

## Chapter 4

# Atmospheric Influences

The sun and the atmosphere drive directly or indirectly almost all dynamical processes in the ocean. Geothermal heating of the oceans from below barely influences the deep layers 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. Deep mixing drives to some extent the deeper currents in the ocean.

The oceans, in turn, help drive the atmospheric circulation. The uneven distribution of heat loss and gain by the ocean leads to winds in the atmosphere. Sunlight warms the tropical oceans, which evaporate, transferring heat in the form of moisture to the atmosphere. Winds and ocean currents carry heat poleward, where it is lost to space. In some regions, cold dry air blows over warm water further extracting heat from the ocean.

The response of the ocean to the atmosphere is not passive because oceanic processes help drive the atmospheric circulation. To understand ocean dynamics, we must consider the ocean and the atmosphere as a coupled dynamic system. In this chapter we will look at the exchange of heat and water between the atmosphere and the ocean. Later, we will explore the influence of the wind on the ocean and the exchange of momentum leading to wind-driven ocean currents.

### 4.1 The Earth in Space

The Earth's orbit about the sun is nearly circular at a mean distance of  $1.5 \times 10^8$  km. The eccentricity of the orbit is small, 0.0168. Thus Earth is 103.4% further from the Sun at aphelion than at perihelion, the time of closest approach to the sun. Perihelion occurred on 3 January in 1995, and it slowly changes by about 20 minutes per year. Earth's axis of rotation is inclined  $23.45^\circ$  to the plane of earth's orbit around the sun (Figure 4.1). 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.

The latitudes of  $23.45^\circ$  North and South are the Tropics of Cancer and

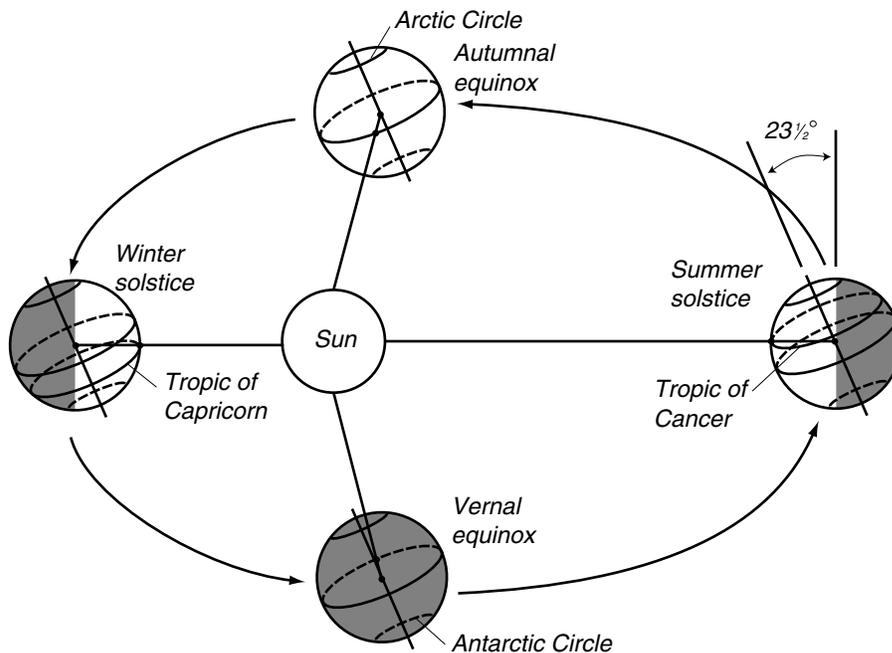


Figure 4.1 The Earth in space. The ellipticity of Earth's orbit around the sun and the tilt of Earth's axis of rotation relative to the plane of Earth orbit leads to an unequal distribution of heating and to the seasons.

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 in the northern hemisphere occurs in the summer, around 21 June. Maximum insolation in the southern hemisphere occurs in December.

If the insolation were rapidly and efficiently 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. The two processes are  $180^\circ$  out of phase in the northern hemisphere; but because the Earth's climate system is non-linear, it can phase lock to either frequency. Hence, which will dominate? Recent work by Thomson (1995) shows that either process can dominate in some regions for some times.

## 4.2 Atmospheric Wind Systems

Figure 4.2 shows the distribution of sea-level winds and pressure averaged over the year 1989. The map shows strong winds from the west between  $40^\circ$  to  $60^\circ$  latitude, the roaring forties, weak winds in the subtropics near  $30^\circ$  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.

## Annual Wind Speed and Sea Level Pressure (hPa) For 1989

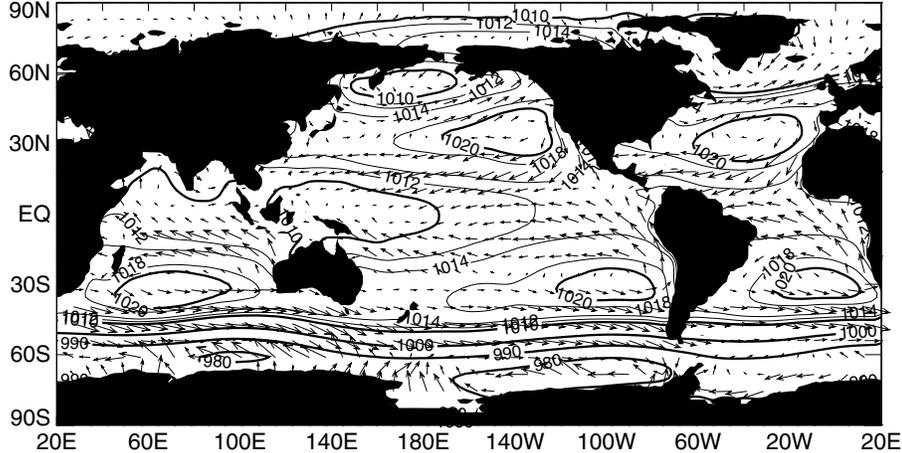


Figure 4.2 Map of mean annual wind velocity calculated from Trenberth (1990) and sea-level pressure for 1989 from the NASA Goddard Space Flight Center's Data Assimilation Office (Schubert et al. 1995).

The mean value of winds over the ocean is (Wentz et al. 1984):

$$U_{10} = 7.4 \text{ m/s} \quad (4.1)$$

Each component of the wind vector has a Gaussian distribution with zero mean, so the magnitude of the wind vector has a Rayleigh distribution (Freilich, 1997).

A simple cartoon (Figure 4.3) shows distribution of winds in the atmosphere, including equatorial convection, trade winds in the tropics, and westerly winds at higher latitudes. The distribution of surface winds strongly influences the properties of the upper ocean.

The simple picture of the winds changes somewhat with the seasons. The largest changes are in the Indian Ocean and the western Pacific Ocean (Figure 4.4). 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.

### 4.3 The Planetary Boundary Layer

The atmosphere immediately above the ocean is influenced by the turbulent drag of the wind on the sea surface and the fluxes of heat through the surface. This layer of the atmosphere that is closely coupled to the surface is the *atmospheric boundary layer*. The thickness of the layer  $Z_i$  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 structure of the layer influences the exchange of momentum and heat between the surface and the atmosphere. (See Dabberdt et al. 1993 for a review of the subject.)

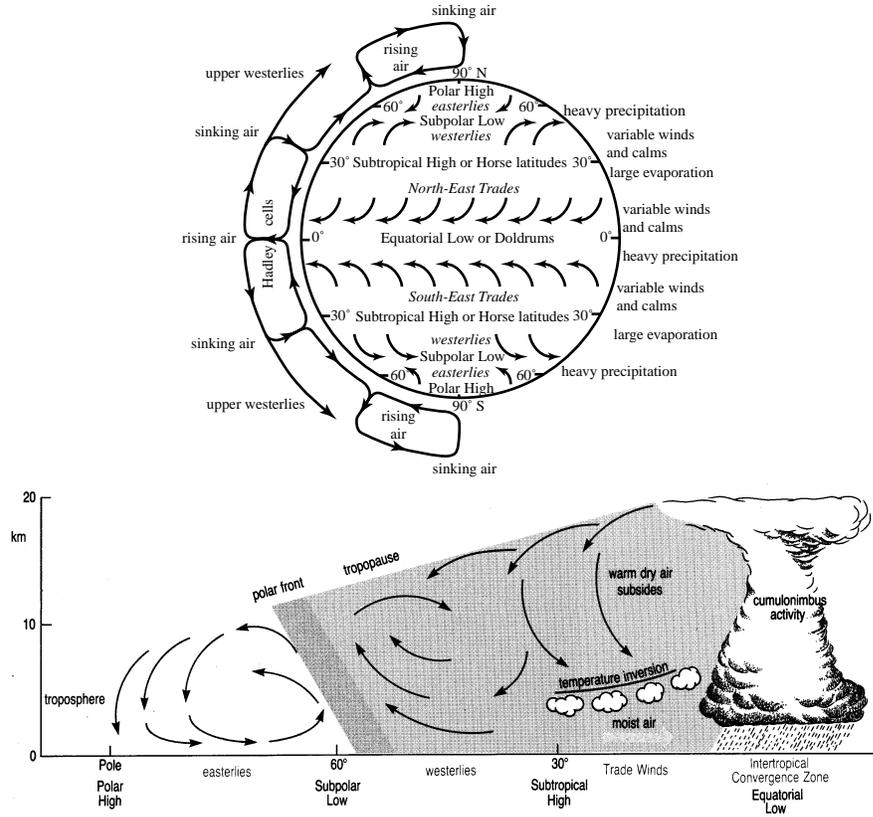


Figure 4.3 Simplified schematic of Earth's atmospheric circulation driven by solar heating in the tropics and cooling at high latitudes. **Upper:** The meridional cells in the atmosphere and the influence of Earth's rotation on the winds. **Bottom:** Cross-section through the atmosphere showing the two major cells of meridional circulation. From The Open University (1989a).

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

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, given by  $U_{10}$ .

#### 4.4 Measurement of Wind

Wind at sea has been measured for centuries. Maury (1847) 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 *Combined*

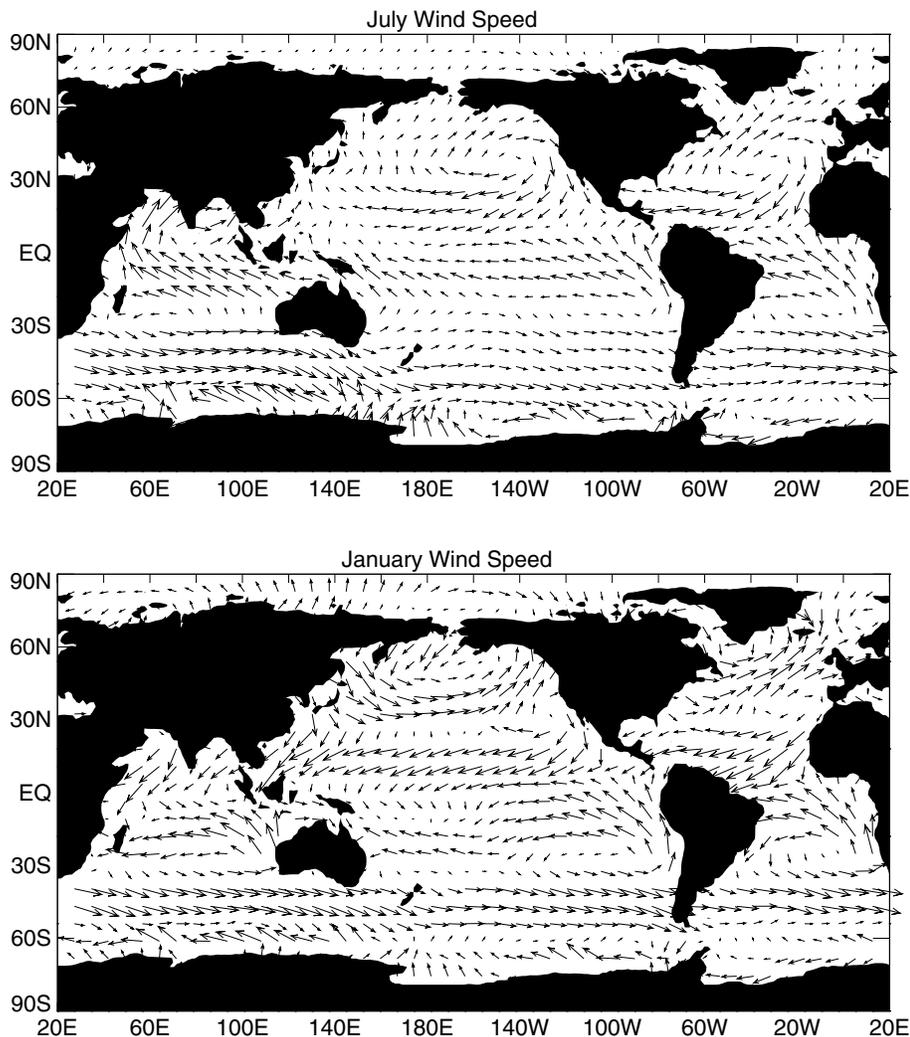


Figure 4.4 Mean, sea-surface winds for July and January calculated by Trenberth's (1990) data set calculated from the ECMWF reanalyses of weather data from 1980 to 1989.

*Ocean, Atmosphere Data Set* COADS is widely used for studying atmospheric forcing of the ocean.

Our knowledge of winds at the sea surface come from many types of instruments or observations. Here are the more important sources, listed in a crude order of importance to the historical record:

**Beaufort Scale** By far the most common source of wind data have been reports of speed based on the Beaufort scale. Even in 1990, 60% of winds reported from the North Atlantic used the Beaufort scale. The scale is based on features, such as foam coverage and wave shape, seen by an observer on a ship, that are influenced by wind speed (Table 4.1).

**Table 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.
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.

From Kent and Taylor (1997)

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 empirical 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 4.1.

Observers on ships usually report weather observations, including Beaufort force, four times per day, at midnight (0000Z), 0600Z, noon (1200Z), and 1800Z Greenwich Mean Time. The reports are coded and reported by radio to national meteorological agencies. The reports have important errors:

1. Ships are unevenly distributed over the ocean. Ships tend to avoid high

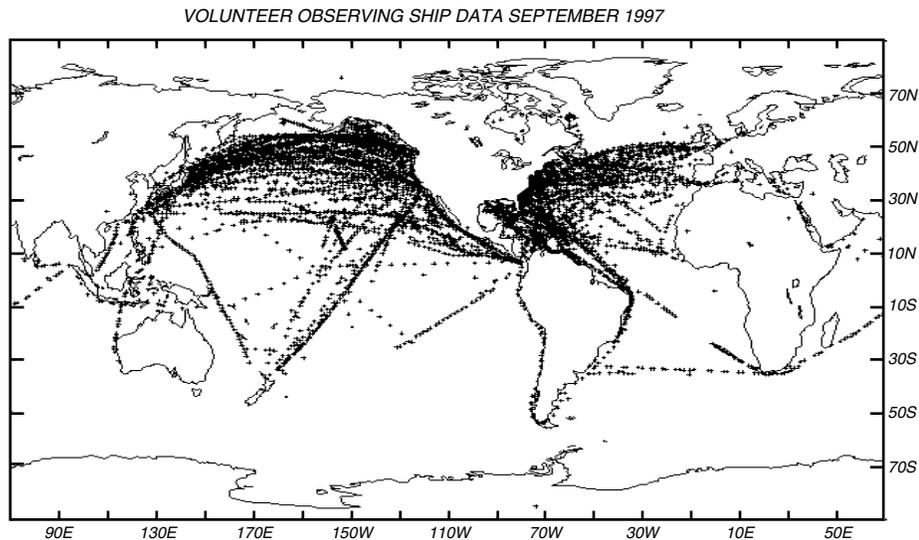


Figure 4.5 Location of surface observations made from volunteer observing ships and reported to national meteorological agencies. (From NOAA, National Ocean Service)

latitudes in winter and hurricanes in summer, and few ships cross the southern hemisphere (Figure 4.5).

2. Observers may fail to take care in observing oceanic conditions on which the Beaufort scale is based.
3. The coding of the data may have errors, which can result in the reports having the wrong location. See Figure 4.5 which shows ship positions in the Sahara Desert.
4. Overall, the accuracy is probably no better than around 10%.

**Scatterometers** Observations of winds at sea are coming more and more from instruments on satellites, and scatterometers are the most common source of the observations. The scatterometer is an 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, and their amplitude depends on wind speed and direction. The scatterometer measures scatter from 2–3 directions, from which wind speed and direction are calculated. Because the instrument cannot distinguish winds blowing from right to left relative to the radio beam from winds blowing from left to right, the observation of direction is ambiguous. The ambiguity can be removed with a few surface observations or by using the data with numerical weather models. For example, wind must blow counterclockwise around lows in the northern hemisphere and clockwise around lows in the southern hemisphere.

The scatterometers on ERS-1 and ERS-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.

Freilich and Dunbar (1999) report that, overall, the NASA scatterometer on ADEOS measured wind speed with an accuracy of  $\pm 1.3$  m/s. For wind speed exceeding 6 m/s, fewer than 3% of the wind values had a significant ambiguity error. For those winds with no ambiguity error, the error in wind direction was  $\pm 17^\circ$ . Spatial resolution was 25 km. The errors in calculated velocity are due to lack of knowledge of scatter vs wind speed, the unknown influence of surface films, and sampling error (Figure 4.6).

**Special Sensor Microwave SSM/I** Another satellite instrument that is widely used for measuring 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  $60^\circ$  from the vertical. The emission 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. As with the scatterometer, the wind direction is ambiguous, and the ambiguity is removed using surface observations or by using the data with numerical weather models.

Winds measured by the instrument have an accuracy of  $\pm 2$  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  $2.5^\circ$  longitude by  $2.0^\circ$  latitude grid every 6 hours (Atlas et al, 1996).

**Anemometers on Ships** The next most common source of winds reported to meteorological agencies come from observers reading the output of an anemometers on ships. The output of the anemometer is read four times a day at the standard Greenwich times and reported via radio to meteorological agencies. These reports also have important errors:

1. The reports are sparse in time and space. Very few ships report anemometer winds.
2. Anemometer may never be calibrated after installation.
3. The observer usually observes the output of the anemometer for a few seconds, and thus the observation is an instantaneous value of wind speed and direction rather than an average over several minutes to an hour. Remember that winds can be gusty, and the observation can have errors of 10–30%.
4. The observations are reported by coded radio messages, and the message can have coding errors. Such errors cause ship winds to be reported from over land as shown in Figure 4.4.

**Calibrated Anemometers on Ships** Few ships carry calibrated anemometers. Those that do tend to be commercial ships participating in the Volunteer Observing Ship program. 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. Errors are due to airflow about the ship and incorrect correction for ship motion. The best accuracy is about  $\pm 2$  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 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 8 minutes before the hour, and the observations are transmitted to shore via satellite links.

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

**Surface Analysis from Numerical General Circulation Models** 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 one-day period and determine the winds over the ocean on say a fixed grid each day?

The best source of gridded winds over the ocean is the output from numerical models of the atmospheric circulation. The strategy used to produce the 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). Usually, all available measurements are used, including surface observations made from meteorological stations on land, ship and buoy reports of pressure and temperature, and meteorological satellite data. The model interpolates the measurements to produce the initial conditions consistent with previous and present observations.

1. The surface fluxes from the European Centre for Medium-range Weather Forecasts ECMWF are perhaps the most widely used fluxes for surface forcing of the ocean. Surface winds and fluxes are calculated every six hours on a  $1^\circ \times 1^\circ$  grid from an explicit boundary-layer model. The fluxes include not only wind stress but also heat fluxes discussed in the next chapter. Calculated values are archived on a  $2.5^\circ$  grid.
2. Accuracy of northern-hemisphere winds calculated by the ECMWF is relatively good. Freilich and Dunbar (1999) estimated that the accuracy for wind speed at 10 meters is  $\pm 1.5$  m/s after removing 0.9% of the values (which were more than three standard deviations from the value reported

by anemometers on buoys), and  $\pm 18^\circ$  for direction. The speed was only 90% of the wind speed observed by buoys in the analysis area.

3. Accuracy in the southern hemisphere is not as good as in the northern hemisphere, but accuracy is improving. The use of scatterometer winds from ADEOS, ERS-1 and 2 have made significant improvements.

Daley (1991) describes the strategies and techniques in considerable detail.

Other surface-analysis data sets of special use in oceanography include: 1) the Planetary Boundary-Layer Data set from the U.S. Navy's Fleet Numerical Oceanography Center FNOC; and 2) surface wind maps for the tropics produced at Florida State University (Goldenberg and O'Brien, 1981).

**Reanalyzed Output from Numerical General Circulation Models** The output from numerical models of the atmospheric circulation has been available for decades. Throughout this period, the models have been constantly changed as meteorologists strive to obtain ever more accurate forecasts. The calculated fluxes 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 measurements for long periods and reanalyzed them using the best numerical models now available to produce a uniform, internally-consistent, surface analysis.

The reanalyzed data sets are now being used to study oceanic and atmospheric dynamics. The surface analysis data sets 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.

Because reanalysed data sets have been made available only recently, their accuracy is not yet firmly established. We do know however that the data sets are more accurate in the northern hemisphere because more surface observations are available from the northern hemisphere. To obtain estimates of their accuracy, various teams are intercomparing output from various reanalyses; and their conclusions will soon be available.

**Sources of Reanalyzed Data** Analysed 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, the NCEP/NCAR reanalysis, have reanalysed 40 years of weather data from 1958 to 1998 using the 25 January 1995 version of their forecast model. The reanalysis period will be extended backward to include the 1948–1957 period; and it is being extended forward to include all date up to the present with about a six month delay in producing data sets. The reanalysis uses surface and ship observations plus sounder data from satellites. Reanalysed products are available every six hours on a T62 grid having  $192 \times 94$  grid points with a spatial resolution of 209 km and with 24 vertical levels. Important subsets

of the reanalysed data, including surface fluxes, are available on CD-ROM (Kalnay et al. 1996; Kistler et al. 1999).

2. The European Centre for Medium-range Weather Forecasts ECMWF have reanalysed 17 years of weather data from 1979 to 1993. The reanalysis uses mostly the same surface, ship and satellite data used by the NCEP/NCAR reanalysis. The European Centre is extending the reanalysis to cover a 40-year period from 1957–1997. Spatial resolution will be 83 km; temporal resolution will be 6 hours. The reanalysis will use most available satellite data sets, including data from the ERS-1 and ERS-2 satellites and SSM/I. The analysis will include an ocean-wave model and it will calculate ocean wave heights.
3. The Data Assimilation Office at NASA's Goddard Space Flight Center has completed a reanalysis for the period 1 March 1980 to 13 December 1993 later extended to February 1995. The analysed data are available every six hours on a  $2^\circ \times 2.5^\circ$  ( $91 \times 144$  point) grid with 20 vertical levels. The analysis uses the NCEP real-time in-situ observations plus TOVS data from the NOAA meteorological satellites and cloud-drift winds (Schubert, Rood, and Pfaendtner, 1993). The analysis places special emphasis on the assimilation of satellite data using the Goddard Earth Observing System general circulation model.

#### 4.5 The Sampling Problem in Scatterometry

Monthly maps of surface winds made from satellite scatterometer observations of the ocean frequently show bands parallel to the satellite track. Zeng and Levy (1995) used data from the scatterometer on the ERS-1 satellite and found that the bands are due to sampling errors. The satellite observed areas on the sea surface 8-12 times per month (Figure 4.6), and the distribution of samples was not uniform. Sometimes the satellite missed storms winds. An example is shown in the figure. A weak storm passed through regions A & B in the Pacific between 11 and 18 September 1992 as shown in panel (a) of the figure. The satellite observed storm winds in A on 15 and 18 September, but it did not observe storm winds in region B. As a result, the monthly mean value of wind speed calculated from the satellite data at B differed by 6 m/s from the mean value at A. Further analysis of ERS-1 scatterometer data showed that monthly mean value of wind speed calculated from the data have a sampling error of 1-2 m/s in mid-latitudes.

#### 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:

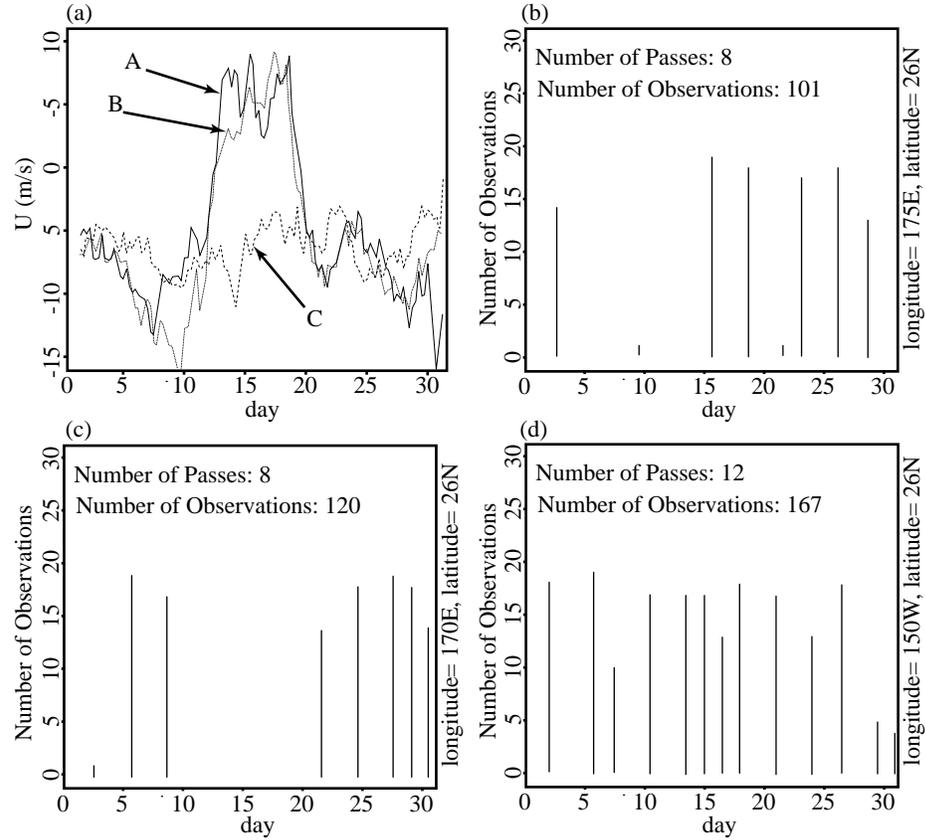


Figure 4.6 (a) Time series of ECMWF zonal winds (in meters per second) at nearby regions A and B during September 1992. Point C is in another region of the Pacific. (b) Number of ERS-1 observations in the  $1^\circ$  by  $1^\circ$  box centered at A versus time. (c) Same as (b) except at point B. (d) Same as (b) except at point C. A and B are in and just outside a strong band in the map of wind speed; C is in a region with no banding. (From Zeng & Levy, 1995).

$$T = \rho C_D U_{10}^2 \quad (4.2)$$

where  $\rho$  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 §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 4.7). The coefficient can also be directly calculated from measurements of fluctuations of the horizontal velocity using a technique, called the *dissipation method*. The technique is complicated, and its description is beyond the scope of this book.

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

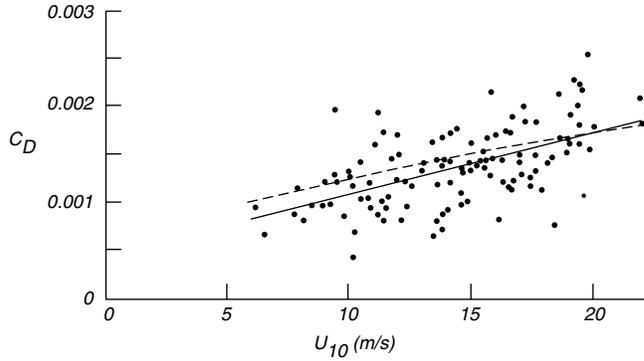


Figure 4.7 Observations of the drag coefficient as a function of wind speed  $U_{10}$  ten meters above the sea. The solid line is  $1000 C_D = 0.44 + 0.063U_{10}$  proposed by Smith (1980) and the dashed line follows from Charnock (1955). From Smith, 1980.

stress to wind velocity on a global scale. Perhaps the most recently published value is that of Yelland and Taylor (1996), who give:

$$1000 C_D = 0.29 + \frac{3.1}{U_{10}} + \frac{7.7}{U_{10}^2} \quad (3 \leq U_{10} \leq 6 \text{ m/s}) \quad (4.3a)$$

$$1000 C_D = 0.60 + 0.070 U_{10} \quad (6 \leq U_{10} \leq 26 \text{ m/s}) \quad (4.3b)$$

for neutrally stable boundary layer. Other values are listed in their table 1 and in figure 4.7.

Useful monthly maps of oceanic wind stress suitable for studies of the ocean circulation have been published by Trenberth et al. (1989) who used analysed wind fields from ECMWF to produce maps with  $2.5^\circ$  resolution, and by Hellerman and Rosenstein (1983) who used 35 million observations made from surface ships between 1870 and 1976 to produce maps with  $2^\circ$  resolution.

#### 4.7 Important Concepts

1. Sunlight is the primary energy source driving the atmosphere and oceans.
2. There is a boundary layer at the bottom of the atmosphere where wind speed decreases with height, 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 are from observations made at sea of the Beaufort force of the wind. Wind is measured from space using scatterometers and microwave radiometers. The output from atmospheric circulation models is perhaps the most useful source of global wind velocity.
4. The flux of momentum from the atmosphere to the ocean, the wind stress, is calculated from wind speed using a drag coefficient.