Chapter 1

Introduction: What drives the ocean currents?

Sixty years ago, this textbook would have been titled "Introductory Geography of the Oceans". Physical oceanography then was a close relative of physical geography and shared its descriptive character. This period culminated in textbooks such as *Geographie des Atlantischen Ozeans* (1912) and *Geographie des Indischen und Stillen Ozeans* (1935) by G. Schott, books which conveyed to the reader through a passioned yet accurate description of its features the fascination which the oceanic environment exerts on the oceanographer. Oceanography has come a long way since then, having concentrated on understanding the physical principles that drive the oceans and using the tools of mathematics and theoretical fluid dynamics to forecast their behaviour. Students of oceanography now spend more time trying to come to grips with vorticity, inverse methods and normal mode analysis than learning about the features of the deep sea basins and marginal seas or about the climatic regions of the oceans. And this is rightly so, for little is learned in science through mere description; analysis and conclusion are required before anyone can claim to understand.

As it turns out, understanding the ocean circulation is impossible without knowledge of geographical details - the depth of certain ocean sills, for example, or the peculiarities of the wind field and its seasonal variation. To separate the facts about the geography of the ocean from acquired knowledge about its dynamics would be like separating the memorizing of the vocabulary of a new language from the learning of its grammar. What we propose to do is describe the features of the world ocean both as a systematic exercise in geography and as examples of physical principles at work.

These physical principles are sufficiently powerful and all-pervasive that it is worth introducing them first. This and the following four chapters are a summary of some of the principles and their consequences, and will serve as a reference for all chapters to come. Students of oceanography who are using this book as their introduction to the discipline will find them essential reading. Advanced students who use the book because of a need to brush up their knowledge on the geography of the oceans can skip the first five chapters and go straight to the chapter of interest to them. Both should take particular notice of the figures which accompany the text. With a bit of guidance, much can be learnt by looking closely at observational data. Our text will provide the guidance, but it will not go into detailed descriptions of what can be seen more easily in figures. The figures are therefore not illustrations of the text; they are an integral part of this book.

Some knowledge of the geography of the oceans is essential in regional oceanography. While we attempted to include as much relevant geographical information as possible, clarity of figures must rank higher than detail in an introductory text. Sometimes the location of a feature can be determined by consulting the index. In other cases the use of an elementary geographical atlas may be required.

One final note on the use of geographic and oceanographic nomenclature, before some readers turn the page and proceed to other chapters. Although the use of geographical names and the rules on naming newly discovered geographic features are regulated by an international advisory body, use of geographical names in oceanography is not uniform. This is particularly true for features such as currents, fronts, or water masses, which are not covered by the international regulations on the use of geographical names. Oceanographers have an unfortunate habit of trying to make their mark by putting names to their liking on features which may or may not have been named before (probably to someone else's liking). In this text we adopt the principle that geographic features are referred to under the names used on the GEBCO charts (IHO/IOC/CHS, 1984). In references to currents and fronts we use generally accepted names where they exist, preferring names which include a reference to geographic features (e.g. Peru/Chile Current, not Humboldt Current). Universally accepted names for water masses exist only for the major oceanic water masses; other water masses can be found under a variety of names in the literature. Our usage of water mass names is based partly on historical use, partly on the systematic approach to water mass analysis described in Chapter 5. Wherever possible we use names already introduced by others and do not invent our own.

We return to our discussion of the most important physical principles. A discussion of what in essence are elements of geophysical fluid dynamics is not to everyone's liking; nevertheless, some of the principles determining ocean flows turn out to be quite simple, and by understanding them it is possible to go a long way towards understanding the role of the ocean in climate variations, both natural or man-made. The ocean is unique in this respect; it can absorb heat in one region, and restore it to the atmosphere (perhaps decades or centuries later) at a quite different place. This has become a topic of widespread interest and intensive research in recent years, and by spending some effort on understanding the underlying principles readers will find that they can gain an understanding of much of the modern literature on this topic.

If we exclude tidal forces, which have little effect on the long-term mean properties of the ocean, the oceanic circulation is driven by three external influences: wind stress, heating and cooling, and evaporation and precipitation - all of which, in turn, are ultimately driven by radiation from the sun. To understand why temperature, salinity and all other properties of the oceans' waters are distributed the way they are, a basic knowledge about these external forces is necessary. We therefore begin our description of the geography of the oceans with a brief look at the atmosphere, which holds the key to the question how the energy received from the sun keeps the ocean circulation going.

We note at the outset that this approach ignores the fact that the circulation of the atmosphere is in turn influenced by the distribution of oceanic properties, such as sea surface temperature (in oceanography often abbreviated as SST) and the distribution of sea ice. In particular, the amount of evaporation from the ocean depends strongly on the sea surface temperature; and when the evaporated water is returned as rain it releases its latent heat into the surrounding air. This heating is probably the strongest driving force for the atmospheric winds. To understand the oceanic and atmospheric circulation fully we should treat them as a single system of two interacting components, coupled at the air-sea interface through the fluxes of momentum, heat, and mass. This of course complicates the task and could not be achieved with traditional oceanographic or meteorological tools. Although the stage has now been reached where treatment of the ocean and the atmosphere as a coupled system is becoming more and more feasible, it seems good advice for an introductory text to follow the traditional approach and consider the state of the atmosphere as determined independently of the state of the ocean. We shall return to the question of interaction between ocean and atmosphere in the last three chapters when we discuss interannual climate fluctuations and long-term climate change.

The amount of heat radiation received by the outer atmosphere varies from the equator to the poles. The difference varies with the seasons, but on average the equatorial regions receive much more heat than the polar regions. The cold air at the poles is denser than the warm air at the equator; and since the air pressure at the sea surface or on land is determined by the weight of the air above the observation point, air pressure at sea level is higher at the poles than at the equator - in other words, a pressure gradient is set up which is directed from the poles toward the equator. The pressure gradient in the upper part of the atmosphere has the opposite sign.

In fluids and gases, pressure gradients produce flow from regions of high pressure to regions of low pressure. If the earth were not rotating, the response to these pressure gradients would be direct and simple. Two circulation cells would be set up, one in either hemisphere, by the differential solar heating. At sea level, winds would blow from the poles to the equator; the air would then rise and recirculate back to the poles at great height. On a rotating earth this pattern is modified quite strongly, in two ways. Firstly, as air moves towards the equator, the rotation of the earth shifts ocean and land eastward under it. An observer moving with the land experiences the air movement as an "easterly" wind, i.e. a wind blowing from the east, with an equatorward component. In the tropics and subtropics this wind is known as the Trade Wind, in polar latitudes it occurs as the Polar Easterlies. The outcome is that the wind no longer blows from regions of high pressure to regions of low pressure along the most direct route but tends to follow contours of constant pressure (isobars) - hence the usefulness of isobars on the daily weather map in the television news.

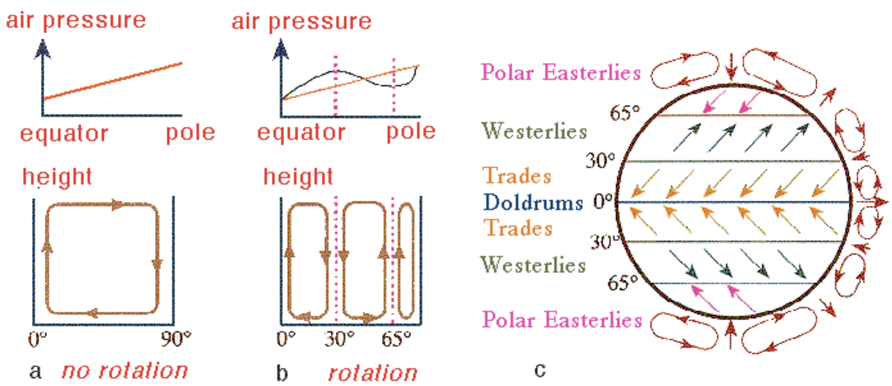
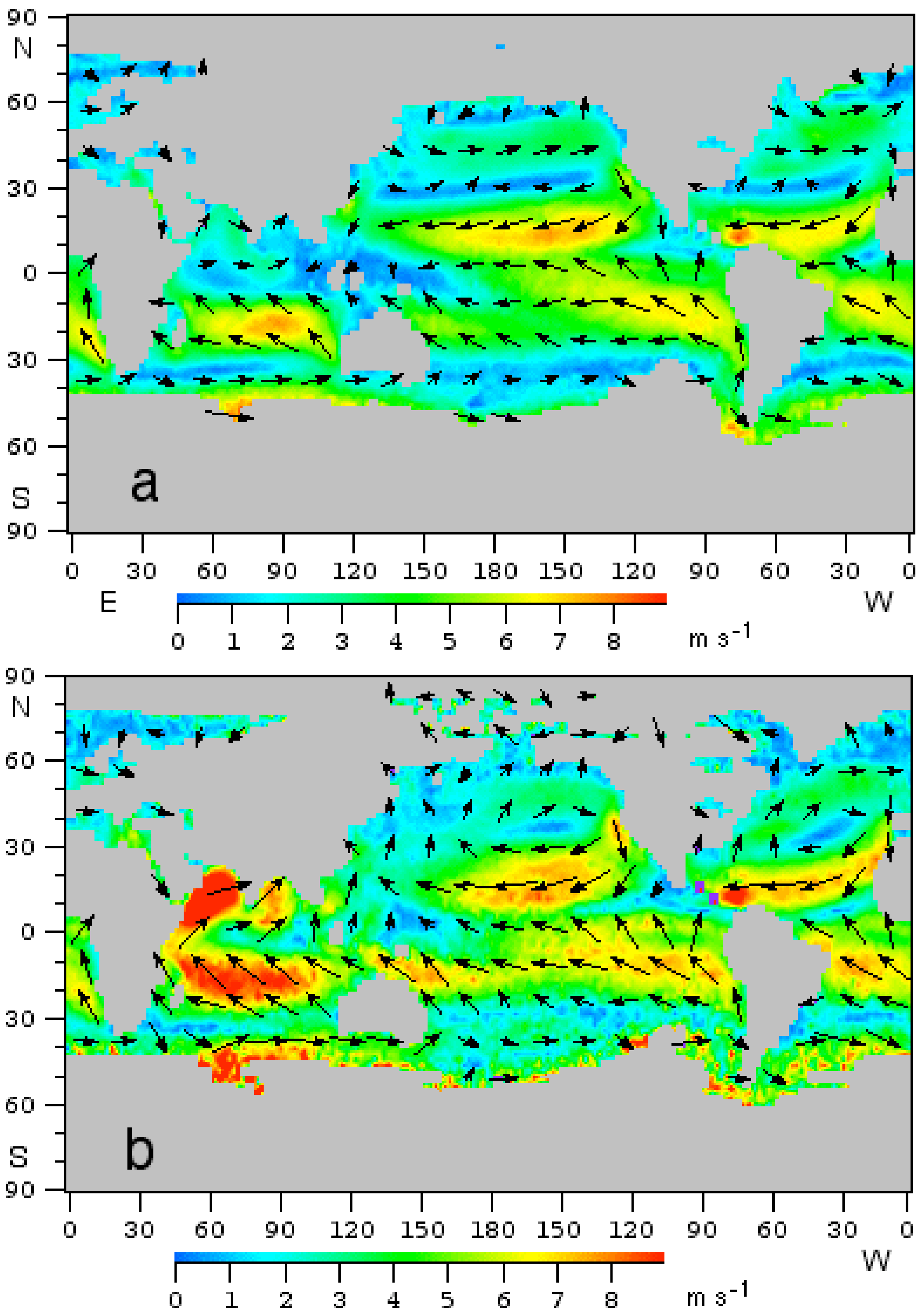


Fig. 1.1. Schematic diagram of the meridional air pressure distribution and associated air movement (a) on a non-rotating earth, (b) on a rotating earth without continents, (c) viewed from above.

Because on a rotating earth the flow of air is more zonal (directed east-west) than meridional (directed north-south), the importance of the vertical component of air movement is reduced: the flow can circle the earth with great speed without need for uplift or sinking. This produces the second, more drastic modification of the simple hemispheric cell arrangement. It turns out that zonal flow of high speed becomes unstable, creating eddies which in turn reshape the air pressure distribution. As a result, an intermediate air pressure maximum is established at mid-latitudes. The reversal of the meridional pressure gradient establishes a band of "westerly" wind, i.e. wind blowing from the west (Fig 1.1).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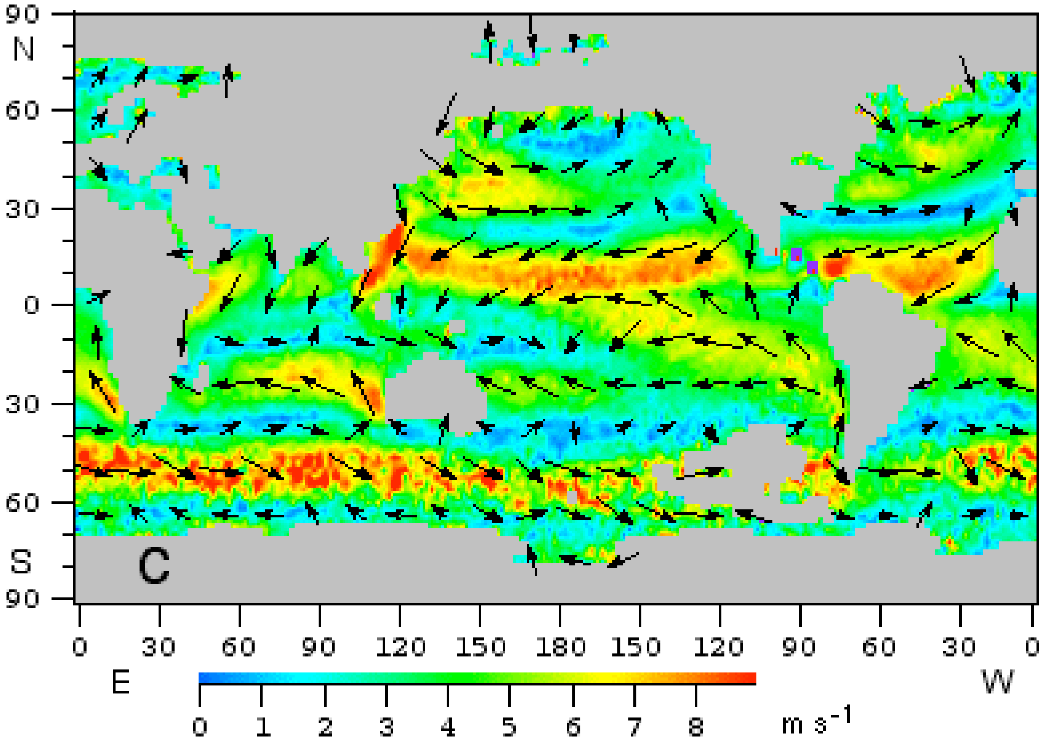
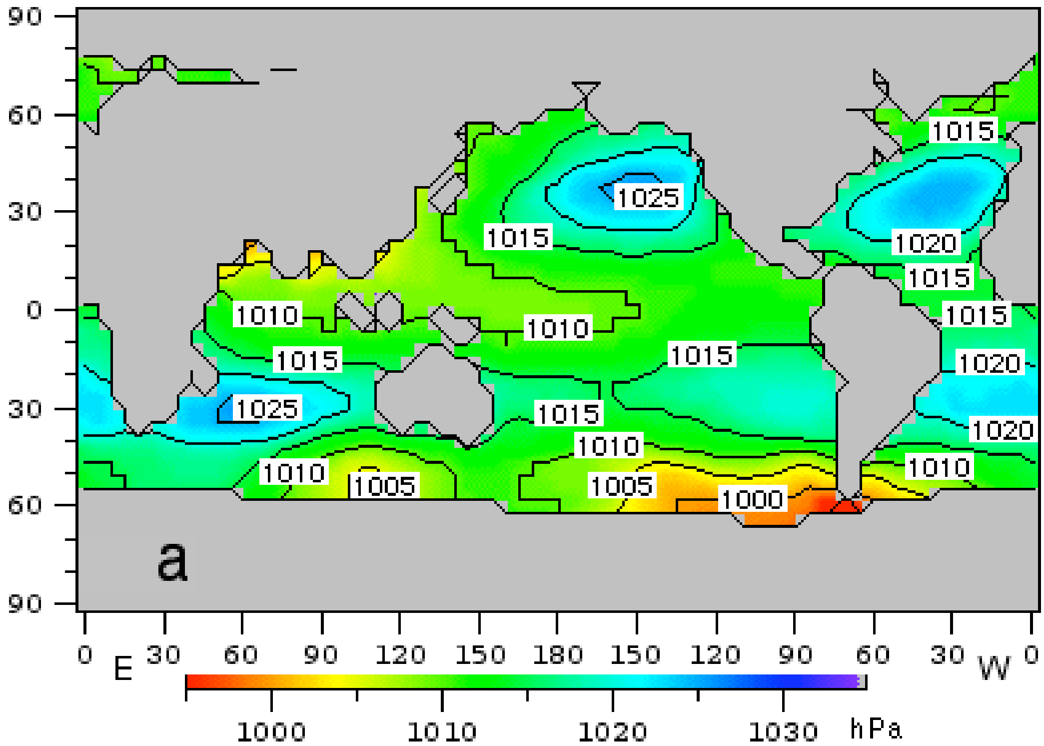


Fig. 1.2. Surface winds over the World Ocean. (a, page 4) Annual mean, (b, page 4) July mean, (c) January mean. Data from http://ferret.wrc.noaa.gov/las/, the NOAA Live Climate Data Server using the Comprehensive Atmosphere/Ocean Data Set (COADS) climatology. COADS is based on ship observations; regions without data are gray. Colour indicates wind speed.

Seafarers know them well and refer to them as the Roaring Forties, thus expressing their experience that between 40° and 50° latitude the winds are usually strong, highly variable and very gusty.

Figure 1.2 gives the winds at sea level in the real world. The features seen in Figure 1.1 come out clearly, but the presence of continental land masses modifies the atmospheric circulation further. Because air heats up faster over the continents than over the oceans during summer, and cools faster during winter, large land masses are characterized by low air pressure in summer - relative to air pressure over the ocean at the same latitude - and high air pressure in winter. This results in a deviation of average wind direction from mainly easterly or mainly westerly over some parts of the oceans. Some ocean regions experience strong seasonal variations in wind direction, including complete reversal. Such wind systems are known as monsoons.

In passing, it should be noted that the convention for indicating the direction of ocean currents differs from the convention used for wind directions. A "westerly" wind is a wind which blows from the west and goes to the east; a "westward" current is a current which



comes from the east and flows towards west. This can cause confusion to people who rarely, if ever, go to sea; but it is easily understood and remembered when related to practical experience with winds and ocean currents. On land, it is important to know from where the wind blows: any windbreak must be erected in this direction. Where the wind goes is of no consequence. At sea, the important information is where the current goes: a ship exposed to current drift has to stay well clear from obstacles downstream. Where the water comes from is irrelevant.

From the point of view of oceanography, knowledge of the planetary wind field above the sea surface has always been unsatisfactory. It is difficult to obtain quantitatively accurate wind data from the oceans, particularly from regions remote from major shipping routes. Advances in numerical modelling of the atmosphere and the use of drifting buoys equipped with pressure sensors greatly improved our knowledge of winds over the Antarctic ocean, but the data are still not adequate for many oceanographic purposes. What is needed in oceanography is accurate measurement of wind gradients rather than pure wind strength, which places much more stringent quality requirements on the individual data. However, significant progress has been made, and will be made over the next decade.

We include for completeness information on the distribution of air pressure at sea level. Air pressure variations affect the ocean only indirectly, through the associated wind systems, and oceanographers are not usually concerned with air pressure maps. Meteorologists need air pressure maps as a basic tool of their trade; but they prepare them for their own purposes. In contrast to physical oceanography, where (with the exception of

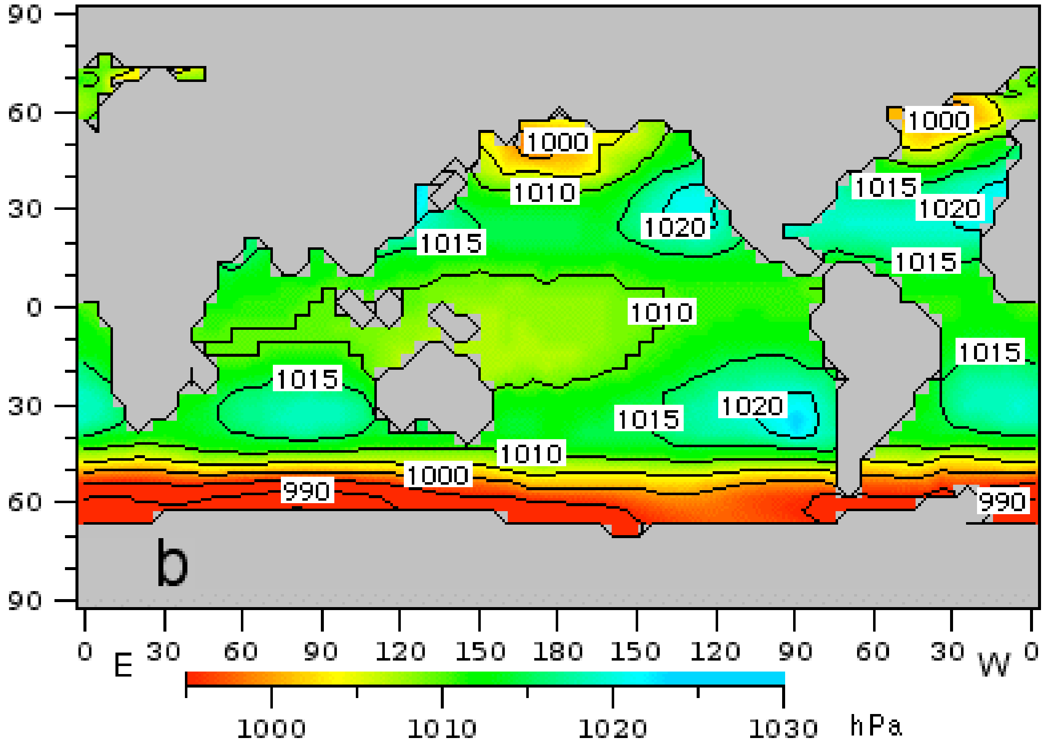


Fig. 1.3. Air pressure (hPa) at sea level. (a, page 6) July mean, (b) January mean. Broken lines show isobars at 5 hPa separation. Data from http://ferret.wrc.noaa.gov/las/, the NOAA Live Climate Data Server using the Comprehensive Atmosphere/Ocean Data Set (COADS) climatology. COADS is based on ship observations.

the northern Indian Ocean) a discussion of the oceanic circulation starts from a well-defined annual mean, meteorology rarely looks at the annual mean atmospheric circulation. This has to do with the low thermal capacity of air, which results in much larger seasonal changes in the atmosphere than in the ocean (see Chapter 18) and makes the annual mean a rather irrelevant quantity. Figure 1.3 therefore shows only the January and July situation. Nevertheless, it gives some useful and instructive information; a comparison with the corresponding wind fields of Figure 1.2 in later chapters will document that the same rules which will be derived for the ocean in chapters 3 - 5 operate in the atmosphere.

The modifications of the basic air pressure pattern of Figure 1.1 by the distribution of land and water are the outstanding features in the seasonal pressure maps. The zonal arrangement of high and low air pressure is seen most clearly in the southern hemisphere. Alternation between low pressure over continents and high pressure over the oceans during summer is particularly evident in the subtropics, but it is easy to see that the basic pattern is preserved in the zonal average. In the northern hemisphere the zonal distribution is disturbed by the Asian land mass which produces a summer low in northern Pakistan and an intense winter high over Mongolia and is responsible for the monsoon winds which dominate the Indian Ocean.

When the wind field is compared with the pressure field it is seen that the nearly zonal pressure distribution in the southern hemisphere produces strong and persistent Westerlies between 40° and 60°S. The remainder of the ocean is dominated by wind systems characterized by wind movement around centres of high and low pressure. During northern summer, for example, the Trade Wind and the Westerlies over the North Pacific Ocean are elements of a wind system in which air circulates around an atmospheric high in a clockwise manner. It is a general rule that air moves around atmospheric highs in a clockwise direction in the northern hemisphere and in an anti-clockwise direction in the southern hemisphere. Likewise, movement around atmospheric lows is anti-clockwise in the northern hemisphere, clockwise in the southern hemisphere. In meteorology and oceanography, circulation around a centre of low pressure is called cyclonic, circulation around centres of high pressure is called anti-cyclonic.

Winds drive ocean currents by releasing some of their momentum to the oceanic surface layer. The important quantity in this process is the wind stress, which is roughly a quadra-

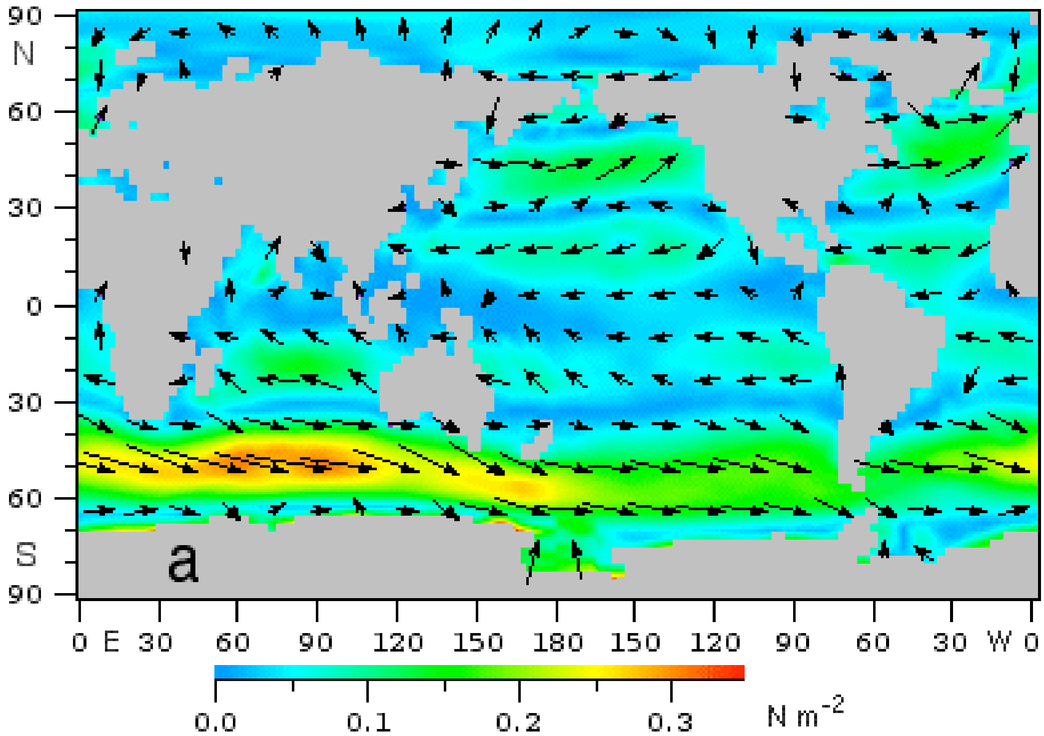
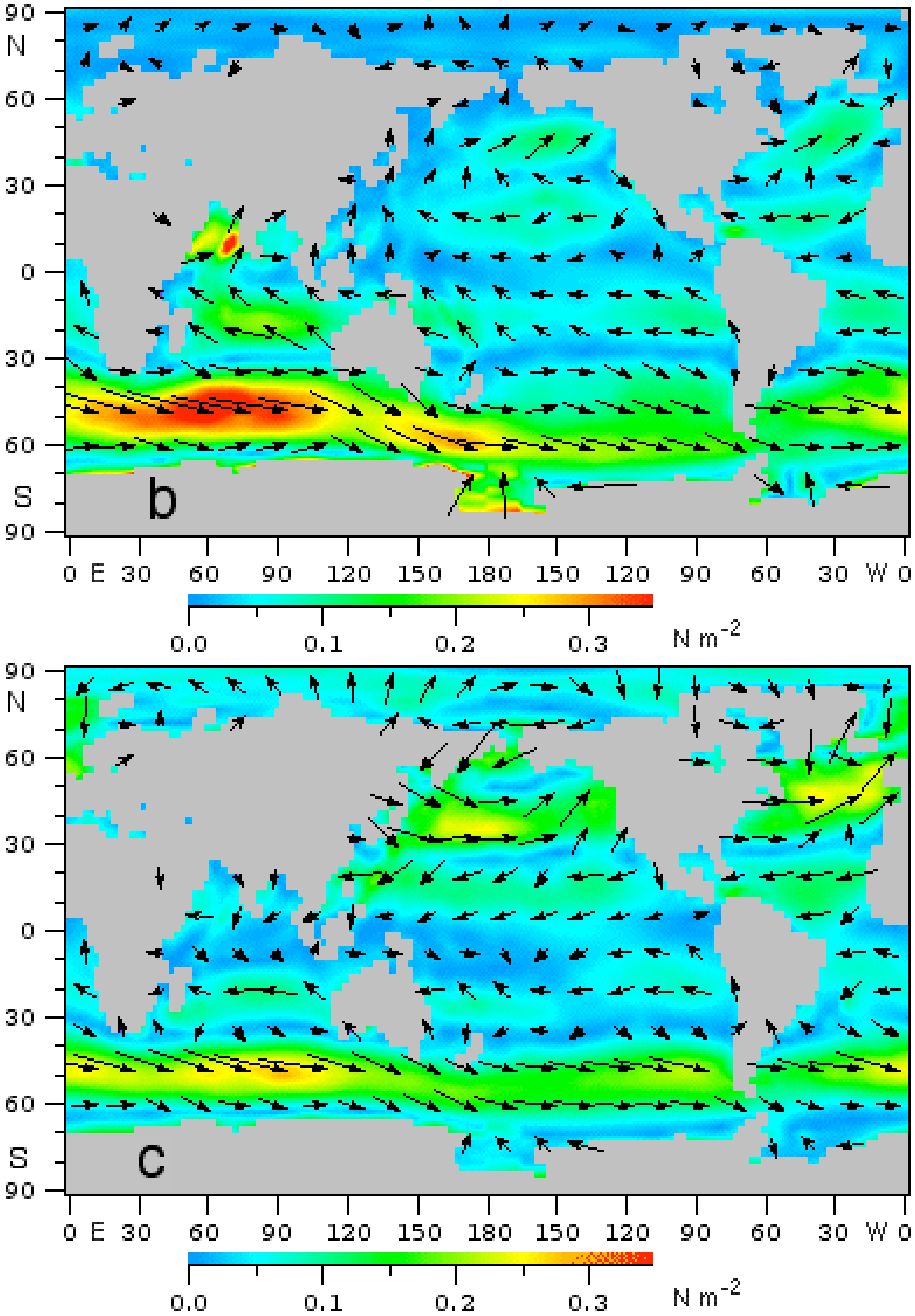


Fig. 1.4. Wind stress over the World Ocean. (a) Annual mean, (b, page 9) June - August, (c, page 9) December - February. Data from http://ferret.wrc.noaa.gov/las/, the NOAA Live Climate Data Server using the Trenberth *et al.* (1989) climatology. Colour indicates wind stress magnitude.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tic function of wind speed. Our knowledge of the wind stress distribution over the ocean is even less well established than our knowledge of the wind field. Most winds contain a considerable amount of turbulence, experienced as short gusts interspersed with periods of relative calm. Because of the quadratic relationship between wind speed and wind stress, gusty winds create larger stresses than would a steady wind of the same average speed. It is possible that our standard measuring equipment does not resolve all wind gusts adequately and that as a consequence our estimates of oceanic wind stress are too low. Direct measurement of the wind stress is difficult; it requires special equipment and has only been done in a small number of locations. The few direct observations were used to develop a formula useful for routine estimation of wind stress. The formula links the stress $\boldsymbol{\tau}$ to routine merchant ship observations of wind speed, air and sea temperatures, wave state and other relevant quantities. $\boldsymbol{\tau}$ is a vector with units of force per unit area (kg m^{-1} s^{-2}, or Newton per square metre) that points directly downwind; its magnitude is calculated from

$$| \boldsymbol{\tau} | = C_d \, \rho_a \boldsymbol{U}^2 \, , \qquad (1.1)$$

where ρ_a is air density (about 1.2 kg m^{-3} at mid-latitudes), $\boldsymbol{U}$ is wind speed at 10 m

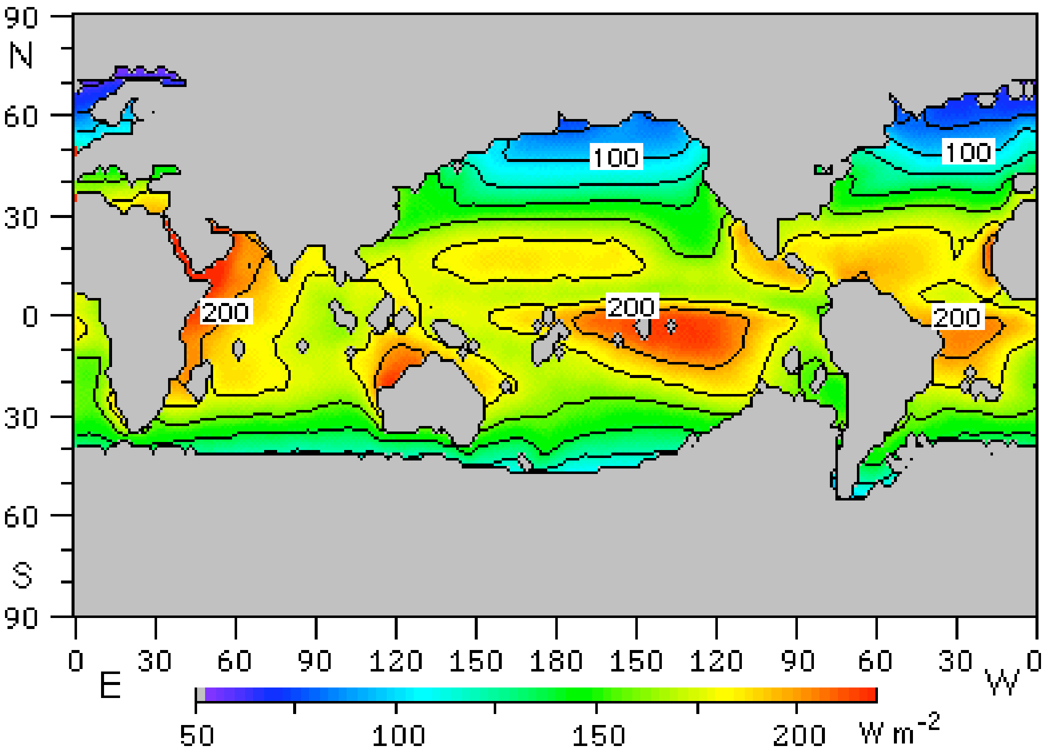

Fig. 1.5. Annual mean solar radiation (W m^{-2}) received at sea level. Data from Oberhuber (1988). 200 W m^{-2} will warm a layer of water 50 m deep by about 2.5°C per month if unopposed by heat losses from other effects. Regions with insufficient data to construct an annual mean are gray. The contouring interval is 20 W m^{-2}.

above sea level, and C_d is the dimensionless "drag coefficient". (Here and in the following, we use bold characters to denote vectors and normal italics for scalars and constants.) Appropriate values for C_d are still the subject of active research, and the uncertainty about its value adds to the lack of precise knowledge about the wind stress distribution over the ocean. C_d varies from about 0.001 to 0.0025 depending on the air-sea temperature difference, the water roughness, and on the wind speed itself. A median value is about 0.0013. Figure 1.4 shows a recent representation of the oceanic wind stress field calculated from eqn (1.1) on the basis of merchant ship data and often used in numerical ocean models. Note that the mean wind stress is not necessarily parallel to the mean wind but is determined by the direction of the strongest winds. Around Antarctica, for example, mean winds are westerlies (Figure 1.2) but the mean wind stresses follow the northwesterly direction of the strong winds in the storm systems. In the northern hemisphere the gusty Westerlies produce larger stresses than the strong but less gusty Trades.

We conclude this chapter by briefly reviewing the atmospheric conditions imposed on the fluxes of heat and mass. Figure 1.5 gives annual mean solar radiation as received at the sea surface; 93% of it is absorbed by the ocean. Solar radiation is naturally largest in the tropics and in cloud-free regions. Again, the observed field shows significant departures from a simple zonal pattern as a result of the distribution of land and water, mainly through

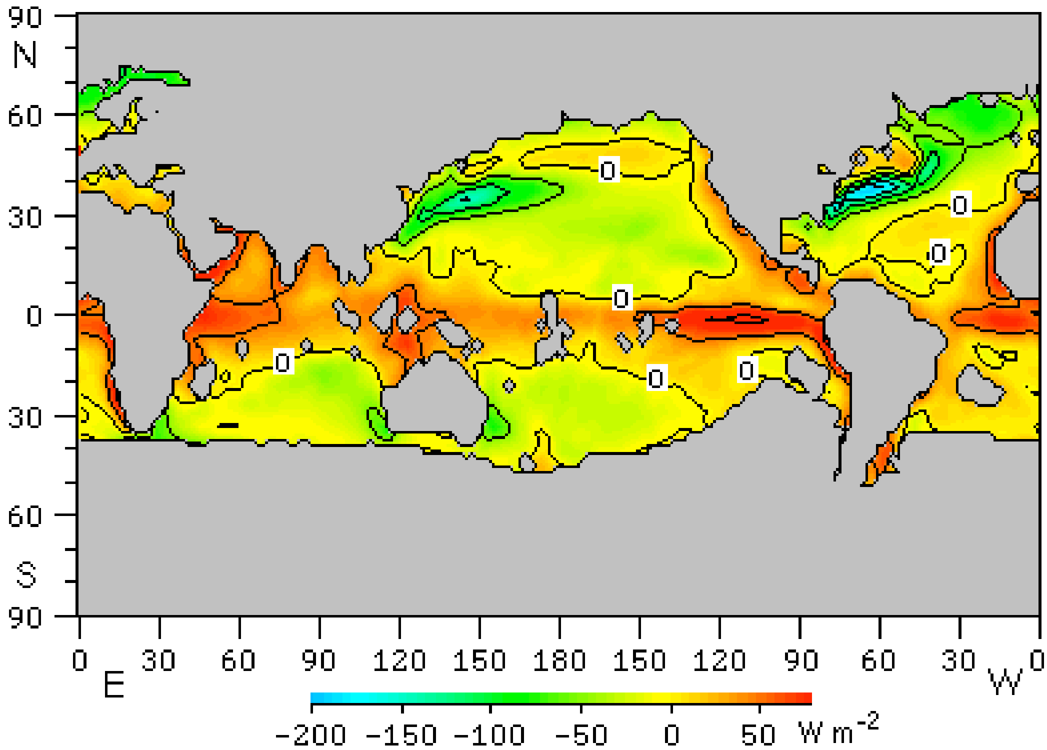

Fig. 1.6. Annual mean heat flux into the ocean (W m^{-2}). Minimum values in the Kuroshio exceed –150 W m^{-2}, in the Gulf Stream –200 W m^{-2}.Data from Oberhuber (1988). Regions with insufficient data to construct an annual mean are gray. The contouring interval is 50 W m^{-2}.

the effect this distribution has on the distribution of atmospheric water vapour and clouds. As an example, the wind convergence between the northern and southern hemispheres' Trades, known as the Intertropical Convergence Zone or ITCZ, consistently shows strong cloud cover and high rainfall (this will be discussed in more detail in Chapter 18) and is thus characterized by a regional minimum of solar radiation. The heat flux through the ocean surface is determined by the balance between four components - incoming solar radiation, outgoing back radiation, heat loss from evaporation, and mechanical heat transfer between the ocean and the atmosphere (for details see Dietrich *et al.*, 1980, or Pond and Pickard, 1983), and their sum can be positive or negative. Each of the four contributions are hard to estimate accurately, so their balance is not very accurately established. Nevertheless, the need for heat flux values as input for ocean models caused several researchers to draw world maps of the balance (Figure 1.6).

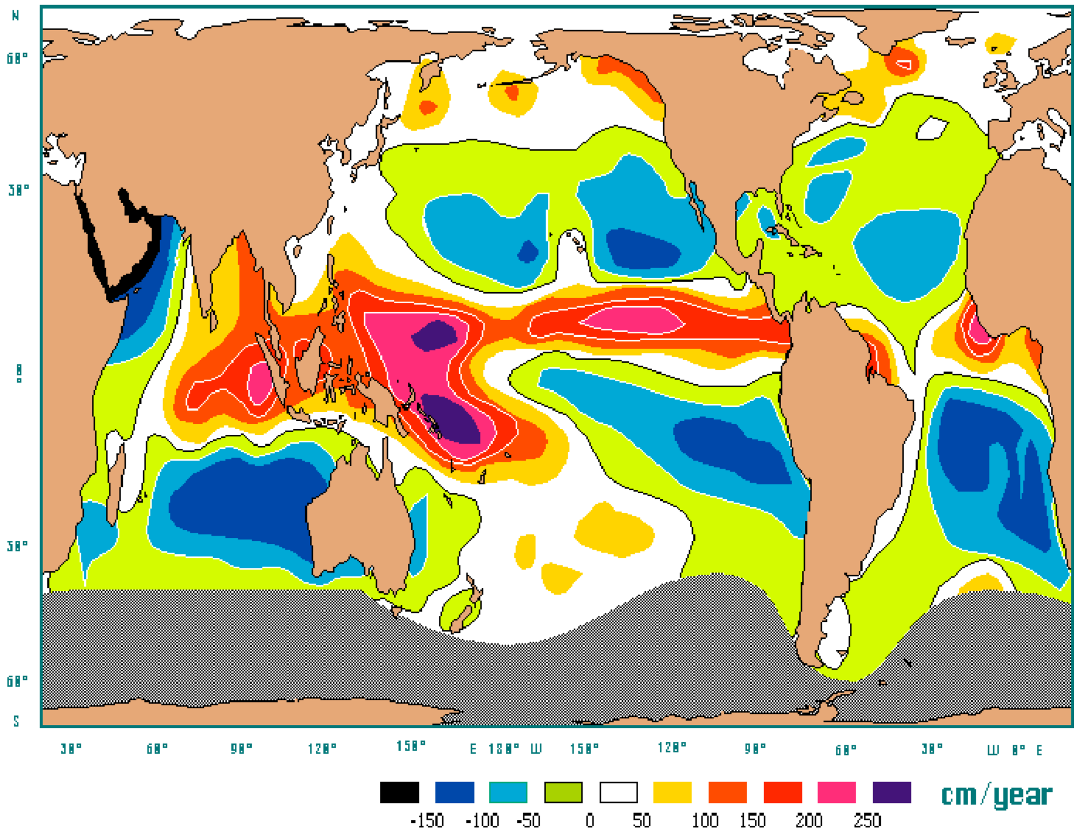

Fig. 1.7. Annual mean difference precipitation - evaporation (P - E, cm per year). Data from Oberhuber (1988). Positive values indicate freshwater gain. The quantity E-P, often seen in oceanography, is the negative of the quantity displayed here.

Generally speaking, the ocean gains heat in the tropics (between 20°S and 20°N) and loses heat in the temperate and polar regions. Departures from this simple zonal distribution are, however, so large that this generalization becomes rather meaningless. Cool water must flow into the regions of net ocean heat gain, and the warmed water must flow away from these regions; this advection does not occur uniformly at all longitudes but in currents of limited longitudinal extent, e.g. along the coasts of Peru and Somalia. Similarly the large heat losses in the Kuroshio and Gulf Stream regions along the coasts of Japan and the eastern USA are caused by rapid poleward advection of warm water. These processes will be addressed in detail in the discussion of individual oceans.

The mass or freshwater flux, i.e. the transport of water between the ocean and the atmosphere, is controlled by the difference between rainfall and runoff from land on one hand and evaporation from the ocean surface on the other hand. (Evaporation from land need not be considered here, since it does not represent a gain or loss to the ocean.) Figure 1.7 shows a recent estimate of the annual mean distribution of precipitation minus evaporation (P-E). Maximum P-E values are found in the ITCZ (known to mariners as the Doldrums) where moist air rises to great height, releasing its water vapour; values of 500 cm/year and more are observed east of Indonesia. Mean sea surface salinity, shown in Figure 2.5b, clearly reflects the mass flux field, most notably in the generally zonal arrangement of the isohalines: lowest salinities tend to occur in regions of maximum P-E, although the relationship is not that simple in detail. Again, modifications are generated by the distribution of land and water and by air and ocean currents. An obvious example is the effect of the limited communication between Indian Ocean and Red Sea waters which produces extreme surface salinities in the latter. Further discussion of these and other aspects is postponed to the appropriate chapters.