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Atmosphere and ocean interactions

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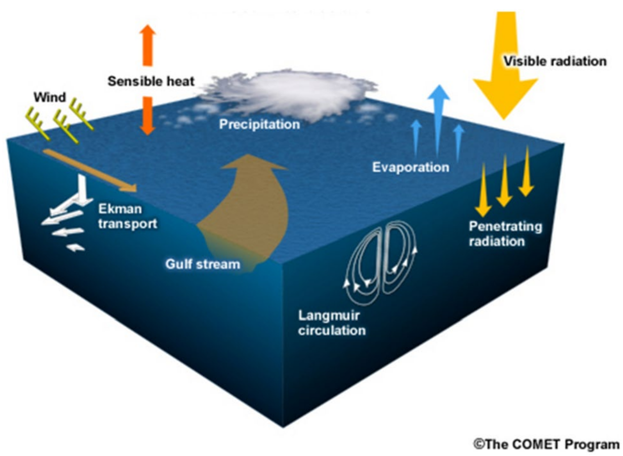
Abstract
The nature and extent of the main interactions between the Earth's atmosphere and the ocean are briefly reviewed, introducing the main properties of these two planetary systems, the hierarchy of timescales and spatial scales characterizing their processes, and the main mechanisms involved. To clarify these mechanisms, a few examples of exchanges of momentum, energy, and mass exchanges are briefly outlined. The paper aims at stimulating students to discover the fascinating interplay among two of the most relevant components of our planet, such as the atmosphere and the ocean, and suggests resources for further exploration.

Graphical abstract



Keywords Ocean · Atmosphere · Climate · Heat · Momentum · Energy · Flow

Foreword

This peer-reviewed paper belongs to the Topical Collection motivated by the Conference "The Mediterranean System: a Hotspot for Climate Change and Adaptation" organized in Rome at the Accademia Nazionale dei Lincei on March 21–22, 2023.

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This paper summarizes the contents of an invited talk, with the same title, given by the author at the conference "The Mediterranean: a hot spot for climate change and adaptation" organized by the Accademia Nazionale dei Lincei in Rome on 20–22 March 2023. The talk was intended for high-school and undergraduate students. Accordingly, this paper is not an academic review on the state of the art, but rather aims at offering the readers an overview of the main interplay between the ocean and the atmosphere, the different mechanisms involved at the various scales, and suggestions for further reading.

1 Introduction

Atmosphere–ocean interactions are a complex and interconnected system that regulates Earth's climate and weather. The exchange of energy, heat, moisture, and gases between the atmosphere and oceans influences precipitation patterns, ocean currents, storm formation, and long-term global climate change. Indeed, the atmosphere and the ocean are two key components of the complex system of our planet. From school studies and personal experience all of us have an idea of what they are and how they evolve. However, it may be worth briefly introducing their properties here, to lay down a common basis for the subsequent reasoning.

1.1 The atmosphere

The atmosphere is the layer, mostly composed of a mixture of gases commonly named air, that surrounds our planet, being retained there by gravity. We say "mostly" because the atmosphere also contains a myriad of particles, either solid or liquid, of different sizes. Solid particles are collectively referred to as particulate matter: they include all kinds of dust, as well as tiny ice bodies, such as the variety of ice crystals composing many cold clouds, and snowflakes. Liquid particles include the constituents of clouds, haze and fog, as well as raindrops. This myriad of particles, of different size and distribution, play key roles in a variety of atmospheric processes—such as nucleation, cloud formation, precipitation dynamics, radiation budgets, and atmospheric energetics. However, for the purpose of the present article, we may neglect these nongaseous constituents, and concentrate on the purely gaseous part of the atmosphere. The composition of the atmosphere is rather homogeneous up to a height of about 100 km above mean sea level (the so-called *homosphere*) and contains approximately 78% nitrogen (N_2), 21% oxygen (O_2), 0.9% argon (Ar), and smaller amounts of other gases, as listed in Table 1.

Table 1 Composition of dry air (Source Wikipedia, s. v., "Atmosphere of Earth")

Dry air		
Gas		Mole fraction
Name	Formula	In %
Nitrogen	N_2	78.084
Oxygen	O_2	20.946
Argon	Ar	0.9340
Carbon dioxide	CO_2	0.0417
Neon	Ne	0.001818
Helium	He	0.000524
Methane	CH_4	0.000187
Krypton	Kr	0.000114

Components listed in Table 1 are almost ubiquitously found in the homosphere with the same fraction indicated therein. Instead, in the upper atmosphere (*heterosphere*) lighter constituents, such as hydrogen (H_2), helium (He) and atomic oxygen (O), dominate. However, the dynamics of these upper levels are quite beyond the scopes of the present paper, as their connections with the oceans are rather limited. Also, within the homosphere, other trace constituents can be found, whose concentration may vary under different situations, being higher in some regions than in others.

This is the case of ozone (O_3), whose concentration is maximum in the *ozonosphere*, a layer between 15 and 35 km above the earth surface, and varies with the seasonal change in incoming solar radiation, as ozone is produced from atmospheric oxygen through the absorption of ultraviolet radiation.

Air also contains a highly variable amount of water vapor. Its variability is closely connected to the frequent changes from liquid water or ice to vapor, and vice versa, that huge amounts of water substance continuously undergo over the earth. This fact is peculiar of the planet Earth, and follows from the range of temperatures, around the freezing/melting point of water, that are commonly found over its surface, as a result of the energy budget of the planet, as discussed below. The frequent and ubiquitous conversions of huge masses of water from one state to another are a key factor for the hydrological cycle, that will be discussed below either.

Thermodynamic properties of the atmosphere, such as air temperature and density, atmospheric pressure and humidity, as well as dynamical variables, such as winds strength and direction, continuously vary in time and space. However, some underlying long-term basic structures may be identified, which constitute the basic state and the general circulation of the atmosphere, respectively.

In particular, air temperature exhibits a peculiar vertical structure (Fig. 1), resulting from the combination of radiation budgets and thermal energy redistribution operated by atmospheric and oceanic circulations (see below). Among the various layers identified in the thermal structure outlined in Fig. 1, the closest to the earth surface is the troposphere. It exhibits an average depth of about 11 km and includes 80% of the atmospheric mass. Practically, all the weather phenomena occur there, as well as surface–atmosphere exchanges, including interactions with the biosphere, e. g., the photosynthesis by terrestrial plants and breathing of terrestrial animals.

1.2 The ocean

The ocean is the body of salt water that covers approximately 70% of our planet. The term ocean is also used to

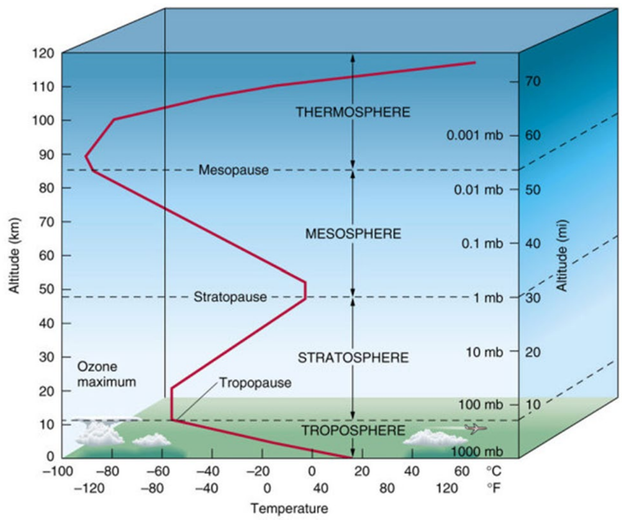

Fig. 1 Average vertical thermal structure of the Earth's atmosphere

refer individually to any of the five large bodies of water into which the worldwide ocean is conventionally divided, i.e., Pacific, Atlantic, Indian, Southern, and Arctic Ocean (Fig. 2). For the purpose of the present paper, we will refer to the ocean as to any large body of salt water, extensively including seas, such as the Mediterranean.

The ocean is the primary component of the Earth's hydrosphere, which also includes freshwater bodies (lakes, rivers, etc.) as well as underground water and water vapor in the atmosphere (Table 2).

Unlike the atmosphere, that covers uniformly all the earth surface, the ocean is constrained to fill some regions above the earth crust, as allowed by the underlying topography (Fig. 3). This determines quite a varied distribution of depth, since the ocean surface level, apart from local oscillations connected to the cycles of tides and the effects of atmospheric pressure and winds, is quite uniformly distributed all over the earth surface by the equalizing effect of gravity. In fact, the *mean sea level* is usually adopted as a reference horizontal level for atmospheric quantities, such as atmospheric pressure. Accordingly, for weather stations based in regions where the earth's surface is above sea level, it is standard observational practice to *reduce* the measured surface pressure to the sea level. This is done by calculating, assuming hydrostatic balance, the pressure that the atmosphere would exhibit directly below the measurement level all the way down to the sea level.

The mean vertical thermal structure of the ocean is simpler than that of the atmosphere, although nontrivial. In particular, the upper part of the ocean is the most exposed to incoming solar radiation, which is absorbed in the upper few hundreds of meters, depending on water absorbance and inclination of the radiation beam over the surface. This heat exchanges produce the *thermocline*, an upper layer, right below the water surface, where the temperature gradient is appreciably greater than the gradients below it. The so-called *permanent thermocline* refers to the layer unaffected by the seasonal and diurnal changes in the surface forcing, whereas the *seasonal thermocline* is the thermocline unaffected by the diurnal changes in the surface forcing, as shown in Fig. 4. In general, the latter is established each year by heating of the surface water in the summer and is reduced or destroyed the following winter by cooling at the surface and wind-driven mixing. As opposed to the above layers, the *diurnal thermocline* refers to the layer that is usually established every day by the heating of the surface water and destroyed the following night by the cooling and/or wind-driven mixing.

Fig. 2 Conventional partitioning given by geographers and scientists to the different areas of the continuous body of water that comprises Earth's Ocean. (Source: https://scied.ucar.edu/learning-zone/earth-system/hydrosphere)

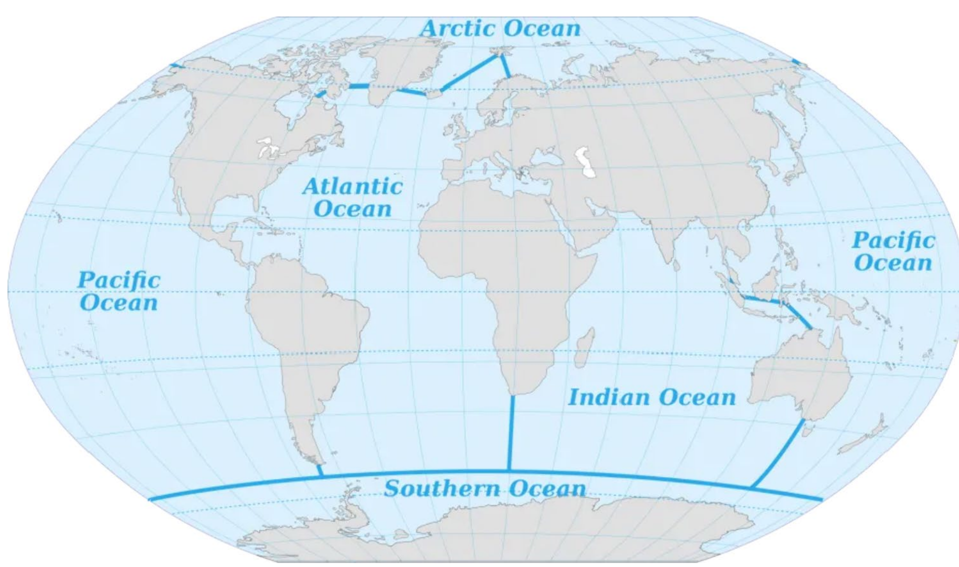

Table 2 Partitioning of water in the various components of the hydrosphere

Water	Quantity (km^3)	Quantity (%)
Total water (pure and saline)		
Oceans	1,348,000,000	97.39
Snow caps, icebergs, glaciers	227,820,000	2.01
Groundwater and soil moisture	8,062,000	0.58
Lakes and rivers	225,000	0.02
Atmosphere	13,000	0.001
Total	*1,584,120,000*	*100.00*

Fig. 3 Map of the oceanic depth, as resulting from the varied topography of the ocean's floor. Colors denote both ocean floor (blues) and land (browns). (Reproduced with permission from Weatherall et al. (2015))

As for many liquids, water mostly expands when it gets warmer,[1] and hence its density decreases. As a consequence of gravity, lighter water tends to rise for buoyancy and to occupy upper levels. Hence the structure outlined in Fig. 5 implies a persistently stable stratification.

The atmosphere and the ocean of course interact, especially through their boundary surfaces: indeed, the ocean covers almost 3/4 of the planet, and its surface is openly exposed to the atmosphere there. This exchange plays a crucial role in regulating the Earth's climate system. Atmosphere–ocean interactions also influence ocean currents and circulation patterns (Huisman et al. 2010). The transfer of energy, heat, and moisture between the atmosphere and the oceans drives the flow of water masses on a global scale. For example, the uneven heating of the Earth's surface by the sun creates temperature and density gradients in the oceans, which may affect the wind-driven large-scale circulation patterns like the Gulf Stream and the Antarctic Circumpolar Current (Huisman et al. 2010).

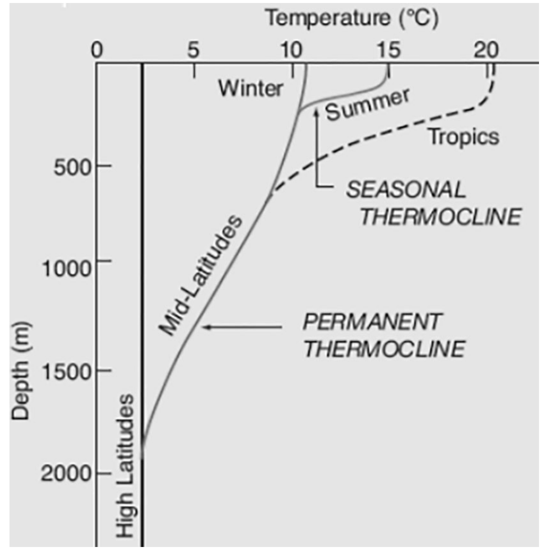

Fig. 4 Mean vertical thermal structure of the upper ocean and typical ranges of its latitudinal and seasonal variations (https://commons.wikimedia.org/wiki/File:ThermoclineSeasonDepth.png)

In addition to these direct interactions, atmosphere–ocean interactions also shape long-term climate patterns. The oceans, with their high heat capacity, act as a buffer, absorbing and storing heat over long periods of time. This helps to regulate global climate by reducing temperature extremes. Changes in atmospheric conditions, such as increased greenhouse gas concentrations, can disrupt this balance and lead to changes in ocean circulation patterns, sea surface temperatures, and climate regimes such as El Niño and La Niña (cf. Trenberth and Branstator 1992).

In the following section, the mechanisms controlling such interactions will be briefly reviewed in more detail.

2 The main factors affecting atmospheric and ocean dynamics

To understand atmospheric and ocean dynamics, it is important to identify the main factors affecting them, and to what extent they do so. Indeed, as we will see below, these dynamics encompass a variety of spatial and time scales. Here we will start from an overview of these two systems as a whole, as part of the planet; hence, we will first focus on the so-called *planetary* scale (a more complete outline of the scales of planetary motions will be provided below).

At a first approximation, our planet can be represented as a sphere with a radius of about 6370 km.[2] With a total mass

[1] A rather peculiar property of the relationship between density and temperature for water will be discussed later.

[2] A better approximation would include the effect of polar flattening, which makes our planet better represented by a rotation ellipsoid with a polar radius of 6359 km and an equatorial radius of 6378 km. How-

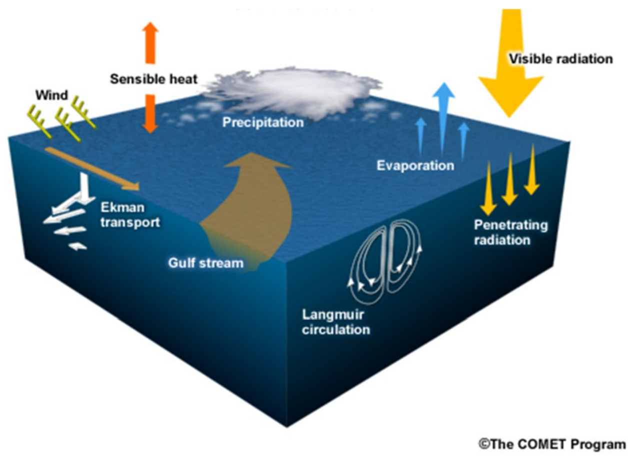

Fig. 5 Schematic of various interactions between the atmosphere and the ocean. (Source: COMET Program). The source of this material is the COMET® Website at http://meted.ucar.edu/ of the University Corporation for Atmospheric Research (UCAR), sponsored in part through cooperative agreement(s) with the National Oceanic and Atmospheric Administration (NOAA), U.S. Department of Commerce (DOC). ©1997-2024 University Corporation for Atmospheric Research. All Rights Reserved

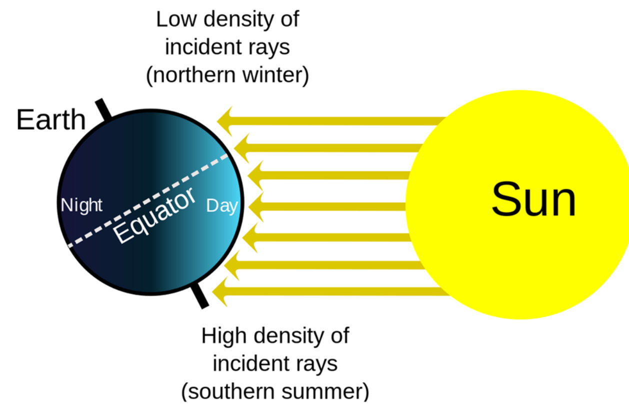

Fig. 6 Effect of latitude in the inclination of sun rays onto the Earth surface

estimated in $M = 5.97\ 10^{24}$kg, mostly due to its denser solid and interior components, the Earth exerts a strong gravitational attraction onto both the waters forming the ocean and the gaseous components of the atmosphere,[3] and keep them quite close to the surface: both atmospheric (Fig. 1) and oceanic depths (Fig. 3) are about 2–3 orders of magnitude smaller than the earth making their layers a rather shallow fluid coating of the planet. As a consequence, the trajectories of water and air parcels composing the planetary flows, in the atmosphere and in the ocean respectively, need to follow the planetary curvature. Another remarkable property of our earth is that it revolves around its axis with a turnover period $T = 86164s$ (sideral day) and, hence, an angular frequency $\Omega = \frac{2\pi}{T} = 7.292\ 10^{-5}$Hz. This rotation has two main consequences for the terrestrial components. First, any reference system fixed with the Earth, as it is common for observers who are based on earth, is a non-inertial one. This means that fictitious forces arise, such as the *centrifugal force* (slightly modifying the pure gravitation attraction into an effective gravity) and the *Coriolis force* acting on every moving parcel of mass m moving at a velocity $\vec{v}$ to deviate it toward a direction orthogonal to its velocity, as described by the formula:

$$\vec{F}_C = -2m\vec{\Omega} \times \vec{v}$$

where $\vec{\Omega}$ is the rotation vector, aligned with the rotation axis and pointing northward, with modulus Ω. Second, the rotation time is short enough to expose all the areas in a zone[4] of comparable latitude to comparable incoming solar radiation in a 24-h time. This explains why the variation of most quantities with longitude is rather weak, and mostly depends on the uneven distribution on continents and ocean. On the other hand, incoming solar radiation strongly depends on latitude, as a consequence of the different angles made by parallel-beam radiation when they hit the curved earth surface at different latitudes[5] (Fig. 6). Also, the tilting by 23.5° above the ecliptic plane of the earth rotation axis makes the seasonal dependence of direct radiation more and more relevant for higher latitudes, with extreme situations in polar areas, which are completely deprived of incoming radiation during the long polar nights in wintertime.

Footnote 2 (continued)

ever, this distinction is not strictly required for the purpose of the present papers, hence it will be neglected hereinafter.

[3] Notice that this attraction is stronger on heavier elements (such as molecular nitrogen N_2 and oxygen O_2) and weaker on lighter constituent (such as hydrogen H_2 and helium He), which explains why the earth lower atmosphere is richer in the former and poorer in the latter, as outlined above.

[4] In geophysics the word *zone* (from the Greek ζώνη, i. e. *belt*) indicates surface bands included between two parallels. Accordingly, the flow direction running along parallels is named *zonal*. Instead, a flow running either poleward or equatorward along the meridians is named *meridional*.

[5] The word *climate* (from the Greek κλίμα, inclination) reminds how climatic zones depend on the angles under which, on average, sun rays hit the earth surface (for a historical review on the origin of the terms *climate* and climatology see Barry 2013).

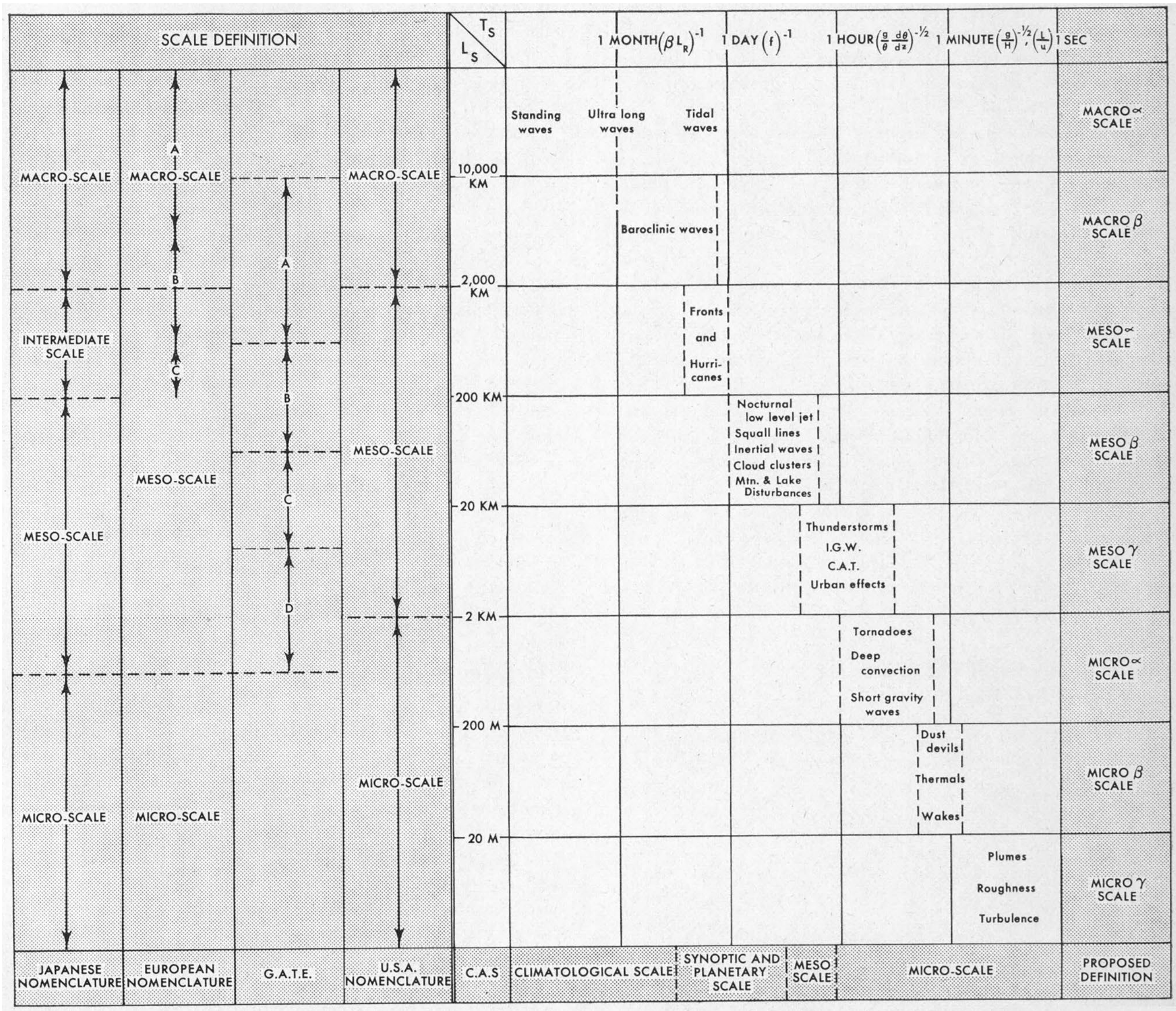

Fig. 7 A rational subdivision of the scales of horizontal atmospheric motions. (Reproduced with permission from Orlanski 1975)

3 Exchanges between atmosphere and ocean

Atmosphere–ocean interactions are a consequence of exchanges between the Earth's atmosphere and the oceans. These exchanges can be explained in terms of fluxes of energy, momentum, and mass of substances (e. g., moisture, gases, etc.) across the ocean surface, and their redistribution within the ocean and within the atmosphere, respectively (Fig. 5). In the following subsections we will briefly review examples of such exchanges.

3.1 Scales of atmospheric and oceanic motions

As most of the exchanges between the ocean and the atmosphere are promoted by, and have consequences on, their motions, it is of utmost importance to remind that motions of large planetary fluid bodies are organized in a hierarchy of spatial and time scales. Figure 7 provides an overview of such scales following the rational subdivision proposed by Orlanski (1975).

In the following, we will mostly concentrate on the planetary and synoptic scale.

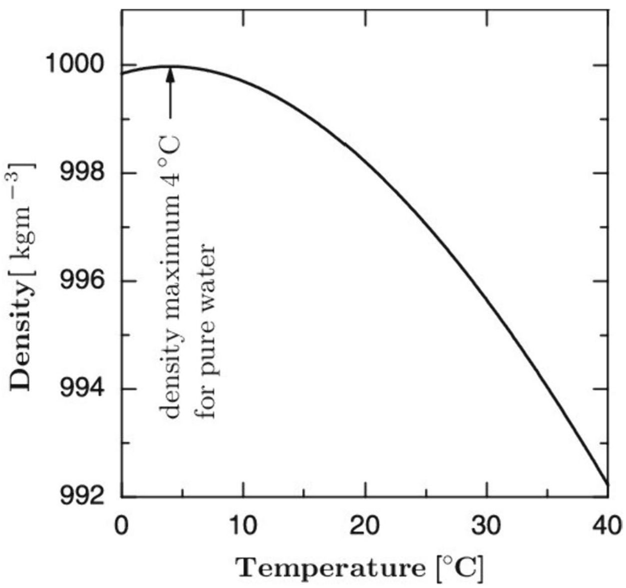

Fig. 8 Changes of water density with temperature. (Reproduced with permission form Hutter et al. 2011)

3.2 Momentum exchanges

The hydrosphere and the atmosphere altogether are commonly referred to as the *fluid earth*, as opposed to the **solid earth**, which includes the earth crust and all the other solid layers beneath, such as the lithosphere. Indeed, oceans are composed of salty water, which can be assumed to a good approximation as an incompressible fluid, whereas the atmosphere is mainly composed of air, i. e., a compressible fluid, whose physical properties are very well amenable to those of perfect gases. Hence both the ocean and the atmosphere are fluid bodies, and as such they exhibit the distinctive property that they cannot withstand any tendency by applied forces to deform them in a way which leaves the shape unchanged: they may resist attempts to deform them, but resistance cannot prevent the deformation from occurring, or, equivalently, the resisting force vanishes with the rate of deformation (Batchelor 1967).

A key factor controlling momentum exchanges in the planetary fluids is their density, i.e., the mass per unit volume. Density controls both the inertia of fluid parcels, and the effect of gravity onto them. As for most of the liquids, the density of water changes with temperature. However, water is somehow peculiar as the temperature increases above the freezing point, rising to a peak at about 4 °C, and then decreases (Fig. 8). This initial increase is unusual, as most liquids only exhibit thermal expansion, i. e., for them density always decreases with increasing temperature. The density of oceanic water also depend on the concentration of salt: since sodium chloride (NaCl) and other salts commonly found in marine water are heavier than water, the higher their concentration, the higher the density of salt water. Changes in temperature and concentration reflects in changes of buoyancy, i. e., the forcing effect of gravity on water masses: this is the reason why circulations controlled by density contrasts, originating from varied combinations of either temperature or salinity, or from both, are also named *thermohaline circulations*.[6]

The density of dry air also depends on temperature: at equal pressure, dry air density decreases with increasing temperature. It also depends on pressure, although to a lesser extent. On the other hand, water vapor also contributes to modifying density of moist air. Water is a species with a molar weight lesser that the effective molar weight of dry air, and this results in moist air being less dense the dry air at equal temperature and pressure. The combination of these variables is well expressed by the perfect gas equation for moist air:

$$\rho = \frac{p}{R_d T(1 + 0.61q)} \tag{1}$$

where ρ is the moist air density, p is pressure, $R_d = 287 \mathrm{JK}^{-1}\mathrm{kg}^{-1}$ is the constant for the state equation of dry air, T is the absolute temperature and q is the specific humidity. Changes of density are crucial for the vertical atmospheric motions associated with atmospheric convection (Saravanan and McCreary 1999).

Conservation of momentum in fluid mechanics is expressed by the Navier–Stokes equations, which is the extension of Newton's second law of dynamics for a pointwise material body to fluid continua. Their full expression and their derivation from basic principles can be found in many textbooks of fluid dynamics (cf. Batchelor 1967) or geophysical fluid mechanics (cf. Gill 1981; Pedlosky 1987; Holton and Hakim 2012; Vallis 2017).

Given the almost spherical shape of our planet, for the geophysical flows it is convenient to write these equations in spherical polar coordinates. The resulting mathematical expressions include a variety of terms, accounting on one side for acceleration, and on the other side for the various forces acting on fluid parcels, either from long-range forces (e.g., gravity), or from short-range interactions with neighboring fluid parcels. A very convenient property of the above equations is that they reproduce adequately well the dynamics of both the atmosphere and the ocean over a surprisingly broad spectrum of scales (cf. Charney and Flierl 1981). However, not all the factors accounted for in those equations have the same importance at all the scales. On the contrary, at each scale of motion the dominating terms are typically fewer, and quite different for the different scales. Concentrating on the dominating terms, and neglecting all

[6] From the Greek words θέρμος i. e. *warm* and ἅλς i. e. *salt*.

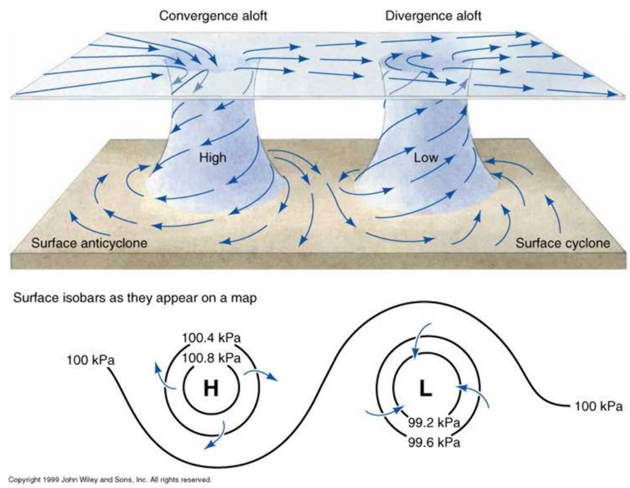

Fig. 9 Airflows associated with the geostrophic winds around pressure highs (H) or lows (L) in the northern hemisphere

the others, allowed much progress in our understanding of key mechanisms behind the most relevant atmospheric and oceanic flows (Charney 1948; Burger 1958).

Also, different factors, and hence different scales, characterize the vertical structure differently from the horizontal ones, either zonal or meridional. Gravity acts on the vertical direction, and it is so strong that the average vertical structure of pressure, both in the ocean and in the atmosphere, is dominated by the hydrostatic balance, i.e., the asset that the fluid mass would exhibit at rest. Even when the atmosphere or the ocean are not at rest, still pressure and density anomalies associated with the flows are mostly connected by hydrostatic balance. Significant departures from the hydrostatic distribution only occur in connection with very strong vertical accelerations, e.g., as a consequence of strong convection, or from topographic effects, such as uplift upstream to a steep mountain range or downdraft downstream to it.

On the other hand, horizontal motions, either in the zonal or in the meridional directions, both at planetary and at synoptic scale, away from the equatorial zone, are dominated by a balance between horizontal pressure gradients ∇p and Coriolis force:

$$-2\vec{\Omega}\times\vec{v}-\frac{1}{\rho}\nabla p\approx 0 \tag{2}$$

This situation is named *geostrophic balance*, and the resulting balanced flow is named *geostrophic flow* (and, more specifically, *geostrophic wind* in the atmosphere, *geostrophic current* in the ocean). This flow is characterized, at any level and at any latitude ϕ (except near the equator, where $\phi\approx 0$) by a horizontal velocity $\vec{v}_G$ given by

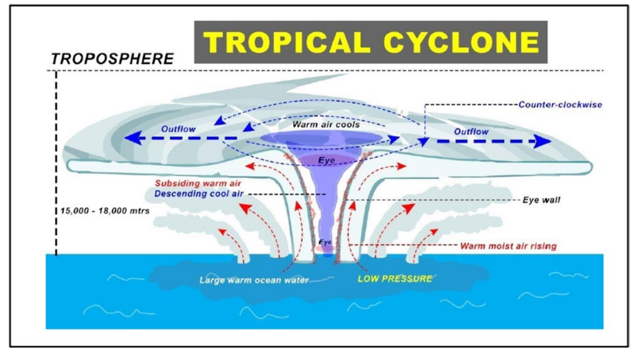

Fig. 10 Schematic of the dynamical and thermal structure of a tropical cyclone

$$\vec{v}_G=\frac{1}{\rho f}\hat{k}\times\nabla p \tag{3}$$

where $f=2\Omega\sin\phi$ is the so-called *Coriolis parameter*, and $\hat{k}$ is the unit vector normal to the earth surface, i. e. indicating the vertical direction. As a result of the vector product in Eq. (3), the geostrophic wind turns out to be always perpendicular to the vertical direction, i. e., purely horizontal, and normal to the pressure gradient, i. e., parallel to the constant-pressure contours (isobars), which can, thus, be adopted as streamlines. Schematic examples of the flows resulting from the above wind field are shown in Fig. 9 for the northern hemisphere. Around high-pressure centers, the flow rotates clockwise (anticyclonic); whereas around low-pressure centers, it rotates counterclockwise (cyclonic). The rotation is reversed in the southern hemisphere.

The Coriolis parameter becomes very small when the latitude angle ϕ approaches zero, i. e., around the equator. This suggests that the geostrophic balance, and hence the approximation of the wind with the geostrophic formula from Eq. (3), are inapplicable at these latitudes. In general, the effects of Coriolis force are nonnegligible whenever the length scale representative of the horizontal extent of the flow, L, and the order of magnitude of the velocity, U, are such that the nondimensional ratio named *Rossby number* $Ro=U/fL$ is significantly smaller than unity. This implies that at midlatitudes the effects of Earth rotation may be significant even for relatively smaller water bodies (e.g., large lakes) when creeping velocities are considered (cf. Amadori et al. 2018).

Apart from intertropical zones, where the geostrophic balance is inapplicable,[7] there is another planetary region where the geostrophic wind is inaccurate in providing the main wind features: this region is the *atmospheric boundary*

[7] Near the equator latitude angles ϕ become very small, and so does $\sin\phi$, making f smaller and smaller, and hence the expression in Eq. (3) singular: this suggests that geostrophic balance does not correctly reproduce the dominant dynamics in the tropical regions, as it does in the extratropical ones.

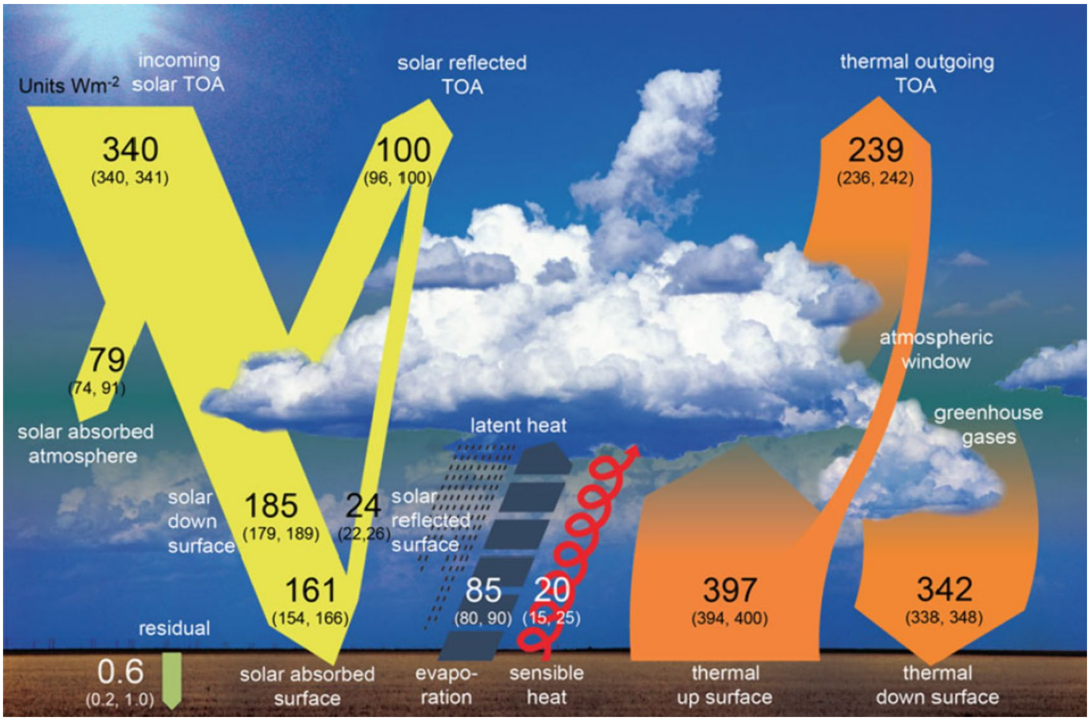

Fig. 11 The annual mean global budget of the earth–atmosphere system. Numbers (in W m^{-2}) indicate best estimates for the magnitudes of the globally averaged energy balance components together with their uncertainty ranges, representing present day climate conditions at the beginning of the twenty first century. (Reproduced with permission from Wild et al. (2013))

layer, i. e. the atmospheric layer closest to the earth surface (either land or ocean) and more strongly interacting with it, responding to surface forcing on timescales of order 1 h or even less. Indeed, in this atmospheric layer closer to the surface the strong interaction with either the terrain or the ocean favors the production of turbulent motions, which enhance exchanges across the surface, including momentum loss to the surface, i.e., enhanced friction. This resistance assumes an order of magnitude comparable to the other two terms in Eq. (3).

As a result, the near-surface wind around a low-pressure center is not any more blowing perfectly parallel to the isobars, but rather exhibits a cross-isobar component, producing a net convergence of the flow toward the center. Hence, as a consequence of the continuity of the flow, deriving from the requirement of mass conservation, a compensating upward vertical motion develops in the column above the center. This upward motion promotes vertical transport of moisture and heat form the surface to the upper levels. This upward flow has two consequences: on one hand, it may fill the column with lighter air,[8] and hence further deepens the low-pressure center; on the other hand, the lifting of moist air promotes condensation and the formation of cloud and precipitations. This mechanism holds both for extratropical (Fig. 9) and for tropical cyclones (Fig. 10), the latter being favored by warmer oceanic surface water.

For this reason, an intensification of tropical cyclone intensity is to be expected under a changing climate (Knutson and Tuleya 2004).

3.3 Energy exchanges

One of the main processes of atmosphere–ocean interactions is the transfer of heat. The oceans exchange heat with the atmosphere through radiation, conduction, and evaporation. This heat transfer is an essential component of the energy budget controlling the Earth's temperature, as the oceans act as a heat sink, absorbing and storing large amounts of heat. Conversely, the oceans also release heat back into the atmosphere, which influences atmospheric circulation patterns (Weller and Anderson 1996).

[8] This is more typical or tropical cyclones, whereas sometimes for extratropical cyclones air at surface level may have a lower enthalpy than air at upper levels.

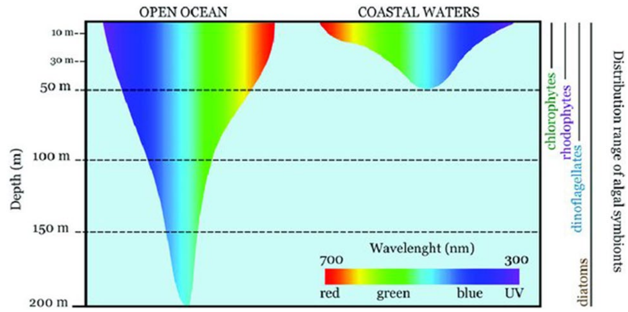

Fig. 12 An example of different penetration of spectral components of shortwave radiation underneath the ocean surface depending on wavelength and water location (open ocean or coastal waters)

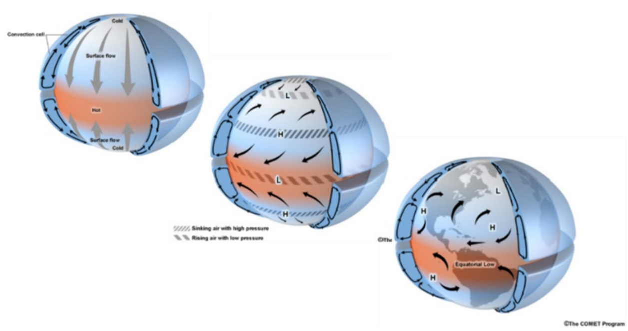

Fig. 13 The general circulation of the atmosphere resulting from the combination of the meridional thermal-convective flow, compensating the uneven heating and cooling of the equatorial vs. the polar zones, and the earth rotation

The primary source of thermal energy is the solar radiation. Apart from long-term fluctuations of the sun activity, the energy emission from the sun is quite regular over the years, and solar radiation reaches the top of the atmosphere with an average irradiance, normal to the beam direction, known a solar constant $S = 1368$ W m^{-2}. Figure 11 shows how this energy input interacts with the atmospheric and terrestrial components and is then partitioned in the overall energy budget of the earth–ocean–atmosphere system.

Atmospheric gases are almost transparent to the incoming shortwave radiation, but clouds, aerosol and reflecting surfaces on the Earth collectively bounce back about 30% of the incoming flux. Most of the heat energy of the sunlight that strikes the Earth is absorbed in the first few millimeters of the soils, which heats up very quickly during daytime, and in turn exchanges heat with the overlying atmosphere. On the other hand, radiation penetrates much deeper below the ocean's surface (Fig. 12), which heats during the day and cools at night as heat is lost to space by radiation. Waves contribute to mixing the water near the surface layer and distribute heat to deeper water such that the temperature may be relatively uniform in the mixed layer, depending on wave strength and on surface turbulence caused by currents. Below this mixed layer, the temperature remains relatively stable over day/night cycles, as shown in Fig. 4.

The diverse response to radiation budgets between water surface and land surface determines significant thermal contrasts among lower atmospheric layer overlying neighboring offshore water and inland surfaces along coastlines. Such contrasts produce density imbalances between neighboring airmasses and hence promote compensating density currents moving airmasses for tens of kilometers across the coastline in the form of sea and land breezes (Rotunno 1983; Simpson 1994). This effect also occurs in extended inland water bodies such as large lakes (Laiti et al. 2014; Giovannini et al 2015).

Values shown in Fig. 11 are globally averaged annual mean, to be considered as reference values: there is a well-known imbalance in the Earth's energy (Trenberth et al. 2014), which is the basis of diverse budgets in different zones and interannual variability. In particular heat is received more in the near equatorial zone, and less in the polar areas. As a consequence of this imbalance, air in the intertropical regions is expected to be warmer, and then less dense, than at the poles. The resulting buoyancy effects are expected to promote a thermal-convective compensating circulation moving air at upper levels form the equator to the poles, and in the opposite direction at lower levels (Fig. 13, left panel). However, observations have shown that this simple idea of compensating flows arranged in a single axial-symmetric cell connecting the equatorial belt with the poles in each hemisphere is not realistic. Rather, the system seems to be better represented with a scheme organized in three annular cells as represented in Fig. 13(central panel). The couple of cells closer to the equator, one in the northern and the other in the southern hemisphere respectively, are usually named *Hadley Cells*.[9] Also, purely meridional flows cannot occur, as motions on the earth surface are also subject to Coriolis force, arising from earth rotation, which deviates the trajectories of moving air parcels toward the right-hand direction of an observer moving with them in the northern hemisphere (while toward the left in the southern). Hence, a meridional component of the flow needs to be accompanied by a zonal one, as shown in Fig. 13 central panel. Indeed, the resulting zonal flows represent the permanently blowing easterly *Trade Winds* in the intertropical areas, and the westerly winds in the midlatitude zones. In fact, the uneven distribution of land and sea adds up further disturbance to this regular picture, producing a scattered distribution of rather persistent high- and low-pressure centers,

[9] After the British mathematician and astronomer John Hadley (1682–1744) who first suggested an explanation of trade winds (Hadley 1735).

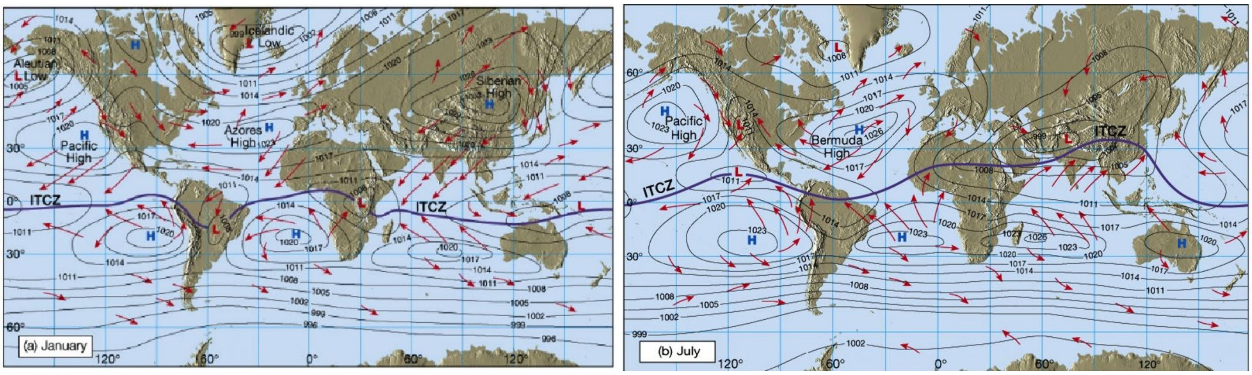

Fig. 14 The general circulation of the atmosphere

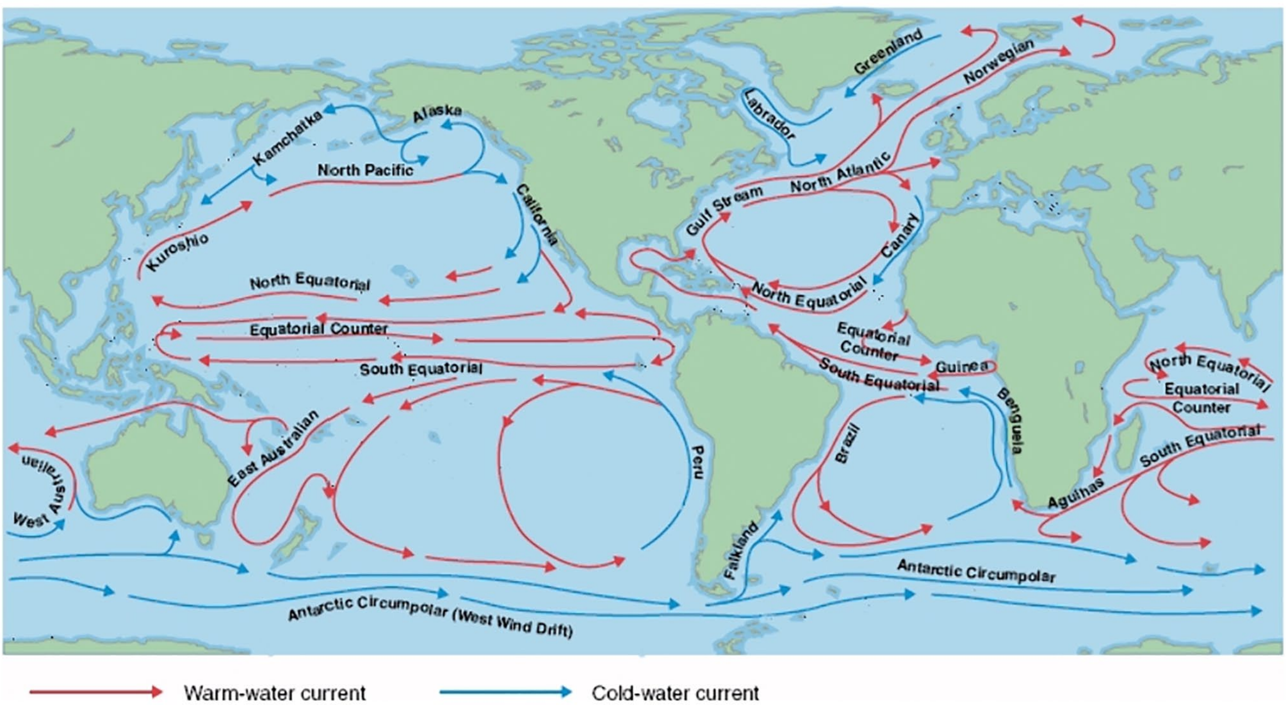

Fig. 15 The general circulation of the ocean

promoting either anticyclonic or cyclonic circulation around them, respectively.

The resulting overall flow constitutes the so-called *general circulation of the atmosphere*: its patterns at mean sea level horizontal surface are well represented by the surface pressure and wind structure outlined in Fig. 14 for the two extreme seasons, summer and winter, respectively.

The wind exerts a drag force on the underlying ocean surface. The force exerted by the wind per unit area in the direction parallel to the surface is named wind stress. The wind stress at the sea surface is the primary driving force for many oceanic phenomena including surface gravity waves, storm surges, and ocean currents. Wind stress τ is normally estimated from the wind speed using the bulk formula in terms of a drag coefficient C_{D}

$$\tau = \rho_a C_{\mathrm{D}} U^2 \quad (2)$$

where ρ_a is the air density, U is the mean wind speed at a reference height above the mean sea level, and C_{D} is the drag coefficient at the same height. The drag coefficient is usually parameterized as a function of the mean wind speed and of the roughness of the sea surface, as determined by the waves (Munk 1955; Charnock 1955; Lin and Sheng 2020). The combination of currents driven by the wind drag results in the surface oceanic circulation patterns shown in Fig. 15.

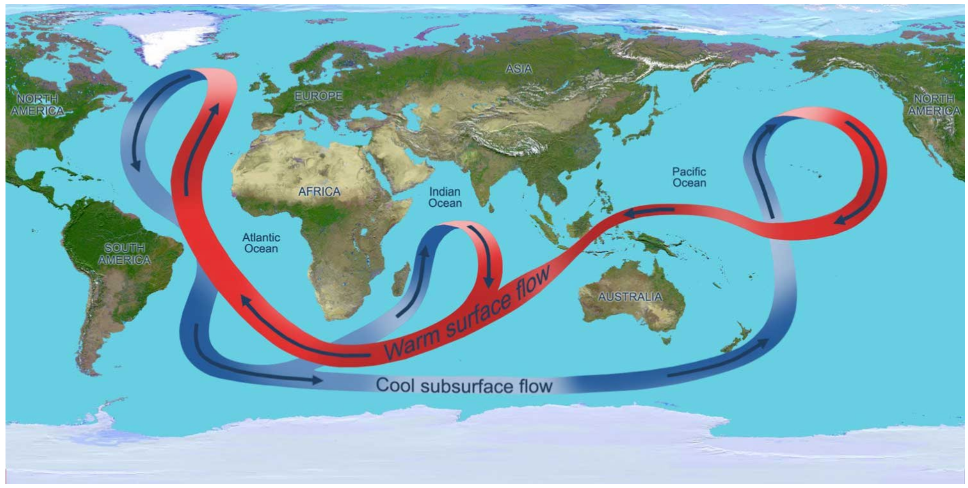

Fig. 16 Overview of the overturning meridional oceanic circulation

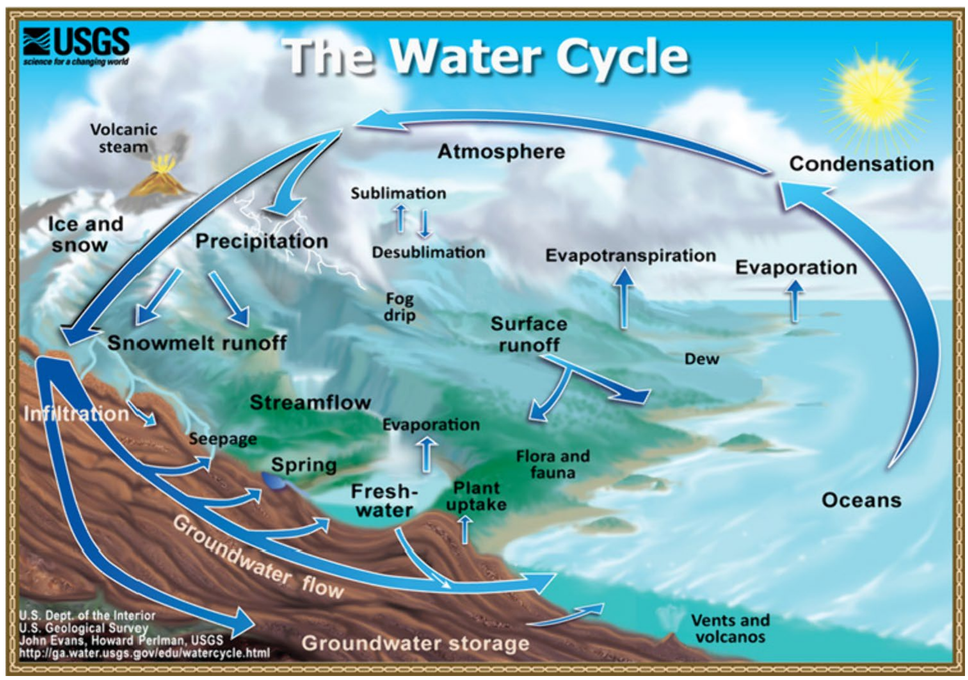

Fig. 17 Schematic of the hydrological cycle

The above surface forcing exerted by the wind combines with the varied oceanic bathymetry and with the changes in water density associated with heating/cooling and increased salt concentration from surface evaporation to produce the so-called general circulation of the ocean. In particular, currents that move up and down in the water column, also called vertical currents, are created by differences in the density of water masses, where heavier waters sink, and lighter waters rise (Fig. 16).

This buoyancy-driven effects combine in the so-called *overturning meridional oceanic circulation* (sometimes also named *thermohaline circulation*), because winds move warm surface waters from the equator toward the poles, where the water cools and increases in density. Some of this water gets so cold that it freezes, leaving its salt behind in the remaining water, thus further increasing the density of this water. This cold, salty water near the poles (primarily in the North Atlantic and near Antarctica) sinks and spreads along the bottom and eventually rises back toward the surface of the ocean. In the absence of such mixing, the circulation would be confined within a very shallow layer, and all the interior of the ocean would become filled with very dense water (Vallis 2017). Notice however that it takes about 1000 years for water to circulate around what is called the global conveyor belt that moves water three dimensionally throughout the world's ocean basins.

These ocean currents can have significant effects on climate and weather patterns. For example, the Gulf Stream

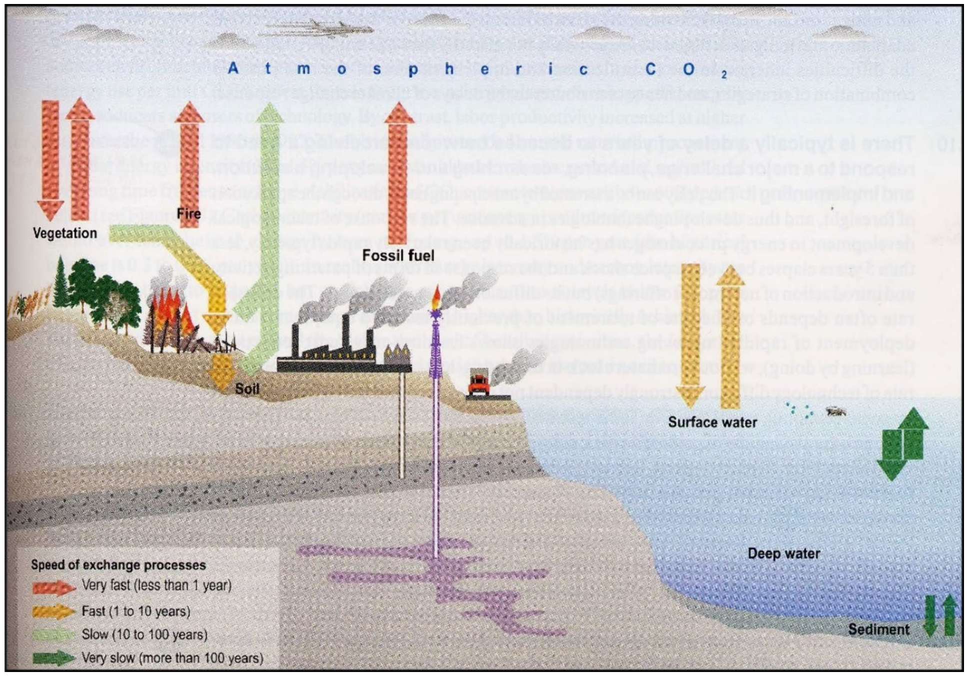

Fig. 18 Schematic of exchanges of carbon among soil, atmosphere, biosphere, ocean, and lithosphere

transports warm water from the tropics to the North Atlantic, which helps to moderate temperatures in regions such as Western Europe. The flow of these currents can also influence the distribution of nutrients, affecting the productivity of marine ecosystems and the distribution of marine species.

Furthermore, atmosphere–ocean interactions play a role in the formation and intensity of tropical cyclones, also known as hurricanes or typhoons (Solomon et al. 2008). These storms develop over warm ocean waters, where the interaction between the atmosphere and the ocean surface provides the necessary energy for their formation and intensification. The heat and moisture transferred from the ocean to the atmosphere fuel the storm, while the storm's strong winds and turbulent winds mix the ocean waters, bringing cooler waters to the surface and potentially weakening the storm.

4 Mass exchanges

Another important aspect of atmosphere–ocean interactions is the exchange of matter. Here, we will focus in particular on the exchange of two key species, namely water and carbon, for the special relevance they have in atmospheric and climate processes.

Evaporation from the ocean's surface adds moisture to the atmosphere, which can then condense and form clouds and precipitation. The amount of moisture in the atmosphere is a key factor in determining weather patterns and can influence the formation of storms. The hydrological, cycle also known as water cycle, is schematically reproduced in Fig. 17.

The oceans also play a role in the exchange of gases, particularly carbon dioxide. The ocean's surface acts as a sink for excess atmospheric carbon dioxide (CO_2) of anthropogenic origin (Canadell et al. 2021), absorbing and storing large amounts of carbon. This process, known as carbon sequestration, helps to regulate the Earth's carbon cycle and mitigate the effects of global warming. Carbon dissolved in oceanic water can be used by microorganism to build up their own shells and skeletons, and the end of their lifetime sinks to the ocean floor and becomes part of the mineral carbon. However, the timescale of these exchanges are quite different, as shown in Fig. 18. In particular, the rate of removal of carbon performed by this process is very slow, excessive carbon dioxide absorption by the oceans can lead to ocean acidification, which negatively impacts marine ecosystems (Doney er al. 2009).

5 Conclusions

Understanding and studying atmosphere–ocean interactions is crucial for protecting our environment, and for predicting and mitigating the impacts of climate change. Scientists use both numerical models and observational data from a variety of sources, including ground based direct and remote sensing devices, as well as satellites, to investigate these interactions and project future scenarios. By improving our understanding of how the atmosphere and oceans interact, we can better anticipate changes in weather patterns, sea level rise, and the overall health of ecosystems (Collins et al. 2013).

Readers who are interested in learning more about these dynamics will find further contents in two comprehensive books, a classical one by Gill (1981), specifically meant to cover both ocean and atmosphere dynamics, and a more recent one by Vallis (2017), a comprehensive overview of dynamical oceanography and meteorology. The mathematical treatment of geophysical flows is explained in depth in the well-known textbook by Pedlosky (1987). Two further textbooks, more focusing on atmospheric processes, are Wallace and Hobbs (2006), which offers an introductory overview on the variety of physical processes involved in the atmosphere, and Holton and Hakim (2012), a classical textbook on dynamical meteorology. Finally, Marshall and Plumb (2007) provides an introductory and comprehensive overview on processes controlling climate and climate change.

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References

Amadori M, Piccolroaz S, Giovannini L, Zardi D, Toffolon M (2018) Wind variability and Earth's rotation as drivers of transport in a deep, elongated subalpine lake: the case of Lake Garda. J Limnol 77(3):505–521. https://doi.org/10.4081/jlimnol.2018.1814

Barry RG (2013) A brief history of the terms climate and climatology. Int J Climatol 33:1317–1320. https://doi.org/10.1002/joc.3504

Batchelor GK (1967) An introduction to fluid dynamics. Cambridge University Press, Cambridge, p 615

Burger AP (1958) Scale consideration of planetary motions of the atmosphere. Tellus 10(2):195–205. https://doi.org/10.3402/tellusa.v10i2.9236p

Canadell JG, Monteiro PMS, Costa MH, Cotrim da Cunha L, Cox PM, Eliseev AV, Henson S, Ishii M, Jaccard S, Koven C, Lohila A, Patra PK, Piao S, Rogelj J, Syampungani S, Zaehle S, Zickfeld K (2021) Global carbon and other biogeochemical cycles and feedbacks. In: Masson-Delmotte V, Zhai P, Pirani A, Connors SL, Péan C, Berger S, Caud N, Chen Y, Goldfarb L, Gomis MI, Huang M, Leitzell K, Lonnoy E, Matthews JBR, Maycock TK, Waterfield T, Yelekçi O, Yu R, Zhou B (eds) Climate change 2021: the physical science basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge and New York, pp 673–816. https://doi.org/10.1017/9781009157896.007

Charney JG (1948) On the scales of atmospheric motions. Geofysiske Publikasjoner 17:3–17

Charney JG, Flierl GR (1981) Oceanic analogues of large-scale atmospheric motions. Evol Phys Oceanogr 504:548

Charnock H (1955) Wind stress on a water surface. QJR Meteorol Soc 81:639–640. https://doi.org/10.1002/qj.49708135027

Collins M, Knutti R, Arblaster JM, Dufresne JL, Fichefet T, Friedlingstein P, Yu Y et al (2013) Long-term climate change: projections, commitments and irreversibility. Climate change 2013: the physical science basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, pp 1029–1136

Doney SC, Fabry VJ, Feely RA, Kleypas JA (2009) Ocean acidification: the other CO_2 problem. Annu Rev Mar Sci 1:169–192

Gill AE (1981) Atmosphere-ocean dynamics. Elsevier Science, Amsterdam, p 662

Giovannini L, Laiti L, Zardi D, de Franceschi M (2015) Climatological characteristics of the Ora del Garda wind in the Alps. Int J Climatol 35:4103–4115. https://doi.org/10.1002/joc.4270/epdf

Hadley G (1735) Concerning the cause of the general trade winds. Phil Trans R Soc 29:58–62

Holton R, Hakim GJ (2012) An introduction to dynamic meteorology, 5th edn. Academic Press, New York

Huisman SE, den Toom M, Dijkstra HA (2010) The climatic importance of the ocean's overturning circulation. Nat Commun 1(7):1–8

Hutter K, Wang Y, Chubarenko IP (2011) Phenomenological coefficients of water. In: Physics of lakes. Advances in geophysical and environmental mechanics and mathematics. Springer, Berlin, Heidelberg. https://doi.org/10.1007/978-3-642-15178-1_10

Knutson TR, Tuleya RE (2004) Impact of CO2-induced warming on simulated hurricane intensity and precipitation: sensitivity to the choice of climate model and convective parameterization. J Clim 17(18):3477–3495

Laiti L, Zardi D, Giovannini L, de Franceschi M, Rampanelli G (2014) Analysis of the diurnal development of a lake-valley circulation in the Alps based on airborne and surface measurements. Atmos Chem Phys 14: 9771–9786. 10.5194/acp-14-9771-2014. http://www.atmos-chem-phys.net/14/9771/2014/acp-14-9771-2014.pdf

Lin S, Sheng J (2020) Revisiting dependences of the drag coefficient at the sea surface on wind speed and sea state. Continent Shelf Res 207:104188. https://doi.org/10.1016/j.csr.2020.104188

Marshall J, Plumb RA (2007) Atmosphere ocean and climate dynamics. Elsevier Science, Amsterdam, p 344

Munk WH (1955) Wind stress on water: an hypothesis. QJR Meteorol Soc 81:320–332. https://doi.org/10.1002/qj.49708134903

Orlanski I (1975) A rational subdivision of scales for atmospheric processes. Bull Am Meteorol Soc 56:527–530

Pedlosky J (1987) Geophysical fluid dynamics. Springer, New York

Rotunno R (1983) On the linear theory of the land and sea breeze. J Atmos Sci 40:1999–2009. https://doi.org/10.1175/1520-0469(1983)040%3c1999:OTLTOT%3e2.0.CO;2

Saravanan R, McCreary JP (1999) Impact of localized convective activity on sea surface temperature and heat fluxes. J Phys Oceanogr 29(10):2503–2522

Simpson JE (1994) Sea breeze and local winds. Cambridge University Press, Cambridge

Solomon A, Michaels PJ, Knappenberger PC (2008) Sea-surface temperatures and tropical cyclones in the Atlantic basin. J Clim 21(13):3471–3482

Trenberth KE, Branstator GW (1992) Effects of El Niño-Southern Oscillation on the winter climate of California. J Clim 5(9):929–948

Trenberth KE, Fasullo JT, Balmaseda MA (2014) Earth's energy imbalance. J Clim 27(9):3129–3144

Vallis G (2017) Atmospheric and oceanic fluid dynamics—fundamentals and large-scale circulation. Cambridge University Press, Cambridge

Wallace JM, Hobbs PV (2006) Atmospheric science: an introductory survey, 2nd edn. Academic Press, New York

Weatherall P, Marks KM, Jakobsson M, Schmitt T, Tani S, Arndt JE, Rovere M, Chayes D, Ferrini V, Wigley R (2015) A new digital bathymetric model of the world's oceans. Earth Space Sci 2:331–345. https://doi.org/10.1002/2015EA000107

Weller RA, Anderson SP (1996) The importance of ocean-atmosphere coupling to variations in the wintertime midlatitude circulation. J Clim 9(5):1098–1120

Wild M, Folini D, Schär C et al (2013) The global energy balance from a surface perspective. Clim Dyn 40:3107–3134. https://doi.org/10.1007/s00382-012-1569-8