Meteorology and ecoclimatology

Course XXXX

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Contents

Prerequisites

Part I

Basic weather elements

Chapter 1

Introduction, Energy and Light

1.1 The earth's atmosphere

The atmosphere and changes in the atmosphere are central in this course.

1.1.1 The earth' spheres

The **atmosphere** is the air above the earth surface (for several km) which interacts with all the other spheres (pedosphere – soil, hydrosphere – water, cryosphere – ice on land, anthroposhere – part of the earth controlled by humans and biosphere – biotic part of the earth).

We will talk about the atmosphere and its interaction with the biosphere (last part of this course) but also its interaction with the hydrosphere, pedosphere and cryosphere.

Links between the different spheres are made though fluxes of energy, water, gases (e.g. CO2) and the biogeochemical cycles.

1.1.2 The atmosphere's composition

The atmosphere/air is gas mixture of almost 80% N2 and more than 20% O2. There are also some noble gases which, together with N2 and O2, are present at constant concentrations in the atmosphere (no mather where on eath you take an air sample). That's why they are called **permanent gases** in the atmosphere.

In the contrary, the **variable gases** are present in really low concentrations (< 0.1 volume percentage except for H2O vapor which can vary considerably between 0 and 4 volume percentages). The other variable gases, which depend on the location on earth and time of the day, have variable concentrations. These are some **greenhouse gases** such as H2O, CO2, CH4, N2O and O3. It is important to know that greenhouse gases such as CO2 are present in very low concentrations (4 ppm $<$ 1%) but nevertheless have a very large impact on radiation balance of the earth and thus climate change. We can also find **aerosols** suspended in the air and CFKs which are responsible for the hole in the ozone layer (again, low concentrations but very reactive).

Table 1.1: Example of Table

Johnson Family	Transformation	Parameter Conditions	X Condition
S_B	$Z = \gamma + \eta ln(\frac{X-\epsilon}{\lambda+\epsilon-X})$	$\eta, \lambda > 0, -\infty < \gamma, \epsilon < \infty$	$\epsilon < X < \epsilon + \lambda$
S_L	$Z = \gamma + \eta ln(X - \epsilon)$	$\eta > 0, -\infty < \gamma, \epsilon < \infty$	$X > \epsilon$
S_U	$Z = \gamma + \eta \sinh^{-1}(\frac{X-\epsilon}{\lambda})$	$\eta, \lambda > 0, -\infty < \gamma, \epsilon < \infty$	$-\infty < X < \infty$

We can easily refer to Table 1.1.

Very important for weather is **water** in the atmosphere. It is the gas in the atmosphere which is present at the most variable concentrations. It is continuously present in the atmosphere in three phases: as a gas (water vapour), solid particles (ice – high white clouds) and liquid particles (water droplets – lower gray clouds). But it is also an important greenhouse gas (in abundance the most important greenhouse gas but it has not shown the recent exponential increasing trend like CO2 and CH4). But most importantly, H2O is responsible for a very large part of the energy transfer on the planet through phase transitions (latent heat). Evaporation of water consumes a lot of energy while condensation of water releases a lot of energy.

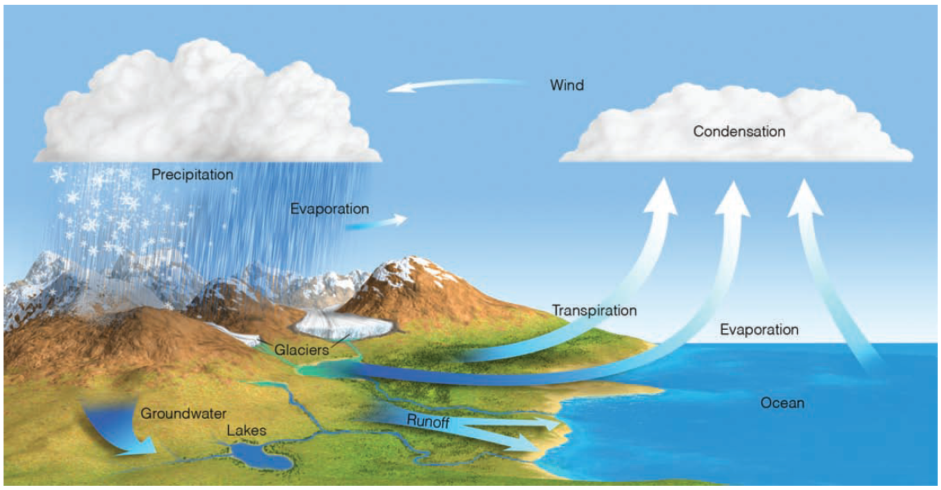


Figure 1.1: Here is the first figure

We can easily refer to Figure 1.1.

Aerosols are all tiny solid and liquid suspended particles in the atmosphere which play a very important role as condensation nuclei for cloud formation but also have an impact on radiation balance. They can originate from anthropogenic (e.g. industry) sources and natural sources (e.g. volcanic eruptions). They cause the so-called global dimming effect. Generally, they are present in low concentrations but these concentrations vary strongly in space and time depending on different individual events or trends such as volcanic eruptions or industrial activities (the latter is cause cleaner air in Europe and more aerosols over China over the past decade).

1.1.3 The atmosphere's layers

Because of gravity almost all air particles are in the first km above the earth surface. The air becomes very thin very fast. The **air density** will decrease exponentially with height. The **air pressure** will decrease exponentially because air pressure is the weight of the air above it. If you are on top of the mount Everest (5.5 km high) you are above 50% of the air molecules. In this way we can look at the atmosphere vertically.

However, in meteorology we mostly look at **the vertical temperature profile**. Based on this profile the atmosphere is divided in layers. One of the only ways to measure a vertical temperature profile is to use a radiosonde (i.e. weather balloon). Typically, the temperature declines when the balloon goes higher till it reaches -50°C at an altitude of around 10 km (airplanes fly at this height). The temperature declines in this first layer (the **troposphere**) because the sun heats the surface, so the further from the surface the colder it gets. The altitude where the temperature eventually stops decreasing is called the **tropopause**. Above the tropopause there is a **permanent temperature inversion** where the temperature will increase with height, which is called the **stratosphere**. The inversion is caused by the ozone layer, where ozone captures UV-radiation of the sun, heating up the layer from the top to the bottom. The stratosphere is a very stable layer with cold air the bottom and warm air at the top (no turbulence, no transfer between layers – lid on the troposphere, see chapter on atmospheric stability). Troposphere is most important for us because this is where the weather is. Everything we will discuss in this course are variations in the troposphere. You don't have any cloud formation above the tropopause. The stratosphere/inversion reaches till 50 km high where there is again a stabilisation and the temperature will decrease with the height again (**mesosphere**). Eventually we reach the **thermosphere** where the temperature increases very quickly because molecules will react very strongly with incoming solar radiation, and solar winds. Dependent on the solar activity the temperature curve will look differently (there is no buffer effect from the stratosphere here).

Homosphere is the lower part of the atmosphere where the chemical composition is very constant (except ozone in ozone layer) while in the **heterosphere** is very variable because of the interaction with solar radiation.

10 km is the average **thickness of the troposphere**. Because of the earth rotation the height of the tropopause is highest at the equator (18 km) and lowest at the poles. This thickness also varies with the seasons, in the summer (in the NH) there is an expansion of the lower layer because there is more warming of the earth surface. Therefore, the highest clouds found in Belgium are at around 10 km high while in the tropics this will be almost double the height (till 18 km high) and at the poles only 6 or 7 km high.

1.2 Meteorology and ecoclimatology

1.2.1 Difference bteween weather and climate

What is the difference between weather & climate? Both terms are pointing at the condition of the atmosphere, the difference lies in the temporal perspective. **Weather** is about the short term, the variation in the atmosphere from day to day, within days. **Climate** is average weather (e.g. "what is the average temperature in September in Belgium", "How will this climate vary from month to month, over the years"). Climate change is a directional change of the average weather (e.g. "Is our climate becoming on average warmer?") within timescales of decades, centuries.

If the figure below presents the probability function of temperature (or wind speed or air humidity), then in meteorology we want to know where we are on this curve next Tuesday for example. We want to predict or understand why it was 25 °C yesterday and will be 15°C next Tuesday. Climatology is what is the shape of this curve for our city, where is the average for our city and how do the tails look for our city. If we study climate change, then we want to know if the curve will shift, will the shape change, will the average T be higher in our city, will the chance to have extreme temperatures change in time? So, there is a difference in perspective. Weather models want to predict where exactly we are on this curve, while climate models want to predict how this full curve will look at the end of the century, and not where we will be on the curve on the 25 of October 2098.

1.2.2 History of meteorology

Aristotle was a philosopher but also the first meteorologist who wrote a **descriptive book** (Meteorologica – "everything that happens in the air") about clouds but also about falling stars and celestial bodies. The **next step** happened more than one thousand years later with the invention of **measuring equipment**. Galilei was the first person who made an instrument to measure the temperature, the thermoscope (bubbles that rise or descend in liquids depending on the temperature). Several decades later the barometer was invented. This evolved till we had a set of instruments to measure weather variables and we could start to continuously measure weather (first observations in Ukkel in 1833). Then the first **computers** evolved to super computers and were used quite rapidly for weather predictions and climate modelling. After the second world war the first **radars** were used for cloud observation and **weather satellites** were launched. Today, ground observations are still key and are combined with remote sensing and simulation models. What we see in weather reports is based on the combination of these different components (e.g. weather stations, remote sensing, climate models). Finality of the part meteorology in this course is reading and interpreting weather maps that synthesize al key weather elements.

1.2.3 Weather and climate in our daily lives

Weather and climate are very important and determine a lot of aspects in our lives (e.g. agriculture, forestry, environmental issues, housing, economy, clothing) especially extreme events as well as day to day weather.

1.2.4 Ecoclimatology

Ecoclimatology is an interdisciplinary science which links ecology and climatology. It is about the link between ecosystems and the climate, between the biosphere and the atmosphere and especially it is about the interactions. These interactions are determined by fluxes of energy, water and chemical elements which are exchanged between the vegetation on the earth surface and the atmosphere.

In the ecoclimatology part of the course, we are going to talk about biogeography, how does the climate determine which vegetation occurs on certain places on the planet, but also about the impact of climate variations on crops, plants, natural ecosystems and what are the feedbacks (how do ecosystems affect the climate in their term). We will also discuss and use vegetation models, which are important tools to study these interactions.

1.2.5 History of Ecoclimatology

Ecoclimatology is a younger and less developed scientific branch which started with Theoprasthus (student of Aristotle) who wrote a descriptive, observational book about plants and where they were found, linked with weather patterns of that place. In the 1800s (when measurements were possible), Alexander **von Humboldt** was the first one who really made the link between climate and the presence of certain plants. Later, others continued his work (e.g. vegetation zones, Köppen classification) and now there is also a lot of modelling.

1.2.6 Biogeoscience

Biogeoscience is closely related to ecoclimatology and is situated on the intersection of the different spheres and studies interactions between the different spheres. A lot of the current environmental issues/problems have to be studied within the biogeosciences, especially when we want to look at the anthropogenic impact.

A good example of the role of vegetation in land-atmopshere interactions is the given by forests. Forest provide ecosystem services, but these are dependent on the type of forest (e.g. a tropical rainforest versus a boreal forest will affect the climate in different ways). For example, the impact on the albedo (reflectivity earth surface) will be greater when a boreal forest grows than the contribution of a tropical forest. In the contrary, the evaporation and carbon storage of one hectare of tropical forest will be greater than one hectare of boreal forest ((-) cooling effect (+) warming effect).

1.2.7 Key land-atmosphere interactions

Interaction between land and climate has an influence on a lot of biophysical processes. The reflectivity (**albedo**) of the earth surface will influence the energy balance. The **roughness** of the earth surface is determined by the type of vegetation, causing different wind patterns and turbulence, which in its turn impacts the **heat and gas exchange**. The physiology of the stomata of plants will have an influence on **water exchange**. Soil moisture will also have an impact. The **carbon cycle** has an impact on the CO2 in the atmosphere and will depend on the type of vegetation. **Nitrogen exchange** (N-deposition from industry or N2O as greenhouse gas from soils) will be different in an agricultural area compared to a forest. **Aerosols** will determine the solar radiation a forest gets which has an impact on photosynthesis and the carbon balance. Forests will also emit **volatile organic carbons** (e.g. isoprene a lot in tropical forests) which are precursors of aerosols, so forests have an impact on the amount of aerosols in the atmosphere (also when a forest burns this brings a lot of particles in the air). Ecoclimatology studies many of these elements.

1.3 Energy, temperature and heat

1.3.1 Definitions

Energy is defined as the capacity to do work. Potential (static) and kinetic energy (dynamic) are types of energy. Energy can take on different forms. **Temperature** is a measure for kinetic energy (e.g. air temperature is the kinetic energy of the air molecules in our atmosphere). Therefore, making an energy balance is essential to understand the climate. **Heat** is the exchange, transfer of energy from one medium to another, it is a flux of energy.

1.3.2 Temperature

Temperature is a **measure for kinetic energy** (e.g. how much will the molecules collide with each other and against the edges of the volume it is confined in). Temperature is measured in ° Celsius, Fahrenheit, Kelvin (most scientific scale). The average temperature on earth is typically 15 °C, but weather stations typically measure variations between -30 °C up to 40 °C.

1.3.3 Specific heat

Different media which exist in natural systems can have very different specific heats expressed as J per kg per °C, it is the amount of energy needed to increase the temperature of 1 kg of a substance with 1 °C (e.g. a rock or sand will heat up faster than ice or wet soil – see table x). Water needs a very high amount of energy to heat up 1 °C (you need four times more energy to heat up a kg of water than a kg of air). Ice needs less energy than liquid water which is important for the climate system (e.g. oceans heat up slower than land and humid areas have a large buffering capacity). This is also why there is so much heat transfer involved with evaporation and condensation of water (latent heat).

1.3.4 Latent and sensible heat

Sensible heat is the energy used to change the temperature of the air. Sensible heat flux can be measured by a temperature change. **Latent heat** is the energy used to change the phase of a substance while the temperature does not change (e.g. energy needed to evaporate water or melt ice). **Evaporation is a cooling process** for the environment because the substance takes up heat from the environment to evaporate while **condensation is a warming process** for the environment because the process releases heat to the environment. Important to understand these concepts to understand the climate system (e.g. when a cloud forms water vapor will condensate to form water droplets which is accompanied by a release of energy to the environment.)

1.3.5 Heat transfer in the atmosphere

There are three ways in which heat is transferred in the atmosphere: **convection, conduction and radiation**. Radiation is based on radiation from the sun or other objects (every object with a temperature higher than 0 K emits radiation). Radiation does not heat the medium, air (imagine feeling the radiative heat of the sun on your skin on a cold winter day). Convection is the transfer heat via a fluid which is typically air in meteorology (hot air bubbles which are moving to transfer energy in the atmosphere). Convection does heat the medium. Conduction is heat transfer through a solid substance. This can be neglected in meteorology because air has a really low heat conductivity. There is only a small amount of heat that is transferred via conduction from the soil to the first layers of air above it. The largest heat fluxes on earth are governed by radiation and convection.

Convection happens when the sun heats a specific spot on the earth (for example a dark plowed field that heats up more than the grassland surrounding it), the air above this hot surface heats up and this hot air bubble rises and moves in the atmosphere (**thermal**). A lot of the energy transfer on earth happens through these thermals, through convection. This is not the same as **advection** which is the horizontal transfer of any property, this can be energy a gas or pollutants (e.g. cold air or pollutant sliding of a mountain) while convection is the three dimensional transfer of heat through hot air bubbles.

1.4 Radiation

Radiation is electromagnetic waves which don't heat the medium (air). Direct sun rays lose almost no energy before reaching earth. The **wavelength** of radiation determines its energy. A quantum of light with a high frequency (short wavelength) has more energy than one with a high wavelength.

1.4.1 Important laws

A first important law describes the energy of a photon (ep) with a certain wavelength (lambda):

$$e_p = h\frac{c}{\Lambda} \tag{1.1}$$

We can also easily refer to Equations, see Eq. (1.1).

Secondly, **Planck's law** relates the radiant flux density per unit wavelength emitted by a black body (W m-2 m-1) to the wavelength (lambda) and temperature (T) (e.g. spectrum of the sun when you fill in T of the sun):

Thirdly, **Wien's displacement law** relates the wavelength of maximum emission to the temperature of a black body:

Lastly, the **Stefan-Boltzmann law** relates the radiant flux density emitted by an object (E) to its temperature obtained by integrating over all wavelengths:

These laws are summarized in the figure which gives the spectrum of emitted energy for the sun and the earth (unit: W m-2 µm-1 – the energy intensity per spectral band spectral intesity). The curve is determined by the temperature of the sun (6000 K) and earth (288 K) (Planck's law). The colder the object the longer the wavelength at which this maximum energy intensity is reached (Wien's law). The amount of total radiation is given by the surface under the curve (Stefan-Boltzmann law). The sun emits **shortwave** radiation (lambda max at 0.5 µm) while the earth emits **longwave** (invisible) radiation (lambda max at 10 µm) (both short and longwave radiation are important radiation components). Thus, the sun emits compared to the earth, radiation of a different quality and quantity.

1.4.2 The sun's EM spectrum

This figure shows the percentages of solar-energy in the different wavelength bands. Almost everything is infrared (37+11) and visible light (44). Only a few percentages are UV light, however, containing a lot of energy (ozon layer protects us from this). This is the spectrum that reaches the earth before entering the atmosphere.

1.4.3 Absorption, reflection and transmission

Once the radiation enters the atmosphere, there is a lot of interaction of the radiation with the molecules in the atmosphere and the earth surface (this changes the nice curve from before). The radiation can be reflected, absorbed or transmitted (e.g. light transmitted through the leaves of a tree).

1.4.4 Scattering

Scattering happens when the sun rays collide with molecules and are reflected resulting in diffuse radiation. There are two mayor ways of diffusion. Firstly, **Rayleigh scattering** is the scattering of light by gas molecules which have a smaller diameter than the wavelength of the light. This is a continous form of scattering of which the intensity is inversely proportional to the wavelength (~ lambda-4). So, short wave lengths (high energy) will be scattered more than long wavelengths. This is the reason blue light is scattered more and the sky looks white at noon looking straight at the sun, blue when not looking straight at it and red in the evening (looking straight at the sun) as all the blue light is already scattered away. Secondly there is **Mie scattering**, which is the scattering of light by molecules with a diameter larger than the wavelength of the light (e.g. aerosols, water droplets in clouds, ice, smoke). For Mie scattering, the intensity is not proportional to the wavelength, every wavelength is scattered equally in all directions (this is why clouds look white or grey).

1.4.5 Direct and diffuse radiation

The **shortwave direct** (Sh) and **diffuse** (Sd) **radiation** coming from the sun are measured with a **pyranometer**. This sensor measures the total shortwave radiation (St=Sh+Sd) in watts per square meter, so the radiation measured is also dependent on the solar angle (low angles, greater spread over the surface). When using a pyranometer which tracks the solar activity and always measures the direct shortwave radiation perpendicular (Sb), we correct for the solar elevation ( ): Sh=Sb x sin . On average we would measure 740 Watt/m2 at noon on a sunny day in Belgium.

This figure shows the shortwave radiation coming from the sun measured at the earth surface (during a cloudy day). Compared to the radiation measured at the top of the atmosphere certain wavelengths have vanished. These wavelengths were absorbed by molecules in the air. The energy quantity coming from diffuse radiation is typically less than the energy coming from direct radiation. The quality is also different, the peak in diffuse radiation can be found in the shorter wavelengths (blue light). Plants can efficiently use these wavelengths present in diffuse radiation.

These figures present the sensitivity of the human eye and the sensitivity of plants to light. Plants typically absorb (pigments in the leaf) blue and red light and less green light which is why they look green. This total spectrum is slightly different than the spectrum which the human eye is sensitive to but it spans the same wavelengths (400 nm -700 nm). **PAR (photosynthetic active radiation)** is the radiation within these wavelengths which plants can use for photosynthesis (µmol photons m-2 s-1). This is approximately half of the incoming shortwave radiation. However, this also depends on whether the radiation is direct or diffuse and the solar elevation. In diffuse radiation there is relatively more PAR radiation and at lower solar elevations there is relatively less PAR radiation. On a sunny day there can be 2000 µmol m-2 s-1 PAR radiation. The PAR fraction in diffuse and direct light depends on the solar elevation (Table) at lower solar elevation the path length through the atmosphere of (especially direct) light is longer, blue light is scatter out, less PAR remains.

1.4.6 Selective absorbers/emitters

The spectrum of light that we receive at the surface of the earth looks different than the spectrum received at the top of the atmosphere because of selective absorbers. This figure shows the different **absorption spectra** for different gas molecules in the atmosphere. Ozon mainly absorbs UV light but also infrared light (making it a greenhouse gas). **Greenhouse gases** typically absorb infrared light. CO2, N2O but also water vapor and methane absorb a lot of infrared radiation. When we look at the total spectrum we see that a lot of wavelengths are filtered out before they can reach the earth surface. The visible light does reach the earth surface while UV (short wavelengths) and a lot of infrared radiation (coming from the sun but also from the earth) are absorbed. The **atmospheric window** (a zone with little absorption) is the group of infrared wavelengths that can leave the earth surface and the atmosphere again (this is how the earth loses energy via longwave radiation). However, this atmospheric window can be closed by clouds which is why a clear night is cooler than a cloudy night.

1.4.7 Greenhouse effect

When there would be no greenhouse effect (no selective absorbers), the earth would lose a lot of infrared radiation and the average temperature on earth would be -18°C. Luckily, there are greenhouse gases which cause this greenhouse effect resulting in a livable temperature (15°C on average) on earth. However, the problem is the increased greenhouse effect and not the greenhouse gases as such.

1.5 Energy balance

1.5.1 Radiation balance

When making an energy balance of the planet, we consider the solar constant. This is the energy that we continuously receive from the sun at the top of the atmosphere (= on average during the day we would measure at the top of the atmosphere 1360 W per square meter).

As we have seen before, when this energy from the sun enters our atmosphere there is interaction with molecules in the atmosphere (diffusion, reflection, etc). Firstly we make a radiation balance and secondly we make an energy balance. When making the radiation balance we calculate how much radiation energy the earth, or for example a grassland, receives and loses. We are calculating the **net radiation**, what is left at the earth surface. The radiation balance is made up off a shortwave ((1-)

St) and a longwave radiation balance (Ld – T04). The shortwave balance is the total shortwave radiation you measure with a pyranometer minus the reflected shortwave radiation. The reflected shortwave radiation is determined by the **albedo** ( ), the reflectivity of the earth surface. The longwave balance is the balance of infrared longwave radiation. There is longwave radiation because objects with a temperature higher than 0 K emit radiation and relatively cold objects (such as the clouds and molecules in the atmosphere (Ld) and the earth surface (T04)) emit longwave radiation. Finally, the net radiation balance is:

The net radiation (the energy the system receives – energy the system loses) is the radiation which is available for the system to for example heat the air or for evaporation.

The **albedo**, the reflectivity of the earth surface, is variable and depends on the color of a surface. Forests (dark) and water surfaces have a low albedo (they absorb a lot) while snow has a really high albedo (reflects a lot). Albedo measured with satellites (e.g. MODIS albedo) show us that the highest albedos can be found in places covered with snow and ice or in the desert, while places with a lot of vegetation like forests in the tropics or temperate areas have a low albedo.

The next two figures are examples of the diurnal cycle of the components of the radiation balance, where we can see the different components change during the day. At night there is no shortwave radiation from the sun. When the sun rises the incoming shortwave radiation increases reaching a maximum at noon. The reflected shortwave radiation is a fraction ( ) of the incoming shortwave radiation. The outgoing and incoming longwave radiation are quite constant (increasing a little bit when the surface and air heat up during the day). Most importantly, there is a **longwave deficit**, which means less longwave radiation is received than lost. So, during the night the earth is losing energy (no incoming shortwave, only longwave deficit) while during the day energy is won, when the net shortwave radiation (surface between the two shortwave curves) is larger than the longwave deficit.

When we make the **radiation balance** for the planet with incoming solar constant equalling 100 units, we see that only 51 units will reach the earth because 30 units are reflected in the atmosphere, by clouds or the earth surface and 19 units are absorbed by the atmosphere and clouds. This is the average radiation balance of the earth (because there is a large variation in local rardiation balances for different places and at different moments in time).

1.5.2 Energy balance

The net radiation we calculate from the radiation balance is part of the input of the energy balance.

The **energy balance of the earth** describes what happens with the **net radiation** that the earth receives from the sun. So on the left side we have the energy gain and on the right side we have the energy loss. You can make this energy balance for an object (house, the hman body, …), an ecosystem, for a region or for the whole planet. Another energy input could be metabolism but this is neglected in an ecosystem because the metabolic energy (for example from photosynthesis and respiration reactions) is only a small fraction of the total energy. The energy input is measured with a **net radiation sensor**. Energy is lost through **latent heat** (E, evaporation), **sensible heat** (H, increasing air temperature), through the **ground heat flux** (G, increasing soil temperature) or it is **stored** in the system (S, residual storage term because difficult to measure). These components are expressed in Watt per square meter or energy flux per second (W m-2 or J s-1 m-2). The latent and sensible heat are measured using the eddy covariance (see later in this course). Depending on the vegetation and water availability more energy will go to evaporation or increasing the air temperature. This is why a forest is cooler as more energy goes to evaporation than heating the forest compared to other systems with less vegetation. So, the energy balance is based on the conservation of energy (energy gained has to go somewhere). On the long term the ground heat flux and the storage term can be neglected in the energy balance as everything that is taken up during the summer is released during the winter.

This figure shows the average energy balance of the planet which is more or less in balance. It is not in balance in certain places or on certain points of time, this is why we continuously have temperature and weather variations. But on average the system is quite stable with a net balance at different levels (at the top of the atmosphere, at the earth surface). The shortwave balance is the shortwave energy received and reflected (like in the radiation balance) but the longwave balance also includes the latent and sensible heat in addition to the longwave radiation.

The energy budget can be considered in different ways (see figures, say the same thing but with different numbers). The last figure is not based on 100 units of solar radiation coming in, but on 341 W m-2 which is the average incoming solar radiation over a full year globally.

So the energy balance of the earth is on average in balance but this **balance can shift (Climate change)**. Climate change (natural or anthropogenic) is always related to one of three factors: **radiation, atmospheric chemistry and albedo**. Solar radiation (the input of the energy balance) can change in time when the sun is more or less active, or because of changes in the geometry of the earth and the sun (i.e. Milankovich cycles). The atmospheric composition can change which is mainly related to anthropogenic factors, greenhouse gases, but also aerosols. The albedo, the reflectivity of the earth surface, can change in time (vegetation cover, urbanisation, melting polar ice caps). So the energy balance is in balance when considering years but can change on the long term.

However, locally there is no balance (there is an **imbalance according to latitude**). Locally, we receive more energy at the tropics and less at the poles. Long wave deficit is larger at the equator because the surface is hotter but net we still get more energy at the equator and lose energy at the poles. So there is a continuous surplus at the tropics and deficit at the poles. Therefore, there are continuous heat transfers from the equator to the poles which drives the climate system.

1.6 Extra: Aurora borealis

Polar light is a phenomenon which you can observe in northern areas (aurora borealis) or on the south pole (aurora australis). It is a visual effect related to the magnetic field of the earth. Solar storms emit charged particles. These particles are deflected and reach the atmosphere near the poles. They react with molecules in the atmosphere which then are excited. When the electrons fall back, they emit light (polar light).

Chapter 2

Temperature, humidity and clouds

2.1 Seasonal temperature variation

Temperature is a measure for the **kinetic energy** of the atmosphere (more energy in the atmosphere, means a higher temperature). We're going to talk about the variations in temperature, the seasonal, diurnal and spatial temperature variations.

2.1.1 Seasons: why?

There are seasons on earth because of the variation in solar radiation in time. The main driving factor is the **tilted axis of the earth** (the earth's axis is not perpendicular to the plane of the earth's orbit around the sun, but is tilted). This results in a variation of the amount of solar radiation throughout the year but also depending on the location on earth. The amount of solar radiation depends on the **solar angle** (lower sun, solar radiation is spread over a larger surface, lower solar intensity and more scattering) and on the **day light hours**. Both of these factors are determined by the tilt of the earth's axis and the geometry of the earth's path around the sun. This path is elliptical (so the earth is not always at a same distance from the sun). It is the tilt which causes the seasons e.g. during summer here (northern hemisphere), we are furthest away from the sun but tilted towards the sun which causes a higher insolation (due to the solar angle and day light hours). During the spring and autumn we are closer to the sun. The figure here is an exaggeration of the elliptical path, in reality it looks more like a circle which is compressed just a little. Moreover, the sun is not located in the center of the ellipse so during the winter we are closer to the sun than during the summer (in NH). This might all be counterintuitive but shows that this distance to the sun is less important than the inclination of the earth's axis. You would expect that in the southern hemisphere the seasons are more pronounced because they are closer to the sun during summer and further away during the winter, however, this is compensated by the fact that the **southern hemisphere is covered by a larger amount of water which acts as a buffer**. Therefore, seasonality in temperature is less pronounced in the Southern hemisphere.

If the earth axis wouldn't be tilted, there wouldn't be any seasons. Every day would be the same, would be like 20th of March or the 22nd of September. But this is not the case, there is an inclination of about 23,5° which can vary a little over the years (see course Climate Change Processes).

On the 21st of June the angle of 90° is above the tropic of Cancer while on the 22nd of September it is above the equator and on the 21st of December it is above the tropic of Capricorn causing the dynamics of the seasons.

Above the polar circles there is also a special situation where during the summer the north pole has 24 hours daylight while during the winter it has none. So exactly on the poles there's six months of day and six months of night but going further away from the poles the shorter the polar night.

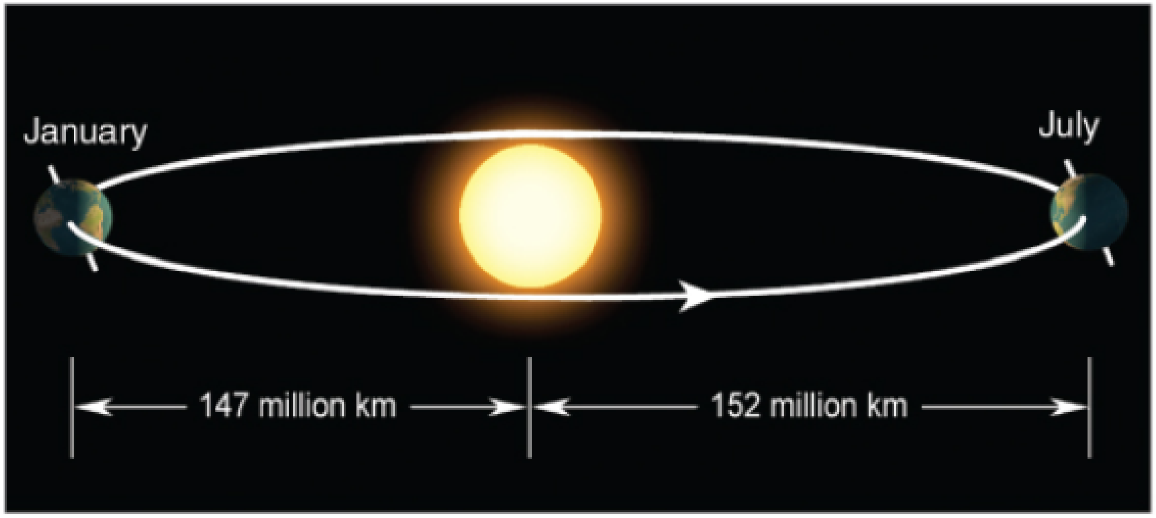


Figure 2.1: Figure caption

2.1.2 Impact on insolation

The above described variations have an impact on the total solar radiation which is available. This is calculated from the intensity multiplied by the daylength. The **intensity depends on the solar angle and the daylength** also depends on the solar geometry. The figure presents the insolation throughout the year for different locations on earth (different latitudes). At 60°N (a little more north than Belgium) we can see a typical strong **seasonal pattern** of insolation. When going to the north pole (90°N) this pattern is more extreme (six month with and without insolation). At 30°N (region around the Sahara), there is also strong seasonality but there is not a deep decline in winter due to the high solar angles. On the equator (0°N), and everywhere between the tropic of Cancer and the tropic of Capricorn (= the **tropics**), there is a completely different pattern, a **bimodal pattern**, because the sun is perpendicular to the equator at two times (September and March). On the equator the two peaks are nicely divided but going to the tropics the peaks come closer to the 21st of June. This is the resulting seasonality in insolation but this does not mean the temperatures will follow the same pattern. The temperature is not only determined by the incoming solar radiation but also by all the other factors of the radiation balance.

This figure shows the situation on the 21st of June (summer solstice). The blue line presents the insolation at the top of the atmosphere while the red line presents the insolation at the earth's surface, for the different latitudes. At the top of the atmosphere the insolation increases from the equator to 23.5°N where the sun is perpendicular to the earth (so mainly the solar angle is the driving factor here). Going further north the solar angle decreases again but the length of the day increases causes the total insolation to keep increasing. At the poles there is an extra increase because there is 24 hours sun there. Therefore the highest insolation over a day is reached at the north pole. However, at the earth surface the insolation decreases after the initial peak because with a lower solar angle comes longer paths through the atmosphere and a lot of scattering. At the poles there is also a lot of reflection from the polar caps (losing insolation). Clouds also play a role, they are the reason why the insolation increases between 23.5°N and 30°N, because there are no clouds above the deserts lying in between these latitudes while at 23.5°N there were clouds.

The **radiation balance** is in balance globally but not locally and not **according to latitude**. The net shortwave radiation is shown in blue which is high at the equator and decreases to the poles. The longwave radiation loss is shown in red and has two peaks at 23.5°N and 23.5°S as these are the warmest regions (see Stefan-Boltzmann). This results in a net **surplus at the equator** and a **deficit at the poles**. To compensate, a continuous flux of energy is needed from the equator to the poles which is realised through **ocean currents, wind circulation and latent heat** (more evaporation at the equator and more condensation at the poles). These three component each transport 1/3 of the energy.

2.1.3 Apparent path of the sun

If we look at the geometry of the path of the earth around the sun from the perspective of someone standing on the earth surface, we can see the apparent path of the sun. If you stand on the north pole you see the sun going in a circle, with this circle descending when winter is coming closer. The closer to the equator the more perpendicular the sun will relative to the earth. At the equator the sun will apparently rise straight up with a little variation to the north or south depending on the time of the year. This also causes the fact that in the tropics the transition between day and night will be very fast.

2.1.4 Solar geometry

We can also quantify solar elevation by using the **zenith angle**, this is the angle between the zenith (line perpendicular on the earth surface) and the Sun. The **altitude angle**, the angle between the horizontal and the line to the Sun is also used. The **azimuth angle** is measured as the angular distance from the north. These angles can be calculated with the help of several formulas.

$$cosZ = sinB = sin\phi \cdot sin\delta + cos\phi \cdot cos\delta \cdot cosh \tag{2.1}$$

- Z is the zenith angle
- B is the altitude angle
- is the latitude
- is the solar declination of the earth (varies around 23°)
- H is the solar hour angle (the angle between the solar noon and where it is at the moment of observation, it is expressed in ° or in hours, 1 hour before solar noon the hour angle is -1 or -15°)

$$cosA_{sun} = \frac{(sin\delta \cdot cos\phi - cos\delta \cdot sin\phi \cdot cosh)}{sinZ} \tag{2.2}$$

Asun is the azimuth angle.

The total amount of radiation received at the atmosphere is the **solar constant** of 1370 Watt per square meter (seen before). We can correct this for solar angle and the radius vector rv, relative distance between the earth and sun, the relative ratio between the current and average distance to the sun.

$$S_H = S_p cosZ \tag{2.3}$$

$$S_H = \frac{S_c}{r_v^2} cosZ \tag{2.4}$$

We can also calculate the total daylength, defined as the period during which the sun is above the horizon. See more details during the practical.

$$\frac{24}{\pi} cos^{-1}\left(-tan\phi \cdot tan\delta\right) \tag{2.5}$$

2.2 Daily (diurnal) temperature variation

The daily temperature variation is driven by the rotation of the earth and the day-night pattern. The rotation of the earth causes the alternation of day and night but also the rising and falling solar elevation.

2.2.1 Temperature profiles

When we measure the temperature at different heights above the ground we see an **exponential profile** during a calm day while on a windy day a **linear profile** is obtained because of the **turbulence** and mixing of the different layers due the wind. On a calm day this exponential profile is caused mainly by convection (hot air bubbles are created at the earth surface due to the heating of the ground by the sun, but very slow process) and a little bit by conduction (only in the very low layers).

During the night, cooling is strong at the earth surface (long wave radiation is leaving the surface = radiational cooling), causing a **nocturnal inversion**. This results in an increasing temperature profile with the coolest, heaviest air at the surface (which does not have the tendency to mix with the upper layers) and thus causes a stable atmosphere. Therefore, when comparing day and night the **temperature variation at the surface will be very large** (at night, the lowest temperatures are reached while during the day the hottest temperatures are obtained). This is why we don't measure the temperature at the surface but at **1.5 m high in a thermometer hut** (above the diurnal variation), so we can measure the variation from day to day. The temperature also depends on humidity, wind, albedo, vegetation, which is also why a thermometer is placed in a hut with shielding so it is not determined by direct radiation. Also during the night the profile is linear for a windy night and exponential when it is a calm night.

2.2.2 Temperature–radiation link

This figure shows how the temperature will vary during a sunny day (diurnal pattern of 24 hours) but it also shows the net radiation balance (net shortwave incoming & net long wave outgoing). The highest temperatures on a sunny day are not reached at the solar noon but a few hours later. The peak of temperature is not at the same time as the peak of solar energy because the temperature is the result of the total energy balance. During the day the incoming energy will be larger than the outgoing energy (positive net energy), which means that also after the solar noon the systems is receiving energy and thus keeps warming the atmosphere. Only when the incoming solar energy is lower than the outgoing energy (in the afternoon), there is a energy deficit and the atmosphere starts cooling. The temperature will decline as long as the outgoing energy is larger than the incoming energy, which is until a few moments after the sun rises (the first moments after the sun rises the incoming solar radiation is not yet enough to compensate the long wave deficit). In the early morning the coolest temperature is measured. The temperature profile is determined by the difference between incoming solar energy and outgoing energy but the outgoing long wave energy profile is also determined by the earth surface temperature. Therefore, the peak in long wave deficit will also be in the afternoon when the surface is hottest. In conclusion, temperature and radiation are linked in both directions. On a cloudy day the incoming solar energy will fluctuate more, causing the temperature (and thus the long wave deficit) not to follow this nice, theoretic pattern.

2.2.3 Thermal belt in valleys

The temperature profile has an impact on agriculture. For example, at night there will be advection (horizontal transport) of cold air downhill in the valley while warm air is stays on top. But because the temperature also decreases with height there will be a combination of these two effects and the highest temperatures are reached in the middle. This is called a **thermal belt**, a zone on hillsides where there is a moderate climate (it doesn't really cool off that much here) and where certain plants can grow which cannot grow below or above this zone. Vineyards will typically be located in these zones (not on the bottom of these valleys). Another consequence is that there is cold and stable air on the bottom of the valley where smoke and pollution will linger increasing the environmental and **pollution problem of cities located in valleys**.

2.2.4 Crop protection

The peak of low temperature in the early morning can be a problem for a lot of crops, especially in spring when the crops are growing. Therefore, there are several methods to deal with this spring frost. Firstly, there are **orchard heaters** which blow warm air but are not energy efficient. Secondly, a smarter method is the use of **wind generators** which cause turbulence and mixing of the cold air with the warmer air above (changing the exponential temperature profile into a linear profile with warmer temperatures near the ground). Thirdly, there is a method where you spray the crops with water just before the temperatures drop below freezing, which creates an **ice coating** and releases latent heat. Lastly, **plastic foils** are used to keep the heat inside (green house effect).

2.3 Regional temperature variation

2.3.1 Main temperature controls

The temperature varies worldwide (see world map which shows the average temperature over 30 years). From this map we can derive the **four main temperature controls** on earth. Firstly, the most important control is the **latitude** (~ solar angle, higher latitudes are colder). However, this map is not linear with latitude because of the three other temperature controls. Secondly, **land and water** transition also play an important role (lines deflect near land water transitions). Thirdly, **ocean currents** play an important role and explain for example why it is warmer in Europe than on the same latitude in North America. Lastly, **elevation** explains some anomalies such as the one near the Himalayas, Alpes, Rocky mountains and Andes. These four factors determine for the most part which temperatures there are on earth. But of course, there are also local factors which determine the temperature locally and regionally.

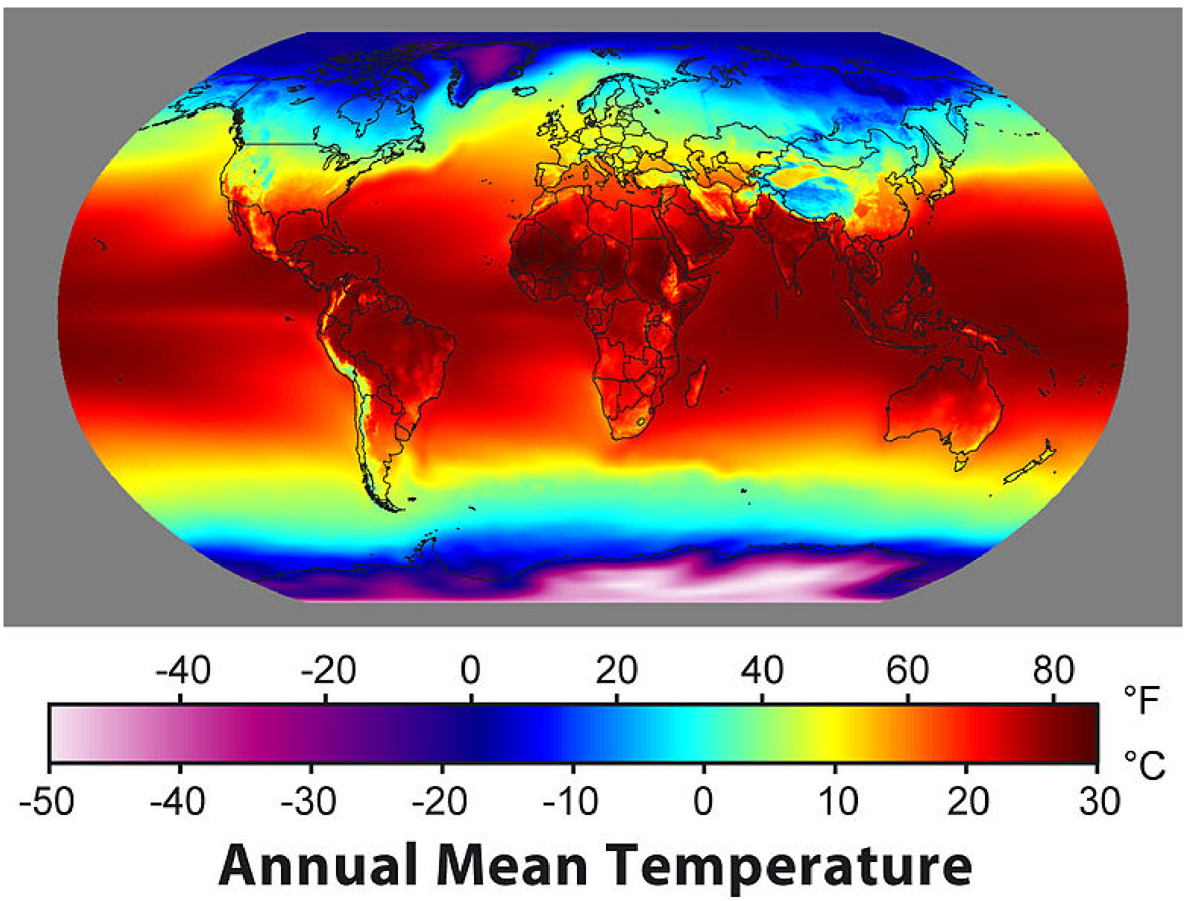


Figure 2.2: Figure caption MAT

2.3.2 Global isotherm maps (°F)

On this global isotherm map for January we can see the highest temperatures south of the equator. We see the **isotherms are closer to each other in the northern hemisphere** (e.g. strong temperature gradients and thus very variable weather). In the contrary in July, the isotherms have shifted up but also the isotherms are further away from each other meaning weather will be more stable.

We can also see that near **water-land transitions**, there are kinks in the isotherms. This is caused by the different energy balance between land and water due to the higher heat capacity of water. But these kinks are also related to the **ocean currents** (e.g. higher temperatures near Europe caused by the golf stream). Lastly, we see that in the **southern hemisphere the situation is stable throughout the year** (the isotherms are practically parallel and not close together). This is because there is significantly less land mass in the southern hemisphere.

2.4 Air temperature data

2.4.1 Temperature scales

$$K = \text{ř}C + 273.13 \tag{2.6}$$

$$\text{ř}C = 5/9\,(\text{ř}F - 32) \tag{2.7}$$

2.4.2 Thermometer huts

How do we measure these air temperatures? In standard circumstances, temperature is measured in a thermometer hut which is well ventilated and cut off from radiation. This thermometer hut is also placed on a standard height to be above the diurnal temperature variations. In the past this was done manually (someone went out two times a day to measure the temperature). With a minimum-maximum thermometer the minimum and maximum temperature of the last 24 hours can also be measured.

2.4.3 MAT (Mean annual temperature) vs seasonality

What do we do with this data? Often the mean annual temperature (MAT) of a location is used. But this does not say everything about the climate. For example, two different locations can have the same mean annual temperature but completely different climates.

2.4.4 Growing degree days

Growing degree days is a cumulative temperature sum. You choose a basis temperature (for example 0°) and measure the maximum temperature everyday (for example 5° today, 10° tomorrow, 6° the day after that). Then you make the cumulative sum of the differences between the maximum temperatures and the basis temperature (so 5+10+6=21 degree days). During the growing season you can further accumulate this temperature sum which is used to study or estimate different phenological phenomena of plants (for example, you need a sum of 1200-1300 to harvest certain crops).

2.4.5 Wind chill index

The temperature we feel depends on the wind and is determined through the wind chill index. Higher wind speeds gives us lower wind chill index (so it feels colder when there is more wind).

2.5 Atmospheric moisture (psychometry)

The study of air humidity is also called psychometry.

2.5.1 Three phases of water

Water is present in the air in three phases: solid (ice), liquid (water droplets), gas (water vapor). Above a water surface there is a dynamic equilibrium, where water evaporates and condensates continuously.

2.5.2 Saturation and condensation

If you have a closed jar with water and air, an equilibrium will be set, where there will be and equal amount of evaporation and condensation. In this situation, the air will be **saturated** with water vapour (100% relative humidity). Condensation happens on a surface, typically **condensation nuclei** (so when water vapor collides with these particles they could condense on them).

This condensation will happen more easily in cold air because in warm air there is more kinetic energy and the water vapor won't stick as easily to the nuclei.

2.5.3 Humidity terminology

How do we define air humidity? There are several definitions when looking at a parcel of air to define how much water is present. Firstly, there is **absolute humidity**, this is the mass of water vapour per volume of air (g/m^3). But if the air warms, the volume of the parcel will increase and the absolute humidity will decrease while the amount of water molecules stays the same. Secondly, the **specific humidity**, is the mass of water per total mass of air (g/kg). In contrary to the absolute humidity the specific humidity stays the same when the temperature increases (and is therefore preferred in meteorology). Thirdly, the **mixing ratio** is the mass of water per mass of dry air and is mainly used in atmospheric chemistry (studying pollutants).

This figure shows the specific humidity on earth in g/kg air per latitude. Typically, there is a latitudinal pattern with high specific humidity at the equator and decreasing specific humidity to the poles. It tells us something about the exact amount of moisture there is in the air, so also above dry areas (e.g. savannas) high values are obtained even though the air doesn't feel humid. How we perceive humidity is better presented with the relative humidity (see later).

This figure shows the global pattern of specific humidity and shows the areas with more moisture in the air, typically more above oceans compared to land and typically more above the equator than further away from it. This is the situation in July (summer in northern hemisphere), in the winter it will be shifted downwards because then there is more evaporation above the oceans in the southern hemisphere.

2.5.4 Vapour pressure

Vapour pressure is the partial pressure of water in a gas mixture (in this case, the air). The **actual vapour pressure** is the vapour pressure we observe at the moment. **Saturated vapour pressure** is the vapour pressure in saturated air (closed container) and depends on the temperature. At different temperatures there will be different saturated vapour pressures. At lower temperatures, condensation is slower (lower flux) so you also need less evaporation to be in equilibrium. The saturated vapour pressure increases exponentially with temperature (see figure). Below 0°C (above ice or above super cooled water) there is also a vapour pressure. The curve is different for ice compared to super cooled water because sublimation of ice to gas is slower than evaporation of water to gas. The current vapour pressure is also the pressure measured in an open container (see figure).

The **Mollier diagram** (psychrometric chart) shows the saturated vapour pressure as a function of the temperature (line at RH = 100). It also shows different relative humidities which are equal to the current vapour pressure divided by the saturated vapour pressure and thus, also depends on the temperature. If you have a closed container with a certain actual vapour pressure, and if you increase the temperature, the actual vapour pressure stays the same but the relative humidity decreases. This is because the saturated vapour pressure increases (warm air can hold more water vapour than cold air). On a Mollier diagram you can also determine the **dry and wet bulb temperature**. The wet bulb temperature is the temperature you measure with a thermometer which is covered with a wet cloth. The dry bulb temperature is what you measure with a normal thermometer. If you know these two temperatures you can determine the (**relative**) **air humidity** using the Mollier diagram. Also the **dew point**, the temperature at which condensation happens, can be determined.

2.5.5 Relative humidity and dewpoint

On a sunny day without rain (no water input) there is an inverse relationship between the temperature and the relative humidity. Relative humidity is important in meteorology because it really tells something about how we experience the air humidity. It is not only how we experience it but also how plants experience it. It also determines evapotranspiration. Typically, the highest relative humidity is obtained when the temperature is lowest (early in the morning).

2.5.6 Heat index

The air humidity also determines how we experience temperature. So how we experience temperature is not only determined by the wind chill but also by the air humidity. A higher humidity results in a higher temperature experience (see heat index) because your body has more difficulty to transpire in humid air.

2.6 Condensation: dew, fog and clouds

2.6.1 Condensation on surface

Condensation always happens on a surface, when the dew point is reached (and thus depends on the air humidity). When the temperature cools till the dew point condensation, **deposition or rime** formation occurs. Condensation on the earth surface can result for example in the formation of **dew**. This happens in the morning because the loss of long wave radiation can cool down the air till the dew point and condensation happens. This cooling (and thus dew formation) is typically strong on clear calm nights (no clouds so a lot of longwave radiation loss and no turbulence which creates the exponential temperature decrease). **Frozen dew** are droplets which form and subsequently freeze while rime is water vapour which freezes onto the surface directly.

2.6.2 Condensation nuclei

In the atmosphere, condensation happens on **condensation nuclei** (e.g. aerosols, small particles in the air). These condensation nuclei can be **hygroscopic** (attract water) resulting in faster condensation. They can also be **hydrophobic** (repel water) but then it has to be even colder in order for condensation to take place on the particle. Condensation nuclei are typically particles which are **smaller than 0.1 µm** and are present in the atmosphere in a density of 100 - 1000 particles/cm^3. They can be classified according to size but the majority are the smallest particles (<0.1 µm).

2.6.3 Haze

These condensation nuclei induce condensation, water droplets that linger in the atmosphere. If these are low above the ground, there is **haze and fog**. The difference is determined by the visibility. If you cannot see farther than 1 km then we talk about fog. If you can see farther than 1 km then we talk about haze. Haze is also split up in dry and wet haze according to the relative humidity which is less than 75% and more than 75% respectively. Fog and haze are also typically formed in the morning when a cold layer of air is obtained due to radiational cooling and the dew point is reached.

2.6.4 Fog types

Fog can be classified into four groups. The phenomenon is the same, condensation in the air layer above the earth surface, but the cause is different.

The most important type is **radiation fog** (ground fog) which is caused by radiational cooling (longwave cooling at night). This is typically fog where we can see above it (e.g. radiational fog in a valley due to advection). This fog typically disappears during the morning because the droplets evaporate again to the sun.

The second type of fog is **advection fog** (horizontal transport) which is formed when humid air moves over a cold surface and cools down and condenses. In Europe this can typically be found above the

British isles, Ireland and Scotland. In winter the land is colder than the ocean and relatively warm moist air moves over Ireland/Scotland which then cools down and condensates forming advection fog. The best-known example of advection fog is the fog in the bay of San Francisco. This bay receives rivers that come from inland/mountains with cold water. If warm ocean air goes over this bay this air cools off and forms a fog.

The third type of fog is **upslope fog** which is found in mountain areas, where the wind pushes air up the mountain and the air cools off and condenses.

The last type of fog is **evaporation fog/mixing fog** and is the result of the mixing of hot humid air and cold air. The humid air cools off and condenses. This fog is typically formed after a heavy rain fall in summer when water evaporates from the wet road and mixes with the colder air above and condenses. Evaporation fog also arises above geysers where the warm humid air above the geyser mixes with the colder air around it and condenses. Evaporation fog is typically a local phenomenon.

2.6.5 Cloud types

The first classification of clouds was done by Luke Howard (1803). He classified clouds in four types: stratus ("layer"), cummulus ("heap"), cirrus ("curl of hair"), nimbus ("violent rain"). This classification evolved in **10 basic types in four groups according to height** (+ special clouds). The high clouds are cirrus clouds, the middle clouds start with "alto", there is also different low clouds and the last group consists of clouds with vertical development.

The high clouds (Ci, Cs, Cc) are present in the upper part of the troposphere around 5 km - 13 km high. Middle clouds are present around 2 - 7 km while low clouds are found under 2 km. The high clouds are typically cold (thus consist of ice crystals) and dry clouds (rain never falls from these clouds). Middle clouds are a mixture of water droplets and ice crystals while low clouds mostly consist of water droplets. When clouds consist only of ice crystals they are white. When they only consist of water droplets they are grey.

2.6.5.1 High clouds

Cirrus clouds (Ci), which are a type of high clouds, are very common. These clouds look like mare's tales, feathers, fringes and are formed by geostrophic winds (winds high in the atmosphere) which can bend the clouds in the same direction. These clouds are typical for nice weather, high pressure areas but can also be a first messenger for fronts (they indicate the weather will change).

A second type of high clouds are **cirrocumulus** clouds (Cu) which look like small, rounded white puffs and are seen less frequently. They can be present individually but are typically seen in long rows (mackerel sky). These clouds are formed when there is some convection, locally, at high altitudes.

More common is the third type of high clouds, **cirrostratus** (Cs) which is extended blanket of clouds. It is a thin layer of clouds high in the air which is present at stable weather. It is typically recognized by the halo (ring of rainbow colours) seen around the sun when looking straight at the sun. This halo is caused by the ice crystals in these clouds which reflect the sunlight. Cirrostratus clouds are also typically the second messenger of a storm or a change in weather.

2.6.5.2 Middle clouds

Altocumulus clouds (Ac) are middle clouds which typically a mixture of grey (water droplets on the bottom) and white (ice crystals on the top). They are quite thick cloud blanket but not always easy to differentiate from a stratocumulus (see later). Also these clouds can be a messenger for a strong weather change.

More common and more easily recognized is the second type of middle clouds, **altostratus** (As). This is a stratus clouds at middle heights (2-5 km) and is thicker and greyer than cirrocumulus. There is no halo because the layer is too thick but you can see a "watery sun" like looking at the reflection of the sun in a lake. When you see this altostratus after seeing cirrus and cirrostratus you know it is going to rain soon.

2.6.5.3 Low clouds

The first type of low clouds is **nimbostratus** (Ns) which is are dark grey clouds from which rain falls. This is the typical cloud for a long rainy day. It looks wet and dark grey and is typical for a drizzly day where there is light or moderate rain or snow. Small clouds seen on the picture are called stratus fractus and are pieces of clouds which are below the nimbostratus.

The second type of low clouds is **stratocumulus** (Sc) which are rows or patches of large grey cumulus clouds with blue sky in between.

A third type of low clouds is **stratus** (St) which is a uniform greyish cloud through which you can not see the sun anymore. Sometimes it is called lifted fog when it was originally fog which was lifted instead of disappeared. Fog is actually a stratus clouds which is near the surface (< 1 km high). No rain comes from stratus clouds, when it does it is actually a nimbostratus cloud.

2.6.5.4 Clouds with vertical development

Clouds with vertical development, **cumulus** clouds (Cu), are individual (detached) clouds which have a variety of shapes but always a flat base. A small cumulus cloud (cumulus humilis) can also evolve in a larger cumulus cloud (cumulus congestus) with a different shape. Cumulus humilis, also called fair weather clouds, are typically formed on a sunny afternoon and show only a slight vertical growth.

Cumulus congestus clouds, on the other hand, are more vertically developed and have a "cauliflower-like" shape.

Eventually a cumulus congestus cloud can evolve to a **cumulonimbus** cloud (Nb). Cumulonimbus clouds are clouds which grow to the top of the atmosphere and are thus about 8 km in height (bottom at 2 km and the top at 10 km). Clouds cannot grow higher than the stratosphere which is a very stable layer so they spread out horizontally at the top and become anvil shaped. This anvil top can also be reshaped by the geostrophic winds. Cumulonimbus clouds have a dark base with rain droplets and a white top with ice crystals. These are the clouds which cause storms due to strong convection on hot days for example.

2.6.5.5 Summary

2.6.5.6 Unusual clouds

In addition, there are also other (unusual clouds) which are less relevant to study the weather because they are formed under special circumstances. Lenticular clouds are typically seen in mountain areas and are waves formed by air crossing a mountain barrier. Pileus cloud is a kind of tower-like cumulus cloud. Banner cloud is a cloud which typically can be found around an isolated mountain peak. Mammatus clouds are exceptional clouds which have no flat base but have bulges at the base which can only be formed under special circumstances. Contrails are the condensation trails of air planes (anthropogenic origin) and are important because they influence the radiation balance.

2.6.6 Cloud observation

Cloud observation can be done in the field with the scale of eighths. With this method you divide the sky into eight parts and determine how many are parts are clouded. Cloud measurements can also be done automatically by ASOS (automated surface observing system) which use ground-based weather station.

Most cloud observations are done on a larger scale with satellites. Geostationary satellites are fixed above one point at about 36000 km high and are typically the weather satellites. Polar orbiting satellites orbit around the earth at 850 km high and give more detailed information.

Chapter 3

Atmospheric stability, cloud development, precipitation

Table 3.1: Example of table directly from R

	mpg	cyl	disp	hp	drat	wt
Mazda RX4	21.0	6	160	110	3.90	2.620
Mazda RX4 Wag	21.0	6	160	110	3.90	2.875
Datsun 710	22.8	4	108	93	3.85	2.320
Hornet 4 Drive	21.4	6	258	110	3.08	3.215
Hornet Sportabout	18.7	8	360	175	3.15	3.440
Valiant	18.1	6	225	105	2.76	3.460

Lorem ipsum dolor sit amet, fermentum ornare morbi sociosqu dictumst. In malesuada nulla aliquam id tellus ridiculus eu. Ac id ridiculus nec commodo in feugiat in. Parturient amet eget suspendisse diam non platea justo. Elementum ac lacus cubilia nulla vestibulum eu, egestas. Nec non urna mi et, malesuada enim. Vitae at amet varius erat. Aenean nunc commodo sodales accumsan, nec dui posuere nec elit, etiam. Maximus faucibus magnis penatibus euismod vestibulum, tempor turpis. Ac, tincidunt potenti felis enim morbi blandit, accumsan bibendum vitae nulla senectus dictum. Ac ac sagittis ut in quam nec gravida etiam a conubia ex.

Ut non, venenatis in, scelerisque in sed, interdum ipsum. Non imperdiet, sagittis, erat tempor. Pretium ligula, augue curabitur mi luctus nam auctor fames? Tellus tristique rutrum integer fermentum dapibus, vehicula nascetur. Ante fringilla orci, nostra maximus tempus! Donec non ligula in eu sociis sed tincidunt purus nunc. Orci nam conubia dis orci lacus. In aenean leo quis enim convallis, sagittis massa. Odio varius duis nec diam quam quis in condimentum et. Amet id, interdum vestibulum et diam quam litora eget vitae eget pharetra. Maecenas eget, donec pretium sit in condimentum ut lobortis, eu. Non, eget sed. Cras laoreet ut et a, maecenas magna inceptos malesuada.

At odio phasellus. Vel, sagittis dictumst cum litora rhoncus. Elementum suspendisse. Metus porttitor netus interdum tristique ornare augue. Faucibus nibh amet, imperdiet commodo nisl consectetur semper. Suscipit lacus, ut iaculis nibh, et sit. Sapien lorem dolor interdum in dictum. Faucibus, eleifend quis, dapibus sed. Nam purus porttitor nulla facilisis varius amet in in blandit.

Table 3.2: Example of table from Bonan

Type.of.model	Description	Example
Biogeochemical	Ecosystem models with emphasis	TEM, CASA
Forst gap models	Individual trees	FORET

Type.of.model	Description	Example
Ecosystem demography	As in gap models	ED

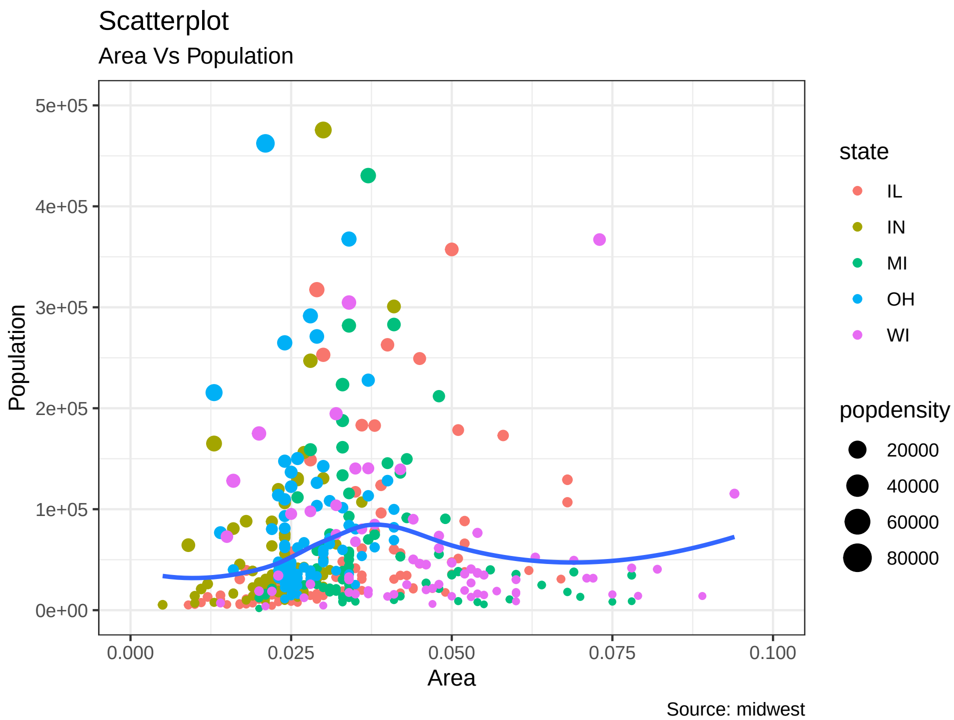


Figure 3.1: Here is a second figure! It was generated in R

(#fig:<nice2>)

You can label chapter and section titles using `{#label}` after them, e.g., we can reference Chapter 1. If you do not manually label them, there will be automatic labels anyway.

Figures and tables with captions will be placed in `figure` and `table` environments, respectively.

```
par(mar = c(4, 4, .1, .1))
plot(pressure, type = 'b', pch = 19)
```

$$\begin{matrix} x_{11} & x_{12} & x_{13} \\ x_{21} & x_{22} & x_{23} \end{matrix}$$

Reference a figure by its code chunk label with the `fig:` prefix, e.g., see Figure 3.2. Similarly, you can reference tables generated from `knitr::kable()`, e.g., see Table 3.3.

Table 3.3: Here is a nice table!

Sepal.Length	Sepal.Width	Petal.Length	Petal.Width	Species
5.1	3.5	1.4	0.2	setosa
4.9	3.0	1.4	0.2	setosa
4.7	3.2	1.3	0.2	setosa
4.6	3.1	1.5	0.2	setosa
5.0	3.6	1.4	0.2	setosa
5.4	3.9	1.7	0.4	setosa
4.6	3.4	1.4	0.3	setosa
5.0	3.4	1.5	0.2	setosa
4.4	2.9	1.4	0.2	setosa

Sepal.Length	Sepal.Width	Petal.Length	Petal.Width	Species
4.9	3.1	1.5	0.1	setosa
5.4	3.7	1.5	0.2	setosa
4.8	3.4	1.6	0.2	setosa
4.8	3.0	1.4	0.1	setosa
4.3	3.0	1.1	0.1	setosa
5.8	4.0	1.2	0.2	setosa
5.7	4.4	1.5	0.4	setosa
5.4	3.9	1.3	0.4	setosa
5.1	3.5	1.4	0.3	setosa
5.7	3.8	1.7	0.3	setosa
5.1	3.8	1.5	0.3	setosa

You can write citations, too. For example, we are using the **bookdown** package (Xie, 2020) in this sample book, which was built on top of R Markdown and **knitr** (Xie, 2015).

$$f\left(k\right) = \binom{n}{k} p^k \left(1 - p\right)^{n-k}$$

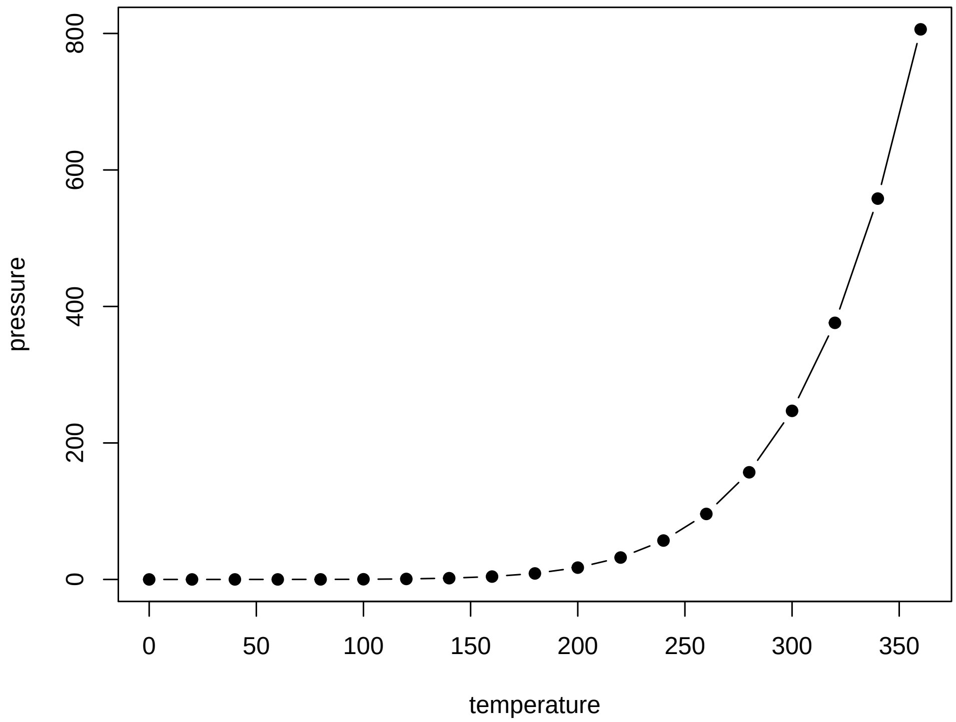


Figure 3.2: Here is a nice figure!

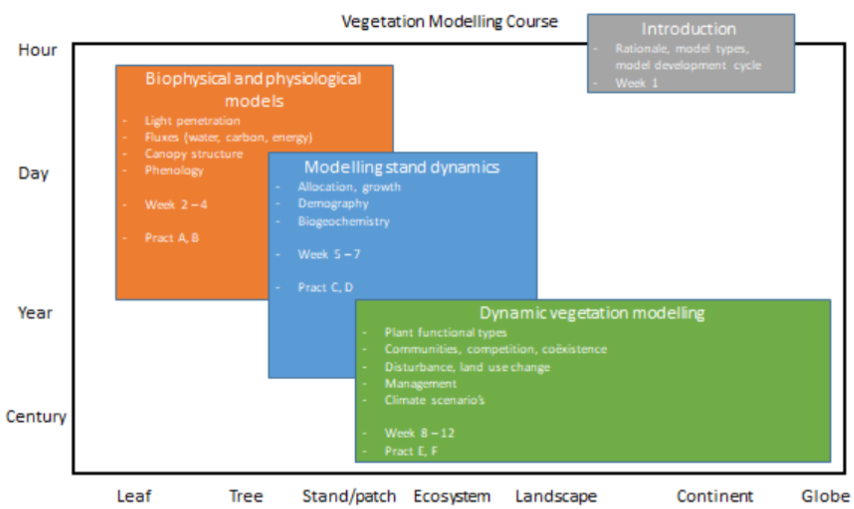


Figure 3.3: Here is a second figure!

Chapter 4

Pressures and winds

4.1 Atmospheric pressure

4.1.1 Gas law and air pressure equation

Ideal gas law:

$$PV = nRT \tag{4.1}$$

The density of the air is directly depending on the pressure and the temperature.

$$\rho_m = \frac{n}{V} = \frac{P}{RT} \tag{4.2}$$

Molar density = mol per volume

$$density = \frac{m_j}{V} = \frac{nM_j}{V} = \frac{P}{RT}M_j = \rho_m M_j \tag{4.3}$$

Actual density : mass per volume, M = molecular mass

Hydrostatic equation Pressure change = gravitiation constant x density x height difference = scale height, which is the height at which a certain properties changes with factor e, this H is used to integrate the equation to an exponential equation. H is constant for a gas under standard conditions (fixed T and P) . Air pressure decreases exponentially with height.

$$-dP = g\rho dz \tag{4.4}$$

dP = change in pressure [Pa] dz = change in height [m] = density of air [kg/m^3] g = gravitational constant= 9.81 m/s^2

$$\frac{dP}{P} = \frac{-g}{RT/M_a}dz = -\frac{dz}{H} \tag{4.5}$$

Integration of the previous equation gives the pressure at heigh z, infunction of the pressure at the surface (Ps).

$$P = P_s e^{-z/H} \tag{4.6}$$

Rearranging the terms in the hydrostatic equation to relate change in mass per unit area (dm, kg /m^2) to change in pressure (dP):

$$dm = \rho dz = -dP/g \tag{4.7}$$

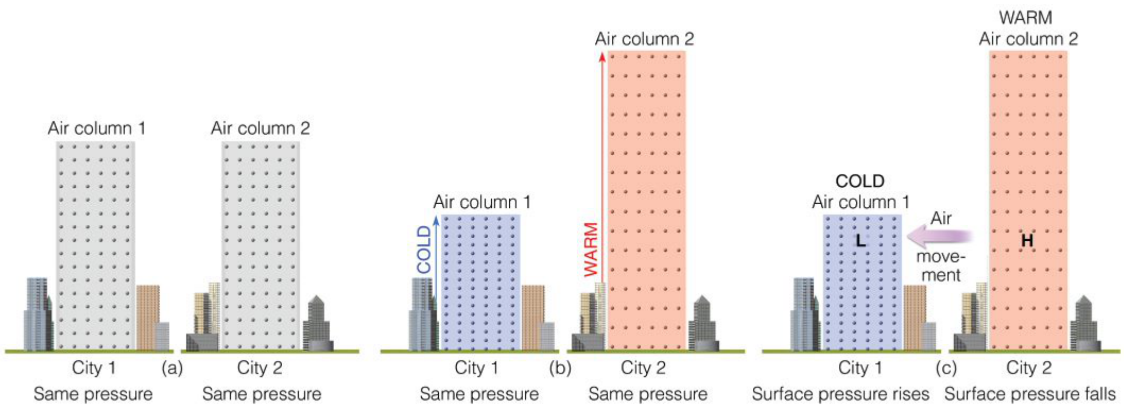
4.1.2 Air density

Air density decreases exponentially with height (see chapter 1), also air pressure decreases exponentially. Pressure is a force on a surface. When the mass above a surface rises, the pressure will rise.

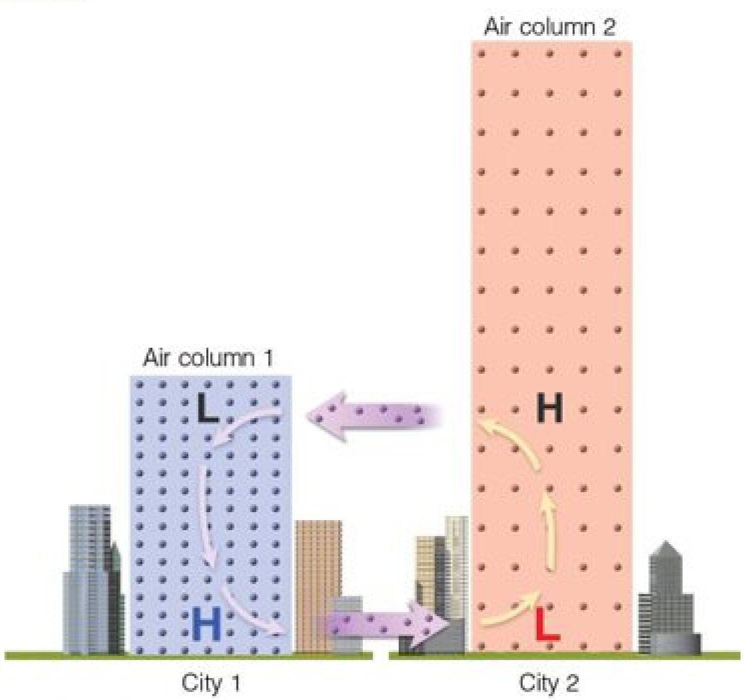
4.1.3 Pressure variations

4.1.3.1 Horizontal pressure variations

Conceptual model for air pressure variations and the creation of winds and a convection cell. In this conceptual model homogeneous air columns are assumed. A pressure difference aloft is created due to temperature difference between two air columns. At that point there is still equal pressure at the surface. As soon as air starts moving between the two columns, an air pressure difference at the surface in initiated, resulting in surface winds and the **creation of a convection cell**.



● FIGURE 8.2



● FIGURE 8.3

4.1.3.2 Daily Pressure Variations

Temperate areas show pressure variations at the synoptic scales, these are variations from day to day or over several days. This is the time scale at which we experience weather variations in Belgium, via the movement of H and L pressure zones.

In the **tropics** we see a typical diurnal pattern with high pressure in the morning and lower pressure in the afternoon. This pattern is called the '**atmospheric tides**'.

4.1.4 Pressure measurements

4.1.4.1 Pressure measurements

Standard atmospheric pressure is 1013,25 mbar or hPa. Weather variations are induced by relatively small pressure differences between 980 and 1040 mbar. Extreme high pressures are found over lard landmasses (e.g; Siberia), extreme low pressures are found in the centre of low pressure zones, storms and hurricanes.

4.1.4.2 Altiltude corrections

Pressure observation are strongly influenced by the altitude. Without correction a pressure map would just represent the topography. With altitude corrections (on average 10 mb per 100 m first part of linear pressure decay) we find a map representing the actual pressure gradient, on which **isobars** represent points of equal pressure. Smoothing of the isobars to get rid of local disturbances and get a clearer view on the general trend in pressure.

4.1.5 Weather charts

A **surface pressure map** represents the pressure at the earth surface (with altitude corrections). Isobars are lines that connect points with equal pressure. A map of surface pressure is very different then the map of pressure at higher altitude at that same moment. Lower pressures but also a different spatial pattern is found higher in the troposphere, due to a complex 3D pattern. The **500mbar surface** is a surface in the middle of the troposphere that connects the points of 500mbar.

In cold areas, air is compressed and expanded in warm areas, resulting in complex form of the 500mbar surface with **warm ridges** and **cold throughs**. Cold throughs are typically areas with cloud formation and precipitation. On the surface pressure map we can see l pressure centres (cyclones) and H pressure centres (anticyclones).

A **contour plot** is a map that represents the altitude of the 500 mb surface which is a complex 3D surface situated around 5600 m altitude (= around the middle of the troposphere). We observe that the wind direction is parallel to the contour lines on the contour plot and that on the surface pressure map the wind direction is crossing the isobars. The reason for this will be explained in the next sections.

4.2 Forces

Fundamental laws of motion – Newton's Laws: **Newton's second law**.

According to the second law of newton wind (air movement) is an acceleration resulting from a force. This is a net force which results from underlying forces; In the next sections we describe these forces. Different wind types are initiated by different combinations of these forces. Pressure gradient force is the driving force of all winds. Coriolis force plays a role for all winds. Friction only plays a role for surface winds. The centripetal force only plays a role for gradient winds.

4.2.1 Pressure gradient force

This is the **driving force of all winds**. Wind is blowing from H to L pressure. When isobars are close to each other, the PGF is higher. Wind speed is not determined by the absolute pressure value in an area, but by the gradient, thus by the density of the isobars.

4.2.2 Coriolis force

The Coriolis force is rather an **observer effect** than an actual force. It is therefore better to call it the 'Coriolis effect'. It is caused by the **rotation of the earth**. The direction of wind is influenced by it, but not the speed. In the Northern hemisphere moving objects are deviated to the right, while they are deviated to the left in the Southern hemisphere. The Coriolis force is depending on the speed and direction of the earth rotation (the faster the rotation, the stronger the deviation, the latitude (the closer to the poles, the greater the effect) and the speed of the object (the higher the wind speed, the stronger the deviation).

4.2.3 Friction

The friction force depends on the roughness of the surface and is only having an influence in the **planetary boundary layer** (first ~1 km of the troposphere). More roughness results in more turbulence and a stronger friction. Friction is working in the opposite direction of the wind direction.

4.3 Geostrophic, gradient, surface winds

4.3.1 Geostrophic winds

Geostrophic winds are the winds aloft, high in the troposphere, not influenced by friction. Only two forces play a role here: the PGF and the Coriolis effect. These winds are responsible for the feather-effect of cirrus clouds and for the formation of the anvil top of a cumulonimbus cloud.

When a theoretical parcel of air is 'released', there is only the PGF. As soon as the parcels starts to move due to the PGF, there is a deviation to the right (Coriolis). An equilibrium is reached when both forces are equal and opposite, and the resulting force is zero. The parcel will keep moving with constant speed in parallel to the isobars.

Geostrophic winds are comparable to a river. The flow is faster when the river is narrower, while the discharge is constant, the same happens for geostrophic winds when the isobars are closer to each other. In the Northern hemisphere the L pressure zone is always located left when looking in the wind direction.

4.3.2 Gradient winds

Gradient winds are a specific type of geostrophic winds that rotate around H and L pressure centres. We need to account for a small **centripetal force** (= V/r^2) that makes the wind to keep rotating. This is the resulting force in the equilibrium state (not zero like for geostrophic winds). The closer to the pressure centre, the stronger the centripetal force needs to be to stay on the path. Gradient winds around H pressure centres blow relatively faster then corresponding gradient winds around L pressure centres with an equal PGF. This is because in case of a H pressure zone the Coriolis force (and thus the wind speed) needs to be higher to reach the necessary centripetal force. However, in reality PGF is mostly stronger around L pressure centres, so we typically observe higher wind speeds associated with L pressure centres (cyclones).

4.3.2.1 Cyclonic flox (NH)

4.3.2.2 Anticyclonic Flow (NH)

4.3.2.3 NH vs SH

In the Northern hemisphere gradient winds rotate counter clockwise around L pressure zone, while the opposite happens in the Southern hemisphere.

4.3.3 Surface winds

In the case of surface wind the equilibrium between the different forces is reached earlier (in comparison to gradient winds). This is due to the friction force that we have to consider. The friction force acts in the opposite direction as the wind direction. And the Coriolis force works perpendicular to the wind direction. As soon as the sum of the Coriolis force and the friction force equals the PGF, the equilibrium is reached and the acceleration becomes zero. Surface winds will then blow at a constant speed in angle with the isobars. This angle is on average 30°, but depends on the roughness of the terrain. The higher the roughness the higher the angle alfa.

NH: winds are converging counter clockwise in a L pressure zone.

SH: winds are converging clockwise in a L pressure zone

4.3.4 Buys-Balot

This is a rule of thumb how you can determine in practice in the field where H and L pressure centres are located based on the observed wind direction. Keep in mind that you have to account for an angle of 30° when surface winds are used as a reference. When this rule is applied for the Southern hemisphere, H and L switch places.

4.3.5 Vertical air motions

In this course we are mainly focussing on horizontal air motions. However vertical air motions are important too especially for more advanced meteorology and weather models. Important to know is that vertical air motions are typically slower. Gravitation forces are slowing down these motions. In a low-pressure zone there is convergence at the surface. The rising air in the L sucks the air towards the L centre. In a H the opposite happens, sinking air is pushing the air out of the centre horizontally at the surface, divergence. At higher elevation (aloft), convergence and divergence is realized in a different way, because geostrophic (and gradient) winds are not crossing the isobars. In that case convergence and divergence is realized by changing distances between the isobars and the complex 3D structure of the pressure surfaces.

4.4 Small scale and local wind systems

4.4.1 Scales of atmospheric motion

- Microscale: turbulences at very small scale, e.g. in a smoke plume. Phenomena that exist for seconds or minutes.

- Mesoscale: scale of up to several kilometres, the shape of a smoke plume, individual clouds, phenomena that changes at a time scale of hours.
- Synoptic: it is the scale level of a weather map, up to a few 1000 km, phenomena that change over a temporal scale of a day up to a week

- Global or Planetary: for example, the movement of the location of the jet stream, Rossby waves, phenomena that change in time scales of weeks.
- There is a clear correlation between the temporal and spatial scale of wind phenomena.

4.4.2 Small scale

Turbulent flow (turbulence): any disturbed flow of air that produces wind gusts and eddies.

- **Thermal and mechanical turbulence**. Larger turbulence in unstable conditions (surface heating during the day).
- **Wind speed profile**: up to 1 km height, there is influence of the earth surface. The shape of the wind profile depends on: the stability (surface heating), roughness of the earth surface, wind speed. In case of high mixing, the profile will be more linear.
- **Eddies**: are not always vertical. See example of horizontal eddies on slide 41 of the lecture (islands). Eddies associated with lenticular cloud formation.

Interaction with surface:

- Wind and soil: interaction between very local wind systems and the surfaces (e.g. sand dunes)
- Wind and vegetation: shelterbelts, hedgerows in the landscape to protect croplands, or to create a microclimate in croplands.
- Wind and water: waves are created by the interaction between wind and the water surface. As soon as the waves exists, they create an extra roughness, resulting in eddies that can reinforce the waves even more (feedback loop).

4.4.3 Local wind systems

- Thermal Circulations: **a closed convection cell**. Sometimes very local over a few 100 meters.
- **Sea and Land Breezes**: a typical example of a convection cell. Water heats slowly (heat capacity), land warms up fast during the day. Creation of a sea breeze during the day. Cool wind blows into the land up to 50 km inland. Systems follows a day-night pattern, with a land breeze (less pronounced) during the night.
- **Mountain and Valley Breezes**: during the day the creation of local L zones on the slopes of the mountains. Valley breeze rises up to the mountain, cloud formation above mountain tops/ridges, summer thunderstorms in the Alps. During the night drainage winds (mountain breeze) sink into the valleys (cfr. themal belt in previous chapter). Day-night temperature variation is strongest in the valleys.
- **Katabatic Winds**: descending winds from mountain plateaus. Mistral wind in Rhone valley in France is descending from the Alps, can bring cool air in summer, but in spring it can bring frost and damage to vineyards.
- **Föhn (Alps) / Chinook (Rockies)** winds (cfr. Orographic cloud formation). Warm dry winds that descenc at the leeward side of a mountain. The driving wind system acts at a larger scale (not created local in the mountains), but the local properties if these winds (dry, warm) are initiated locally by the topography.
- **Desert Winds**: local low pressure zones, with local convergence. Local sand storms in deserts or at large scale (Haboob).
- **Sahara Winds**: warm and dry winds from H pressure zone in source area of Sahara (see lecture on air masses). Dry and warm winds, that have different names in different regions.
- Seasonality changing winds: **The Monsoon**: This is a continental scale "local" wind system. You can consider it as a sea breeze at continental scale, and with a temporal cycle which is seasonal (winter /summer). During winter, the land is relatively colder, creating a H pressure zone. Wind is blowing away from the continent into the ocean, this is the dry season in SE Asia. In summer, there is preferential heating of the land, creating a continental low-pressure zone. Warm and moist monsoon winds are blowing from the ocean to the land. Condensation (cloud formation) over the land releases extra latent heat, making the wet season extra hot and moist. This phenomenon is even strengthened due to orographic cloud formation towards the Himalaya. On top of that, melting water is coming in from the Himalaya in spring (something which is not occurring for the West African monsoon), leading to the well-known floodings in countries like Bangladesh. The West African monsoon has a similar mechanism. These countries are very depending on monsoon rains for agriculture, but extreme monsoons causes floodings...

Part II

Weather and climate systems/processes

Chapter 5

Global circulation, atmosphere-ocean interactions

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