Climate change and its implications to biodiversity and environment

General introductory course
L-T-P-C: 2-0-0-2

1. **Course Objectives:**
   The following are the course objectives:
   a. Climate change and its impact on biodiversity and the environment is an introductory course on climate, climate change and its possible impact on biodiversity and the environment.
   b. Students will learn about the basic concept of climate, possible natural and anthropogenic influences on climate change, and global climate models (GCMs), interpretation of GCM outputs.
   c. Students will also get a general awareness of the impact of changing climate on our biodiversity, environment, and projected trends.
   d. At the end of the course, the students will learn the possible adaptation measures for climate change.

2. **Syllabus:**
   Unit – 1 [5 Hours]: Introduction to climate and climate change: weather and climate, important meteorological variables, global warming, possible reasons for global warming, greenhouse gases and human contributions, black carbon and global warming, sources of GHGs and black carbon

   Unit – 2 [5 Hours]: Evidence of climate change: climate since industrial revolution, climate modelling, models and future projections, representative concentration pathways, their importance.
   Hands-on training: Interpretation of global climate model output, QGIS

   Unit – 3 [3 Hours]: Projected future trends and impact: Impact of climate change: global & Indian scenario, surface temperature, precipitation, ocean pH, sea-level, Arctic sea-ice extent.
   ***Tutorials*:** Trend analysis of climate data and its interpretation

   Unit – 4 [5 Hours]: Climate change and biodiversity: biodiversity, importance of biodiversity, pressure on biodiversity from human activities, possible impact, vulnerable species and ecosystems, adaptation, and mitigation.

   Unit – 5 [3 Hours]: Climate change and agriculture: Indian agriculture, impact of climate change on agriculture and models, agricultural policies in context of climate change, initiatives of Government of India for climate change adaptation.

   Unit – 6 [3 Hours]: Climate change and water resources: global and national water budget; outline of impact of climate change on water, climate change-drought & flood, mitigation and adaptation measures.

3. **Course Outcomes:**
   i. Students will understand climate and climate change, anthropogenic influences on global warming, and climate change.

ii. Students will be familiar with global and regional climate models, representative concentration pathways, and their importance.
iii. Students can analyse the climate model projections using QGIS, future trends in climatic variables.
iv. Students can understand biodiversity, its importance, the possible impact of climate change on biodiversity, and adaptation measures.
v. Students can understand the impact of climate change on Indian agriculture and government policies to mitigate the changing climate.
vi. Students will get a general awareness about the water resources of India, the impact of climate change on water, and mitigation measures.

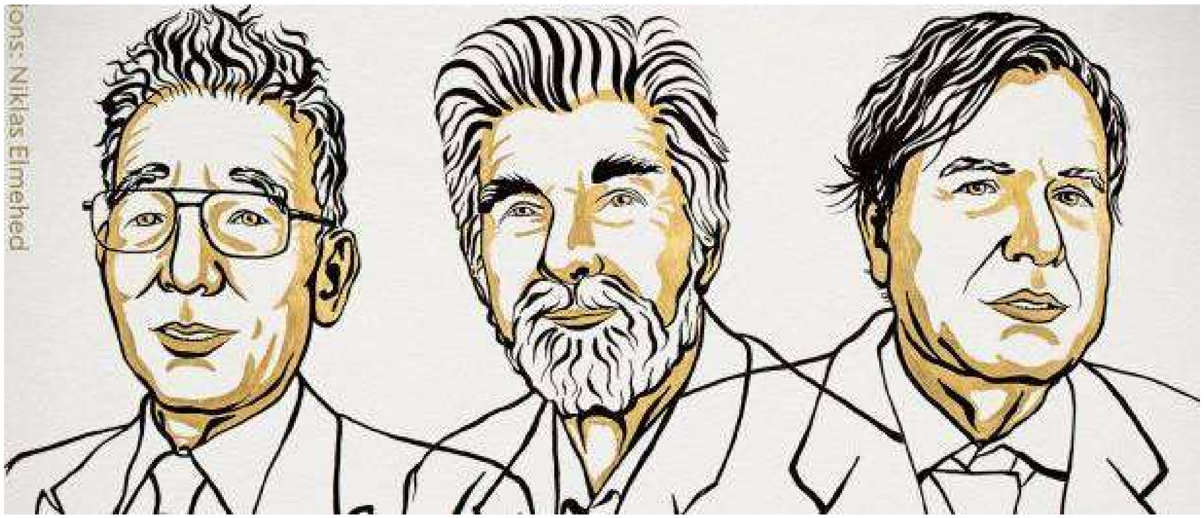
4. **Text Books**: (1-2 text books)
   a) Lawrence M Krauss, The physics of climate change, 2021, Published by Post Hill Press.
   b) Cynthia E. Rosenzweig, Daniel Hillel, Handbook of climate change and agroecosystems: Impacts, adaptation, and mitigation, 2010, ISBN-13 -978-1783265633
5. **Reference Books:** (2-4 text books)
   a) IPCC, 2014: Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Core Writing Team, R.K. Pachauri and L.A. Meyer (eds.)]. IPCC, Geneva, Switzerland, 151 pp
   b) IPCC: Climate change and biodiversity. Technical Paper V, 2002. ISBN:92-9169-104-7
   c) Jan C. Van Dam. Impacts of Climate Change and Climate Variability on Hydrological Regimes, 2003, Cambridge University Press.
   d) IPCC Report Technical Paper VI, 2008, Climate change and water.
   e) ICAR-Policy paper. Climate Change and Indian Agriculture: Impacts, Coping Strategies, Programs and Policy, 2019.

Climate change and its implications

Dr. Raji P

(raji.p@iiits.in)

Lecture-1



Syukuro Manabe Klaus Hasselmann Giorgio Parisi

Course outline: 2 credit

- Introduction to climate & climate change
- Evidence of climate change
- Global climate models & future climate projections
- Implications to

  : Biodiversity, mitigation practices

  : Agriculture, mitigation practices

  : Water resources, mitigation practices

QGIS-Quantum GIS

https://qgis.org/en/site/

Tinn R

Class outline:

Earth system Components

- Atmosphere
- Oceans
- Cryosphere
- Biosphere
- Earth’s crust and mantle

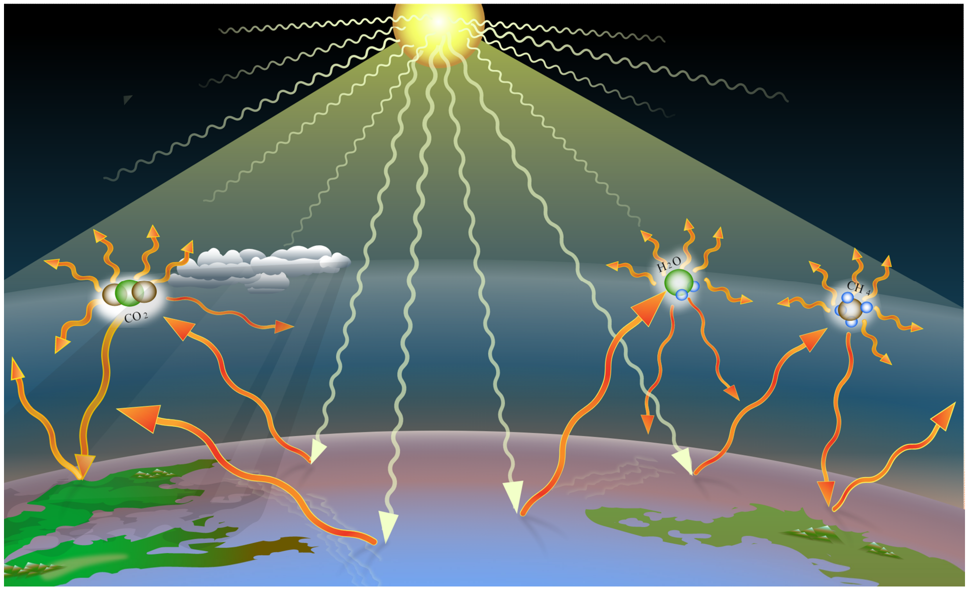
Class outline:

Earth system Components: 1) Atmosphere

Characteristics of Earth's atmosphere

- Optical properties
- Mass
- Composition
- Vertical structure: temperature, general circulations

Optical properties



- Earth's atmosphere is transparent to incoming solar radiation
- The outgoing radiation emitted by Earth is absorbed by the atmosphere (green house effect), and this makes the Earth's atmosphere warm
- About 22% of incoming solar rad is backscattered to space without absorption

Mass of the atmosphere

Mass of the atmosphere

- At any point on the earth's surface, the atmosphere exerts a downward force on the underlying surface due to earth's gravitational attraction.
- The downward force (i.e., the weight) of a unit volume of air with density ρ is given by,

  $F = \rho g$

  Where g is the acceleration due to gravity.
- Integrating the equ from earth's surface to the top of the atmosphere, we obtain the atmospheric pressure on the earth's surface (Ps) due to the weight (per unit area) of the air in the overlying column.

  i.e., $P_s = \int_0^\infty \rho g \, dz$
- Neglect the small variation of g with lat, long, and height, we can take the mean value g, which is equal to 9.807 m^2/s, we can take it outside the integral.

Then Ps= g $\int_0^{\infty} \rho \, dz$,

which is again equal to = g m

where m is the vertically integrated **mass per unit area of the overlying air**

Exercise 1

- The globally averaged surface pressure is 985 hPa. Estimate the mass of the atmosphere.

Earth's radius=$6.37x10^6$ m

Chemical composition

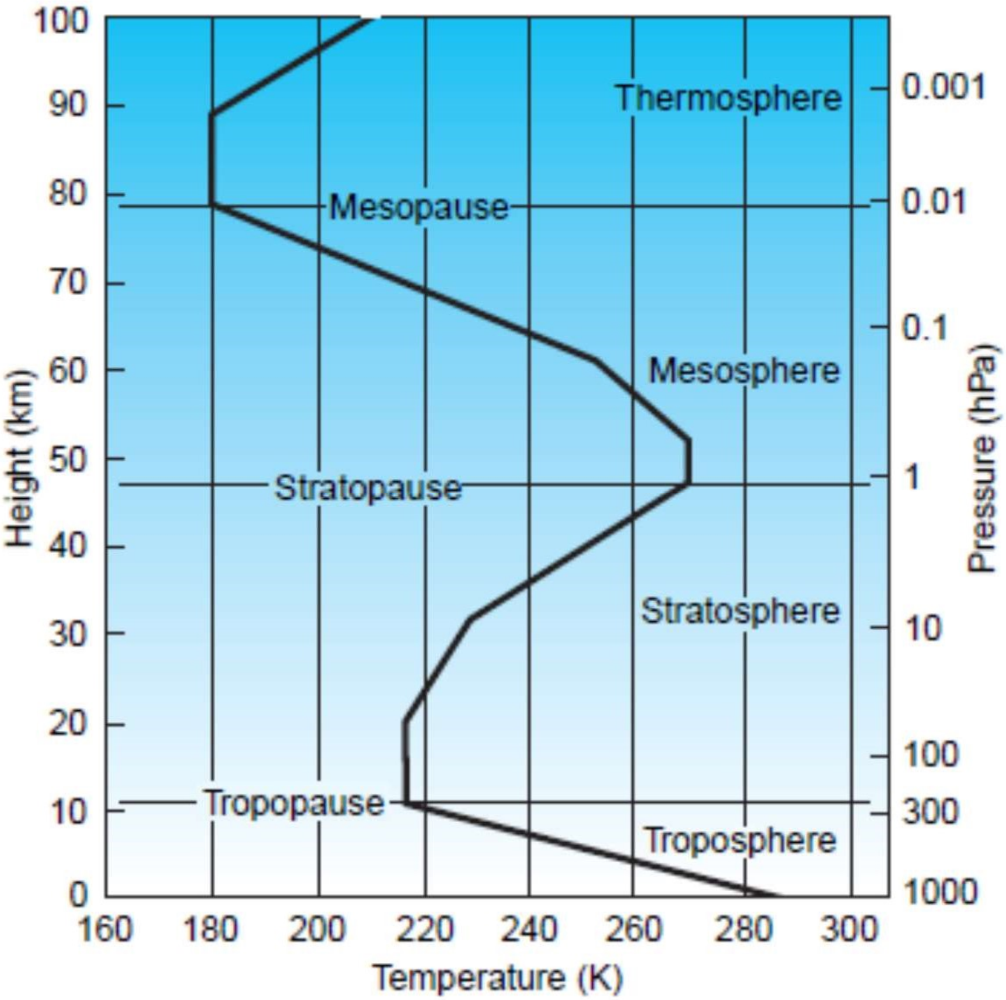
Constituent[a]	Molecular weight	Fractional concentration by volume
Nitrogen (N_2)	28.013	78.08%
Oxygen (O_2)	32.000	20.95%
Argon (Ar)	39.95	0.93%
Water vapor (H_2O)	18.02	0–5%
Carbon dioxide (CO_2)	44.01	380 ppm
Neon (Ne)	20.18	18 ppm
Helium (He)	4.00	5 ppm
Methane (CH_4)	16.04	1.75 ppm
Krypton (Kr)	83.80	1 ppm
Hydrogen (H_2)	2.02	0.5 ppm
Nitrous oxide (N_2O)	56.03	0.3 ppm
Ozone (O_3)	48.00	0–0.1 ppm

- Water vapour accounts for 0.25% of the mass of the atmosphere
- Exposure to ozone concentration>0.1 ppmv is considered hazardous to human health
- Gas molecules with certain structures are highly effective in trapping outgoing radiation and are called green house gases

eg. CH4, N_2O, CO, and chloroflurocarbons (CFCs) – enter into the atmosphere via burning of plant matter, fossil fuels, emission from plants, decay of plants and animals etc.

Vertical structure of atmosphere

Vertical structure



- Tropo-(turning or changing/vertical mixing) -sphere
- Temperature decreases with height (~6.5˚C)
- Troposheric air accounts for the 80% of the mass of the atmosphere

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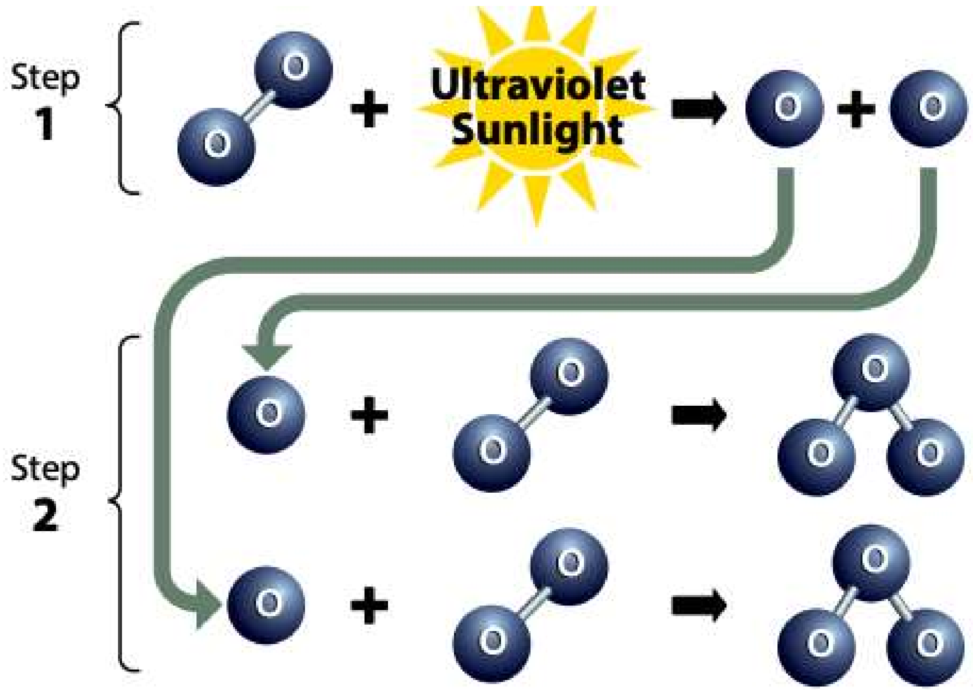
- Strato-(layered)-sphere, vertical mixing is prohibited due to the increase of temp with height
- Residence time of particles are longer
- Air is extremely dry and ozone rich
- They absorb the UV from the spectrum
- This increases the temperature

- Meso-(inbetween)-sphere: temperature decreases with height

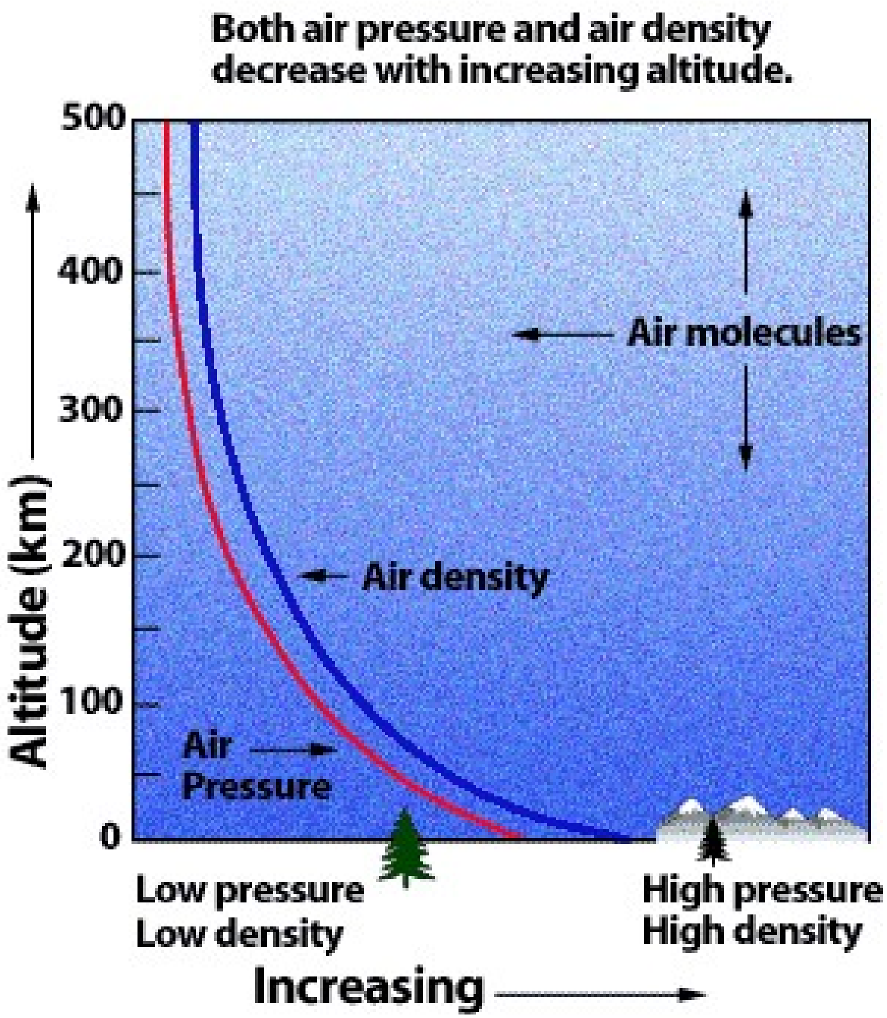
- Thermosphere-temp increases with height due to the absorption of solar rad, and lots of ionization processes occurs



Ozone formation in stratosphere

Anvil cloud (Cloud flattening) at the tropopause

- All the weather and climate activities are under the tropopause



Density of air @ sea level is 1.25 kg/m^3

Pressure at any height:

$$\boxed{p \simeq p_0 e^{-z/H}}$$

P_0 is the pressure @ sea level (reference level)

H- scale height; e-folding depth (height at which pressure becomes 1/e times p_0), 7 to 8 Km

$$\boxed{\ln \frac{p}{p_0} \simeq -\frac{z}{H}}$$

$$\boxed{z = H \ln(p_0/p)}$$

Introduction and Overview

1

1.1 Scope of the Subject and Recent Highlights

Atmospheric science is a relatively new, applied discipline that is concerned with the structure and evolution of the planetary atmospheres and with the wide range of phenomena that occur within them. To the extent that it focuses mainly on the Earth's atmosphere, atmospheric science can be regarded as one of the *Earth* or *geo*sciences, each of which represents a particular fusion of elements of physics, chemistry, and fluid dynamics.

The historical development of atmospheric sciences, particularly during the 20th century, has been driven by the need for more accurate weather forecasts. In popular usage the term "meteorologist," a synonym for atmospheric scientist, means "weather forecaster." During the past century, weather forecasting has evolved from an art that relied solely on experience and intuition into a science that relies on numerical models based on the conservation of mass, momentum, and energy. The increasing sophistication of the models has led to dramatic improvements in forecast skill, as documented in Fig. 1.1. Today's weather forecasts address not only the deterministic, day-to-day evolution of weather patterns over the course of the next week or two, but also the likelihood of hazardous weather events (e.g., severe thunderstorms, freezing rain) on an hour-by-hour basis (so called "nowcasting"), and departures of the climate (i.e., the statistics of weather) from seasonally adjusted normal values out to a year in advance.

Weather forecasting has provided not only the intellectual motivation for the development of atmospheric science, but also much of the infrastructure. What began in the late 19th century as an assemblage of regional collection centers for real time teletype transmissions of observations of surface weather variables has evolved into a sophisticated *observing system* in which satellite and in situ measurements of many surface and upper air variables are merged (or *assimilated*) in a dynamically consistent way to produce optimal estimates of their respective three-dimensional fields over the entire globe. This global, real time atmospheric dataset is the envy of oceanographers and other geo- and planetary scientists: it represents both an extraordinary technological achievement and an exemplar of the benefits that can derive from international cooperation. Today's global weather observing system is a vital component of a broader Earth observing system, which supports a wide variety of scientific endeavors, including climate monitoring and studies of ecosystems on a global scale.

A newer, but increasingly important organizing theme in atmospheric science is *atmospheric chemistry*. A generation ago, the principal focus of this field was urban air quality. The field experienced

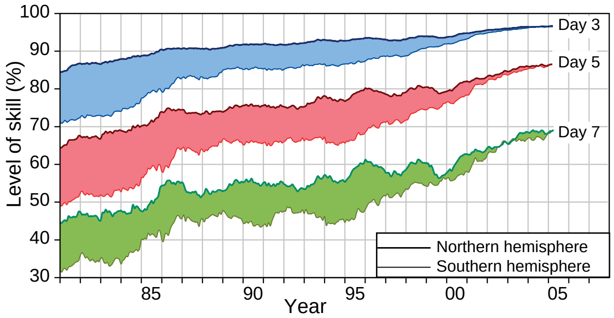


Fig. 1.1 Improvement of forecast skill with time from 1981 to 2003. The ordinate is a measure of forecast skill, where 100% represents a perfect forecast of the hemispheric flow pattern at the 5-km level. The upper pair of curves is for 3-day forecasts, the middle pair for 5-day forecasts, and the lower pair for 7-day forecasts. In each pair, the upper curve that marks the top of the band of shading represents the skill averaged over the northern hemisphere and the lower curve represents the skill averaged over the southern hemisphere. Note the continually improving skill levels (e.g., today's 5-day forecasts of the northern hemisphere flow pattern are nearly as skillful as the 3-day forecasts of 20 years ago). The more rapid increase in skill in the southern hemisphere reflects the progress that has been made in assimilating satellite data into the forecast models. [Updated from *Quart. J. Royal Met. Soc.*, **128**, p. 652 (2002). Courtesy of the European Centre for Medium-Range Weather Forecasting.]

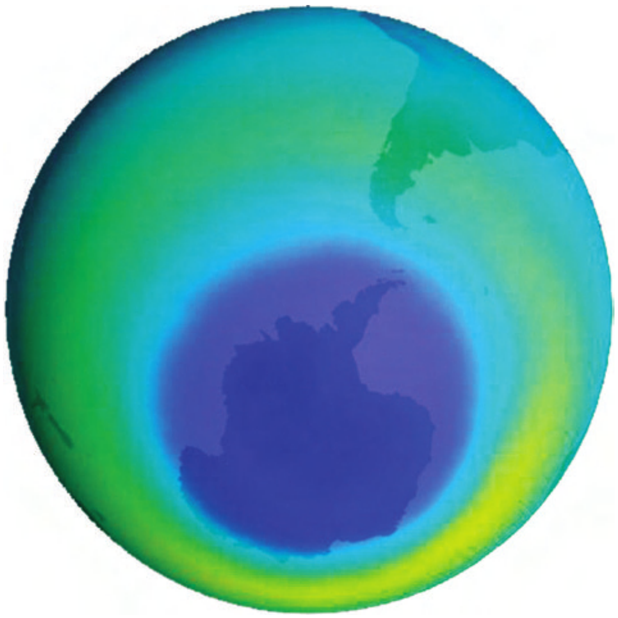


Fig. 1.2 The Antarctic ozone hole induced by the buildup of synthetic chlorofluorocarbons, as reflected in the distribution of vertically integrated ozone over high latitudes of the southern hemisphere in September, 2000. Blue shading represents substantially reduced values of total ozone relative to the surrounding region rendered in green and yellow. [Based on data from NASA TOMS Science Team; figure produced by NASA's Scientific Visualization Studio.]

a renaissance during the 1970s when it was discovered that forests and organisms living in lakes over parts of northern Europe, the northeastern United States, and eastern Canada were being harmed by *acid rain* caused by sulfur dioxide emissions from coal-fired electric power plants located hundreds and, in some cases, thousands of kilometers upwind. The sources of the acidity are gaseous oxides of sulfur and nitrogen (SO_2, NO, NO_2, and N_2O_5) that dissolve in microscopic cloud droplets to form weak solutions of sulfuric and nitric acids that may reach the ground as raindrops.

There is also mounting evidence of the influence of human activity on the composition of the global atmosphere. A major discovery of the 1980s was the *Antarctic "ozone hole"*: the disappearance of much of the stratospheric ozone layer over the southern polar cap each spring (Fig. 1.2). The ozone destruction was found to be caused by the breakdown of chlorofluorocarbons (CFCs), a family of synthetic gases that was becoming increasingly widely used for refrigeration and various industrial purposes. As in the acid rain problem, heterogeneous chemical reactions involving cloud droplets were implicated, but in the case of the "ozone hole" they were taking place in wispy polar stratospheric clouds. Knowledge gained from atmospheric chemistry research has been instrumental in the design of policies to control and ultimately reverse the spread of acid rain and the ozone hole. The unresolved scientific issues surrounding *greenhouse warming* caused by the buildup of carbon dioxide (Fig. 1.3) and other trace gases released into the atmosphere by human activities pose a new challenge for atmospheric chemistry and for the broader field of geochemistry.

Atmospheric science also encompasses the emerging field of *climate dynamics*. As recently as a generation ago, climatic change was viewed by most atmospheric scientists as occurring on such long timescales that, for most purposes, today's climate could be described in terms of a standard set of

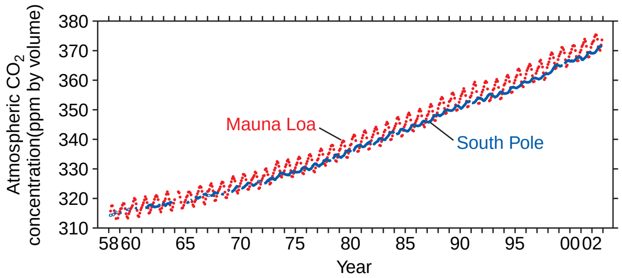


Fig. 1.3 Time series showing the upward trend in monthly mean atmospheric CO_2 concentrations (in parts per million by volume) at Mauna Loa and the South Pole due to the burning of fossil fuels. A pronounced annual cycle is also evident at Mauna Loa, with minimum values in the summer. [Based on data of C. D. Keeling. Courtesy of Todd P. Mitchell.]

statistics, such as January climatological-mean (or "normal") temperature. In effect, climatology and climate change were considered to be separate subfields, the former a branch of atmospheric sciences and the latter largely the province of disciplines such as geology, paleobotany, and geochemistry. Among the factors that have contributed to the emergence of a more holistic, dynamic view of climate are:

- documentation of a coherent pattern of year-to-year climate variations over large areas of the globe that occurs in association with El Niño (Section 10.2).
- proxy evidence, based on a variety of sources (ocean sediment cores and ice cores, in particular), indicating that large, spatially coherent climatic changes have occurred on time scales of a century or even less (Section 2.6.4).
- the rise of the global-mean surface air temperature during the 20th century and projections of a larger rise during the 21st century due to human activities (Section 10.4).

Like some aspects of atmospheric chemistry, climate dynamics is inherently multidisciplinary: to understand the nature and causes of climate variability, the atmosphere must be treated as a component of the *Earth system*.

1.2 Some Definitions and Terms of Reference

Even though the Earth is not perfectly spherical, atmospheric phenomena are adequately represented in terms of a spherical coordinate system, rotating with the Earth, as illustrated in Fig. 1.4. The coordinates are latitude ϕ, longitude λ, and height z above sea level, z.[1] The angles are often replaced by the distances

$$dx \equiv r\, d\lambda \cos\phi \tag{1.1}$$

and

$$dy \equiv r\, d\phi$$

where x is distance east of the Greenwich meridian along a latitude circle, y is distance north of the

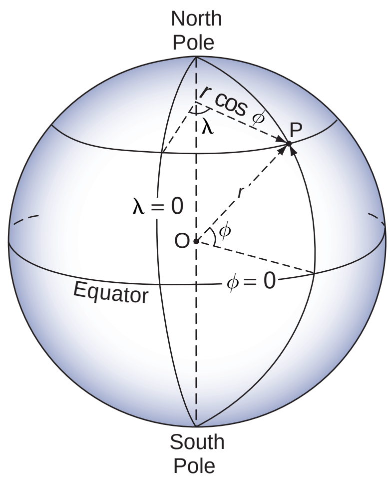


Fig. 1.4 Coordinate system used in atmospheric science. Angle ϕ is latitude, defined as positive in the northern hemisphere and negative in the southern hemisphere, and λ is longitude relative to the Greenwich meridian, positive eastward. The radial coordinate (not shown) is height above sea level.

[1] Oceanographers and applied mathematicians often use the colatitude $\theta = \pi/2 - \phi$ instead of ϕ.

Fig. 1.5 The limb of the Earth, as viewed from space in visible satellite imagery. The white layer is mainly light scattered from atmospheric aerosols and the overlying blue layer is mainly light scattered by air molecules. [NASA Gemini-4 photo. Photograph courtesy of NASA.]

equator, and r is the distance from the center of the Earth. At the Earth's surface a degree of latitude is equivalent to 111 km (or 60 nautical miles). Because 99.9% of the mass of the atmosphere is concentrated within the lowest 50 km, a layer with a thickness less than 1% of the radius of the Earth, r, is nearly always replaced by the mean radius of the Earth (6.37×10^6 m), which we denote by the symbol R_E. Images of the limb of the Earth (Fig. 1.5) emphasize how thin the atmosphere really is.

The three velocity components used in describing atmospheric motions are defined as

$$u \equiv \frac{dx}{dt} = R_E \cos\phi \frac{d\lambda}{dt} \quad \text{(the } \textit{zonal velocity component}\text{)} \tag{1.2}$$

$$v \equiv \frac{dy}{dt} = R_E \frac{d\phi}{dt} \quad \text{(the } \textit{meridional velocity component}\text{)},$$

and

$$w \equiv \frac{dz}{d} = \frac{dr}{dt} \qquad \text{(the } \textit{vertical velocity component}\text{)}.$$

where z is height above mean sea level. The adjectives *zonal* and *meridional* are also commonly used in reference to averages, gradients, and cross sections. For example, a *zonal average* denotes an average around latitude circles; a *meridional cross section* denotes a north–south slice through the atmosphere. The *horizontal velocity vector* $\mathbf{V}$ is given by $\mathbf{V} \equiv u\mathbf{i} + v\mathbf{j}$, where $\mathbf{i}$ and $\mathbf{j}$ are the unit vectors in the zonal and meridional directions, respectively. Positive and negative zonal velocities are referred to as *westerly* (from the west) and *easterly* (from the east) winds, respectively; positive and negative meridional velocities are referred to as *southerly* and *northerly* winds (in both northern and southern hemispheres, respectively.[2] For scales of motion in the Earth's atmosphere in excess of 100 km, the length scale greatly exceeds the depth scale, and typical magnitudes of the horizontal velocity component $\mathbf{V}$ exceed those of the vertical velocity component w by several orders of magnitude. For these scales the term *wind* is synonymous with *horizontal velocity component*. The SI unit for velocity (or speed) is m s^{-1}. One meter per second is equivalent to 1.95 knots (1 knot = 1 nautical mile per hour). Vertical velocities in large-scale atmospheric motions are often expressed in units of cm s^{-1}: 1 cm s^{-1} is roughly equivalent to a vertical displacement of 1 kilometer per day.

Throughout this book, the local derivative $\partial/\partial t$ refers to the rate of change at a fixed point in rotating (x, y, z) space and the total time derivative d/dt refers to the rate of change following an air parcel as it moves along its three-dimensional trajectory through the atmosphere. These so-called *Eulerian*[3]

[2] Dictionaries offer contradictory definitions of these terms, derived from different traditions.

[3] **Leonhard Euler** (1707–1783) Swiss mathematician. Held appointments at the St. Petersburg Academy of Sciences and the Berlin Academy. Introduced the mathematical symbols e, i, and $f(x)$. Made fundamental contributions in optics, mechanics, electricity, and magnetism, differential equations, and number theory. First to describe motions in a rotating coordinate system. Continued to work productively after losing his sight by virtue of his extraordinary memory.

and *Lagrangian*[4] rates of change are related by the chain rule

$$\frac{d}{dt} = \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y} + w\frac{\partial}{\partial z}$$

which can be rewritten in the form

$$\frac{\partial}{\partial t} = \frac{d}{dt} - u\frac{\partial}{\partial x} - v\frac{\partial}{\partial y} - w\frac{\partial}{\partial z} \tag{1.3}$$

The terms involving velocities in Eq. (1.3), including the minus signs in front of them, are referred to as *advection terms*. At a fixed point in space the Eulerian and Lagrangian rates of change of a variable ψ differ by virtue of the advection of air from upstream, which carries with it higher or lower values of ψ. For a hypothetical *conservative tracer*, the Lagrangian rate of change is identically equal to zero, and the Eulerian rate of change is

$$\frac{\partial}{\partial t} = -u\frac{\partial}{\partial x} - v\frac{\partial}{\partial y} - w\frac{\partial}{\partial z}$$

The fundamental thermodynamic variables are pressure p, density ρ, and temperature T. The SI unit of pressure is $1\ \mathrm{N\ m^{-2}} = 1\ \mathrm{kg\ m^{-1}\ s^{-2}} = 1$ pascal (Pa). Prior to the adoption of SI units, atmospheric pressure was expressed in millibars (mb), where $1\ \mathrm{bar} = 10^6\ \mathrm{g\ cm^{-1}\ s^{-2}} = 10^6$ dynes. In the interests of retaining the numerical values of pressure that atmospheric scientists and the public have become accustomed to, atmospheric pressure is usually expressed in units of hundreds of (i.e., hecto) pascals (hPa).[5] Density is expressed in units of $\mathrm{kg\ m^{-3}}$ and temperature in units of °C or K, depending on the context, with °C for temperature differences and K for the values of temperature itself. Energy is expressed in units of joules ($\mathrm{J} = \mathrm{kg\ m^2\ s^{-2}}$).

Atmospheric phenomena with timescales shorter than a few weeks, which corresponds to the theoretical limit of the range of deterministic (day-by-day) weather forecasting, are usually regarded as relating to *weather*, and phenomena on longer timescales as relating to *climate*. Hence, the adage (intended to apply to events a month or more in the future): "Climate is what you expect; weather is what you get." Atmospheric variability on timescales of months or longer is referred to as *climate variability*, and statistics relating to conditions in a typical (as opposed to a particular) season or year are referred to as *climatological-mean* statistics.

1.1 Atmospheric Predictability and Chaos

Atmospheric motions are inherently unpredictable as an initial value problem (i.e., as a system of equations integrated forward in time from specified initial conditions) beyond a few weeks. Beyond that time frame, uncertainties in the forecasts, no matter how small they might be in the initial conditions, become as large as the observed variations in atmospheric flow patterns. Such exquisite *sensitivity to initial conditions* is characteristic of a broad class of mathematical models of real phenomena, referred to as *chaotic nonlinear systems*. In fact, it was the growth of errors in a highly simplified weather forecast model that provided one of the most lucid early demonstrations of this type of behavior.

In 1960, Professor Edward N. Lorenz in the Department of Meteorology at MIT decided to rerun an experiment with a simplified atmospheric model in order to extend his "weather forecast" farther out into the future. To his surprise, he found that he was unable to duplicate his previous forecast. Even though the code and the prescribed initial conditions in the two experiments were identical, the states of the model in the two fore-

Continued on next page

[4] **Joseph Lagrange** (1736–1813) French mathematician and mathematical physicist. Served as director of the Berlin Academy, succeeding Euler in that role. Developed the calculus of variations and also made important contributions to differential equations and number theory. Reputed to have told his students "Read Euler, read Euler, he is our master in everything."

[5] Although the pressure will usually be expressed in hectopascals (hPa) in the text, it should be converted to pascals (Pa) when working quantitative exercises that involve a mix of units.

1.1 Continued

casts diverged, over the course of the first few hundred time steps, to the point that they were no more like one another than randomly chosen states in experiments started from entirely different initial conditions. Lorenz eventually discovered that the computer he was using was introducing round-off errors in the last significant digit that were different each time he ran the experiment. Differences between the "weather patterns" in the different runs were virtually indistinguishable at first, but they grew with each time step until they eventually became as large as the range of variations in the individual model runs.

Lorenz's model exhibited another distinctive and quite unexpected form of behavior. For long periods of (simulated) time it would oscillate around some "climatological-mean" state. Then, for no apparent reason, the state of the model would undergo an abrupt "regime shift" and begin to oscillate around another quite different state, as illustrated in Fig. 1.6. Lorenz's model exhibited two such preferred "climate regimes." When the state of the model resided within one of these regimes, the "weather" exhibited quasi-periodic oscillations and consequently was predictable quite far into the future. However, the shifts between regimes were abrupt, irregular, and inherently unpredictable beyond a few simulated days. Lorenz referred to the two climates in the model as *attractors*.

The behavior of the real atmosphere is much more complicated than that of the highly simplified model used by Lorenz in his experiments. Whether the Earth's climate exhibits such regime-like behavior, with multiple "attractors," or whether it should be viewed as varying about a single state that varies in time in response to solar, orbital, volcanic, and anthropogenic forcing is a matter of ongoing debate.

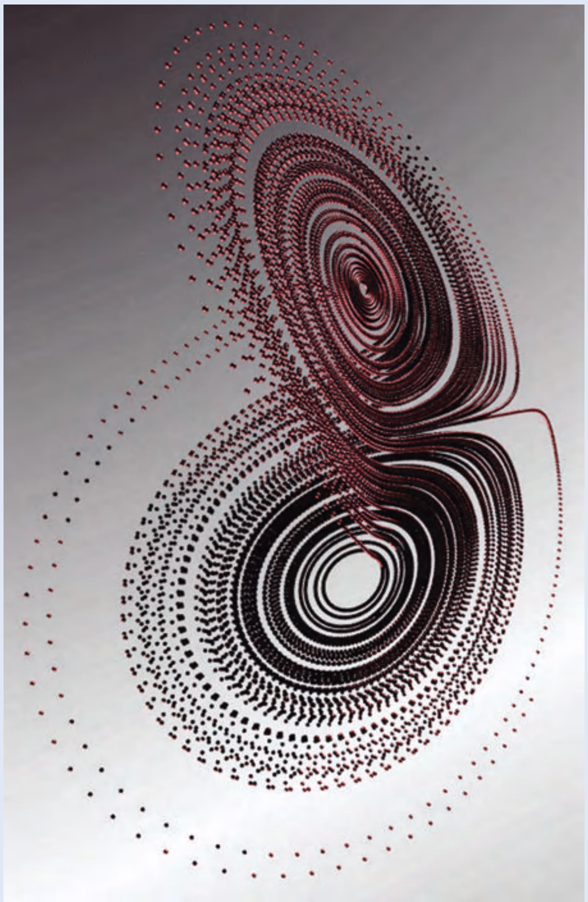


Fig. 1.6 The history of the state of the model used by Lorenz can be represented as a trajectory in a three-dimensional space defined by the amplitudes of the model's three dependent variables. Regime-like behavior is clearly apparent in this rendition. Oscillations around the two different "climate attractors" correspond to the two, distinctly different sets of spirals, which lie in two different planes in the three-dimensional phase space. Transitions between the two regimes occur relatively infrequently. [Permission to use figure from *Nature*, **406**, p. 949 (2000). © Copyright 2000 Nature Publishing Group. Courtesy of Paul Bourke.]

1.3 A Brief Survey of the Atmosphere

The remainder of this chapter provides an overview of the optical properties, composition, and vertical structure of the Earth's atmosphere, the major wind systems, and the climatological-mean distribution of precipitation. It introduces some of the terminology that will be used in subsequent chapters and some of the conventions that will be used in performing calculations involving amounts of mass and rates of movement.

1.3.1 Optical Properties

The Earth's atmosphere is relatively transparent to incoming solar radiation and opaque to outgoing radiation emitted by the Earth's surface. The blocking of outgoing radiation by the atmosphere, popularly referred to as the *greenhouse effect*, keeps the surface of the Earth warmer than it would be in the absence of an atmosphere. Much of the absorption and reemission of outgoing radiation are due to air molecules, but cloud droplets also play a significant role. The radiation emitted to space by air molecules and cloud droplets provides a basis for remote sensing of the three-dimensional distribution of temperature and various atmospheric constituents using satellite-borne sensors.

The atmosphere also scatters the radiation that passes through it, giving rise to a wide range of optical effects. The blueness of the outer atmosphere in Fig. 1.5 is due to the preferential scattering of incoming short wavelength (solar) radiation by air molecules, and the whiteness of lower layers is due to scattering from cloud droplets and atmospheric aerosols (i.e., particles). The backscattering of solar radiation off the top of the deck of low clouds off the California coast in Fig. 1.7 greatly enhances the whiteness (or reflectivity) of that region as viewed from space. Due to the presence of clouds and aerosols in the Earth's atmosphere, ~22% of the incoming solar radiation is backscattered to space without being absorbed. The backscattering of radiation by clouds and aerosols has a cooling effect on climate at the Earth's surface, which opposes the greenhouse effect.

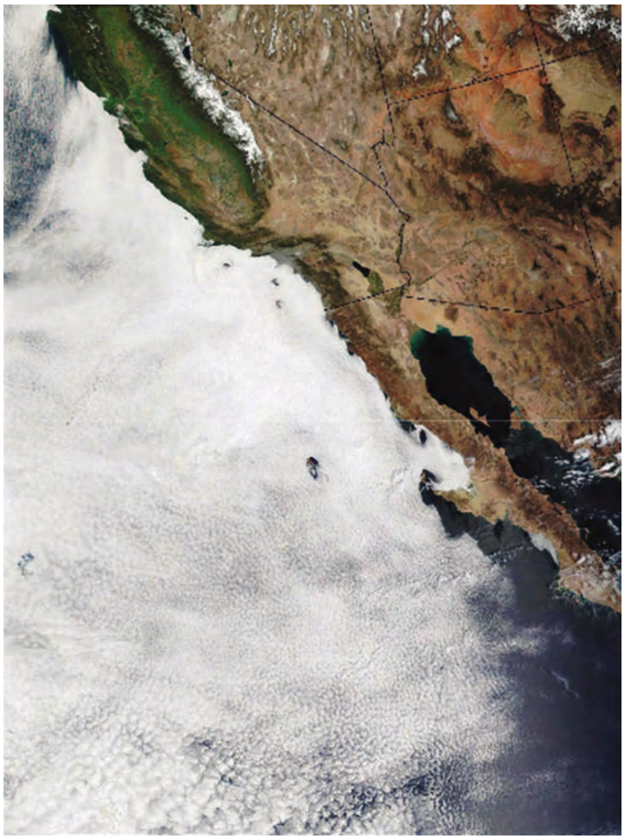


Fig. 1.7 A deck of low clouds off the coast of California, as viewed in reflected visible radiation. [NASA MODIS imagery. Photograph courtesy of NASA.]

1.3.2 Mass

At any point on the Earth's surface, the atmosphere exerts a downward force on the underlying surface due to the Earth's gravitational attraction. The downward force, (i.e., the *weight*) of a unit volume of air with density ρ is given by

$$F = \rho g \tag{1.4}$$

where g is the acceleration due to gravity. Integrating Eq. (1.4) from the Earth's surface to the "top" of the atmosphere, we obtain the atmospheric pressure on the Earth's surface p_s due to the weight (per unit area) of the air in the overlying column

$$p_s = \int_0^\infty \rho g dz \tag{1.5}$$

Neglecting the small variation of g with latitude, longitude and height, setting it equal to its mean value of $g_0 = 9.807 \text{ m s}^{-2}$, we can take it outside the integral, in which case, Eq. (1.5) can be written as

$$p_s = m g_0 \tag{1.6}$$

where $m = \int_0^\infty \rho dz$ is the vertically integrated mass per unit area of the overlying air.

Exercise 1.1 The globally averaged surface pressure is 985 hPa. Estimate the mass of the atmosphere.

Solution: From Eq. (1.6), it follows that

$$\overline{m} = \frac{\overline{p_s}}{g_0}$$

where the overbars denote averages over the surface of the Earth. In applying this relationship the pressure must be expressed in pascals (Pa). Substituting numerical values we obtain

$$\overline{m} = \frac{985 \times 10^2 \text{ Pa/hPa}}{9.807} = 1.004 \times 10^4 \text{ kg m}^{-2}$$

The mass of the atmosphere is

$$\begin{aligned} M_{atm} &= 4\pi R_E^2 \times \overline{m} \\ &= 4\pi \times (6.37 \times 10^6)^2 \text{ m}^2 \times 1.004 \times 10^4 \text{ kg m}^{-2} \\ &= 5.10 \times 10^{14} \text{ m}^2 \times 1.004 \times 10^4 \text{ kg m}^{-2} \\ &= 5.10 \times 10^{18} \text{ kg}^6 \end{aligned}$$ ■

1.3.3 Chemical Composition

The atmosphere is composed of a mixture of gases in the proportions shown in Table 1.1, where fractional concentration *by volume* is the same as that based on numbers of molecules, or partial pressures exerted by the gases, as will be explained more fully in Section 3.1. The fractional concentration *by mass* of a constituent is computed by weighting its fractional concentration *by volume* by its molecular weight, i.e.,

$$\frac{m_i}{\Sigma m_i} = \frac{n_i M_i}{\Sigma n_i M_i} \tag{1.7}$$

where m_i is the mass, n_i the number of molecules, and M_i the molecular weight of the ith constituent, and the summations are over all constituents.

Diatomic nitrogen (N_2) and oxygen (O_2) are the dominant constituents of the Earth's atmosphere, and argon (Ar) is present in much higher concentrations than the other noble gases (neon, helium, krypton, and xenon). Water vapor, which accounts for roughly 0.25% of the mass of the atmosphere, is a highly variable constituent, with concentrations ranging from around 10 parts per million by volume (ppmv) in the coldest regions of the Earth's atmosphere up to as much as 5% by volume in hot, humid air masses; a range of more than three orders of magnitude. Because of the large variability of water vapor concentrations in air, it is customary to list the percentages of the various constituents in relation to dry air. Ozone concentrations are also highly variable. Exposure to ozone concentrations >0.1 ppmv is considered hazardous to human health.

For reasons that will be explained in §4.4, gas molecules with certain structures are highly effective at trapping outgoing radiation. The most important of these so-called *greenhouse gases* are water vapor, carbon dioxide, and ozone. Trace constituents CH_4, N_2O, CO, and chlorofluorocarbons (CFCs) are also significant contributors to the greenhouse effect.

Among the atmosphere's trace gaseous constituents are molecules containing carbon, nitrogen, and sulfur atoms that were formerly incorporated into the cells of living organisms. These gases enter the atmosphere through the burning of plant matter and fossil fuels, emissions from plants, and the decay of plants and animals. The chemical transformations that remove these chemicals from the atmosphere involve oxidation, with the hydroxyl (OH) radical playing an important role. Some of the nitrogen and sulfur compounds are converted into particles that are eventually "scavenged" by raindrops, which contribute to acid deposition at the Earth's surface.

Table 1.1 Fractional concentrations by volume of the major gaseous constituents of the Earth's atmosphere up to an altitude of 105 km, with respect to dry air

Constituent[a]	Molecular weight	Fractional concentration by volume
Nitrogen (N_2)	28.013	78.08%
Oxygen (O_2)	32.000	20.95%
Argon (Ar)	39.95	0.93%
Water vapor (H_2O)	18.02	0–5%
Carbon dioxide (CO_2)	44.01	380 ppm
Neon (Ne)	20.18	18 ppm
Helium (He)	4.00	5 ppm
Methane (CH_4)	16.04	1.75 ppm
Krypton (Kr)	83.80	1 ppm
Hydrogen (H_2)	2.02	0.5 ppm
Nitrous oxide (N_2O)	56.03	0.3 ppm
Ozone (O_3)	48.00	0–0.1 ppm

[a] So called *greenhouse gases* are indicated by bold-faced type. For more detailed information on minor constituents, see Table 5.1.

[6] When the vertical and meridional variations in g and the meridional variations in the radius of the earth are accounted for, the mass per unit area and the total mass of the atmosphere are ~0.4% larger than the estimates derived here.

Although aerosols and cloud droplets account for only a minute fraction of the mass of the atmosphere, they mediate the condensation of water vapor in the atmospheric branch of the hydrologic cycle, they participate in and serve as sites for important chemical reactions, and they give rise to electrical charge separation and a variety of atmospheric optical effects.

1.3.4 Vertical structure

To within a few percent, the density of air at sea level is 1.25 kg m^{-3}. Pressure p and density ρ decrease nearly exponentially with height, i.e.,

$$p \simeq p_0 e^{-z/H} \tag{1.8}$$

where H, the e-folding depth, is referred to as the *scale height* and p_0 is the pressure at some reference level, which is usually taken as sea level ($z = 0$). In the lowest 100 km of the atmosphere, the scale height ranges roughly from 7 to 8 km. Dividing Eq. (1.8) by p_0 and taking the natural logarithms yields

$$\ln \frac{p}{p_0} \simeq -\frac{z}{H} \tag{1.9}$$

This relationship is useful for estimating the height of various pressure levels in the Earth's atmosphere.

Exercise 1.2 At approximately what height above sea level $\overline{z}_m$ does half the mass of the atmosphere lie above and the other half lie below? [Hint: Assume an exponential pressure dependence with $H = 8$ km and neglect the small vertical variation of g with height.]

Solution: Let $\overline{p}_m$ be the pressure level that half the mass of the atmosphere lies above and half lies below. The pressure at the Earth's surface is equal to the weight (per unit area) of the overlying column of air. The same is true of the pressure at any level in the atmosphere. Hence, $\overline{p}_m = \overline{p}_0/2$ where $\overline{p}_0$ is the global-mean sea-level pressure. From Eq. (1.9)

$$\overline{z}_m = -H \ln 0.5 = H \ln 2$$

Substituting $H = 8$ km, we obtain

$$\overline{z}_m = 8 \text{ Km} \times 0.693 \sim 5.5 \text{ km}$$

Because the pressure at a given height in the atmosphere is a measure of the mass that lies above that level, it is sometimes used as a vertical coordinate in lieu of height. In terms of mass, the 500-hPa level, situated at a height of around 5.5 km above sea level, is roughly halfway up to the top the atmosphere. ■

Density decreases with height in the same manner as pressure. These vertical variations in pressure and density are much larger than the corresponding horizontal and time variations. Hence it is useful to define a *standard atmosphere*, which represents the horizontally and temporally averaged structure of the atmosphere as a function of height only, as shown in Fig. 1.8. The nearly exponential height dependence of pressure and density can be inferred from the fact that the observed vertical profiles of pressure and density on these semilog plots closely resemble straight lines. The reader is invited to verify in Exercise 1.14 at the end of this chapter that the corresponding 10-folding depth for pressure and density is ~17 km.

Exercise 1.3 Assuming an exponential pressure and density dependence with $H = 7.5$ km, estimate the heights in the atmosphere at which (a) the air density is equal to 1 kg m^{-3} and (b) the height at which the pressure is equal to 1 hPa.

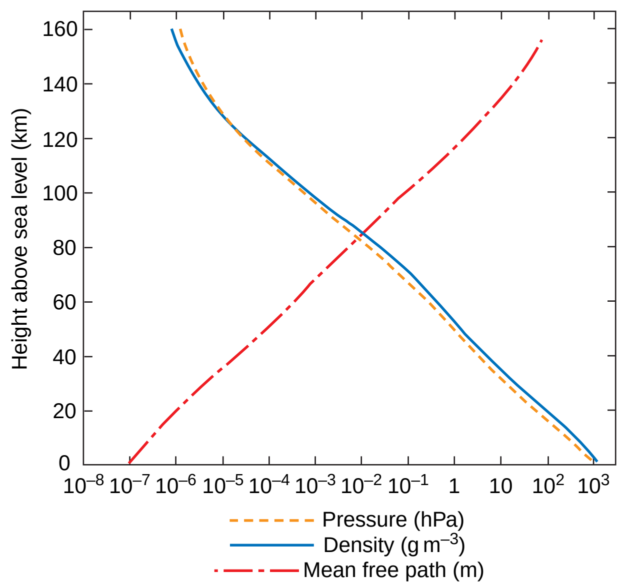


Fig. 1.8 Vertical profiles of pressure in units of hPa, density in units of kg m^{-3}, and mean free path (in meters) for the *U.S. Standard Atmosphere*.

Solution: Solving Eq. (1.9), we obtain $z = H \ln(\rho_0/\rho)$, and similarly for density. Hence, the heights are (a)

$$7.5 \text{ km} \times \ln\left(\frac{1.25}{1.00}\right) = 1.7 \text{ km}$$

for the 1-kg m^{-3} density level and (b)

$$7.5 \text{ km} \times \ln\left(\frac{1000}{1.00}\right) = 52 \text{ km}$$

for the 1-hPa pressure level. Because H varies with height, geographical location, and time, and the reference values ρ_0 and p_0 also vary, these estimates are accurate only to within ~10%. ■

Exercise 1.4 Assuming an exponential pressure and density dependence, calculate the fraction of the total mass of the atmosphere that resides between 0 and 1 scale height, 1 and 2 scale heights, 2 and 3 scale heights, and so on above the surface.

Solution: Proceeding as in Exercise 1.2, the fraction of the mass of the atmosphere that lies between 0 and 1, 1 and 2, 2 and 3, and so on scale heights above the Earth's surface is $e^{-1}, e^{-2}, \ldots e^{-N}$ from which it follows that the fractions of the mass that reside in the 1st, 2nd . . ., N^{th} scale height above the surface are $1 - e^{-1}, e^{-1}(1 - e^{-1}), e^{-2}(1 - e^{-1}) \ldots, e^{-N}(1 - e^{-1})$, where N is the height of the base of the layer expressed in scale heights above the surface. The corresponding numerical values are 0.632, 0.233, 0.086 . . . ■

Throughout most of the atmosphere the concentrations of N_2, O_2, Ar, CO_2, and other long-lived constituents tend to be quite uniform and largely independent of height due to mixing by turbulent fluid motions.[7] Above ~105 km, where the mean free path between molecular collisions exceeds 1 m (Fig. 1.8), individual molecules are sufficiently mobile that each molecular species behaves as if it alone were present. Under these conditions, concentrations of heavier constituents decrease more rapidly with height than those of lighter constituents: the density of each constituent drops off exponentially with height, with a scale height inversely proportional to molecular weight, as explained in Section 3.2.2. The upper layer of the atmosphere in which the lighter molecular species become increasingly abundant (in a relative sense) with increasing height is referred to as the *heterosphere*. The upper limit of the lower, well-mixed regime is referred to as the *turbopause*, where *turbo* refers to turbulent fluid motions and *pause* connotes limit of.

The composition of the outermost reaches of the atmosphere is dominated by the lightest molecular species (H, H_2, and He). During periods when the sun is active, a very small fraction of the hydrogen atoms above 500 km acquire velocities high enough to enable them to escape from the Earth's gravitational field during the long intervals between molecular collisions. Over the lifetime of the Earth the leakage of hydrogen atoms has profoundly influenced the chemical makeup of the Earth system, as discussed in Section 2.4.2.

The vertical distribution of temperature for typical conditions in the Earth's atmosphere, shown in Fig. 1.9, provides a basis for dividing the atmosphere into four layers (*troposphere, stratosphere,*

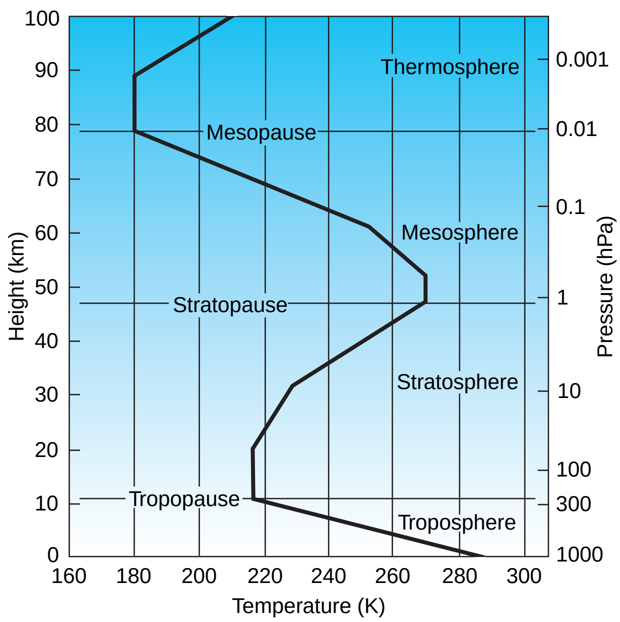


Fig. 1.9 A typical midlatitude vertical temperature profile, as represented by the U.S. Standard Atmosphere.

[7] In contrast, water vapor tends to be concentrated within the lowest few kilometers of the atmosphere because it condenses and precipitates out when air is lifted. Ozone are other highly reactive trace species exhibit heterogeneous distributions because they do not remain in the atmosphere long enough to become well mixed.

mesosphere, and *thermosphere*), the upper limits of which are denoted by the suffix *pause*.

The *tropo*(turning or changing)*sphere* is marked by generally decreasing temperatures with height, at an average *lapse rate*, of $\sim 6.5\ °\text{C km}^{-1}$. That is to say,

$$\Gamma \equiv \frac{\partial T}{\partial z} \sim 6.5\ °\text{C}\ \ \text{km}^{-1} = 0.0065\ °\text{C m}^{-1}$$

where T is temperature and Γ is the lapse rate. Tropospheric air, which accounts for $\sim$80% of the mass of the atmosphere, is relatively well mixed and it is continually being cleansed or scavenged of aerosols by cloud droplets and ice particles, some of which subsequently fall to the ground as rain or snow. Embedded within the troposphere are thin layers in which temperature increases with height (i.e., the lapse rate Γ is negative). Within these so-called *temperature inversions* it is observed that vertical mixing is strongly inhibited.

Within the *strato*-(layered)-*sphere*, vertical mixing is strongly inhibited by the increase of temperature with height, just as it is within the much thinner temperature inversions that sometimes form within the troposphere. The characteristic anvil shape created by the spreading of cloud tops generated by intense thunderstorms and volcanic eruptions when they reach the tropopause level, as illustrated in Fig. 1.10, is due to this strong stratification.

Cloud processes in the stratosphere play a much more limited role in removing particles injected by volcanic eruptions and human activities than they do in the troposphere, so residence times of particles tend to be correspondingly longer in the stratosphere. For example, the hydrogen bomb tests of the 1950s and early 1960s were followed by hazardous radioactive fallout events involving long-lived stratospheric debris that occurred as long as 2 years after the tests.

Fig. 1.10 A distinctive "anvil cloud" formed by the spreading of cloud particles carried aloft in an intense updraft when they encounter the tropopause. [Photograph courtesy of Rose Toomer and Bureau of Meteorology, Australia.]

Stratospheric air is extremely dry and ozone rich. The absorption of solar radiation in the ultraviolet region of the spectrum by this *stratospheric ozone layer* is critical to the habitability of the Earth. Heating due to the absorption of ultraviolet radiation by ozone molecules is responsible for the temperature maximum $\sim$50 km that defines the stratopause.

Above the ozone layer lies the *mesosphere* (meso connoting "in between"), in which temperature decreases with height to a minimum that defines the *mesopause*. The increase of temperature with height within the *thermosphere* is due to the absorption of solar radiation in association with the dissociation of diatomic nitrogen and oxygen molecules and the stripping of electrons from atoms. These processes, referred to as *photodissociation* and *photoionization*, are discussed in more detail in Section 4.4.3. Temperatures in the Earth's outer thermosphere vary widely in response to variations in the emission of ultraviolet and x-ray radiation from the sun's outer atmosphere.

At any given level in the atmosphere temperature varies with latitude. Within the troposphere, the *climatological-mean* (i.e., the average over a large number of seasons or years), *zonally averaged* temperature generally decreases with latitude, as shown in Fig. 1.11. The meridional temperature gradient is substantially stronger in the winter hemisphere where the polar cap region is in darkness. The tropopause is clearly evident in Fig. 1.11 as a discontinuity in the lapse rate. There is a break between the tropical tropopause, with a mean altitude $\sim$17 km, and the extratropical tropopause, with a mean altitude $\sim$10 km. The tropical tropopause is remarkably cold, with temperatures as low as -80 °C. The remarkable dryness of the air within the stratosphere is strong evidence that most of it has entered by way of this "cold trap."

Exercise 1.5 Based on data shown in Fig. 1.10, estimate the mean lapse rate within the tropical troposphere.

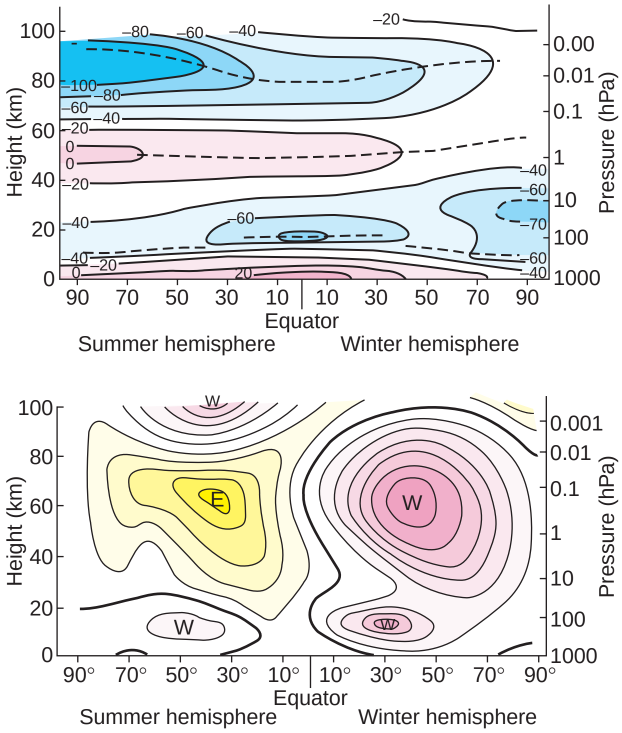


Fig. 1.11 Idealized meridional cross sections of zonally averaged temperature (in °C) (Top) and zonal wind (in m s^{-1}) (Bottom) around the time of the solstices, when the meridional temperature contrasts and winds are strongest. The contour interval is 20 °C; pink shading denotes relatively warm regions, and cyan shading relatively cold regions. The contour interval is 10 m s^{-1}; the zero contour is bold; pink shading and "W" labels denote westerlies, and yellow shading and "E" labels denote easterlies. Dashed lines indicate the positions of the tropopause, stratopause, and mesopause. This representation ignores the more subtle distinctions between northern and southern hemisphere climatologies. [Courtesy of Richard J. Reed.]

Solution: At sea level the mean temperature of the tropics is ~27 °C, the tropopause temperature is near −80 °C, and the altitude of the tropopause altitude is ~17 km. Hence the lapse-rate is roughly

$$\frac{[27-(-80)]\ ^\circ\mathrm{C}}{17\ \mathrm{km}} = 6.3\ ^\circ\mathrm{C\ km^{-1}}$$

Note that a decrease in temperature with height is implicit in the term (and definition of) *lapse rate*, so the algebraic sign of the answer is positive. ■

1.3.5 Winds

Differential heating between low and high latitudes gives rise to atmospheric motions on a wide range of scales. Prominent features of the so-called *atmospheric general circulation* include planetary-scale west-to-east (westerly) midlatitude *tropospheric jet streams*, centered at the tropopause break around 30° latitude, and lower *mesospheric jet streams*, both of which are evident in Fig. 1.11. The winds in the tropospospheric jet stream blow from the west throughout the year; they are strongest during winter and weakest during summer. In contrast, the mesospheric jet streams undergo a seasonal reversal: during winter they blow from the west and during summer they blow from the east.

Superimposed on the tropospheric jet streams are eastward propagating, *baroclinic waves* that feed upon and tend to limit the north–south temperature contrast across middle latitudes. Baroclinic waves are one of a number of types of *weather systems* that develop spontaneously in response to *instabilities* in the large-scale flow pattern in which they are embedded. The low level flow in baroclinic waves is dominated by *extratropical cyclones*, an example of which is shown in Fig. 1.12. The term *cyclone* denotes a closed circulation in which the air spins in the same sense as the Earth's rotation as viewed from above (i.e., counterclockwise in the northern hemisphere). At low levels the air spirals inward toward the center.[8] Much of the significant weather associated with extratropical cyclones is concentrated within narrow *frontal zones*, i.e., bands, a few tens of kilometers in width, characterized by strong horizontal temperature contrasts. Extratropical weather systems are discussed in Section 8.1.

Tropical cyclones (Fig. 1.13) observed at lower latitudes derive their energy not from the north–south temperature contrast, but from the release of latent heat of condensation of water vapor in deep convective clouds, as dicussed in Section 8.3. Tropical cyclones tend to be tighter and more axisymmetric than extratropical cyclones, and some of them are much more intense. A distinguishing feature of a well-developed tropical cyclone is the relatively calm, cloud-free *eye* at the center.

[8] The term *cyclone* derives from the Greek word for "coils of a snake."

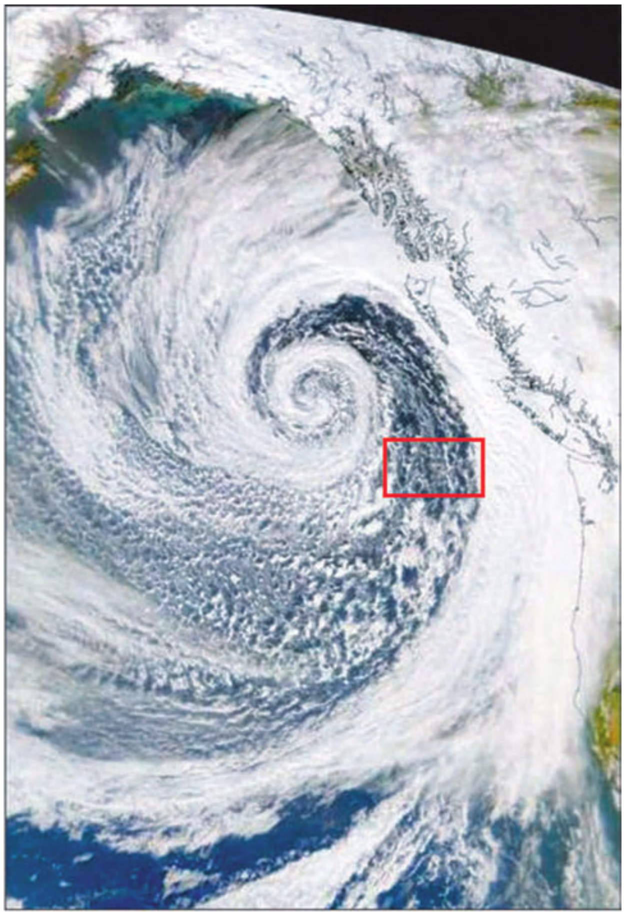


Fig. 1.12 An intense extratropical cyclone over the North Pacific. The spiral cloud pattern, with a radius of nearly 2000 km, is shaped by a vast counterclockwise circulation around a deep low pressure center. Some of the elongated cloud bands are associated with frontal zones. The region enclosed by the red rectangle is shown in greater detail in Fig. 1.21. [NASA MODIS imagery. Photograph courtesy of NASA.]

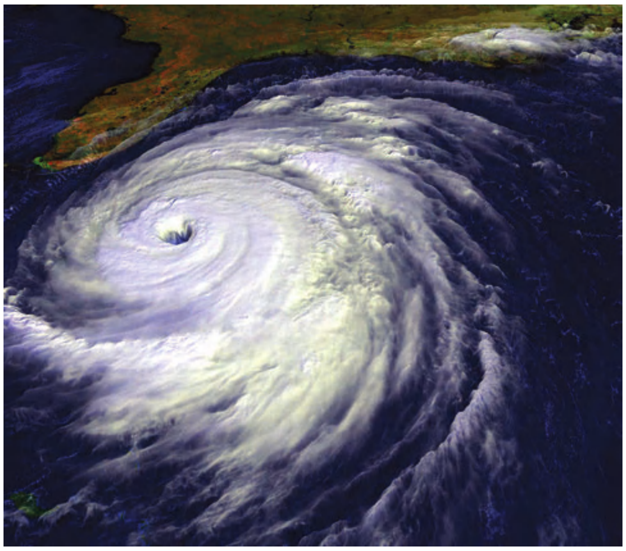


Fig. 1.13 The cloud pattern associated with an intense tropical cyclone approaching Florida. The eye is clearly visible at the center of the storm. The radius of the associated cloud system is ~600 km. [NOAA GOES imagery. Photograph courtesy of Harold F. Pierce, NASA Goddard Space Flight Center.]

a. Wind and pressure

The pressure field is represented on weather charts in terms of a set of *isobars* (i.e., lines or contours along which the pressure is equal to a constant value) on a horizontal surface, such as sea level. Isobars are usually plotted at uniform increments: for example, every 4 hPa on a sea-level pressure chart (e.g., . . . 996, 1000, 1004 . . . hPa). Local maxima in the pressure field are referred to as *high pressure centers* or simply *highs*, denoted by the symbol **H**, and minima as *lows* (**L**). At any point on a pressure chart the local *horizontal pressure gradient* is oriented perpendicular to the isobars and is directed from lower toward higher pressure. The strength of the horizontal pressure gradient is inversely proportional to the horizontal spacing between the isobars in the vicinity of that point.

With the notable exception of the equatorial belt (10 °S–10 °N), the winds observed in the Earth's atmosphere closely parallel the isobars. In the northern hemisphere, lower pressure lies to the left of the wind (looking downstream) and higher pressure to the right.[9,10] It follows that air circulates counterclockwise around lows and clockwise around highs, as shown in the right-hand side of Fig. 1.14. In the southern hemisphere the relationships are in the opposite sense, as indicated in the left-hand side of Fig. 1.14.

This seemingly confusing set of rules can be simplified by replacing the words "clockwise" and

[9] This relationship was first noted by Buys-Ballot in 1857, who stated: If, in the northern hemisphere, you stand with your back to the wind, pressure is lower on your left hand than on your right.

[10] **Christopher H. D. Buys-Ballot** (1817–1890) Dutch meteorologist, professor of mathematics at the University of Utrecht. Director of Dutch Meteorolgical Institute (1854–1887). Labored unceasingly for the widest possible network of surface weather observations.

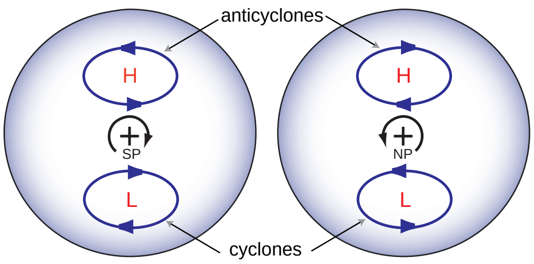


Fig. 1.14 Blue arrows indicate the sense of the circulation around highs (H) and lows (L) in the pressure field, looking down on the South Pole (left) and the North Pole (right). Small arrows encircing the poles indicate the sense of the Earth's rotation.

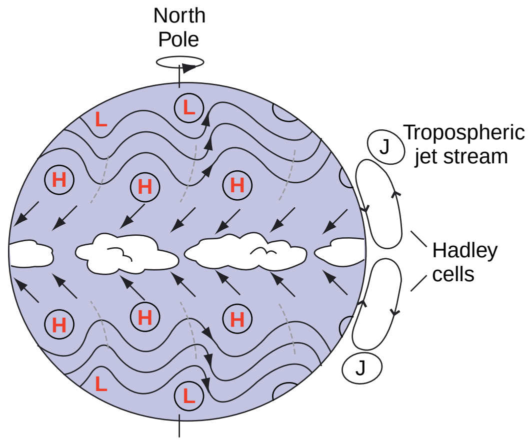


Fig. 1.15 Schematic depiction of sea-level pressure isobars and surface winds on an idealized *aqua planet*, with the sun directly overhead on the equator. The rows of H's denote the subtropical high-pressure belts, and the rows of L's denote the subpolar low-pressure belt. Hadley cells and tropospheric jet streams (J) are also indicated.

"counterclockwise" with the terms *cyclonic* and *anticyclonic* (i.e., in the same or in the opposite sense as the Earth's rotation, looking down on the pole). A *cyclonic circulation* denotes a counterclockwise circulation in the northern hemisphere and a clockwise circulation in the southern hemisphere. In either hemisphere the circulation around low pressure centers is cyclonic, and the circulation around high pressure centers is anticyclonic: that is to say, in reference to the pressure and wind fields, the term *low* is synonymous with *cyclone* and *high* with *anticyclone*.

In the equatorial belt the wind tends to blow straight down the pressure gradient (i.e., directly across the isobars from higher toward lower pressure). In the surface wind field there is some tendency for *cross-isobar flow* toward lower pressure at higher latitudes as well, particularly over land. The basis for these relationships is discussed in Chapter 7.

b. The observed surface wind field

This subsection summarizes the major features of the geographically and seasonally varying *climatological-mean* surface wind field (i.e., the background wind field upon which transient weather systems are superimposed). It is instructive to start by considering the circulation on an idealized ocean-covered Earth with the sun directly overhead at the equator, as inferred from simulations with numerical models.

The main features of this idealized "aqua-planet, perpetual equinox" circulation are depicted in Fig. 1.15. The extratropical circulation is dominated by *westerly wind belts*, centered around 45 °N and 45 °S. The westerlies are disturbed by an endless succession of eastward migrating disturbances called *baroclinic waves*, which cause the weather at these latitudes to vary from day to day. The average wavelength of these waves is ~4000 km and they propagate eastward at a rate of ~10 m s^{-1}.

The tropical circulation in the aqua-planet simulations is dominated by much steadier *trade winds*,[11] marked by an easterly zonal wind component and a component directed toward the equator. The *northeasterly trade winds* in the northern hemisphere and the *southeasterly trade winds* in the southern hemisphere are the surface manifestation of overturning circulations that extend through the depth of the troposphere. These so-called *Hadley*[12] *cells* are characterized by (1) equatorward flow in the boundary layer, (2) rising motion within a few degrees of the equator, (3) poleward return flow in the tropical upper troposphere, and (4) sinking motion in the

[11] The term *trade winds* or simply *trades* derives from the steady, dependable northeasterly winds that propelled sailing ships along the popular trade route across the tropical North Atlantic from Europe to the Americas.

[12] **George Hadley** (1685–1768) English meteorologist. Originally a barrister. Formulated a theory for the trade winds in 1735 which went unnoticed until 1793 when it was discovered by John Dalton. Hadley clearly recognized the importance of what was later to be called the Coriolis force.

subtropics, as indicated in Fig. 1.15. Hadley cells and trade winds occupy the same latitude belts.

In accord with the relationships between wind and pressure described in the previous subsection, trade winds and the extratropical westerly wind belt in each hemisphere in Fig. 1.15 are separated by a *subtropical high-pressure belt* centered ~30° latitude in which the surface winds tend to be weak and erratic. The jet streams at the tropopause (12 km; 250 hPa) level are situated directly above the subtropical high pressure belts at the Earth's surface. A weak minimum in sea-level pressure prevails along the equator, where trade winds from the northern and southern hemispheres converge. Much deeper lows form in the extratropics and migrate toward the poleward flank of the extratropical westeries to form the *subpolar low pressure belts*.

In the real world, surface winds tend to be stronger over the oceans than over land because they are not slowed as much by surface friction. Over the Atlantic and Pacific Oceans, the surface winds mirror many of the features in Fig. 1.15, but a longitudinally dependent structure is apparent as well. The subtropical high-pressure belt, rather than being continuous, manifests itself as distinct high-pressure centers, referred to as *subtropical anticyclones*, centered over the mid-oceans, as shown in Fig. 1.16.

In accord with the relationships between wind and pressure described in the previous subsection, surface winds at lower latitudes exhibit an equatorward component on the eastern sides of the oceans and a poleward component on the western sides. The equatorward surface winds along the eastern sides of the oceans carry (or *advect*) cool, dry air from higher latitudes into the subtropics; they drive coastal ocean currents that advect cool water equatorward; and they induce coastal upwelling of cool, nutrient-rich ocean water, as explained in the next chapter. On the western sides of the Atlantic and Pacific Oceans, poleward winds advect warm, humid, tropical air into middle latitudes.

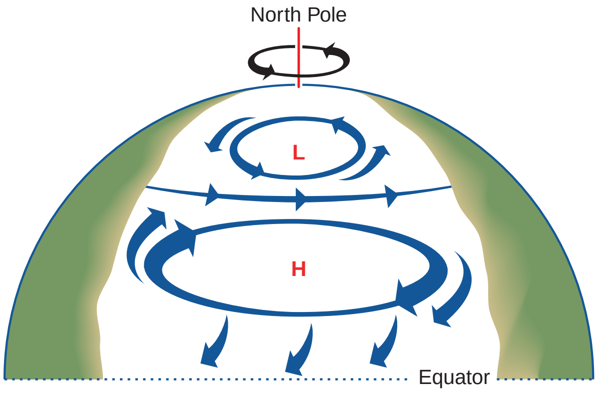


Fig. 1.16 Schematic of the surface winds and sea-level pressure maxima and minima over the Atlantic and Pacific Oceans showing subtropical anticyclones, subpolar lows, the midlatitude westerly belt, and trade winds.

In an analogous manner, the subpolar low-pressure belt manifests itself as mid-ocean cyclones referred to, respectively, as the *Icelandic low* and the *Aleutian low*. The poleward flow on the eastern flanks of these semipermanent, subpolar cyclones moderates the winter climates of northern Europe and the Pacific coastal zone poleward of ~40 °N. The subtropical anticyclones are most pronounced during summer, whereas the subpolar lows are most pronounced during winter.

The idealized tropical circulation depicted in Fig. 1.15, with the northeasterly and southeasterly trade winds converging along the equator, is not realized in the real atmosphere. Over the Atlantic and Pacific Oceans, the trade winds converge, not along the equator, but along ~7 °N, as depicted schematically in the upper panel of Fig. 1.17. The belt in which the convergence takes place is referred to as the *intertropical convergence zone (ITCZ)*. The asymmetry with respect to the equator is a consequence of the land–sea geometry, specifically the northwest–southeast orientation of the west coastlines of the Americas and Africa.

Surface winds over the tropical Indian Ocean are dominated by the seasonally reversing *monsoon circulation*,[13] consisting of a broad arc originating as a westward flow in the winter hemisphere, crossing the equator, and curving eastward to form a belt of moisture-laden westerly winds in the summer hemisphere, as depicted [for the northern hemisphere (i.e., *boreal*) summer] in the lower panel of Fig. 1.17. The monsoon is driven by the presence of India and southeast Asia in the northern hemisphere subtropics versus the southern hemisphere subtropics. Surface temperatures over land respond much more strongly to the seasonal variations in solar heating than those over ocean. Hence, during July the

[13] From *mausin*, the Arabic word for season.

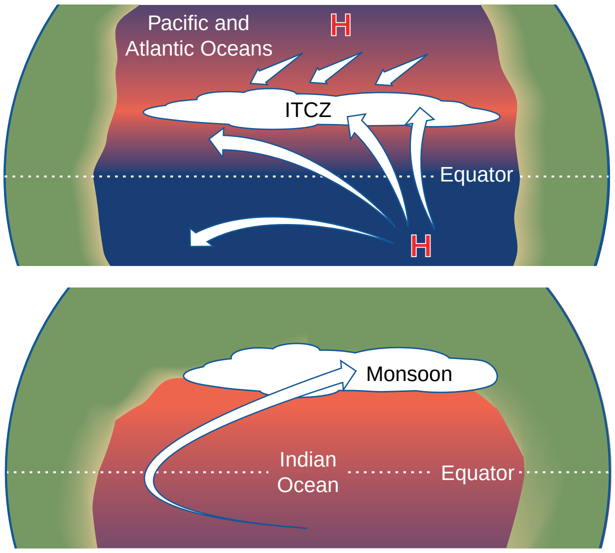


Fig. 1.17 Schematic depicting surface winds (arrows), rainfall (cloud masses), and sea surface temperature over the tropical oceans between ~30 °N and 30 °S. Pink shading denotes warmer, blue cooler sea surface temperature, and khaki shading denotes land. (Top) Atlantic and Pacific sectors where the patterns are dominated by the intertropical convergence zone (ITCZ) and the equatorial dry zone to the south of it. (Bottom) Indian Ocean sector during the northern (boreal) summer monsoon, with the Indian subcontinent to the north and open ocean to the south. During the austral summer (not shown) the flow over the Indian Ocean is in the reverse direction and the rain belt lies just to the south of the equator.

subtropical continents of the northern hemisphere are much warmer than the sea surface temperature over the tropical Indian Ocean. It is this temperature contrast that drives the monsoon flow depicted in the lower panel of Fig. 1.17. In January, when India and southeast Asia are cooler than the sea surface temperature over the tropical Indian Ocean, the monsoon flow is in the reverse sense (not shown).

The reader is invited to compare the observed climatological-mean surface winds for January and July shown in Figs. 1.18 and 1.19 with the idealized flow patterns shown in the two previous figures. In Fig. 1.18, surface winds, based on satellite data, are shown together with the rainfall distribution, indicated by shading, and in Fig. 1.19 a different version of the surface wind field, derived from a blending of many datasets, is superimposed on the climatological-mean sea-level pressure field.

By comparing the surface wind vectors with the shading in Fig. 1.18, it is evident that the major rain belts, which are discussed in the next subsection, tend to be located in regions where the surface wind vectors flow together (i.e., converge). Convergence at low levels in the atmosphere is indicative of ascending motion aloft. Through the processes discussed in Chapter 3, lifting of air leads to condensation of water vapor and ultimately to precipitation. Figure 1.19 provides verification that the surface winds tend to blow parallel to the isobars, except in the equatorial belt. At all latitudes a systematic drift across the isobars from higher toward lower pressure is also clearly apparent.

The observed winds over the southern hemisphere (Figs. 1.18 and 1.19) exhibit well-defined extratropical westerly and tropical trade wind belts reminiscent of those in the idealized aqua-planet simulations (Fig. 1.15). Over the northern hemisphere the surface winds are strongly influenced by the presence of high latitude continents. The subpolar low-pressure belt manifests itself as oceanic pressure minima (the *Icelandic* and *Aleutian lows*) surrounded by cyclonic (counterclockwise) circulations, as discussed in connection with Fig. 1.16. These features and the belts of westerly winds to the south of them are more pronounced during January than during July. In contrast, the northern hemisphere oceanic subtropical anticyclones are more clearly discernible during July.

c. Motions on smaller scales

Over large areas of the globe, the heating of the Earth's surface by solar radiation gives rise to buoyant plumes analogous to those rising in a pan of water heated from below. As the plumes rise, the displaced air subsides slowly, creating a two-way circulation. Plumes of rising air are referred to by glider pilots as *thermals*, and when sufficient moisture is present they are visible as cumulus clouds (Fig. 1.20). When the overturning circulations are confined to the lowest 1 or 2 km of the atmosphere (the so-called *mixed layer* or *atmospheric boundary layer*), as is often the case, they are referred to as *shallow convection*. Somewhat deeper, more vigorous convection gives rise to showery weather in cold air masses flowing over a warmer surface (Fig. 1.21).

Under certain conditions, buoyant plumes originating near the Earth's surface can break through the weak temperature inversion that usually caps the mixed layer, giving rise to towering clouds that extend all the way to the tropopause, as shown in Fig. 1.22. These clouds are the signature of *deep convection*,

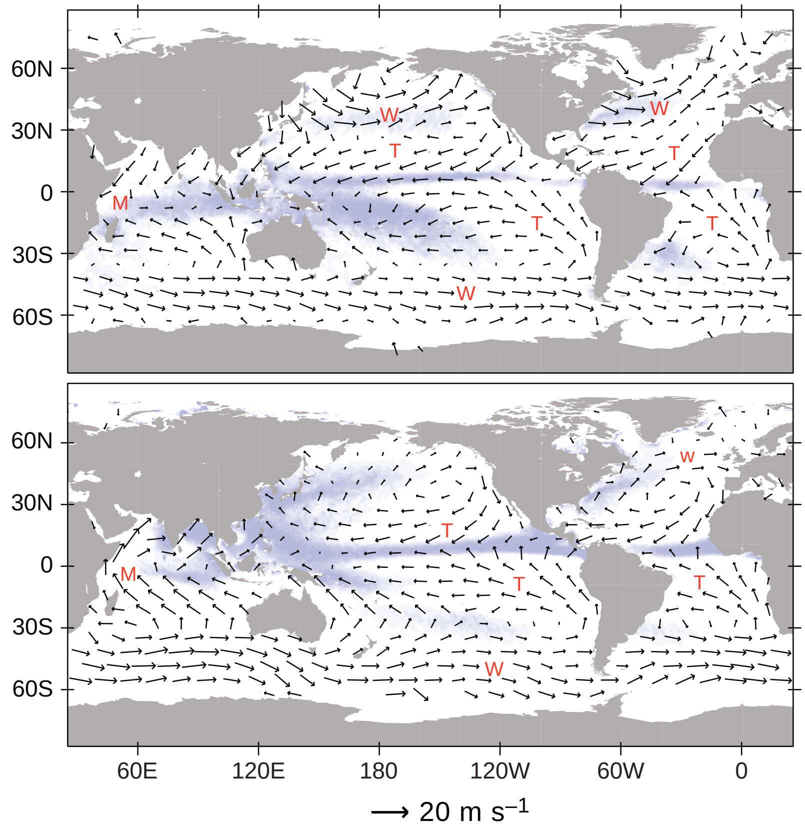


Fig. 1.18 December–January–February and June–July–August surface winds over the oceans based on 3 years of satellite observations of capillary waves on the ocean surface. The bands of lighter shading correspond to the major rain belts. **M**'s denote monsoon circulations, **W**'s westerly wind belts, and **T**'s trade winds. The wind scale is at the bottom of the figure. [Based on QuikSCAT data. Courtesy of Todd P. Mitchell.]

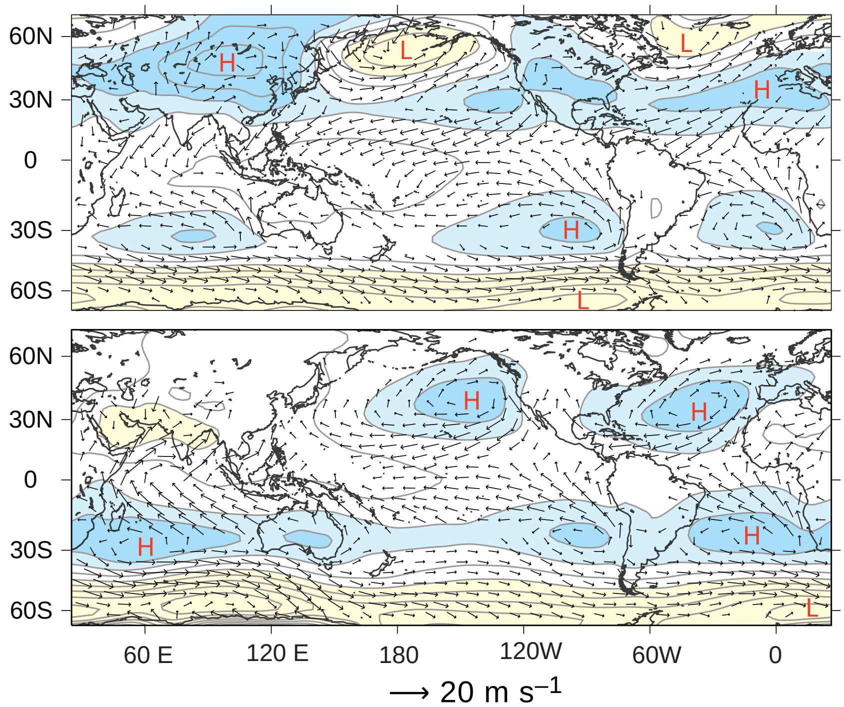


Fig. 1.19 December–January–February (top) and June–July–August (bottom) surface winds, as in Fig. 1.18, but superimposed on the distribution of sea-level pressure. The contour interval for sea-level pressure is 5 hPa. Pressures above 1015 hPa are shaded blue, and pressures below 1000 hPa are shaded yellow. The wind scale is at the bottom of the figure. [Based on the NCEP/NCAR reanalyses. Courtesy of Todd P. Mitchell.]

Fig. 1.20 Lumpy cumulus clouds reveal the existence of shallow convection that is largely confined to the atmospheric boundary layer. [Photograph courtesy of Bruce S. Richardson.]

Fig. 1.21 Enlargement of the area enclosed by the red rectangle in Fig. 1.12 showing convection in a cold air mass flowing over warmer water. The centers of the convection cells are cloud free, and the cloudiness is concentrated in narrow bands at the boundaries between cells. The clouds are deep enough to produce rain or snow showers. [NASA MODIS imagery. Photograph courtesy of NASA.]

Fig. 1.22 Clouds over the south China Sea as viewed from a research aircraft flying in the middle troposphere. The foreground is dominated by shallow convective clouds, while deep convection is evident in the background. [Photograph courtesy of Robert A. Houze.]

which occurs intermittently in the tropics and in warm, humid air masses in middle latitudes. Organized deep convection can cause locally heavy rainstorms, often accompanied by lightning and sometimes by hail and strong winds.

Convection is not the only driving mechanism for small-scale atmospheric motions. Large-scale flow over small surface irregularities induces an array of chaotic waves and eddies on scales ranging up to a few kilometers. Such *boundary layer turbulence*, the subject of Chapter 9, is instrumental in causing smoke plumes to widen as they age (Fig. 1.23), in limiting the strength of the winds in the atmosphere, and in mixing momentum, energy, and trace constituents between the atmosphere and the underlying surface.

Turbulence is not exclusively a boundary layer phenomenon: it can also be generated by flow instabilities higher in the atmosphere. The cloud pattern shown in Fig. 1.24 reveals the presence of waves that develop spontaneously in layers with strong *vertical wind shear* (layers in which the wind changes rapidly with height in a vectorial sense). These waves amplify and break, much as ocean waves do when they encounter a beach. *Wave breaking* generates smaller scale waves and eddies, which, in turn, become unstable. Through this succession of instabilities, kinetic energy extracted from the large-scale wind field within the planetary boundary layer and within patches of strong vertical wind shear in the free atmosphere gives rise to a spectrum of small-scale

Fig. 1.23 Exhaust plume from the NASA space shuttle launch on February 7, 2001. The widening of the plume as it ages is due to the presence of small-scale turbulent eddies. The curved shape of the plume is due to the change in horizontal wind speed and direction with height, referred to as *vertical wind shear*. The bright object just above the horizon is the moon and the dark shaft is the shadow of the upper, sunlit part of the smoke plume. [Photograph courtesy of Patrick McCracken, NASA headquarters.]

motions extending down to the molecular scale, inspiring Richardson's[14] celebrated rhyme:

> Big whirls have smaller whirls that feed on their velocity, and little whirls have lesser whirls, and so on to viscosity. . . . in the molecular sense.

Within localized patches of the atmosphere where wave breaking is particularly intense, eddies on

Fig. 1.24 Billows along the top of this cloud layer reveal the existence of breaking waves in a region of strong vertical wind shear. The right-to-left component of the wind is increasing with height. [Courtesy of Brooks Martner.]

scales of tens of meters can be strong enough to cause discomfort to airline passengers and even, in exceptional cases, to pose hazards to aircraft. Turbulence generated by shear instability is referred to as *clear air turbulence (CAT)* to distinguish it from the turbulence that develops within the cloudy air of deep convective storms.

1.3.6 Precipitation

Precipitation tends to be concentrated in space and time. Annual-mean precipitation at different points on Earth ranges over two orders of magnitude, from a few tens of centimeters per year in dry zones to several meters per year in the belts of heaviest rainfall, such as the ITCZ. Over much of the world climatological-mean precipitation exhibits equally dramatic seasonal variations. The global-mean, annual-mean precipitation rate is ~1 m of liquid water per year or ~0.275 cm per day or 1 m per year.

Climatological-mean distributions of precipitation for the months of January and July are shown in Fig. 1.25. The narrow bands of heavy rainfall that dominate the tropical Atlantic and Pacific sectors coincide with the ITCZ in the surface wind field. In the Pacific and Atlantic sectors the ITCZ is flanked by expansive *dry zones* that extend westward from the continental deserts and cover much of the subtropical oceans. These features coincide with the

[14] **Lewis F. Richardson** (1881–1953). English physicist and meteorologist. Youngest of seven children of a Quaker tanner. Served as an ambulance driven in France during World War I. Developed a set of finite differences for solving differential equations for weather prediction, but his formulation was not quite correct and at that time (1922) computations of this kind could not be performed quickly enough to be of practical use. Pioneer in the causes of war, which he described in his books "Arms and Insecurity" and "Statistics of Deadly Quarrels," Boxward Press, Pittsburg, 1960. Sir Ralph Richardson, the actor, was his nephew.

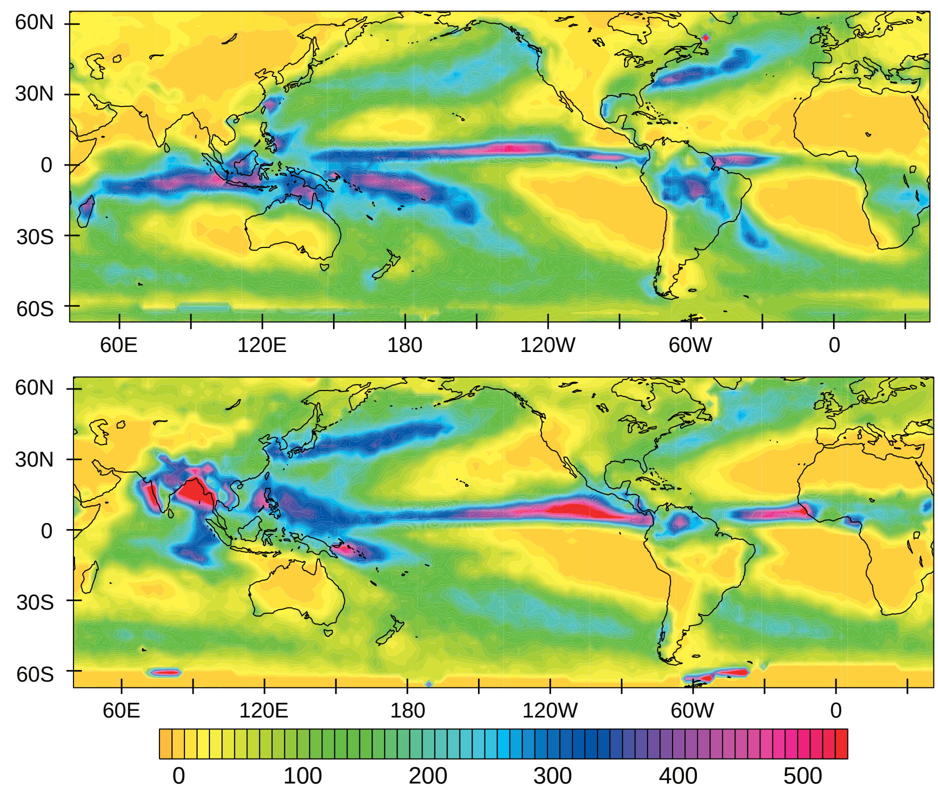


Fig. 1.25 January and July climatological-mean precipitation. [Based on infrared and microwave satellite imagery over the oceans and rain gauge data over land, as analyzed by the NOAA National Centers for Environmental Prediction CMAP project. Courtesy of Todd P. Mitchell.]

subtropical anticyclones and, in the Pacific and Atlantic sectors, they encompass equatorial regions as well.

Small seasonal or year-to-year shifts in the position of the ITCZ can cause dramatic local variations in rainfall. For example, at Canton Island (3 °S, 170 °W) near the western edge of the equatorial dry zone, rainfall rates vary from zero in some years to over 30 cm per month (month after month) in other years in response to subtle year-to-year variations in sea surface temperature over the equatorial Pacific that occur in association with El Niño, as discussed in Section 10.2.2.

Over the tropical continents, rainfall is dominated by the monsoons, which migrate northward and southward with the seasons, following the sun. Most equatorial regions receive rainfall year-round, but the belts that lie 10 - 20° away from the equator experience pronounced dry seasons that correspond to the time of year when the sun is overhead in the opposing hemisphere. The rainy season over India and southeast Asia coincides with the time of year in which the surface winds over the northern Indian Ocean blow from the west (Figs. 1.17 and 1.18). Analogous relationships exist between wind and rainfall in Africa and the Americas. The onset of the rainy season, a cause for celebration in many agricultural regions of the subtropics, is remarkably regular from one year to the next and it is often quite dramatic: for example, in Mumbai (formerly Bombay) on the west coast of India, monthly mean rainfall jumps from less than 2 cm in May to ~50 cm in June.

The flow of warm humid air around the western flanks of the subtropical anticyclones brings copious summer rainfall to eastern China and Japan and the eastern United States. In contrast, Europe and western North America and temperate regions of the southern hemisphere experience dry summers. These regions derive most of their annual precipitation from wintertime extratropical cyclones that form within the belts of westerly surface winds over the oceans and propagate eastward over land. The rainfall maxima extending across the Pacific and Atlantic at latitudes ~45 °N in Fig. 1.25 are manifestations of these oceanic *storm tracks*.

Rainfall data shown in Fig. 1.25, which are averaged over 2.5° latitude × 2.5° longitude grid boxes, do not fully resolve the fine structure of the distribution

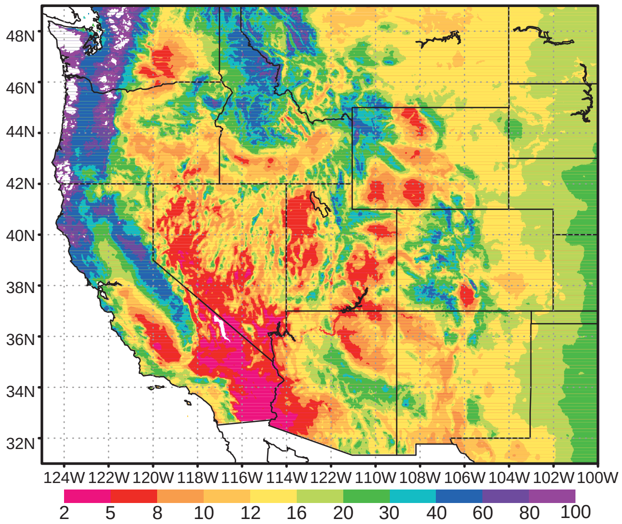


Fig. 1.26 Annual-mean precipitation over the western United States resolved on a 10-km scale making use of a model. The color bar is in units of inches of liquid water (1 in. = 2.54 cm). Much of the water supply is derived from a winter snow pack, which tends to be concentrated in regions of blue, purple, and white shading. [Map produced by the NOAA Western Regional Climate Center using PRISM data from Oregon State University. Courtesy of Kelly Redmond.]

of precipitation in the presence of *orography* (i.e., terrain). Flow over and around mountain ranges imparts a fractal-like structure to the precipitation distribution, with enhanced precipitation in regions where air tends to be lifted over terrain features and suppressed precipitation in and downstream of regions of descent (i.e., *subsidence*).

The distribution of annual-mean precipitation over the western United States, shown in Fig. 1.26, illustrates the profound influence of orography. Poleward of ~35°, which corresponds to the equatorward limit of the extratropical westerly flow regime that prevails during the wintertime, annual-mean precipitation tends to be enhanced where moisture-laden marine air is lifted as it moves onshore and across successive ranges of mountains. The regions of suppressed precipitation on the lee side of these ranges are referred to as *rain shadows*.

On any given day, the cloud patterns revealed by global satellite imagery exhibit patches of deep convective clouds that can be identified with the ITCZ and the monsoons over the tropical continents of the summer hemisphere; a relative absence of clouds in the subtropical dry zones; and a succession of comma-shaped, frontal cloud bands embedded in the baroclinic waves tracking across the mid-latitude oceans. These features are all present in the example shown in Fig. 1.27.

1.4 What's Next?

The brief survey of the atmosphere presented in this chapter is just a beginning. All the major themes introduced in this survey are developed further in subsequent chapters. The first section of the next chapter provides more condensed surveys of the other components of the Earth system that play a role in climate: the oceans, the crysophere, the terrestrial biosphere, and the Earth's crust and mantle.

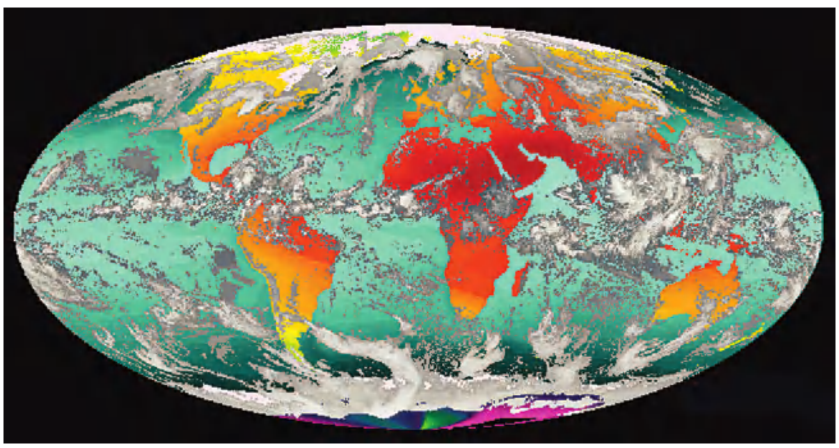


Fig. 1.27 Composite satellite image showing sea surface temperature and land surface air temperature and clouds. [Courtesy of the University of Wisconsin Space Science and Engineering Center.]

Climate change and its implications

Dr. Raji P

Lecture-2

1st lecture: Earth’s Atmosphere

- Optical properties of Earth’s atmosphere
- Mass of Earth’s atmosphere
- Vertical structure of the atmosphere: troposphere, stratosphere, mesosphere, thermosphere
- Temperature, pressure, density variations in the atmosphere

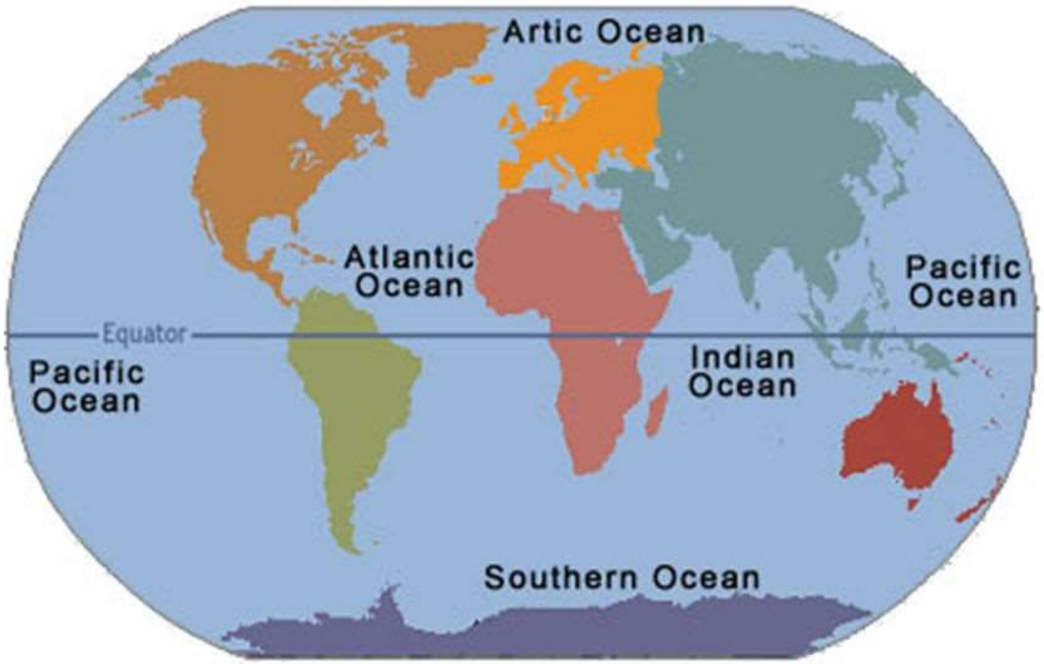
Class outline: Introduction (Conti…)

Earth system components

- Oceans
- Cryosphere
- Biosphere
- Earth’s crust and mantle

- Climate depends on atmosphere as well as physical, chemical, and biological processes involving other components of earth system

The Oceans

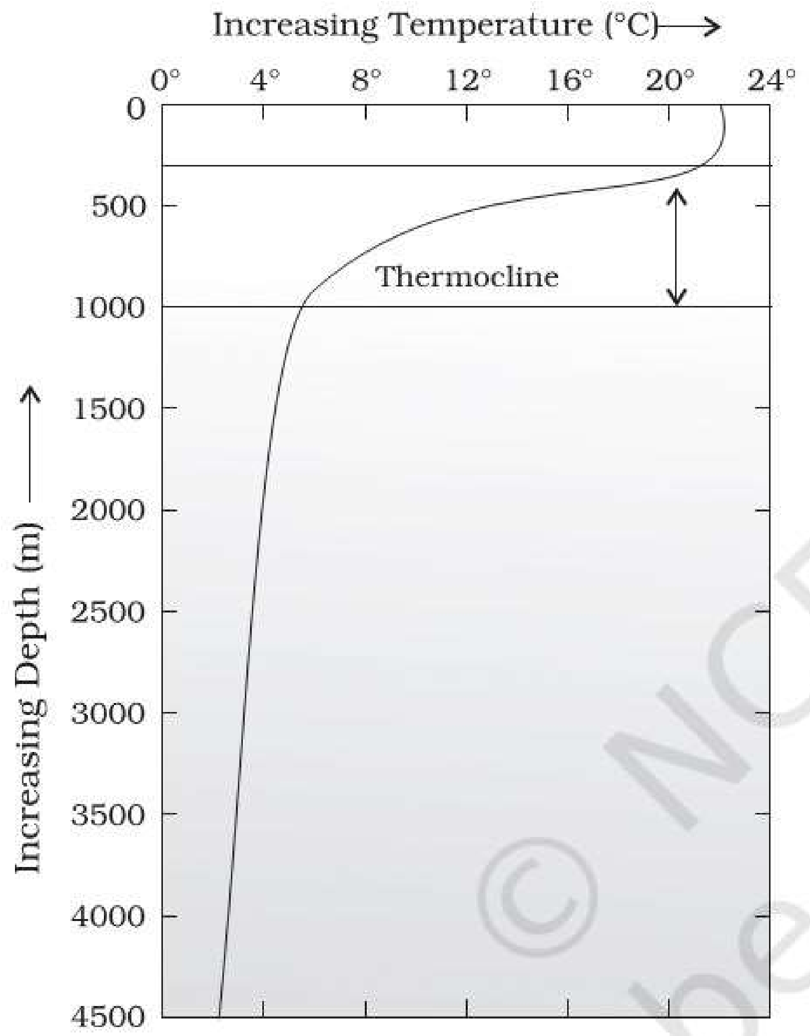


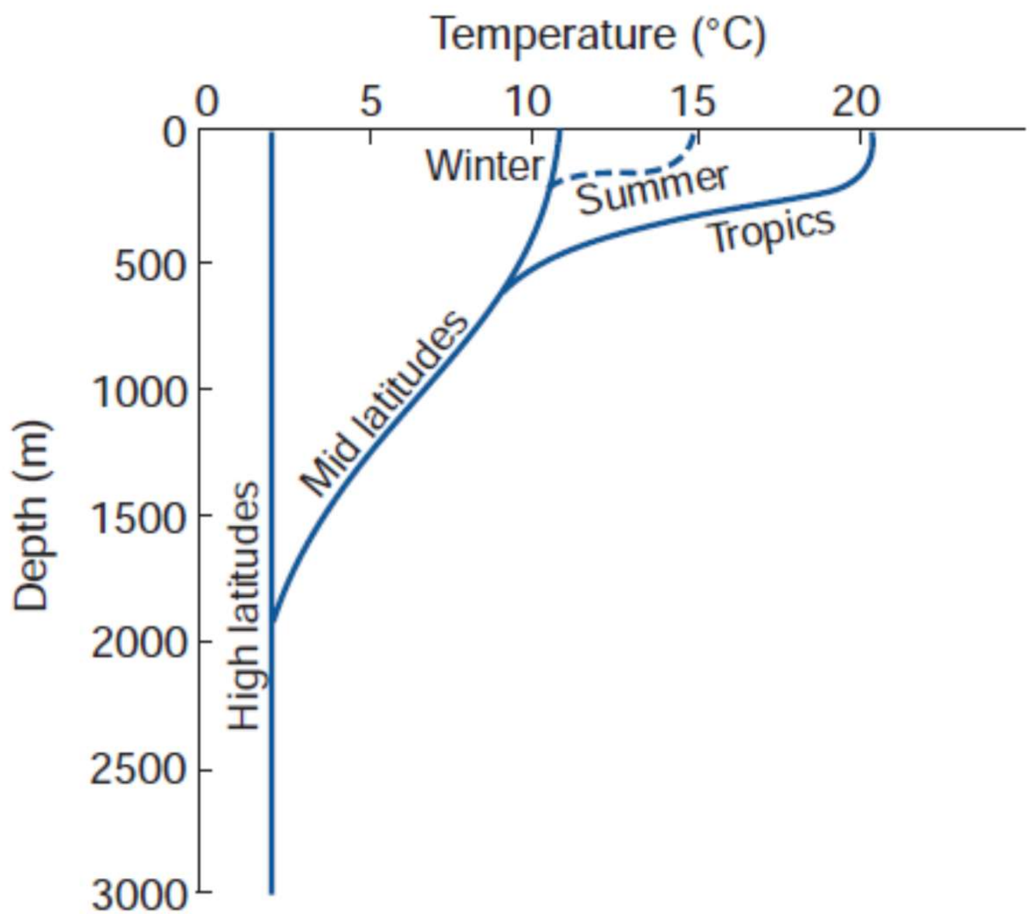
There are 5 Main Oceans
The Pacific Ocean is the largest ocean – by far!

- Oceans cover 72% of the area of the earth’s surface
- Reaches to an extreme depth of 11 km
- Mass of the ocean is approx. 250 times as that of atmosphere

Composition and vertical structure of ocean:

- Density of sea water linearly proportional to the concentration of dissolved salt
- Sea water contains salt ~34 -36 g/kg of fresh water
- Sea water is ~ 2.4% denser than fresh water @ same temperature


Increasing Temperature (°C)
0°
4°
8°
12°
16°
20°
24°
0
500
1000
1500
2000
2500
3000
3500
4000
4500
Thermocline
Increasing Depth (m)


Temperature (°C)
0
5
10
15
20
0
500
1000
1500
2000
2500
3000
Depth (m)
Winter
Summer
Tropics
Mid latitudes
High latitudes

- The density of sea water ranges from 1.02 to 1.03 kg/m^3

- Density of water in the wind-stirred layer (mixed layer) is smaller by a few tenths of a percent than density of water below it

- Thermocline: Layer in which there is a strong temperature gradient exist with respect to depth

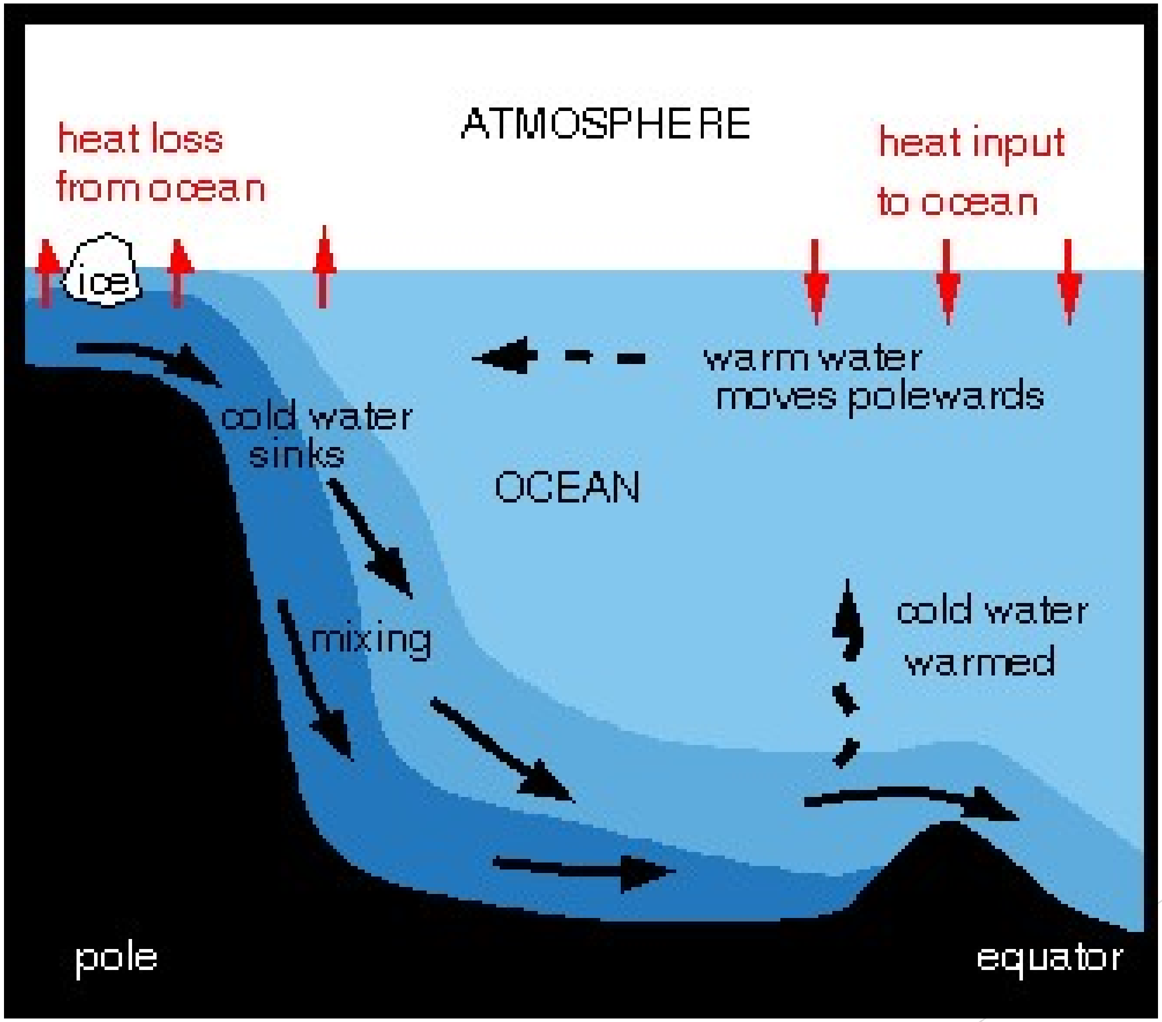
- Precipitation lowers the salinity by diluting the salts that are present in the oceanic mixed layer

- But evaporation increases the salinity by removing fresh water and thereby concentrating the residual salts

Exercise-1

- A heavy tropical storm dumps 20 cm of rainfall in a region of the ocean in which the salinity is 35 g kg-1 and the mixed layer depth is 50 m. Assuming that the water is well mixed, by how much does the salinity decrease?

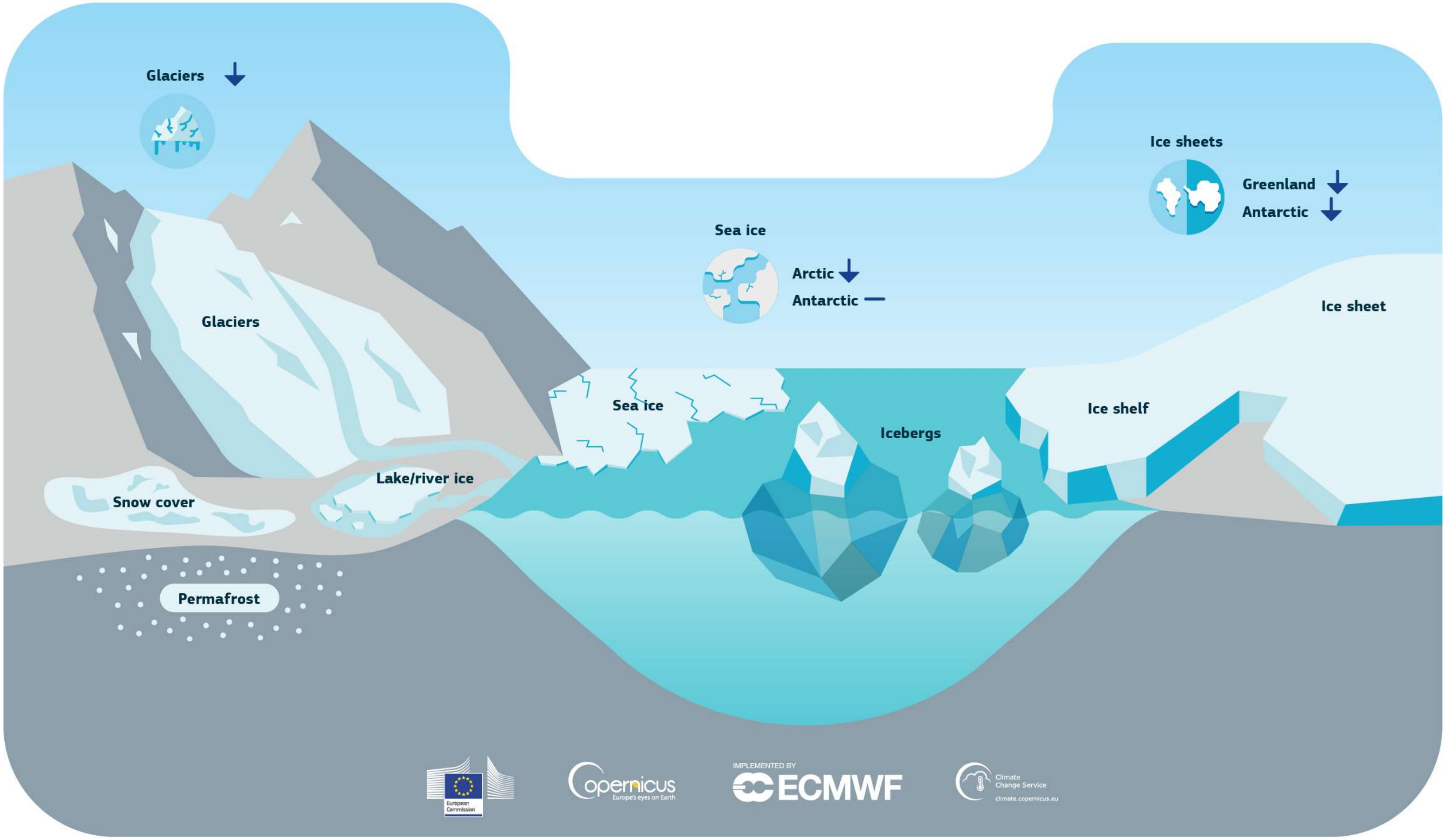
Ocean circulation




0°C

- The ocean circulation is composed of a wind-driven component and a thermohaline (density-dependent) component
- The wind driven circulation dominates the surface currents, and it is restricted to the topmost few hundred meters
- The circulation deeper in the oceans is dominated by the slower thermohaline circulation
- Velocities in wind driven currents are on the order of 10 cm/s
- The timescale in which a parcel completes a circuit of this thermohaline circulation is on the order of hundreds of years

Cryosphere



Glaciers
Ice sheets
Greenland
Antarctic
Sea ice
Arctic
Antarctic
Ice sheet
Glaciers
Sea ice
Icebergs
Ice shelf
Lake/river ice
Snow cover
Permafrost
European Commission
Copernicus
Europe's eyes on Earth
IMPLEMENTED BY
ECMWF
Climate Change Service
climate.copernicus.eu

- Cryo (frozen)-sphere refers to the components of the earth system comprised of water in its solid state
- Taking up and releasing fresh water in the polar regions and influences oceanic thermohaline circulation
- It stores enough water to significantly influence global sea level
- The continental ice sheets dominated by Antartica and Greenland are the most massive elements of the cryosphere
- Ice sheets are replenished by snowfall

Cryospheric component	Area	Mass
Antarctic ice sheet	2.7	53
Greenland ice sheet	0.35	5
Alpine glaciers	0.1	0.2
Arctic sea ice (March)	3	0.04
Antarctic sea ice (September)	4	0.04
Seasonal snow cover	9	<0.01
Permafrost	5	1

Area is expressed in percentage of the area of the surface of Earth; Mass is expressed in 10^3 kg/m^2
Total surface area of Earth (m^2)=5.12x10^{14}
Land area (m^2)=1.45x10^{14}

- Permafrost is any ground that remains completely frozen (0°C) or colder—for at least two years straight
- These permanently frozen grounds are most common in regions with high mountains and in Earth's higher latitudes—near the North and South Poles

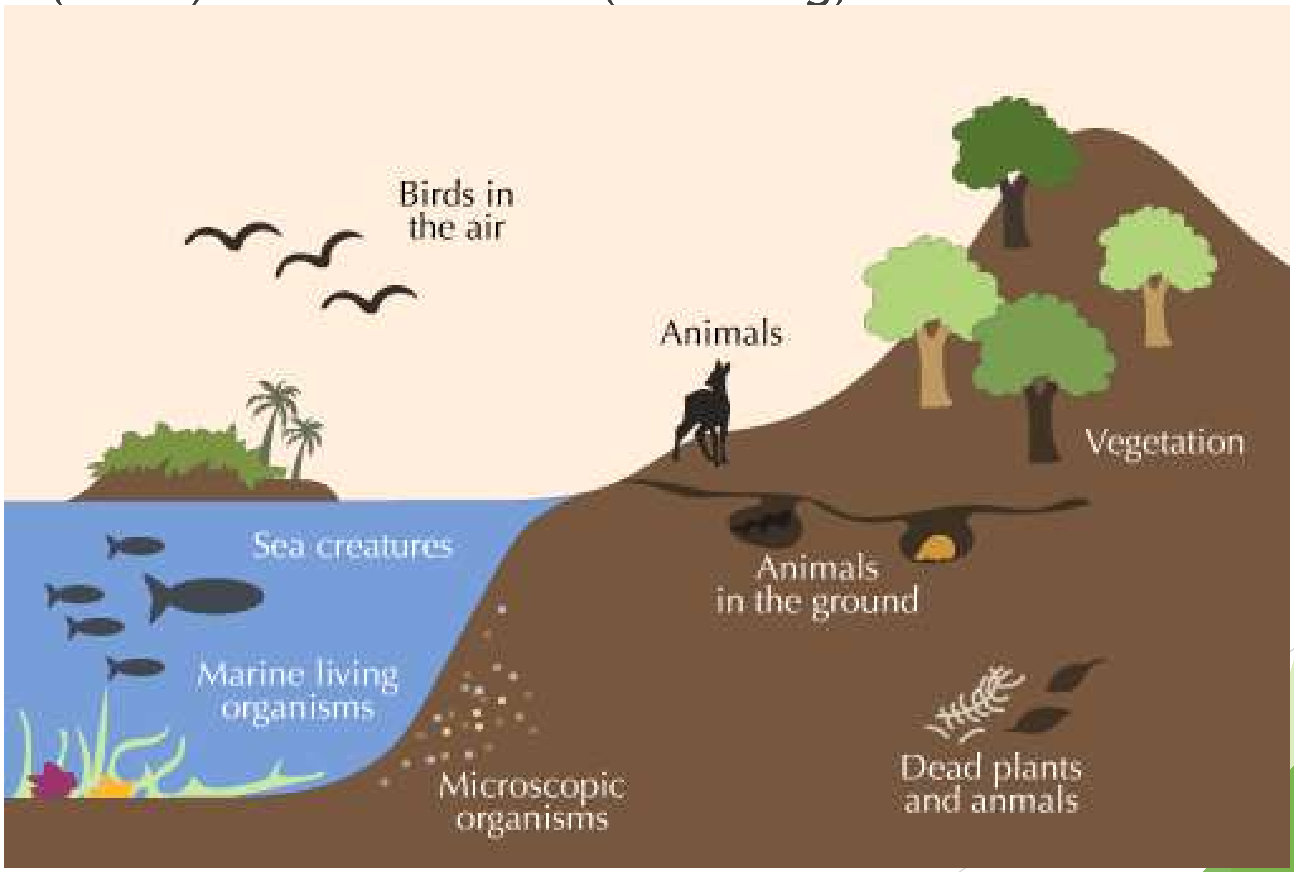
Exercise-2

Estimate how much the sea level would rise if the entire Arctic ice sheet were to melt. Area covered by Arctic sea ice is 3% of the area of the surface of the Earth, land area is 29.5% of the surface of Earth. [Earth's radius=6371 km; mass of Artic ice sheet=0.04×10^3 kg/m^2]

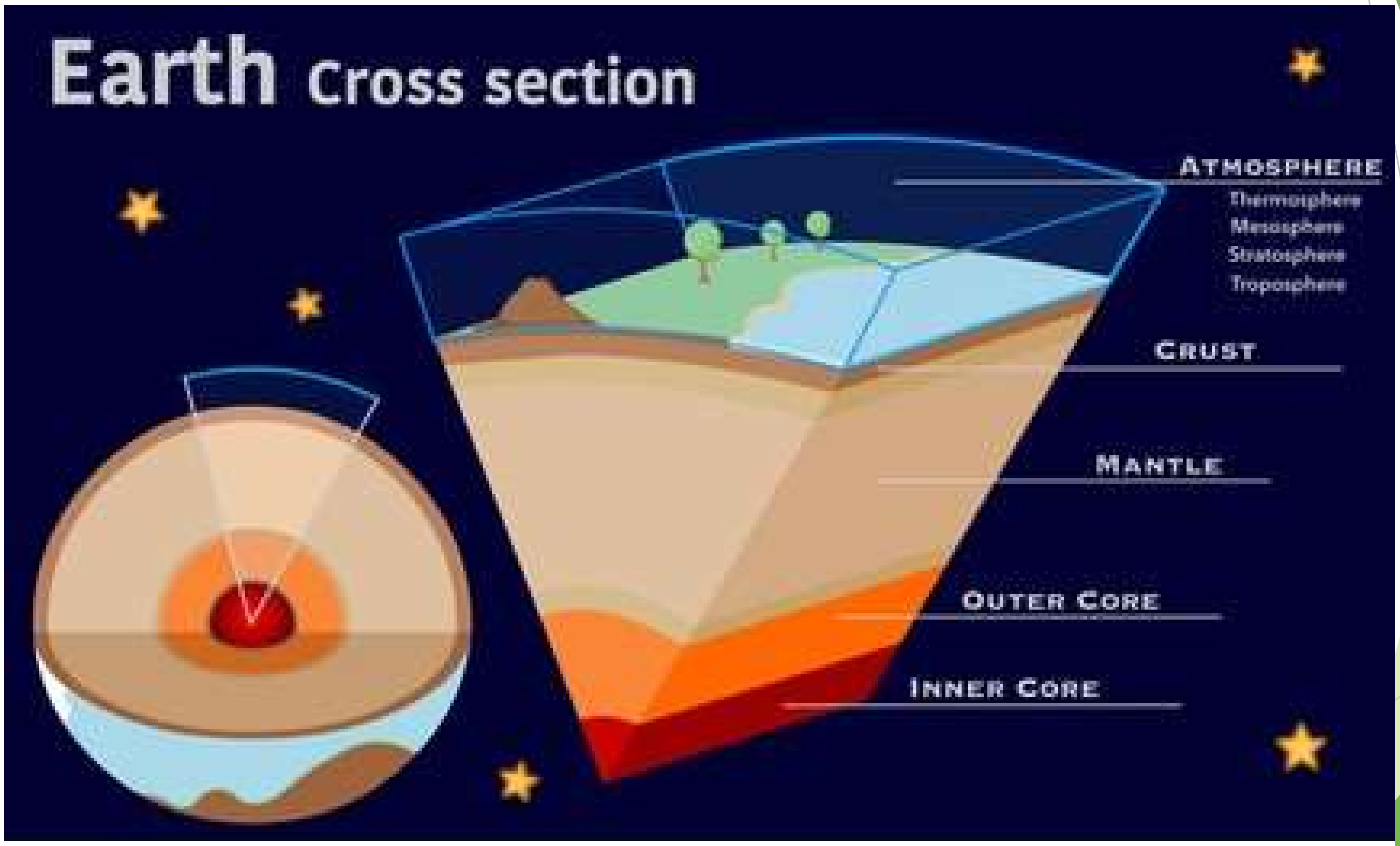
Estimate how much the sea level would rise if the entire permafrost were to melt. Area covered by permafrost is 5% of the area of the surface of the Earth, land area is 28.5% of the surface of Earth. [Earth's radius=6371 km; mass of Artic ice sheet=1×10^3 kg/m^2]

Biosphere

- The biosphere is a global ecosystem composed of living organisms (biota) and the abiotic (nonliving) factors



Earth’s crust and mantle



The Earth System

2

Climate depends not only on atmospheric processes, but also on physical, chemical, and biological processes involving other components of the Earth system. This chapter reviews the structure and behavior of those other components. We show how the cycling of water, carbon, and oxygen among the components of the Earth system has affected the evolution of the atmosphere. Drawing on this background, we summarize the history of climate over the lifetime of the Earth, with emphasis on causal mechanisms. The final section discusses why Earth is so much more habitable than its neighbors in the solar system. Chapter 10 revisits some of these same topics in the context of climate dynamics, with a quantitative discussion of feedbacks and climate sensitivity.

2.1 Components of the Earth System

This section introduces the cast of characters and briefly describes their roles and interrelations in the ongoing drama of climate. The atmosphere, which in some sense plays the starring role, has already been introduced in Chapter 1. The interplay between atmospheric radiation and convection regulates the temperature at the Earth's surface, setting the limits for snow and ice cover and for the various life zones in the biosphere. The stratospheric ozone layer protects the biosphere from the lethal effects of solar ultraviolet radiation. Atmospheric wind patterns regulate the patterns of oceanic upwelling that supplies nutrients to the marine biosphere, they determine the distribution of water that sustains the terrestrial (land) biosphere, and they transport trace gases, smoke, dust, insects, seeds, and spores over long distances. Rain, frost, and wind erode the Earth's crust, wearing down mountain ranges, reshaping the landscape, and replenishing the soils and the supply of metallic ions needed to sustain life.

Other components of the Earth system also play important roles in climate. The oceans are notable for their large "thermal inertia" and their central role in the cycling of carbon, which controls atmospheric carbon dioxide concentrations. Extensive snow and ice-covered surfaces render the Earth more reflective, and consequently cooler, than it would be in their absence. By evaporating large quantities of water through their leaves, land plants exert a strong moderating influence on tropical and extratropical summer climate. Living organisms on land and in the sea have been instrumental in liberating oxygen and sequestering of carbon in the Earth's crust, thereby reducing the atmospheric concentration carbon dioxide. On timescales of millions of years or longer, plate tectonics exerts an influence on climate through continental drift, mountain building, and volcanism. This section describes these processes and the media in which they occur.

2.1.1 The Oceans

The oceans cover 72% of the area of the Earth's surface and they reach an extreme depth of nearly 11 km. Their total volume is equivalent to that of a layer 2.6 km deep, covering the entire surface of the Earth. The mass of the oceans is ~250 times as large as that of the atmosphere.

a. Composition and vertical structure

The density of sea water is linearly dependent on the concentration of dissolved salt. On average, sea water in the open oceans contains ~35 g of dissolved salts per kg of fresh water, with values typically ranging from 34 to 36 g kg^{-1} (or parts per thousand by mass, abbreviated as o/oo). Due to the presence of these dissolved salts, sea water is ~2.4% denser than fresh water at the same temperature.

The density σ of sea water (expressed as the departure from 1 in g kg^{-1} or o/oo) typically ranges from 1.02 to 1.03. It is a rather complicated function of temperature T, salinity s, and pressure p; i.e., $\sigma = \sigma(T, s, p)$. The pressure dependence of density in liquids is much weaker than in gases and, for purposes of this qualitative discussion, will be ignored.[1] As in fresh water, $\partial\sigma/\partial T$ is temperature dependent, but the fact that sea water is saline makes the relationship somewhat different: in fresh water, density increases with increasing temperature between 0 and 4 °C, whereas in sea water, density decreases monotonically with increasing temperature.[2] In both fresh water and sea water, $\partial\sigma/\partial T$ is smaller near the freezing point than at higher temperatures. Hence, a salinity change of a prescribed magnitude δs is equivalent, in terms of its effect on density, to a larger temperature change δT in the polar oceans than in the tropical oceans, as illustrated in Fig. 2.1.

Over most of the world's oceans, the density of the water in the wind-stirred, *mixed layer* is smaller, by a few tenths of a percent, than the density of the water below it. Most of the density gradient tends to be concentrated within a layer called the *pycnocline*, which ranges in depth from a few tens of meters to a few hundred meters below the ocean surface. The density gradient within the pycnocline tends to inhibit vertical mixing in the ocean in much the same manner that the increase of temperature with height inhibits vertical mixing in atmospheric temperature inversions and in the stratosphere. In particular, the pycnocline strongly inhibits the exchange of heat and salt between the mixed layer, which is in direct contact with the atmosphere, and the deeper layers of the ocean. At lower latitudes, *pycnocline* is synonymous with the *thermocline* (i.e., the layer in which temperature increases with height), but in polar oceans, *haloclines* (layers with fresher water above and saltier water below) also play an important role in inhibiting vertical mixing. The strength and depth of the thermocline vary with latitude and season, as illustrated in the idealized profiles shown in Fig. 2.2.

Within the oceanic mixed layer, temperature and salinity (and hence density) vary in response to

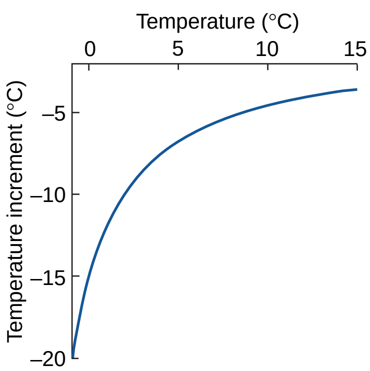


Fig. 2.1 The change in temperature of a water parcel required to raise the density of sea water at sea level as much as a salinity increase of 1 g kg^{-1}, plotted as a function of the temperature of the parcel. For example, for sea water at a temperature of 10 °C, a salinity increase of 1 g kg^{-1} would raise the density as much as a temperature decrease of ~5 °C, whereas for sea water at 0 °C the same salinity increase would be equivalent to a temperature change of ~17 °C. [Adapted from data in M. Winton, Ph.D. thesis, University of Washington, p. 124 (1993).]

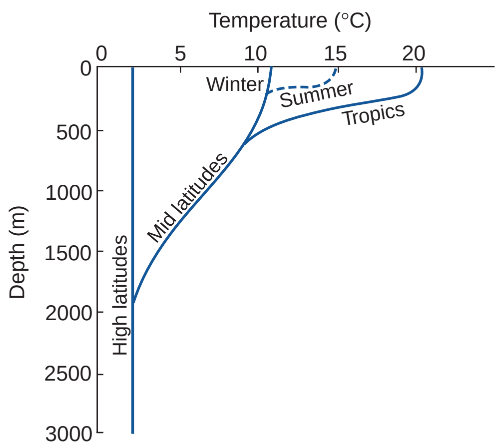


Fig. 2.2 Idealized profiles of the temperature plotted as a function of depth in different regions of the world's oceans. The layer in which the vertical temperature gradient is strongest corresponds to the thermocline. [From J. A. Knauss, *Introduction to Physical Oceanography*, 2nd Edition, p. 2, © 1997. Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ.]

[1] The small effect of pressure upon density is taken into account through the use of *potential density*, the density that a submerged water parcel would exhibit if it were brought up to sea level, conserving temperature and salinity. (See Exercise 3.54.)

[2] Ice floats on lakes because the density of fresh water decreases with temperature from 0 to 4 °C. In contrast, sea ice floats because water rejects salt as it freezes.

exchanges of heat and water with the atmosphere. Precipitation lowers the salinity by diluting the salts that are present in the oceanic mixed layer, and evaporation raises the salinity by removing fresh water and thereby concentrating the residual salts, as illustrated in the following example.

Exercise 2.1 A heavy tropical storm dumps 20 cm of rainfall in a region of the ocean in which the salinity is 35.00 g kg^{-1} and the mixed layer depth is 50 m. Assuming that the water is well mixed, by how much does the salinity decrease?

Solution: The volume of water in a column extending from the surface of the ocean to the bottom of the mixed layer is increased by a factor

$$\frac{0.2 \text{ m}}{50 \text{ m}} = 4 \times 10^{-3}$$

and (ignoring the small difference between the densities of salt water and fresh water) the mass of the water in the column increases by a corresponding amount. The mass of salt dissolved in the water remains unchanged. Hence, the salinity drops to

$$\frac{35.00 \text{ g of salt}}{1.004 \text{ kg of water}} = 34.86 \text{ g kg}^{-1}. \qquad ■$$

Water parcels that are not in contact with the ocean surface tend to conserve temperature and salinity as they move over long distances. Hence, *water masses* (layers of water extending over large areas that exhibit nearly uniform temperature and salinity) can be tracked back to the regions of the mixed layer in which they were formed by exchanges of heat and mass with the atmosphere. Among the important water masses in the Atlantic Ocean, in order of increasing density, are:

- *Mediterranean outflow*, which is conspicuously warm and saline due to the excess of evaporation over precipitation in the Mediterranean Sea.
- *North Atlantic deep water (NADW)*, formed by the sinking of water along the ice edge in the Greenland, Iceland, and Norwegian Seas.
- *Antarctic bottom water (AABW)*, formed by sinking along the ice edge in the Weddell Sea.

The NADW and AABW, each marked by its own distinctive range of temperatures and salinities, are

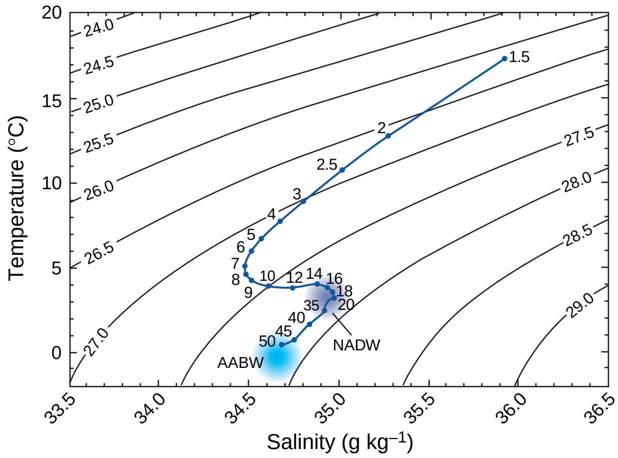


Fig. 2.3 Vertical sounding of water temperature and salinity in a vertical sounding in the subtropical Atlantic Ocean. Numbers along the sounding indicate depths in hundreds of meters. Potential (i.e., pressure-adjusted) density (in *o/oo*) is indicated by the contours. Characteristic temperature and salinity ranges for North Atlantic deep water (NADW) and Antarctic bottom water (AABW) are indicated by shading. [Reprinted from *Seawater: Its Composition, Properties and Behavior*, The Open University in association with Pergamon Press, p. 48 (1989), with permission from Elsevier.]

both clearly evident near the bottom of the tropical sounding shown in Fig. 2.3. The AABW is slightly colder and fresher than the NADW. When both temperature and salinity are taken into account, the AABW is slightly denser than the NADW, consistent with its placement at the bottom of the water column.

b. The ocean circulation

The ocean circulation is composed of a *wind-driven* component and a *thermohaline* component. The wind-driven circulation dominates the surface currents, but it is largely restricted to the topmost few hundred meters. The circulation deeper in the oceans is dominated by the slower thermohaline circulation.

By generating ocean waves, surface winds transfer horizontal momentum from the atmosphere into the ocean. The waves stir the uppermost layer of the ocean, mixing the momentum downward. The momentum, as reflected in the distribution of surface currents shown in Fig. 2.4, mirrors the pattern of surface winds shown in Figs. 1.18 and 1.19, with closed anticyclonic circulations (referred to as *gyres*) at subtropical latitudes and cyclonic gyres at subpolar latitudes. Another notable feature of the wind-driven

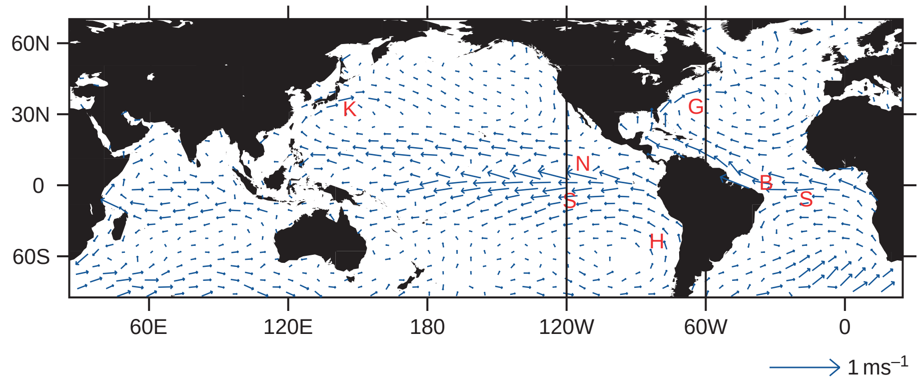


Fig. 2.4 Annual mean ocean surface currents based on the rate of drift of ships. The *Gulf Stream* (G) and the *Kuroshio Current* (K) are warm, *western boundary currents*. The *Humboldt Current* (H) is the most prominent of the cold, equatorward currents driven by the winds along the eastern flanks of the subtropical anticyclones. The westward *South Equatorial Current* (S) is driven by the easterlies along the equator and the weaker eastward *North Equatorial Countercurrent* (N) is a response to the winds in the vicinity of the ITCZ. [Data courtesy of Philip Richardson, WHOI; graphic courtesy of Todd P. Mitchell.]

circulation is the west-to-east *Antarctic circumpolar current* along 55 °S, the latitude of the Drake passage that separates Antarctica and South America. Velocities in these wind-driven currents are typically on the order of 10 cm s^{-1}, a few percent of the speeds of the surface winds that drive them, but in the narrow *western boundary currents* such as the *Gulf Stream* off the east coast of the United States (Figs. 2.4 and 2.5) velocities approach 1 m s^{-1}. The relatively warm water transported poleward by the western boundary currents contributes to moderating winter temperatures over high latitude coastal regions.

Over certain regions of the polar oceans, water in the mixed layer can become sufficiently dense, by virtue of its high salinity, to break through the pycnocline and sink all the way to the ocean floor to become what oceanographers refer to as *deep water* or *bottom water*. In some sense, these negatively buoyant plumes are analogous to the plumes of warm, moist air in low latitudes that succeed in breaking through the top of the atmospheric mixed layer and continue ascending until they encounter the tropopause. The presence of CFCs[3] in NADW and AABW indicates that these water

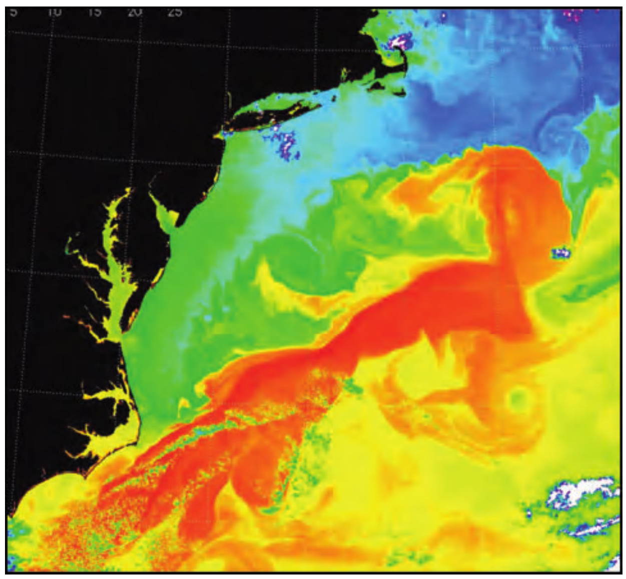


Fig. 2.5 Eddies along the landward edge of the Gulf Stream, as revealed by the pattern of sea surface temperature. Temperatures range from ~20 °C in the orange regions down to ~6 °C in the darkest blue regions. Note the sharpness of the boundary and the indications of turbulent mixing between the waters of the Gulf Stream and the colder Labrador Current to the north of it. [Based on NASA Terra/MODIS imagery. Courtesy of Otis Brown.]

[3] The term *chlorofluorocarbons (CFCs)* refers to a family of gaseous compounds that have no natural sources; first synthesized in 1928. Atmospheric concentrations of CFCs rose rapidly during the 1960s and 1970s as these gases began to be used for a widening range of purposes.

masses were in relatively recent contact with the atmosphere.

By virtue of their distinctive chemical and isotopic signatures, it is possible to track the flow of water masses and to infer how long ago water in various parts of the world's oceans was in contact with the atmosphere. Such chemical analyses indicate the existence of a slow overturning characterized by a spreading of deep water from the high latitude sinking regions, a resurfacing of the deep water, and a return flow of surface waters toward the sinking regions, as illustrated in Fig. 2.6. The timescale in which a parcel completes a circuit of this so-called *thermohaline circulation* is on the order of hundreds of years.

The resurfacing of deep water in the thermohaline circulation requires that it be *ventilated* (i.e., mixed with and ultimately replaced by less dense water that has recently been in contact with the ocean surface). Still at issue is just how this ventilation occurs in the presence of the pycnocline. One school of thought attributes the ventilation to mixing along sloping isopycnal (constant density) surfaces that cut through the pycnocline. Another school of thought attributes it to irreversible mixing produced by tidal motions propagating downward into the deep oceans along the continental shelves, and yet another to vertical mixing in restricted regions characterized by strong winds and steeply sloping isopycnal surfaces, the most important of which coincides with the *Antarctic circumpolar current*, which lies beneath the ring of strong westerly surface winds that encircles Antarctica.

Although most of the deep and bottom water masses are formed in the Atlantic sector, the thermohaline circulation involves the entire world's oceans, as illustrated in Fig. 2.7. Within the Atlantic sector itself, the thermohaline circulation is comprised of two different cells: one involving NADW and the other involving AABW, as illustrated in Fig. 2.8.

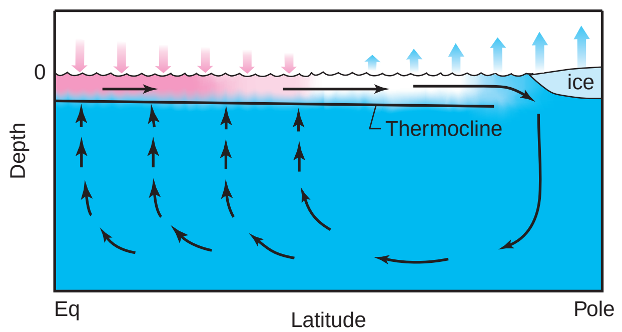


Fig. 2.6 Idealized schematic of the thermohaline circulation in an equatorially symmetric ocean. The domain extends from the sea floor to the ocean surface and from equator to pole. Pink shading indicates warmer water and blue shading indicates colder water. The shaded arrows represent the exchange of energy at the air–sea interface: pink downward arrows indicate a heating of the ocean mixed layer and blue upward arrows indicate a cooling. The role of salinity is not specifically represented in this schematic but it is the rejection of salt when water freezes along the ice edge that makes the water dense enough to enable it to sink to the bottom.

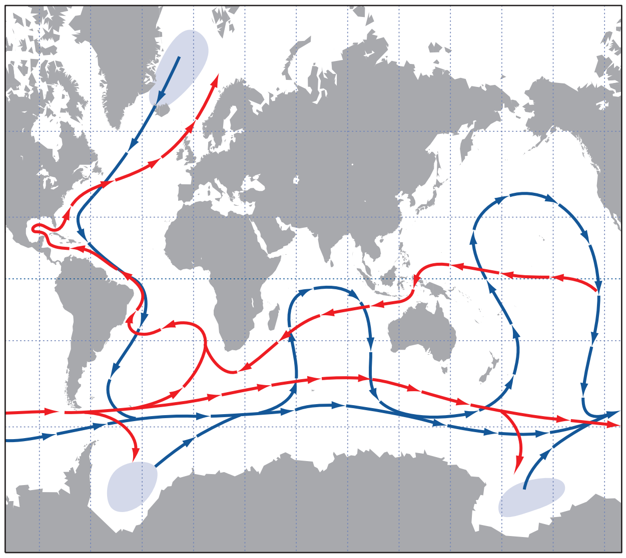


Fig. 2.7 Highly simplified schematic of the thermohaline circulation. Shading denotes regions of downwelling, blue arrows denote transport of bottom water, and red arrows denote the return flow of surface water. [Adapted from W. J. Schmitz, Jr., "On the interbasin-scale thermohaline circulation," *Rev. Geophys.*, **33**, p. 166, Copyright 1995 American Geophysical Union.]

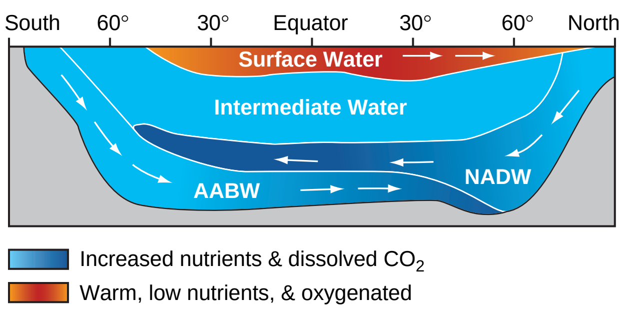


Fig. 2.8 Idealized cross section of the thermohaline circulation in the Atlantic Ocean. In this diagram, *Intermediate Water* comprises several different water masses formed at temperate latitudes. Note the consistency with Fig. 2.3. [Courtesy of Steve Hovan.]

c. The marine biosphere

Virtually all the sunlight that reaches the surface of the ocean is absorbed within the topmost hundred meters. Within this shallow *euphotic zone*,[4] life abounds wherever there are sufficient nutrients, such as phosphorous and iron, to sustain it. In regions of the ocean where the marine biosphere is active, the uppermost layers are enriched in dissolved oxygen (a product of photosynthesis) and depleted in nutrients and dissolved carbon, as illustrated in Fig. 2.9. *Phyto-* (i.e., plant) *plankton* are capable of consuming the nutrients in the euphotic zone within a matter of days. Hence, the maintenance of high *primary productivity* (i.e., photosynthesis) requires a continual supply of nutrients. The most productive regions of the oceans tend to be concentrated in regions of upwelling, where nutrient-rich sea water from below the euphotic zone is first exposed to sunlight.

Nutrients consumed within the euphotic zone by phytoplankton return to the deeper layers of the oceans when marine plants and animals that feed on them die, sink, and decompose. The continual exchange of nutrients between the euphotic zone and the deeper layers of the ocean plays an important role in the carbon cycle, as discussed in Section 2.3. The distribution of upwelling, in turn, is controlled by the pattern of surface winds discussed earlier. The distribution of *ocean color* (Fig. 2.10) shows evidence of high biological productivity and, by inference, upwelling

- beneath cyclonic circulations such as Aleutian and Icelandic lows,
- along the eastern shores of the oceans at subtropical latitudes,
- in a narrow strip along the equator in the equatorial Atlantic and Pacific Oceans.

In contrast, the ocean regions that lie beneath the subtropical anticyclones are biological deserts. The dynamical basis for these relationships is discussed in Section 7.2.5. Through their effect in mediating the geographical distribution of upwelling and the depth of the mixed layer, year-to-year changes in the atmospheric circulation, such as those that occur in association with El Niño, perturb the entire food chain that supports marine mammals, seabirds, and commercial fisheries.

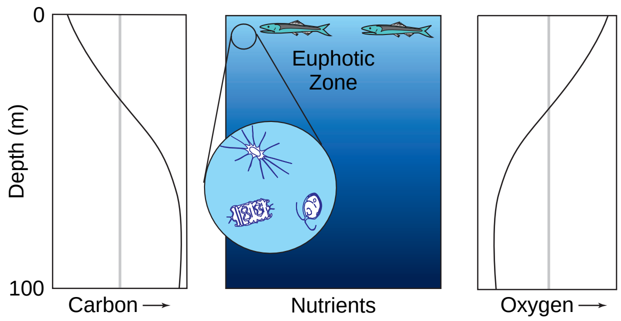


Fig. 2.9 Idealized vertical profiles of dissolved carbon (left) and oxygen (right) in biologically active regions of the oceans. The intensity of sunlight is indicated by the depth of the shading in the middle panel.

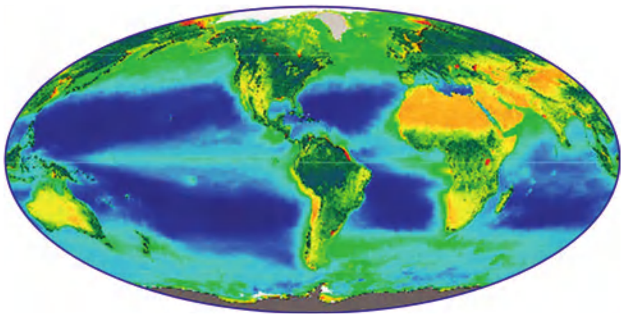


Fig. 2.10 Distribution of primary productivity in the marine and terrestrial biosphere, averaged over a 3-year period. Over the oceans the dark blue areas are indicative of very low productivity and the green and yellow areas are relatively more productive. Over land dark green is indicative of high productivity. [Imagery courtesy of SeaWiFS Project, NASA/GSFC and ORBIMAGE, Inc.]

d. Sea surface temperature

The global distribution of sea surface temperature is shaped by both radiative and dynamical factors relating to the pattern of seasonally varying, climatological-mean surface wind field over the oceans (Fig. 1.18). Radiative heating is the dominant factor. That incident solar radiation is so much stronger in the tropics than in the polar regions gives rise to a strong north–south temperature gradient, which dominates the annual-mean field shown in Fig. 2.11 (top).

The effects of the winds on the sea surface temperature pattern become more clearly apparent when the zonally averaged sea surface temperature at each latitude is removed from the total field, leaving just

[4] Greek: *eu*-good and *photic*-light.

the departures from the zonal-mean, shown in Fig. 2.11 (bottom). The coolness of the eastern oceans relative to the western oceans at subtropical latitudes derives from circulation around the subtropical anticyclones (Fig. 1.16). The equatorward flow of cool air around the eastern flanks of the anticyclones extracts a considerable quantity of heat from the ocean surface, as explained in Section 9.3.4, and drives cool, southward ocean currents (Fig. 2.4). In contrast, the warm, humid poleward flow around their western flanks extracts much less heat and drives warm western boundary currents such as the Gulf Stream. At higher latitudes the winds circulating around the subpolar cyclones have the opposite effect, cooling the western sides of the oceans and warming the eastern sides. The relative warmth of the eastern Atlantic at these higher latitudes is especially striking.

Wind-driven upwelling is responsible for the relative coolness of the equatorial eastern Pacific and Atlantic, where the southeasterly trade winds protrude northward across the equator (Fig. 1.18). Wind-driven upwelling along the coasts of Chile, California, and continents that occupy analogous positions with respect to the subtropical anticyclones, although not well resolved in Fig. 2.11, also contributes to the coolness of the subtropical eastern oceans, as do the highly reflective cloud layers that tend to develop at the top of the atmospheric boundary layer over these regions (Section 9.4.4).

The atmospheric circulation feels the influence of the underlying sea surface temperature pattern, particularly in the tropics. For example, from a comparison of Figs. 1.25 and 2.11 it is evident that the intertropical convergence zones in the Atlantic and Pacific sectors are located over bands of relatively warm sea surface temperature and that the dry zones lie over the *equatorial cold tongues* on the eastern sides of these ocean basins.

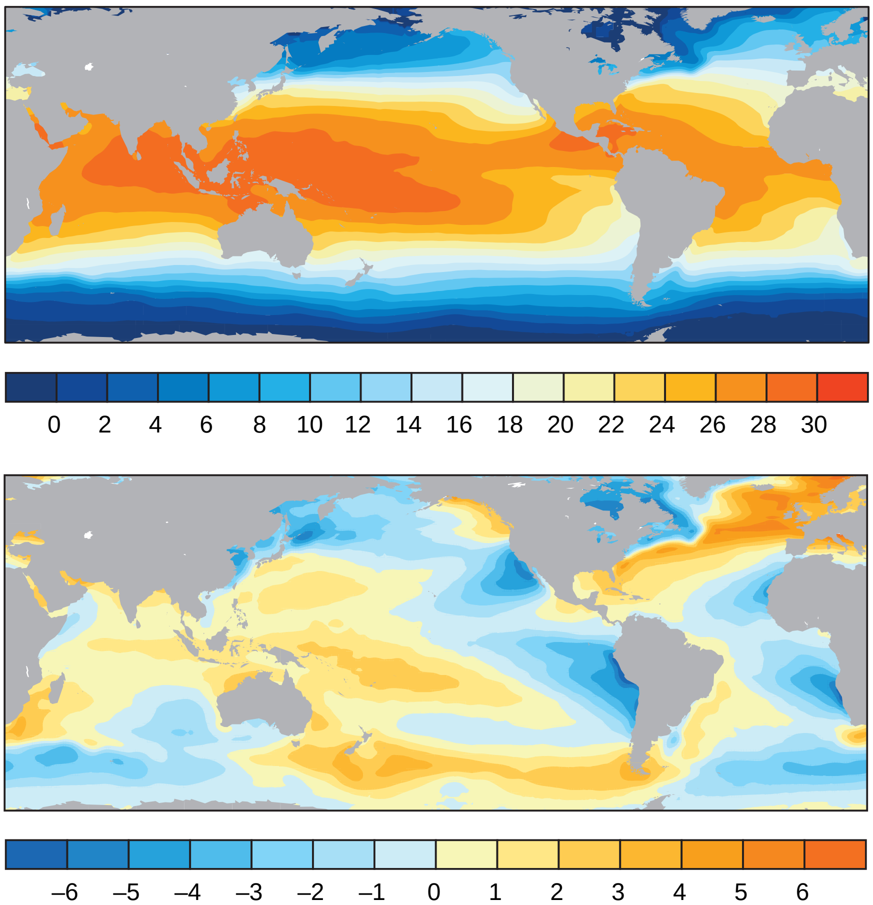


Fig. 2.11 Annual mean sea surface temperature. (Top) The total field. (Bottom) Departure of the local sea surface temperature at each location from the zonally average field. [Based on data from the U.K. Meteorological Office HadISST dataset. Courtesy of Todd P. Mitchell.]

2.1.2 The Cryosphere

The term *cryo-* (frozen) *sphere* refers to components of the Earth system comprised of water in its solid state, or in which frozen water is an essential component. The cryosphere contributes to the thermal inertia of the climate system; it contributes to the reflectivity or *albedo* of the Earth; by taking up and releasing fresh water in the polar regions, it influences oceanic thermohaline circulation; and it stores enough water to significantly the influence global sea level. The elements of the cryosphere are listed in Table 2.1 and all of them, with the exception of alpine glaciers, are represented in Fig. 2.12.

The *continental ice sheets*, dominated by Antarctica and Greenland, are the most massive elements of the cryosphere. The ice sheets are continually replenished by snowfall; they lose mass by sublimation, by the calving of icebergs, and, in summer, by runoff in streams and rivers along their periphery. The net *mass balance* (i.e., the balance between the mass sources and sinks) at any given time determines whether an ice sheet is growing or shrinking.

Over periods of tens of thousands of years and longer, annual layers of snow that fall in the relatively flat interior of the ice sheets are compressed by the accumulation of new snow on top of them. As the pressure increases, snow is transformed into ice. Due to the dome-like shape of the ice sheets and the plasticity of the ice itself, the compressed layers of ice gradually creep downhill toward the periphery of the ice sheet, causing the layer as a whole to spread out horizontally and (in accordance with the conservation of mass) to thin in the vertical dimension. Much of the flow toward the periphery tends to be concentrated in relatively narrow, fast-moving ice streams tens of kilometers in width (Fig. 2.13).

Table 2.1 Surface area and mass of the various components of the cryosphere[a]

Cryospheric component	Area	Mass
Antarctic ice sheet	2.7	53
Greenland ice sheet	0.35	5
Alpine glaciers	0.1	0.2
Arctic sea ice (March)	3	0.04
Antarctic sea ice (September)	4	0.04
Seasonal snow cover	9	<0.01
Permafrost	5	1

[a] Surface area is expressed as percentage of the area of the surface of the Earth. Mass is expressed in units of 10^3 kg m^{-2} (numerically equivalent to meters of liquid water) averaged over the entire surface area of the Earth. For reference, the total surface area of the Earth and the area of the Earth covered by land are 5.12 and 1.45×10^{14} m^2, respectively. [Courtesy of S. G. Warren.]

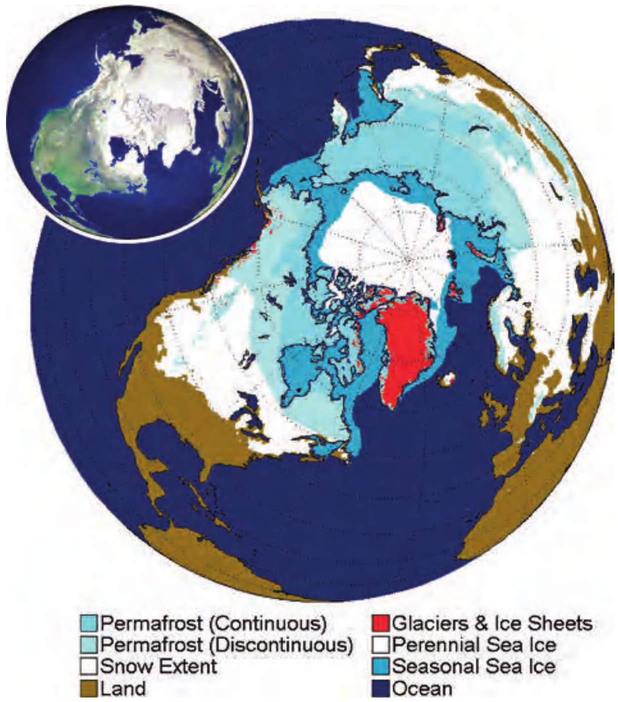


Fig. 2.12 Elements of the northern hemisphere cryosphere. The equatorward edge of the snow cover corresponds to ~50% coverage during the month of maximum snow extent. [Courtesy of Ignatius Rigor.] The inset at the upper left shows a NASA RADARSAT image highlighting these features.

Along the divides of the ice sheets the movement is very slow and the layering of the ice is relatively undisturbed. In *ice cores* extracted from these regions, the age of the ice increases monotonically with depth to ~100,000 years in the Greenland ice sheet and over 500,000 years in the Antarctic ice sheet. Analysis of air bubbles, dust, and chemical and biological tracers embedded within these ice cores is providing a wealth of information on the climate of the past few hundred thousand years, as discussed later in this chapter.

In many respects, *alpine* (i.e., mountain) *glaciers* behave like continental ice sheets, but they are much smaller in areal coverage and mass. Their fate is also determined by their mass balance. Parcels of ice within them flow continually downhill from an upper dome-like region where snow and ice accumulate toward their snouts where mass is lost continually due to melting. Because of their much smaller masses, glaciers respond much more quickly to climate change than continental ice sheets, and ice cycles through them much more rapidly. Some alpine glaciers also exhibit time-dependent behavior that is not climate

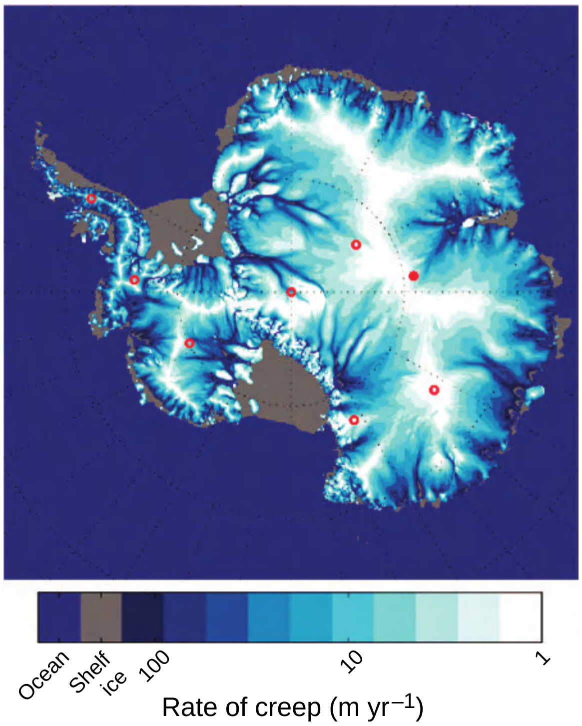


Fig. 2.13 Satellite image of the Antarctic ice sheet showing rate of creep of the ice (in m year^{-1}) on a logarithmic scale. Dots show the locations of ice core sites. Vostok, the site of the ice core shown in Fig. 2.31, is indicated by the solid red dot. [Adapted with permission from Bamber, J. L., D. G. Vaughan and I. Joughin, "Widespread Complex Flow in the Interior of the Antarctic Ice Sheet," *Science*, **287**, 1248–1250. Copyright 2000 AAAS. Courtesy of Ignatius Rigor.]

related: episodic surges of a few months' to a few years' duration interspersed with much longer periods of slow retreat.

Sea ice covers a larger area of the Earth's surface area than the continental ice sheets (Table 2.1) but, with typical thicknesses of only 1–3 m, is orders of magnitude less massive. The ice is not a continuous surface, but a fractal field comprised of ice floes (pieces) of various of shapes and sizes, as shown in Figs. 2.14 and 2.15. The individual floes are separated by patches of open water (called *leads*) that open and close as the ice pack moves, dragged by surface winds.

Seasonal limits of the northern hemisphere pack ice are shown in Fig. 2.12. During winter, ice covers not only the Arctic, but also much of the Bering Sea and the Sea of Okhotsk, but during the brief polar summer the ice retreats dramatically and large leads are sometimes observed, even in the vicinity of the North Pole. Antarctic pack ice also advances and retreats with the seasons.

The annual-mean sea ice motion, shown in Fig. 2.16, is dominated by the clockwise *Beaufort Gyre* to the north of Alaska and the *transpolar drift stream* from Siberia toward Greenland and Spitzbergen.[5] Some ice floes remain in the Arctic for a decade or more, circulating around and around the Beaufort Gyre, whereas others spend just a year or two in the Arctic before they exit either through the Fram Strait between Greenland and Spitzbergen or through the Nares Strait into Baffin Bay along the west side of Greenland. Ice floes exiting the Arctic make a one-way trip into warmer waters, where they are joined by much thicker *icebergs* that break off the Greenland ice sheet.

New pack ice is formed during the cold season by the freezing of water in newly formed leads and in regions where offshore winds drag the pack ice away from the coastline, exposing open water. The new ice thickens rapidly at first and then more gradually as it begins to insulate the water beneath it from the subfreezing air above. Ice thicker than a meter is formed, not by a thickening of newly formed layer of

[5] The existence of a transpolar drift stream was hypothesized by Nansen[6] when he learned that debris from a shipwreck north of the Siberian coast had been recovered, years later, close to the southern tip of Greenland. Motivated by this idea, he resolved to sail a research ship as far east as possible off the coast of Siberia and allow it to be frozen into the pack ice in the expectation that it would be carried across the North Pole along the route suggested by Fig. 2.16. He supervised the design and construction of a research vessel, the *Fram* ("Forward"), with a hull strong enough to withstand the pressure of the ice. The remarkable voyage of the *Fram*, which began in summer of 1893 and lasted for 3 years, confirmed the existence of the transpolar drift stream and provided a wealth of scientific data.

[6] **Fridtjof Nansen** (1861–1930). Norwegian scientist, polar explorer, statesman, and humanitarian. Educated as a zoologist. Led the first traverse of the Greenland ice cap on skis in 1888. The drift of his research vessel the *Fram* across the Arctic (1893–1896) was hailed as a major achievement in polar research and exploration. Midway through this voyage, Nansen turned over command of the *Fram* to Harald Sverdrup and set out with a companion on what proved to be a 132-day trek across the pack ice with dog-drawn sledges and kayaks, reaching 86 °N before adverse conditions forced them to turn southward.

Sacrificed his subsequent aspirations for Antarctic exploration to serve the needs of his country and to pursue humanitarian concerns. Was instrumental in peacefully resolving a political dispute between Norway and Sweden in 1905–1906 and negotiating a relaxation of an American trade embargo that threatened Norwegian food security during World War I. Awarded the Nobel Peace Prize in 1922 in recognition of his extensive efforts on behalf of war refugees and famine victims.

Fig. 2.14 Ice floes and leads in Antarctic pack ice. The lead in the foreground is 4–5 m across. The floe behind it consists of multi-year ice that may have originated as an iceberg; it is unusually thick, extending from ~15 m below to ~1 m above sea level. Most of the portion of the floe that extends above sea level is snow. At the time this picture was taken, the pack ice in the vicinity was under lateral pressure, as evidenced by the fact that a *pressure ridge* had recently developed less than 100 m away. [Photograph courtesy of Miles McPhee.]

ice, but by mechanical processes involving collisions of ice floes. *Pressure ridges* up to 5 m in thickness are created when floes collide, and thickening occurs when part of one floe is pushed or *rafted* on top of another.

When sea water freezes, the ice that forms is composed entirely of fresh water. The concentrated salt water known as *brine* that is left behind mixes with the surrounding water, increasing its salinity. Brine rejection is instrumental in imparting enough negative buoyancy to parcels of water to enable them to break through the pycnocline and sink to the bottom. Hence, it is no accident that the sinking regions in the oceanic thermohaline circulation are in high latitudes, where sea water freezes.

Land snow cover occupies an even larger area of the northern hemisphere than sea ice and it varies much more widely from week to week and month to month than does sea ice. With the warming of the land surface during spring, the snow virtually disappears, except in the higher mountain ranges.

Permafrost embedded in soils profoundly influences terrestrial ecology and human activities over large areas of Siberia, Alaska, and northern Canada. If the atmosphere and the underlying land surface

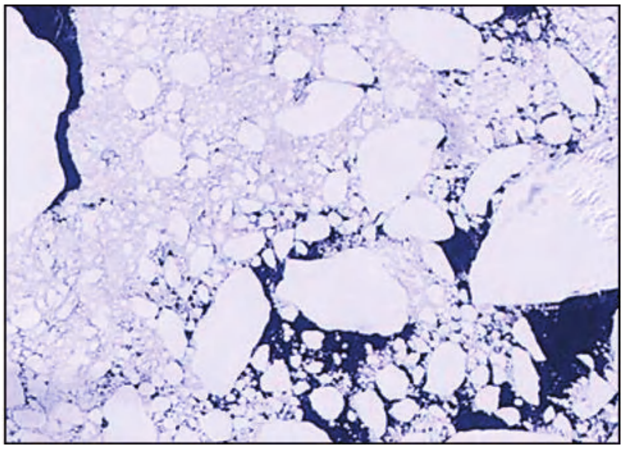


Fig. 2.15 Floes in pack ice streaming southward off the east coast of Greenland. The white area at the upper left is *landfast ice* that is attached to the coast, and the black channel adjacent to it is open water, where the mobile pack ice has become detached from the landfast ice. [NASA MODIS imagery.]

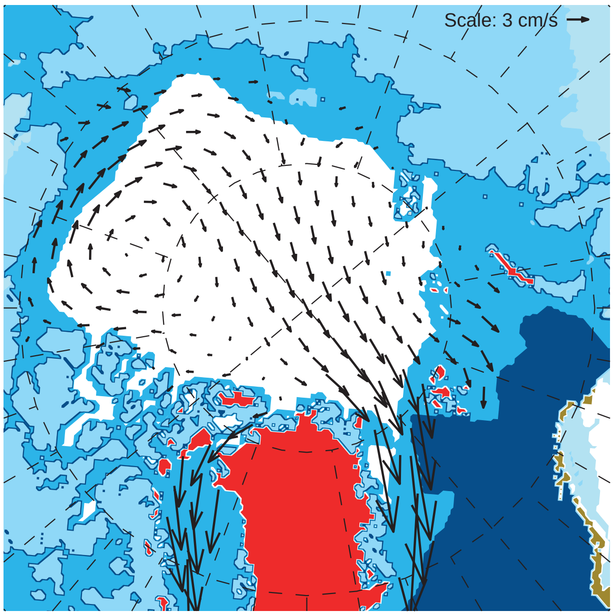


Fig. 2.16 Wintertime Arctic sea ice motion as inferred from the tracks of an array of buoys dropped on ice floes by aircraft. [Courtesy of Ignatius Rigor.]

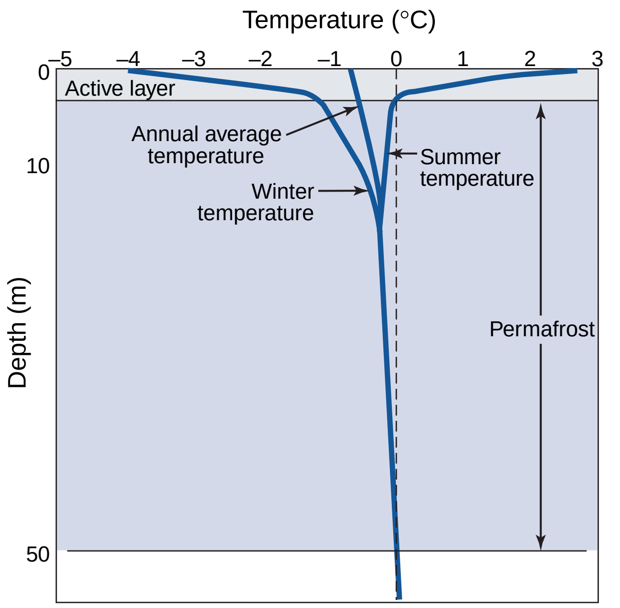


Fig. 2.17 Schematic vertical profile of summer and winter soil temperatures in a region of permafrost. The depth of the permafrost layer varies from as little as a few meters in zones of intermittent permafrost to as much as 1 km over the coldest regions of Siberia.

were in thermal equilibrium, the zones of continuous and intermittent permafrost in Fig. 2.12 would straddle the 0 °C isotherm in annual-mean surface air temperature. There is, in fact, a close correspondence between annual-mean surface air temperature and the limit of continuous permafrost, but the critical value of surface air temperature tends to be slightly above 0 °C due to the presence of snow cover, which insulates the land surface during the cold season, when it is losing heat.

Even in the zone of continuous permafrost, the topmost few meters of the soil thaw during summer in response to the downward diffusion of heat from the surface, as shown in Fig. 2.17. The upward diffusion of heat from the Earth's interior limits the vertical extent of the permafrost layer. Because the molecular diffusion of heat in soil is not an efficient heat transfer mechanism, hundreds of years are required for the permafrost layer to adjust to changes in the temperature of the overlying air.

2.1.3 The Terrestrial Biosphere

Much of the impact of climate upon animals and humans is through its role in regulating the condition and geographical distribution of forests, grasslands, tundra, and deserts, elements of the *terrestrial* (land) *biosphere*. A simple conceptual framework for relating climate (as represented by annual-mean temperature and precipitation) and vegetation type is shown in Fig. 2.18. The boundary between tundra and forest corresponds closely to the limit of the permafrost zone, which, as noted earlier, is determined by annual-mean temperature. The other boundaries in Fig. 2.18 are determined largely by the water requirements of plants. Plants utilize water both as raw material in producing chlorophyll and to keep cool on hot summer days, as described later. Forests require more water than grasslands, and grasslands, in turn, require more water than desert vegetation. The water demands of any specified type of vegetation increase with temperature.

Biomes are geographical regions with climates that favor distinctive combinations of plant and animal species. For example, tundra is the dominant form of vegetation in regions in which the mean temperature of the warmest month is ≤10 °C, and sparse, desert vegetation prevails in regions in which potential evaporation (proportional to the quantity of solar radiation reaching the ground) exceeds precipitation. The global distribution of biomes is determined by the *insolation* (i.e., the incident solar radiation) at the top of the atmosphere and by the climatic variables:

- annual-mean temperature,
- the annual and diurnal temperature ranges,
- annual-mean precipitation, and
- the seasonal distributions of precipitation and cloudiness.

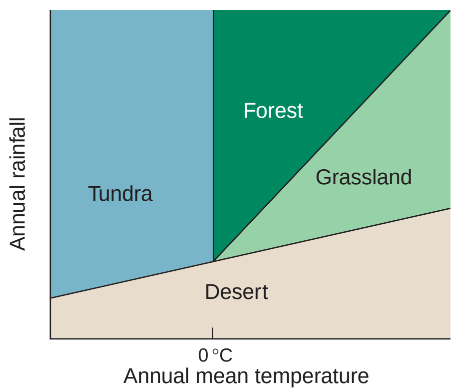


Fig. 2.18 A conceptual framework for understanding how the preferred types of land vegetation over various parts of the globe depend on annual-mean temperature and precipitation.

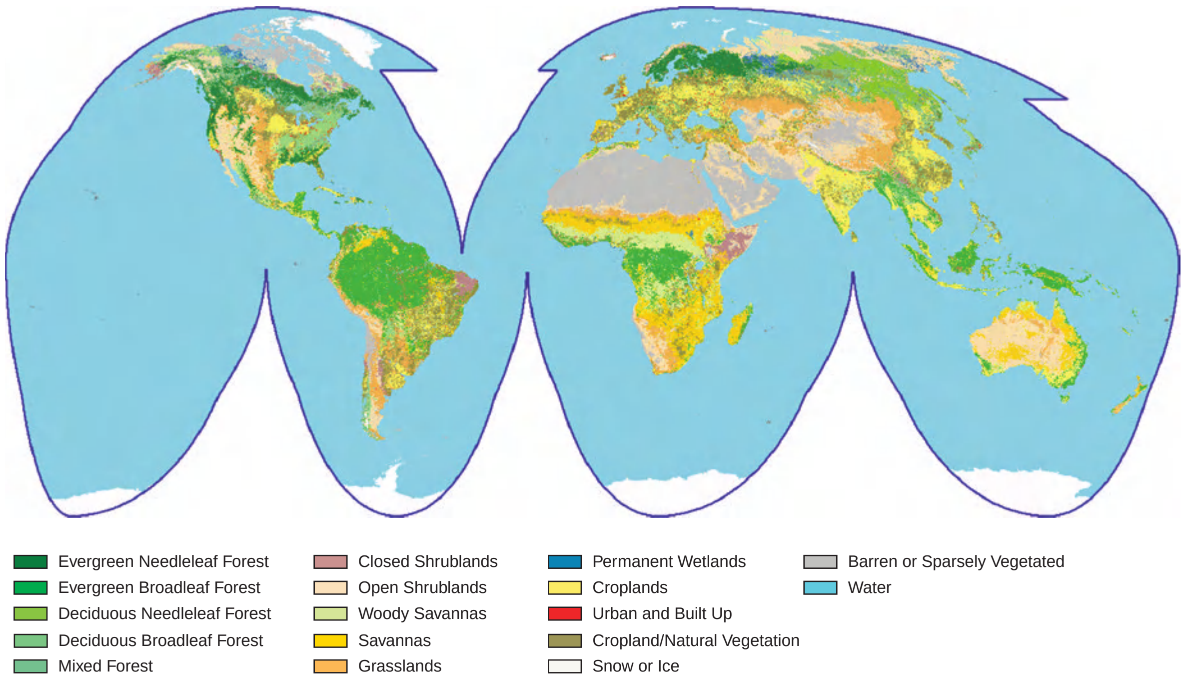


Fig. 2.19 Global land cover characterization, as inferred from NASA AVHRR NDVI satellite imagery and ground-based data relating to ecological regions, soils, vegetation, land use, and land cover. [From USGS Land Processes DAAC.]

Insolation and climate at a given location, in turn, are determined by latitude, altitude, and position with reference to the land–sea configuration and terrain. The combined influence of altitude upon temperature (Fig. 1.9), terrain upon precipitation (Fig. 1.25), and local terrain slope upon the incident solar radiation (Exercise 4.16) gives rise to a variegated distribution of biomes in mountainous regions.

Several different systems exist for assigning biomes, each of which consists of a comprehensive set of criteria that are applied to the climate statistics for each geographical location.[7] The "ground truth" for such classification schemes is the observed distribution of land cover, as inferred from ground-based measurements and high-resolution satellite imagery. An example is shown in Fig. 2.19.

The state of the terrestrial biosphere feeds back upon the climate through its effects on

- the hydrologic cycle: for example, during intervals of hot weather, plants control their temperatures by *evapo-transpiration* (i.e., by giving off water vapor through their leaves or needles). Energy derived from absorbed solar radiation that would otherwise contribute to heating the land surface is used instead to evaporate liquid water extracted from the soil by the roots of the plants. In this manner, the solar energy is transferred to the atmosphere without warming the land surface. Hence, on hot summer days, grass-covered surfaces tend to be cooler than paved surfaces and vegetated regions do not experience as high daily maximum temperatures as deserts and urban areas.
- the local albedo (the fraction of the incident solar radiation that is reflected, without being absorbed): for example, snow-covered tundra is more reflective, and therefore cooler during the daytime, than a snow-covered forest.
- the roughness of the land surface: wind speeds in the lowest few tens of meters above the ground tend to be higher over bare soil and tundra than over forested surfaces.

[7] These systems are elaborations of a scheme developed by Köppen[8] a century ago.

[8] **Wladimir Peter Köppen** (1846–1940) German meteorologist, climatologist, and amateur botanist. His Ph.D. thesis (1870) explored the effect of temperature on plant growth. His climate classification scheme, which introduced the concept of biomes, was published in 1900. For many years, Köppen's work was better known to physical geographers than to atmospheric scientists, but in recent years it is becoming more widely appreciated as a conceptual basis for describing and modeling the interactions between the atmosphere and the terrestrial biosphere.

2.1.4 The Earth's Crust and Mantle

The current configuration of continents, oceans, and mountain ranges is a consequence of plate tectonics and continental drift.[9] The Earth's crust and mantle also take part in chemical transformations that mediate the composition of the atmosphere on timescales of tens to hundreds of millions of years.

The Earth's crust is broken up into plates that float upon the denser and much thicker layer of porous but viscous material that makes up the Earth's mantle. Slow convection within the mantle moves the plates at speeds ranging up to a few centimeters per year (tens of kilometers per million years). Plates that lie above regions of upwelling in the mantle are spreading, whereas plates that lie above regions of downwelling in the mantle are being pushed together. Earthquakes tend to be concentrated along plate boundaries.

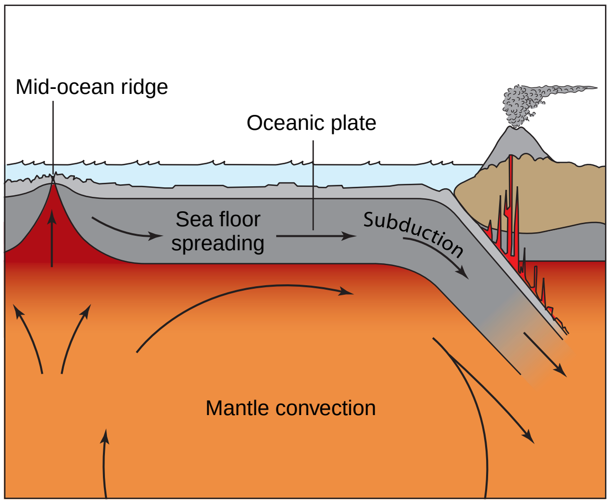


Fig. 2.20 Schematic showing subduction, sea floor spreading, and mountain building. [Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ. Edward J. Tarbuck, Frederick K. Lutgens and Dennis Tasa, Earth: An Introduction to Physical Geology, 8th Edition, © 2005, p. 426, Fig. 14.9.]

Oceanic plates are thinner, but slightly denser than continental plates so that when the two collide, the ocean plate is *subducted* (i.e., drawn under the continental plate) and incorporated into the Earth's mantle, as shown schematically in Fig. 2.20. Rocks in the subducted oceanic crust are subjected to increasingly higher temperatures and pressures as they descend, giving rise to physical and chemical transformations.

Collisions between plate boundaries are often associated with volcanic activity and with the uplift of mountain ranges. The highest of the Earth's mountain ranges, the Himalayas, was created by folding of the Earth's crust following the collision of the Indian and Asian plates, and it is still going on today. The Rockies, Cascades, and Sierra ranges in western North America have been created in a similar manner by the collision of the Pacific and North American plates. These features have all appeared within the past 100 million years.

Oceanic plates are continually being recycled. The Pacific plate is being subducted along much of the extent of its boundaries, while new oceanic crust is being formed along the mid-Atlantic ridge as magma upwelling within the mantle rises to the surface, cools, and solidifies. As this newly formed crust diverges away from the mid-Atlantic ridge, the floor of the Atlantic Ocean is spreading, pushing other parts of the crust into the spaces formerly occupied by the subducted portions of the Pacific plate. As the Atlantic widens and the Pacific shrinks, the continents may be viewed as drifting away from the Atlantic sector on trajectories that will, in 100–200 million years, converge over what is now the mid-Pacific. A similar congregation of the continental plates is believed to have occurred about 200 million years ago, when they were clustered around the current position of Africa, forming a supercontinent called *Pangaea* (all Earth).

Some of the material incorporated into the mantle when plates are subducted contains *volatile substances* (i.e., substances that can exist in a gaseous form, such as water in hydrated minerals). As the temperature of these materials rises, pressure builds

[9] The theory of continental drift was first proposed by Alfred Wegener[10] in 1912 on the basis of the similarity between the shapes of coastlines, rock formations, and fossils on the two sides of the Atlantic. Wegener's radical reinterpretation of the processes that shaped the Earth was largely rejected by the geological community and did not become widely accepted until the 1960s, with the advent of geomagnetic evidence of sea-floor spreading.

[10] **Alfred Wegener** (1880–1930). German meteorologist, professor at University of Graz. Began his career at the small University of Marburg. First to propose that ice particles play an important role in the growth of cloud droplets. Set endurance record for time aloft in a hot air balloon (52 h) in 1906. Played a prominent role in the first expeditions to the interior of Greenland. Died on a relief mission on the Greenland icecap. The Alfred Wegener Institute in Bremerhaven is named in his honor. Son-in-law of Vladimir Köppen and co-authored a book with him.

up beneath the Earth's crust, leading to volcanic eruptions. As will be explained later in this chapter, gases expelled in volcanic eruptions are the source of the Earth's present atmosphere, and they are continually renewing it.

2.1.5 Roles of Various Components of the Earth System in Climate

Atmospheric processes play the lead role in determining such fundamental properties of climate as the disposition of incoming solar radiation, temperatures at the Earth's surface, the spatial distribution of water in the terrestrial biosphere, and the distribution of nutrients in the euphotic zone of the ocean. However, other components of the Earth system are also influential. Were it not for the large storage of heat in the ocean mixed layer and cryosphere during summer, and the extraction of that same heat during the following winter, seasonal variations in temperature over the middle and high latitude continents would be much larger than observed and, were it not for the existence of widespread vegetation, summertime daily maximum temperatures in excess of 40 °C would be commonplace over the continents. The oceanic thermohaline circulation warms the Arctic and coastal regions of Europe by several degrees, while wind-driven upwelling keeps the equatorial eastern Pacific cool enough to render the Galapagos Islands a suitable habitat for penguins!

Plate tectonics shaped the current configuration of continents and topography, which, in turn, shapes many of the distinctive regional features of today's climate. The associated recycling of minerals through the Earth's upper mantle is believed to have played a role in regulating the concentration of atmospheric carbon dioxide, which exerts a strong influence upon the Earth's surface temperature.

These are but a few examples of how climate depends not only on atmospheric processes, but on processes involving other components of the Earth system. As explained in Section 10.3, interactions between the atmosphere and other components of the Earth system give rise to feedbacks that can either amplify or dampen the climatic response to an imposed external forcing of the climate system, such as a change in the luminosity of the sun or human-induced changes in atmospheric composition.

The next three sections of this chapter describe the exchanges and cycling of water, carbon, and oxygen among the various components of the Earth system.

2.2 The Hydrologic Cycle

Life on Earth is critically dependent on the cycling of water back and forth among the various *reservoirs* in the Earth system listed in Table 2.2, which are collectively known as the *hydrosphere*. In discussing the exchanges between the smaller reservoirs, we make use of the concept of *residence time* of a substance within a specified reservoir, defined as the mass in the reservoir divided by the *efflux* (the rate at which the substance exits from the reservoir).[11] Residence time provides an indication of amount of time that a typical molecule spends in the reservoir between visits to other reservoirs. Long residence times are indicative of large reservoirs and/or slow rates of exchange with other reservoirs, and vice versa.

Based on current estimates, the largest reservoir of water in the Earth system is the mantle. The rate at which water is expelled from the mantle in volcanic emissions is estimated to be $\sim 2 \times 10^{-4}$ kg m^{-2} year^{-1} averaged over the Earth's surface, which is the basis for the 10^{11} year residence time in Table 2.2. At this

Table 2.2 Masses of the various reservoirs of water in the Earth system (in 10^3 kg m^{-2}) averaged over the surface of the Earth, and corresponding residence times

Reservoirs of water	Mass	Residence time
Atmosphere	0.01	Days
Fresh water (lakes and rivers)	0.6	Days to years
Fresh water (underground)	15	Up to hundreds of years
Alpine glaciers	0.2	Up to hundreds of years[a]
Greenland ice sheet	5	10,000 years[b]
Antarctic ice sheet	53	100,000 years
Oceans	2,700	
Crust and mantle	20,000	10^{11} years

[a] Estimated by dividing typical ice thicknesses of a large alpine glacier (~300 m) by the annual rate of ice accumulation (~1 m).
[b] Estimated by dividing typical ice thicknesses in the interior of the Greenland ice sheet (2000 m) by the annual rate of ice accumulation (~0.2 m).

[11] The concept of residence time is developed more fully in Chapter 5.1.

rate of exchange, only roughly 5% of the water estimated to reside in the mantle would be expelled over the $\sim 4.5 \times 10^9$-year lifetime of the Earth—not even enough to fill the oceans.

After the mantle and oceans, the next largest reservoir of water in the Earth system is the continental ice sheets, the volumes of which have varied widely on timescales of tens of thousands of years and longer, causing large variations in global sea level.

Exercise 2.2 Based on data provided in Table 2.1, estimate how much the sea level would rise if the entire Greenland ice sheet were to melt.

Solution: The mass of the Greenland ice sheet is equal to its mass per unit area averaged over the surface of the Earth (as listed in Table 2.1) times the area of the Earth or

$$(5 \times 10^3 \text{ kg m}^{-2}) \times (5.10 \times 10^{14} \text{ m}^2) = 2.55 \times 10^{18} \text{ kg}$$

If the ice cap were to melt, this mass would be distributed uniformly over the ocean-covered area of the Earth's surface. Hence, if x is the sea level rise, we can write

$$\text{(Area of oceans)} \times \text{(density of water)}\, x = \text{mass of ice sheet}$$

$$((5.10 - 1.45) \times 10^{14} \text{ m}^2) \times (10^3 \text{ kg m}^{-2})\, x = 2.55 \times 10^{18} \text{ kg}$$

Solving, we obtain $x = 7$ m.

Because the masses given in Table 2.2 are expressed in units numerically equivalent to the depth (in m) of a layer covering the entire surface of the Earth, we could have written simply

$$(5.10 - 1.45)\, x = 5.10 \times 5 \text{ m}$$
$$x = 7 \text{ m}$$

■

Of the reservoirs listed in Table 2.2, the atmosphere is by far the smallest and it is the one with the largest rates of exchange with the other components of the Earth system. The residence time of water in the atmosphere, estimated by dividing the mass of water residing in the atmosphere ($\sim$30 kg m^{-2}, equivalent to a layer of liquid water $\sim$3 cm deep) by the mean rainfall rate averaged over the Earth's surface (roughly 1 m per year or 0.3 cm day^{-1}), is $\sim$10 days. By virtue of the large exchange rate and the large latent heat of vaporization of water, the cycling of water vapor through the atmospheric branch of the hydrologic cycle is effective in transferring energy from the Earth's surface to the atmosphere.

Averaged over the globe, the rate of precipitation P equals the rate of evaporation E: any appreciable imbalance between these terms would result in a rapid accumulation or depletion of atmospheric water vapor, which is not observed. However, in analyzing the water balance for a limited region, the horizontal transport of water vapor by winds must also be considered. For example, within the region of the ITCZ, $P >> E$: the excess precipitation is derived from an influx of water vapor carried by the converging trade winds shown in Fig. 1.18. Conversely, in the region of the relatively dry, cloud-free subtropical anticyclones, $E >> P$: the excess water vapor is carried away, toward the ITCZ on the equatorward side and toward the midlatitude storm tracks on the poleward side, by the diverging low-level winds. For the continents as a whole, $P > E$: the excess precipitation returns to the sea in rivers. Local evapotranspiration E, as described in Section 2.1.3, accounts for an appreciable fraction of the moisture in summer rainfall P over the continents.

Under steady-state conditions, the mass balance for water vapor over in a column of area A, extending from the Earth's surface to the top of the atmosphere, can be written in the form

$$\overline{E} - \overline{P} = \overline{Tr} \tag{2.1}$$

where overbars denote averages over the area of the column and $\overline{Tr}$ denotes the horizontal transport (or flux) of water vapor out of the column by the winds, as discussed in the previous paragraph. Figure 2.21 shows the distributions of the export of water vapor (i.e., divergence of water vapor transport) by the winds over the low latitude oceans together with the observed distribution of $E - P$. Two aspects of Fig. 2.21 are worthy of note.

1. Apart from the sign reversal, the distribution of $E - P$ in the lower panel resembles the rainfall distribution in Fig. 1.25. That $-P$ and $E - P$ exhibit similar distributions indicates that the horizontal gradients of P must be much stronger than those in E. It follows that the strong observed gradients in climatological-mean rainfall are due to wind patterns rather than to gradients in local evaporation.

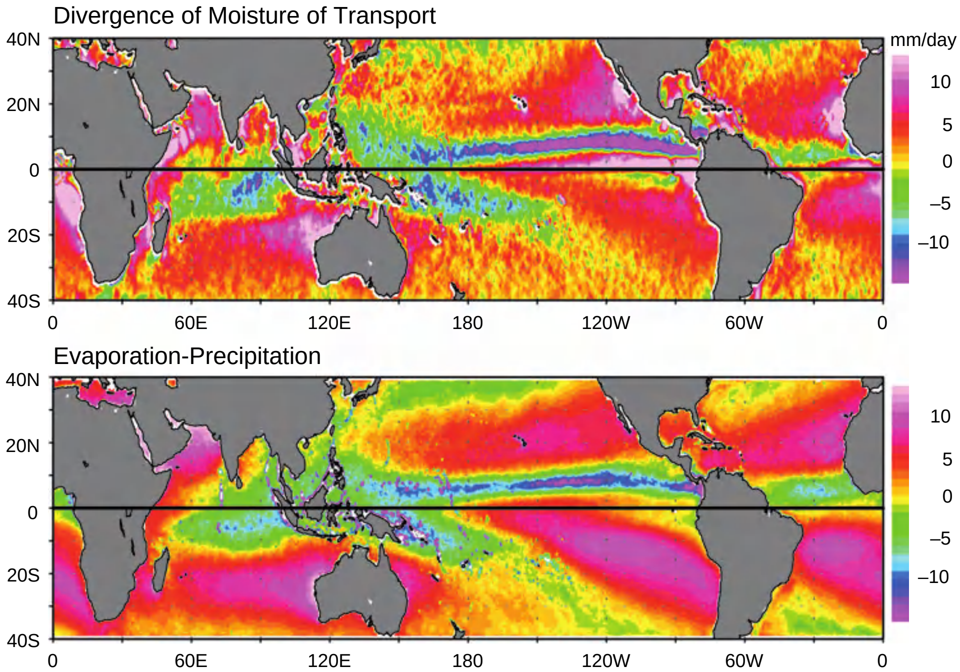


Fig. 2.21 Terms in the annual mean mass balance of atmospheric water vapor in units of mm day^{-1} of liquid water. (Top) The local rate of change of vertically integrated water vapor due to horizontal transport by the winds. (Bottom) Difference between local evaporation and local precipitation. If the estimates were perfect, the maps would be identical. [Based on data from NASA's QuikSCAT and Tropical Rain Measuring Mission (TRMM). Courtesy of W. Timothy Liu and Xiaosu Xie.]

2. In accordance with (2.1), the geographical distributions of $E - P$ and *Tr* in Fig. 2.21 are similar. The agreement is noteworthy because the measurements used in constructing these two maps are entirely different. The distribution of *Tr* is constructed from data on winds and atmospheric water vapor concentrations, without reference to evaporation and precipitation.

Also of interest is the time-dependent hydrologic mass balance over land for a layer extending from the land surface downward to the base of the deepest aquifers. In this case

$$\frac{d\overline{St}}{dt} = \overline{P} - \overline{E} - \overline{T} \tag{2.2}$$

where $\overline{St}$ is the area averaged storage of water within some prescribed region and the transport term involves the inflow or outflow of water in rivers and subsurface aquifers. For the special case of a land-locked basin from which there is no inflow or outflow of surface water, the transport term vanishes, and

$$\frac{d\overline{St}}{dt} = \overline{P} - \overline{E} \tag{2.3}$$

Hence the storage of water within the basin, which is reflected in the level of the lake into which the rivers within the basin drain, increases and decreases in response to time variations in $\overline{P} - \overline{E}$.

Figure 2.22 shows how the level of the Great Salt Lake in the reat Basin of the western United States has varied in response to variations in precipitation. From the time of its historic low[12] in 1963 to the time of its high in 1987, the level of the Great Salt Lake rose by 6.65 m, the area of the lake increased by a factor of 3.5, and the volume increased by a factor of 4. The average precipitation during this 14-year interval was heavier than the long-term average, but there were large, year-to-year ups and downs. It is notable that the lake level rose smoothly and

[12] The historical time series dates back to 1847.

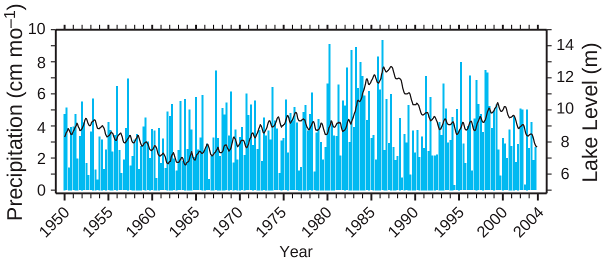


Fig. 2.22 The black curve shows variations in the depth of the Great Salt Lake based on a reference level of 4170 feet above sea level (in m). Depth scale (in m) at right. Blue bars indicate seasonal-mean precipitation at nearby Logan, Utah (in cm month^{-1}). [Lake level data from the U.S. Geological Survey. Courtesy of John D. Horel and Todd P. Mitchell.]

monotonically despite the large year-to-year variations in the precipitation time series. Exercises 2.11–2.13 at the end of this chapter are designed to provide some insight into this behavior.

2.3 The Carbon Cycle

Most of the exchanges between reservoirs in the hydrologic cycle considered in the previous section involve phase changes and transports of a single chemical species, H_2O. In contrast, the cycling of carbon involves chemical transformations. The carbon cycle is of interest from the point of view of climate because it regulates the concentrations of two of the atmosphere's two most important greenhouse gases: carbon dioxide (CO_2) and methane (CH_4).

The important carbon reservoirs in the Earth system are listed in Table 2.3 together with their masses and the residence times, in the same units as in Table 2.2. The atmospheric CO_2 reservoir is intermediate in size between the active biospheric reservoir (green plants, plankton, and the entire food web) and the gigantic reservoirs in the Earth's crust. The exchange rates into and out of the small reservoirs are many orders of magnitude faster than those that involve the large reservoirs. The carbon reservoirs in the Earth's crust have residence times many orders of magnitude longer than the atmospheric reservoirs, reflecting not only their larger sizes, but also the much slower rates at which they exchange carbon with the other components of the Earth system. Figure 2.23 provides an overview of the cycling of carbon between the various carbon reservoirs.

Table 2.3 Major carbon reservoirs in the Earth system and their present capacities in units of kg m^{-2} averaged over the Earth's surface and their residence times[a]

Reservoir	Capacity	Residence time
Atmospheric CO_2	1.6	10 years
Atmospheric CH_4	0.02	9 years
Green part of the biosphere	0.2	Days to seasons
Tree trunks and roots	1.2	Up to centuries
Soils and sediments	3	Decades to millennia
Fossil fuels	10	–
Organic C in sedimentary rocks	20,000	2×10^8 years
Ocean: dissolved CO_2	1.5	12 years
Ocean CO_3^{2-}	2.5	6,500 years
Ocean HCO^-	70	200,000 years
Inorganic C in sedimentary rocks	80,000	10^8 years

[a] Capacities based on data in Fig. 8.3 (p. 150) of Kump, Lee R.; Kasting, James F.; Crane, Robert G., The Earth System, 2nd Edition, © 2004. Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ.

Exercise 2.3 Carbon inventories are often expressed in terms of gigatons of carbon (Gt C), where the prefix *giga* indicates 10^9 and *t* indicates a metric ton or 10^3 kg. (Gt is equivalent to Pg in cgs units, where the prefix *peta* denotes 10^{15}.) What is the conversion factor between these units and the units used in Table 2.3?

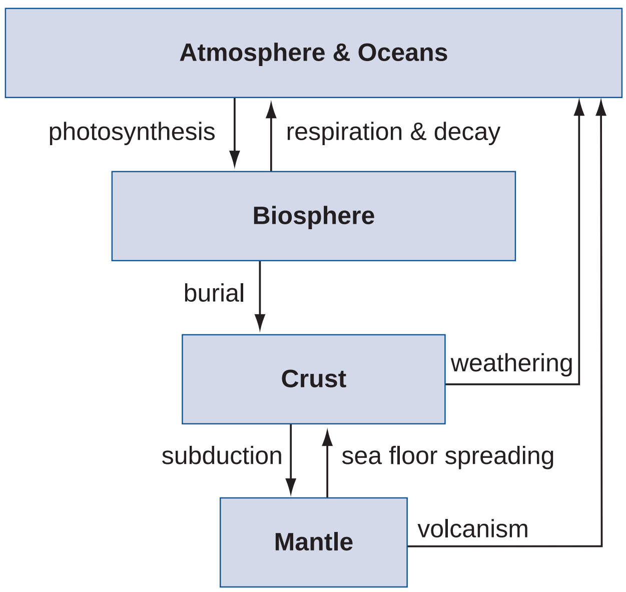


Fig. 2.23 Processes responsible for the cycling of carbon between the various reservoirs in the Earth system.

Climate Change And Its Implications (CCI)

Dr. Raji P

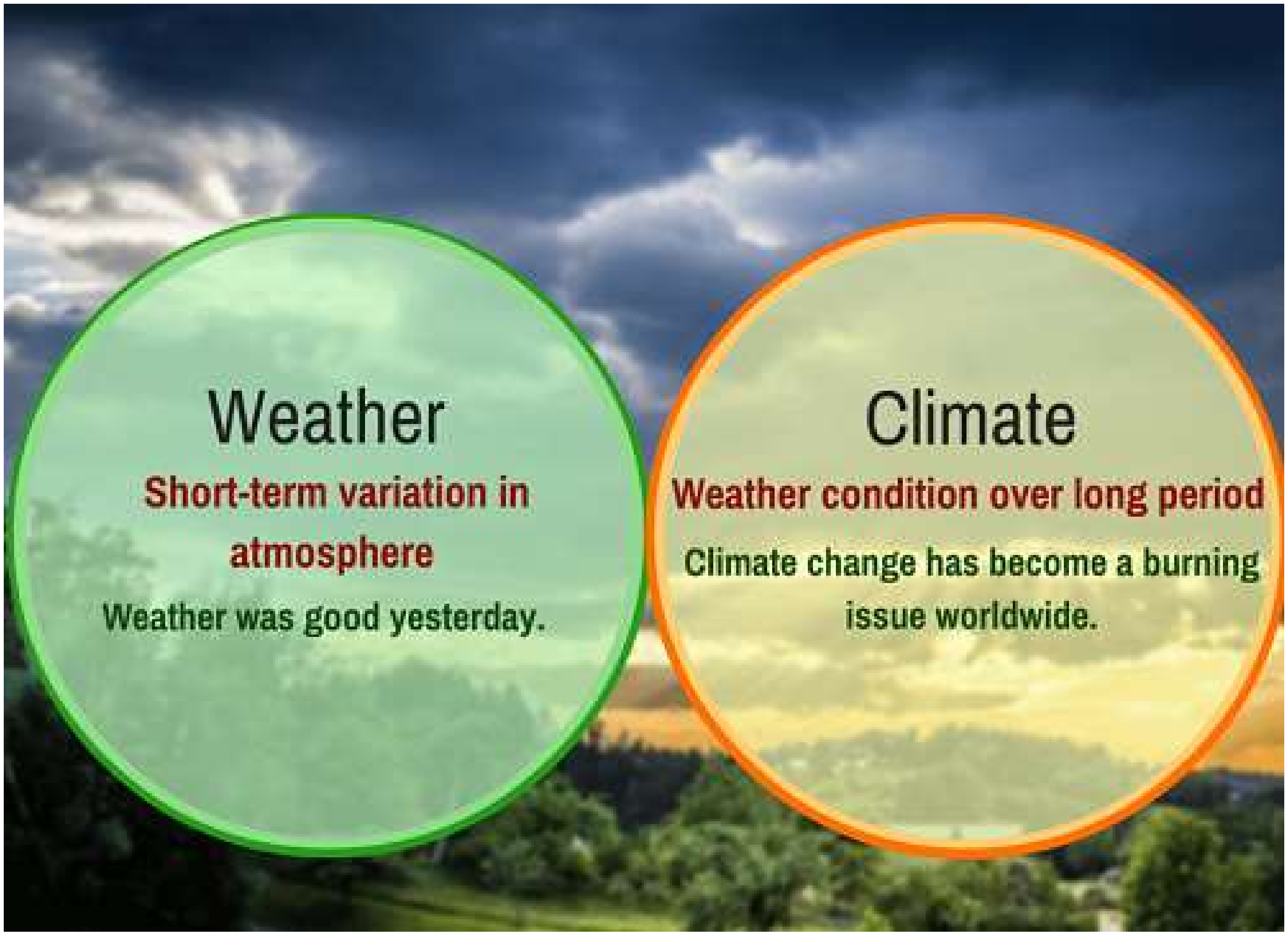
Lecture-3

Lecture-1&2

- Earth’s atmosphere
- Oceans
- Cryosphere
- Biosphere
- Earth’s crust and mantle

Class outline

- Introduction to weather & climate
- Weather parameters
- Measurements and analysis of weather parameters -interpretation


Weather
Short-term variation in atmosphere
Weather was good yesterday.
Climate
Weather condition over long period
Climate change has become a burning issue worldwide.

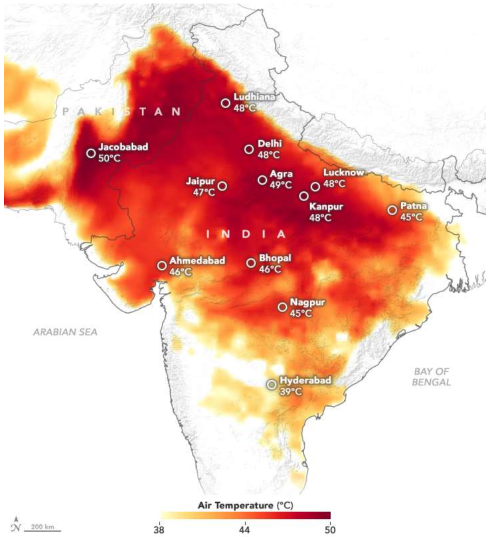
Weather & Climate

- Weather is the state of the atmosphere experienced at a given time
- It is defined by variables such as temperature, wind, rainfall, pressure, and other dynamical variables –Meteorology

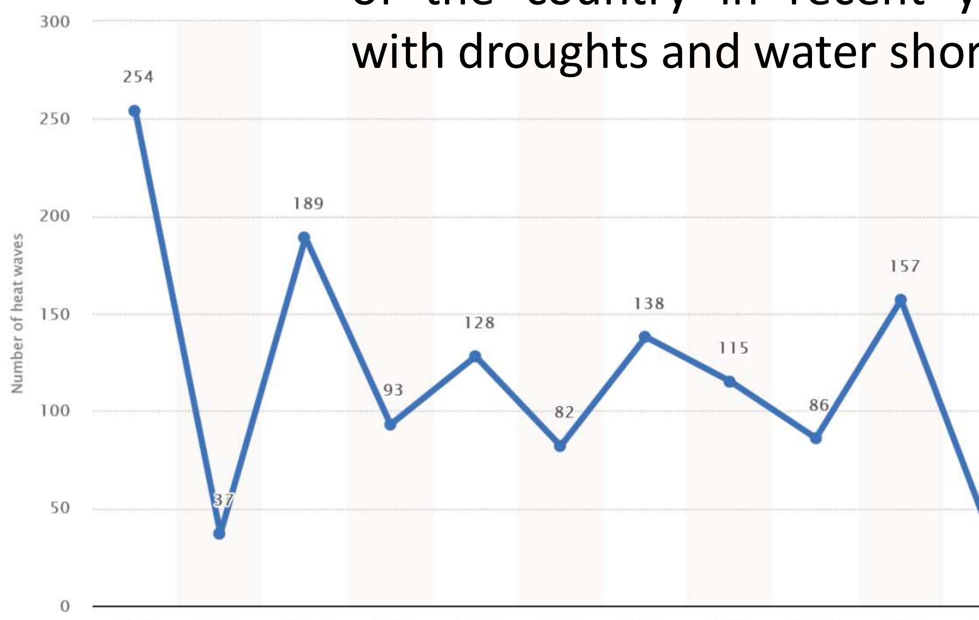
- Climate is the averages of weather elements obtained from their time series for a location or any region
- Climate refers to the monthly, seasonal or annual mean distributions of temperature, rainfall or any other weather parameter
- Any change in the incoming and outgoing radiations would affect its climate

Where do we use weather information?

Heat waves

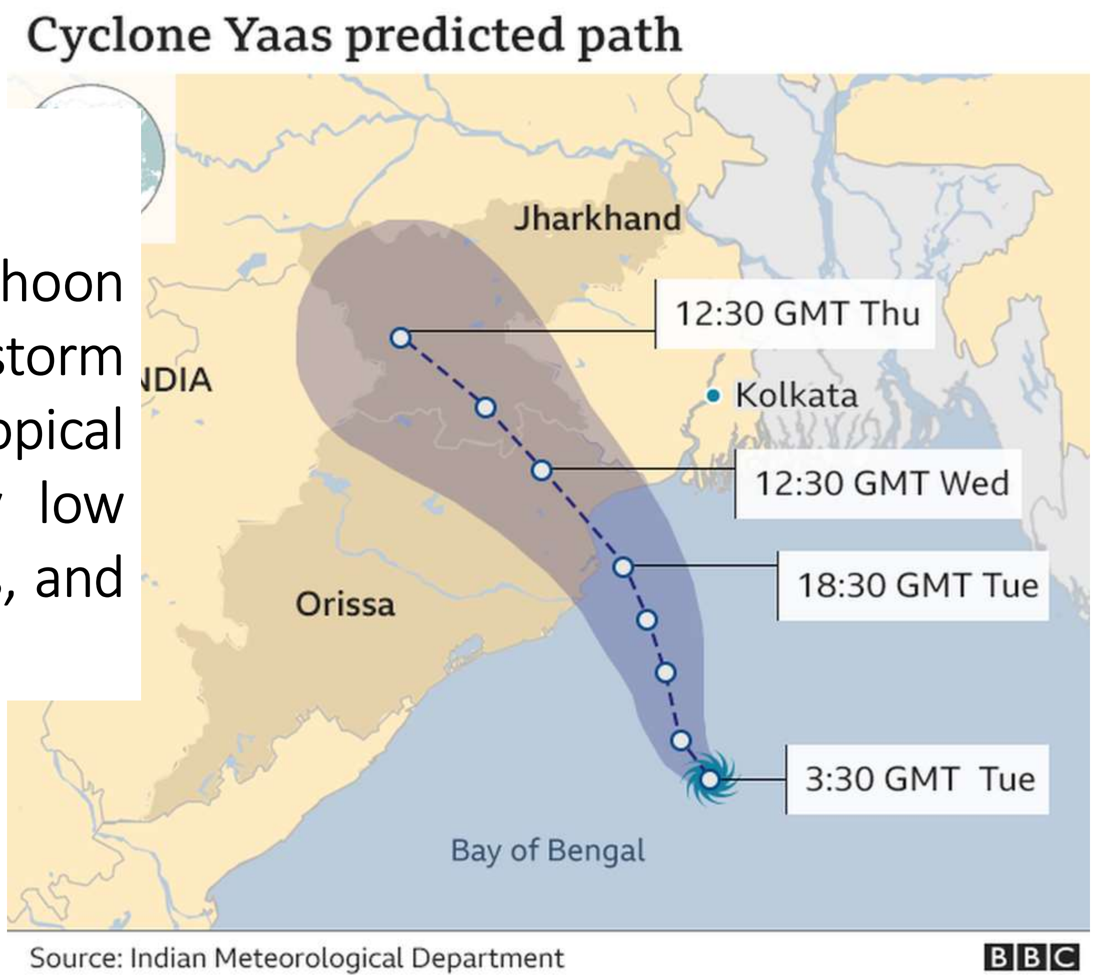


- Heat wave is considered if maximum temperature of a station reaches at least 40^0C or more for Plains and at least 30^0C or more for Hilly regions.
- About 42 heat wave days are reported in India in 2020
- These are more intense in northern regions of the country in recent years, coinciding with droughts and water shortage



Tropical cyclones

- Tropical cyclone, also called typhoon or hurricane, an intense circular storm that originates over warm tropical oceans and is characterized by low atmospheric pressure, high winds, and heavy rain



Extreme rainfall

Tornadoes

- A tornado is a narrow, violently rotating column of air that extends from a thunderstorm to the ground


USAF
DELTA

Weather Parameters

Temperature

- Maximum and minimum temperatures (most important for agriculture)
- Average temperature
- Diurnal temperature (Tmax-Tmin)

Measurements: Thermometers

Stevenson's screen

Satellite measurements

- Satellite measures the atmosphere in radiance (W/m^2), and then using mathematical and statistical equations, temperature is derived from this

INSAT-3D is a meteorological, data relay and satellite aided search and rescue satellite developed by the Indian Space Research Organisation (in 2013)

Precipitation


Hail


Sleet


Rain


Glaze

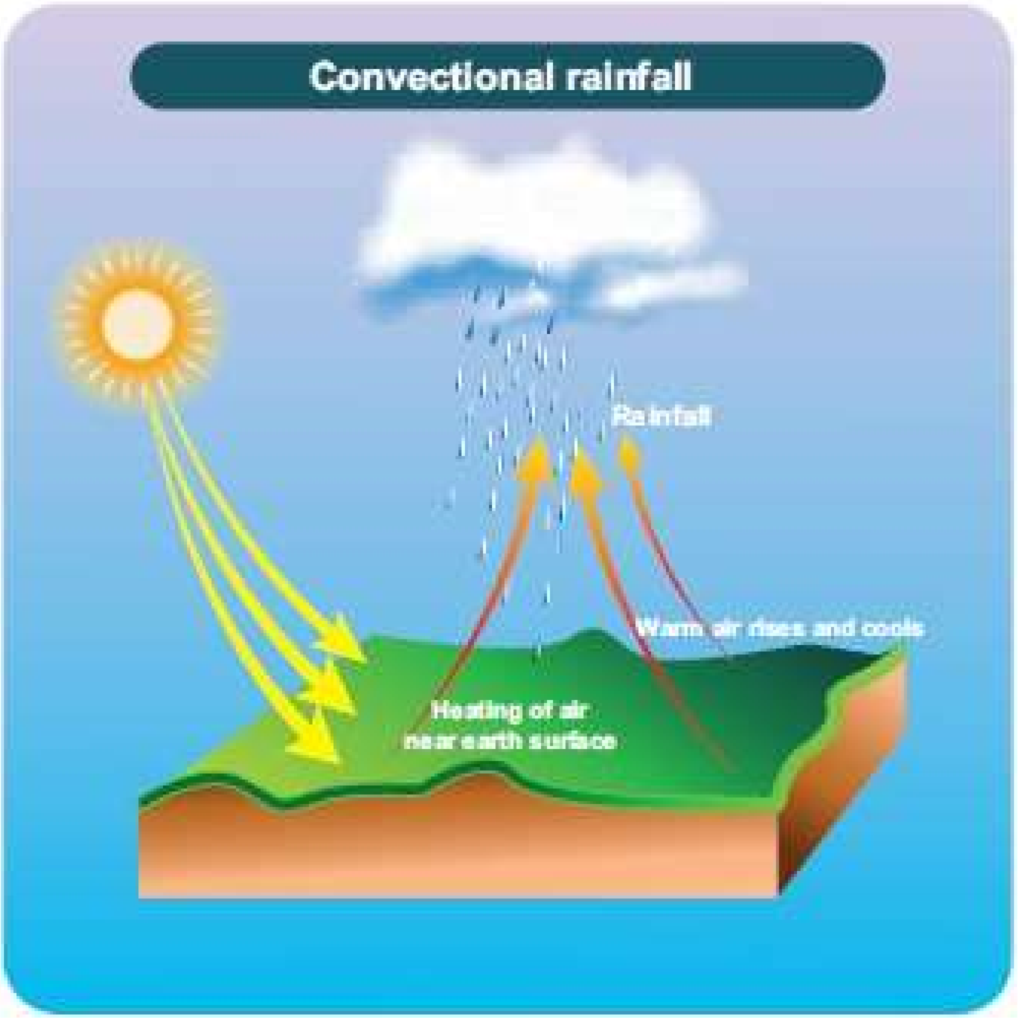

Snow

Forms of precipitation

- Rain – It is the main form of precipitation with water drop size >0.5 mm
- Snow – Ice crystals, which combines to form flakes
- Drizzle-Fine sprinkle of numerous water droplets of size <0.5 mm
- Glaze- when rain or drizzle comes in contact with cold ground at 0°C, the water drops freeze to form an ice coating called freezing rain
- Sleet – frozen raindrops forms when rain falls through air at subfreezing temperature
- Hail – precipitation in the form of ice crystals of size >8 mm

Weather systems for precipitation

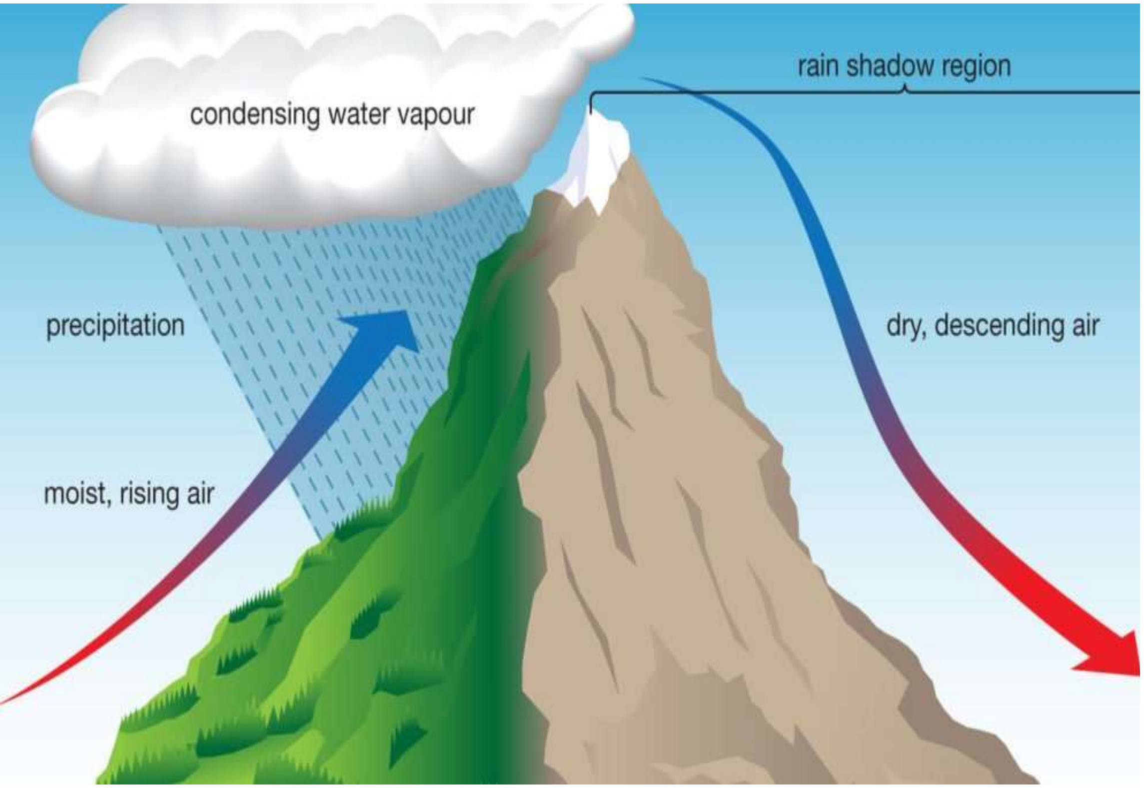
Convective precipitation



- On a hot day the ground surface becomes heated, as does the air in contact with it

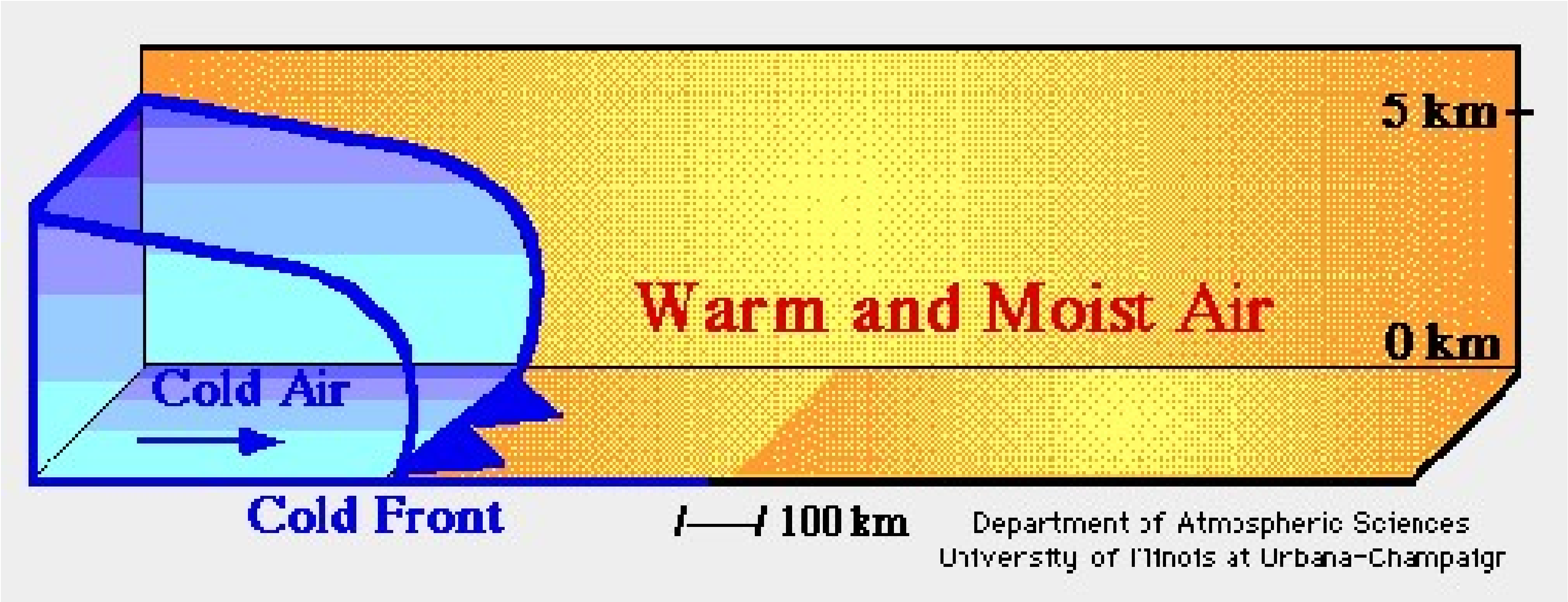
- This causes the air to rise, expand, and cool dynamically, causing condensation and precipitation

Orographic precipitation



- Lifting an air mass occurs when air flows up and over a topographic feature such as a mountain barrier
- Orographic barriers often supply the lift to set off precipitation
- For this reason, precipitation is heavier on windward slopes, with rain shadows (areas of lighter precipitation) on leeward slopes
- Orographic precipitation is associated with low intensity with relatively long durations

<u>Precipitation along cold front</u>: A cold front is defined as the transition zone where a cold air mass is replacing a warmer air mass



Characteristics of precipitation in India

1) **South-West Monsoon (June-September)**

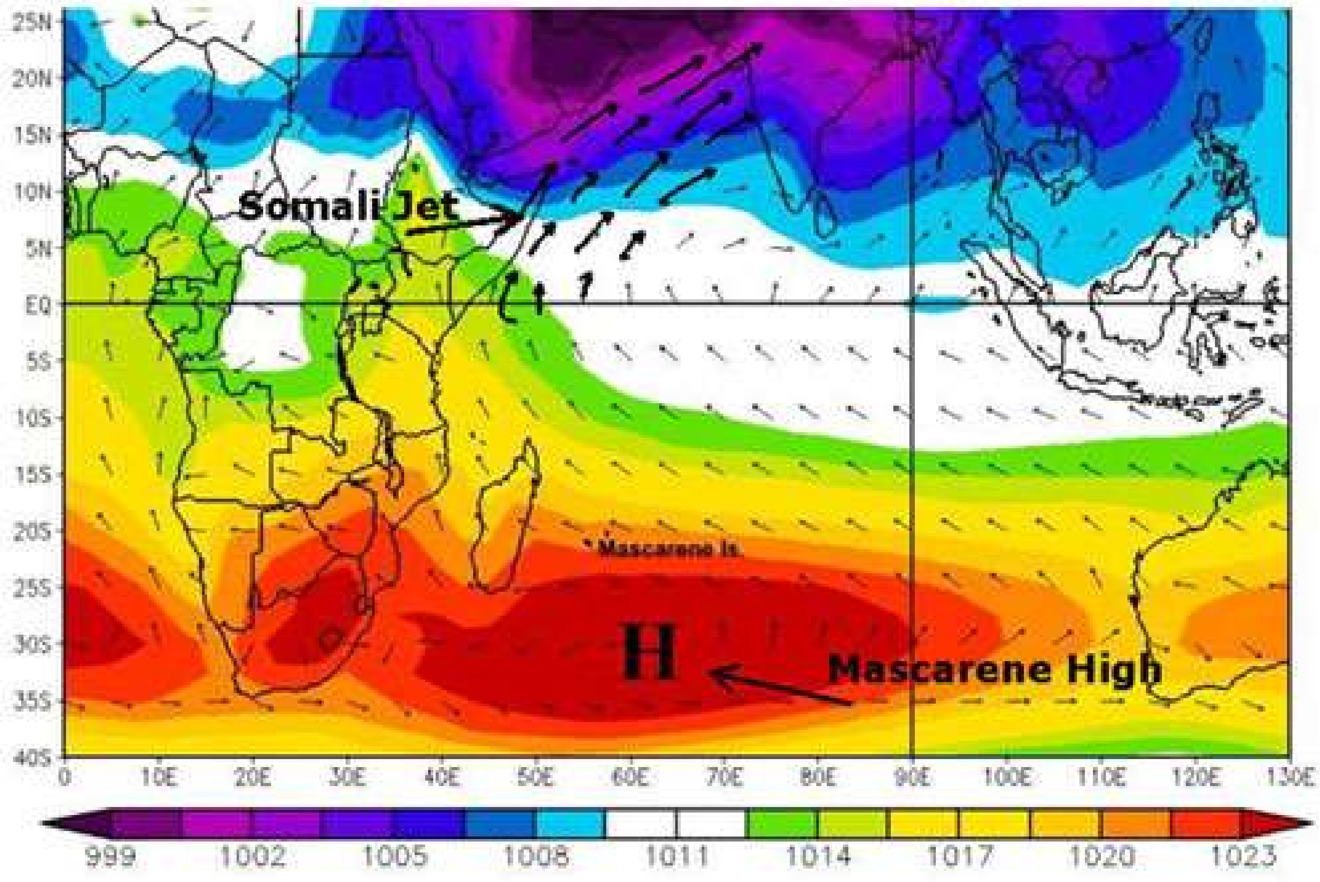
- It originates in Indian ocean
- Receives 75% of annual rainfall
- Starts from Kerala, and extends towards all states except Tamil Nadu and Jammu & Kashmir

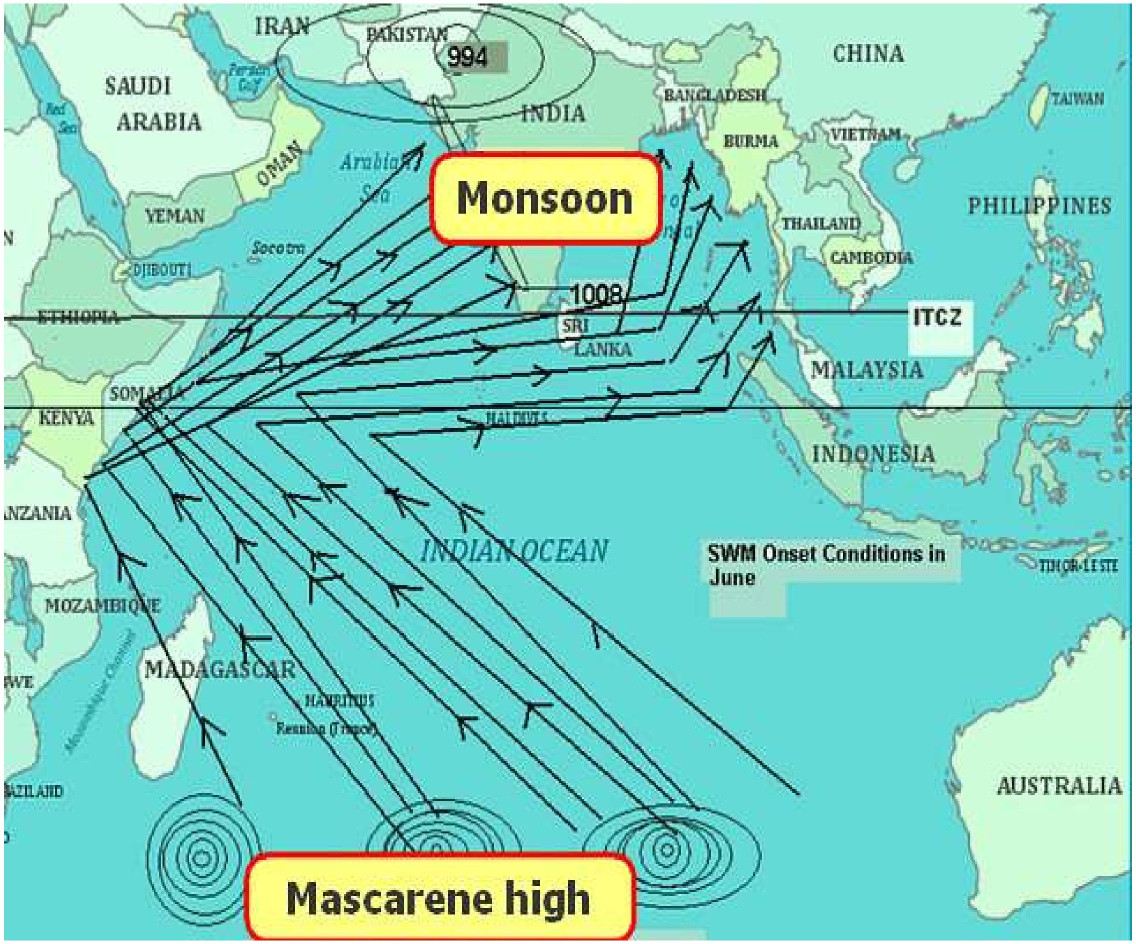
2) Post-Monsoon or NE Monsoon (October-November)

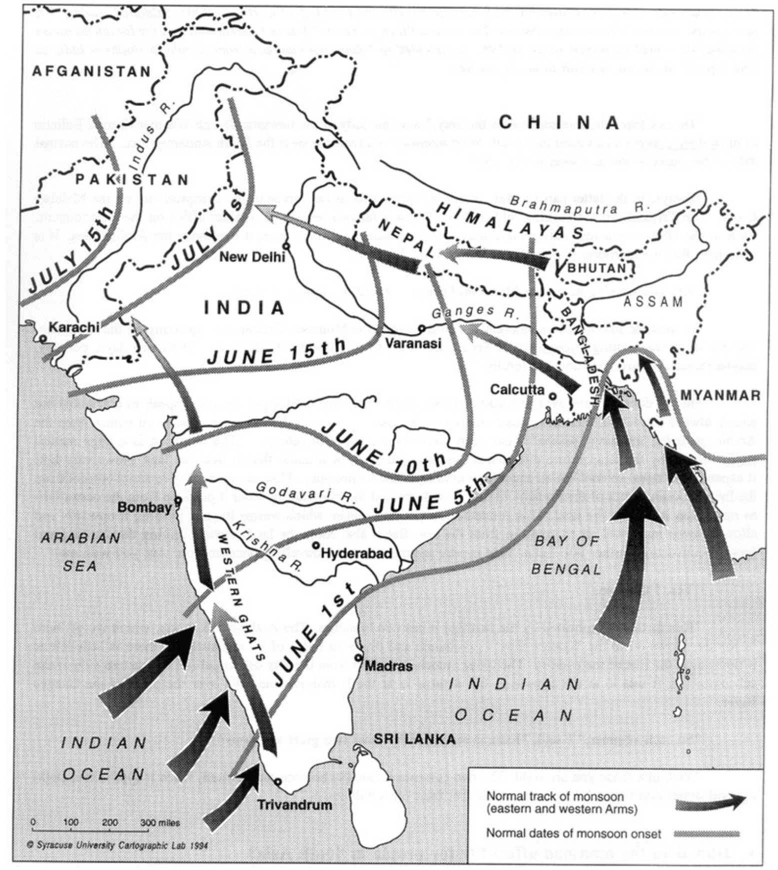
- It mainly strikes the east coast of Southern Peninsula (Tamil Nadu)

3) Pre-Monsoon (March-May)

Indian monsoon and its friends







ITCZ-Inter Tropical Convergence Zone: The region that circles the Earth, near the equator, where the trade winds of the Northern and Southern Hemispheres come together

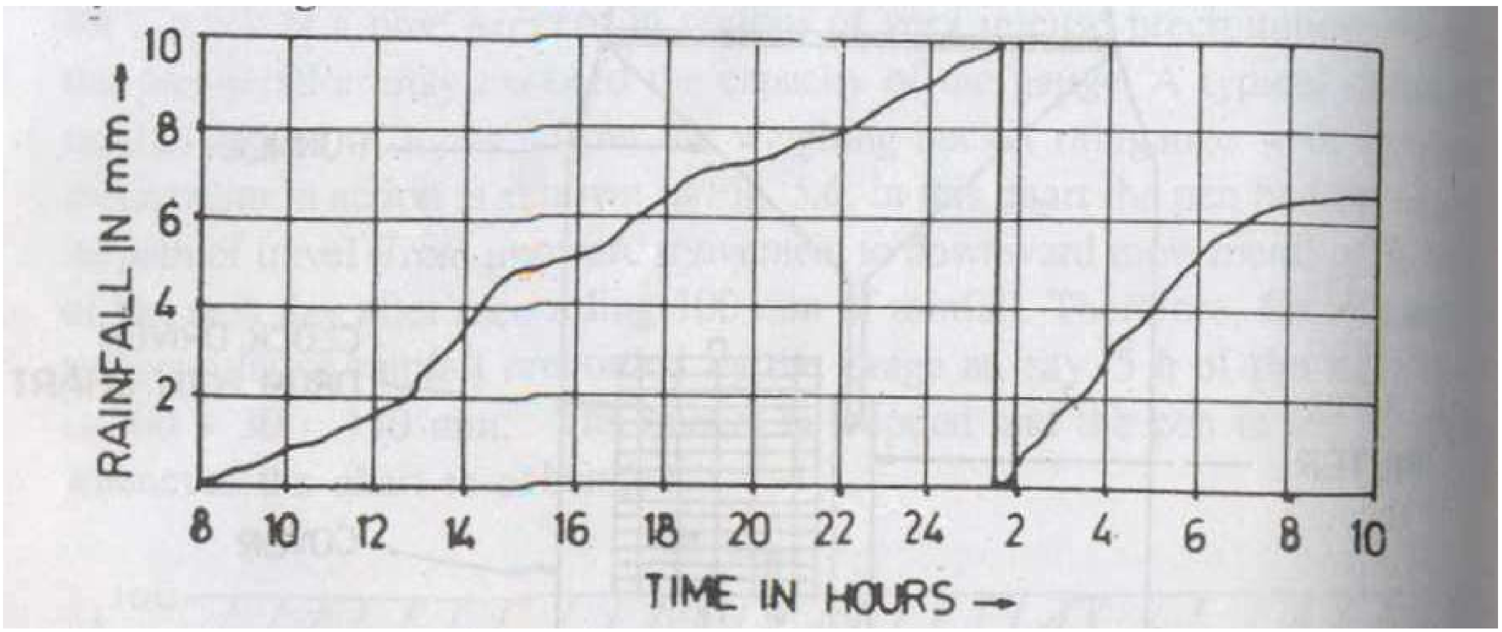
- The **Mascarene High** (MH) is a semi-permanent subtropical high-pressure zone in the South Indian Ocean
- A trough extends from this low over Pakistan (994 hPa) to Head Bay with strong pressure gradient to the south. This trough is often referred to as the 'monsoon trough'.

INDIA
BLANK MAP
North East
Trade winds
Arabian Sea
Bay of Bengal
INDIAN OCEAN

Measurement of rainfall

- Rainfall is expressed in terms of depth (mm)
- Rainfall is measured using rain gauges: Recording & non-recording gauges
- Radar/satellite measurement

Rain gauges



Rain gauge chart

Tropical Rainfall Measuring Mission (TRMM)-1997

- A joint space mission between NASA and the Japan Aerospace Exploration Agency JAXA designed to monitor and study tropical rainfall

Solar radiation

Meaurement (W/m^2)

- **<u>Pyranometer</u>**: It is designed to measure the solar radiation flux density (W/m^2) from the hemisphere above within a wavelength range 0.3 µm to 3 µm

- **Satellite**

Meteorological observatories

IMD (Indian Meteorological Department)

http://weather.uwyo.edu/upperair/sounding.html- University of Wyoming

https://power.larc.nasa.gov/data-access-viewer/- NASA Power data

Climate Change And Its Implications (CCI)

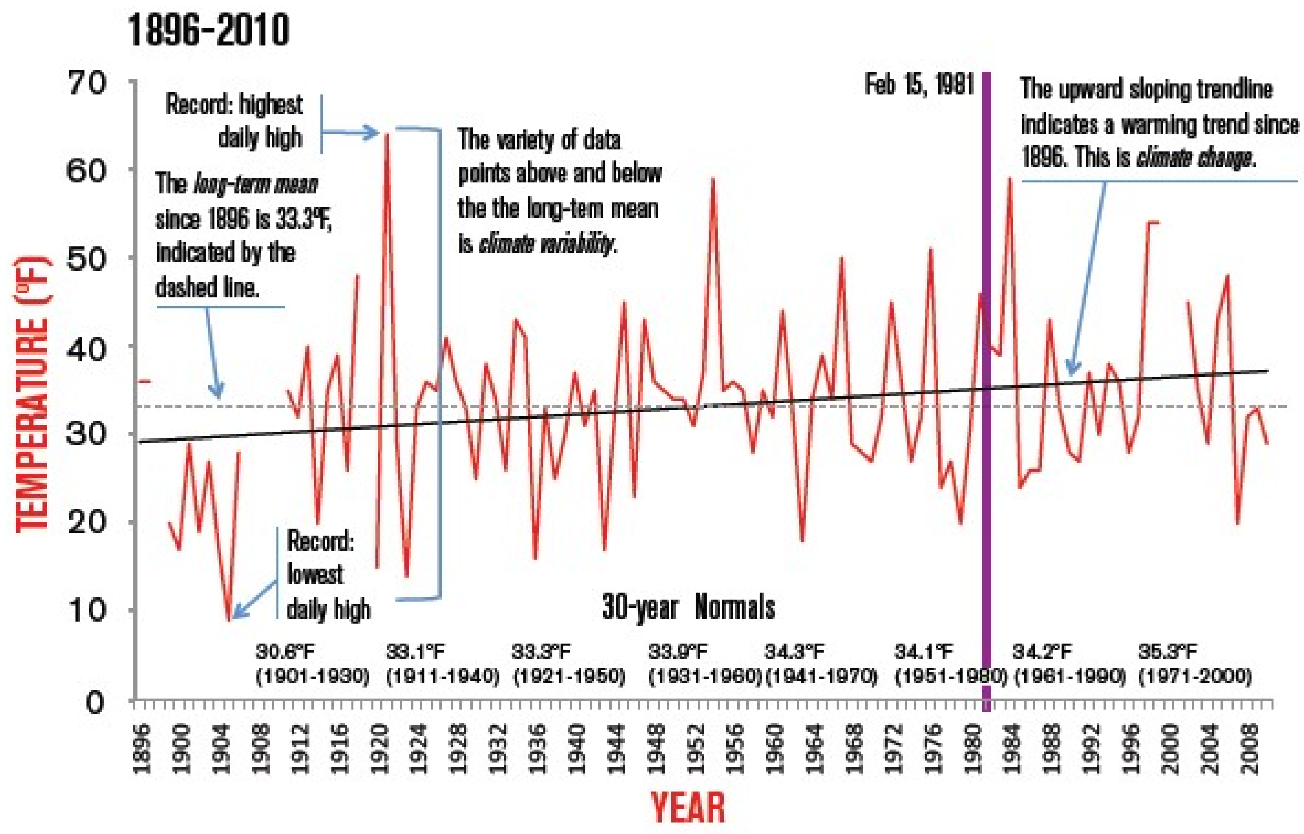
Dr. Raji P

Lecture-4

Class outline

- Climate variability and climate change
- Reasons for climate variability & change

Climate Variability & Climate Change



- Climate varies over seasons and years instead of day to day like weather
- Some summers are colder than others and some years precipitations are higher than others

- Climate variability: The way the climate fluctuates yearly above or below a long term average

- Climate change: Long term continuous change to average weather conditions

- Climate change is slow and continuous unlike variability

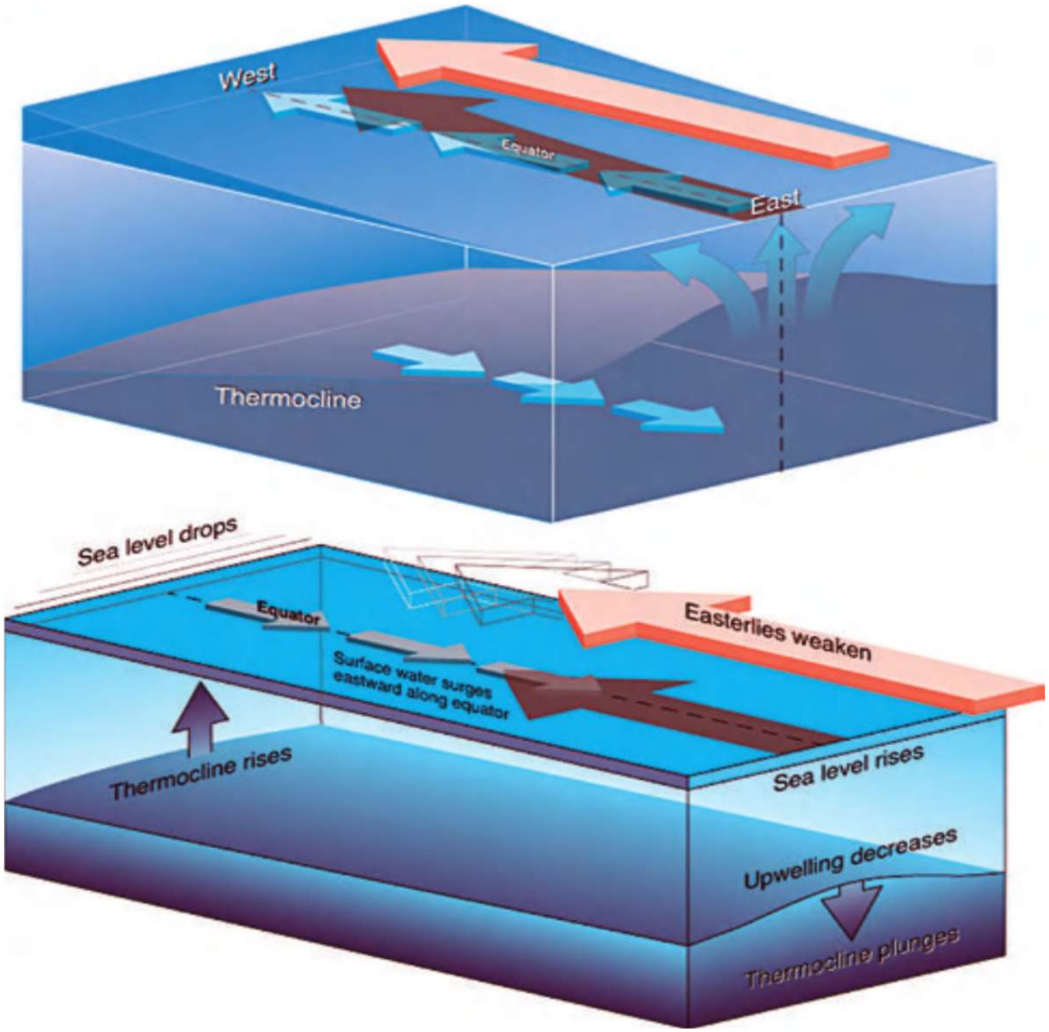
Assignment-1

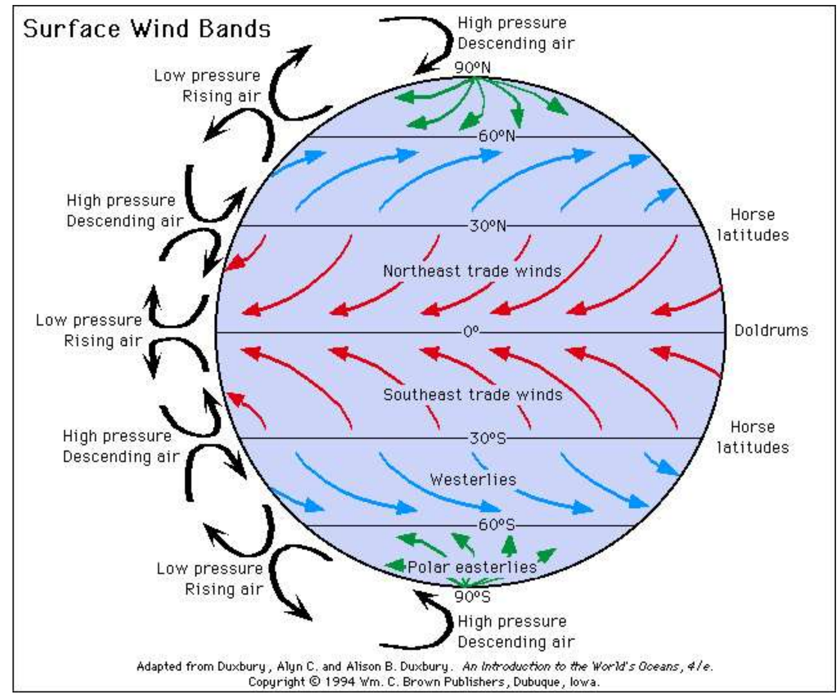
- Identify the trend in temperature and rainfall in your location from the historical data.

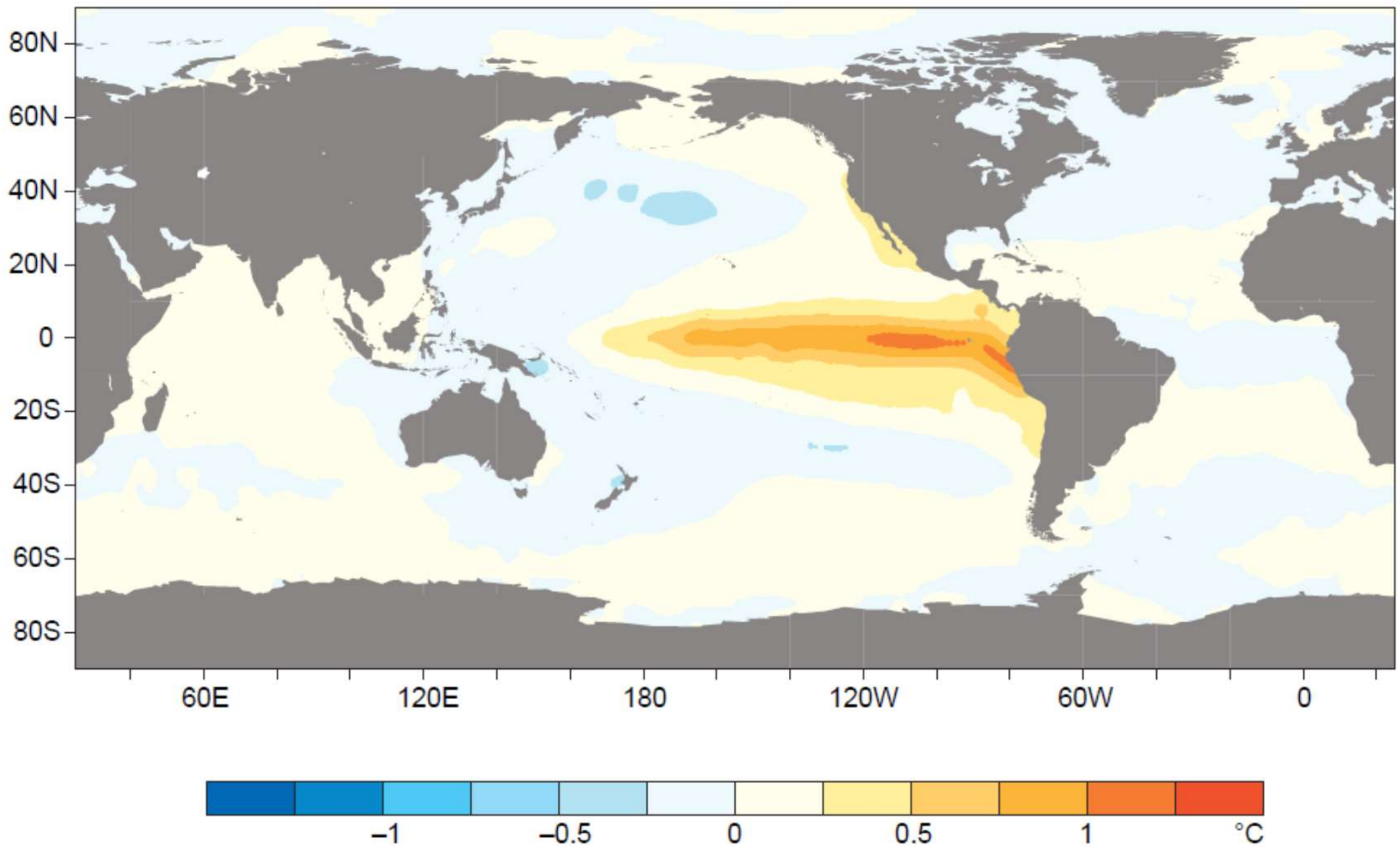
Download the data from 1980 to 2020 (30 years daily data) from NASA Power data (https://power.larc.nasa.gov/data-access-viewer/)

Submission due date: 10.02.2022

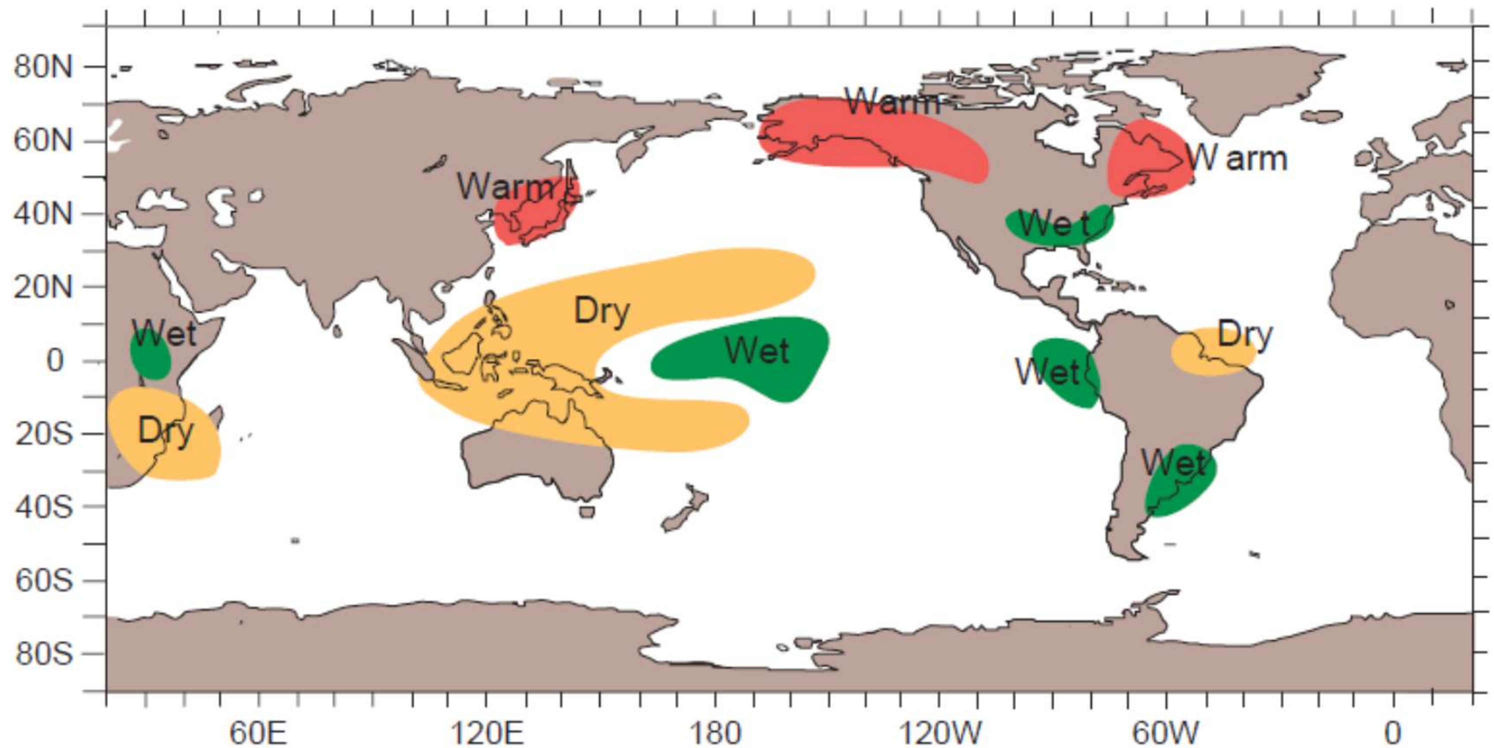
General circulation of the atmosphere



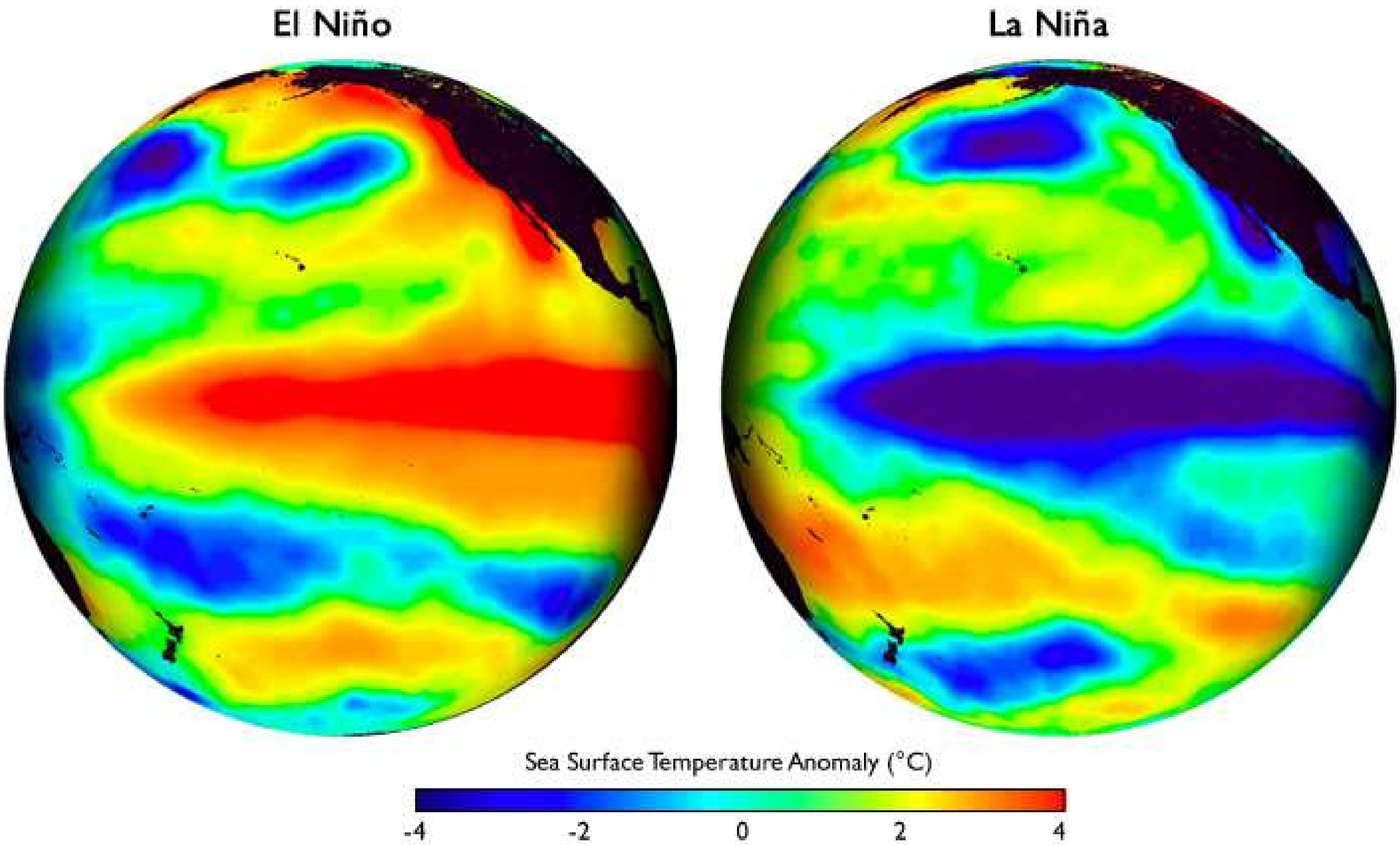




Global pattern of sea surface temperature (°C) anomalies observed during El Niño years

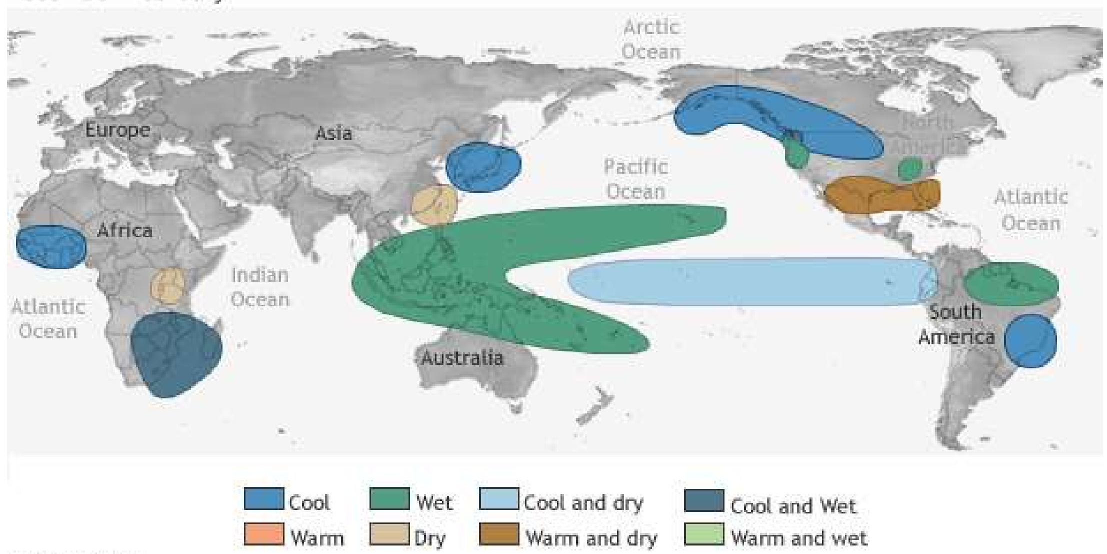


Impacts of El Niño on weather and climate

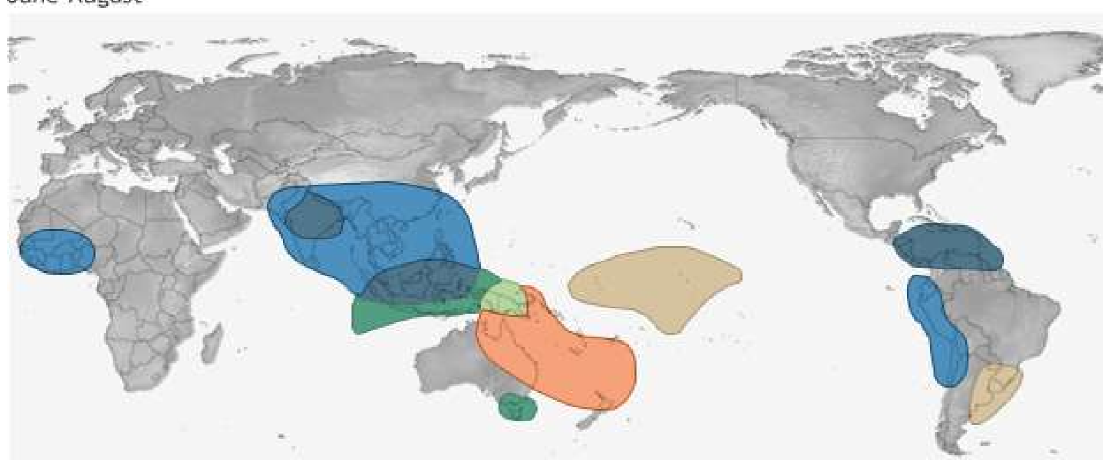

El Niño
La Niña
Sea Surface Temperature Anomaly (°C)
-4
-2
0
2
4

LA NIÑA CLIMATE IMPACTS

December-February



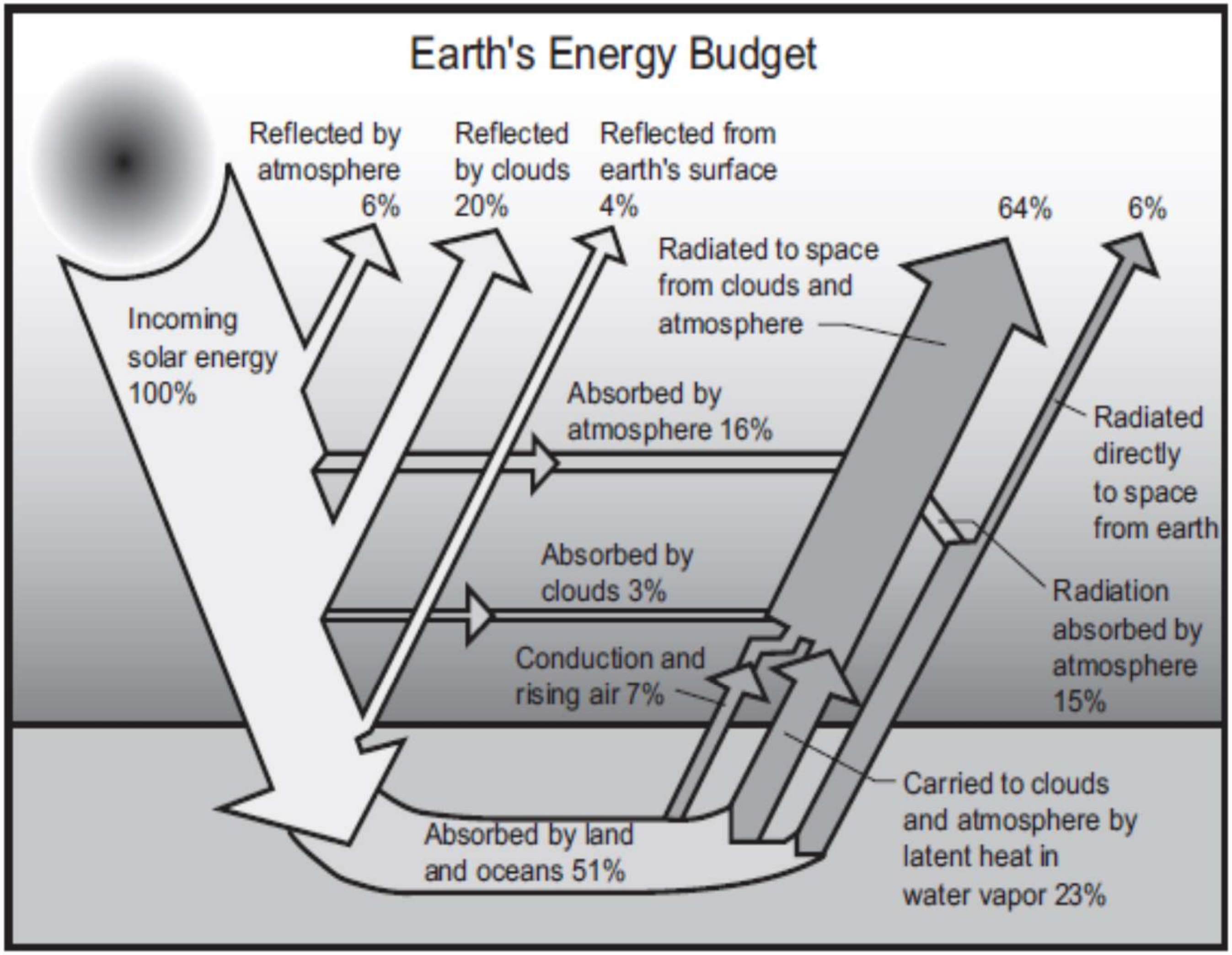
June-August



NOAA Climate.gov

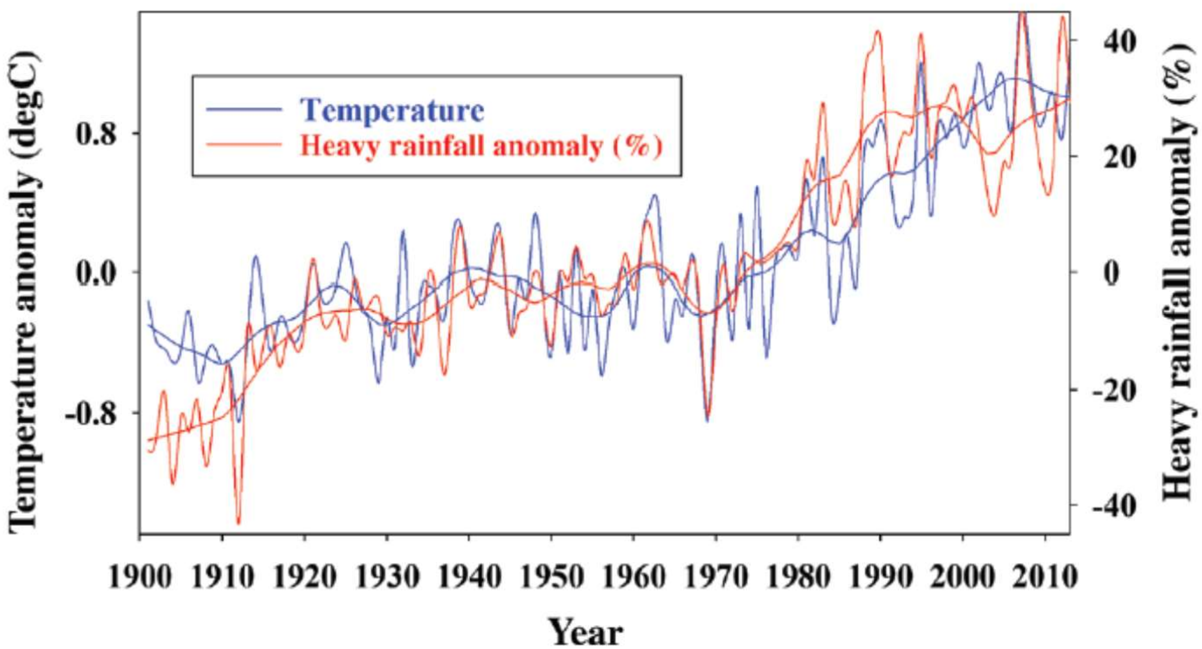
- El Niño and La Niña are opposite phases of a natural climate pattern across the tropical Pacific Ocean that swings back and forth every 3-7 years on average
- Together, they are called ENSO (pronounced “en-so”), which is short for **E**l **N**iño-**S**outhern **O**scillation
- El Niño (the warm phase) and La Niña (the cool phase) lead to significant differences from the average ocean temperatures, winds, surface pressure, and rainfall across parts of the tropical Pacific
- Climate Change is making El Niños more intense, leading to intensifying droughts, worsening floods, and shifting hurricane patterns
- Strong El Niños can cause severe drought in dry climates such as Australia and India, intense flooding in wetter climates such as the Pacific Northwest and Peru, and causes more hurricanes to form in the Pacific and fewer in the Atlantic

How the climate is changing?



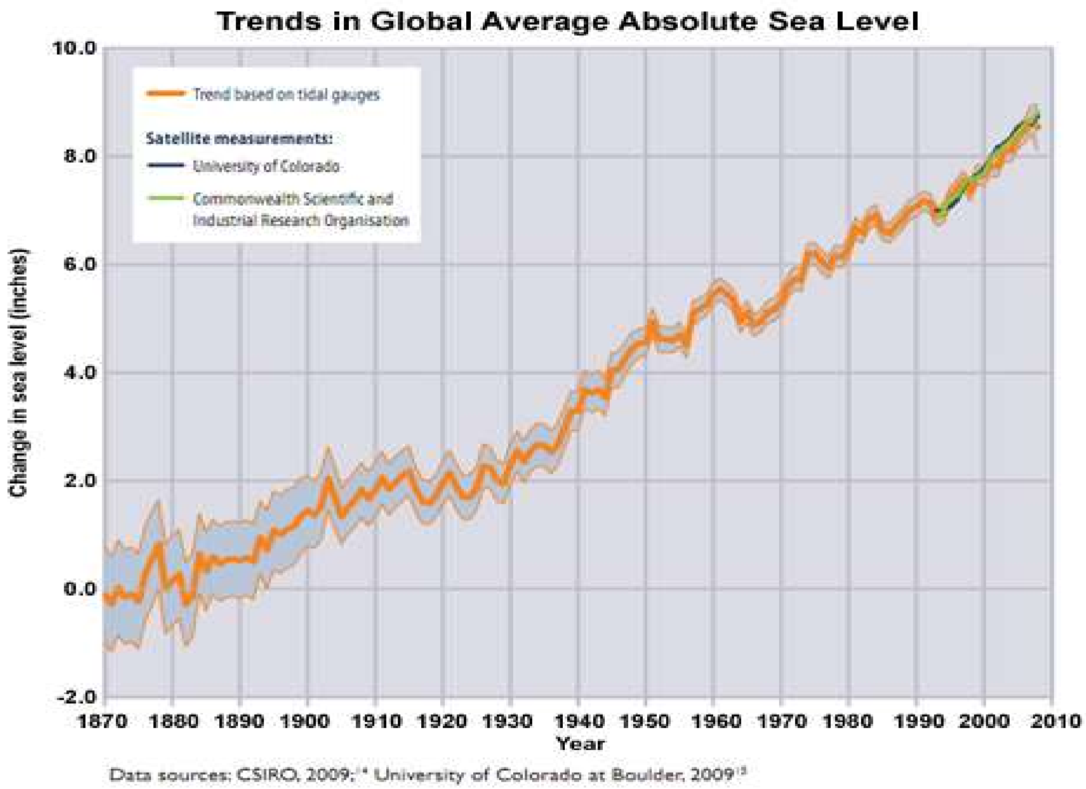
Atmosphere is responsible for radiating ~90% of total absorbed solar energy back to space!!

Climate Change: Is it real ?



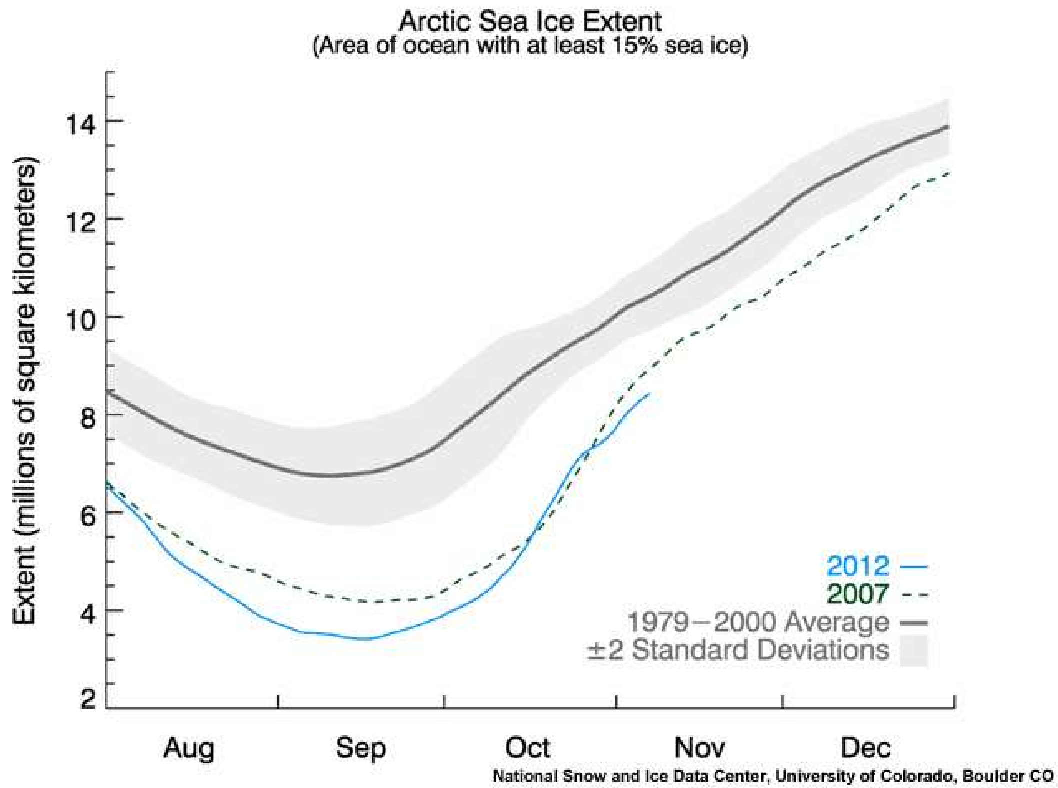
Earth is getting warmer and the temperature has been well above normal for more than 25 years.

Evidence of change: Oceans



The IPCC estimates that the oceans rose 4 to 10 inches (10-25 cm) in the 20th century from melting ice and snow and the physical expansion of warmer water.

Evidence of change: Sea Ice

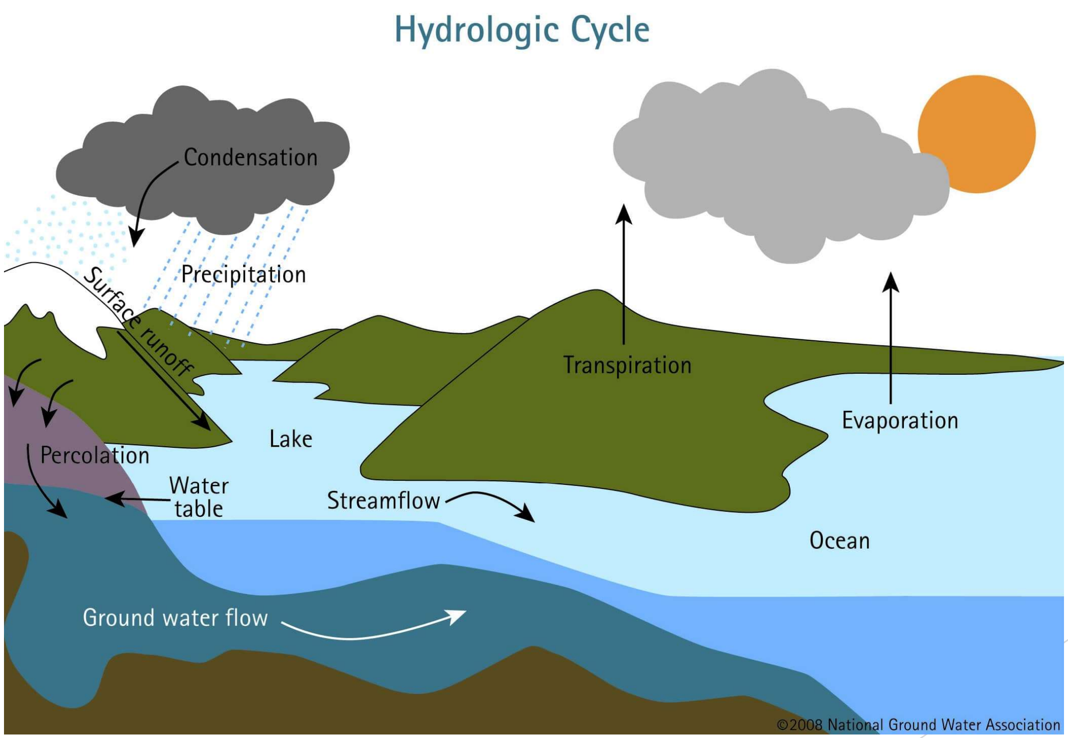


Sea ice is diminishing in the Arctic. Satellites have observed winter Arctic sea ice shrink by about 3-4% per decade from 1979, and an even higher rate in summer.

Natural reasons for climate change

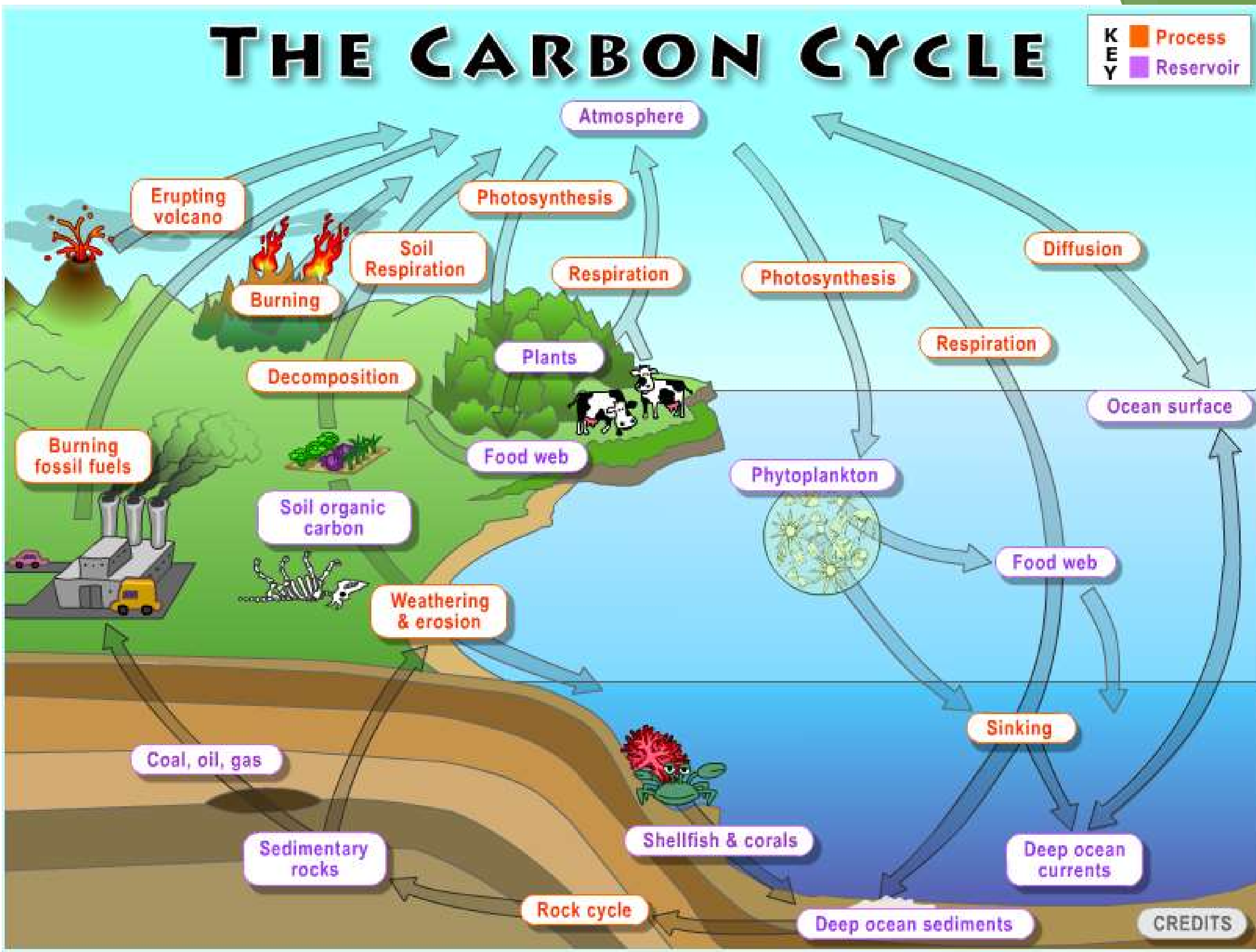
Hydrologic cycle –water cycle

- Life on Earth is dependent on the cycling of water back and forth among various reservoirs in the Earth system, and are hydrosphere

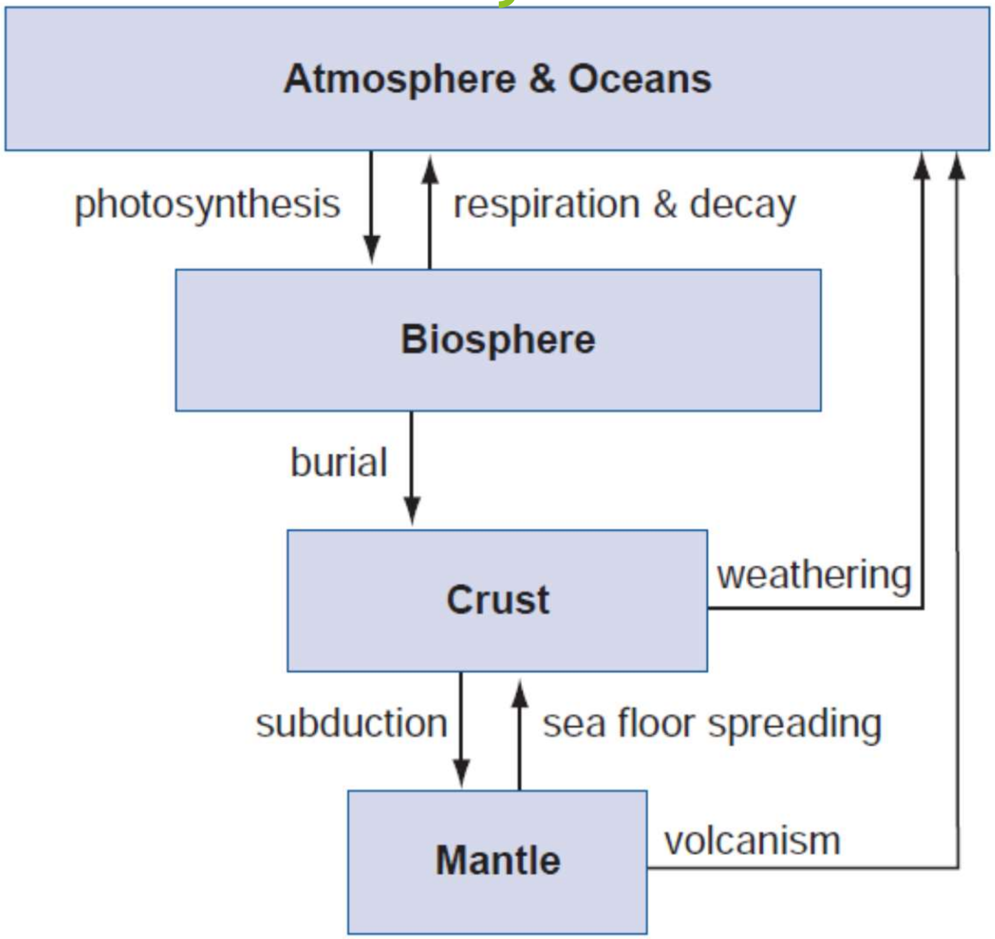


Reservoirs of water	Mass	Residence time
Atmosphere	0.01	Days
Fresh water (lakes and rivers)	0.6	Days to years
Fresh water (underground)	15	Up to hundreds of years
Alpine glaciers	0.2	Up to hundreds of years[a]
Greenland ice sheet	5	10,000 years[b]
Antarctic ice sheet	53	100,000 years
Oceans	2,700	
Crust and mantle	20,000	10^{11} years

Mass in 10^3 kg/m^2


THE CARBON CYCLE
KEY
Process
Reservoir
Atmosphere
Erupting volcano
Photosynthesis
Soil Respiration
Burning
Respiration
Photosynthesis
Diffusion
Plants
Respiration
Decomposition
Ocean surface
Burning fossil fuels
Food web
Phytoplankton
Soil organic carbon
Food web
Weathering & erosion
Sinking
Coal, oil, gas
Shellfish & corals
Sedimentary rocks
Deep ocean currents
Rock cycle
Deep ocean sediments
CREDITS

Carbon cycle



- Carbon cycle is important because it regulates two important green house gases CO_2 and CH_4

- Subduction: geological process in which the oceanic lithosphere is recycled into the Earth's mantle at convergent boundaries

➢ Lithosphere: Rocky outer part of the Earth

- Seafloor spreading is a geologic process in which tectonic plates—large slabs of Earth's lithosphere—split apart from each other

CHAPTER 2

CLIMATIC ELEMENTS

2.1 Weather elements

The weather elements that are used to describe climate are also the elements that determine the type of climate for a region. The most important elements of weather which in different combinations make up the climate of a particular place or area are: solar radiation, air temperature, air pressure, wind velocity and wind direction, humidity and precipitation, and amount of cloudiness. The climatic elements of temperature, precipitation, and wind are the most significant elements used to express the climate of a region.

Temperature:

The temperature of an area is dependent upon

a) latitude or the distribution of the incoming and outgoing radiation,
b) the nature of the surface (land or water),
c) the altitude and the
d) prevailing winds.

The air temperature normally used in climatology is that recorded at the surface. Moisture or lack of moisture modifies the temperature. The more moisture in a region, the smaller the temperature range, and the drier the region, the greater the temperature range. Moisture is also influenced by temperature. Warmer air can hold more moisture than a cooler air, resulting in increased evaporation and a higher probability of clouds and precipitation. Moisture, when coupled with condensation and evaporation, is an extremely important climatic element. It ultimately determines the type of climate for a specific region.

Precipitation:

Precipitation is the second most important climatic element. In most studies, precipitation is defined as water reaching Earth's surface by falling either in a liquid or a solid state. The most significant forms are rain and snow. Precipitation has a wide range of variability over the earth's surface. Because of its variability, a longer series of observations is generally required to establish mean or an average. It often becomes necessary to include such factors as average number of days with precipitation and average amount per day. Precipitation is expressed in millimeters. Since precipitation amounts are directly associated with amounts and type of clouds, cloud cover must also be considered with precipitation. Cloud climatology also includes phenomena such as fog and thunderstorms.

Wind:

Wind is the climatic element that transports heat and moisture into a region. Climatologists are mostly interested in wind with regards to its direction, speed, and gustiness. Wind is therefore usually discussed in terms of prevailing direction, average speeds, and maximum gusts. Some climatological studies use resultant wind, which is the vectorial average of all wind directions and speeds for a given level, at a specific place, and for a given period.

2.2 Expression of climatic elements

Climatic elements are observed over long periods of time; therefore, specific terms must be used to express these elements so they have a definite meaning.

Mean (Average):

The mean is the most commonly used climatological parameter. The term mean normally refers to a mathematical averaging obtained by adding the values of all factors or cases and then dividing by the number of items. For example, the average daily temperature would be the sum of the hourly temperatures divided by 24.

The mean temperature of 1 day has been devised by simply adding the maximum and minimum temperature values for that day and dividing by 2. In analyzing weather data, the terms average and mean are often used interchangeably.

Normal:

In climatology, the term normal is applied to the average value over aperiod of time, which serves as a standard with which values (occurring on a date or during a specified time) may be compared. These periods of time may be a particular month or other portion of the year. They may refer to a season or to a year as a whole. The normal is usually determined over a 20- or 30-year period.

Absolute:

In climatology, the term absolute is usually applied to the extreme highest and lowest values for any given meteorological element recorded at the place of observation and are most frequently applied to temperature.

Extreme:

The term extreme is applied to the highest and lowest value for a particular meteorological element occurring over a period of time. This period of time is usually a matter of months, seasons or years. The term may be used for a calendar day only, for which it is particularly applicable to temperature. At time the term is applied to the average of the highest and lowest temperatures as mean monthly or mean annual extremes.

Range:

Range is the difference between the highest and lowest values and reflects the extreme variations of these values. Since it has a high variability, this statistic is not recommended for precise work. Range is related to extreme values of record and can be useful in determining the extreme range for the records available.

Frequency:

Frequency is defined as the number of times a certain value occurs within a specified period of time. A frequency distribution may be used to present a condensed presentation of large data.

Mode:

Mode is defined as the value occurring with the greatest frequency or the value about which the most cases occur.

Median:

The median is the value at the midpoint in an array. It is determined by arranging all values in order of size. Rough estimates of the median may be obtained by taking the middle value of an ordered series; if there are two middle values they may be averaged to obtain the median. The median is not widely used in climatological computations. A longer period of record might be required to formulate an accurate median.

Degree-Day:

A degree-day is the number of degrees the mean daily temperature is above or below a standard temperature base. Degree days are accumulated over a season. At any point in the season, the total can be used as an index of past temperature effect upon some quantity, such as plant growth, fuel consumption, power output, etc. Degree-days are frequently applied to fuel and power consumption in the form of heating degree-days and cooling degree-days.

Average and standard deviations

In the analysis of climatological data, it may be desirable to compute the deviation of all items from a central point. This can be obtained from a computation of either the mean (or average) deviation or the standard deviation. These are termed measures of dispersion and are used to determine whether the average is truly representative or to determine the extent to which data vary from the average.

Average deviation :

Average deviation is obtained by computing the arithmetic average of the deviations from an average of the data.

Average deviation = $\sum$ d / n

where d (the deviations) and n is the number of items.

Standard deviation:

Standard deviation is the measure of the scatter or spread of all values in a series of observations.

Standard deviation = SQRT($\sum d^2$ / n)

where d^2 is the sum of the squared deviations from the arithmetic average, and n is the number of items in the group of data.

2.2 FORMS OF PRECIPITATION

Some of the common forms of precipitation are : rain, snow, drizzle, glaze, sleet and hail.

Rain

It is the principal form of precipitation in India. The term *rainfall* is used to describe precipitations in the form of water drops of sizes larger than 0.5 mm. The maximum size of a raindrop is about 6 mm. Any drop larger in size than this tends to break up into drops of smaller sizes during its fall from the clouds. On the basis of its intensity, rainfall is classified as :

Type	Intensity
1. Light rain	trace to 2.5 mm/h
2. Moderate rain	2.5 mm/h to 7.5 mm/h
3. Heavy rain	> 7.5 mm/h

Snow

Snow is another important form of precipitation. Snow consists of ice crystals which usually combine to form flakes. When new, snow has an initial density varying from 0.06 to 0.15 g/cm^3 and it is usual to assume an average density of 0.1 g/cm^3. In India, snow occurs only in the Himalayan regions.

Drizzle

A fine sprinkle of numerous water droplets of size less than 0.5 mm and intensity less than 1 mm/h is known as drizzle. In this the drops are so small that they appear to float in the air.

Glaze

When rain or drizzle comes in contact with cold ground at around 0° C, the water drops freeze to form an ice coating called *glaze* or *freezing rain.*

Sleet

It is frozen raindrops of transparent grains which form when rain falls through air at subfreezing temperature. In Britain, *sleet* denotes precipitation of snow and rain simultaneously.

Hail

It is a showery precipitation in the form of irregular pellets or lumps of ice of size more than 8 mm. Hails occur in violent thunderstorms in which vertical currents are very strong.

2.3 WEATHER SYSTEMS FOR PRECIPITATION

For the formation of clouds and subsequent precipitation, it is necessary that the moist air masses cool to form condensation. This is normally accomplished by adiabatic cooling of moist air through a process of being lifted to higher altitudes. Some of the terms and processes connected with the weather systems associated with precipitation are given below.

Front

A front is the interface between two distinct air masses. Under certain favourable conditions when a warm air mass and cold air mass meet, the warmer air mass is lifted over the colder one with the formation of a front. The ascending warmer air cools adiabatically with the consequent formation of clouds and precipitation.

Cyclone

A cyclone is a large low pressure region with circular wind motion. Two types of cyclones are recognised: tropical cyclones and extratropical cyclones.

Tropical cyclone: A tropical cyclone, also called *cyclone* in India, *hurricane* in USA and *typhoon* in South-East Asia, is a wind system with an intensely strong depression with MSL pressures sometimes below 915 mbars. The normal areal extent of a cyclone is about 100–200 km in diameter. The isobars are closely spaced and the winds are anticlockwise in the northern hemisphere. The centre of the storm, called the *eye*, which may extend to about 10–50 km in diameter, will be relatively quiet. However, right outside the eye, very strong winds reaching to as much as 200 kmph exist. The wind speed gradually decreases towards the outer edge. The pressure also increases outwards (Fig. 2.1). The rainfall will normally be heavy in the entire area occupied by the cyclone.

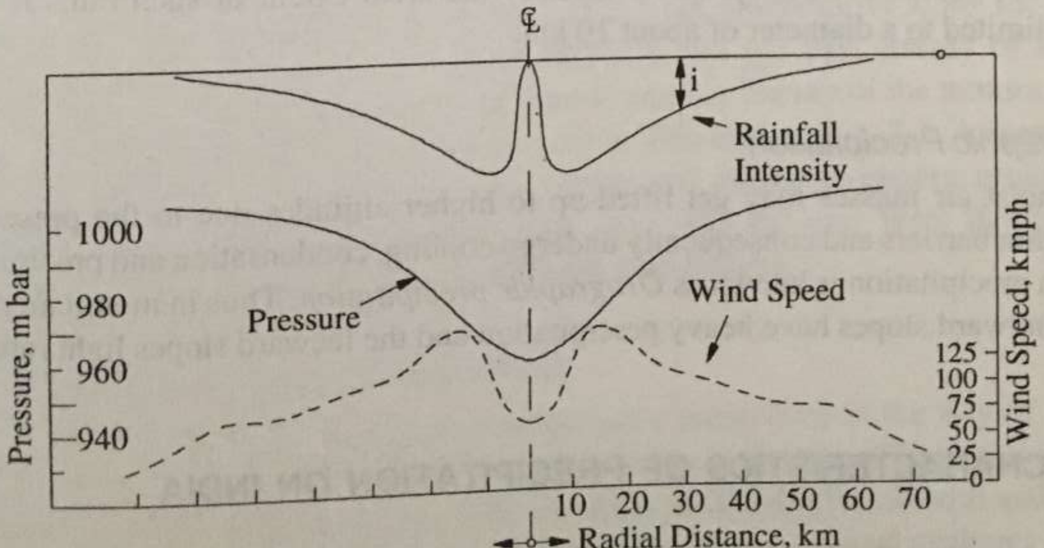


Fig. 2.1 Schematic section of a tropical cyclone

During summer months, tropical cyclones originate in the open ocean at around 5–10° Latitude and move at speeds of about 10–30 kmph to higher latitudes in an irregular path. They derive their energy from the latent heat of condensation of ocean water vapour and increase in size as they move on oceans. When they move on land the source of energy is cut off and the cyclone dissipates its energy very fast. Hence, the intensity of the storm decreases rapidly. Tropical cyclones cause heavy damage to life and property on their land path and intense rainfall and heavy floods in streams are its usual consequences. Tropical cyclones give moderate to excessive precipitation over very large areas, of the order of 10^3 km^2, for several days.

Extratropical cyclone: These are cyclones formed in locations outside the tropical zone. Associated with a frontal system, they possess a strong counter-clockwise wind circulation in the northern hemisphere. The magnitude of precipitation and wind velocities are relatively lower than those of a tropical cyclone. However, the duration of precipitation is usually longer and the areal extent also is larger.

Anticyclones

These are regions of high pressure, usually of large areal extent. The weather is usually calm at the centre. Anticyclones cause clockwise wind circulations in the northern hemisphere. Winds are of moderate speed, and at the outer edges, cloudy and precipitation conditions exist.

Convective Precipitation

In this type of precipitation a packet of air which is warmer than the surrounding air due to localised heating rises because of its lesser density. Air from cooler surroundings flows to take up its place thus setting up a convective cell. The warm air continues to rise, undergoes cooling and results in precipitation. Depending upon the moisture, thermal and other conditions light showers to thunderstorms can be expected in convective precipitation. Usually the areal extent of such rains is small, being limited to a diameter of about 10 km.

Orographic Precipitation

The moist air masses may get lifted-up to higher altitudes due to the presence of mountain barriers and consequently undergo cooling, condensation and precipitation. Such a precipitation is known as *Orographic precipitation*. Thus in mountain ranges, the windward slopes have heavy precipitation and the leeward slopes light rainfall.

2.4 CHARACTERISTICS OF PRECIPITATION ON INDIA

From the point of view of climate the Indian subcontinent can be considered to have two major seasons and two transitional periods as :

1. South-west monsoon (June–Setember)
2. Transition-I, post-monsoon (October–November)
3. Winter season (December–February)
4. Transition-II, Summer, (March–May)

The chief precipitation characteristics of these seasons are given below.

South–West Monsoon (June–September)

The south-west monsoon (popularly known as *monsoon*) is the principal rainy season of India when over 75% of the annual rainfall is received over a major portion of the country. Excepting the south-eastern part of the peninsula and Jammu and Kashmir, for the rest of the country the south-west monsoon is the principal sources of rain with July as the rainiest month. The monsoon originates in the Indian ocean and heralds its appearance in the southern part of Kerala by the end of May. The onset of monsoon is accompanied by high south-westerly winds at speeds of 30–70 kmph and low-pressure regions at the advancing edge. The monsoon winds advance across the country in two branches: (i) the Arabian sea branch and (ii) the Bay of Bengal branch. The former sets in at the extreme southern part of Kerala and the latter at Assam, almost simultaneously in the first week of June. The Bay branch first covers the north eastern regions of the country and turns westwards to advance into Bihar and UP. The Arabian sea branch moves northwards over Karnataka, Maharashtra and Gujarat. Both the branches reach Delhi around the same time by about the fourth week of June. A low-pressure region known as *monsoon trough* is formed between the two branches. The trough extends from the Bay of Bengal to Rajasthan and the precipitation pattern over the country is generally determined by its position. The monsoon winds increase from June to July and begin to weaken in September. The withdrawal of the monsoon, marked by a substantial rainfall activity starts in September in the northern part of the country. The onset and withdrawal of the monsoon at various parts of the country are shown in Fig. 2.2.

The monsoon is not a period of continuous rainfall. The weather is generally cloudy with frequent spells of rainfall. Heavy rainfall activity in various parts of the country owing to the passage of low pressure regions is common. Depressions formed in the Bay of Bengal at a frequency of 2-3 per month move along the trough causing excessive precipitation of about 100-200 mm per day. Breaks of about a week in which the rainfall activity is the least is another feature of the monsoon. The south-west monsoon rainfall over the country is indicated in Fig. 2.3. As seen from this figure, the heavy rainfall areas are Assam and the north-eastern region with 200-400 cm; west coast and western ghats with 200-300 cm; West Bengal with 120-160 cm, UP, Haryana and the Punjab with 100-120 cm.

Post–Monsoon (October–November)

As the south-west monsoon retreats, low-pressure areas form in the Bay of Bengal and a north-easterly flow of air that picks up moisture in the Bay of Bengal is formed. This air mass strikes the east coast of the southern peninsula (Tamilnadu) and causes rainfall. Also, in this period, specially in November, severe tropical cyclones form in the Bay of Bengal and the Arabian sea. The cyclones formed in the Bay of Bengal are

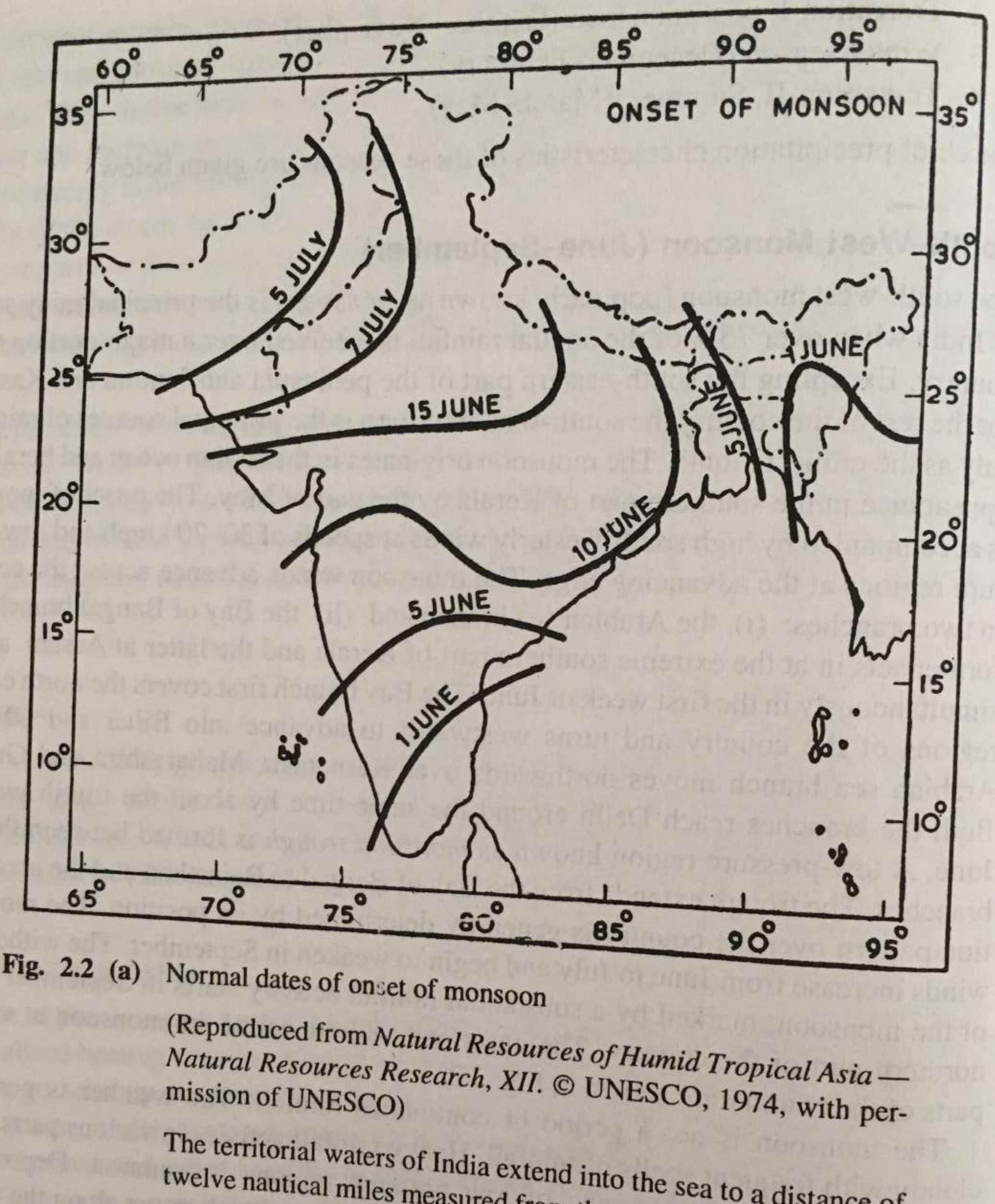


Fig. 2.2 (a) Normal dates of onset of monsoon

(Reproduced from *Natural Resources of Humid Tropical Asia — Natural Resources Research, XII*. © UNESCO, 1974, with permission of UNESCO)

The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate baseline

Responsibility for the correctness of the internal details on the map rests with the publisher.

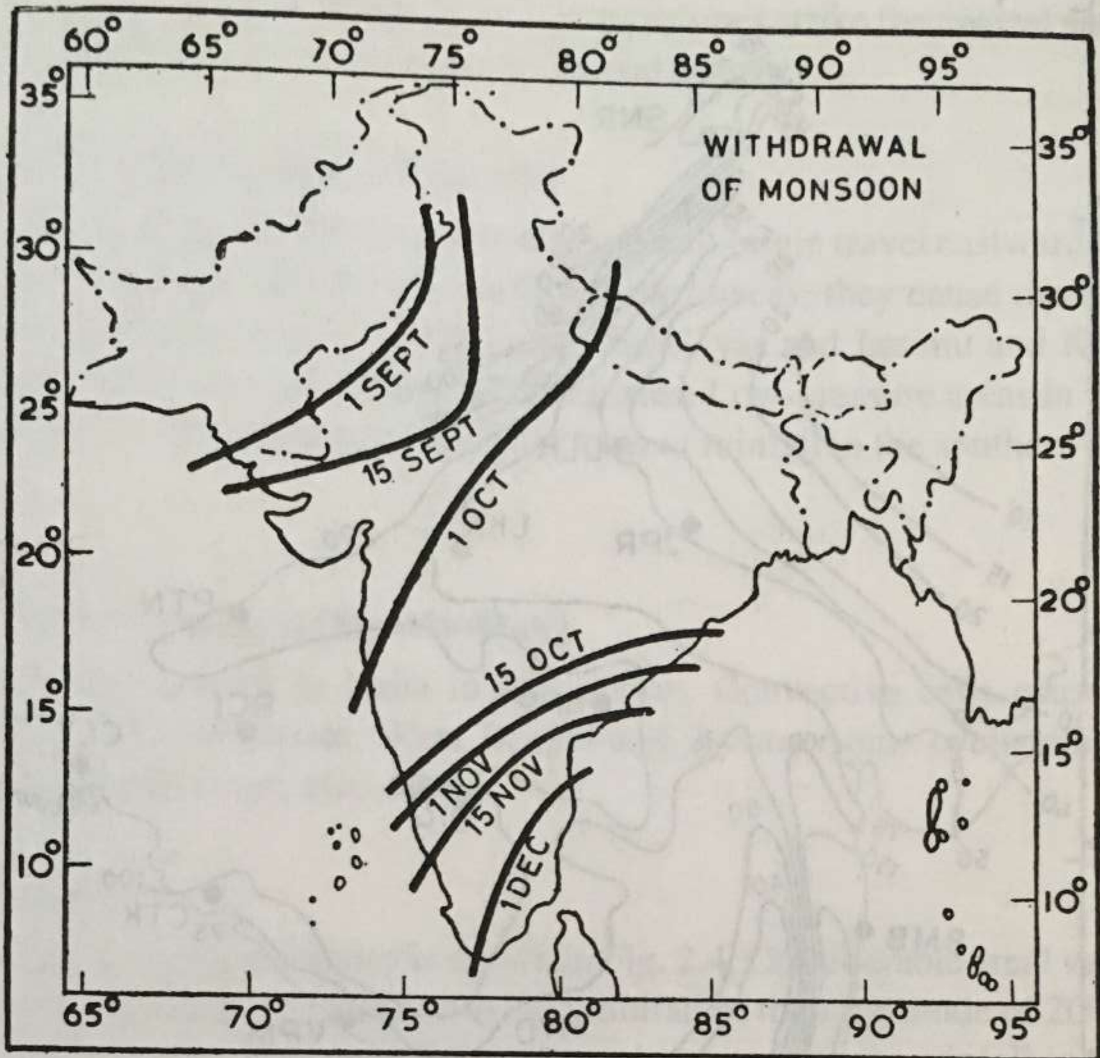


Fig. 2.2 (b) Normal dates of withdrawal of monsoon

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The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate baseline

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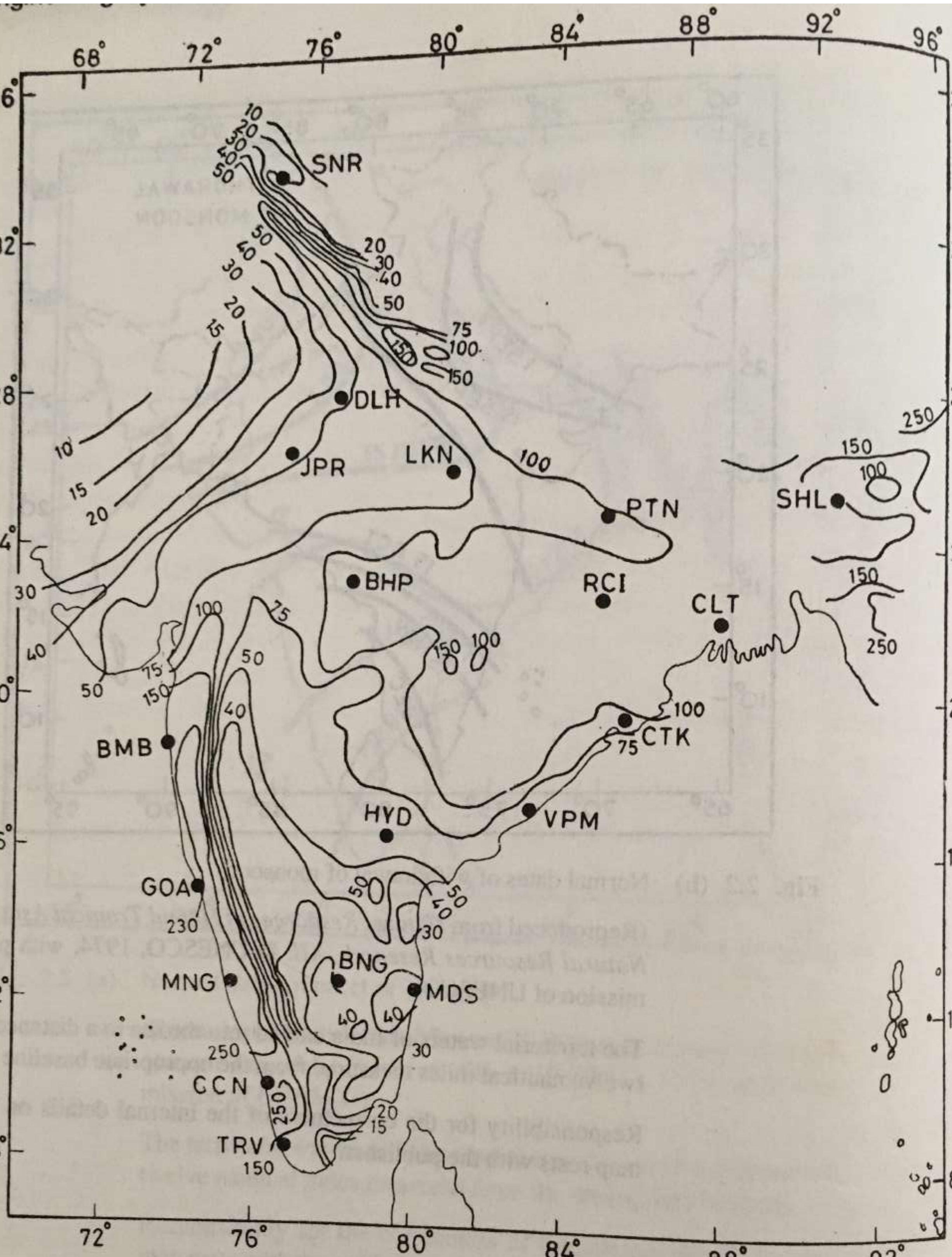

Fig. 2.3 Southwest monsoon rainfall (cm) over India and neighbourhood

(Reproduced with permission from India Meteorological Department) about twice as many as in the Arabian sea. These cyclones strike the coastal areas and cause intense rainfall and heavy damage to life and property.

Winter Season (December–February)

By about mid-December, disturbances of extra tropical origin travel eastwards across Afghanistan and Pakistan. Known as *western disturbances*, they cause moderate to heavy rain and snowfall (about 25 cm) in the Himalayas and Jammu and Kashmir. Some light rainfall also occurs in the northern plains. Low-pressure areas in the Bay of Bengal formed in these months cause 10–12 cm of rainfall in the southern parts of Tamilnadu.

Summer (Pre–monsoon) (March–May)

There is very little rainfall in India in this season. Convective cells cause some thunderstorms mainly in Kerala, West Bengal and Assam. Some cyclone activity, dominantly on the east coast, also occurs.

Annual Rainfall

The annual rainfall over the country is shown in Fig. 2.4. Considerable areal variation exists for the annual rainfall in India with high rainfall of the magnitude of 200 cm in Assam and north-eastern parts and the western ghats, and scanty rainfall in eastern Rajasthan and parts of Gujarat, Maharashtra and Karnataka. The average annual rainfall for the entire country is estimated[5] as 119 cm.

It is well-kown that there is considerable variation of annual rainfall in time at a place. The coefficient of variation,

$$C_v = \frac{100 \times \text{standard deviation}}{\text{mean}}$$

of the annual rainfall varies between 15 to 70, from place to place with an average value of about 30. Variability is least in regions of high rainfall and largest in regions of scanty rainfall. Gujarat, Haryana, Punjab and Rajasthan have large variability of rainfall.

2.5 MEASUREMENT

Precipitation is expressed in terms of the depth to which rainfall water would stand on an area if all the rain were collected on it. Thus 1 cm of rainfall over a catchment area of 1 km^2 represents a volume of water equal to 10^4 m^3. In the case of snowfall, an equivalent depth of water is used as the depth of precipitation. The precipitation is collected and measured in a *raingauge*. Terms such as *pluviometer*, *ombrometer* and *hyetometer* are also sometimes used to designate a raingauge.

A raingauge essentially consists of a cylindrical-vessel assembly kept in the open to collect rain. The rainfall catch of the raingauge is affected by its exposure conditions. To enable the catch of raingauge to accurately represent the rainfall in the

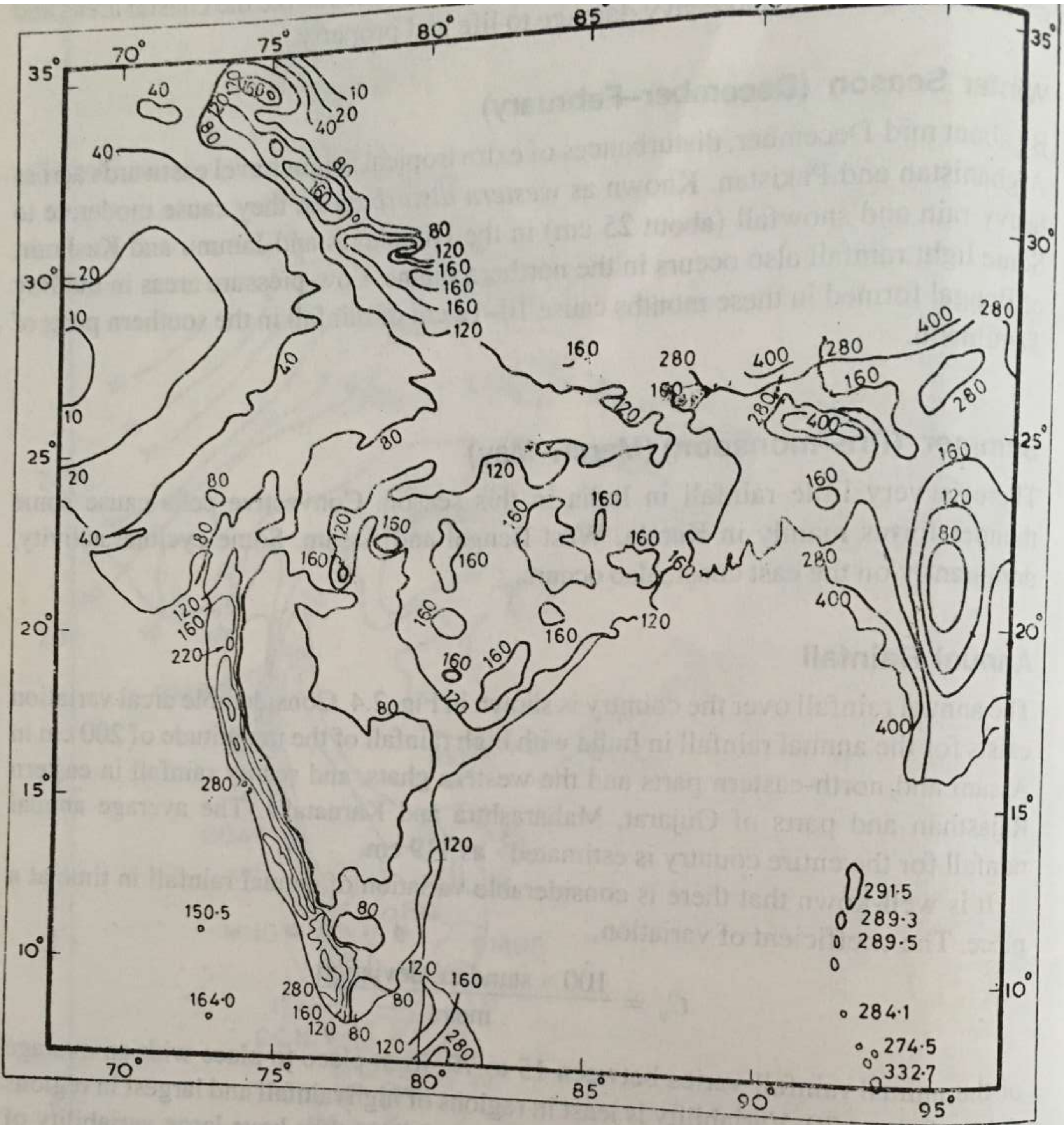

Fig. 2.4 Annual rainfall (cm) over India and neighbourhood

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Based upon Survey of India map with the permission of the Surveyor General of India © Government of India Copyright 1984

The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate baseline

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area surrounding the raingauge standard settings are adopted. For siting a raingauge the following considerations are important:

1. The ground must be level and in the open and the instrument must present a horizontal catch surface.
2. The gauge must be set as near the ground as possible to reduce wind effects but it must be sufficiently high to prevent splashing, flooding etc.
3. The instrument must be surrounded by an open fenced area of at least $5.5 \text{ m} \times 5.5 \text{ m}$. No object should be nearer to the instrument than 30 m or twice the height of the obstruction.

Raingauge can be broadly classified into two categories as (i) nonrecording raingauges and (ii) recording gauges.

Nonrecording Gauges

The nonrecording gauge extensively used in India is the *Symons' gauge*. It essentially consists of a circular collecting area of 12.7 cm (5.0 inch) diameter connected to a funnel. The rim of the collector is set in a horizontal plane at a height of 30.5 cm above the ground level. The funnel discharges the rainfall catch into a receiving vessel. The funnel and receiving vessel are housed in a metallic container. Figure 2.5 shows the details of the installation. Water contained in the receiving vessel is measured by a suitably graduated measuring glass, with an accuracy up to 0.1 mm.

Recently, the India Meteorological Department (IMD) has changed over to the use of fibreglass reinforced polyester raingauges, which is an improvement over the Symons' gauge. These come in different combinations of collector and bottle. The collector is in two sizes having areas of 200 and 100 cm^2 respectively. Indian standard (IS : 5225–1969) gives details of these new raingauges.

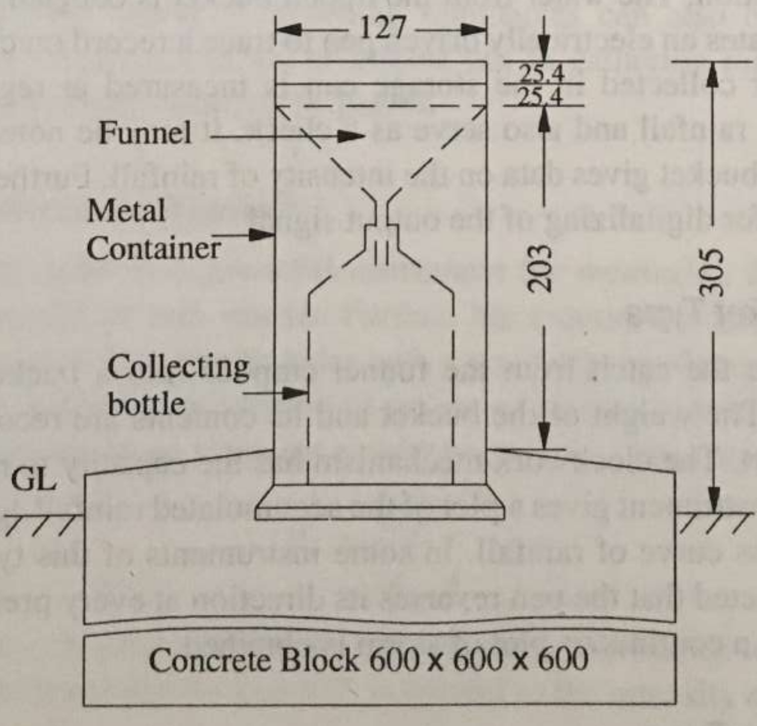

Fig. 2.5 Nonrecording raingauge (Symons' gauge)

For uniformity, the rainfall is measured every day at 8.30 AM (IST) and is recorded as the rainfall of that day. The receiving bottle normally does not hold more than 10 cm of rain and as such in the case of heavy rainfall the measurements must be done more frequently and entered. However, the last reading must be taken at 8.30 AM and the sum of the previous readings in the past 24 h entered as total of that day. Proper care, maintenance and inspection of raingauges, especially during dry weather to keep the instrument free from dust and dirt is very necessary. The details of installation of nonrecording raingauges and measurement of rain are specified in Indian Standard (IS : 4986–1968).

This raingauge can also be used to measure snowfall. When snow is expected, the funnel and receiving bottle are removed and the snow is allowed to collect in the outer metal container. The snow is then melted and the depth of resulting water measured. Antifreeze agents are sometimes used to facilitate melting of snow. In areas where consideable snowfall is expected, special snowgauges with shields (for minimizing the wind effect) and storage pipes (to collect snow over longer durations) are used.

Recording Gauges

Recording gauges produce a continuous plot of rainfall against time and provide valuable data of intensity and duration of rainfall for hydrological analysis of storms. The following are some of the commonly used recording raingauges.

Tipping–Bucket Type

This is a 30.5 cm size raingauge adopted for use by the US Weather Bureau. The catch from the funnel falls onto one of a pair of small buckets. These buckets are so balanced that when 0.25 mm of rainfall collects in one bucket, it tips and brings the other one in position. The water from the tipped bucket is colleted in a storage can. The tipping actuates an electrically driven pen to trace a record on clockwork-driven chart. The water collected in the storage can is measured at regular intervals to provide the total rainfall and also serve as a check. It may be noted that the record from the tipping bucket gives data on the intensity of rainfall. Further, the instrument is ideally suited for digitalizing of the output signal.

Weighing–Bucket Type

In this raingauge the catch from the funnel empties into a bucket mounted on a weighing scale. The weight of the bucket and its contents are recorded on a clock-work-driven chart. The clockwork mechanism has the capacity to run for as long as one week. This instrument gives a plot of the accumulated rainfall against the elapsed time, i.e. the mass curve of rainfall. In some instruments of this type the recording unit is so constructed that the pen reverses its direction at every preset value, say 7.5 cm (3 in.) so that a continuous plot of storm is obtained.

Natural–Syphon Type

This type of recording raingaguge is also known as *float-type gauge*. Here the rainfall collected by a funnel-shaped collector is led into a float chamber causing a float to rise. As the float rises, a pen attached to the float through a lever system records the elevation of the float on a rotating drum driven by a clockwork mechanism. A syphon arrangement empties the float chamber when the float has reached a pre-set maximum level. This type of raingauge is adopted as the standard recording-type raingauge in India and its details are described in Indian Standard (IS : 5235–1969).

A typical chart from this type of raingauge is shown in Figure 2.6. This chart shows a rainfall of 53.8 mm in 30 h. The vertical lines in the pen-trace correspond to the sudden emptying of the float chamber by syphon action which resets the pen to zero level. It is obvious that the natural syphon-type recording raingauge gives a plot of the mass curve of rainfall.

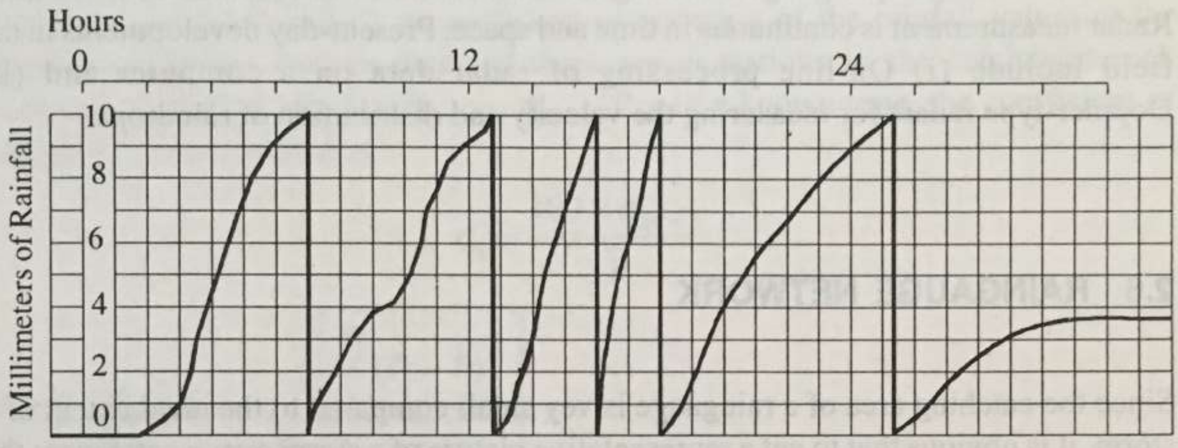

Fig. 2.6 Recording from a natural syphon–type gauge (schematic)

Telemetering Raingauges

These raingauges are of the recording type and contain electronic units to transmit the data on rainfall to a base station both at regular intervals and on interrogation. The tipping-bucket type raingauge, being ideally suited, is usually adopted for this purpose. Any of the other types of recording raingauges can also be used equally effectively. Telemetering gauges are of utmost use in gathering rainfall data from mountainous and generally inaccessible places.

Radar Measurement of Rainfall

The meteorological radar is a powerful instrument for measuring the areal extent, location and movement of rain storms. Further, the amounts of rainfall over large areas can be determined through the radar with a good degree of accuracy.

The radar emits a regular succession of pulses of electromagnetic radiation in a narrow beam. When raindrops intercept a radar beam, it has been shown that

$$P_r = \frac{C Z}{r^2} \tag{2.1}$$

where P_r = average echopower, Z = radar-echo factor, r = distance to target volume and C = a constant. Generally the factor Z is related to the intensity of rainfall as

$$Z = a I^b \tag{2.2}$$

rainfall depth equal to or greater than 280 mm at Madras occurring (a) once in 20 successive years (b) two times in 15 successive years and (c) at least once in 20 successive years.

SOLUTION : Here $P = \frac{1}{50} = 0.02$

By using Eq. (2.12) :

(a) $n = 20, r = 1$

$$P_{1,20} = \frac{20\,!}{19\,!\,1\,!} \times 0.02 \times (0.98)^{19}$$

$$= 20 \times 0.02 \times 0.68123 = 0.272$$

(b) $n = 15, r = 2$

$$P_{2,15} = \frac{15\,!}{13\,!\,2!} \times (0.02)^2 \times (0.98)^{13}$$

$$= 15 \times \frac{14}{2} \times 0.0004 \times 0.769 = 0.323$$

By Eq. (2.13)

(c)

$$P_1 = 1 - (1 - 0.02)^{20} = 0.332$$

Plotting Position

The purpose of the frequency analysis of an annual series is to obtain a relation between the magnitude of the event and its probability of exceedence. The probability analysis may be made either by empirical or by analytical methods.

A simple empirical technique is to arrange the given annual extreme series in descending order of magnitude and to assign an order number m. Thus for the first entry $m = 1$, for the second entry $m = 2$ and so on till the last event for which $m = N$ = Number of years of record. The probability P of an event equalled to or exceeded is given by the Weibull formula

$$P = \left(\frac{m}{N+1}\right) \tag{2.14}$$

The recurrence interval $T = 1/P = (N+1)/m$.

Equation (2.14) is an empirical formula and there are several other such empirical formulae available to calculate P (Table 2.2). The exceedence probability of the event obtained by the use of an empirical formula, such as Eq. (2.14) is called *plotting position*. Equation (2.14) is the most popular plotting position formula and hence only this formula is used in further sections of this book.

TABLE 2.2 PLOTTING POSITION FORMULAE

Method	P
California	m/N
Hazen	$(m-0.5)/N$
Weibull	$m/(N+1)$
Chegodayev	$(m-0.3)/(N+0.4)$
Blom	$(m-0.44)/(N+0.12)$
Gringorten	$(m-3/8)/(N+1/4)$

Having calculated P (and hence T) for all the events in the series, the variation of the rainfall magnitude is plotted against the corresponding T on a semi-log paper (Fig. 2.14) or log-log paper. By suitable extrapolation of this plot, the rainfall magnitude of specific duration for any recurrence interval can be estimated. This simple empirical procedure can give good results for small extrapolations and the errors increase with the amount of extrapolation. For accurate work, various analytical calculation procedures using frequency factors are available. Gumbel's extreme value distribution and Log Pearson Type III method are two commonly used analytical methods and are described in Chap. 7 of this book.