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Atmospheric Sciences Department

Climatology

Lectures for 3rd Class Students

1st Semester

By

Dr. ALAA M. AL-LAMI

Assistant Professor

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References

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Chapter 1. Description of the climate system and its components

1.1 Introduction

Climate is traditionally defined as the description in terms of the mean and variability of relevant atmospheric variables such as temperature, precipitation and wind. Climate can thus be viewed as a synthesis or aggregate of weather. This implies that the portrayal of the climate in a particular region must contain an analysis of mean conditions, of the seasonal cycle, of the probability of extremes such as severe frost and storms, etc. Following the World Meteorological Organisation (WMO), 30 years is the classical period for performing the **statistics** used to define climate. This is well adapted for studying recent decades since it requires a reasonable amount of data while still providing a good sample of the different types of weather that can occur in a particular area. However, when analysing the most distant past, such as the last glacial maximum around 20 000 years ago, climatologists are often interested in variables characteristic of longer time intervals. As a consequence, the 30-year period proposed by the WMO should be considered more as an indicator than a norm that must be followed in all cases. This definition of the climate as representative of conditions over several decades should, of course, not mask the fact that climate can change rapidly. Nevertheless, a substantial time interval is needed to observe a difference in climate between any two periods. In general, the less the difference between the two periods, the longer is the time needed to be able to identify with confidence any changes in the climate between them.

We must also take into account the fact that the state of the atmosphere used in the definition of the climate given above is influenced by numerous processes involving not only the atmosphere but also the ocean, the sea ice, the vegetation, etc. Climate is thus now more and more frequently defined in a wider sense as the statistical description of the **climate system**. This includes the analysis of the behavior of its five major components: the atmosphere (the gaseous envelope surrounding the Earth), the **hydrosphere** (liquid water, i.e. ocean, lakes, underground water, etc), the **cryosphere** (solid water, i.e. **sea ice**, glaciers, ice sheets, etc), the land surface and the **biosphere** (all the living organisms), and of the interactions between them (IPCC 2007, Fig. 1.1). We will use this wider definition when we use the word **climate**. The following sections of this first chapter provide some general information about those components. Note that the climate system itself is often considered as part of the broader **Earth System**, which includes all the parts of the Earth and not only the elements that are directly or indirectly related to the temperature or precipitation.

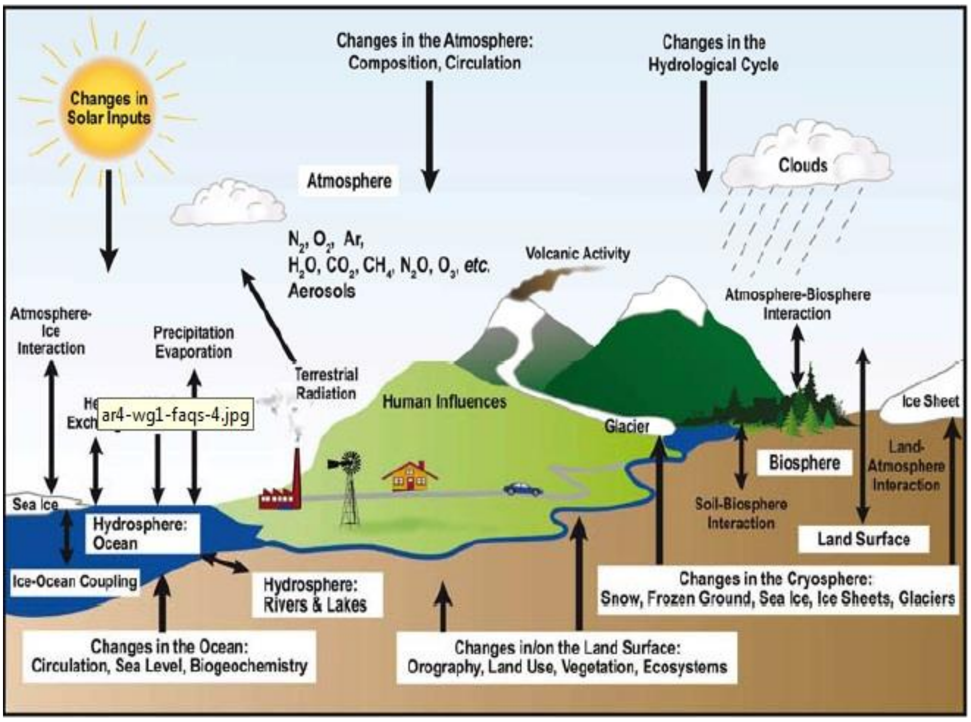


Figure 1.1: *Schematic view of the components of the climate system and of their potential changes.*

1.2 The Atmosphere

1.2.1 Composition and temperature

Dry air is mainly composed of nitrogen (78.08 % in volume), oxygen (20.95% in volume), argon (0.93% in volume) and to a lesser extent carbon dioxide1 (380 ppm or 0.038% in volume). The remaining fraction is made up of various trace constituents such as neon (18 ppm), helium (5 ppm), methane1 (1.75 ppm), and krypton (1 ppm). In addition, a highly variable amount of water vapour is present in the air. This ranges from approximately 0% in the coldest part of the atmosphere to as much as 5% in moist and hot regions. On average, water vapour accounts for 0.25% of the mass of the atmosphere.

On a large-scale, the atmosphere is very close to **hydrostatic quilibrium**, meaning that at a height *z*, the force due to the pressure *p* on a 1 m^2 horizontal surface balances the force due to the weight of the air above *z*. The atmospheric pressure is thus at its maximum at the Earth's surface and the surface pressure *ps* is directly related the mass of the whole air column at a particular location. Pressure then decreases with height, closely following an exponential law:

$$p \simeq p_s e^{-z/H} \tag{1.1}$$

where *H* is a scale height (which is between 7 and 8 km for the lowest 100 km of the atmosphere). Because of this clear and monotonic relationship between height and pressure, pressure is often used as a vertical coordinate for the atmosphere. Indeed, pressure is easier to measure than height and choosing a pressure coordinate simplifies the formulation of some equations.

The temperature in the **troposphere**, roughly the lowest 10 km of the atmosphere, generally decreases with height. The rate of this decrease is called the **lapse rate** Γ:

$$p \simeq p_s e^{-z/H} \qquad (1.1)$$

where *T* is the temperature. The lapse rate depends mainly on the adiative balance of the atmosphere (see section 2.1) and on **convection** as well as on the horizontal heat transport. Its global mean value is about 6.5 K km^{-1}, but Γ varies with the location and season.

The lapse rate is an important characteristic of the atmosphere. For instance, it determines its vertical stability. For low values of the lapse rate, the atmosphere is very stable, inhibiting vertical movements. Negative lapse rates (i.e. temperature increasing with height), called temperature inversions, correspond to highly stable conditions. When the lapse rate rises, the stability decreases, leading in some cases to vertical instability and convection. The lapse rate is also involved in **feedbacks** playing an important role in the response of the climate system to a perturbation.

At an altitude of about 10 km, a region of weak vertical temperature **gradients**, called the **tropopause**, separates the **troposphere** from the **stratosphere** where the temperature generally increases with height until the stratopause at around 50 km (Fig. 1.2). Above the stratopause, temperature decreases strongly with height in the mesosphere, until the mesopause is reached at an altitude of about 80 km, and then increases again in the thermosphere above this height. The vertical **gradients** above 10 km are strongly influenced by the absorption of solar **radiation** by different atmospheric constituents and by chemical reactions driven by the incoming light. In particular, the warming in the stratosphere at heights of about 30-50 km is mostly due to the absorption of ultraviolet **radiation** by stratospheric **ozone**, which protects life on Earth from this dangerous **radiation**.

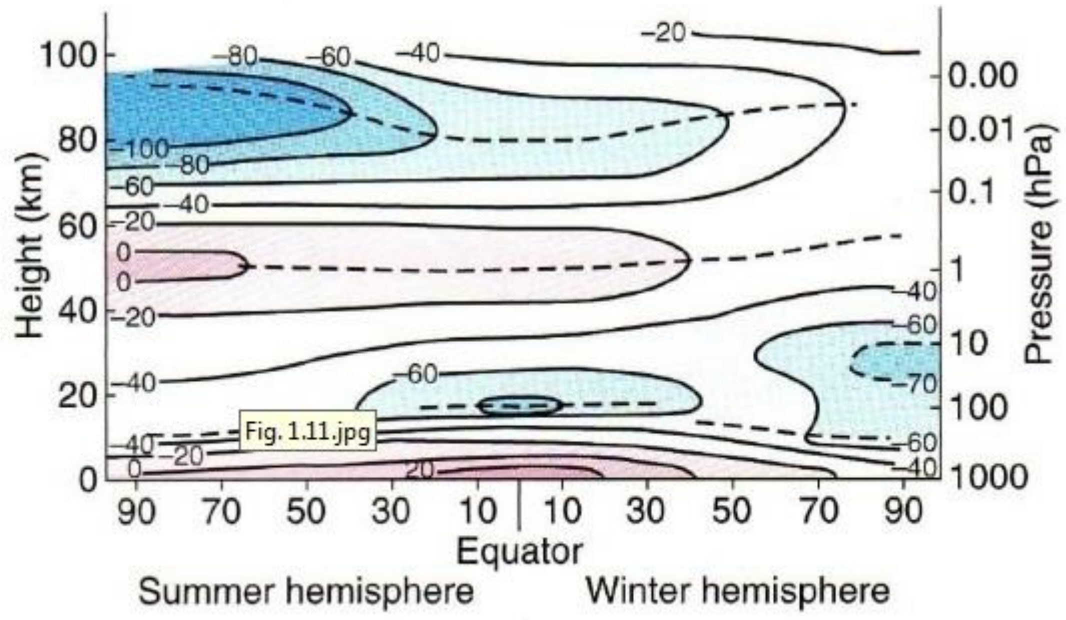


Figure 1.2: *Idealised* ***zonal*** *mean temperature (in °C) in the atmosphere as a function of the height (or of the pressure). The dashed lines represent schematically the location of the tropopause, stratopause and mesopause.*

Atmospheric **specific humidity** also displays a characteristic vertical profile with maximum values in the lower levels and a marked decrease with height. As a consequence, the air above the tropopause is nearly dry. This vertical distribution is mainly due to two processes. First, the major source of atmospheric water vapour is evaporation at the surface. Secondly, the warmer air close to the surface can contain a much larger quantity of water before it becomes saturated than the colder air further away; saturation that leads to the formation of water or ice droplets, clouds and eventually precipitation.

At the Earth's surface, the temperature reaches its maximum in equatorial regions (Fig. 1.3) because of the higher incoming radiations. In those regions, the temperature is relatively constant throughout the year. Because of the much stronger seasonal cycle at mid and high latitudes, the north-south gradients are much larger in winter than in summer. The distribution of the surface temperature is also influenced by atmospheric and oceanic heat transport as well as by the thermal inertia of the ocean. Furthermore, the role of topography is important, with a temperature decrease at higher altitudes associated with the positive lapse rate in the troposphere.

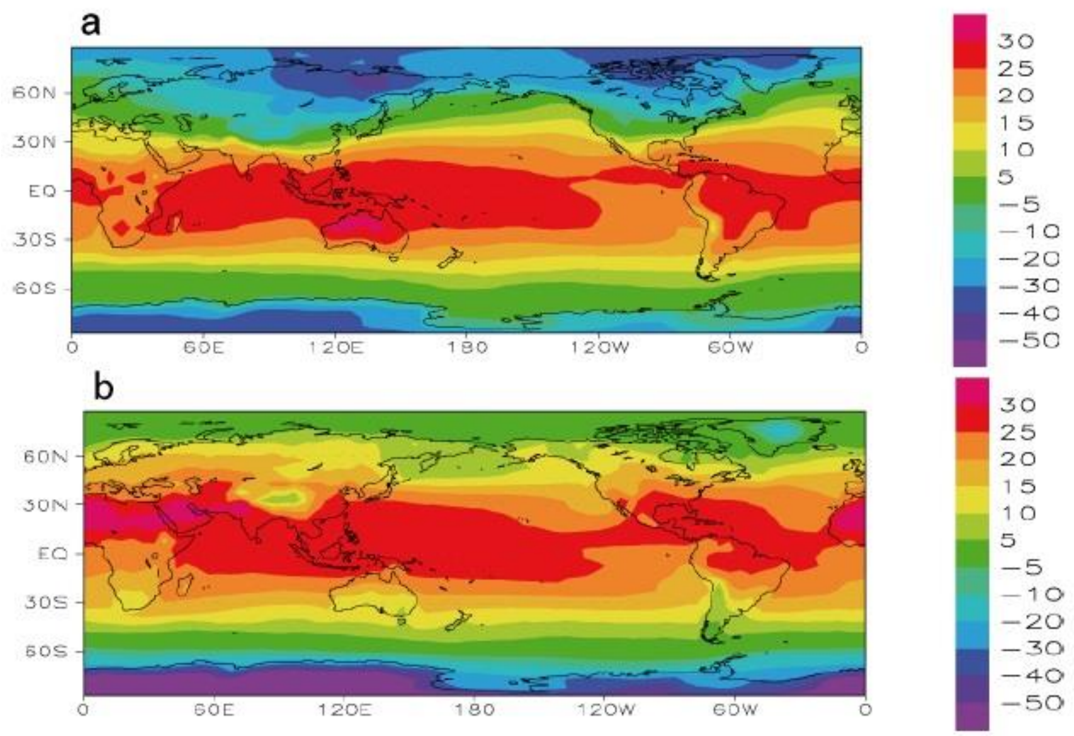


Figure 1.3: *Surface air temperature (in °C) averaged over (a) December, January, and February and (b) June, July, and August.*

1.3 The ocean

1.3.1 Composition and properties

The ocean has a major impact on the climate of the atmosphere. It covers approximately 71 per cent of the Earth's surface and thus has a dominant role for transfers of energy and other properties between the atmosphere and the Earth's surface. Its large heat capacity, made accessible for surface energy transfers by circulations within the ocean, provides a moderating effect on temperature variability in the atmosphere. Oceanic currents transfer large amounts of heat energy away from equatorial regions. Finally, the ocean is an important source for atmospheric water vapour, as well as a source and sink for other greenhouse gases.

The ocean's heat capacity exceeds that of the atmosphere by a factor of the order of 1000. This is due to differences both in heat capacity per unit mass (the specific heat of liquid water is about four times that of air), and in total mass between the ocean and atmosphere. The ocean's heat capacity bears upon atmospheric temperature through oceanic transports (both horizontal and vertical) that produce and maintain surface water temperatures warmer or colder than the atmosphere resulting in large heat transfers. The depth to which the oceans interact with the atmosphere depends on the time scale under consideration. For diurnal variations the depth is small, of the order of five to 10 meters. For seasonal variations the depth is 20-200 meters (the depth of the well-mixed oceanic surface layer). The ocean is a major component in determining the climate and its variations for annual, annually-averaged and longer period conditions.

The strong influence of the ocean on surface air temperature is clearly evident.

the oceanic condition is different from the atmosphere in two fundamental ways. First, the primary forcing of the ocean is at the upper boundary, whereas the primary forcing for the atmosphere is at its lower boundary. Atmospheric winds above the ocean are a major factor in causing ocean surface currents through surface friction processes. In contrast, for the atmosphere frictional conditions at its lower boundary tend to reduce atmospheric motion. Second, the density of the ocean water is determined primarily by its salinity and temperature instead of pressure, temperature and water vapour content as in the atmosphere. The water vapour factor is relatively unimportant for atmospheric density except in hot and humid conditions. On the other hand, salinity can play a major role for ocean density especially when temperatures are near freezing in which case density changes very little with temperature. Salinity conditions in polar ocean regions are important for determining whether or not significant vertical motions occur in local areas (see figure 1.4).

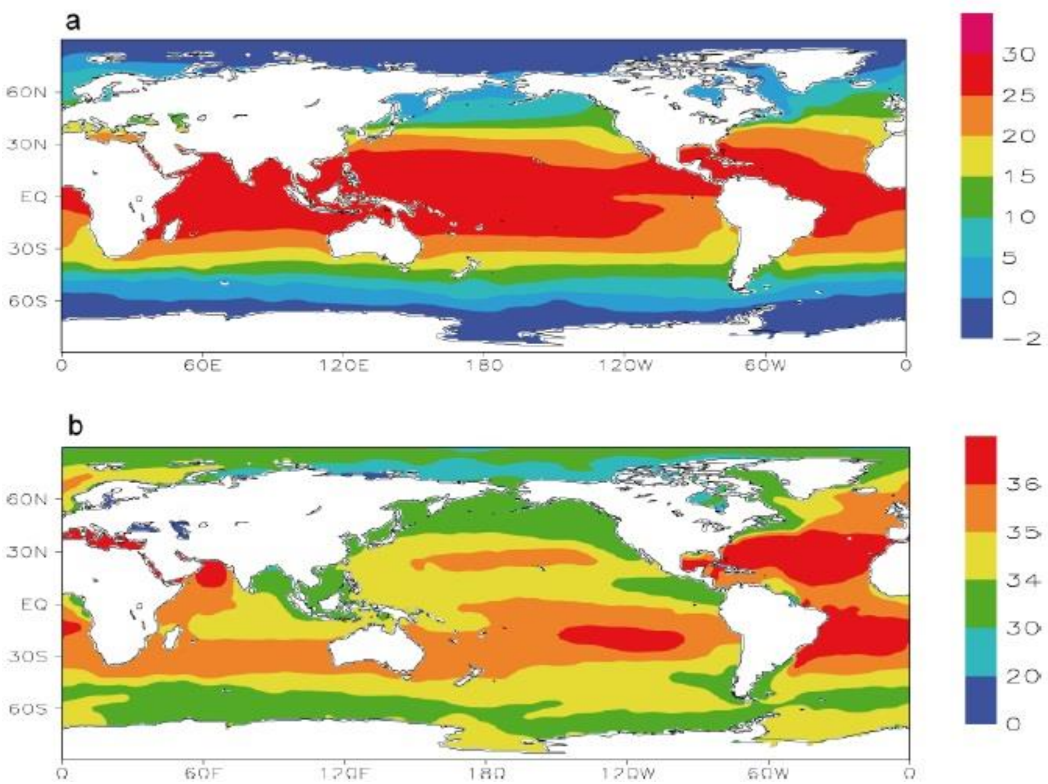


Figure 1.4: *(a) Annual mean sea surface temperature (°C) and (b) surface salinity (psu). Data source: Levitus (1998).*

1.4 The cryosphere

1.4.1 Components of the cryosphere

The cryosphere is the portion of the Earth's surface where water is in solid form. It thus includes sea ice, lake ice and river-ice, snow cover, glaciers, ice caps and ice sheets, and frozen ground. The snow cover has the largest extent, with a maximum area of more than 45 106 km2 (Table 1.1). Because of the present distribution of continents, land surfaces at high latitudes are much larger in the Northern Hemisphere than in the Southern Hemisphere. As a consequence, the large majority of the snow cover is located in the Northern Hemisphere (Figs. 1.5 and 1.6). The same is true for the freshwater ice that forms on rivers and lakes in winter. Both the snow cover and freshwater ice have a very strong seasonal cycle, as they nearly disappear in summer in both hemispheres (Table 1.1).

Table 1.1: *Areal extent and volume of snow cover and sea ice. Data compiled in Climate and Cryosphere (CliC) project science and co-ordination plan (2001).*

Component	Maximum area (10^6 km^2)	Minimum area (10^6 km^2)	Maximum Ice volume (10^6 km^3)	Minimum Ice volume (10^6 km^3)
Northern Hemisphere Snow cover	46.5 (late January)	3.9 (late August)	0.002	
Southern Hemisphere Snow cover	0.83 (late July)	0.07 (early May)		
Sea ice in the Northern Hemisphere	14.0 (late March)	6.0 (early September)	0.05	0.02
Sea ice in the Southern Hemisphere	15.0 (late September)	2.0 (late February)	0.02	0.002

The cryosphere — the ice component — has significant impacts on the climate system in several ways. It affects radiative and sensible heat transfers at the Earth's surface. It influences temperatures in the ocean and at the Earth's surface due to transfers between latent and sensible energy during melting and freezing. Finally, its melting and freezing influences water runoff from land and ocean salinity. Ice and snow exist primarily in the latitudes poleward of 30 degrees latitude and are thus are unfamiliar to the majority of the world's human population. Although only about two per cent of all the water on Earth is frozen, it covers an average of 11 per cent of the world's land surface and seven per cent of its oceans. There are many constituents to the cryosphere: land ice in polar ice sheets, glaciers, permafrost, frozen ground, seasonal snow cover and sea ice.

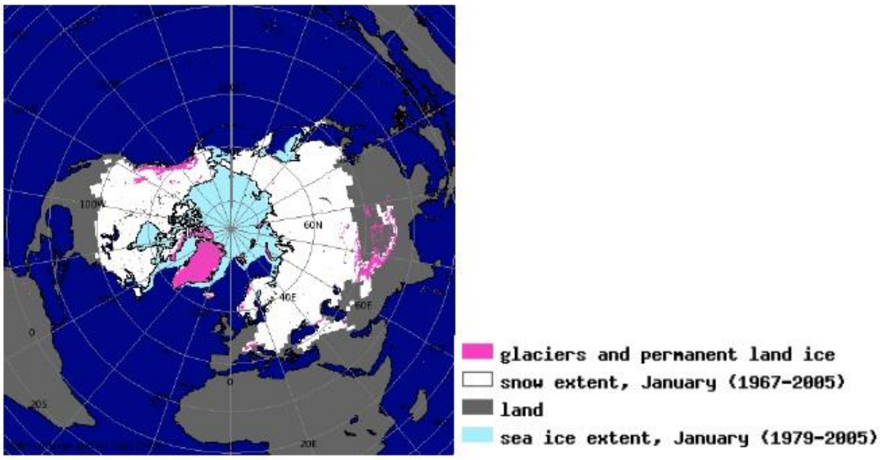


Figure 1.5: *The distribution of sea ice, snow and land ice in January in the Northern Hemisphere.*

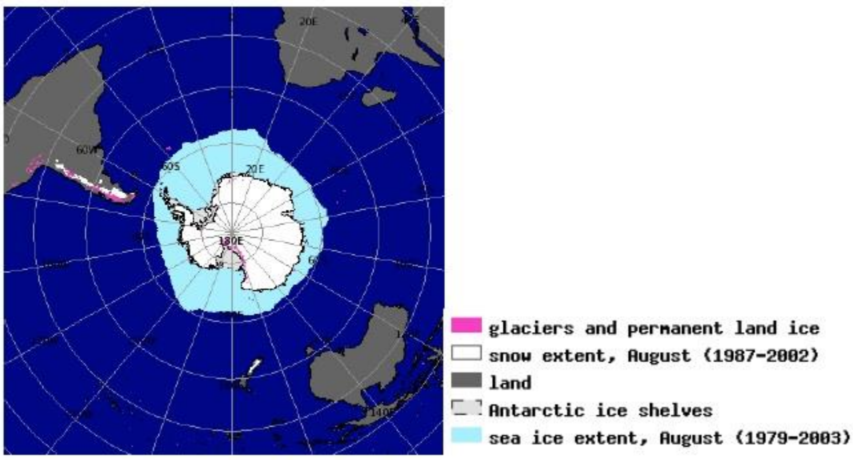


Figure 1.6: *Location of sea ice, snow and land ice in August in the Southern Hemisphere. Source;*

1.5 The land surface (Lithosphere)

Land surface is an important interactive component of the climate system. It covers 29 per cent of the Earth's surface. Significant exchanges of heat, moisture, and momentum occur between the atmosphere and the land surface, including its biosphere. It is also the surface on which people live. The heat storage factor of land surface with respect to atmospheric temperature variations is much less than that for the oceans. Land has a lower specific heat than the ocean, and its rigidity restricts heat transport to deeper levels. As a result, the depth of the soil layer which is important for energy exchange interactions with the atmosphere is only several meters for the annual-cycle time scale. A cave 20 meters underground will remain at the same temperature all year round. Because of the small heat capacity of the land surface, variations in atmospheric temperature just above the surface are much larger over the land than over the ocean. The energy and momentum exchanges between land surfaces and the atmosphere are similar to those for an ocean surface. Heat and latent heat (water vapour) exchanges depend on temperature and water vapour pressure differences between the land surface and the lower atmosphere, roughness of the land surface, and surface atmospheric wind speed. The latter may be characterized by wind conditions in the lowest ten meters of the atmosphere (the atmospheric 'mixed layer').

Radiation transfer is the other important energy exchange. The amount of solar radiation absorbed by a land surface depends on both the amount of solar radiation coming through the atmosphere (a highly variable quantity as discussed before) and the albedo (reflectivity) of the land surface which is also highly variable. The albedo ranges from five to 90 per cent and depends on the type of cover for the land surface as shown in Table 1.2. The infrared radiation transfer is the net of the infrared radiation emitted by the land surface (which is close to the maximum 'black body' value and thus dependent only on temperature) and the total downward infrared radiation produced by the atmosphere. Because of the small heat capacity of the land surface, the radiative, sensible, and latent energy transfers come close to balancing most of the time.

Table 1.2: *Typical range of the albedo of various surfaces.*

Surface type	**Albedo**
Ocean	0.05-0.15
Fresh snow	0.75-0.90
Old snow	0.40-0.70
Sea ice	0.3-0.6
Soil	0.05-0.40
Desert	0.20-0.45
Cropland	0.18-0.25
Grassland	0.16-0.26
Deciduous forest	0.15-0.20
Coniferous forest	0.05-0.15
Snow covered coniferous forest	0.13-0.3

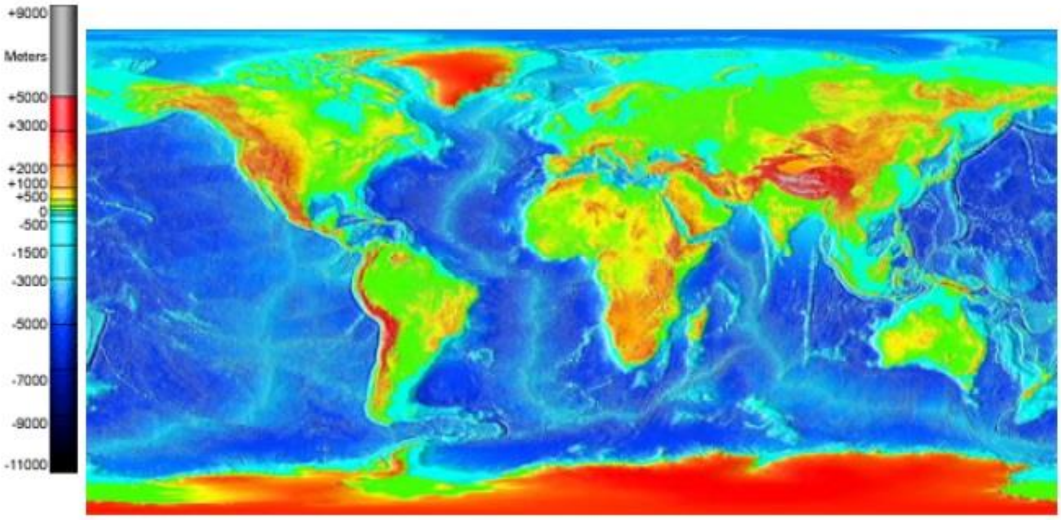


Figure 1.7: High resolution map of the surface topography.

Topography of the land surface has a pronounced effect on large-scale atmospheric circulations. particularly in the Northern Hemisphere. The Rocky Mountains, which are oriented north-south transect the Northern Hemisphere westerlies, and the Tibetan Plateau with its extreme height and aerial extent affects flow over a large area. Topography is a factor in the wave patterns in the upper tropospheric horizontal wind flow (shown in Figure 1.7), and also has major effects on surface temperature and rainfall. Alteration of land surface by human activity is an important factor in climate change that adds to the effects of human-produced changes in the radiative characteristics of the atmosphere.

Urbanization, cultivation for agriculture, irrigation, and deforestation change the albedo of land surfaces and the surface sensible and latent heat transfers. These factors can also greatly influence the local aspects of climate change.

1.6 The Biosphere

The biosphere is a component of the climate system that has a distinct role in the interactions of both the oceans and land surface with the atmosphere. Vegetation on the land surface and both plant and animal life in the oceans are all relevant elements of the biosphere component that interact with the atmosphere.

Climate conditions of the atmosphere have a direct effect on the type of terrestrial plant growth at the Earth's surface. The nature of the plant cover in turn feeds back on the atmospheric condition by influencing the sensible and latent energy transfers from a land surface, as well as surface layer turbulence in the atmosphere (through its roughness properties). Furthermore, land vegetation is a significant reservoir for carbon with a total carbon content nearly equal to that in the atmosphere. Changes in the amount of land vegetation due, for instance, to forest cutting and burning or simply seasonal changes have a direct impact on the carbon dioxide concentration in the atmosphere. Along with dissolved inorganic carbon and calcium carbonate solids, plant and animal life have key roles in the ocean, in sensible heat. Recall that in the global mean, the latent energy transfer from the Earth to the atmosphere was much larger than the sensible heat transfer (see figure 1.8).

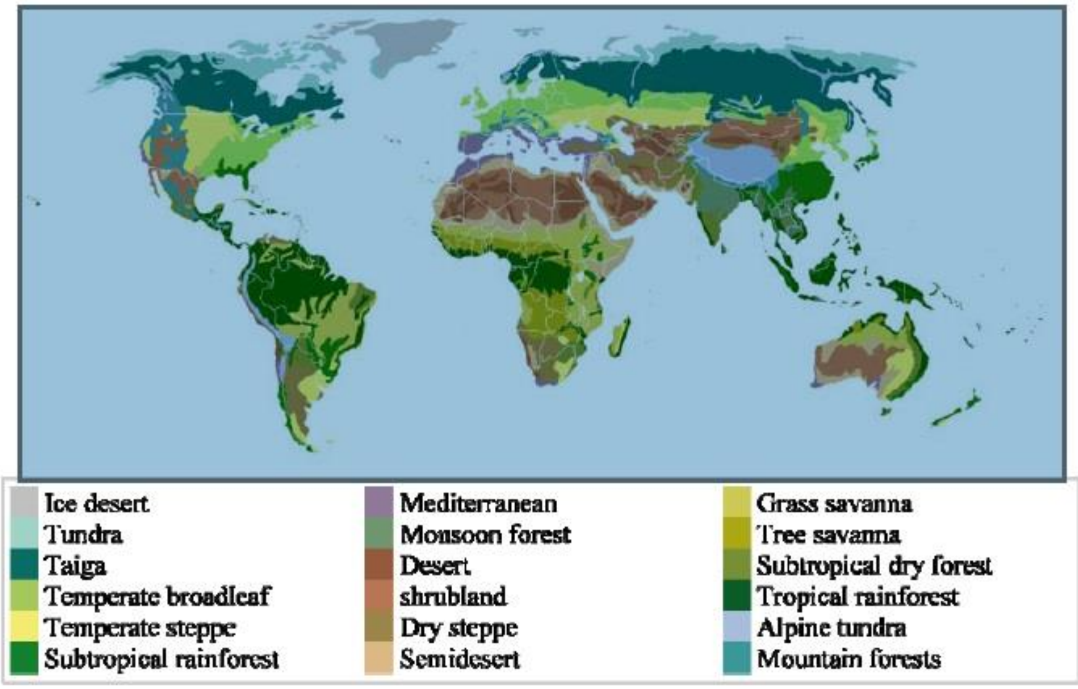


Figure (1.8): *Terrestrial biomes.*

Chapter Two

Radiation, energy, and climate

2.1 Heat Budget of the atmosphere

Almost all energy affecting the earth is derived from solar radiation, which is of short wavelength (<4μm) due to the high temperature of the sun (6000K) (i.e., Wien's Law). The solar constant has a value of approximately 1366W m–2. The sun and the earth radiate almost as black bodies (Stefan's Law, F =σT4), whereas the atmospheric gases do not. Terrestrial radiation, from an equivalent black body, amounts to only about 270W m–2 due to its low radiating temperature (263K); this is infrared (longwave) radiation between 4 and 100μm. Water vapor and carbon dioxide are the major absorbing gases for infrared radiation, whereas the atmosphere is largely transparent to solar radiation (the greenhouse effect). Trace-gas increases are now augmenting the 'natural'greenhouse effect (33K). Solar radiation is lost by reflection, mainly from clouds, and by absorption (largely by water vapor). The planetary albedo is 31 percent; 49 percent of the extraterrestrial radiation reaches the surface. The atmosphere is heated primarily from the surface by the absorption of terrestrial infrared radiation and by turbulent heat transfer. Temperature usually decreases with height at an average rate of about 6.5°C/km in the troposphere. In the stratosphere and thermosphere, it increases with height due to the presence of radiation absorbing gases.

The excess of net radiation in lower latitudes leads to a poleward energy transport from tropical latitudes by ocean currents and by the atmosphere. This is in the form of sensible heat (warm air masses/ocean water) and latent heat (atmospheric water vapor). Air temperature at any point is affected by the incoming solar radiation and other vertical energy exchanges, surface properties (slope, albedo, heat capacity), land and sea distribution and elevation, and also by horizontal advection due to air mass movements and ocean currents.

We can summarize the net effect of the transfers of energy in the earth–atmosphere system averaged over the globe and over an annual period:

The incident solar radiation averaged over the globe is:

$$\text{Solar constant} \times \pi r^2 / 4\pi r^2$$

where r = radius of the earth and $4\pi r^2$ is the surface area of a sphere. This figure is approximately 342W m^{-2}, or 11 · 109J m^{-2} yr^{-1} (10^9J = 1GJ); for convenience we will regard it as 100 units. Referring to Figure 2.1, incoming radiation is absorbed in the stratosphere (3 units), by ozone mainly, and 20 units are absorbed in the troposphere by carbon dioxide (1), water vapour (13), dust (3) and water droplets in clouds (3). Twenty units are reflected back to space from clouds, which cover about 62 percent of the earth's surface, on average. A further 9 units are similarly reflected from the surface and 3 units are returned by atmospheric scattering. The total reflected radiation is the *planetary albedo* (31 percent or 0.31). The remaining 49 units reach the earth either directly (Q = 28) or as diffuse radiation (q =21) transmitted via clouds or by downward scattering. The pattern of outgoing terrestrial radiation is quite different (see Figure 2.2). The black-body radiation, assuming a mean surface temperature of 288K, is equivalent to 114 units of infrared (longwave) radiation. This is possible because most of the outgoing radiation is reabsorbed by the atmosphere; the *net* loss of infrared radiation at the surface is only 19 units. These exchanges represent a time-averaged state for the whole globe. Recall that solar radiation affects only the sunlit hemisphere, where the incoming radiation exceeds 342W m^{-2}. Conversely, no solar radiation is received by the night-time hemisphere. Infrared exchanges continue, however, due to the accumulated heat in the ground. Only about 12 units escape through the atmospheric window directly from the surface. The atmosphere itself radiates 57 units to space (48 from the emission by atmospheric water vapor and CO2 and 9 from cloud emission), giving a total of 69 units (*Lu*); the atmosphere in turn radiates 95 units back to the surface (*Ld*); thus, *Lu* + *Ld* = *Ln* is negative.

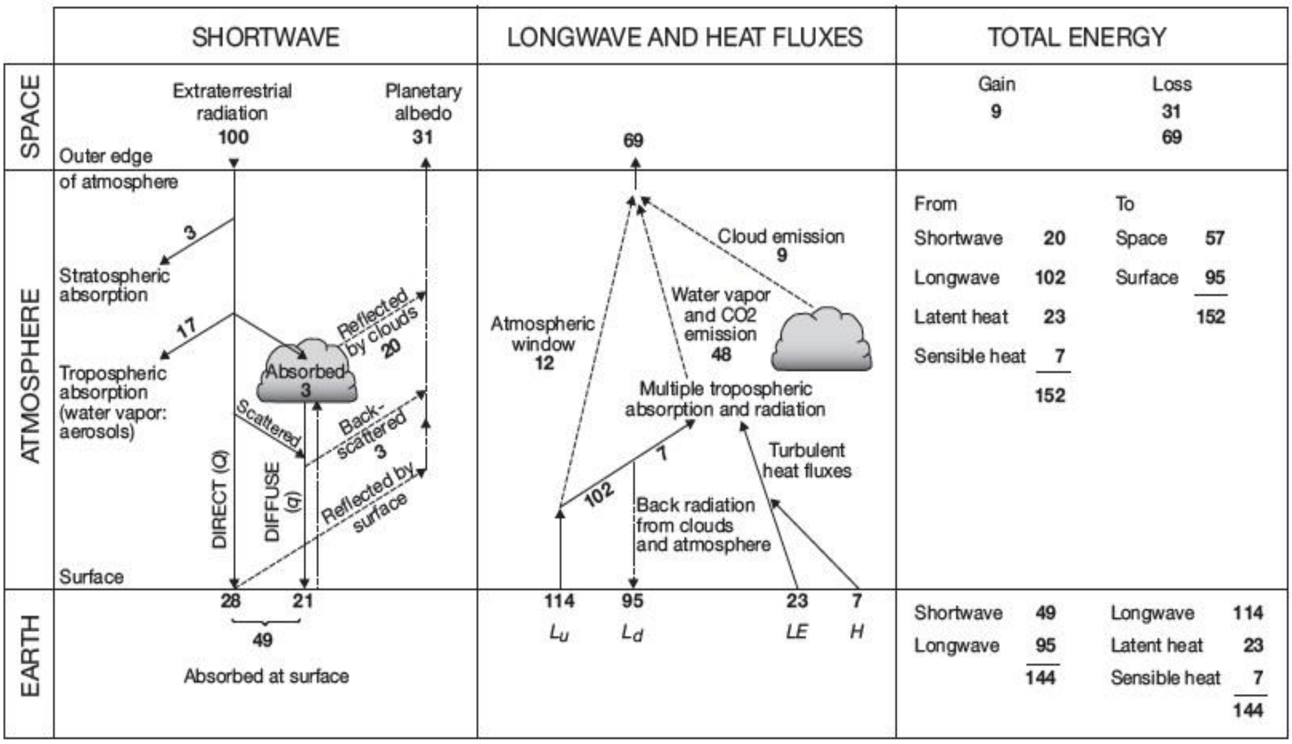


Figure 2.1 *The balance of the atmospheric energy budget.*

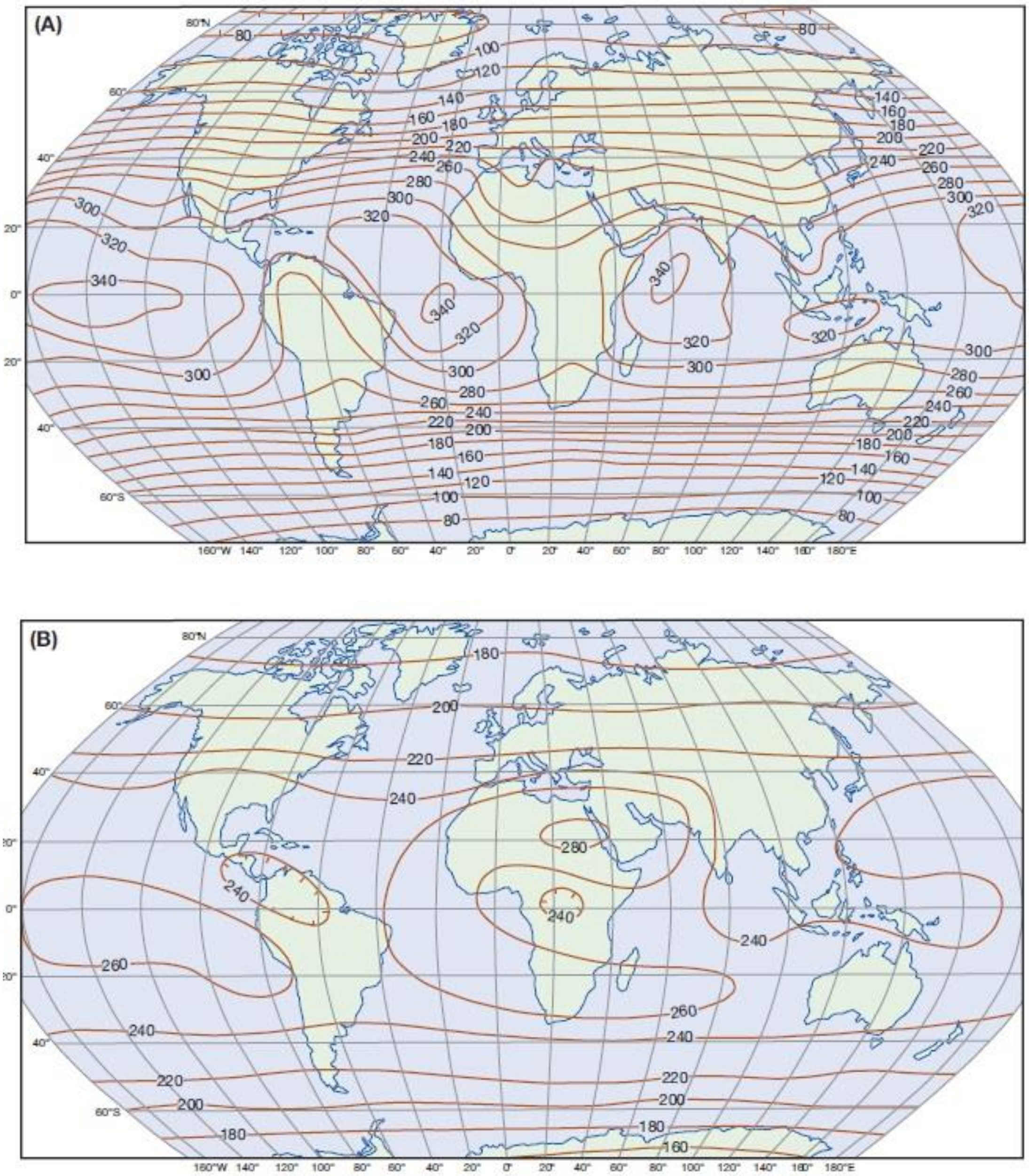


Figure 2.2 *Planetary short and longwave radiation (W m–2): (A) Mean annual absorbed shortwave radiation for the period April 1979 to March 1987; B Mean annual net planetary longwave radiation (Ln) on a horizontal surface at the top of the atmosphere.*

2.2 Energy flow representation

The exchanges and flows associated with energy inputs into the Earth-atmosphere system is represented by a series of symbolic equations. Use of the equations permits easy calculation once values are input. Shortwave solar radiation (K↓) reaching the surface is made up of the vertical radiation (S) and diffuse radiation (D):

$$K\downarrow = S + D$$

Some of the energy is reflected back to space ($K\uparrow$) so that net shortwave radiation (K^*) is the difference between the two:

$$K^* = K\downarrow - K\uparrow$$

Net longwave, terrestrial radiation (L^*) comprises downward atmospheric radiation ($L\downarrow$) less upward terrestrial radiation ($L\uparrow$):

$$L^* = L\downarrow - L\uparrow$$

The amount of energy available at any surface is thus the sum of K^* and L^*. This is net all-wave radiation (Q^*):

$$Q^* = K^* + L^*$$

which may also be given as:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow)$$

Q^* may be positive or negative.

High positive values will occur during high sun periods when $K\downarrow$ is at its maximum and atmospheric radiation, $L\downarrow$, exceeds outgoing radiation, $L\uparrow$. Negative values require outgoing values to be greater than incoming. This happens, for example, on clear nights when $L\uparrow$is larger than other values. On a long-term basis, Q^* will vary with latitude and surface type.

2.2.1 The heat budget

Consider a column of the Earth's surface extending down to where vertical heat exchange no longer occurs (Figure 2.3). The net rate (G) at which heat in this column changes depends upon the following:

Net radiation $(K\uparrow - K\downarrow) + (L\uparrow - L\downarrow)$

Latent heat transfer (LE)

Sensible heat transfer (H)

Horizontal heat transfer (S)

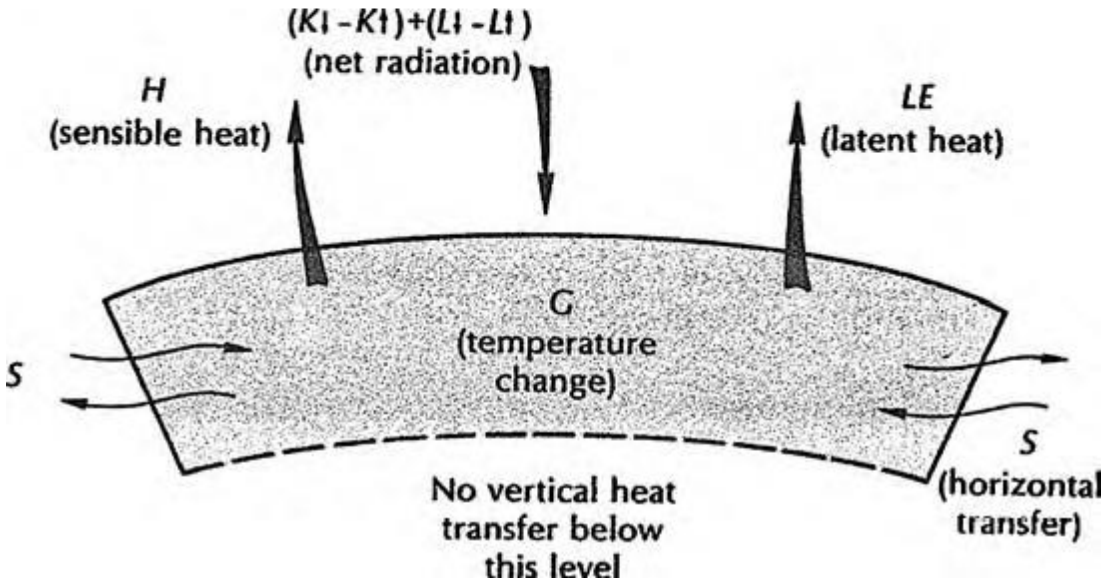


Figure (2.3): *Model of energy transfer in the atmospheric system.*

In symbolic form:

$$G = (K\uparrow - K\downarrow) + (L\uparrow - L\downarrow) - LE - H \pm S$$

Since

$$(K\uparrow - K\downarrow) + (L\uparrow - L\downarrow) = Q^*$$

then

$$G = Q^* - LE - H \pm S$$

in terms of Q*

$$Q^* = G + LE + H \pm S$$

that is, it is neither gaining nor losing heat over that time, so G=0 and can be dropped from the equation.

$$Q^* = LE + H \pm S$$

This equation will apply to a mobile column, such as the oceans. On land, where subsurface flow of heat is negligible, S will be unimportant. The land heat budget becomes

$$Q^* = LE + H$$

The ratio between LE and H is given as the Bowen Ratio.

2.3 Spatial pattern of the heat budget components

The mean latitudinal values of the heat budget components discussed above conceal great spatial variations. Figure 2.4 shows the global distribution of the annual net radiation at the surface. Broadly, its magnitude decreases poleward from about 25° latitude. However, as a result of the high absorption of solar radiation by the sea, net radiation is greater over the oceans – exceeding 160W m^{-2} in latitudes 15–20° – than over land areas, where it is about 80–105W m^{-2} in the same latitudes. Net radiation is also lower in arid continental areas than in humid areas, because in spite of the increased insolation

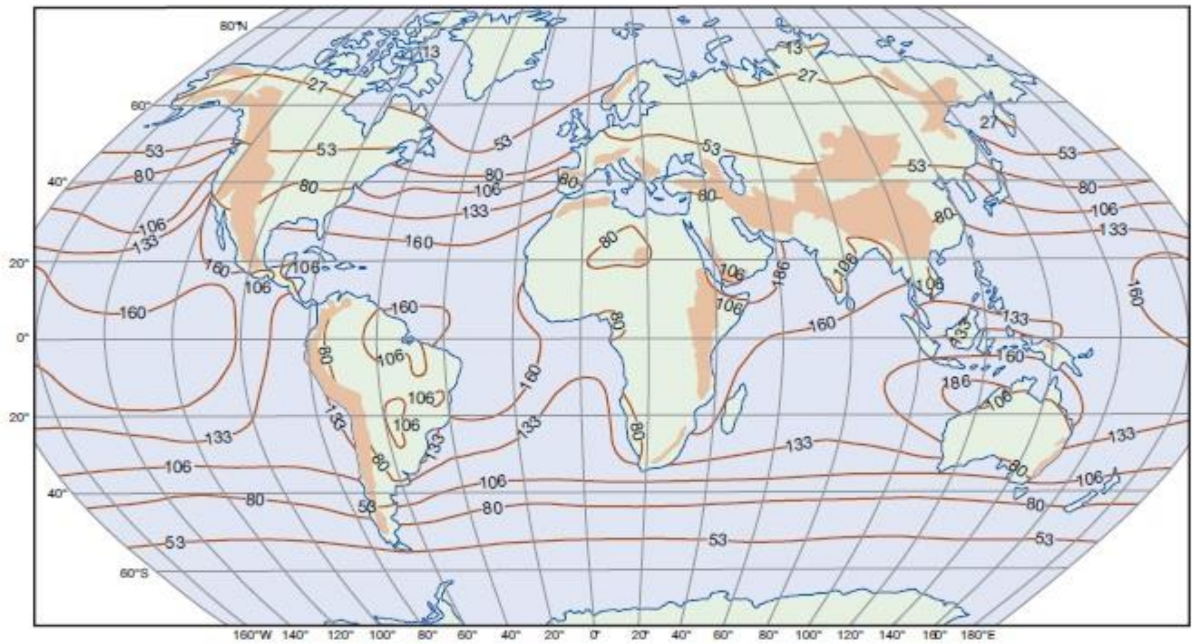


Figure 2.4*: Global distribution of the annual net radiation at the surface, in W m^{-2}.*

receipts under clear skies there is at the same time greater net loss of terrestrial radiation. Figures 2.5 and 2.6 show the annual vertical transfers of latent and sensible heat to the atmosphere. Both fluxes are distributed very differently over land and seas. Heat expenditure for evaporation is at a maximum in tropical and subtropical ocean areas, where it exceeds 160W m^{-2}. It is less near the equator, where wind speeds are somewhat lower and the air has a vapor pressure close to the saturation value . It is clear from Figure 2.5 that the major warm currents greatly increase the evaporation rate. On land, the latent heat transfer is largest in hot, humid regions. It is least in arid areas with low precipitation and in high latitudes, where there is little available energy or moisture. The largest exchange of sensible heat occurs over tropical deserts, where more than 80W m^{-2} is transferred to the atmosphere (see Figure 2.6). In contrast to latent heat, the sensible heat flux is generally small over the oceans, only reaching 25–40W m^{-2} in areas of warm currents. Indeed, negative values occur (transfer *to* the ocean) where warm continental air masses move offshore over cold currents.

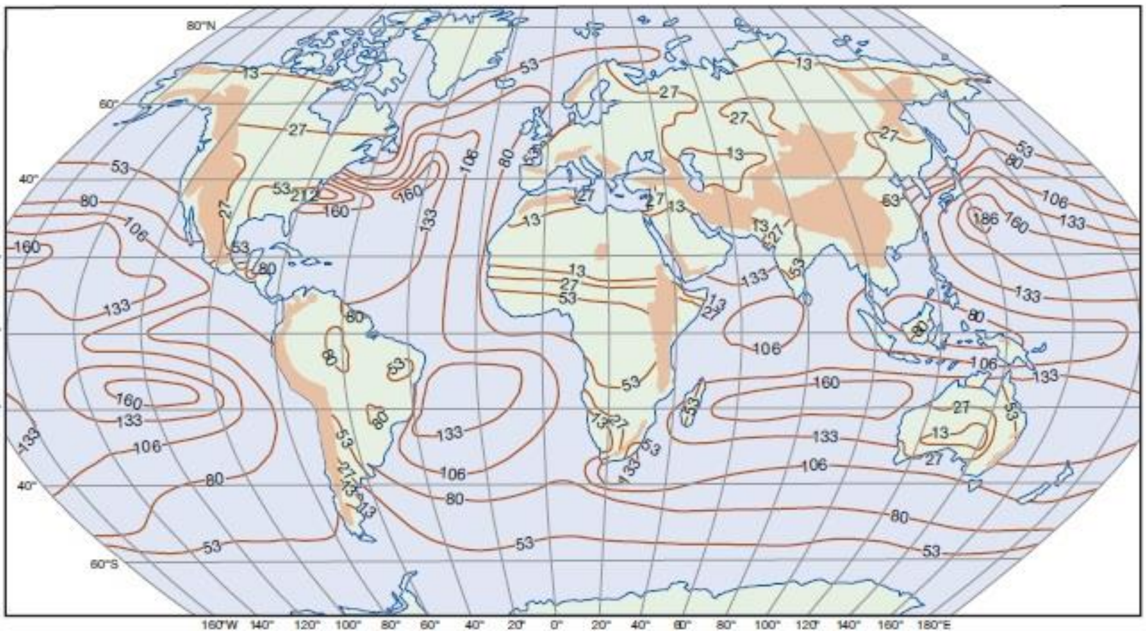


Figure 2.5: *Global distribution of the vertical transfer of latent heat, in W m^{-2}.*

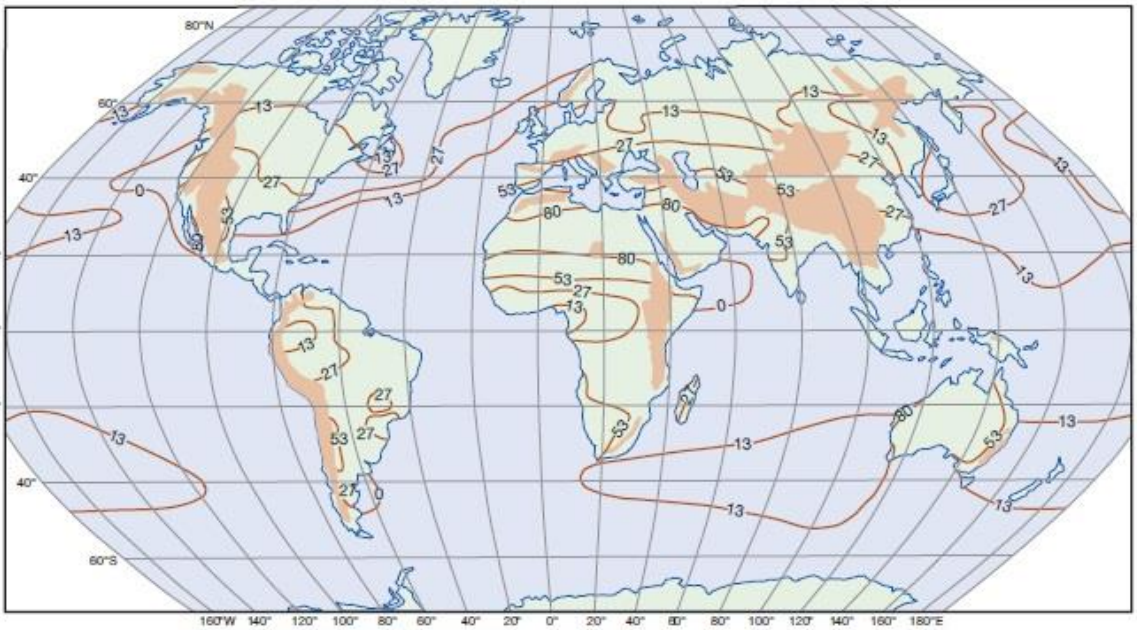


Figure 2.6*: Global distribution of the vertical transfer of sensible heat, in W m^{-2}.*

2.4 The greenhouse effect

The natural greenhouse effect of the earth's atmosphere is attributable primarily to water vapor. It accounts for 21K of the 33K difference between the effective temperature of a dry atmosphere and the real atmosphere through the trapping of infrared radiation. Water vapor is strongly absorptive around 2.4–3.1μm, 4.5–6.5μm and above 16μm. The concept of greenhouse gas-induced warming is commonly applied to the effects of the increases in atmospheric carbon dioxide concentrations resulting from anthropogenic activities, principally the burning of fossil fuels. Sverre Arrhenius in Sweden drew attention to this possibility in 1896, but observational evidence was only forthcoming some 40 years later (Callendar, 1938, 1961). However, a careful record of atmospheric concentrations of carbon dioxide was lacking until Charles Keeling installed calibrated instruments at the Mauna Loa Observatory, Hawaii, in 1957. Within a decade, these observations became the global benchmark.

They showed an annual cycle of about 5ppm at the Observatory, caused by the biospheric uptake and release, and the ca. 0.4 percent annual increase in CO_2, from 315ppm in 1957 to 383ppm in 2007, due to fossil fuel burning. The annual increase is about half of the total emission due to CO_2 uptake by the oceans and the land biosphere. The principal absorption band for radiation by carbon dioxide is around 14–16μm, but there are others at 2.6 and 4.2μm. Most of the effect of increasing CO_2 concentration is by enhanced absorption in the latter, as the main band is almost saturated. The sensitivity of mean global air temperature to a doubling of CO_2 is in the range 2–5°C, while a removal of all atmospheric CO_2 might lower the mean surface temperature by more than 10°C.

The important role of other trace greenhouse gases (methane, nitrous oxide, fluorocarbons) was recognized in the 1980s and many additional trace gases began to be monitored. The latest is nitrogen trifluoride used during the manufacture of liquid crystal flat-panel displays, thin-film solar cells and microcircuits. Although concentrations of the gas are currently only 0.454 parts per trillion, it is 17,000 times more potent as a global warming agent than a similar mass of carbon dioxide. The past histories of greenhouse gases, reconstructed from ice core records, show that the preindustrial level of CO_2 was 280ppm and methane 750ppb compared with 383ppm and 1790ppb, respectively, today. Their concentrations decreased to about 180 ppm and 350ppb, respectively, during the maximum phases of continental glaciation in the Pleistocene Ice Age. The positive feedback effect of CO_2, which involves greenhouse gas-induced warming leading to an enhanced hydrological cycle with a larger atmospheric vapor content and therefore further warming, is still not well resolved quantitatively.

2.5 Greenhouse Gases

Although the Earth's atmosphere consists mainly of oxygen and nitrogen, neither plays a significant role in enhancing the greenhouse effect because both are essentially transparent to terrestrial radiation. The greenhouse effect is primarily a function of the concentration of water vapor, carbon dioxide, and other trace gases in the atmosphere that absorb the terrestrial radiation leaving the surface of the Earth. Changes in the atmospheric concentrations of these greenhouse gases can alter the balance of energy transfers between the atmosphere, space, land, and the oceans. A gauge of these changes is called radiative forcing, which is a simple measure of changes in the energy available to the Earth-atmosphere system. Holding everything else constant, increases in greenhouse gas concentrations in the atmosphere will produce positive radiative forcing (i.e., a net increase in the absorption of energy by the Earth).

Naturally occurring greenhouse gases include water vapor, carbon dioxide (CO2), methane (CH_4), nitrous oxide (N_2O), and ozone (O_3). Several classes of halogenated substances that contain fluorine, chlorine, or bromine are also greenhouse gases, but they are, for the most part, solely a product of industrial activities. Chlorofluorocarbons (CFCs) and hydrochlorofluorocarbons (HCFCs) are halocarbons that contain chlorine, while halocarbons that contain bromine are referred to as bromofluorocarbons (i.e., halons). Because CFCs, HCFCs, and halons are stratospheric ozone depleting substances, they are covered under the Montreal Protocol on Substances that Deplete the Ozone Layer.

Carbon dioxide, methane, and nitrous oxide are continuously emitted to and removed from the atmosphere by natural processes on Earth. Anthropogenic activities, however, can cause additional quantities of these and other greenhouse gases to be emitted or sequestered, thereby changing their global average atmospheric concentrations. Natural activities such as respiration by plants or animals and seasonal cycles of plant growth and decay are examples of processes that only cycle carbon or nitrogen between the atmosphere and organic biomass. Such processes – except when directly or indirectly perturbed out of equilibrium by anthropogenic activities – generally do not alter average atmospheric greenhouse gas concentrations over decadal timeframes. Climatic changes resulting from anthropogenic activities, however, could have positive or negative feedback effects on these natural systems. Atmospheric concentrations of these gases, along with their rates of growth and atmospheric lifetimes, are presented in Table 2.1

Table 2.1: *Global atmospheric concentration (ppm unless otherwise specified), rate of concentration change (ppb/year) and atmospheric lifetime (years) of selected greenhouse gases.*

	CO_2	CH_4	N_2O	SF_6(ppt)	CF_4(ppty)
Pre-industrial concentration	278	0.700	0.270	0	40
Atmospheric concentration 1998	365	1.745	0.314	4.3	80
Rate of concentration change (90-99)	1.5	0.007	0.0008	0.24	1.0
Atmospheric lifetime	50-200	12	114	3,200	>50,000

2.5.1 Water Vapor (H_2O)

Overall, the most abundant and dominant greenhouse gas in the atmosphere is water vapor. Water vapor is neither long-lived nor well mixed in the atmosphere, varying spatially from 0 to 2 percent. In addition, atmospheric water can exist in several physical states including gaseous, liquid, and solid. Human activities are not believed to directly affect the average global concentration of water vapor; however, the radiative forcing produced by the increased concentrations of other greenhouse gases may indirectly affect the hydrologic cycle. A warmer atmosphere has an increased water holding capacity; yet, increased concentrations of water vapor affects the formation of clouds, which can both absorb and reflect solar and terrestrial radiation.

2.5.2 Carbon Dioxide (CO2)

. In nature, carbon is cycled between various atmospheric, oceanic, land biotic, marine biotic, and mineral reservoirs. The largest fluxes occur between the atmosphere and terrestrial biota, and between the atmosphere and surface water of the oceans. In the atmosphere, carbon predominantly exists in its oxidized form as CO_2. Atmospheric carbon dioxide is part of this global carbon cycle, and therefore its fate is a complex function of geochemical and biological processes.

Carbon dioxide concentrations in the atmosphere increased from approximately 280 parts per million by volume (ppmv) in pre-industrial times to 367 ppmv in 1999, a 31percent increase. The IPCC notes that "[t]his concentration has not been exceeded during the past 420,000 years, and likely not during the past 20 million years. The rate of increase over the past century is unprecedented, at least during the past 20,000 years." The IPCC definitively states that "the present atmospheric CO_2 increase is caused by anthropogenic emissions of CO_2". Forest clearing, other biomass burning, and some non-energy production processes (e.g., cement production) also emit notable quantities of carbon dioxide. In its second assessment, the IPCC also stated that "[t]he increased amount of carbon dioxide [in the atmosphere] is leading to climate change and will produce, on average, a global warming of the Earth's surface because of its enhanced greenhouse effect – although the magnitude and significance of the effects are not fully resolved".

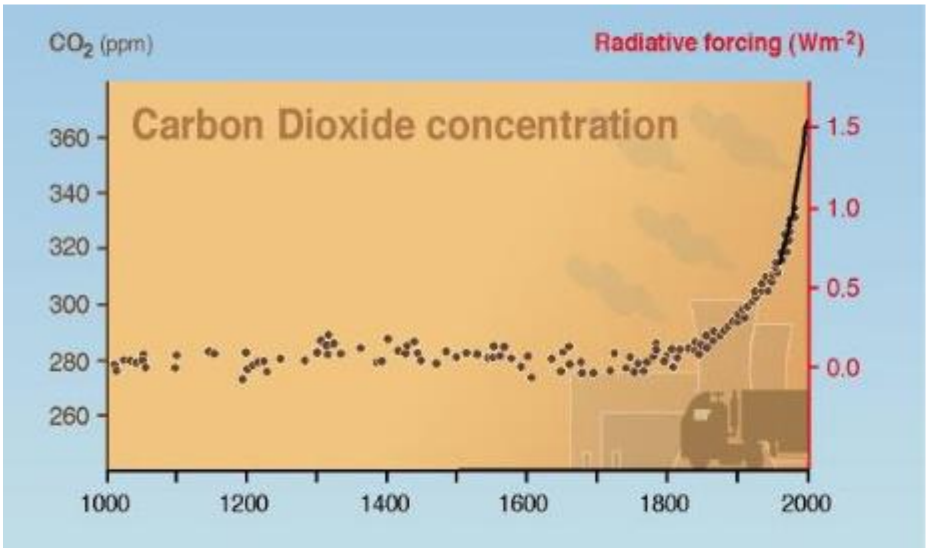


Figure 2.7: *Concentration change and radiative forcing of CO2 in the last thousand years.*

2.5.3 Methane (CH4).

Methane is primarily produced through anaerobic decomposition of organic matter in biological systems. Agricultural processes such as wetland rice cultivation, enteric fermentation in animals, and the decomposition of animal wastes emit CH4,as does the decomposition of municipal solid wastes. Methane is also emitted during the production and distribution of natural gas and petroleum, and is released as a by-product of coal mining and incomplete fossil fuel combustion. Atmospheric concentrations of methane have increased by about 150 percent since pre-industrial times, although the rate of increase has been declining. The IPCC has estimated that slightly more than half of the current CH 4 flux to the atmosphere is anthropogenic, from human activities such as agriculture, fossil fuel use and waste disposal. Methane is removed from the atmosphere by reacting with the hydroxyl radical (OH) and is ultimately converted to CO2. Minor removal processes also include reaction with Cl in the marine boundary layer, a soil sink, and stratospheric reactions. Increasing emissions of methane reduce the concentration of OH, a feedback which may increase methane's atmospheric lifetime.

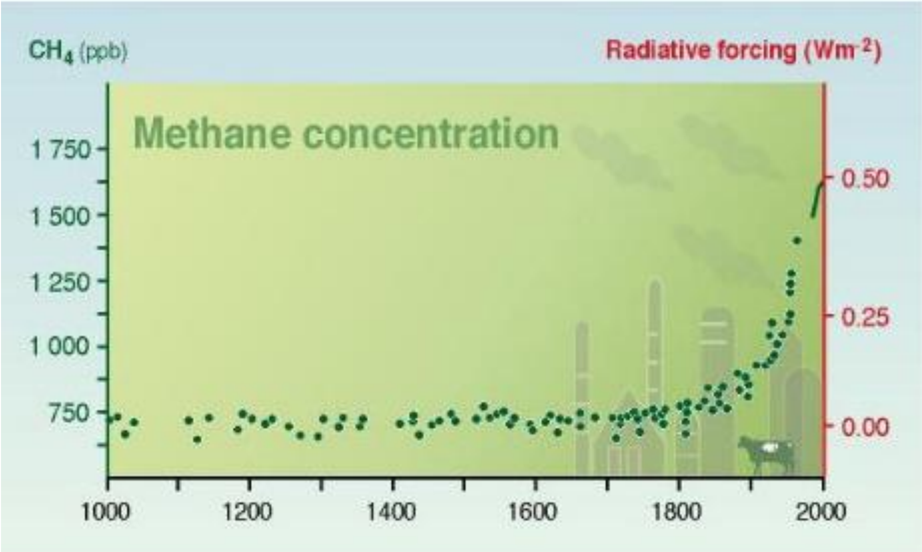


Figure 2.8: *Concentration change and radiative forcing of CH_4 in the last thousand years.*

2.5.4 Nitrous Oxide (N_2O)

Anthropogenic sources of N2O emissions include agricultural soils, especially the use of synthetic and manure fertilizers; fossil fuel combustion, especially from mobile combustion; adipic (nylon) and nitric acid production; wastewater treatment and waste combustion; and biomass burning. The atmospheric concentration of nitrous oxide (N_2O) has increased by 16 percent since 1750, from a pre industrial value of about 270 ppb to 314 ppb in 1998, a concentration that has not been exceeded during the last thousand years. Nitrous oxide is primarily removed from the atmosphere by the photolytic action of sunlight in the stratosphere.

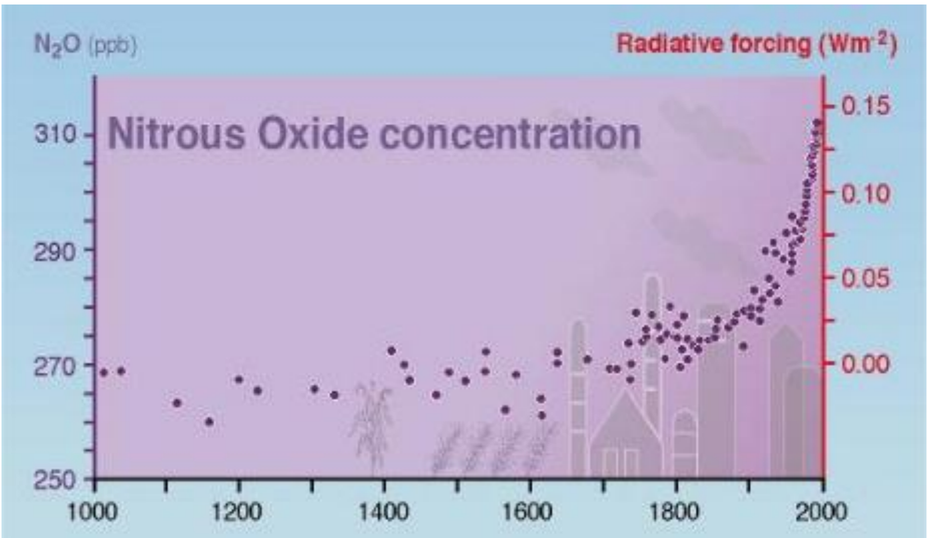


Figure 2.9: *Concentration change and radiative forcing of N_2O in the last thousand years.*

2.5.5 Ozone (O_3)

Ozone is present in both the upper stratosphere, where it shields the Earth from harmful levels of ultraviolet radiation, and at lower concentrations in the troposphere, where it is the main component of anthropogenic photochemical "smog." During the last two decades, emissions of anthropogenic chlorine and bromine-containing halocarbons, such as chlorofluorocarbons (CFCs), have depleted stratospheric ozone concentrations. This loss of ozone in the stratosphere has resulted in negative radiative forcing, representing an indirect effect of anthropogenic emissions of chlorine and bromine compounds. The depletion of stratospheric ozone and its radiative forcing was expected to reach a maximum in about 2000 before starting to recover, with detection of such recovery not expected to occur much before 2010. The past increase in tropospheric ozone, which is also a greenhouse gas, is estimated to provide the third largest increase in direct radiative forcing since the pre-industrial era, behind CO_2 and CH4. Tropospheric ozone is produced from complex chemical reactions of volatile organic compounds mixing with nitrogen oxides (NO_x) in the presence of sunlight. Ozone, carbon monoxide (CO), sulfur dioxide (SO_2), nitrogen dioxide (NO_2) and particulate matter are included in the category referred to as "criteria pollutants" in the United States under the Clean Air Act and its subsequent amendments. The tropospheric concentrations of ozone and these other pollutants are short-lived and, therefore, spatially variable

2.5.6 Halocarbons, Perfluorocarbons, and Sulfur Hexafluoride (SF6)

Halocarbons are, for the most part, man-made chemicals that have both direct and indirect radiative forcing effects. Halocarbons that contain chlorine – chlorofluorocarbons (CFCs), hydro-chlorofluoro-carbons (HCFCs), methyl chloroform, and carbon tetrachloride – and bromine – halons, methyl bromide, and hydrobromofluorocarbons (HBFCs) – result in stratospheric ozone depletion and are therefore controlled under the Montreal Protocol on Substances that Deplete the Ozone Layer. Although CFCs and HCFCs include potent global warming gases, their net radiative forcing effect on the atmosphere is reduced because they cause stratospheric ozone depletion, which is itself an important greenhouse gas in addition to shielding the Earth from harmful levels of ultraviolet radiation. Under the Montreal Protocol, the United States phased out the production and importation of halons by 1994 and of CFCs by 1996. PFCs and SF_6 are predominantly emitted from various industrial processes including aluminum smelting, semiconductor manufacturing, electric power transmission and distribution, and magnesium casting. Currently, the radiative forcing impact of PFCs and SF_6 is also small; however, they have a significant growth rate, extremely long atmospheric lifetimes, and are strong absorbers of infrared radiation, and therefore have the potential to influence climate far into the future.

2.5.7 Carbon Monoxide (CO)

Carbon monoxide has an indirect radiative forcing effect by elevating concentrations of CH4 and tropospheric ozone through chemical reactions with other atmospheric constituents (e.g., the hydroxyl radical, OH) that would otherwise assist in destroying CH4 and tropospheric ozone. Carbon monoxide is created when carbon-containing fuels are burned incompletely. Through natural processes in the atmosphere, it is eventually oxidized to CO2 . Carbon monoxide concentrations are both short-lived in the atmosphere and spatially variable.

2.5.8 Nitrogen Oxides (NOx).

The primary climate change effects of nitrogen oxides (i.e., NO and NO_2) are indirect and result from their role in promoting the formation of ozone in the troposphere and, to a lesser degree, lower stratosphere, where it has positive radiative forcing effects. Additionally, NOx emissions from aircraft are also likely to decrease methane concentrations, thus having a negative radiative forcing effect. Nitrogen oxides are created from lightning, soil microbial activity, biomass burning –both natural and anthropogenic fires – fuel combustion, and, in the stratosphere, from the photo-degradation of nitrous oxide (N2O). concentrations of NO_x are both relatively short-lived in the atmosphere and spatially variable.

2.5.8 Non-methane Volatile Organic Compounds (NMVOCs)

Non-methane volatile organic compounds include compounds such as propane, butane, and ethane. These compounds participate, along with NOx , in the formation of tropospheric ozone and other photochemical oxidants. NMVOCs are emitted primarily from transportation and industrial processes, as well as biomass burning and non-industrial consumption of organic solvents. Concentrations of NMVOCs tend to be both short-lived in the atmosphere and spatially variable.

Chapter Three

Climate classification

Introduction

The purpose of any classification system is to obtain an efficient arrangement of information in a simplified and generalized form. Climate statistics can be organized in order to describe and delimit the major types of climate in quantitative terms. Obviously, any single classification can serve only a few purposes satisfactorily and many different schemes have therefore been developed. Many climatic classifications are concerned with the relationships between climate and vegetation or soils and rather few attempt to address the direct effects of climate on humans.

Only the basic principles of generic classifications related to plant growth or vegetation are summarized here.

3.1 Generic classification related to plant growth or vegetation

Numerous schemes have been suggested for relating climate limits to plant growth or vegetation groups. They rely on two basic criteria – the degree of aridity and of warmth. Aridity is not simply a matter of low precipitation, but of the 'effective precipitation' (i.e., precipitation minus evaporation). The ratio of rainfall/temperature is often used as an index of precipitation effectiveness, since higher temperatures increase evaporation. W. Köppen developed the pre-eminent example of such a classification. Between 1900 and 1936, he devised several classification schemes that involve considerable complexity in their full detail. The system has been used extensively in geographical teaching. The key features of Köppen's approach are temperature criteria and aridity criteria.

3.1.1Temperature criteria

Five of the six major climate types are based on monthly mean temperature values.

1 Tropical rainy climate: coldest month >18°C.

2 Dry climates.

3 Warm temperate rainy climates: coldest month between –3° and +18°C, warmest month >10°C.

4 Cold boreal forest climates: coldest month <–3°, warmest month >10°C. Note that many American workers use a modified version with 0°C as the C/D boundary.

5 Tundra climate: warmest month 0–10°C.

6 Perpetual frost climate: warmest month <0°C.

3.1.2 Aridity criteria

The criteria imply that, with winter precipitation, arid (desert) conditions occur where $r/T < 1$, semi-arid conditions where $1 < r/T < 2$. If the rain falls in summer, a larger amount is required to offset evaporation and maintain an equivalent effective precipitation.

Subdivisions of each major category are made with reference, first, to the seasonal distribution of precipitation. The most common of these are:

f = no dry season; *m*= monsoonal, with a short, dry season and heavy rains during the rest of the year; *s* = summer dry season; *w* = winter dry season. Second, there are further temperature criteria based on seasonality. Twenty-seven subtypes are recognized, of which 23 occur in Asia. The ten major Köppen types each have distinct annual energy budget regimes, as illustrated in Figure 3.1.

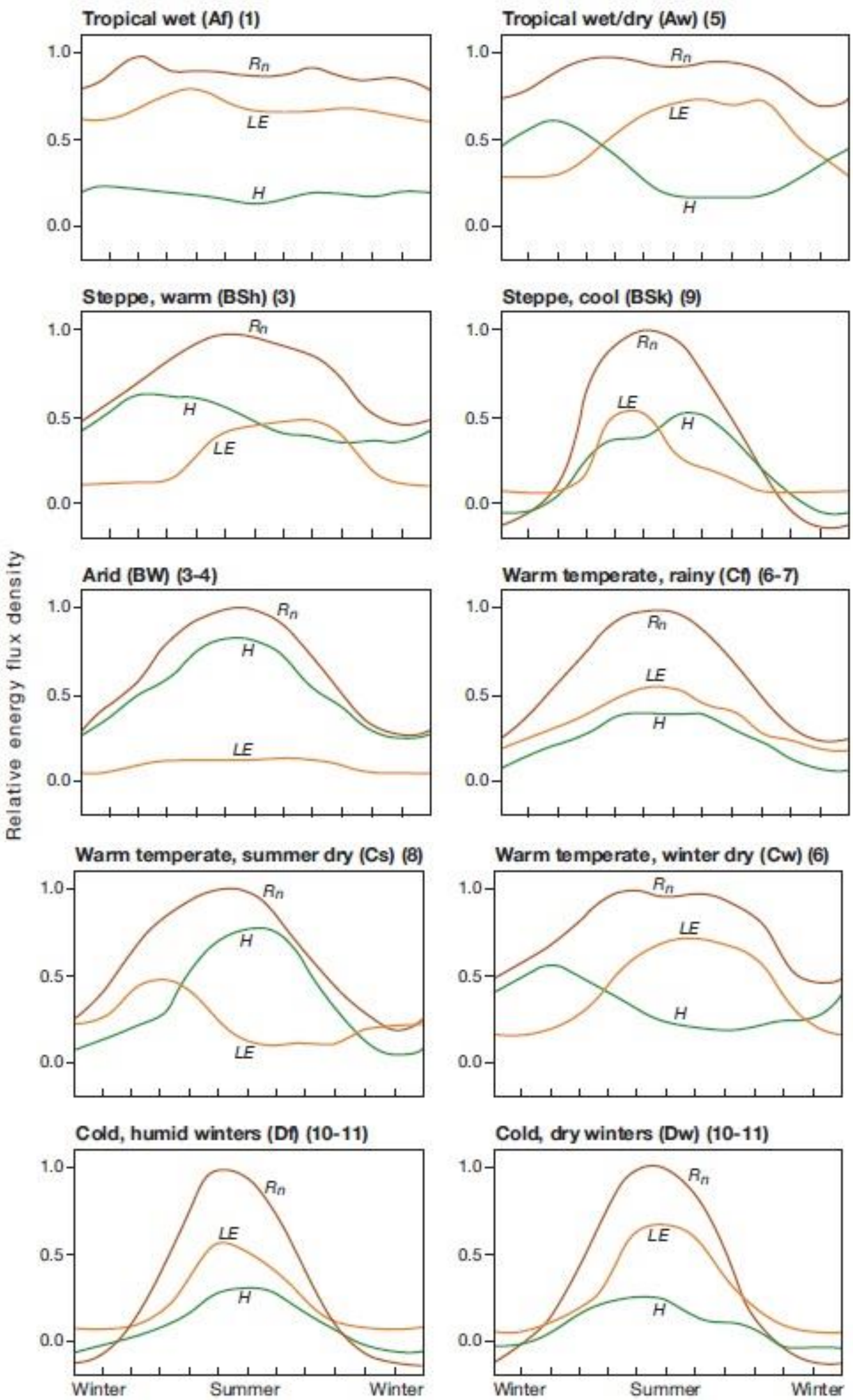
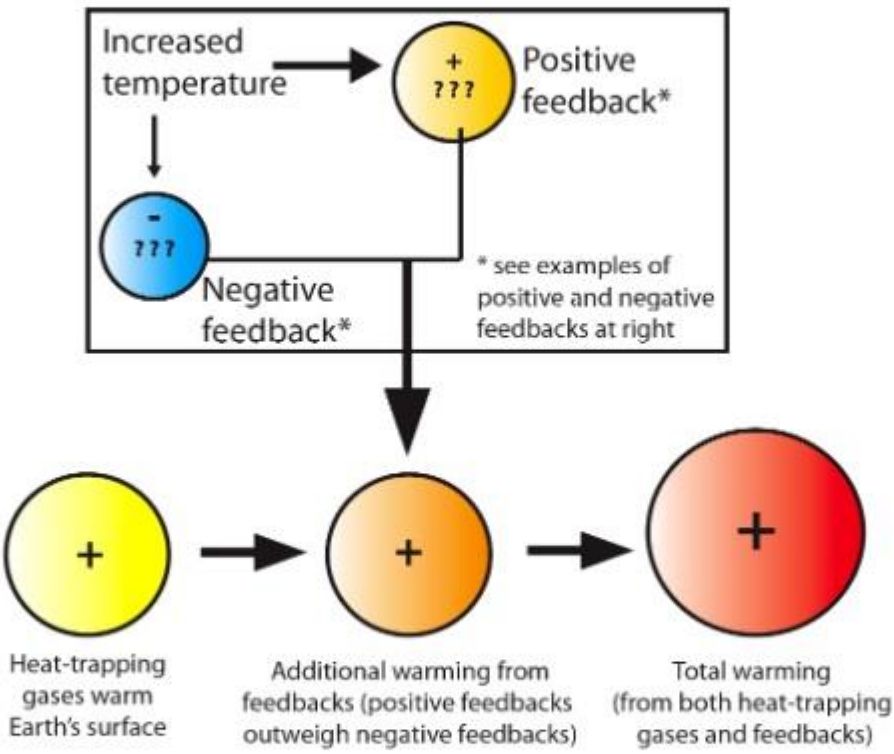


Figure 3.1 *Characteristic annual energy balances for ten different climatic types (Köppen symbols and Strahler classification numbers shown).*

3.2 Feedback in climate system

As human activities increase the concentration of heat-trapping gases in the atmosphere, Earth's surface warms. This initial warming causes many changes in the atmosphere, on land, and at sea. These changes, in turn, may cause additional warming (positive feedbacks) or reduce the rate of warming (negative feedbacks). The actual rate of warming of Earth's surface is determined by the warming caused by heat-trapping gases, plus the effects of the feedbacks to this warming.

Positive feedbacks to warming are expected to outweigh negative feedbacks, leading to additional warming. Scientists estimate that feedbacks may increase warming by 15 – 78% over this century.



3.2.1 Positive feedback

Warming leads to positive feedbacks by:

Reducing ice and snow cover , exposing soil or ocean Water . Ice reflects much more solar radiation than soil or water . Loss of ice and snow means Earth's surface absorbs more energy , increasing global warming. See below:

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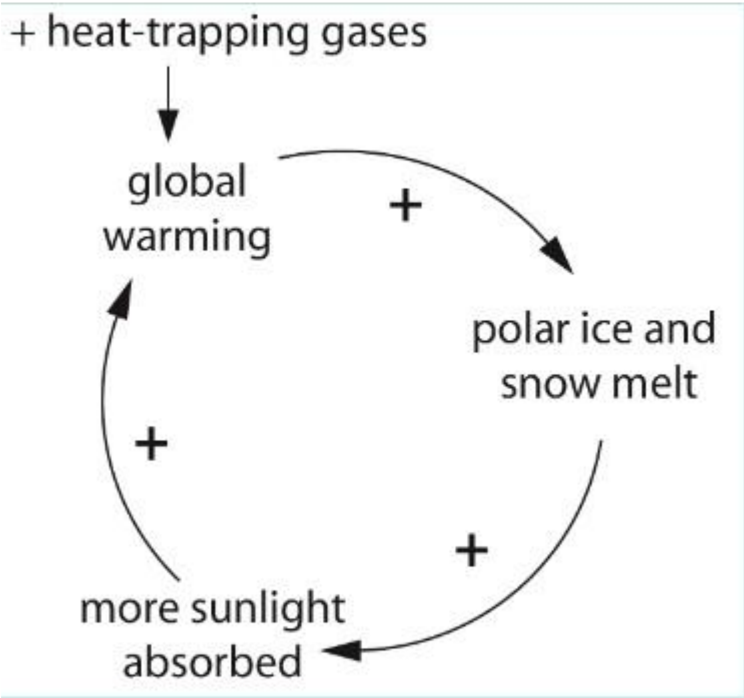


Increasing soil respiration rates Causing permafrost to melt . permafrost, the frozen soil commonly found in Earth's coldest regions, could release large amounts of CO2 and methane if it thaws.

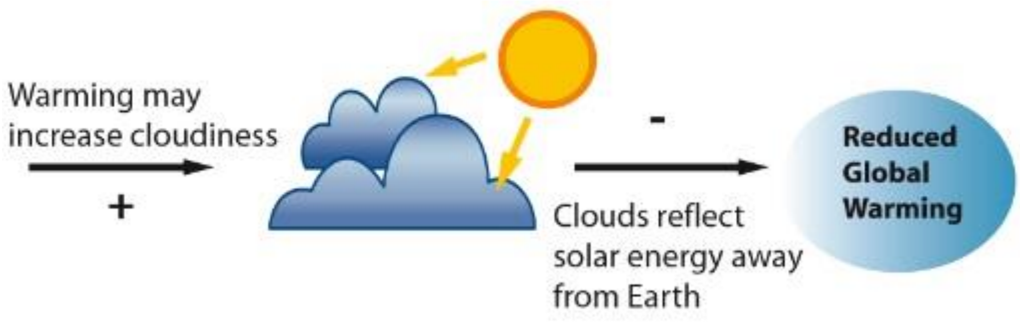
Increasing water vapor concentrations. Warmer weather leads to more evaporation. Water vapor is a powerful greenhouse gas.

Positive feedbacks act to destabilize the climate system.

3.2.2 Negative feedback

Warming leads to negative feedbacks by:

Increasing cloud cover. Warming leads to more evaporation, which can increase cloudiness. Clouds reflect solar radiation and decrease warming. However it is uncertain if cloud production will increase or decrease with warming. Overall, scientists believe that clouds will act as a negative feedback.



Increasing plant growth. As warming and increasing CO2 stimulate plant growth, carbon storage on land may increase.

Negative feedbacks act to stabilize the climate system.

Examples of possible positive and negative feedback in physical systems

1. As carbon dioxide levels in the atmosphere rise:

• Temperature of Earth rises

As Earth warms:

• the rate of photosynthesis in plants increases

• more carbon dioxide is therefore removed from the atmosphere by plants,

reducing the greenhouse effect and reducing global temperatures

2. As Earth warms:

• Ice cover melts, exposing soil or water

• Albedo decreases

• More energy is absorbed by Earth’s surface

• Global temperature rises

• More ice melts

3. As Earth warms, upper layers of permafrost melt, producing waterlogged soil above frozen ground:

• Methane gas is released in anoxic environment

• Greenhouse effect is enhanced

• Earth warms, melting more permafrost

4. As Earth warms, increased evaporation:

• Produces more clouds

• Clouds increase albedo, reflecting more light away from Earth

• Temperature falls

• Rates of evaporation fall

5. As Earth warms, organic matter in soil is decomposed faster:

• More carbon dioxide is released

• Enhanced greenhouse effect occurs

• Earth warms further

• Rates of decomposition increase

6. As Earth warms, evaporation increases:

• Snowfall at high latitudes increases

• Icecaps enlarge

• More energy is reflected by increased albedo of ice cover

• Earth cools

• Rates of evaporation fall

3.3 Climate and the General Circulation

3.3.1 General Circulation of the Atmosphere

Before we study the general circulation, we must remember that it only represents the average air flow around the world. Actual winds at any one place and at any given time may vary considerably from this average. Nevertheless, the average can answer why and how the winds blow around the world the way they do—why. The average can also give a picture of the driving mechanism behind these winds, as well as a model of how heat and momentum are transported from equatorial regions poleward, keeping the climate in middle latitudes tolerable.

The underlying cause of the general circulation is the unequal heating of the earth's surface. We have learned that averaged over the entire earth, incoming solar radiation is roughly equal to outgoing earth radiation. However, we also know that this energy balance is not maintained for each latitude, since the tropics experience a net gain in energy, while polar regions suffer a net loss. To balance these inequities, the atmosphere transports warm air poleward and cool air equatorward. Although seemingly simple, the actual flow of air is complex; certainly not everything is known about it. In order to better understand it, we will first look at some models (that is, artificially constructed simulations) that eliminate some of the complexities of the general circulation.

3.1.2 Single-cell Model

The first model is the single-cell model, in which we assume that the earth's surface is uniformly covered with water, so that differential heating between land and water does not come into play. We will further assume that the sun is always directly over the equator, so that the winds will not shift seasonally. Finally, we assume that the earth does not rotate, so that the only force we need deal with is the pressure gradient force. With these assumptions, the general circulation of the atmosphere would look much like Figure 3.2, a huge thermally driven convection cell in each hemisphere.

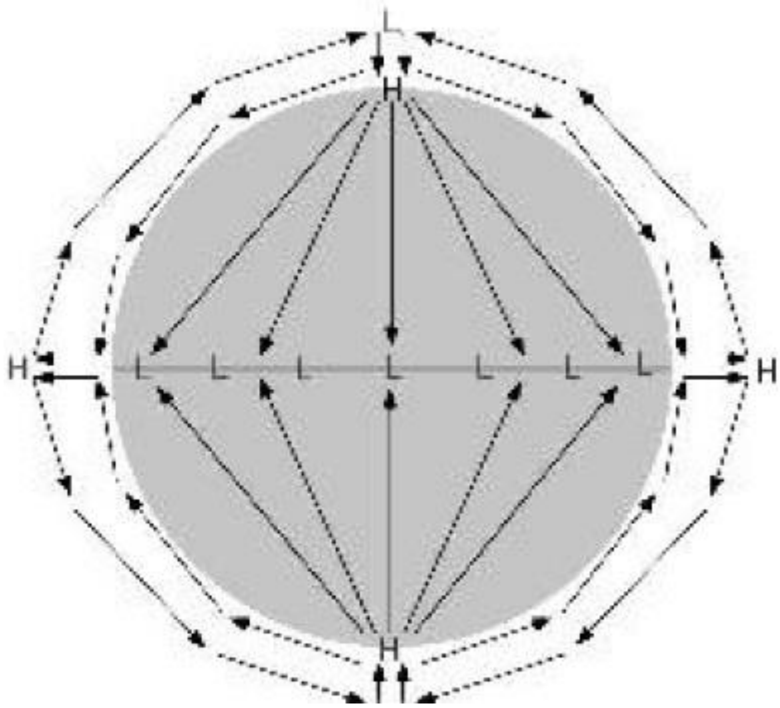


Figure 3.2: *Simplified one-cell global air circulation patterns.*

This is the **Hadley cell** (named after the eighteenth-century English meteorologist George Hadley, who first proposed the idea). It is driven by energy from the sun Excessive heating of the equatorial area produces a broad region of surface low pressure, while at the poles excessive cooling creates a region of surface high pressure. In response to the horizontal pressure gradient, cold surface polar air flows equatorward, while at higher levels air flows toward the poles. The entire circulation consists of a closed loop with rising air near the equator, sinking air over the poles, an equatorward flow of air near the surface, and a return flow aloft. In this manner, some of the excess energy of the tropics is transported as sensible and latent heat to the regions of energy deficit at the poles.

Such a simple cellular circulation as this does not actually exist on the earth. For one thing, the earth rotates, so the Coriolis force would deflect the southward-moving surface air in the Northern Hemisphere to the right, producing easterly surface winds at practically all latitudes, These winds would be moving in a direction opposite to that of the earth's rotation and, due to friction with the surface, would slow down the earth's spin. We know that this does not happen and that prevailing winds in middle latitudes actually blow from the west. Therefore, observations alone tell us that a closed circulation of air between the equator and the poles is not the proper model for a rotating earth. (Models that simulate air flow around the globe have also verified this.) How, then, does the wind blow on a rotating planet? To answer, we will keep our model simple by retaining our first two assumptions—that is, that the earth is covered with water and that the sun is always directly above the equator.

3.1.3 Three-cell Model

If we allow the earth to spin, the simple convection system breaks into a series of cells as shown in Figure 3.3. Although this model is considerably more complex than the single-cell model, there are some similarities. The tropical regions still receive an excess of heat and the poles a deficit. In each hemisphere, three cells instead of one have the task of energy redistribution. A surface high-pressure area is located at the poles, and a broad trough of surface low pressure still exists at the closely resembles that of a Hadley cell. Let's look at this model more closely by examining what happens to the air' above the equator.

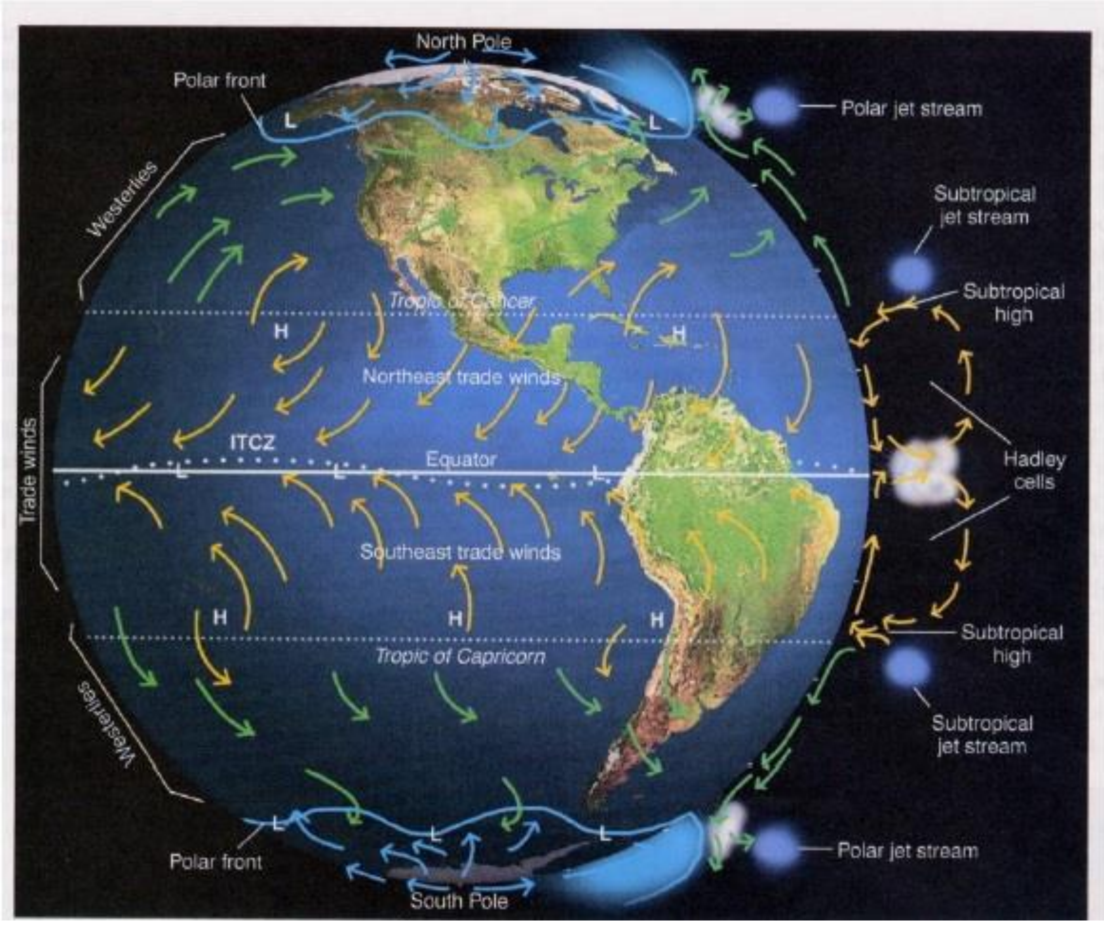


Figure 3.3: *General circulation schematic.*

Over equatorial waters, the air is warm, horizontal pressure gradients are weak, and winds are light. This region is referred to as the **doldrums**. Here, warm air rises, often condensing into huge cumulus clouds and thunderstorms called convective “hot” towers because of the enormous amount of latent heat they liberate. This heat makes the air more buoyant and provides energy to drive the Hadley cell. The rising air reaches the tropopause, which acts like a barrier, causing the air to move laterally toward the poles. The Coriolis force deflects this poleward flow toward the right in the Northern Hemisphere and to the left in the Southern Hemisphere, providing westerly winds aloft in both hemispheres. (We will see later that these westerly winds reach maximum velocity and produce jet streams near 30_ and 60_ latitudes.)

As air moves poleward from the tropics it constantly cools by radiation, and at the same time it also begins to converge, especially as it approaches the middle latitudes. This convergence (piling up) of air aloft increases the mass of air above the surface, which in turn causes the air pressure at the surface to increase. Hence, at latitudes near 30_, the convergence of air aloft produces belts of high pressure called **subtropical highs** (or anticyclones). As the converging, relatively dry air above the highs slowly descends, it warms by compression. This subsiding air produces generally clear skies and warm surface temperatures; hence, it is here that we find the major deserts of the world. Over the ocean, the weak pressure gradients in the center of the high produce only weak winds. According to legend, sailing ships traveling to the New World were frequently becalmed in this region, and, as food and supplies dwindled, horses were either thrown overboard or eaten. As a consequence, this region is sometimes called the **horse latitudes**.

From the horse latitudes, some of the surface air moves back toward the equator. It does not flow straight back, however, because the Coriolis force deflects the air, causing it to blow from the northeast in the Northern Hemisphere and from the southeast in the Southern Hemisphere. These steady winds provided sailing ships with an ocean route to the New World; hence, these winds are called the **trade winds**. Near the equator, the northeast trades converge with the southeast trades along a boundary called the **intertropical convergence zone** (ITCZ). In this region of surface convergence, air rises and continues its cellular journey.

Meanwhile, at latitude 30_, not all of the surface air moves equatorward. Some air moves toward the poles and deflects toward the east, resulting in a more or less westerly air flow—called the **prevailing westerlies,** or, simply **westerlies**—in both hemispheres Consequently, from Texas northward into Canada, it is much more common to experience winds blowing out of the west than from the east. The westerly flow is not constant; migrating areas of high and low pressure break up the surface flow pattern from time to time.

As this mild air travels poleward, it encounters cold air moving down from the poles. These two air masses of contrasting temperature do not readily mix. They are separated by a boundary called the **polar front**, a zone of low pressure—the **subpolar low**—where surface air converges and rises and storms develop. Some of the rising air returns at high levels to the horse latitudes, where it sinks back to the surface in the vicinity of the subtropical high. In this model, the middle cell (called the Ferrel cell, after the American meteorologist William Ferrel) is completed when surface air from the horse latitudes flows poleward toward the polar front. Behind the polar front, the cold air from the poles is deflected by the Coriolis force, so that the general flow of air is northeasterly. Hence, this is the region of the **polar easterlies**.

We can summarize all of this by referring back to Fig. 3.3 and 3.4 and noting that, at the surface) there are two major areas of high pressure and two major areas of low pressure. Areas of high pressure exist near latitude 30_ and the poles; areas of low pressure exist over the equator and near 60_ latitude in the vicinity of the polar front. By knowing the way the winds blow around these systems, we have a generalized picture of surface winds throughout the world. The trade winds extend from the subtropical high to the equator, the westerlies from the subtropical high to the polar front, and the polar easterlies from the poles to the polar front.

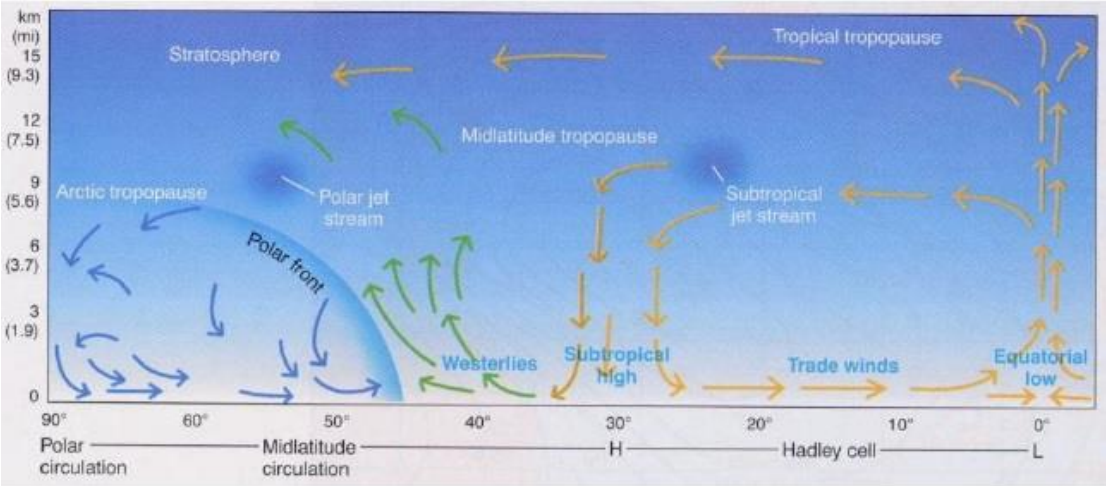


Figure 3.4: *Equator-to-pole cross section of the Northern Hemisphere. It shows the Hadley cell, subtropical highs, the subpolar low-pressure cell, and approximate locations of the subtropical and polar jet streams.*