Poseidon (1992–2006), Jason (2002–), and Envisat (2002). Topex/Poseidon and Jason were specially designed to make extremely accurate measurements of sea-surface height. They measure sea-surface height with an accuracy of ± 0.05 m.

Satellite Altimeter Maps of the Sea-floor Topography

Seasat, GEOSAT, ERS-1, and ERS-2 were operated in orbits with ground tracks spaced 3–10 km apart, which was sufficient to map the geoid. By combining data from echo-sounders with data from GEOSAT and ERS-1 altimeter systems, Smith and Sandwell (1997) produced maps of the sea floor with horizontal resolution of 5–10 km and a global average depth accuracy of ± 100 m.

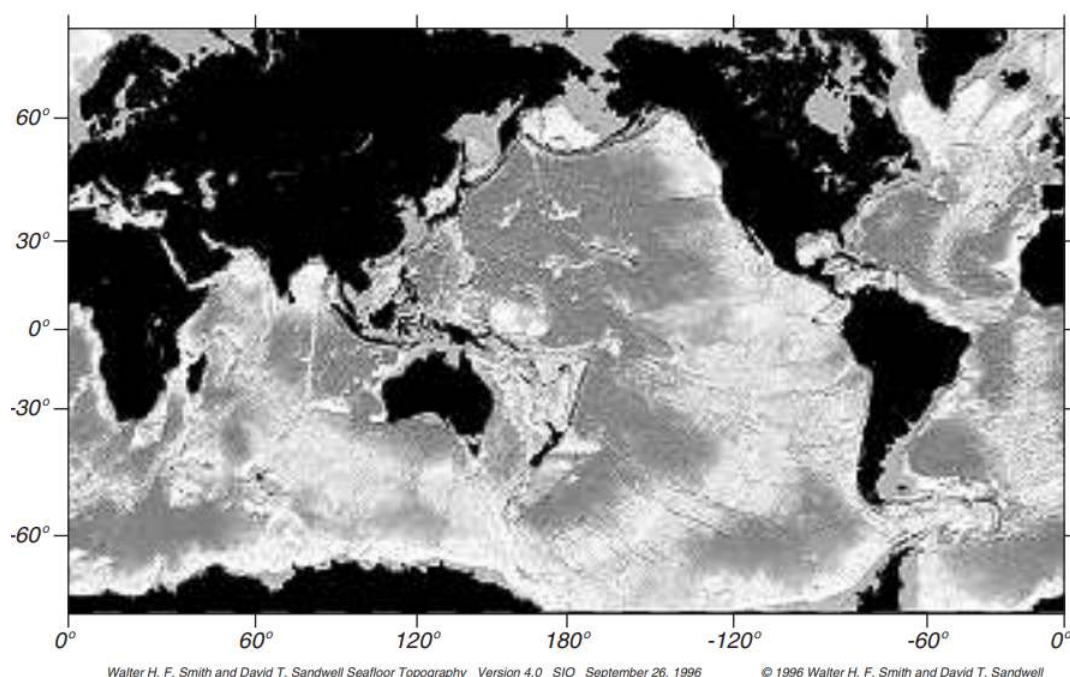


Figure 3.4.5: The sea-floor topography of the ocean with 3 km resolution produced from satellite altimeter observations of the shape of the sea surface. From Smith and Sandwell.

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3.5: Sea Floor Charts and Data Sets

Almost all echo-sounder data have been digitized and combined to make seafloor charts. Data have been further processed and edited to produce digital data sets which are widely distributed in CD-ROM format. These data have been supplemented with data from altimetric satellites to produce maps of the sea floor with horizontal resolution around 3 km.

The British Oceanographic Data Centre publishes the General Bathymetric Chart of the ocean (GEBCO) Digital Atlas on behalf of the Intergovernmental Oceanographic Commission of UNESCO and the International Hydrographic Organization. The atlas consists primarily of the location of depth contours, coastlines, and tracklines from the GEBCO 5th Edition published at a scale of 1:10 million. The original contours were drawn by hand based on digitized echo-sounder data plotted on base maps.

The U.S. National Geophysical Data Center publishes the ETOPO-2 CD-ROM containing digital values of oceanic depths from echo sounders and altimetry and land heights from surveys. Data are interpolated to a 2-minute (2 nautical mile) grid. Ocean data between 64°N and 72°S are from the work of Smith and Sandwell (1997), who combined echo-sounder data with altimeter data from GEOSAT and ERS-1. Seafloor data northward of 64°N are from the International Bathymetric Chart of the Arctic Ocean. Seafloor data southward of 72°S are from the US Naval Oceanographic Office's Digital Bathymetric Data Base Variable Resolution. Land data are from the GLOBE Project, that produced a digital elevation model with 0.5-minute (0.5 nautical mile) grid spacing using data from many nations.

National governments publish coastal and harbor maps. In the USA, the NOAA National Ocean Service publishes nautical charts useful for navigation of ships in harbors and offshore waters.

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3.6: Sound in the Ocean

Sound provides the only convenient means for transmitting information over great distances in the ocean. Sound is used to measure the properties of the sea floor, the depth of the ocean, temperature, and currents. Whales and other ocean animals use sound to navigate, communicate over great distances, and find food.

Sound Speed

The sound speed in the ocean varies with temperature, salinity, and pressure (MacKenzie, 1981; Munk et al. 1995: 33):

$$C = 1448.96 + 4.591t - 0.05304t^2 + 0.0002374t^3 + 0.0160Z \\ + (1.340 - 0.01025t)(S - 35) + 1.675 \times 10^{-7}Z^2 - 7.139 \times 10^{-13}tZ^3$$

where C is speed in m/s, t is temperature in Celsius, S is salinity (see [Chapter 6](#) for a definition of salinity), and Z is depth in meters. The equation has an accuracy of about 0.1 m/s (Dushaw et al. 1993). Other sound-speed equations have been widely used, especially an equation proposed by Wilson (1960) which has been widely used by the U.S. Navy.

For typical oceanic conditions, C is usually between 1450 m/s and 1550 m/s (figure 3.6.1). Using Equation (3.6.1), we can calculate the sensitivity of C to changes of temperature, depth, and salinity typical of the ocean. The approximate values are: 40 m/s per 10°C rise of temperature, 16 m/s per 1000 m increase in depth, and 1.5 m/s per 1 increase in salinity. Thus the primary causes of variability of sound speed is temperature and depth (pressure). Variations of salinity are too small to have much influence.

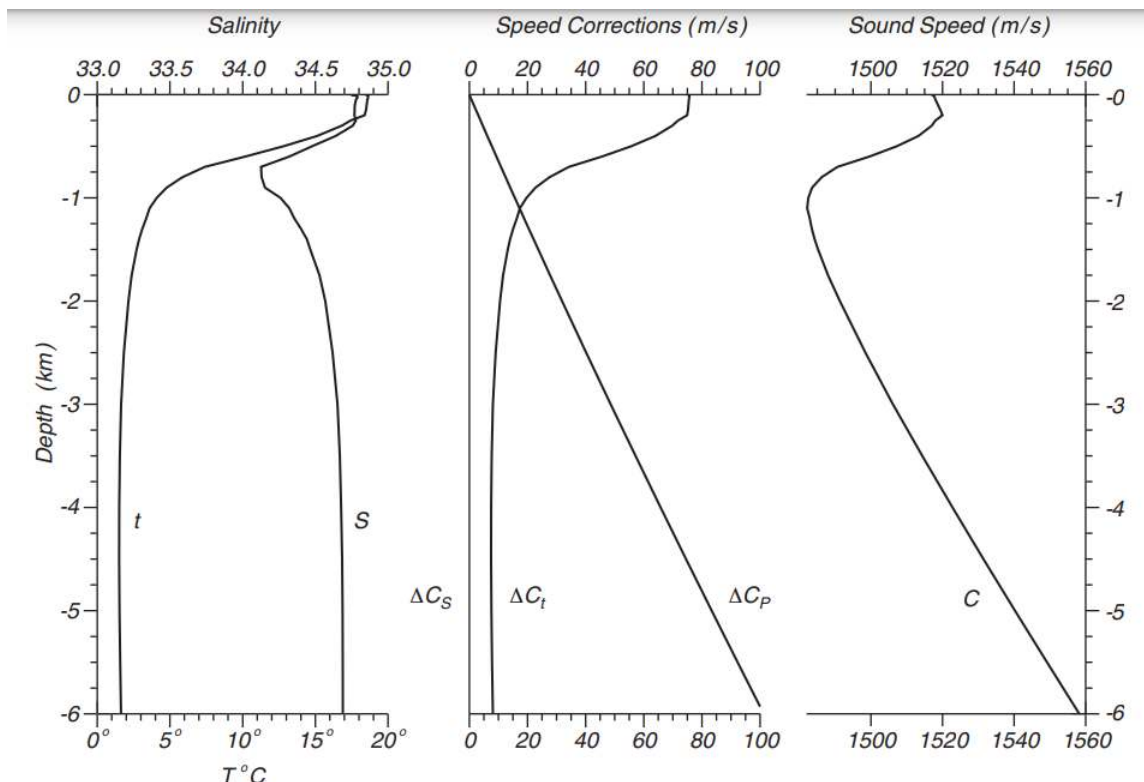


Figure 3.6.1: Processes producing the sound channel in the ocean. **Left:** Temperature T and salinity S measured as a function of depth during the R.V. *Hakuho Maru* cruise KH-87-1, station JT, on 28 January 1987 at Latitude $33^\circ 52.90' \text{ N}$, Long $141^\circ 55.80' \text{ E}$ in the North Pacific. **Center:** Variations in sound speed due to variations in temperature, salinity, and depth. **Right:** Sound speed as a function of depth, showing the velocity minimum near 1 km depth which defines the sound channel in the ocean. (Data from JPOTS Editorial Panel, 1991).

If we plot sound speed as a function of depth, we find that the speed usually has a minimum at a depth around 1000 m (figure 3.6.2). The depth of minimum speed is called the **sound channel**. It occurs in all oceans, and it usually reaches the surface at very high latitudes. The sound channel is important because sound in the channel can travel very far, sometimes halfway around the earth. Here is how the channel works: Sound rays that begin to travel out of the channel are refracted back toward the center of the channel. Rays propagating upward at small angles to the horizontal are bent downward, and rays propagating downward at small

angles to the horizontal are bent upward (figure 3.6.2). Typical depths of the channel vary from 10 m to 1200 m depending on geographical area.

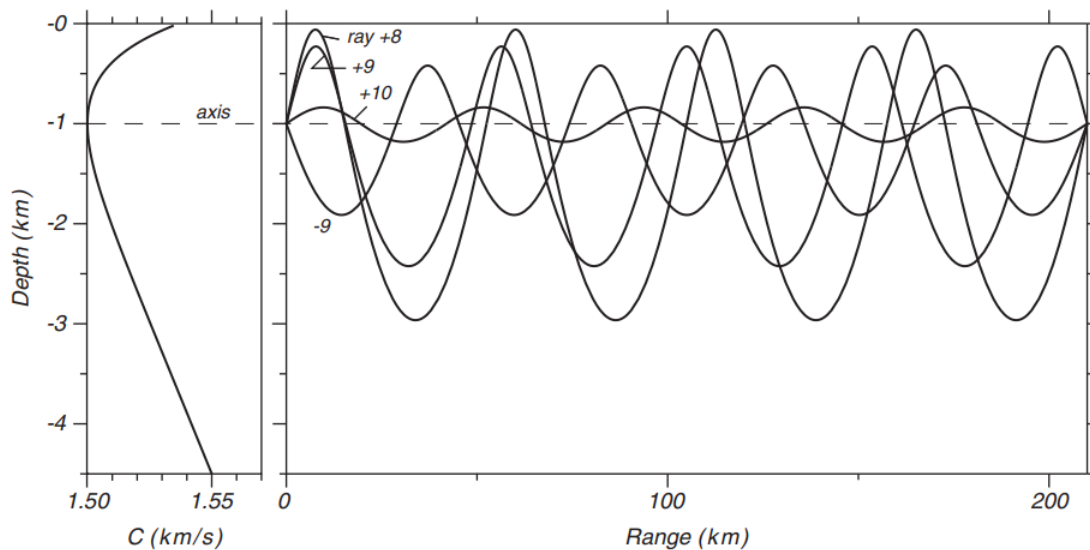


Figure 3.6.2: Ray paths of sound in the ocean for a source near the axis of the sound channel. After Munk et al. (1995).

Absorption of Sound

Absorption of sound per unit distance depends on the intensity I of the sound:

$$dI = -kI_0 dx$$

where I_0 is the intensity before absorption and k is an absorption coefficient which depends on frequency of the sound. The equation has the solution:

$$I = I_0 \exp(-kx)$$

Typical values of k (in decibels, or dB, per kilometer) are 0.08 dB/km at 1000 Hz, and 50 dB/km at 100,000 Hz. Decibels are calculated from: $dB = 10 \log(I/I_0)$, where I_0 is the original acoustic power and I is the acoustic power after absorption.

For example, at a range of 1 km a 1000 Hz signal is attenuated by only 1.8%: $I = 0.982I_0$. At a range of 1 km a 100,000 Hz signal is reduced to $I = 10^{-5}I_0$. The 30,000 Hz signal used by typical echo sounders to map the ocean's depths are attenuated little going from the surface to the bottom and back.

Very low-frequency sounds in the sound channel, those with frequencies below 500 Hz, have been detected at distances of megameters. In 1960 15-Hz sounds from explosions set off in the sound channel off Perth Australia were heard in the sound channel near Bermuda, nearly halfway around the world. Later experiment showed that 57-Hz signals transmitted in the sound channel near Heard Island (75°E, 53°S) could be heard at Bermuda in the Atlantic and at Monterey, California in the Pacific (Munk et al. 1994).

Use of Sound

Because low-frequency sound can be heard at great distances, in the 1950s the U.S. Navy placed arrays of microphones on the sea floor in deep and shallow water and connected them to shore stations. The Sound Surveillance System SOSUS, although designed to track submarines, has found many other uses. It has been used to listen to and track whales up to 1,700 km away, and to find the location of sub-sea volcanic eruptions.

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3.7: Important Concepts

1. If the ocean were scaled down to a width of 8 inches it would have depths about the same as the thickness of a piece of paper. As a result, the velocity field in the ocean is nearly 2-dimensional. Vertical velocities are much smaller than horizontal velocities.
 2. There are only three official oceans.
 3. The volume of ocean water exceeds the capacity of the ocean basins, and the ocean overflows on to the continents creating continental shelves.
 4. The depths of the ocean are mapped by echo sounders which measure the time required for a sound pulse to travel from the surface to the bottom and back. Depths measured by ship-based echo sounders have been used to produce maps of the sea floor. The maps have poor horizontal resolution in some regions because the regions were seldom visited by ships and ship tracks are far apart.
 5. The depths of the ocean are also measured by satellite altimeter systems which profile the shape of the sea surface. The local shape of the surface is influenced by changes in gravity due to sub-sea features. Recent maps based on satellite altimeter measurements of the shape of the sea surface combined with ship data have depth accuracy of ± 100 m and horizontal resolutions of ± 3 km.
 6. Typical sound speed in the ocean is 1480 m/s. Speed depends primarily on temperature, less on pressure, and very little on salinity. The variability of sound speed as a function of pressure and temperature produces a horizontal sound channel in the ocean. Sound in the channel can travel great distances. Low-frequency sounds below 500 Hz can travel halfway around the world provided the path is not interrupted by land.
-

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CHAPTER OVERVIEW

4: Atmospheric Influences

The sun and the atmosphere drive, directly or indirectly, almost all dynamical processes in the ocean. The dominant external sources and sinks of energy are sunlight, evaporation, infrared emissions from the sea surface, and sensible heating of the sea by warm or cold winds. Winds drive the ocean's surface circulation down to depths of around a kilometer. Wind and tidal mixing drive the deeper currents in the ocean.

The ocean, in turn, is the dominant source of heat that drives the atmospheric circulation. The uneven distribution of heat loss and gain by the ocean leads to winds in the atmosphere. Sunlight warms the tropical ocean, which evaporate, transferring heat in the form of water vapor to the atmosphere. The heat is released when the vapor condenses as rain. Winds and ocean currents carry heat poleward, where it is lost to space.

Because the atmosphere drives the ocean, and the ocean drives the atmosphere, we must consider the ocean and the atmosphere as a coupled dynamic system. In this chapter we will look mainly at the exchange of momentum between the atmosphere and the ocean. In [the next chapter](#), we will look at heat exchanges. In [Chapter 14](#) we will look at how the ocean and the atmosphere interact in the Pacific to produce El Niño.

[4.1: The Earth in Space](#)

[4.2: Atmospheric Wind Systems](#)

[4.3: The Planetary Boundary Layer](#)

[4.4: Measurement of Wind](#)

[4.5: Calculations of Wind](#)

[4.6: Wind Stress](#)

[4.7: Important Concepts](#)

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4.1: The Earth in Space

Earth's orbit about the sun is nearly circular at a mean distance of 1.5×10^8 km. The eccentricity of the orbit is small, 0.0168. Thus Earth is 3.4% further from the Sun at aphelion than at perihelion, the time of closest approach to the sun. Perihelion occurs every year in January, and the exact time changes by about 20 minutes per year. In 1995, it occurred on 3 January. Earth's axis of rotation is inclined 23.45° to the plane of Earth's orbit around the sun (figure 4.1.1). The orientation is such that the sun is directly overhead at the Equator on the vernal and autumnal equinoxes, which occur on or about 21 March and 21 September each year.

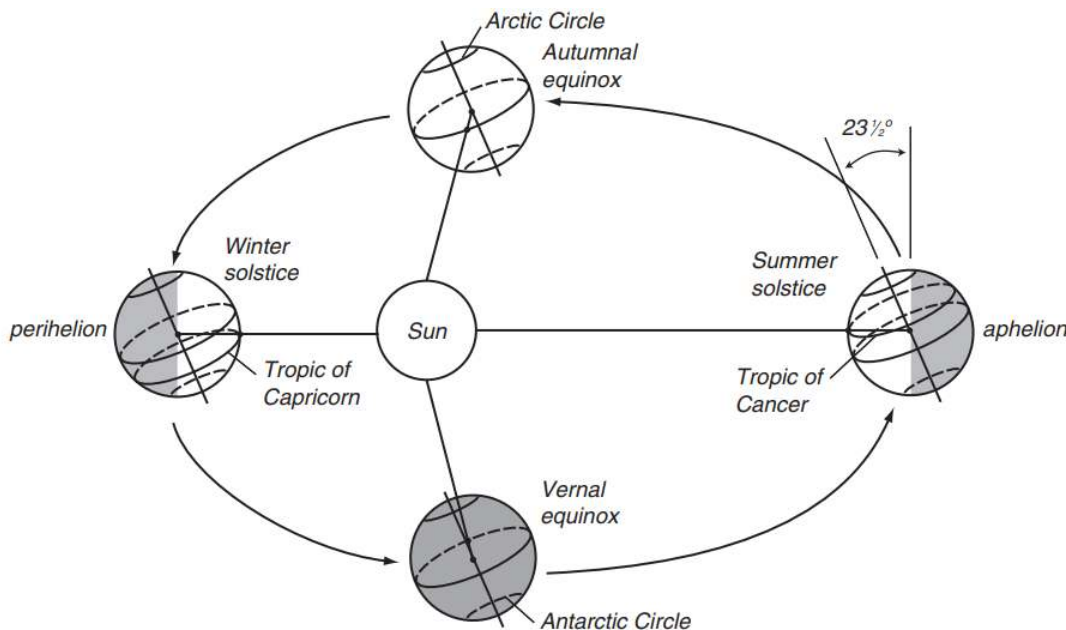


Figure 4.1.1: The Earth in space. The ellipticity of Earth's orbit around the sun and the tilt of Earth's axis of rotation relative to the plane of Earth orbit leads to an unequal distribution of heating and to the seasons. Earth is closest to the sun at perihelion.

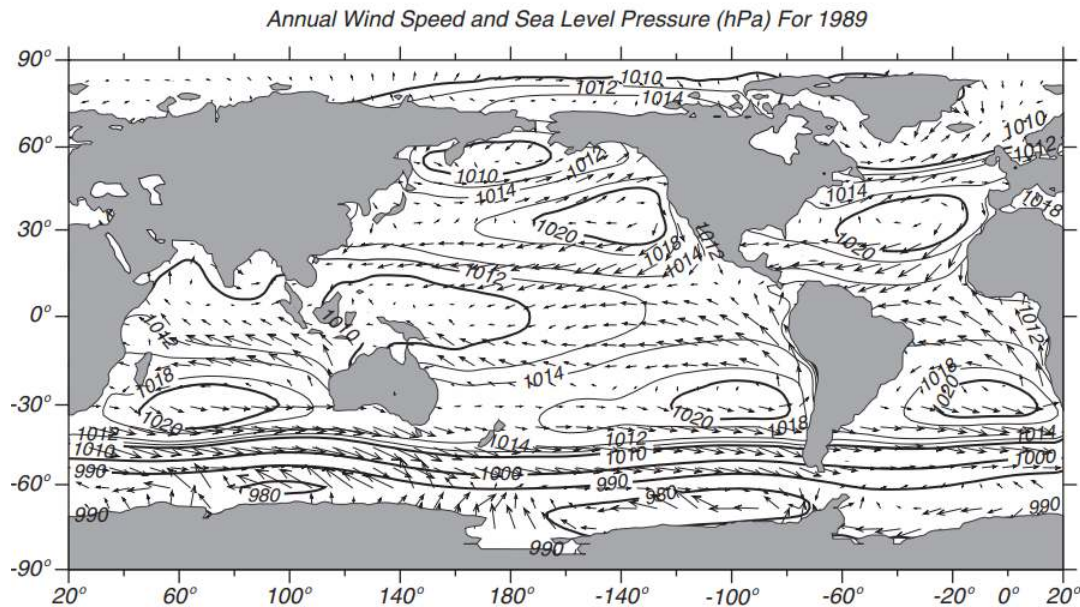
The latitudes of 23.45° North and South are the Tropics of Cancer and Capricorn respectively. The tropics lie equatorward of these latitudes. As a result of the eccentricity of earth's orbit, maximum solar insolation averaged over the surface of the earth occurs in early January each year. As a result of the inclination of earth's axis of rotation, the maximum insolation at any location outside the tropics occurs around 21 June in the northern hemisphere, and around 21 December in the southern hemisphere.

If solar heat was rapidly redistributed over earth, maximum temperature would occur in January. Conversely, if heat were poorly redistributed, maximum temperature in the northern hemisphere would occur in summer. So it is clear that heat is not rapidly redistributed by winds and currents.

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4.2: Atmospheric Wind Systems

Figure 4.2.1 shows the distribution of sea-level winds and pressure averaged over the year 1989. The map shows strong winds from the west between 40° to 60° latitude, the roaring forties, weak winds in the subtropics near 30° latitude, trade winds from the east in the tropics, and weaker winds from the east along the Equator. The strength and direction of winds in the atmosphere is the result of uneven distribution of solar heating and continental land masses and the circulation of winds in a vertical plane in the atmosphere.



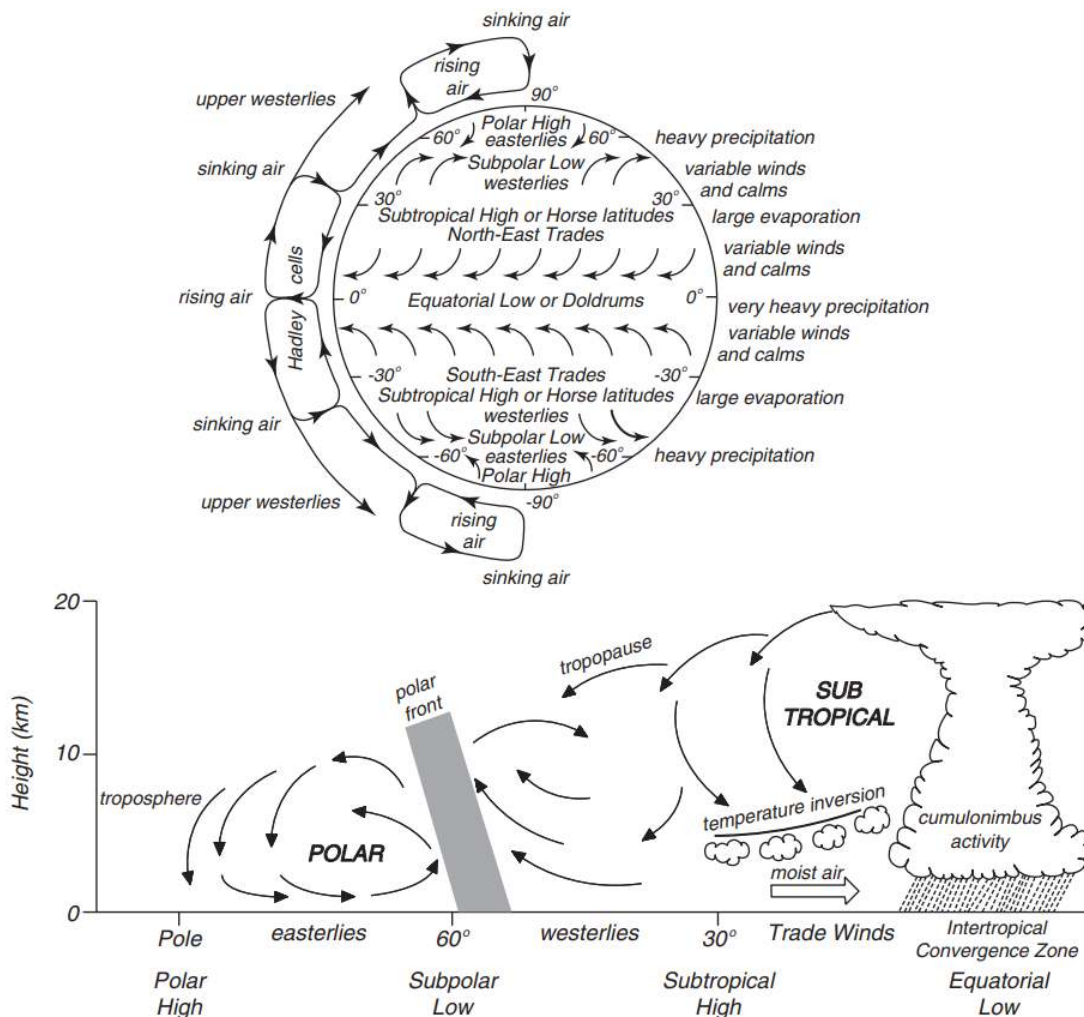


Figure 4.2.2: Sketch of earth's atmospheric circulation driven by solar heating in the tropics and cooling at high latitudes. Upper: The meridional cells in the atmosphere and the influence of earth's rotation on the winds. Bottom: Cross-section through the atmosphere showing the two major cells of meridional circulation. After The Open University (1989a: 14).

Maps of surface winds change somewhat with the seasons. The largest changes are in the Indian Ocean and the western Pacific Ocean (figure 4.2.3). Both regions are strongly influenced by the Asian monsoon. In winter, the cold air mass over Siberia creates a region of high pressure at the surface, and cold air blows southeastward across Japan and on across the hot Kuroshio, extracting heat from the ocean. In summer, the thermal low over Tibet draws warm, moist air from the Indian Ocean leading to the rainy season over India.

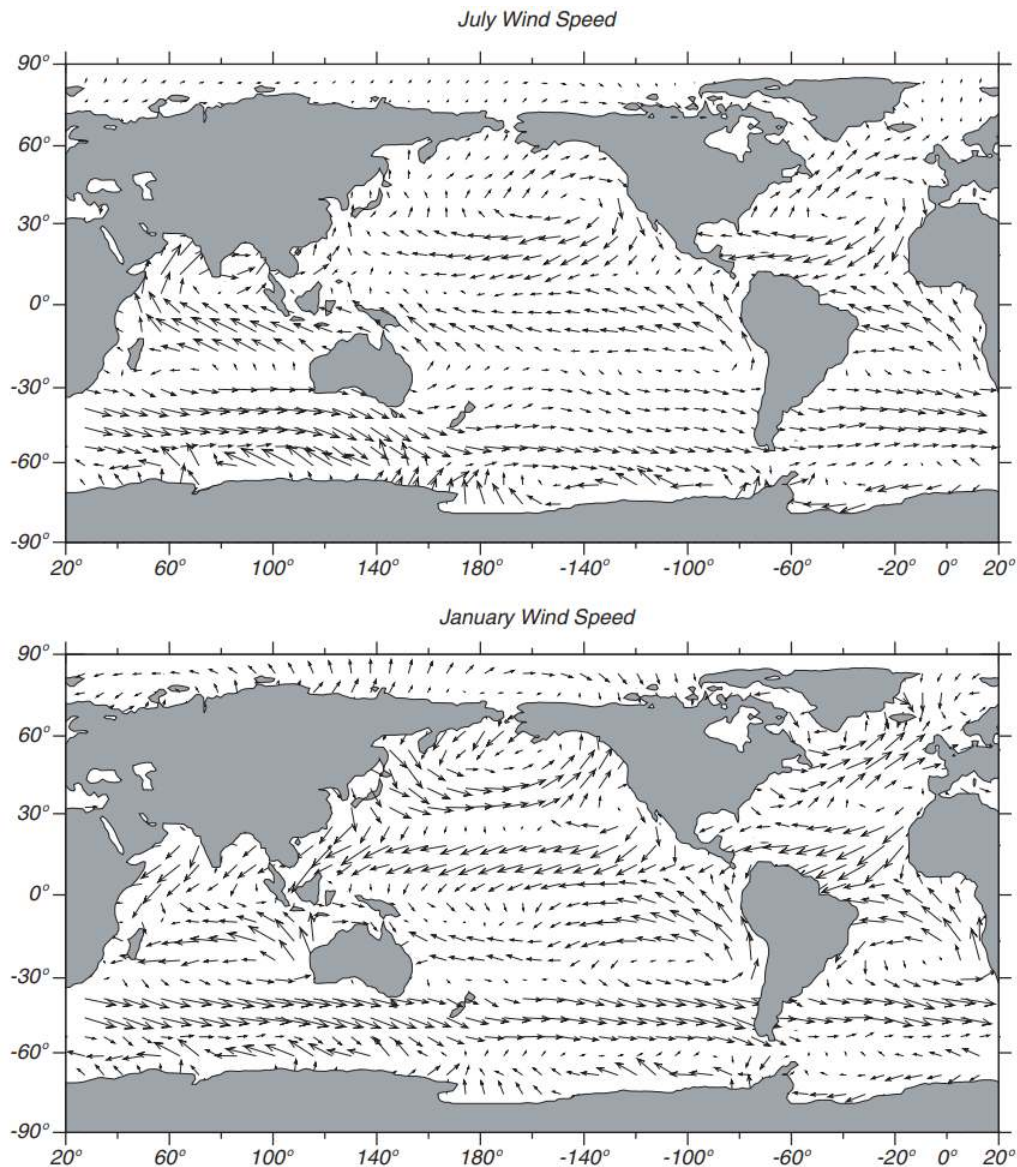


Figure 4.2.3: Mean, sea-surface winds for July and January calculated from the Trenberth et al. (1990) data set, which is based on the ECMWF reanalyses of weather data from 1980 to 1989. The winds near 140°W in the equatorial Pacific are about 8 m/s.

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4.3: The Planetary Boundary Layer

The atmosphere within 100 m of the sea surface is influenced by the turbulent drag of the wind on the sea and the fluxes of heat through the surface. This is the **atmospheric boundary layer**. Its thickness Z_i varies from a few tens of meters for weak winds blowing over water colder than the air to around a kilometer for stronger winds blowing over water warmer than the air.

The lowest part of the atmospheric boundary layer is the surface layer. Within this layer, which has thickness of $\approx 0.1Z_i$, vertical fluxes of heat and momentum are nearly constant.

Wind speed varies as the logarithm of height within the surface layer for neutral stability. See “[The Turbulent Boundary Layer Over a Flat Plate](#)” in Chapter 8. Hence, the height of a wind measurement is important. Usually, winds are reported as the value of wind at a height 10 m above the sea, U_{10} .

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4.4: Measurement of Wind

Wind at sea has been measured for centuries. Maury (1855) was the first to systematically collect and map wind reports. Recently, the US National Atmospheric and Oceanic Administration (NOAA) has collected, edited, and digitized millions of observations going back over a century. The resulting *International Comprehensive Ocean, Atmosphere Data Set* (ICOADS), discussed in [Section 5.5](#), is widely used for studying atmospheric forcing of the ocean.

Our knowledge of winds at the sea surface come from many sources. Here are the more important, listed in a crude order of relative importance:

Beaufort Scale

By far the most common source of wind data up to 1991 have been reports of speed based on the Beaufort scale. The scale is based on features, such as foam coverage and wave shape, seen by an observer on a ship (table 4.4.1).

The scale was originally proposed by Admiral Sir F. Beaufort in 1806 to give the force of the wind on a ship's sails. It was adopted by the British Admiralty in 1838 and it soon came into general use. The International Meteorological Committee adopted the force scale for international use in 1874. In 1926 they adopted a revised scale giving the wind speed at a height of 6 meters corresponding to the Beaufort Number. The scale was revised again in 1946 to extend the scale to higher wind speeds and to give the equivalent wind speed at a height of 10 meters. The 1946 scale was based on the equation $U_{10} = 0.836B^{3/2}$, where B = Beaufort Number and U_{10} is the wind speed in meters per second at a height of 10 meters (List, 1966). More recently, various groups have revised the Beaufort scale by comparing Beaufort force with ship measurements of winds. Kent and Taylor (1997) compared the various revisions of the scale with winds measured by ships having anemometers at known heights. Their recommended values are given in table 4.4.1.

Table 4.4.1. Beaufort Wind Scale and State of the Sea.

Beaufort Number	Descriptive term	m/s	Appearance of the Sea
0	Calm	0	Sea like a mirror.
1	Light air	1.2	Ripples with appearance of scales; no foam crests.
2	Light breeze	2.8	Small wavelets; crests of glassy appearance, not breaking.
3	Gentle breeze	4.9	Large wavelets; crests begin to break; scattered whitecaps.
4	Moderate breeze	7.7	Small waves, becoming longer; numerous whitecaps.
5	Fresh breeze	10.5	Moderate waves, taking longer to form; many whitecaps; some spray.
6	Strong breeze	13.1	Large waves forming; whitecaps everywhere; more spray.
7	Near gale	15.8	Sea heaps up; white foam from breaking waves begins to be blown into streaks.
8	Gale	18.8	Moderately high waves of greater length; edges of crests begin to break into spindrift; foam is blown in well-marked streaks.

Beaufort Number	Descriptive term	m/s	Appearance of the Sea
9	Strong gale	22.1	High waves; sea begins to roll; dense streaks of foam; spray may reduce visibility.
10	Storm	25.9	Very high waves with overhanging crests; sea takes white appearance as foam is blown in very dense streaks; rolling is heavy and visibility reduced.
11	Violent storm	30.2	Exceptionally high waves; sea covered with white foam patches; visibility still more reduced.
12	Hurricane	35.2	Air is filled with foam; sea completely white with driving spray; visibility greatly reduced.

Observers on ships everywhere in the world usually report weather observations, including Beaufort force, at the same four times every day. The times are at 0000Z, 0600Z, 1200Z and 1800Z, where Z indicates Greenwich Mean Time. The reports are coded and reported by radio to national meteorological agencies. The biggest error in the reports is the sampling error. Ships are unevenly distributed over the ocean. They tend to avoid high latitudes in winter and hurricanes in summer, and few ships cross the southern hemisphere (figure 4.4.1). Overall, the accuracy is around 10%.

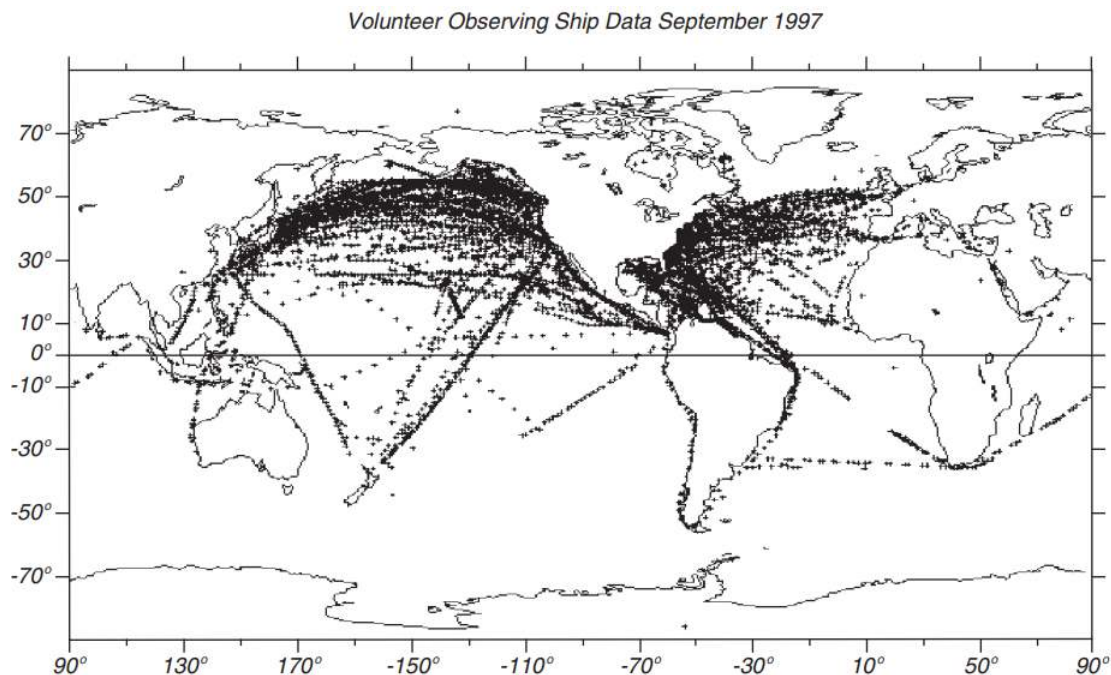


Figure 4.4.1: Location of surface observations made from volunteer observing ships and reported to national meteorological agencies. From NOAA, National Ocean Service.

Scatterometers

Observations of winds at sea now come mostly from scatterometers on satellites (Liu, 2002). The scatterometer is a instrument very much like a radar that measures the scatter of centimeter-wavelength radio waves from small, centimeter-wavelength waves on the

sea surface. The area of the sea covered by small waves, their amplitude, and their orientation depend on wind speed and direction. The scatterometer measures scatter from 2–4 directions, from which wind speed and direction are calculated.

The scatterometers on ERS-1 and -2 have made global measurements of winds from space since 1991. The NASA scatterometer on ADEOS measured winds for a six-month period beginning November 1996 and ending with the premature failure of the satellite. It was replaced by another scatterometer on QuikScat, launched on 19 June 1999. Quikscat views 93% of the ocean every 24 hr with a resolution of 25 km.

Freilich and Dunbar (1999) report that, overall, the NASA scatterometer on ADEOS measured wind speed with an accuracy of ± 1.3 m/s. The error in wind direction was $\pm 17^\circ$. Spatial resolution was 25 km. Data from QuikScat has an accuracy of ± 1 m/s.

Because scatterometers view a specific oceanic area only once a day, the data must be used with numerical weather models to obtain 6-hourly wind maps required for some studies.

Windsat

Windsat is an experimental, polarimetric, microwave radiometer developed by the US Navy that measures the amount and polarization of microwave radiation emitted from the sea at angles between 50° to 55° relative to the vertical and at five radio frequencies. It was launched on 6 January 2003 on the Coriolis satellite. The received radio signal is a function of wind speed, sea-surface temperature, water vapor in the atmosphere, rain rate, and the amount of water in cloud drops. By observing several frequencies simultaneously, data from the instrument are used for calculating the surface wind speed and direction, sea-surface temperature, total precipitable water, integrated cloud liquid water, and rain rate over the ocean regardless of time of day or cloudiness.

Winds are calculated over most of the ocean on a 25-km grid once a day. Winds measured by Windsat have an accuracy of ± 2 m/s in speed and $\pm 20^\circ$ in direction over the range of 5–25 m/s.

Special Sensor Microwave SSM/I

Another satellite instrument that is used to measure wind speed is the Special-Sensor Microwave/Imager (SSM/I) carried since 1987 on the satellites of the U.S. Defense Meteorological Satellite Program in orbits similar to the NOAA polar-orbiting meteorological satellites. The instrument measures the microwave radiation emitted from the sea at an angle near 60° from the vertical. The radio signal is a function of wind speed, water vapor in the atmosphere, and the amount of water in cloud drops. By observing several frequencies simultaneously, data from the instrument are used for calculating the surface wind speed, water vapor, cloud water, and rain rate.

Winds measured by SSM/I have an accuracy of ± 2 m/s in speed. When combined with ECMWF 1000 mb wind analyses, wind direction can be calculated with an accuracy of $\pm 22^\circ$ (Atlas, Hoffman, and Bloom, 1993). Global, gridded data are available since July 1987 on a 0.25° grid every 6 hours. But remember, the instrument views a specific oceanic area only once a day, and the gridded, 6-hourly maps have big gaps.

Anemometers on Ships

Satellite observations are supplemented by winds reported to meteorological agencies by observers reading anemometers on ships. The anemometer is read four times a day at the standard Greenwich times and reported via radio to meteorological agencies.

Again, the biggest error is the sampling error. Very few ships carry calibrated anemometers. Those that do tend to be commercial ships participating in the Volunteer Observing Ship program (figure 4.4.1). These ships are met in port by scientists who check the instruments and replace them if necessary, and who collect the data measured at sea. The accuracy of wind measurements from these ships is about ± 2 m/s.

Calibrated Anemometers on Weather Buoys

The most accurate measurements of winds at sea are made by calibrated anemometers on moored weather buoys. Unfortunately there are few such buoys, perhaps only a hundred scattered around the world. Some, such as Tropical Atmosphere Ocean (TAO) array in the tropical Pacific (figure 14.4.1) provide data from remote areas rarely visited by ships, but most tend to be located just offshore of coastal areas. NOAA operates buoys offshore of the United States and the TAO array in the Pacific. Data from the coastal buoys are averaged for eight minutes before the hour, and the observations are transmitted to shore via satellite links.

The best accuracy of anemometers on buoys operated by the us National Data Buoy Center is the greater of ± 1 m/s or 10% for wind speed and $\pm 10^\circ$ for wind direction (Beardsley et al. 1997).

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4.5: Calculations of Wind

Satellites, ships, and buoys measure winds at various locations and times of the day. If you wish to use the observations to calculate monthly averaged winds over the sea, then the observations can be averaged and gridded. If you wish to use wind data in numerical models of the ocean's currents, then the data will be less useful. You are faced with a very common problem: How to take all observations made in a six-hour period and determine the winds over the ocean on a fixed grid?

One source of gridded winds over the ocean is the **surface analysis** calculated by numerical weather models. The strategy used to produce the six-hourly gridded winds is called **sequential estimation techniques** or **data assimilation**. "Measurements are used to prepare initial conditions for the model, which is then integrated forward in time until further measurements are available. The model is thereupon re-initialized" (Bennett, 1992: 67). The initial condition is called the **analysis**.

Usually, all available measurements are used in the analysis, including observations from weather stations on land, pressure and temperature reported by ships and buoys, winds from scatterometers in space, and data from meteorological satellites. The model interpolates the measurements to produce an analysis consistent with previous and present observations. Daley (1991) describes the techniques in considerable detail.

Surface Analysis from Numerical Weather Models

Perhaps the most widely used weather model is that run by the European Centre for Mediumrange Weather Forecasts ecmwf. It calculates a surface analysis, including surface winds and heat fluxes (see Chapter 5) every six hours on a $1^\circ \times 1^\circ$ grid from an explicit boundary-layer model. Calculated values are archived on a 2.5° grid. Thus the wind maps from the numerical weather models lack the detail seen in maps from scatterometer data, which have a $1/4^\circ$ grid.

ECMWF calculations of winds have relatively good accuracy. Freilich and Dunbar (1999) estimated that the accuracy for wind speed at 10 meters is ± 1.5 m/s, and $\pm 18^\circ$ for direction.

Accuracy in the southern hemisphere is probably as good as in the northern hemisphere because continents do not disrupt the flow as much as in the northern hemisphere, and because scatterometers give accurate positions of storms and fronts over the ocean.

The NOAA National Centers for Environmental Prediction and the US Navy also produce global analyses and forecasts every six hours.

Reanalyzed Data from Numerical Weather Models

Surface analyses of weather over some regions have been produced for more than a hundred years, and over the whole Earth since about 1950. Surface analyses calculated by numerical models of the atmospheric circulation have been available for decades. Throughout this period, the methods for calculating surface analyses have constantly changed as meteorologists worked to make ever more accurate forecasts. Fluxes calculated from the analyses are therefore not consistent in time. The changes can be larger than the interannual variability of the fluxes (White, 1996). To minimize this problem, meteorological agencies have taken all archived weather data and reanalyzed them using the best numerical models to produce a uniform, internally-consistent, surface analysis.

The reanalyzed data are used to study oceanic and atmospheric processes in the past. Surface analyses issued every six hours from weather agencies are used only for problems that require up-to-date information. For example, if you are designing an offshore structure, you will probably use decades of reanalyzed data. If you are operating an offshore structure, you will watch the surface analysis and forecasts put out every six hours by meteorological agencies.

Sources of Reanalyzed Data

Reanalyzed surface flux data are available from national meteorological centers operating numerical weather prediction models.

1. The U.S. National Centers for Environmental Predictions, working with the National Center for Atmospheric Research have produced the NCEP/NCAR reanalysis based on 51 years of weather data from 1948 to 2005 using the 25 January 1995 version of their forecast model. The reanalysis period is being extended forward to include all dates up to the present with about a three-day delay in producing data sets. The reanalysis uses surface and ship observations plus sounder data from satellites. Reanalysis products are available every six hours on a T62 grid having 192×94 grid points with a spatial resolution of 209 km and with 28 vertical levels. Important subsets of the reanalysis, including surface fluxes, are available on CD-ROM (Kalnay et al. 1996; Kistler et al. 2000).

2. The European Centre for Medium-range Weather Forecasts, ECMWF, has reanalyzed 45 years of weather data from September 1957 to August 2002 (ERA-40) using their forecast model of 2001 (Uppala et al. 2005). The reanalysis uses mostly the same surface and ship data used by the NCEP/NCAR reanalysis plus data from the ERS-1 and ERS-2 satellites and SSM/I. The ERA-40 full-resolution products are available every six hours on a N80 grid having 160×320 grid points with a spatial resolution of 1.125° and with 60 vertical levels. The ERA-40 basic-resolution products are available every six hours with a spatial resolution of 2.5° and with 23 vertical levels. The reanalysis includes an ocean-wave model that calculates ocean wave heights and wave spectra every six hours on a 1.5° grid.

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4.6: Wind Stress

The wind, by itself, is usually not very interesting. Often we are much more interested in the force of the wind, or the work done by the wind. The horizontal force of the wind on the sea surface is called the **wind stress**. Put another way, it is the vertical transfer of horizontal momentum. Thus momentum is transferred from the atmosphere to the ocean by wind stress.

Wind stress T is calculated from:

$$T = \rho_a C_D U_{10}^2$$

where $\rho_a = 1.3 \text{ kg/m}^3$ is the density of air, U_{10} is wind speed at 10 meters, and C_D is the drag coefficient. C_D is measured using the techniques described in [Section 5.6](#). Fast response instruments measure wind fluctuations within 10–20 m of the sea surface, from which T is directly calculated. The correlation of T with U_{10}^2 gives C_D (figure 4.6.1).

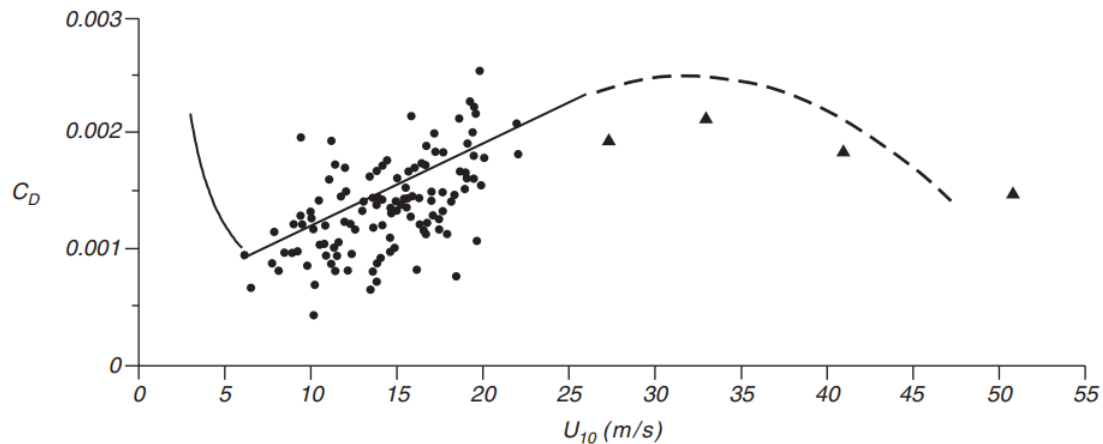


Figure 4.6.1: The drag coefficient as a function of wind speed U_{10} ten meters above the sea. Circles: Measured values from Smith (1980). Triangles: Measured values from Powell, Vickery, and Reinhold (2003). The solid line is from eq (4.6.1) proposed by Yelland and Taylor (1996). The dashed line is from Jarosz (2007).

Various measurements of C_D have been published based on careful measurements of turbulence in the marine boundary layer. Trenberth et al. (1989) and Harrison (1989) discuss the accuracy of an effective drag coefficient relating wind stress to wind velocity on a global scale. Perhaps the best of the recently published values are those of Yelland and Taylor (1996) and Yelland et al. (1998) who give:

$$1000C_D = 0.29 + \frac{3.1}{U_{10}} + \frac{7.7}{U_{10}^2} \quad (3 \leq U_{10} \leq 6 \text{ m/s}) \quad (4.6.1)$$

$$1000C_D = 0.60 + 0.071U_{10} \quad (6 \leq U_{10} \leq 26 \text{ m/s}) \quad (4.6.2)$$

for neutrally stable boundary layer. Other values are listed in their table 1 and in figure 4.6.1.

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4.7: Important Concepts

1. Sunlight is the primary energy source driving the atmosphere and ocean.
 2. There is a boundary layer at the bottom of the atmosphere where wind speed decreases with as the boundary is approached, and in which fluxes of heat and momentum are constant in the lower 10–20 meters.
 3. Wind is measured many different ways. The most common until 1995 was from observations made at sea of the Beaufort force of the wind.
 4. Since 1995, the most important source of wind measurements is from scatterometers on satellites. They produce daily global maps with 25 km resolution.
 5. The surface analysis from numerical models of the atmosphere is the most useful source of global, gridded maps of wind velocity for dates before 1995. It also is a useful source for 6-hourly maps. Resolution is 100-250 km.
 6. The flux of momentum from the atmosphere to the ocean, the wind stress, is calculated from wind speed using a drag coefficient.
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