The Oceans and Climate Second Edition

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PUBLISHED BY THE PRESS SYNDICATE OF THE UNIVERSITY OF CAMBRIDGE The Pitt Building, Trumpington Street, Cambridge, United Kingdom

CAMBRIDGE UNIVERSITY PRESS The Edinburgh Building, Cambridge CB2 2RU, UK 40 West 20th Street, New York, NY 10011–4211, USA 477 Williamstown Road, Port Melbourne, VIC 3207, Australia Ruiz de Alarcón 13, 28014 Madrid, Spain Dock House, The Waterfront, Cape Town 8001, South Africa

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First published 1996 Reprinted 1998, 1999, 2001 Second edition 2003

Printed in the United Kingdom at the University Press, Cambridge

Typeface Times 10/12 pt. System LATEX 2_E [TB]

A catalogue record for this book is available from the British Library

Library of Congress Cataloguing in Publication data

Bigg, Grant R. The oceans and climate / by Grant R. Bigg. – 2nd edn. p cm. Includes bibliographical references and index. ISBN 0 521 81570 3 – ISBN 0 521 01634 7 (pbk.) 1. Climate changes. 2. Ocean–atmosphere interaction. I. Title. QC981.8.C5B54 2003 551.5'246–dc21 2003043956

ISBN 0 521 81570 3 hardback ISBN 0 521 01634 7 paperback

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1 The climate system

Traditionally climate was defined as the average atmospheric state over at least a score of years, modulated by the seasonal cycle.

Such a definition concealed the temporal variability which produced the mean state, and the complexity of the underlying physical, chemical, biological, geographical and astronomical processes contributing to the climate system. It also implicitly assumed that the climate of a locality does not change over decades, while accepting long-term changes such as glacial periods. In recent decades this perspective of climate has changed. The strong coupling of different constituents of the climate system is now widely recognized, as is the fragility of 'stable' climate.

The traditional definition of climate contains two elements which lead us towards the concept of climatic change. First, the seasonal cycle demonstrates one scale of change through the direct impact of the annual change in solar radiation on the atmosphere and the *biosphere* (the collective term for plant and animal life). Contrastingly, the implication of stability over a number of years implies that basic balances exist within the system. It also implies that any decadal changes in the surface forcing of the atmosphere, from the land or ocean, are small; Chapters 5 and 6 will show that this is not always the case but in general this is a reasonable claim.

In recent decades concern over potential climatic changes due to humanity's activities has arisen, although acknowledgement of the possibility was first made early in the nineteenth century. This book will explore the natural climate system, and potential changes, man-made (*anthropogenic*) or otherwise. Its dominant theme will be the contribution of the oceans to these processes. Observable changes to the climate due to anthropogenic inputs could be expected to have already occurred; Chapters 6 and 7 will show that recent climatic shifts are, by the year 2002, increasingly likely to be largely attributable to these inputs. We will see that numerical climate models are an important means towards this recognition. The stability of the ocean's thermal and freshwater responses to change, its absorption of a significant proportion of anthropogenically derived compounds from the atmosphere, and its coupling with the atmosphere, form important pieces of the puzzle of climatic change.

To understand how the ocean affects, and is affected by, the climate we need to briefly consider the climate system as a whole. It is a complex, many-faceted system; Fig. 1.1 illustrates its major constituents and interactions. There are five components: the atmosphere, the ocean, the cryosphere (ice sheets, transient snow and sea-ice), the biosphere, and the geosphere (the solid earth). The system





is driven by short wavelength, principally visible and ultra-violet, solar radiation, with longer wavelength, infra-red, radiation being emitted into space to maintain thermal equilibrium. In the rest of this chapter we will briefly examine the different components of the climate system, including the timescales on which it operates and the principles behind the numerical models that increasingly are required to study it. However, we will begin with a short discussion of the basic energy source driving the climate – radiation from the Sun.

1.1 Solar radiation

The interior of the Sun, where the nuclear reactions occur that ultimately lead to life on Earth, is incredibly hot, at a temperature of several million degrees Celsius. However, the *electromagnetic radiation* (see Appendix A) that provides the energy for the climate system is derived from the outer layers of the Sun. The greatest amount of radiation comes from the *photosphere*, a layer some 300 km thick in the solar atmosphere. This varies in temperature from $10\,000K^1$ at the bottom to 5000K at the top. Outside the photosphere are much less dense regions – the chromosphere and corona. While these outer regions are at much higher temperatures, up to millions of degrees in the corona, their low density means that they radiate relatively little energy. Most of this is at very short, X-ray and gamma-ray wavelengths which affect the upper atmospheres of the planets (see §3.7 – Carbon 14) but do not penetrate into the lower atmosphere.

The Sun appears to us as (almost) a *black body*. That is, the *spectrum* and total energy of electromagnetic radiation emitted from the Sun (as from all surfaces, and indeed molecules) is a function of its temperature. The total energy flux, *E*, emitted by a black body follows the *Stefan–Boltzmann Law*:

$$E = \sigma T^4 \tag{1.1}$$

¹ The absolute scale of temperature is in degrees Kelvin (K). In this scale 0K is the coldest possible temperature when all molecular motion has stopped. The freezing point of water, 0°C, is 273.16K in this scale. Note, however, that a change of 1K is equivalent to a change of 1°C.

Fig. 1.2. The Sun's spectrum, seen from space (broken line). Both scales are logarithmic. For comparison, a Planck spectrum for a temperature of 5785K is shown (solid line). Note the accentuation of long and short wavelength energies emitted by the Sun, particularly during solar flares.



where σ is the Stefan–Boltzmann constant and *T* is the temperature in degrees Kelvin (a list of constants and their values can be found in Appendix A). The energy density, E_{λ} or radiant energy per unit wavelength, λ , per unit volume per second, is given by

$$E_{\lambda} = \frac{8\pi c}{\lambda^5} \left[\frac{1}{e^{hc/\lambda kT} - 1} \right]$$
(1.2)

where *c* is the speed of light, *k* is Boltzmann's constant and *h* is Planck's constant. The Sun's spectra, as observed from space (Fig. 1.2), obeys (1.2) for a temperature near 6000K. However, for very small (X-ray) and very long (microwave) wavelengths the solar spectrum is enhanced due to contributions from the outer regions of the solar atmosphere (see §§3.7 and 7.1.1).

The vast majority of the energy that reaches the Earth comes from the ultraviolet through visible to infra-red part of the spectrum. The peak energy is in the visible, near wavelengths that we see as the colour blue. The variation in the amount of energy emitted by the Sun is probably small on non-geological time scales. At the Earth's distance from the Sun this *solar constant* is about 1.38 kWm^{-2} . On very long time scales, comparable with the life of the planet, astrophysicists believe that the Sun's irradiance varies dramatically as the supply of fuel within the Sun changes. We will see in §1.8 that variation in the Earth's orbit can affect the amount of energy reaching the Earth's surface by a few per cent, on time scales of thousands of years. However, over several decades to centuries solar irradiance is thought to vary by significantly less than this. Satellite measurements extend back only to 1978 and these reveal irradiance changes of only 0.08% between *sunspot* maxima (higher) and minima (lower).

This does not, however, preclude larger changes in more active beats of the 11 year solar cycle, or the existence of frequencies in the Sun's behaviour of which we have only a dim perception (see §7.1.1).

1.1.1 The effective temperature of the Earth

If the Earth was a sterile planet like the Moon, with no atmosphere, oceans or biosphere what temperature would we expect the surface to possess, given the solar constant, *S*, at the Earth's astronomical position? If we think of the Earth as a flat disc, viewed from the Sun, then the surface area illuminated by solar radiation is πr^2 , where *r* is the radius of the Earth. The energy absorbed is thus $(1 - a)S\pi r^2$, where *a* is the *albedo*, or the proportion of the Sun's energy reflected from the Earth back into space (*c*. 30%).² For equilibrium between the absorbed solar radiation and the emitted radiation from the whole Earth's surface of area $4\pi r^2$, the Earth's temperature, *T*_E, will therefore, from (1.1), be

$$T_E = \left[\frac{(1-a)S}{4}\right]^{0.25}$$
(1.3)

Equation (1.3) gives a surface temperature for this hypothetical atmosphereless planet of 255K, or -18° C, much colder than the Earth's average surface temperature of about 16° C. This effective planetary temperature is more typical of the real atmospheric temperature at a height of about 6 km above the surface. The atmosphere clearly has a significant impact on the distribution of the energy contributing to this effective temperature and will thus be the first component of the climate system to be considered.

1.2 The atmosphere

The atmosphere is a largely homogeneous mixture of gases, both horizontally and vertically, over the height range important for climate: namely the *troposphere* and *stratosphere* (Fig. 1.3). The composition of this apparently stable mixture, air, is shown in Table 1.1. The balance of the dominant constituents of air is thought to have evolved considerably over the lifetime of the planet; for instance, oxygen is likely to have been a product, rather than a necessity, of life (the *Gaia hypothesis*). The climate system's immense natural variability will be a recurring theme of our discussion throughout the book.

The temperature of the atmosphere varies strongly both in the vertical and with latitude. The latter is due to an imbalance in the radiation received over the Earth's surface throughout the year due to the planet's orbit and *obliquity* (Fig. 1.4 and §1.8). The circulation of both the atmosphere and ocean are ultimately derived from this energy imbalance; they act to counter it, in the ratio of about 3:2 respectively.

The vertical temperature distribution shown in Fig. 1.3 comes about because the atmosphere is basically heated from two sources: the ground and the upper

² Note that this average planetary albedo assumes that the hypothetical, sterile, Earth has the same net reflectivity as the real Earth–atmosphere system. Thus this albedo is *not* the surface reflectivity (see §1.4).

Fig. 1.3. Zonal mean

during June at 45°N.

vertical profile of temperature

Gaseous constituent	Molecular form	Proportion (%)
Nitrogen	N_2	78.1
Oxygen	O_2	20.9
Argon	Ār	0.93
Water vapour	H_2O	variable: 0.1–1
Carbon dioxide	\tilde{O}_2	0.037
Methane	CH_4	0.000175
Nitric oxide	N ₂ O	0.000032
Ozone	$\tilde{O_3}$	variable c. 0.000005

Table 1.1. The major constituents of the atmosphere



stratosphere (although we will see in the next sub-section that this is a significant simplification). The ground (or ocean surface) is a heat source since some 49% of the incoming solar radiation is absorbed there. There is also an important heat source between 30 and 50 km above the ground, in the *ozone layer*. When ozone, O₃, absorbs a photon of ultra-violet light – denoted hv because this is the energy of a photon of frequency v – the molecule dissociates in the process to form an oxygen molecule and an energetic oxygen atom, O¹D, where one electron is displaced into a higher energy state than in the ordinary oxygen atom (see Appendix B for the Periodic Table of the Elements and a discussion of electron orbitals). This can then react with an oxygen molecule to reform ozone as part of the Chapman cycle:

 $O_3 + h\nu \rightleftharpoons O_2 + O^1 D \qquad O^1 D + O_2 + M \rightleftharpoons O_3 + M$ (1.4)

The air molecule, M (that is, predominantly N_2 or O_2), is necessary in the second reaction in (1.4) as the reaction produces excess energy. This is carried

Fig. 1.4. Contour plot of daily average insolation at the top of the atmosphere as a function of season and latitude. The contour interval is 50 Wm⁻². The heavy broken line indicates the latitude of the sub-solar point at noon. [Fig. 2.6 of Hartmann (1994), *Global Physical Climatology*. Reprinted with permission from Academic Press.]



away by M thereby stabilizing O₃, which would otherwise dissociate. Such a reaction is called *exothermic*. Hence the air becomes warmer, as the temperature of a medium is merely a reflection of the average kinetic energy of its molecules.

The reactions in equation (1.4) are only part of the full Chapman cycle, which also contains reactions involving photo-dissociation of O_2 , and reactions between the excited oxygen atoms themselves and O_2 . Some of these are likewise exothermic, adding to the energy which is transferred, via chemical reaction, from solar radiation to the middle atmosphere. There are many other reactions involving ozone, some of which will be discussed in §7.2.1.

The lower atmosphere is therefore heated both from above and from below. Between these regions is a zone, in the lower stratosphere, where the energy from these heated regions only weakly penetrates. This is strongly stratified, which means that there are large vertical gradients in the concentrations of trace constituents of the air and the *potential temperature* (see Appendix C). The tropopause, at the bottom of the stratosphere where the gradients are greatest, resists penetration by cloud convection, or even diffusion. The wellmixed region below this, the troposphere, is the part of the atmosphere that we will be largely concerned with, because of its direct interaction with the oceans.

The strong heating of the surface at the equator (Fig. 1.4) makes the air less dense, forcing it to rise. Air flows towards this region of rising air, which tends to be concentrated in a narrow band around the globe known as the Inter-Tropical Convergence Zone, or ITCZ. Aloft, the rising air moves polewards to compensate for the surface flow. In the late seventeenth century, when Halley first proposed this mechanism for driving the atmospheric circulation (modified 50 years later by Hadley) it was believed that this *Hadley cell* extended to the polar regions. This seemed logical, as polar air is cold, and so relatively dense, and should therefore flow towards the low pressure regions of the tropics in order to transfer heat from the equator to the poles and so maintain the Earth's thermal equilibrium. By the nineteenth century this idea was seen to be too simplistic. The *Coriolis force*, due to the solid Earth and moving atmosphere revolving at slightly different rates (see §2.5.2), gives this converging near-surface wind a

Fig. 1.5. Schematic cross-section of the zonal mean circulation in the troposphere. The dotted upper region of the Ferrel cell indicates that it is a less constant feature. The crosses (and E) show where the surface flow has an easterly component, while W shows where the surface flow is predominantly westerly.



westward component, resulting in the observed easterly³ Trade winds. Ferrel therefore proposed an intermediate 'Ferrel cell' in mid-latitudes. Modern observations support this, as shown schematically in Fig. 1.5, a zonal cross-section of the tropospheric flow.

The general circulation of the lower troposphere is shown in Fig. 1.6, and the sea level pressure field for northern winter in Fig. 1.7. The ascending air of the equatorial region is shown by the low pressure. To replace this, easterly winds flow equatorwards driven by high pressure in the sub-tropics, where the air in the tropical Hadley cell descends (Fig. 1.5). This latter, relatively calm, zone has strong westerly winds on its poleward side, which, in turn, lie equatorward of another region of low pressure near 60° of latitude. This region of sub-polar low pressure forms the ascending branch of the polar Hadley cell of Fig. 1.5, with easterly winds at the surface due to polar high pressure. The most vigorous part of this system is where the tropical Hadley cell meets the mid-latitude Ferrel cell. Here the upper level convergence of air produces an extremely strong westerly *jet-stream* in the upper troposphere (Fig. 1.5). This often has a secondary maximum over the mid-latitudes, above the polar front (Fig. 1.5). This polar front jet-stream steers the transient pressure systems that we experience on the ground in the mid-latitudes. The latter systems are a significant mechanism in the redistribution of heat from equator to pole.

The mainly zonally symmetric structure of the general circulation is mostly due to the latitudinal distribution of the solar radiation received by the Earth; the distribution of land and sea over the Earth's surface distorts the zonality. Some aspects of this latter interaction will be discussed in Chapters 2 and 5.

1.2.1 The greenhouse effect

The vertical profile in Fig. 1.3 shows a decline in temperature of about 6.5° C per kilometre in the troposphere. It can be shown that ascending 'dry' air, i.e. air without clouds, changes temperature because of expansional cooling by 9.8° C

³ Confusingly, meteorologists and oceanographers follow different conventions when specifying the direction of fluid flow. Meteorologists use the direction from which the wind has come to describe it, while oceanographers take the direction in which the flow is going. Thus, an east*erly* wind to a meteorologist is a west*ward* wind to an oceanographer! This unfortunate difference is too entrenched to be easily altered, and this book will use the convention appropriate to the fluid medium being described.

Fig. 1.6. Mean surface wind field in (a) January, and (b) July. [Data from Oort, 1983.]





Fig. 1.7. Mean sea level pressure field in January. Contours are every 5 mb. The data is a mean of 17 years of National Meteorological Center model analysis fields. for each kilometre of *adiabatic* vertical motion (the latter occurs if a parcel of air does not exchange any heat with its surroundings, as is a good approximation in, for example, the formation of *cumulus* clouds – see Appendix C). Within a cloud the decline of temperature with height in vertical motion can approach the typical value of Fig. 1.3, due to the release of *latent heat* upon condensation of water vapour. However, substantially less than half of the troposphere contains cloud at any one time so other processes must be lowering the *environmental lapse rate* below the *dry adiabatic lapse rate*. Diffusion and advection of heat from the stratosphere, the ground, or surrounding air masses is partially responsible but the major reason for the enhancement of tropospheric temperatures is the *greenhouse effect*.

A number of low concentration, or *trace*, gases in the atmosphere are unresponsive to illumination by short wavelength radiation from the Sun but absorb energy of infra-red wavelengths. The gas molecules do this by increasing their *vibrational* and *rotational* energies, rather than their kinetic energy. How this happens can be illustrated by the water molecule, shown in Fig. 1.8. The bond angle between the hydrogen atoms of an ordinary water molecule is 105° , but if a *photon* of a certain wavelength of infra-red radiation (6.27 µm)⁴ collides with the molecule the energy of the photon can be converted into a vibration of the hydrogen bonds, such that the angle between the hydrogen atoms undergoes rapid oscillation of a few degrees. Other forms of oscillation can be excited by wavelengths of 2.66 or 2.74 µm. The *absorption spectra* of H₂O, shown in Fig. 1.9, is more complex than just these three wavelengths, however,

⁴ 1 μ m (micrometre, often called a micron) = 10⁻⁶ m.



Fig. 1.9. The absorption spectrum of water vapour. Note the region $8-12 \ \mu m$, known as the 'water vapour window', where there is little absorption of infra-red radiation by the water vapour molecule.

as multiples, or *harmonics*, of the principal absorption wavelengths can also be absorbed. In addition wavelengths which are sums, or differences, of these three (and their harmonics) also show a degree of absorption, although generally of reduced intensity.

Equation (1.2) shows that the wavelength of electromagnetic radiation emitted by an object is inversely related to its temperature. Thus the mean wavelength of the radiation emitted by the Earth's surface, and within the atmosphere itself, will be longer than that of the incoming radiation from the Sun, as the latter has a surface temperature of about 6000K compared to a typical Earth surface temperature of 289K. Fig. 1.10 depicts a typical energy spectrum, seen from the tropopause, of the radiation from the Earth's surface, with the absorption by trace gases shown by shading. There are regions of the spectrum, such as wavelengths shorter than 8 μ m and from 15 to 20 μ m, where the infra-red radiation is almost totally absorbed by atmospheric gases. It is this absorption, and the associated re-emission of energy, much of which warms the troposphere, that is called the greenhouse effect. This name is a misnomer as the physical mechanism involved in keeping a greenhouse warm is totally different from this radiative physics. There is a small contribution from glass being transparent to solar radiation, but partially reflective to the outgoing infra-red radiation from the air and soil within the greenhouse. However, greenhouses are warm predominantly because the enclosed space eliminates convection, and hence mixing with cooler air.

The principal greenhouse gases, and their relative contribution to the greenhouse effect, are shown in Table 1.2. The percentages shown are not strictly additive because the absorption ranges of the different gases overlap. Table 1.2 also gives the fundamental absorption wavelengths of these molecules, but the complexity of the absorption spectra, with their harmonics and linear combinations of these fundamental wavelengths, must be remembered (see Fig. 1.9). Water vapour is two to three times as important in the total greenhouse effect as carbon dioxide. This fact is often neglected in discussions of greenhouse warming because water vapour is highly variable in concentration, both in space and

Gas	Basic absorption wavelengths (µm)	Contribution
Water vapour (H_2O)	2.66, 2.74, 6.27	55-70%
Carbon dioxide (CO_2)	4.26, 7.52, 14.99	25%
Methane (CH_4)	3.43, 6.85, 7.27	5%
Nitrous oxide (N ₂ O)	4.50, 7.78, 16.98	2%
Chlorofluorocarbons (CFCs)	typical bonds:	
	9.52, 13.8, 15.4	1%
Ozone (O_3) , sulphur dioxide (SO_2) , other oxides of nitrogen, carbon		
monoxide (CO), etc.		<1% each

Table 1.2. The greenhouse gases and their contribution to the totalgreenhouse effect

Fig. 1.10. Earth's surface radiation spectrum, seen at the tropopause. The broken line is the black body emission for a typical surface temperature of 294K (21°C). The solid line is the observed spectrum, with the shaded region between denoting the energy absorbed by gases in the troposphere.



time, making it difficult to isolate its global effect. It will, however, be vital to much of our later discussion.

The importance of the greenhouse effect to the *heat budget* of the atmosphere, and therefore the climate system, is shown in Fig. 1.11. The *feedback* mechanism between the radiation from the Earth's surface and the greenhouse re-radiation raises the amount of energy available to heat the surface from 70% of the incident solar radiation (in the absence of the atmosphere) to 144%. Present and future changes to the amounts, and proportions, of the trace gases that contribute to this effect may change these figures and so have implications for the global climate. A number of the greenhouse gases have increased significantly in concentration in the last 200 years. This may be linked to the rise in global average surface temperature of about 0.7°C over the twentieth century. Detailed discussion of these variations will be delayed to Chapters 6 and 7, which examine natural and anthropogenic alterations to the climate system.

Fig. 1.11. Global average pathways for energy in the atmosphere. A notional 100 units comes from the Sun. [From Bigg, 1992a with values modified from Kiehl and Trenberth, 1997.]



1.2.2 Reflected radiation

Another major pathway for energy in the atmosphere is reflection from the surface, clouds or airborne particles. About 30% of the incident radiation is so lost from the climate system. The surface accounts for a quarter of this (the surface albedo) but the predominant loss is from the atmosphere. Variation in the *cloud amount* and type, the amount of suspended volcanic debris or solid chemical aggregates, and the characteristics of the Earth's surface can change the magnitude of this energy sink. While the ocean does not directly affect most of these processes we will see several exceptions later.

1.3 The oceans

The oceans cover 361 million square kilometres, or 71% of the surface area of the globe, almost two and a half times the land area. To the surface observer this immense area seems almost featureless compared with the land, with only icebergs and waves to give a vertical dimension. However, beneath the water surface the ocean floor shows all the orographic richness of the land.

Fig. 1.12 shows the percentage of the Earth's surface in different height bands, relative to mean sea level. The first striking feature is the greater average depth of the ocean compared with the land's altitude. Much of the oceans are more than 3000 m deep, while little of the land surface is above 3000 m in altitude. This discrepancy also appears in the extremes of orography: Mount Everest is 8848 m in altitude, but the deepest point in the Marianas Trench, east of the Philippines, is 11 022 m below sea level. Large areas of the ocean have relatively little variation in depth; these regions are known as abyssal plains. They are usually deeper than 3000 m.

Fig. 1.12. Variation of the elevation of the Earth's surface. The mean elevation of the land is 0.84 km, while the mean depth of the ocean is 3.8 km.



'Mountainous' *bathymetry* separates these deep, oceanic plateaux. For example, the Mid-Atlantic Ridge begins north of Eurasia and essentially splits the Atlantic Ocean into two halves. This ridge then continues eastwards across the Southern Ocean, curving northward into the east Pacific as the East Pacific Rise. A spur extends northwards from this main ridge into the western Indian Ocean.

The character of the oceanic perimeters varies considerably. In places the continental shelves adjacent to the coast are hundreds of kilometres across, for example, off part of western Europe and northern Australia. Elsewhere the shelf may be only a few tens of kilometres wide; a good example is off the west coast of South America. The continental slope, which joins the shallow coastal zone to the deep ocean, can be relatively steep in such regions, with average gradients greater than 1 in 10. In §1.3.2 we will see that these various bathymetric structures are important in guiding the oceanic circulation, particularly the deep water flow.

1.3.1 Chemical composition of the oceans

The oceans are, by volume, 96.5% water. The water molecule has properties that are important for the movement of heat, momentum and climatically active gases (including water itself) between the atmosphere and the oceans. The remaining 3.5% of the oceanic solution – dissolved salts, particles, organic material and gases – plays an inordinately large role in such climatic processes and the circulation in the ocean. We will here consider some chemical properties of water and the dissolved salts; more details of the chemical and biological processes in the ocean which contribute to the climate system will be the subject of Chapters 3 and 4.

Water, or H_2O , is a very special molecule. Other compounds with similar molecular weight, such as methane, CH_4 , or ammonia, NH_3 , are gases at room

Latent heat of fusion	$3.33 \times 10^5 \text{ Jkg}^{-1}\text{K}^{-1}$
Latent heat of vaporization	$2.25 \times 10^{6} \mathrm{Jkg^{-1}K^{-1}}$
Specific heat	$4.18 \times 10^3 \mathrm{Jkg^{-1}K^{-1}}$
Surface tension	$7.2 \times 10^9 \text{ Nm}^{-1}$
Maximum density	$1.00 \times 10^3 \text{ kgm}^{-3}$
Heat conductivity (at 290K)	$5.92 \times 10^{-2} \text{ Js}^{-1} \text{m}^{-1} \text{K}^{-1}$
Molecular viscosity (at 293K)	$1.0 \times 10^{-2} \ { m Nsm^{-2}}$

Table 1.3. Some physical properties of liquid water

temperature. Water, by contrast, is a liquid, and is readily found in its solid state (ice) below 0°C. The reason for this unusual behaviour lies in the molecular structure, already seen in Fig. 1.8. A water molecule is composed of an oxygen atom, bonded to two hydrogen atoms separated by an angle of 105°. As a result of the re-arrangement of the atomic electron orbitals (see Appendix B), the oxygen atom accrues a small net negative charge, while the hydrogen atoms gain a small positive charge. This makes the molecule dipolar and allows groups of molecules to form aggregates, with oppositely charged portions of the molecules adjoining each other, held together by *coulombic* forces. These structures resist break-up and permit water to be in a less energetic state than other, similar molecular weight, compounds at ordinary temperatures. Such unusual bonding leads to a number of important physical properties. These are summarized in Table 1.3, but are worth elaboration because we shall see their effects in later chapters.

The latent heat of fusion (the energy required to melt 1 kg of ice) and the latent heat of vaporization (that needed to evaporate 1 kg of water) are among the highest for any substance. This has important implications for the climate system as, conversely, this energy is released to the environment when water changes state to a more ordered structure. For instance, when water vapour condenses to form water droplets in a cloud, the energy latent within the vapour is released as heat and contributes to the driving energy of the cloud-producing process.

Related to these properties is the specific heat, highest of all solids and liquids except ammonia. This is the amount of energy required to increase the temperature of one kilogramme of the substance by 1°C. Dry air requires less than a quarter of the energy water needs to heat 1 kg by 1°C, and when the thousand-fold difference in density is taken into account, it is quickly seen that the ocean will be much slower in responding to heating, or cooling, than the atmosphere. Climatically, this is an extremely important property, as not only does it explain the smaller annual range in temperature of maritime climates, but it also points to the ocean's ability to act as a flywheel for longer term climatic change. Energy can be both stored and released over decades, or even centuries, by the ocean while the atmosphere reacts to energy changes with time delays of only a few weeks.

The heat conductivity and molecular viscosity of water are also strongly affected by the inter-molecular forces, being unusually high and low respectively. These parameters give the mixing ability of the liquid with respect to heat and molecular motion. However, mixing within water principally occurs because

Constituent	Ion	Average concentration in sea water of salinity 35 psu	Average concentration in river water (%)
Chloride	Cl-	19.350	0.0078
Sodium	Na^+	10.760	0.0063
Sulphate	SO_4^{2-}	2.712	0.0012
Magnesium	Mg^{2+}	1.294	0.0041
Calcium	Ca^{2+}	0.412	0.0150
Potassium	K^+	0.399	0.0023
Bicarbonate/	$HCO_3^{-}/$	0.145	0.0588
carbonate	CO_3^{2-}		

Table 1.4. Concentration of major ions in sea water

of stirring by eddies within the fluid, which are on a much bigger scale than the molecular processes for which conductivity and viscosity are appropriate. Therefore these properties are not significant for the processes with which we are concerned. A final physical property of note that derives from liquid water's unique structure is the high surface tension. This is related to the force needed to break the air–water interface: a high value is detrimental to the speed of gas and particle exchanges between the air and water. This will be of importance in later chapters.

Another consequence of the molecular structure of water is its dissolving power. Sea water is a mixture of many compounds; the main ingredients, apart from water itself, are shown in Table 1.4. The addition of these salts has its own effect on the properties of the mixture. The freezing point of sea water is about -1.8° C, rather than 0°C. This lowering of the freezing point of water upon the addition of salts underlies the salting of roads in winter when nearfreezing temperatures are expected. The density is also affected by the addition of salt. A typical surface sea water density is about 1026 kgm⁻³, an increase of 2.6% above that for pure water (see Table 1.3; note that the density of air near sea level is only about 1.2 kgm⁻³). The density of sea water is a complicated function of temperature, *salinity* (the proportion, by weight, of dissolved salts)⁵ and pressure (see, for example, Gill 1982, Appendix 3). However, because of the strength of the inter-molecular forces within water near its freezing point, it is found that salinity has most effect on density at low temperatures, while temperature exerts the predominant influence at higher temperatures.

The main input of particulate or dissolved material to the oceans occurs through riverine input; there is a small contribution from wind-blown (aeolian) deposits and precipitation. The globally averaged riverine chemical composition is distinctly different from the sea water composition shown in Table 1.4. Bicarbonate (HCO_3^-) is the dominant riverine *anion* and calcium (Ca^{2+}) the most prevalent *cation*. Neither sodium (Na^+) nor chloride (Cl^-) contribute large percentages to the total dissolved ion concentration.

⁵ Since 1982 a salinity scale based on the electrical conductivity of sea water has been used. The average salinity of the oceans is 35×10^{-3} , or 35 practical salinity units (psu) in this scale (a dimensionless number). Salinity values in psu are essentially identical to a measure of parts per thousand (‰) by weight.

Sea water appears to be of remarkably stable composition; the salinity may vary but the proportions of the different salts remain almost constant. Therefore, for a considerable time, perhaps hundreds of millions of years, the riverine input has been in balance with processes that remove the salts, such as sedimentation and ejection into the atmosphere. The excess chloride in sea water, in comparison with the riverine source, is thought to have come from volcanism early in the Earth's history. Volcanic eruptions would have emitted large quantities of the very soluble gas HCl. Dissolution of this gas in the sea forms a very weak hydrochloric acid solution. The input of bicarbonate over time has neutralised this, to leave sea water as slightly alkaline (pH c. 8.0). Another non-riverine input to the oceans which has contributed to the concentration of trace constituents is submarine hydrothermal activity on mid-ocean ridges. These sites are sources for dissolved gases like helium, and some metals, for example manganese.

The chloride ion concentration in sea water may be explained by early volcanism, but sodium is not a large component of volcanic gases. Sodium must therefore attain its abundance by other means. This leads us to an important concept in the chemistry of the environment, namely the concept of elemental *cycling*. Many chemical elements cycle repeatedly through various parts of the Earth's outer crust, atmosphere and oceans in such a way that the concentration in each component of the system is stable over long periods of time. The carbon cycle is the best known of these, and will be considered in detail in Chapter 3. Other environmentally important cycles include sulphur, calcium, sodium and lead. The time taken for an atom to complete one cycle is determined by the average *residence time* of an atom of the element concerned within the different components of the cycle. In the ocean, for instance, this is defined as the total amount of the element in the ocean divided by the riverine and aeolian input per year. This definition implicitly assumes the cycle is in long-term balance.

To explain the abundance of sodium in sea water, relative to calcium, we can use the concept of residence time. Calcium has a much shorter residence time in the oceans than sodium. This is because it is a major constituent in the skeletons of marine organisms and is therefore easily lost to the ocean through settling of dead organisms onto the sea floor. It is therefore taken out of the oceans into the sediments sufficiently fast for more sodium, from riverine inputs, to accumulate in the oceans than calcium.

Cycling can also be applied to important molecules, as well as elements. The hydrological cycle is merely the cycle of water through the climate system. This is illustrated in Fig. 1.13. The residence time of water in the ocean is 3220 years. This can be compared with the time taken for material injected into the deep ocean to mix thoroughly around the globe (typically 1000 years). By contrast, the residence time for water vapour in the atmosphere is only 10 days and the mixing time is of the order of a month.

1.3.2 Ocean circulation

A schematic of the global ocean circulation is shown in Fig. 1.14. This is known as the *thermohaline* circulation because it is driven by density contrasts. The basic structure consists of deep water being carried towards the Pacific Ocean, which upwells along route and is transported back in near-surface currents to

Fig. 1.13. Reservoir sizes in the hydrological cycle. The accuracy of several of the components is poor, making it difficult to accurately close the cycle. [Reprinted with permission from Chahine (1992), *Nature*, **359**, 373–9.]



Fig. 1.14. Schematic of the thermohaline circulation of the global ocean – the Conveyor Belt. The broken arrows represent the major surface components of the circulation. The continuous line denotes the deep water circulation, emanating from source regions denoted by open circles. Slow upwelling in the Atlantic, Indian and Pacific Oceans closes the circuit.



the downwelling regions in the North Atlantic (principally the Norwegian– Greenland Sea and the Labrador Sea) and the Weddell Sea. This cycle, known as the *conveyor belt*, is very important in the climate system. We will see in later chapters that it provides a stabilising effect on climate, because of its long time scale (see the last section), but can also cause abrupt climatic change in the space of a few decades if it is disturbed in certain ways.

The conveyor belt mechanism is naturally a gross simplification. The mean surface circulation is shown in Fig. 1.15. It has several shared features in each basin. Sub-tropical *gyres* rotate *anticyclonically* in each of the main ocean basins. The western margins of each of these gyres have strong poleward currents, such as the Gulf Stream in the North Atlantic. Poleward of these gyres there is some evidence in the Northern Hemisphere for *cyclonic* sub-polar gyres, where the westerly winds change to polar easterlies. In the Southern Hemisphere the water is able to flow around the entire globe, driven by the strong westerly winds at these latitudes. Sub-polar gyres exist in the Weddell and Ross Seas, poleward of this Antarctic Circumpolar Current.

The surface flow in the tropics consists of strong westward flowing currents at, and near, the equator. These are extensions of the tropical arm of the sub-tropical gyres. Between these two westward currents a counter-current,



Fig. 1.15. The global surface current system. Cool currents are shown by dashed arrows; warm currents are shown by solid arrows. The map shows average conditions for winter months in the Northern Hemisphere: there are local differences in the summer, particularly in regions affected by monsoonal circulations. [Fig. 3.1 of Open University Course Team (1989). Reprinted with permission from Butterworth-Heinemann.]

flowing eastwards, is usually found. There is also typically a strong eastward current below the surface on the equator, the equatorial under-current. The equatorial currents are intimately coupled with the atmosphere and will be discussed further in Chapters 2 and 5.

The deep circulation, shown in Fig. 1.16, conveys water that has sunk in the polar regions throughout the world oceans. In the Greenland and Norwegian Seas during winter the surface waters are strongly cooled making them denser. This dense water then overturns, probably in very localized regions up to a few tens of kilometres in diameter. In the Southern Hemisphere, particularly in the Weddell Sea, ice formation leaves a greater concentration of salt in the water beneath, as salt tends to be expelled from the ice lattice as it forms. This dense water also sinks. These two distinct types of water, or *water masses*, then travel equatorwards from the polar regions, to form the deep waters of the world's oceans. This deep water circulation is driven by subtle differences in temperature and salinity.

The water at intermediate depths also comes from the sinking of water masses, but those formed in less extreme circumstances. One such important contribution comes from the Mediterranean Sea. Intense evaporation raises the salinity of this basin above that of the North Atlantic. This dense, saline water sinks to form the deep water of the Mediterranean basin and spills out over the sill at the Strait of Gibraltar. Here, surface water flows into the basin from the North Atlantic, to compensate for the intermediate water outflow and the Mediterranean's evaporation. To conserve mass locally, the deep water flows out into the Atlantic beneath. This warm, salty water is evident in the intermediate layers of the North Atlantic for thousands of kilometres, contributing about 6% of the North Atlantic's salinity. Under some climatic situations, although probably Fig. 1.16. The deep circulation of the global ocean. The two main sources of deep water are shown by open circles. The deep water originating from the North Atlantic is slightly less dense than the deep water of Antarctic origin; hence its path is shown by dashed arrows until it merges with the latter (whose path is shown by a continuous line).



not at the present time, it may also act to pre-condition the water entering the Norwegian–Greenland Sea, that will later overturn and produce deep water, by making it denser than it would otherwise be.

1.4 The cryosphere

Ice covers about 5.7% of the Earth's surface and contains 2.05% of the Earth's supply of water (the oceans contain 97.25%). It is very variable seasonally and comes in many forms: continental ice sheets, mountain glaciers, shoaled ice shelves, sea-ice, snow and perma-frost.

In this book we will mainly be concerned with ice as sea-ice. However, landice has a high albedo -0.95 for freshly fallen snow, over 0.4 for old snow and ice – and low temperature. As the average albedo of the Earth is 0.3, and of the oceans 0.08, the ice cover both on land and on the oceans drastically reduces the heat energy entering the climate system (see §2.1 for albedos of other substances). Ice cover also tends to reduce the input of moisture, and hence latent heat, to the atmosphere by evaporation. The quantity of global ice has varied significantly in the past. A hundred million years ago there was probably almost none. As recently as 15 000 years before present (BP) continental ice covered much of Canada, the northern United States, northern Siberia and northern Europe and winter sea-ice may have extended to the latitude of northern Britain in the eastern Atlantic.

Another impact of ice on the global environment is its effect on sea level. Eighteen thousand years ago Britain was joined to western Europe, Australia and New Guinea were one large island, and the Black Sea was isolated from the Mediterranean. The volume of water stored in continental ice sheets resulted in sea level being 120 m lower than today. If all the ice presently on the Earth's land masses⁶ melted sea level would rise by 80 m, flooding most coastal regions. Sea level has been rising over the twentieth century at the rate of about 1.5 mm/year, although about two thirds of this rise is thought to be due to thermal expansion

⁶ Sea-ice is floating on the ocean, as it is less dense than water, and only displaces its own weight of sea water. Therefore, if it melted the melt water would merely replace the volume of ice previously submerged.

Fig. 1.17. Average boundaries of sea-ice at the end of winter and summer in (a) the Northern Hemisphere, (b) the Southern Hemisphere. The average winter extent of icebergs is also shown, and their source region in (a). [Reprinted, with permission, from J. G. Harvey, *Atmosphere and Ocean: Our Fluid Environment* (London: Artemis Press, 1985), pp. 36–7, Fig. 4.6.]



(a)

of sea water (as the global temperature has increased slightly – see \$1.8) and tectonic effects rather than melting of land ice. In \$&6.2.2 and 7.2.4 we will examine the phenomenon of sea level change in more detail.

The previous section demonstrated a significant climatic interaction of polar ice, that of providing a mechanism for deep water formation. This mostly occurs in the Southern Hemisphere (*austral*) winter under the *shelf ice* of the Weddell and Ross Seas, off Antarctica. Here the ice can be tens to hundreds of metres thick.

Sea-ice thermally insulates the ocean from the atmosphere. It also decouples the ocean from direct driving by the wind. *Pack ice* essentially flows in the same direction as the underlying ocean; shelf ice, however, is also subject to motion induced by its contact with continental shelves and land glacier forcing.

The extent of oceanic ice cover varies dramatically with the season. Fig. 1.17 shows seasonal extremes for the two hemispheres. Comparing Fig. 1.15 with Fig. 1.17 shows the impact of the ocean circulation on ice distribution in areas such as east of Greenland, where the local current pushes ice much further south than elsewhere.