5. The climate system

5.1. Weather, climate, and components of the climate system

The weather is characterized by the atmospheric conditions (e.g. temperature, precipitations, cloud cover, wind speed) at a particular place at a particular time. It applies to short-term changes in these conditions (less than a day to a few weeks). Climate is the average atmospheric conditions of a region and applies to longer-term changes (years and longer). Climate can also apply to larger spatial scales like continents or oceans. One can even refer to the climate of a particular planet (i.e. average atmospheric conditions characterizing this planet). Factors influencing the climate include for example the latitude, the presence or absence of mountains, and the distance from the sea (continentality).

The climate system consists of five interacting natural components (Fig. 53, see also section 1) and is powered by ***solar energy***. These natural components are the atmosphere, the hydrosphere, the cryosphere, the lithosphere, and the biosphere. To these natural components, a sixth component can be added: the anthroposphere (sum of all factors related to human activities). The 2007 report of the Intergovernmental Panel on Climate Change (IPCC) stated: “there is *very high confidence* that the global average net effect of human activities since 1750 has been one of warming” and “most of the observed increase in global average temperature since the mid-20th century is *very likely* due to the observed increase in anthropogenic greenhouse gas concentrations”.

Figure 53: Schematic representation of Earth’s climate system and its components: the atmosphere, the hydrosphere, the cryosphere, the lithosphere, the biosphere, and the anthroposphere. Source: adapted from Earth’s climate: past and future, Ruddiman (2001).

5.2. The atmosphere

The atmosphere is the layer of gases surrounding the Earth. Its density decreases with increasing distance from Earth’s surface. The densest atmospheric layer is the troposphere and has an average thickness of 11 km (Fig. 54). That is where clouds form and atmospheric circulation takes place. It is a very mobile and fast-changing environment —think about how fast cloud cover may change and weather may sometimes shift from sunny to rainy in less than an hour—. Above the troposphere lies the stratosphere up to an altitude of 50 km. The stratosphere is a more stable environment. However, large volcanic eruptions can eject small particles in the stratosphere where they may remain several years before gravity removes them. This volcanic dust can potentially lower the amount of sunlight reaching the Earth’s surface. The stratosphere is also where the ozone (O_3) blocks harmful ultraviolet radiation from the Sun. Above the stratosphere, the atmosphere thins out progressively until interplanetary space is reached.

Figure 54: Vertical structure of the atmosphere. The grey line represents the evolution of air temperature with altitude. Source: adapted from Understanding Earth 6th edition.

Atmospheric circulation is driven by the uneven heating of Earth's surface (Fig. 55). The amount of incoming sunlight reaching the Earth's surface per unit area varies with latitude. At higher latitudes, sunlight reaches the Earth's surface at a higher angle and the same amount of solar energy must warm a larger surface than at lower latitudes (Fig. 55A). The amount of light reflected by Earth's surface is referred to as its ***albedo***. The albedo depends on the nature of the substrate and on the angle of incidence of the sunlight. Snow and ice have the largest albedos (60-90% of light reflected) whereas water has the lowest (5-10%). Also, the greater the angle of incidence, the greater the albedo (Fig. 55B). Therefore, a larger proportion of the incoming sunlight is absorbed by Earth's surface at lower latitudes, whereas a larger proportion is reflected at higher latitudes. Cloud cover should also be taken into account as some of the incident sunlight is reflected by the clouds. All in all, the poles reflect more sunlight than the other regions of the globe (Fig. 55C). A maximum solar energy is absorbed at the equator and a minimum at the poles. Therefore, the poles are much colder than the equator.

Figure 55: Factors contributing to the unequal heating of the Earth's surface. (A) the same amount of solar radiation heats a larger area of Earth's surface at higher latitudes, (B) the relationship between the latitude, the angle of incidence of the sunlight, and the proportion of light absorbed/reflected, and (C) graph showing the unequal heating of Earth's surface and the heat loss at the poles vs. the maximum heat gain at the equator. Source: adapted from Earth's climate: past and future, Ruddiman (2001).

As a result, Polar Regions lose heat whereas low-latitudes areas gain heat. Since on average the poles do not get cooler and the equator does not get warmer over time, there must be a transport of heat from the equator to the poles. Heat is transported from low to high latitudes in the air (atmospheric circulation) and in the oceans (oceanic circulation).

The circulation of air and water masses does not follow a straight line from the equator to the poles. Their trajectory is complicated by the Earth's rotation and constrained by the topography of ocean basins and landmasses. To understand the effect of Earth's rotation, let's first consider the example of a fixed object (Fig. 56A). If this object is situated at one of the poles, it will spin around on itself (with no eastward velocity). As the object's position gets farther away from the poles, it will move in the same direction of Earth's rotation faster and faster (its eastward velocity increases). Now let's consider the effect of Earth's rotation on a moving object, say a projectile launched from France in the direction of the equator (Fig. 56B). In the case of a non-rotating Earth, the object would follow a longitudinal line (Fig. 56B-1). In reality, the projectile seen from the perspective of an Earth-bound observer is not following a straight line but is deflected to the right (Fig. 56B-2). This is because Paris is moving eastward more slowly than any point situated on the equator. If the projectile had been launched from somewhere in the southern hemisphere toward the equator, its trajectory would have been deflected to the left. This is due to the ***Coriolis effect***. The Coriolis effect influences the trajectory of any objects or fluids moving in any direction in the air, in the ocean, and even in the liquid outer core. For air or water masses in motion, the result is a deviation to the right in the northern hemisphere and to the left in the southern hemisphere. The Coriolis effect becomes significant for any object traveling over long distances but it affects moving objects at all scales. For instance, “a ball thrown horizontally 100 m in 4 s in the United States will, due to the Coriolis force, deviate 1.5 cm to the right” (quoted from Persson, 1998).

Figure 56: Illustration of the Coriolis effect. The eastward velocity of a fixed point on Earth is maximum at the equator and 0 at the poles (A). From the viewpoint of an Earth-bound observer, an object launched from Paris toward the equator would follow a longitudinal line if the Earth was not rotating (B-1) and follows a curved path deflected to the right in the case of a rotating Earth (B-2). The deflection occurs because the object leaves Paris with an eastward velocity lower than the eastward velocity of the targeted point on the equator. The deflection would be to the left if the object was launched from a location in the southern hemisphere.

Let’s begin our description of the atmospheric circulation by the equatorial region (Fig. 57). In this area, solar heating of Earth’s surface is maximum. Hot, low-density air rises high in the troposphere, producing a low atmospheric pressure belt. As the rising air cools, water vapor — coming from the evaporation of ocean water— condenses and large clouds form, resulting in abundant rainfalls. These rainfalls sustain highly productive rainforests in the equatorial region (e.g. Amazon, Ituri). As the air moves away from the equator at high altitude, it cools further, becomes denser, and sinks around 30º of latitude (area of high atmospheric pressure). Closer to the ground, air rushes toward the equator to replace the hot rising air. This produces the ***trade winds*** which blow west due to the Coriolis force and transport abundant water vapor that feeds the large equatorial clouds. The area where northern and southern trade winds meet is called the ***intertropical convergence zone*** (ITCZ). The combination of the high-altitude movement of air away from the equator and the surface trade winds blowing toward the equator represents a component of the global atmospheric circulation called the ***Hadley Cell*** (Fig. 57).

Near 30º of latitude, dry air sinks and becomes warmer due to the increase in atmospheric pressure at lower altitudes (think about the increase in temperature when you compress air in a bicycle pump). The sinking air is dry because most of its water vapor was released as rainfalls near the equator. Rainfalls are therefore limited in this region. What makes rainfalls even less likely is the increase in temperature which allows the air to hold more water vapor. Consequently these high-pressure subtropical areas are characterized by an arid climate (e.g. Sahara, Kalahari, Gobi, Australian desert). Part of the sinking air returns to the equatorial region (carried by the trade winds) and another part moves poleward. This poleward flow is deflected by the Coriolis force and produces winds called the ***westerlies*** which transport heat from low to high latitudes (blowing in a

SW-NE direction in the northern hemisphere).

The ***polar front*** is a zone where the warm air carried by westerlies meets the much colder air moving away from the poles. The warm air rises above the cold, denser air and produces clouds and rainfalls as water vapor condenses at high altitudes. North of the polar front, the cold polar winds are called ***polar easterlies*** (blowing from the NE in the northern hemisphere).

Figure 57: Simplified representation of Earth's atmospheric circulation. Note that other models include two other cells, one at mid-latitudes and another at high latitudes. Source: adapted from Earth's climate: past and future, Ruddiman (2001)

5.3. The hydrosphere

The oceans too carry heat from the tropics to the poles. There are two types of ocean circulation involved in this heat transport: the ***surface*** —wind-driven— ***circulation*** (Fig. 58) and the ***thermohaline circulation*** (Fig. 59). The former results from the action of winds that push water masses horizontally and affects the top 100 m of the water column. Because of the Coriolis force, surface currents are deflected to the right of the wind direction in the northern hemisphere and to the left in the southern hemisphere. As a result of this and of the topography of ocean basins and landmasses, surface currents delineate large ***gyres*** —clockwise in the northern hemisphere and anticlockwise in the southern hemisphere—. Ocean currents are not limited to the surface. There is also a slow oceanic circulation driven by differences in water density which affects shallow and deep water masses. Ocean water density is a function of temperature and salinity, hence the name thermohaline circulation (*therme* = heat, *halos* = salt). There are areas at high latitudes where dense ocean water (salty and cold) sinks toward the bottom of the ocean (***downwelling***). A major source of deep water is the North Atlantic region near Greenland where water becomes saltier due to evaporation and is chilled by cold winds. The cold, dense water sinks to the bottom of the Atlantic and starts its deep journey in the direction of the southern hemisphere. Another area of deep water formation is the South Atlantic region near Antarctica (e.g. the Weddell Sea, not shown in Fig. 59). There is no deep water formation in the Pacific. Since water masses sink in some areas, there must be other areas where water masses return to the surface (***upwelling***). Upwelling can occur along the equator where trade winds push surface waters in opposite directions (Fig. 60B). It can also occur along the west coast of America and Africa where surface winds blowing along the coast push water masses offshore (Fig. 60A). The thermohaline circulation is very slow. It takes on average more than one thousand years for ocean water to return to the surface after it sinks in the North Atlantic.

Figure 58: Surface, wind-driven circulation. Red lines: warm surface currents, blue lines: cold surface currents. Source: Understanding Earth 6th edition.

Figure 59: Thermohaline circulation. Note that this map is highly simplified and that all the downwelling and upwelling areas are not shown. Source: Understanding Earth 6th edition

Figure 60: Upwelling areas. (A) Upwelling along a coast where winds push surface water away from the coast and (B) upwelling along the equator where winds push surface water away from the equator. Source: Earth's climate: past and future, Ruddiman (2001).

5.4. The cryosphere

The cryosphere comprises the Antarctic and Arctic ice sheets (Fig. 61), continental glaciers and snow, sea ice, and permafrost (permanently frozen soils of high-latitude regions). The cryosphere influences the Earth's climate because of the high albedo of snow and ice and because of its influence on the thermohaline circulation.

Figure 61: Extent of the Antarctic ice sheet (left) and arctic ice sheet (right) in summer. Source: NASA.

To understand its influence on the thermohaline circulation, think about the impact of the melting of the Greenland ice sheet in response to global warming. More fresh water added to the North Atlantic ocean would decrease the salinity —hence the density— of surface water and prevent it from sinking. This would slow down or even shut down the thermohaline circulation. The consequence would be an alteration of the heat transfer between low and high latitudes in the North Atlantic. It would have a great impact on the climate, perhaps leading to a significant cooling of Western Europe. Questions related to the current state of the thermohaline circulation and the impact a slowdown (or shutdown) of the thermohaline circulation would have on the climate are particularly relevant in the present context of global warming.

5.5. The lithosphere

The lithosphere influences the climate in a number of ways at various spatial and time scales. The nature of the land surface determines the albedo. In addition, the land tends to warm and cool more quickly than the ocean. Large bodies of water, particularly the oceans, can store a large amount of heat in summer and release it slowly in winter. Oceans tend to buffer changes in surface temperature. As a consequence, the annual range of surface temperature of inland continental areas is much larger than along the coast or offshore.

In the tropics, the difference between the thermal properties of the land and the ocean gives rise to the ***monsoonal cycle*** (e.g. Indian monsoon). In summer, the land warms quickly whereas the ocean remains relatively cold. Warm air above the land rises and draws humid air masses in from the ocean. The air rises, condenses and produces abundant rainfalls (Fig. 62A). In winter, the land cools more quickly than the ocean. The air tends to sink over the land. It then flows toward the ocean where it warms, becomes humid, rises and produces rainfalls offshore (Fig. 62B).

Figure 62: Monsoonal cycle. (A) Summer monsoon and (B) winter monsoon. See text for explanations. Source: Earth's climate: past and future, Ruddiman (2001).

The effect of land topography is also crucial. When warm air that has swept over the ocean and taken up abundant water vapor meets a mountain chain, the air rises and water vapor condenses, forming clouds and producing rainfalls (orographic precipitation). The air that flows down the opposite flank is dry. This results in a low-precipitation area on the leeward side (opposite to the windward side), a phenomenon referred to as the ***rain shadow effect*** (Fig. 63).

Figure 63: Orographic precipitation and rain shadow effect. See text for explanations. Source: Earth's climate: past and future, Ruddiman (2001).

Volcanoes release greenhouse gases in the atmosphere that can influence the climate (e.g. CO_2 and water vapor, see section 8.8. for more information on the greenhouse effect). Large volcanic eruptions can also produce a tremendous amount of dusts which may block part of the incoming sunlight.
On a much larger time scale (over millions of years), plate tectonics controls the topography and shape of ocean basins and landmasses. Plate tectonics therefore controls ocean circulation patterns and influences the transfer of heat from low to high latitudes. The absence of a landmass at the poles makes it more difficult for snow and ice to accumulate and build up ice sheets.

5.6. The biosphere

Living organisms have a crucial influence on the uptake and release of greenhouse gases. Using photosynthesis to fix inorganic carbon (CO_2), terrestrial and marine plants store a large amount of carbon derived from carbon dioxide. Some of this carbon dioxide is returned to the atmosphere during cellular respiration.

Photosynthesis: $6H_2O + 6CO_2 + \text{solar energy} \rightarrow C_6H_{12}O_6 + 6O_2$

Cellular respiration: $C_6H_{12}O_6 + 6O_2 \rightarrow 6H_2O + 6CO_2 + \text{chemical energy}$

Transpiration, the evaporation of excess water excreted by plants, accounts for as much as 10% of the moisture in the atmosphere, the other 90% being linked to evaporation of the oceans (source:

USGS). ***Evapotranspiration*** (the sum of evaporation and plant transpiration) influences air temperature and the amount of precipitation. The transformation of liquid water to water vapor requires energy and this energy is taken from the environment as heat. Heat is absorbed by evaporating water which tends to decrease the temperature of the environment, hence the cooling effect of plant transpiration (and sweating!). In addition, the water evaporating from plants can condense at high altitude and produce rain, influencing local precipitation patterns.

5.7. Earth's radiation balance

The source of energy driving the climate system is the sunlight (external forcing). The energy radiated by the Sun and reaching the Earth consists mostly of visible light (***shortwave radiation***). The average amount of solar radiation arriving on top of the atmosphere is 342 W/m2 (1 W = 1 Joule per second). What happens to this incoming radiation? (Fig. 64)

70% (240 W/m2) is absorbed by Earth (30% by the clouds and 70% by Earth's surface) and 30% is reflected/scattered back into space (85% by the top of clouds and 15% by Earth's surface). The 240 W/m2 absorbed by the Earth drives the climate system.

Because Earth's temperature is constant, the energy absorbed (240 W/m2) must be radiated back into space. In other words, the radiation balance must be maintained. Any object warmer than absolute zero (-273ºC) emits light. In the case of Earth, the light that is radiated by Earth's surface and by the top of clouds is in the infrared range (***longwave radiation***). What happens to the infrared light emitted by Earth's surface?

The amount of radiation sent back into space (240 W/m2) comes mostly from the infrared light emitted by the top of clouds. Very little of the radiation emitted by Earth's surface makes it into space (only 5%). If we translate the infrared light radiated back into space (240 W/m2) in terms of temperature, we would obtain an average surface temperature of -16ºC. That is far below the actual value of +15ºC at Earth's surface! The difference is due to the infrared light that is absorbed and re-emitted back to Earth's surface by clouds and greenhouse gases. This makes the Earth's surface 31ºC warmer than it would be in the absence of greenhouse effect! Note the dual effect of clouds on climate: (1) cooling due to the reflection of the incoming solar radiation and (2) warming because clouds re-emit infrared light back to Earth's surface. The most important greenhouse gases are water vapor, carbon dioxide (CO_2) and methane (CH_4). In the following two chapters of this course, we will talk more about the impact of human activities on the concentration of greenhouse gases in the atmosphere and their influence on climate.

So far in our discussion on the Earth's heat budget, we have been talking about heat transfer involving thermal radiation (solar radiation in the visible range of light and radiation emitted by Earth's surface and the clouds in the infrared range). There are two other important processes contributing to heat transfer that influence the climate system:

(1) One involves a transport of heat by moving air masses. Hot, low-density air masses tend to move upward whereas cold, high-density air masses tend to move downward. This vertical movement of air masses driven by density differences is called ***convection*** (same process that drives movements of matter in the asthenospheric mantle). Heat carried by moving air masses is called ***sensible heat***.

(2) The other process involves heat released during condensation of water vapor and heat absorbed during evaporation of liquid water. Condensation of water vapor leads to a warming of the environment, whereas evaporation of liquid water leads to a cooling of the environment (remember the effect of plants transpiration). This heat is called ***latent heat***.

Figure 64: Earth's radiation budget. See text for explanation. Source: Earth's climate: past and future, Ruddiman (2001).