Université catholique de Louvain
Faculté des Sciences
Département de physique

What climate change for the 21st century ?

A projection with LMD5-CLIO2 and its sensitivity to freshwater flux from the Greenland ice sheet.

Dissertation doctorale présentée par
Chantal Poncin
en septembre 2003
en vue de l'obtention
du grade de Docteur en Sciences

Promoteur: Pr. T. Fichefet.

Acknowledgements

Sorry for the English speaking people but this part is in French !

Tout d'abord, je tiens à remercier le Professeur André Berger qui m'a accueillie au sein de l'Institut d'Astronomie et de Géophysique G. Lemaître lors de mon mémoire de fin d'étude. Ce travail fut une ouverture vers le monde de la climatologie. Ce fut de plus l'occasion de découvrir une merveilleuse application des matières apprises au cours de mes années d'étude au sein de la Faculté des Sciences Appliquées. L'encadrement de ce mémoire par les Professeurs Thierry Fichefet et Jean-Pascal van Ypersele m'a encouragé à prolonger l'aventure par une thèse de doctorat.

Le suivi de cette thèse a été assuré par le Professeur Thierry Fichefet. Je le remercie pour le temps qu'il y a consacré, pour sa disponibilité, et pour la liberté qu'il a su m'octroyer.

J'ai également apprécié les rencontres au sein de GASTON, groupe français de modélisation du système climatique. Que tous les scientifiques que j'y ai rencontrés soient remerciés pour les discussions conviviales qui n'éludaient pas les problèmes rencontrés par les différentes équipes. On se rend alors mieux compte qu'on n'est pas le seul à être parfois surpris et démuni face à ces modèles complexes. Je remercie particulièrement le Professeur Hervé LeTreut qui a participé au jury de cette thèse.

Grâce à la collaboration du Professeur Philippe Huybrecht et de Ives Janssens de la VUB, le couplage de notre modèle de circulation générale et leur modèle de la calotte de glace groenlandaise a pu être réalisé. Je remercie de plus le Professeur Philippe Huybrecht pour sa participation à mon jury et pour ses commentaires constructifs.

Le manuscrit préliminaire de cette thèse a été relu par plusieurs personnes, en particulier les membres du jury les Professeurs Philippe Huybrechts, Hervé LeTreut, Jean-Pierre Antoine, Guy Schayes, Jean-Pascal van Ypersele et André Berger, je les remercie pour leurs conseils avisés. J'ai aussi apprécié l'effort consenti par ma cousine Maryline Cuvelier pour la vérification de mon anglais malgré l'aspect technique qui en aurait découragé plus d'un. J'espère ne pas l'avoir dégoûtée de la climatologie.

Cette thèse n'aurait pas pu être menée à bien sans une importante aide logistique qui a été mise à ma disposition par l'Université catholique de Louvain avec le soutien du FNRS. Le centre de Calcul Intensif et de Stockage de Masse qui a ainsi été développé était un outil indispensable pour mener à bien cette recherche. Mes remerciements vont plus particulièrement à Bernard Van Renterghem informaticien au Calcul Intensif, Jean Castel-Branco ancien informaticien de l'Institut d'Astronomie et de Géophysique G. Lemaître qui ont tous deux guidé mes premiers pas dans le monde du Calcul Intensif. Je remercie également Jan Dormaels et Raoul De Guchteneere qui ont résolu mes problèmes quotidiens en la matière. Merci aux secrétaires pour tout le travail qu'elles fournissent dans les coulisses de l'Institut.

Un tel travail de longue haleine ne peut être réalisé sans un environnement humain chaleureux et stimulant. Je remercie tant les anciens que les nouveaux doctorants de l'Institut que j'ai cotoyés durant ces années et qui ont su allier fous rires et chaleur humaine à la discussion scientifique motivante. J'ai également apprécié le contact des étudiants que j'ai encadrés lors des mes activités didactiques, ils m'ont fourni un bol d'air indispensable pour relativiser les déceptions qui ne manquent pas dans la recherche.

Finalement, merci Anne avec qui j'ai partagé ces années au bureau et qui m'a enrichie par sa personnalité et sa franche amitié. Bisous à ses proches. J'ai apprécié également le soutien de toute ma famille avec en tête ma maman sans qui cette fin de thèse jumelée aux études et à l'éducation de deux enfants n'aurait pas été possible. Bravo d'ailleurs à mes deux loustiques Jérôme et Marielle qui ont de l'énergie à revendre et qui sont toujours prêts pour un fou rire réparateur, la vie paraît tellement belle quand on la regarde avec leurs yeux. Et j'en ai encore oublié un ... merci Damien !

Abstract

The Earth's climate is changing. This is a conclusion of the Third Assessment Report published by the International Panel on Climate Change. The increasing concentrations of greenhouse gases and sulphate aerosols alter the energy path through the atmosphere. In the future, a likely consequence of the induced global warming is an increased melting of the Greenland ice sheet. This could lead to abrupt climatic modifications associated with the collapse of the thermohaline circulation in the North Atlantic.

To investigate this issue, a coupled model of the climate system has been developed. Most components of this system are taken into account by this model. LMD5, an atmospheric general circulation model originating from the Laboratoire de Météorologie Dynamique in Paris, simulates the atmosphere. CLIO2, an oceanic general circulation model set up at the Institut d'Astronomie et de Géophysique G. Lemaître at the UCL, accounts for the ocean and the sea ice. Finally, GISM is a Greenland ice-sheet model developed at the Vrije Universiteit Brussel (VUB). The atmospheric and oceanic components had already been coupled (*Grenier* [1997]). A new version has been elaborated during this thesis to enable long-term realistic simulations. To restrain the initial drift, adjustements have been made in atmospheric, oceanic and sea-ice parameters in collaboration with the research teams that set up the models. Other modifications have been performed in the framework of climate change experiments to separately handle greenhouse gases and sulphate aerosols. Finally, LMD5-CLIO2 has been coupled to GISM in collaboration with VUB researchers.

Validation of LMD5-CLIO2 implies a 150-year long control simulation under constant 1970 forcings. The validation is twofold: on the one hand, the LMD5-CLIO2 results have been contrasted with observational estimates and on the other hand, these have been compared to other coupled models results. This leads to the conclusions that LMD5-CLIO2 simulates relatively well the present-day climate and that it performs as well as many other coupled models.

Two climate change experiments have been carried out using the IPCC SRES B2 scenario. The first one deals with the impact of greenhouse gases and sulphate aerosols on the 21st century climate. The globally averaged warming reaches 2.4°C at the end of the 21st century compared to 1970. The hydrological cycle is amplified: precipitation increases by 3.3 % over the same period. This corroborates the projections of other coupled models. Consequently, it improves our confidence in the LMD5-CLIO2 simulations. The second climate change experiment performed with LMD5-CLIO2-GISM investigates the impact of the Greenland ice-sheet freshwater flux on the North Atlantic thermohaline circulation. To our knowledge, it is the first time that such a complex coupled model is used to analyse

this issue. In this simulation, deep convection in the North Atlantic ocean shuts down at the end of the 21st century, which is an unusual result. The strong and abrupt reduction of the intensity of the thermohaline circulation implies significant mainly local modifications of sea ice and of the atmosphere.

Cautions should be exercised to this projection. Firstly, LMD5-CLIO2 has deficiencies which probably influence the results. Secondly, modifications of the initial conditions could lead to another projection: such spectacular event not occurring or differences as to when it happens. Consequently, other ensemble simulations should been performed. Thirdly, the current simulation is too short to know if the reduction is temporary or not. Longer experiments are therefore required.

Contents

Introduction

The 1988 Toronto Conference states that "*humanity is conducting an unintended, uncontrolled, globally pervasive experiment whose ultimate consequences could be second only to a global nuclear war*". This assertion concerns the pollution of the atmosphere and especially the emission of greenhouse gases. Each year, billions tons of carbon dioxide are released into the atmosphere because of Man's exploitation of the world natural resources, primarily fossil fuels. In addition, other greenhouse gases, notably methane, nitrous oxide, and chloro-fluoro-carbons, are also being released as a result of human activities. As a consequence, the atmospheric concentration of these gases is increasing, with large implications for the Earth's climate. Shortly before the Toronto conference, the Intergovernmental Panel on Climate Change (IPCC) was jointly established by the World Meteorological Organization (WMO) and the United Nations Environment Programme (UNEP). Since its creation, the IPCC has produced a series of Assessment Reports, Special Reports, Technical Papers, and Summaries for Policymakers which give the scientific point of view on various aspects of climate change. In its Third Assessment Report, the IPCC notes that "*There is new and stronger evidence that most of the warming observed over the last 50 years is attributable to human activities*". This implies that the climate has already changed and that those modifications are due to human activities.

How do these greenhouse gases influence the climate ? The Earth's climate system receives energy from the Sun and emits energy to space in such a way that the incoming energy flux is balanced on average by the outgoing one. Most of the terrestrial radiation is absorbed in the atmosphere by the greenhouse gases, and part of this energy is emitted back to the Earth's surface, contributing to its warming. This phenomenon is usually called "the greenhouse effect". As the greenhouse-gas concentration increases, more energy is trapped in the atmosphere enhancing the greenhouse effect. This tends to warm the Earth's climate. In contrast to greenhouse gases, tropospheric aerosols are believed to cool the Earth's climate. These small airborne particles absorb or reflect solar radiation, and also affect cloud cover as well as cloud properties. The sulphate aerosol loading has increased as a consequence of fossil fuel combustion by Man. Though not sufficiently to compensate the radiative forcing of the greenhouse gases, this restrains the climate warming. According to the IPCC, combination of the greenhouse-gas and sulphate-aerosol effects seems to explain most of the evolution of the climate over the last 50 years.

In addition, *Houghton et al.* [2001] asserts that "*Emissions of greenhouse gases and aerosols due to human activities continue to alter the atmosphere in ways that are expected to affect the climate (...) Human influences will continue to change atmospheric composition throughout the 21st century. (...) Global average temperature and sea level* *are projected to rise under all IPCC SRES scenarios.*" Furthermore, the socio-economic stakes are large, because it is thought that all human activities would be directly or indirectly affected by any warming of the atmosphere by a few degrees and any change in the hydrological cycle. Consequently, climate-change issues claim for an integrated framework. An integrated view of climate change should consider the dynamics of the complete cycle of interlinked causes and effects across all sectors concerned. Population, economy, technology, and governance are the driving forces of greenhouse-gas and aerosol emissions. Through their implication on the atmospheric concentrations, they induce changes in the physical climate system. Climate change has human and natural impacts, which feedback on the socio-economic driving forces. All the parts of the climate change problem interact mutually. The Third Assessment Report of the IPCC tries to provide such an integrated point of view. This is of course far beyond the scope of this thesis, in which we only focus on the modifications of the Earth's climate under increasing greenhouse-gas concentrations and changing sulphate-aerosol loading. No attempt is made to reproduce the past climate, and this study is restricted to the 21st century.

The 21st century climate has already been investigated by other modeling teams. Nevertheless, they are several reasons for which the 21st century climate projections are highly uncertain. Firstly, future emissions of greenhouse gases and aerosols are difficult to predict, partly because they depend on a large number of human activities. At best, several scenarios may be examined. That is what is done by the IPCC. Its Special Report on Emission Scenarios (SRES, *Nakićenovic et al.* [2000]) provides 40 SRES scenarios gathered into 4 storylines. Secondly, the numerical models are not perfect partly because the climate system is so complex that many of its mechanisms are still far from being well understood. Nonetheless, climate modeling plays a leading role to quantify the response of the climate system to these changes in the atmospheric composition. Indeed, it is essential to account for all the complex interactions between the five components of the climate system (i.e., the atmosphere, the hydrosphere, the cryosphere, the lithosphere, and the biosphere). It is not possible to do this reliably using empirical or statistical models because of the complexity of the system, and because the possible outcomes may go well beyond any conditions ever experienced previously. Instead, the response must be found using numerical models of the climate system based upon sound physical principles. The most comprehensive global climate models are coupled atmosphere-ocean general circulation model (AOGCMs). Despite their ability to reproduce some of the features of today's climate, a significant number of shortcomings remain. Consequently, each model produces a somewhat different projection. It is commonly assumed that a multi-model approach should produce a better estimate of the climate change of the real system than the result from any particular model. Furthermore, intercomparison programs attempt to link the simulated biases to the model formulation. In this context, the overall objective of this thesis is to contribute to the international research effort leading to an improved understanding of the climate system and to a better assessment of the impact of human activities on the global climate. Consequently, an AOGCM, called LMD5-CLIO2 (*Grenier* [1997]), is improved and validated under present-day conditions. It is then used to simulate the 21st century climate using the IPCC SRES B2 scenario for greenhouse gases and sulphate aerosols. This scenario describes a world in which the emphasis is on local solutions to economic, social and environmental sustainability. It is a world with continuously increasing global population, intermediate levels of economic development, and diverse technological change. While this scenario is oriented toward environmental protection and social equity, it focuses on local and regional levels. It has been chosen because, as only one scenario is used in this thesis, we opt for a mid-range one.

A likely consequence of global warming is increased melting from the Greenland ice sheet, resulting in a larger freshwater flux into the surrounding ocean, a shrinking ice volume, and a positive contribution to sea-level change. Over the past decade, various modeling studies were conducted to quantify these volume changes under specified climate scenarios for the 21st century and beyond (see *Church et al.* [2001] for a review). Conversely, the enhanced Greenland ice-sheet freshwater flux into the ocean can alter the oceanic circulation. There is consensus today that the ocean thermohaline circulation is to a large extent driven from the high-latitude North Atlantic through the production of North Atlantic Deep Water (NADW). Sinking of surface water in the Greenland-Iceland-Norwegian Seas and in the Labrador Sea initiates an overturning circulation cell on the meridional plane. The northward transport of upper ocean warm water is balanced by a deep return flow of cold water, imposing a strong northward heat flux in both the North and South Atlantic Oceans. Due to this transport, the northern North Atlantic is about $4°C$ warmer than the Pacific at similar latitudes. Changes in NADW circulation therefore have the potential to cause significant climate change over the North Atlantic region. The pioneering work by *Stommel* [1961] and other more recent studies (e.g., *Rahmstorf* [1995]) suggest that the thermohaline circulation is a non-linear system which is highly sensitive to changes in freshwater forcing. It may collapse if a certain threshold is exceeded and can show hysteresis behaviour. There is also growing evidence that some of the past climate shifts recorded in proxies were associated with changes in NADW flow (e.g., *Stocker* [2000]). The impact of the Greenland ice sheet on the 21st century climate is investigated in this thesis using LMD5-CLIO2 coupled to a comprehensive model of the Greenland ice sheet (GISM). To our knowledge, it is the first simulation performed with such an interactive coupled model including the atmosphere, the ocean, the sea ice, and the Greenland ice sheet.

The aim of the first chapter is to provide the reader with useful information on the climate system and its recent evolution. The five components of the climate system are defined. Then, their main characteristics and their composition are briefly depicted. These five components are not unconnected boxes, they interact with each other because various chemical species and energy circulate between them. The second part of this chapter is thus devoted to the description of the main cycles in the climate system. The energy path implies various form of energy: solar radiation, longwave radiation, latent heat, sensible heat, kinetic energy, and potential gravity energy. The second cycle described is the hydrological cycle. It is one of the most important cycle in the climate system. Indeed, it is deeply related to the energy cycle; e.g., water is related to latent heat thought phase transition, water vapour is also a very efficient greenhouse gas which affects the infrared radiative transfer in the atmosphere, ... The next section depicts the carbon cycle. Carbon dioxide also plays a leading role in the climate system; e.g. it contributes to the greenhouse effect, it interacts with the biosphere though photosynthesis and respiration. Finally, other greenhouse gases and aerosols are involved in the atmospheric chemistry and modify the Earth's radiative budget. Their sources and sinks are briefly noticed. After the explanation of these cycles, the recent climate evolution is discussed and climate change is discriminated from the natural climate variability. Detection and attribution studies have been performed under the framework of the IPCC. Using AOGCMs, the scientific community came to the conclusion that the last 50 year temperature evolution could be reproduced by the models only if both the natural and anthropogenic forcings were taken into account.

Chapter 2 portrays the tool used in the present work, an AOGCM coupled to a Greenland ice-sheet model. This chapter starts with the description of the AOGCM components. Our AOGCM, LMD5-CLIO2, is indeed made up of an atmospheric model, LMD5, originated from the Laboratoire of Météorologie Dynamique in Paris, and of a sea-ice and oceanic model, CLIO2, developed at the Institut d'Astronomie et de Géophysique G. Lemaître at the UCL. These programs are linked by a coupler which performs interpolation and orchestrates the exchanges between them. A lot of sensitivity experiments were performed to adjust the model parameters in order to reduce the initial drift of this non-flux corrected model. These experiments are not discussed here, but the selected values for parameters are given. The atmospheric model has also been adapted to take into account each greenhouse gas separately and to reckon with the direct effect of aerosols on the shortwave radiation. The second section is devoted to the Greenland ice-sheet model, GISM (provided by the Vrije Universiteit Brussel), and to its coupling with LMD5-CLIO2. GISM receives temperature and precipitation anomalies from LMD5-CLIO2 and sends back a total freshwater flux to LMD5-CLIO2. A two-way coupling is thus performed, though LMD5 does not take into account modification of the ice-sheet geometry. Then the experimental design is detailed. Three experiments have been achieved: a 150-year run (CONT) with LMD5-CLIO2 under present-day (1970) conditions, and two climate-change simulations (SRESb2 and SRESb2G) using the IPCC SRES B2 scenario for greenhouse-gas and sulphate-aerosol concentrations. The SRESb2 experiment is performed with LMD5-CLIO2 alone (the Greenland freshwater flux is kept at 1970 values). LMD5-CLIO2 is coupled with GISM to perform the SRESb2G simulation. Both start in 1970 and end in 2100. Finally, the initial drift in the CONT run is briefly depicted at the end of this chapter. As all other non-flux corrected AOGCMs, LMD5-CLIO2 is not constrained to the observed climate. Consequently, the first years in coupled mode are characterized by a climate drift.

The results of the CONT simulations are thoroughly analyzed in the third chapter. The climate simulated by LMD5-CLIO2 under present-day conditions is compared to observations. This provide a validation of the model and its main deficiencies are identified. Observational estimates are not available for all variables. In those cases, the climate simulated by LMD5-CLIO2 is compared to the results of forced experiments. Indeed, we assume that the forced models, LMD5 and CLIO2, give a reasonably good image of the real climate as they are constrained by observed forcings. The results are analyzed according to three aspects: the dynamic (motions), thermal (temperature and heat fluxes) and hydrological (humidity, salinity, and water fluxes) states. This type of analysis is applied to the atmospheric and oceanic results. Finally, a section is devoted to sea ice.

The last chapter is dedicated to the 21st century climate. The SRESb2 simulation gives a first view of the climatic change induced by transient modifications of the greenhouse-gas and sulphate-aerosol amounts. The same type of analysis as above is adopted. The thermal, hydrological, and dynamic responses of LMD5-CLIO2 to perturbed atmospheric conditions are described both in the atmosphere and in the ocean. Then, a section is devoted to the sea-ice response. The influence of the Greenland ice-sheet freshwater flux is analyzed in the second half of the chapter. Therefore, the SRESb2G and SRESb2 experiments are compared as their designs only differ by the treatment of the Greenland runoff and calving. While these two climate-change experiments lead to similar results until 2075, they diverge during the last 25 years. The NADW formation shuts down around 2087. The collapse of the thermohaline circulation in the North Atlantic Ocean has been linked to the enhanced Greenland ice-sheet freshwater discharge at the top of the ocean. The resulting decrease of the surface salinity stabilizes the water column and deep convection is reduced. In the last section, implications of this modification of the thermohaline circulation on the climate are detailed.

This thesis thus includes the validation of an AOGCM under present-day conditions owing to the 150-year long CONT simulation and two projections for the 21st century. The SRESb2 simulation corroborates the results obtained by other AOGCMs. The SRESb2G experiment is more original. Using an AOGCM coupled to a Greenland ice-sheet model, it provides us with new information on a plausible future abrupt climate modification. An abrupt reduction of the NADW formation occurs at the end of this SRESb2G simulation. Nevertheless, caution should be kept in mind: our AOGCM (as any model) has deficiencies, another simulation could lead to other results depending on initial state, and this experiment is too short to distinguish whether this reduction is temporary or not.

Chapter 1

The climate system and its recent evolution: an overview

Before we get to the focus of this thesis, this chapter presents background information on the Earth's climate system and its recent evolution. This chapter is not an exhaustive overview of this topic, but the main concepts are shortly described in order to avoid misunderstanding and to help the reader unfamiliar with these concepts.

This chapter begins with a brief description of the climate system components. It then focuses on their interactions: the cycles of energy, carbon, water, and other chemical species. Finally, the topic of this work - climate change during the 21st century - is explained and is discriminated from the natural climate variability.

1.1 The climate system components

The climate system, as currently defined by the scientific community (*Houghton et al.* [2001]), is composed of five main components (figure 1.1):

- the atmosphere, i.e., the air surrounding the Earth, where we are living in;
- the hydrosphere, i.e., the continental water, like lakes or rivers, but also the salty water in the ocean;
- the cryosphere, i.e., the land ice (the seasonal snow cover, glaciers, and ice sheets) and sea ice in polar regions;
- the lithosphere, i.e., the "solid" Earth we are walking on;
- the biosphere, i.e., the life over the continents and in the oceans.

The first questions crossing our mind are "What are these ? What are they made up? And what is their structure ?" The present section attempts to answer these questions.

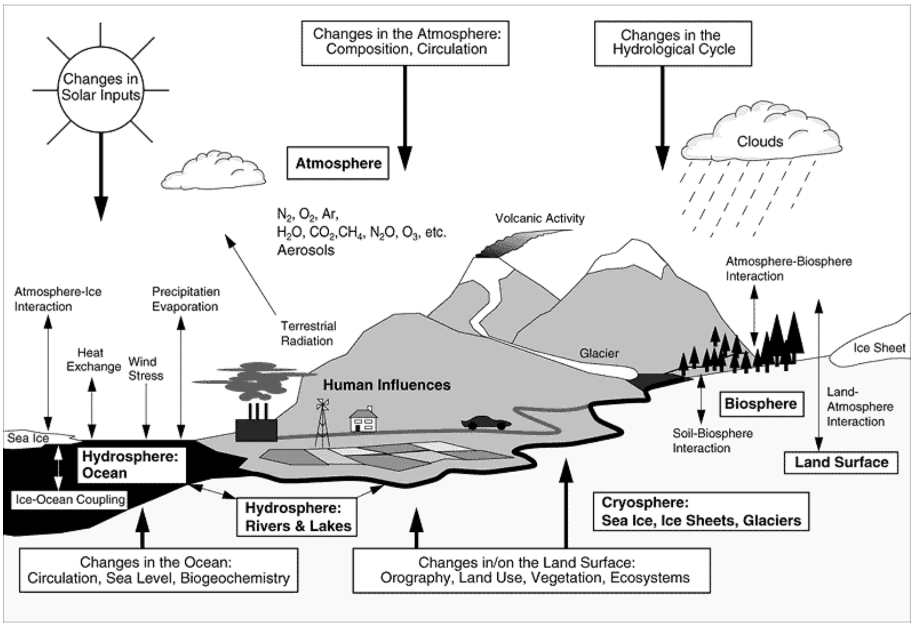

Figure 1.1: Schematic view (*Baede et al.* [2001]) of the components of the global climate system (bold), their processes and interactions (thin arrows) and some aspects that may change (bold arrows).

1.1.1 The atmosphere

As we are living in the atmosphere, we are strongly concerned by its state. Food, water, and energy supply systems are constrained to meet demand and are optimized to the current averaged climatic conditions. For example, a farmer needs to know how precipitation is distributed throughout the year and especially in summer, or a hydroelectric plant is dependent on interannual variability in rainfall and snow accumulation. In this study, we are not dealing with weather (the day-to-day variations of the atmospheric conditions) but with climate (the seasonal or annual mean state and variability). Nonetheless, the important parameters in both cases are temperature, humidity, precipitation, wind speed and direction, etc.

The atmosphere surrounding the Earth is made up of various gases, but liquid or solid particles may also be in suspension. Two main gases prevail: N_2, about 78,1% of the total atmospheric volume, and O_2, about 20,9%. Many other species are present in much smaller amount (see table 1.1 from *Hartmann* [1994]). Some of these are of utmost interest when studying the climate because they influence the energy transfer with the outer space and/or inside of the climate system: water H_2O, carbon dioxide CO_2, nitrous oxide N_2O, methane CH_4, chloro-fluoro-carbons CFCs, ozone O_3, aerosols, etc. In the section 1.2, we will follow the path of energy inside the atmosphere and see where these species play a role.

Water is present in the atmosphere in all phases: liquid or solid water in clouds or precipitation, and water vapour in moist air. The water vapour volume mixing ratio is highly variable, but it is typically of the order of 1%. The saturation level depends on

Constituent	Chemical Formula	Fraction by volume in dry air
Nitrogen	N_2	78.08%
Oxygen	O_2	20.95 %
Argon	Ar	0.936 %
Water vapour	H_2O	variable
Carbon dioxide	CO_2	353 ppmv
Neon	Ne	18.18 ppmv
Krypton	Kr	1.14 ppmv
Helium	He	5.24 ppmv
Methane	CH_4	1.72 ppmv
Xenon	Xe	87 ppbv
Ozone	O_3	variable
Nitrous oxide	N_2O	310 ppbv
Carbon monoxide	CO	120 ppbv
Hydrogen	H_2	500 ppbv
Ammonia	NH_3	100 ppbv
Nitrogen dioxide	NO_2	1 ppbv
Sulfur dioxide	SO_2	200 ppbv
Hydrogen sulfide	H_2S	200 pptv
CFC-12	CCl_2F_2	480 pptv
CFC-11	CCl_3F	280 pptv

Table 1.1: Composition of the atmosphere in 1990 (*Hartmann* [1994]).

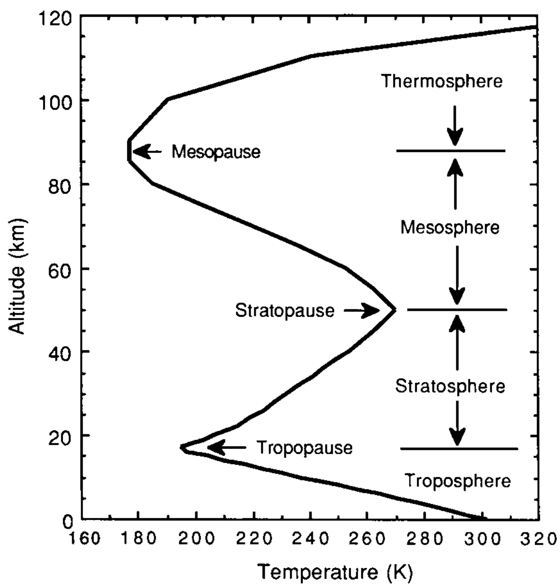

Figure 1.2: The mean thermal structure of the atmosphere. (*Hartmann* [1994]).

temperature according to Claudius-Clapeyron law:

$$\frac{dp_s}{p_s} = \left(\frac{m_v L}{RT}\right)\frac{dT}{T}, \tag{1.1}$$

where p_s is the saturation vapour pressure, $m_v = 18.016$ g mol^{-1} is the water molecular weight, $R = 8.31$ J K^{-1} mol^{-1} is the universal gas constant, and $L = 2.5 \times 10^6$ J kg^{-1} is the latent heat of vaporization. Equation 1.1 shows that the amount of water vapour in saturated air increases as the temperature increases. As $\frac{m_v L}{RT} \approx 20$ under current terrestrial conditions, a 1 % change in temperature (about 3°C) produces a 20 % change in saturation specific humidity.

This study will focus on the lowest 10 − 15 km of the atmosphere: the troposphere (see figure 1.2). In this zone, temperature decreases with altitude. The global mean tropospheric lapse rate is about 6.5 K km^{-1}. This lapse rate is mainly determined by a balance between radiative cooling and convection of heat from the surface as we will see later on. The troposphere is clearly the densest part of the atmosphere, as about half the atmospheric mass is located in the lowest 5 km. Vertical and horizontal motions are important in this region. Higher, in the stratosphere, temperature increases with height up to about 50 km. This is mainly due to absorption of ultra-violet radiation by ozone. These stable conditions inhibit vertical motions. Therefore, the troposphere and the low stratosphere are the atmospheric regions which are most directly linked to the surface climatic conditions (their equilibrium time according to *Henderson-Sellers and McGuffie* [1987] is ∼ 11 days). Then, temperature decreases again in the mesosphere up to about 90 km. The upper part of the atmosphere is the thermosphere where temperature rapidly increases due to absorption of ultraviolet radiation from the Sun. As we are mainly interested in the surface conditions, we will focus on the troposphere.

WATER RESERVOIR	DEPTH IF SPREAD OVER THE ENTIRE SURFACE OF EARTH (M)	PERCENT OF TOTAL
Oceans	2650	97
Ice caps and glaciers	60	2.2
Ground water	20	0.7
Lakes and streams	0.35	0.013
Soil moisture	0.12	0.0044
Atmosphere	0.025	0.0009
Total	2730	100

Table 1.2: Water on Earth (*Hartmann* [1994]).

1.1.2 The hydrosphere

Water is essential to life on our planet. All phases coexist: water vapour in the atmosphere, liquid water (the hydrosphere), and solid water (the cryosphere). This section depicts the hydrosphere, and the next one the cryosphere. The water distribution between various reservoirs is shown in table 1.2[1]. The hydrosphere consists of two components: the oceans and the continental water (ground water, lakes, streams, and soil moisture).

1.1.2.1 Ocean

The ocean is made up of salty water: $\sim 1310 \times 10^6$ km^3 of water with salt. The mean salinity[2] is around 34 psu = 34 g/kg, but it ranges from $\sim$ 10 psu in the Baltic Sea to $\sim 40\,psu$ in the Red Sea. The ocean covers approximatively 70 % of the Earth's surface and its depth varies from a few meters along the coast to more than 5500 m in the deep ocean.

Due to its large heat capacity (for dry air $c_v = 717$ J K^{-1} kg^{-1}, for liquid water $c_v = 4218$ J K^{-1} kg^{-1}), it is able to store a huge amount of energy and may prevent rapid temperature fluctuations, or delay temperature change. Moreover, its motions play an important role in redistributing the energy at the surface of the Earth. For example, the Irish and Icelandic climates are strongly influenced by the Gulf Stream: palm trees grow on the south-west Irish coast and the Reykjavik port is nearly never locked by sea ice in spite of its high latitude. The oceanic motions are forced by the atmospheric winds (the wind-driven circulation) and by salinity and temperature contrasts (the thermohaline circulation). Therefore, we are interested in the seasonal and annual mean and variability of temperature, salinity, and motion.

Temperature generally decreases with depth, except in polar regions. Surface processes, associated to wind or stability, induce turbulence in a thin surface mixed layer, where

[1] While the numbers are uncertain (mainly for the small reservoirs), the orders of magnitude are correct.

[2] Salinity is the mass of all dissolved salts in 1 kg of seawater. A new definition has been given in 1978 which is related to seawater conductivity: the Pracical Salinity Units (psu). In most cases, where the highest accuracy is not required, old and new definitions may be used together indisciminately (*Baretta-Bekker et al.* [1998]).

temperature and salinity are almost independent of depth. Water in this zone is strongly connected to surface conditions. Then, most of temperature or salinity changes occur in the first kilometer in the thermocline or the halocline. If surface conditions would change, the upper and intermediate waters would need $\sim 7-8$ years to reach a new equilibrium (*Henderson-Sellers and McGuffie* [1987]). Locally in polar regions, the very cold surface temperature and salinity processes implying sea ice (see below) trigger deep convection, which brings deep water in contact with the surface. Deep convection being local, the time-scale is much longer to significantly modify the deep-ocean characteristics ($\sim$ 300 years according to *Henderson-Sellers and McGuffie* [1987]).

1.1.2.2 Continental water

When liquid water falls on the ground (due to precipitation), water can be stored in the ground, percolate towards underground water, run off towards rivers or lakes, or evaporate back to the atmosphere. These phenomena are of key importance for mankind as we are land-dwelling creatures. About 80% of our food grows over land, and water supplies set limits to agriculture. Furthermore, flooding or drought produce great damage in some regions. Continental water reservoirs store about 9.8×10^6 km^3 of water. While this amount is negligible by comparison to the amount of water in the ocean, water fluxes with the other reservoirs are not negligible (see section 1.3).

1.1.3 The cryosphere

1.1.3.1 Continental snow or ice

Rainfall turns sometimes into snowfall according to temperature. When snow falls over continents, it strongly modifies the surface albedo. Indeed, snow (or ice) is generally a much more effective reflector of solar radiation than the underlying surface. Consequently, when snow (or ice) cloaks the surface, the solar energy available at the surface is reduced. This hinders snow or ice melting and fosters cold temperature. Moreover, new precipitation change into snowfall and increase the snow cover. This positive feedback is called the "snow- or ice-albedo feedback". As a consequence, the most important factor for climate is not the amount of water stored in the cryosphere, but rather the surface area that it covers.

Seasonal snow covers about 50×10^6 km^2 (*Hartmann* [1994]). In cold conditions, snow accumulates as it does not thaw in summer. This process produced, after tens of thousands of years, the Antarctic and Greenland ice sheets. These constitute a big freshwater reservoir ($\sim 32.7 \times 10^6$ km^3 of water): if the Antarctic ice sheet ($\sim 25 \times 10^6$ km^3) would melt away, the sea level would rise by $\sim$ 80 m, and if the Greenland ice sheet would melt away, the sea level would rise by $\sim$ 7 m. Smaller glaciers are also found in mountains, they store $\sim 0.3 \times 10^6$ km^3 of water. This perennial ice overlays about 16×10^6 km^2. Examining these numbers, while ice sheets form a huge water reservoir, snow-covered land exceeds ice-covered land. Furthermore, while snow has an equilibrium time of a few hours or days, ice sheets react much more slowly, needing $\sim$ 3000 years to reach equilibrium (*Henderson-Sellers and McGuffie* [1987]). As a consequence, snow cover is as important as ice cover for climate studies.

Water is also stored as frozen water in the soil: the permafrost. This does not significantly modify the surface albedo, but the permafrost stores $\sim 0.4 \times 10^6$ km^3 of water.

1.1.3.2 Sea ice

When the sea-surface temperature goes down to the freezing point (which depends on water salinity), water freezes creating sea ice. Convection makes things more complex. At typical ocean salinities, density increases with decreasing temperature. Therefore, when the sea surface cools, the cooling triggers convection bringing warmer deep water towards the surface. Consequently, the whole water column must reach the freezing point before sea ice forms. More precisely, as salinity also influences density, only the upper layer above the pycnocline must reach the freezing point. Then, further cooling produces ice crystals which go up to the surface as water ice is lighter than liquid water. This is known as frazil ice. Frazil ice then agglomerates into "pancakes" under the action of wind, wave, etc. Finally, pancake ice amalgamates into larger blocks of ice, called ice floes, through welding and rafting. These are the basic constituents of the ice pack. They thicken by ice accretion at their base of by transformation of snow into ice at their top. The atmosphere or the ocean can bring heat into the ice pack and induce melting. Wind and ocean currents both shear sea-ice opening or closing open-water areas (called leads).

Sea ice is found in the Arctic Ocean and around Antarctica. Most of it is seasonal. Indeed, the sea-ice cover ranges seasonally between $\sim 3 \times 10^6$ km^2 and $\sim 18 \times 10^6$ km^2 in the Southern Hemisphere, and between $\sim 8 \times 10^6$ km^2 and $\sim 15 \times 10^6$ km^2 in the Northern Hemisphere.

If sea ice would melt away, this would not rise the sea level as sea ice is already floating over the ocean. Nonetheless, sea ice plays an important role in the climate system as it significantly modifies atmosphere-ocean exchanges of heat, water, and momentum.

Sea ice alters the sensible, latent, and solar heat fluxes at the atmosphere-ocean interface. Due to its low thermal conductivity, sea ice isolates the ocean from very cold atmospheric temperatures, reducing turbulent oceanic heat release. This enables the underlying water to remain at the freezing point while surface air temperature goes down to much lower values. Sea-ice albedo being much higher than open-water albedo, when sea ice starts melting, the available solar energy increases and implies further melting (sea ice - albedo feedback). Furthermore, the latent heat released (absorbed) during ice production (destruction) delays the occurrence of air temperature minima (maxima), and therefore strongly influences the seasonal cycle of the surface air temperature.

Interactions between sea ice and the freshwater fluxes are also of utmost importance in the climate system. Sea ice intercepts most of the snow falling during winter, preventing it from immediately contributing to the ocean freshwater balance. Furthermore, during the generation of ice crystals, most of the salt contained in sea water is expelled out of the ice lattice. So, when sea ice forms, salt is released into the ocean increasing water density. This triggers convection, contributing to the thermohaline circulation. On the contrary, when sea ice melts, freshwater is released at the oceanic surface, and a layer of light water can overlap the oceanic column.

The ice pack also hinders the free exchange of momentum between the atmosphere and the ocean. On one hand, wind and ocean currents drive ice motions and shear it creating leads. These open-water areas play a leading role in the atmosphere-ocean exchanges in polar regions. On the other hand, sea ice prevents wind-induced mixing of the oceanic surface. Moreover, the coupling of the ice drift with the ice formation and ablation produces a net annual poleward transport of heat in the atmosphere and of salt in the ocean.

1.1.4 The lithosphere

The solid Earth being the bottom boundary of the atmosphere, the ocean, or the continental cryosphere, it clearly interacts with them.

The surface topography influences the surface climatic conditions in the atmosphere (it is common knowledge that it is colder at the top of a mountain than at its base). It also influences the atmospheric motion as the air must skirt round or rise over the mountains, which produces gravity waves. The slope and the composition of the ground also modify its heat capacity, the way water penetrates into it, and the vegetation growth.

The Earth shape also determines the bottom topography of the ocean. It produces rifts, basins, etc. which constrain oceanic currents.

Furthermore, coast lines are drawn according to the shape of the solid Earth. The arrangement of land and ocean areas plays a role in determining global climate as their physical characteristics are different: for example, land usually has higher albedo, lower heat capacity and less water available for evaporation than the ocean. At present, about 70% of the continental area is in the Northern Hemisphere. This asymmetry causes significant differences in the climates of the Earth's hemispheres.

Some energy is also released from inside the Earth (for example in Iceland, warm geysers or during volcanic events), but it is negligible in global mean by comparison to the energy provided by the Sun at the top of the atmosphere.

1.1.5 The biosphere

The Earth is the only planet in the solar system where life has developed. All living creatures (animals or vegetation) constitute the biosphere. There is an obvious relationship between biosphere and climate: the development of life has been possible thanks to the appropriate climate conditions, and conversely the Earth's climate has been influenced by the cycle of carbon, oxygen, or water through the living organisms (*Ramade* [1987]).

On the one hand, climate influences biosphere through its supplies in water and solar radiation. Indeed, biomes are based on autotrophic vegetation that performs photosynthesis to transform solar energy into chemical energy:

$$n\mathrm{CO_2} + 2\mathrm{nH_2O} + \text{solar energy} \longrightarrow \mathrm{n\,(CH_2O)} + \mathrm{nO_2} + \mathrm{nH_2O}.$$

In most cases, $n = 6$ and the reaction becomes:

$$6\mathrm{CO_2} + 12\mathrm{H_2O} + 2817\,\mathrm{kJ} \longrightarrow \mathrm{C_6H_{12}O_6} + 6\mathrm{O_2} + 6\mathrm{H_2O},$$

where $C_6H_{12}O_6$, fructose or glucose, stores chemical energy. Due to this reaction, biomes depend on precipitation, evaporation, and energy supply. We can therefore easily observe a good correlation between the map of annual mean precipitation, annual mean air surface temperature (associated to the energy supply), and the vegetation types. The annual cycle and the extremes are also very important.

On the other hand, the biosphere has influenced and is still influencing the climate. The development of life, and especially of autotrophic vegetation, has contributed to the current composition of the atmosphere by uptake of CO_2 and release of O_2. The biosphere is still exchanging these species with the atmosphere and ocean, and therefore plays a leading role in the carbon cycle (see section 1.4). The biosphere also exchanges water with its environment and modifies the surface runoff, evaporation, and precipitation (see the section 1.3). Finally, as photosynthesis needs solar energy, vegetation modifies the radiative budget of the surface.

1.2 The energy cycle

The climate system components, that we have just depicted, are not isolated; these are "open boxes" communicating with each other. One key quantity in physics is energy. It is stored and exchanged in the climate system. This section explains the cycle of energy relevant to climate studies.

1.2.1 The various forms of energy

We must remember that the climate system is a huge energy machine. Energy is stored and available in various forms:

- **kinetic energy:** this kind of energy is associated to motion and subsequently, to the winds, depression and anticyclone in the atmosphere, or water currents, subtropical gyres in the ocean, ...:
$$E_K = \frac{1}{2}\vec{U}.\vec{U},$$
where E_K is the kinetic energy per unit mass associated to a velocity $\vec{U}$;

- **internal heat:** the air, water, ice, ground, ... temperature can be seen as storage of internal energy:
$$E_I = c_v T,$$
where E_I is the internal energy per unit mass for a temperature T and a specific heat at constant volume c_v;

- **potential gravity energy:** by rising of air or water masses like in the tropical part of the Hadley cells, the air parcel gets more energy, which is given back in downward motions. This type of energy is called potential gravity energy:
$$E_P = \Phi,$$
where E_P is the potential gravity energy per unit mass of a parcel located at height z from the Earth surface and $\Phi = \int_0^z g\,dz$ is the geopotential with $g \simeq 9.81$ m.s^{-2} is the Earth gravity constant.

Energy is exchanged between all these forms. To illustrate this fact more precisely, let's write the equations for the variation of the total energy and for each type of energy (see for example *Candel* [1996]).

The first law of thermodynamics states that

> "the variation of the ***total energy*** in a closed system - which does not exchange mass with the surrounding - is equal to the heat added to the system plus the work performed by the forces on the system that are not associated to the potential energy".

It can be expressed as in equation 1.2.

$$\begin{aligned}\rho\frac{D\left(E_I+E_K+E_P\right)}{Dt} &= \varepsilon_T-\vec{\nabla}.\left(p\vec{U}\right)-\vec{\nabla}.\left(\overline{\tau}.\vec{U}\right) \\ &= \varepsilon_T-p\vec{\nabla}.\left(\vec{U}\right)-\vec{U}.\vec{\nabla}p-\overline{\tau}:\vec{\nabla}\vec{U}-\vec{U}.\left(\vec{\nabla}.\overline{\tau}\right).\end{aligned} \tag{1.2}$$

On the right hand side of this equation, ε_T is the heat added, the others terms are the work done by the pressure and the friction forces (p is the pressure, $\overline{\tau}$ is the shearing stress tensor).

By definition of the ***geopotential***,

$$\rho\frac{D\Phi}{Dt}=\rho\vec{U}.\vec{\nabla}\Phi. \tag{1.3}$$

Multiplying by velocity the Newton's second law - which links acceleration to forces -, we obtain an equation for the ***kinetic energy***:

$$\rho\frac{DE_K}{Dt}=-\rho\vec{U}.\vec{\nabla}\Phi-\vec{U}.\vec{\nabla}p-\vec{U}.\left(\vec{\nabla}.\overline{\tau}\right), \tag{1.4}$$

where $\rho\vec{\nabla}\Phi$ is the gravity force, $\frac{1}{\rho}\vec{\nabla}p$ is the pressure force, $\vec{\nabla}.\overline{\tau}$ is the friction force, the Coriolis force is perpendicular to the velocity so that it disappears.

When we combine equations 1.2, 1.3, and 1.4, we obtain an equation for the ***internal energy***:

$$\rho\frac{DE_I}{Dt}=\varepsilon_T-p\vec{\nabla}.\vec{U}+\overline{\tau}:\vec{\nabla}\vec{U}. \tag{1.5}$$

These equations clarify the role of forces.

- Part of the work done by the pressure forces is used to change the kinetic energy[3] and another part contributes to decrease the internal energy if the motion contains a divergent component[4].

[3] A pressure difference between two places tends to induce a motion from the high pressure to the low pressure zone. Therefore, velocity increases if the initial motion is in the same direction as the pressure gradient. On the contrary, if the initial motion is in the opposite direction, velocity is reduced.

[4] The compression of a fluid increases its temperature. Conversely, a divergent zone tends to cool. This is not true in vacuum i.e. without pressure.

- The friction forces change the kinetic energy[5] but they also increase the temperature (as the term $\overline{\tau} : \overrightarrow{\nabla}\overrightarrow{U}$ is always positive)[6].
- Some transfers are also illustrated: $\rho\overrightarrow{U}.\overrightarrow{\nabla}\Phi$ is exchanged between the potential energy and the kinetic energy[7].

The heat added (ε_T) encompasses three main heat sources or sinks: radiation, sensible heat flux, latent heat flux.

1. *Radiation.* (no mass is exchanged, and no medium is required) Radiative energy is emitted by all warm objects according to Stefan-Boltzmann's law:

$$E_{BB} = \sigma T^4, \tag{1.6}$$

   where E_{BB} is the energy emitted, per unit area, by a blackbody, $\sigma = 5.67 \times 10^{-8}$ Wm^{-2}K^{-4} is the Stefan-Boltzmann constant, and T is the temperature. For usual bodies (called grey bodies), less energy is emitted so equation 1.6 becomes $E_{GB} = \epsilon\sigma T^4$, where $\epsilon \leq 1$ is called emissivity. Moreover, the Wien's law states that the wavelength of maximum emission (λ_{max}) is inversely proportional to temperature:

$$\lambda_{max} T = 2.898 \times 10^{-3} \text{ mK}. \tag{1.7}$$

2. *Sensible heat flux.* This flux encompasses two kinds of energy transport.

   By conduction: no mass is exchanged, but a medium is required to transfer heat by collisions between atoms or molecules.

   By turbulence: parcels of different energy amounts change places, so that energy is exchanged.

3. *Latent heat release or uptake.* This occurs during all phase transitions:

$$E_L = Lm,$$

   where L is the latent heat release per unit mass during a phase transition of any species present in the climate system, m is the mass involved in the phase transition. This couples the energy cycle and the cycles of all these species. In the climate system, the main species involved in phase transition is water (evaporation, condensation and precipitation, ice formation or melting, and sublimation).

[5]Friction often thwarts the motion, reducing its velocity. Nonetheless, in some conditions, friction assists the motion.

[6]The friction of things over each other produces heat.

[7]When something goes down, its potential gravity energy is progressively used to increase its velocity.

1.2.2 The energy cycle in the climate system

After this description of the various forms of energy, new questions arise "How does energy circulate in the climate system amongst these forms?" or "Does the climate system exchange energy with the outer space ? And how ?".

The outer space may be more or less considered as a vacuum. Therefore, as sensible or latent heat fluxes need a medium, only radiation can transport energy between the Earth and the outer space. In fact, the primary source of energy for the climate system is the ***solar radiation***. The Sun provides the Earth with shortwave radiative energy according to the temperature of its photosphere ($T_{photo} \simeq 6000\,\mathrm{K}$), i.e., following equation 1.7,

$$\lambda_{max} = \frac{2.898 \times 10^{-3}}{T_{photo}} \simeq 0.5\,\mu\mathrm{m},$$

which is clearly in the visible ($0.4 - 0.75\,\mu$m). If we look at the whole spectrum, about half of the radiation of the Sun is in the visible shortwave part of the electromagnetic spectrum. The other half is mostly in the near-infrared part, with some in the ultraviolet part. Moreover, the Sun's luminosity (the total rate at which energy is released by the Sun) is

$$L_S = \sigma T_{photo}^4 4\pi r_{photo}^2 \simeq 3.9 \times 10^{26}\,\mathrm{W},$$

as energy is emitted by the Sun according to Stefan's law (1.6) at the surface of its photosphere. This energy radiates into space and reaches the Earth at a distance equal to $r_{SE} \simeq 1.496 \times 10^{11}$ m. Consequently, the solar constant, which is the amount of radiation from the Sun incident on a surface at the top of the atmosphere perpendicular to the direction of the Sun, can be determined as follow:

$$S_0 = \frac{L_S}{4\pi r_{SE}^2} \simeq 1367\,\mathrm{Wm}^{-2}.$$

Finally, the interception surface is the Earth's circle perpendicular to the beam of the solar light; while the total Earth's surface is the area of a sphere. So, the global mean solar radiation incident at the top of the atmosphere is $\frac{S_0 \pi r_T^2}{4\pi r_T^2} = \frac{S_0}{4} \simeq 342\,\mathrm{Wm}^{-2}$.

The solar radiation enters the atmosphere by its top and is partly reflected or absorbed before it reaches the Earth's surface. Figure 1.3 shows the absorption diagram of the solar radiation in the atmosphere. It can be seen that the atmosphere is mainly transparent at those wavelengths. Solar radiation is reflected by *clouds* and by small particles called *aerosols*. The remaining part is reflected or absorbed by the surface (lithosphere, hydrosphere, biosphere, or cryosphere) and warms it (see figure 1.4). The ratio of the solar radiation that is reflected back to space compared to the incident solar radiation is called the planetary albedo (α_p). Currently, it is equal to about 30 %.

The surface then emits ***longwave infrared radiation***. At a first approximation, the climate system should be at equilibrium. So, the solar radiation entering the system has to be balanced by terrestrial radiation towards space:

$$\frac{S_0}{4}(1 - \alpha_p) = \sigma T_{e,}^4$$

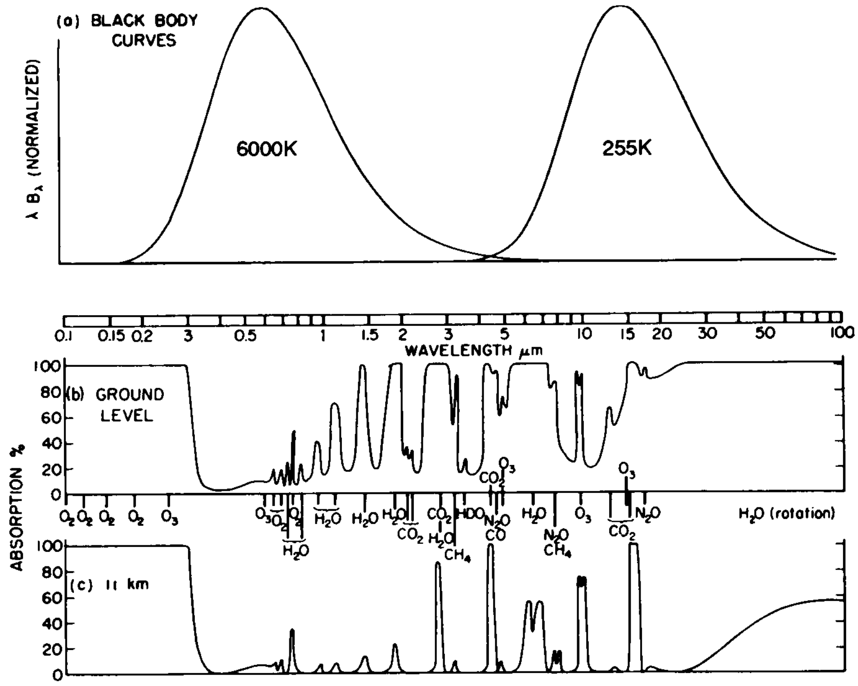

Figure 1.3: The normalized blackbody emission spectra for the Sun (6000 K) and Earth (255 K) as a function of wavelength (top). The fraction of radiation absorbed while passing from the surface to the top of the atmosphere as a function of wavelength (bottom) (*Hartmann* [1994]).

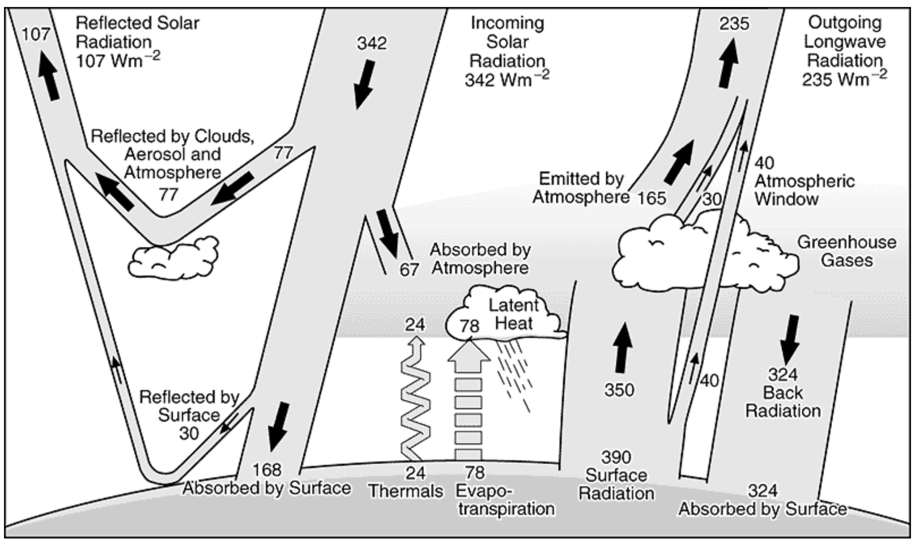

Figure 1.4: Energy (Wm^{-2}) pathway in the atmosphere (*Baede et al.* [2001]).

with $T_e \simeq -18°$C, which is the global mean effective Earth's temperature. However, at those wavelengths the atmosphere is not transparent any more: many gases present in the air (and clouds) are able to absorb this longwave radiation and emit it back to the surface and to space according to their own temperature. Therefore, part of the energy absorbed and re-emitted by the Earth surface is trapped in the atmosphere and sent back to the surface increasing its temperature (figure 1.4). This effect is called the *greenhouse effect*. It is a natural phenomena with maintains the global mean surface temperature close to $+15°C$. As shown in figure 1.3, the gases absorbing the longwave radiation are called "*greenhouse gases*": the most important are water vapour H_2O, carbon dioxide CO_2, nitrous oxide N_2O, methane CH_4, chloro-fluoro-carbons CFCs hydro-chloro-fluoro-carbons HCFCs.

The surface does not only release heat as longwave radiation but also as latent, and sensible heat. The ***sensible heat flux*** and ***latent heat flux*** are mainly due to turbulence[8] in the atmospheric boundary layer (typically about 1 kilometer deep). Turbulence is characterized by chaotic fluctuations in wind velocity producing eddies of various sizes. The parcels moving upwards have different properties than those moving downwards. Therefore, turbulent motions result in a net flux of heat and moisture between the air surrounding the surface and upper layers. Due to turbulence mixing, the heat and moisture contents are almost independent of height in the boundary layer, so that the whole boundary layer quickly responds to changes in the surface conditions. Two processes generate turbulence. One is mechanical: an important wind shear near the surface induces transformation of the mean winds into turbulent motions. The other is thermal: air parcels warmed near the surface are accelerated upwards by their buoyancy. Consequently, one might hypothesize that the sensible heat flux is proportional to the temperature difference between the surface and the upper-lying air[9], and to the mean wind speed[10]; the proportionality factor being dependent on the vertical stability[11] and surface roughness. In the same way, the latent heat flux associated to the energy required for *water* evaporation can be related to the difference of specific humidity between the surface and the air in the boundary layer, to the mean wind speed, to the vertical stability, and to the surface roughness. Nonetheless, surface evaporation can also been constrained by lack of surface moisture. Latent energy is then released inside the atmosphere during condensation. Theses processes are described more precisely in section 1.3 dealing with the hydrological cycle.

Turbulence also occurs in the oceanic boundary layer below the surface. This enables energy to enter into the first upper meters of the ocean.

Moreover, the vertical eddies sometimes penetrate further into the atmosphere leading to deep convection. This occurs when much energy is available at the surface and when the vertical profile is unstable. This phenomenon redistributes energy vertically throughout the atmosphere. The same kind of process can occur in the ocean but there is a notable difference: the atmosphere is heated at its base which contributes to unstable conditions, while the ocean is heated at its top which contributes to stable conditions. A notable

[8] Energy transport by conduction is negligible in the atmosphere, except within about a few millimeters of the surface.

[9] difference between the properties of the parcels going upwards and downwards.

[10] inducing a shearing stress.

[11] quantifying the buoyancy.

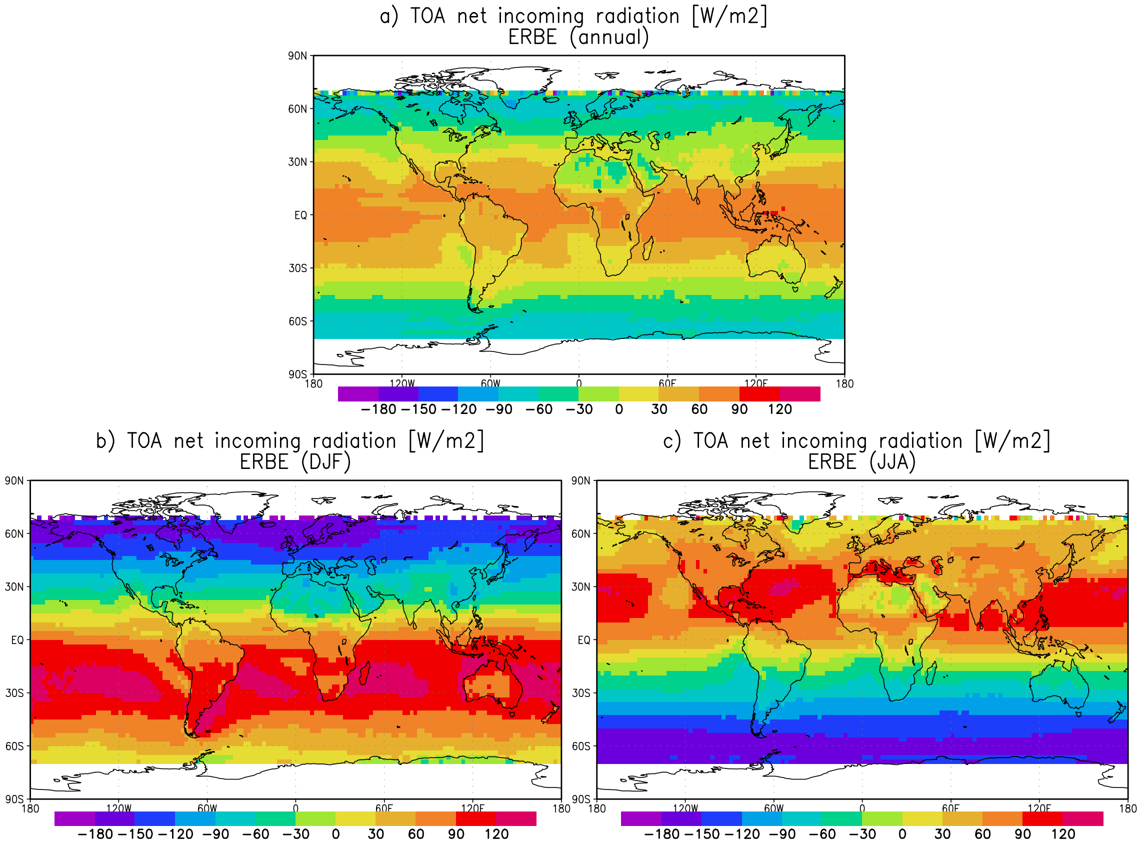

Figure 1.5: Net incoming radiation at the top of the atmosphere from ERBE data (*Barkstrom and Smith* [1986]), (a) annual mean, (b) December-January-February (DJF), (c) June-July-August (JJA).

exception is the polar regions where the ocean can be cooled at the surface inducing deep water formation.

This is a one-dimensional vertical view of the climate system while the energy sources or sinks are unequally spaced over the Earth. Looking at the top of the atmosphere, the net heat flux between the climate system and the outer space is purely radiative (as explained before). The net incoming radiation at the top of the atmosphere (figure 1.5) is measured accurately by satellites. On an annual average, polar regions lose energy, while tropical regions gain energy. This is mainly due to the meridional repartition of the incident solar radiation. A notable departure from this zonal pattern is the Sahara and other dry desert areas. These lose energy in spite of their subtropical latitude. Indeed, their high albedo reduces the net incoming solar radiation. These also emit a lot of longwave radiation according to their high surface temperature and their dry atmosphere is unable to trap it. Besides, the net heat flux at the top of the atmosphere undergoes strong seasonal fluctuations (figure 1.5). The summer hemisphere has a positive energy budget, while the winter hemisphere has a negative one. The highest values occur over the subtropical

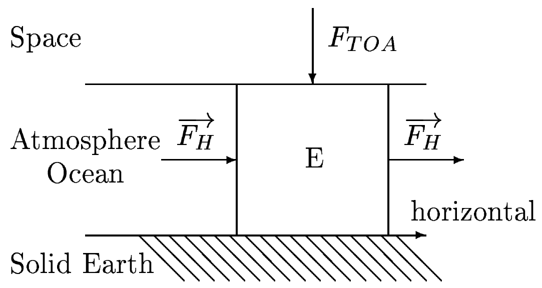

Figure 1.6: Diagram of the energy balance of a vertical column through the climate system.

oceans in the summer hemisphere, where large insolation and relatively low albedo[12] both contribute to large absorbed solar radiation.

The geographical distribution of energy sources and sinks around the Earth induces motions in the atmosphere and in the ocean to redistribute energy. These horizontal and vertical motions (***kinetic energy*** and ***potential gravity energy***) transport energy from regions of positive budget to regions of negative budget. At the same time, they transport ***internal heat*** and water vapour that can imply ***latent energy*** by phase transition. At each location defined by its latitude ϕ and longitude θ, the energy balance of a vertical column through the climate system (figure 1.6) involves the net incoming radiation (F_{TOA}) at the top of the system, the transport through the lateral boundaries and the time rate of change of its energy E:

$$\frac{\partial E}{\partial t} = F_{TOA} - \vec{\nabla}.\overrightarrow{F_H},$$

where $\vec{\nabla}.\overrightarrow{F_H}$ is the divergence of the horizontal heat flux. On an annual average, the storage term is negligible and there is a balance between the net flux at the top of the system and the horizontal transport. The zonally averaged meridional heat transport at a given latitude (F_ϕ) is obtained after integration along longitude and from the South Pole to that latitude:

$$F_\phi = \int_{-\frac{\pi}{2}}^{\phi} \left(\int_0^{\frac{\pi}{2}} F_{TOA}\, d\theta \right) d\phi.$$

This transport is the integral of the zonally averaged flux at the top of the system, starting this integration at the South Pole where the meridional heat transport must be zero due to the spherical shape of the Earth.

Heat is transported at once by the atmosphere and by the ocean (figure 1.7). They transport energy northwards in the Northern Hemisphere and southwards in the Southern Hemisphere. The oceanic transport peaks[13] at about 2 PW near 20°N and at about −1.5 PW near 10°S. The northward atmospheric heat transport is maximum near 40°N reaching about 4.5 PW. In the Southern Hemisphere, a maximum of 5 PW is transported

[12] due to sparse cloudiness and low zenithal angle.
[13] $1\ PW = 10^{15}\ W$

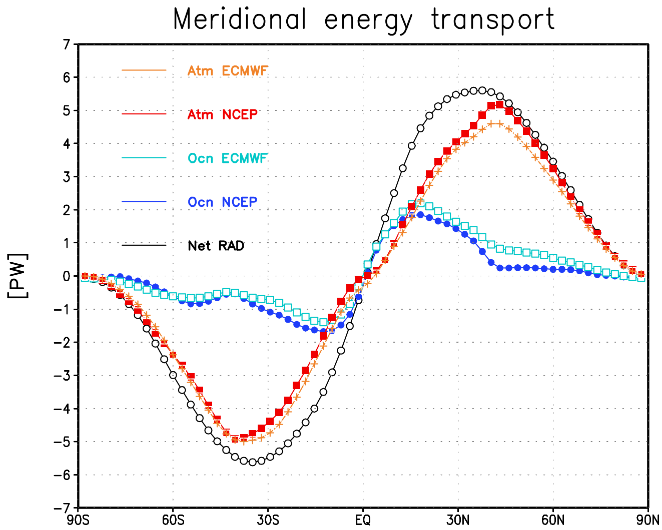

Figure 1.7: Meridional transport of energy in annual and zonal mean. The radiatively required northward heat transport (black), the oceanic transport (blue and cyan), and the atmospheric transport (red and orange). The blue and red values are obtained by the NCEP model and the cyan and orange ones by the ECMWF model (see *Trenberth and Caron* [2001]).

southwards near 35°S. In the atmosphere, this meridional heat transport is performed by the Hadley cells which export energy from the tropical regions to the mid-latitudes. Then, the mid-latitude instabilities transport it to higher latitudes. In the ocean, this transport is partly performed by the thermohaline circulation and also by the meso-scale eddies.

Furthermore, these motions transport all the species present in the atmosphere: greenhouse gases, water vapour, liquid or solid water, and aerosols. This is another interaction between the energy cycle and these species. We will now briefly describe their cycles.

1.3 The hydrological cycle

Water is of utmost importance for the Earth's climate. From section 1.1, it appears that water is present in many parts of the climate system. Furthermore, water influences the energy cycle (see section 1.2) through phase transition and its optical properties. Now, we focus on the hydrological cycle, i.e., the exchanges of water between the climate system components and between solid, liquid, and vapour phases.

At the surface, ***evaporation***, associated to the latent heat flux, introduces water vapour into the atmosphere. All the climate system components contribute to this flux (figure 1.8):

- the hydrosphere by evaporation over the oceans (the most important part) or over lakes and rivers,
- the lithosphere by evaporation of water stored in the ground,

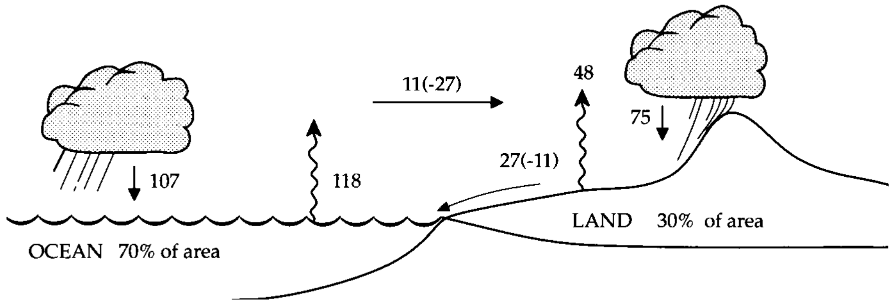

Figure 1.8: The hydrological cycle (*Hartmann* [1994]).

- the cryosphere by snow or ice sublimation,
- the biosphere by evapotranspiration of water stored in the vegetation or pumped by the roots in the ground.

Thanks to turbulence, this water vapour is mixed inside the boundary layer. Locally, convection carries it upwards into the free atmosphere. Dissipation also contributes to this upward motion but it is not as efficient. The atmospheric motions redistribute water vapour around the Earth. Locally, an air parcel can reach saturation leading to ***condensation***. Then, the latent energy stored in the water vapour is released in the atmosphere. Super-saturation is not the only criterion that triggers condensation. Condensation nuclei are needed to set off phase transition. Indeed, as demonstrated by *Houze* [1993], the air should be super-saturated by 300-400% for a drop of pure water to be created directly from vapour. Nevertheless, cloud droplets actually form by collecting water vapour onto aerosol particles. If these cloud condensation nuclei are composed of a material soluble in water, condensation is further enhanced as the saturation vapour pressure over the liquid solution is generally lower than that over a surface of pure water. In cold clouds, water droplets of pure water freeze at temperatures lower than about -35 to -40°C (see *Houze* [1993]). Nonetheless, some particles in the air, called ice nuclei, can trigger the formation of ice crystals at warmer temperature. Whereas the atmosphere has no shortage of wettable nuclei for condensation into liquid water, ice crystals do not form readily on many of the particles found in air. In fact, the ice nuclei must have a lattice structure similar to that of ice. But whenever a super-cooled drop at any temperature lower than 0°C comes into contact with an ice crystal, it immediately freezes leading to a rapid growth of the ice crystal. Aerosols are the more usual condensation nuclei. Their impact on the cloud formation and on the cloud optical properties is therefore significant.

Cloud droplets or crystals are subject to downward gravitational force. This can lead to ***precipitation*** falling out[14]. For small drops, the gravitational force is largely offset by the frictional resistance of the air, which carry it away following the winds. Therefore, cloud particles are suspended in the air unless they get to a threshold size. During their fall, water drops or ice crystals will encounter other water or ice particles; these can merge

[14]Water also condensates directly at the surface and forms dew.

and produce bigger water drops. This process is called coalescence. Precipitation can also evaporate when falling through a layer of under-saturated air.

Evaporation, moisture transport, condensation, and precipitation constitute the atmospheric part of the hydrological cycle. It manages the distribution of vapour, liquid, and solid water in the atmosphere, and alters the radiative budget of the climate system. Indeed, water vapour is a very efficient greenhouse gas, and clouds absorb longwave radiation and reflect solar radiation. The cloud optical properties, both in the shortwave and longwave parts of the electromagnetic spectrum, are influenced by the amount of liquid and/or solid water in the cloud. Reflection is also influenced by the cloud-droplet size or ice-crystal size.

Precipitation reaching the surface can either be in liquid or solid form (rainfall or snowfall). Liquid precipitation directly contributes to the hydrosphere, i.e., to ocean or continental water according to the location. Over land, the soil soaks up some water and the excess runs off, forms lakes and rivers, and goes back to the ocean[15]. Solid precipitation contributes to the cryosphere or falls over free ocean and melts. Accumulation of snow during thousands of years over Greenland and Antarctica has led to the formation of big ice sheets. Melt water from ice sheets runs off to the ocean. Ice sheets also discharge iceberg into the ocean where they drift and melt. Snow also accumulates over sea ice. Under warm summer conditions, it melts. But if the snow/ice interface is depressed below sea level by the weight of the snow pack, snow transforms into snow ice.

The net freshwater flux at the surface of the ocean (taking into account evaporation, precipitation, runoff, iceberg melting, and sea-ice formation and melting) is an important factor driving the oceanic circulation. As previously explained for the meridional heat transport (figure 1.6), the net surface freshwater flux is related to the meridional transport of salt by oceanic currents and to the meridional water transport in the atmosphere.

Over the continents, vegetation influences the water cycle. First, plants and all living organisms take up water. Indeed, these organisms contain a huge amount of water: 63% for the human, 80% for mushrooms, etc. Moreover, thanks to photosynthesis, water provides autotrophic plants with hydrogen needed in the organic matter (see *Ramade* [1987]). Secondly, vegetation releases water by transpiration and respiration. The plant roots harness ground water at depths up to 2 meters, this ground water would be "unavailable" for evaporation otherwise. Therefore, transpiration of vegetation significantly contributes to the amount of water released in the atmosphere. Thirdly, the vegetation in decomposition (humus) contained in the soil and some plants like mosses strongly modifies the retention of water in the ground, altering the amount of water available for evapotranspiration and runoff.

1.4 The carbon cycle

Carbon can be found in each component of the climate system: in the biosphere, but also in the atmosphere and the ocean. It is one of the basic elements which have led to life on the Earth. As CO_2 and CH_4 are two greenhouse gases, their atmospheric concentration has a climatic impact. Therefore, knowledge of the carbon reservoirs and of the carbon

[15] The runoff that occurs in some basins does not go to the ocean, but to in-land lakes or seas.

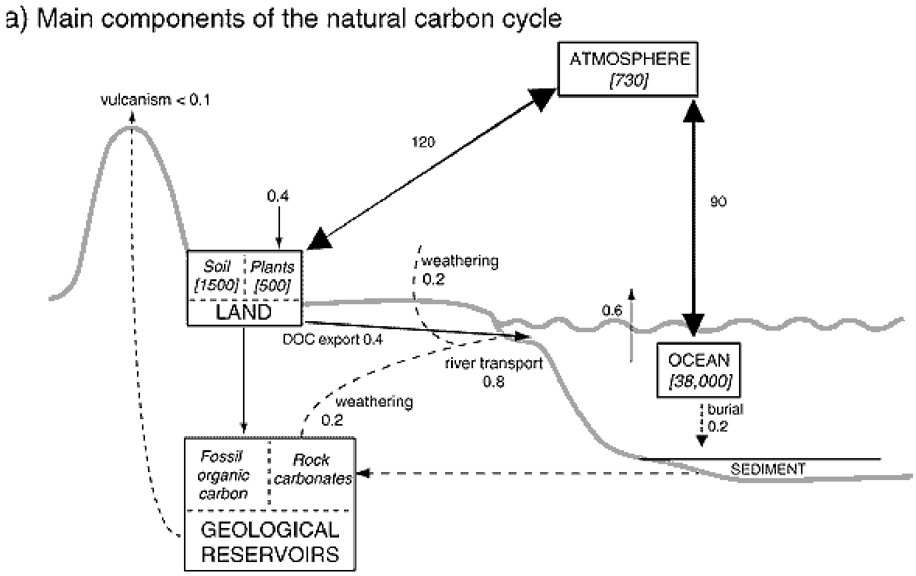

Figure 1.9: The natural carbon cycle (*Prentice et al.* [2001]).

fluxes between them is necessary to understand the evolution of the climate system.

1.4.1 Carbon dioxide

This section is devoted to the description of the carbon dioxide reservoirs and fluxes.

A first step is to identify the reservoirs (see figure 1.9). The carbon cycle is strongly influenced by the biosphere, which contains organic carbon. The living organisms store about 500 GtC[16] over land, and about 3 GtC in the ocean. The other main reservoirs are the atmosphere containing CO_2 (about 730 GtC), the ocean where carbon is dissolved in inorganic and organic forms (about 38000 GtC), the lithosphere containing organic matter in decomposition at the surface (1500 GtC are stored in the soil and detritus over land) and fossil fuels at deeper depths. Finally, sediments at the bottom of the ocean, originating from skeletons and shells of dead oceanic animals, contain calcium carbonates ($CaCO_3$).

Carbon continuously circulates between these reservoirs. It is usually assumed that each reservoir is naturally at equilibrium, so that the incoming and outgoing fluxes balance each other. Nonetheless, an increase of the atmospheric CO_2 concentration has been observed since preindustrial times. This has been attributed to human activities. First, the more or less undisturbed cycle is described to identify the main natural fluxes. Then, we will focus on the recent influence of human activities.

Carbon is exchanged between the atmosphere and the biosphere thanks to photosynthesis and respiration (both of the order of 120 GtC/yr). The photosynthesis produces organic matter and oxygen from CO_2, water, and solar energy:

$$6\,CO_2 + 12\,H_2O \xrightarrow{\text{solar energy}} C_6H_{12}O_6 + 6\,O_2 + 6\,H_2O,$$

[16] GtC is giga tons of carbon, $1\ GtC = 10^{12}\ kgC$.

where $C_6H_{12}O_6$ is an elementary organic component. The efficiency of this reaction depends on various climatic factors:

- The atmospheric concentration of carbon dioxide. Some plants (C_3-plants) are more sensible to this factor as they are less efficient than C_4-plants to trap atmospheric carbon (*Saugier* [1996]). Nevertheless, the uptake of carbon though photosynthesis levels off at high concentration.

- The available solar energy. Plants grow quicker in a sunny place as more light is available for photosynthesis. This effect is more conspicuous for C_4-plants, which need more energy to perform photosynthesis.

- The air temperature. Carbon dioxide enters into the leaves by stomates which tend to close themselves to reduce transpiration when the surrounding air is too warm.

- The soil humidity. The most important part of the water used to produce organic matter is pumped in the soil. Shortage in water resources inhibits photosynthesis.

- The availability of nutriments. The intermediate reactions in the pathway to photosynthesis need various other elements like phosphate and nitrate. If the plants encounter a shortage in one of these, the plant growth is restrained.

While photosynthesis uses solar energy to produce the elementary components of organic matter, respiration exploits organic matter to supply plants with energy in usable form. Respiration provides plants, animals, and decomposers with the heat and energy they need to ensure growth, reproduction, electric work in nervous cells, or mechanical work for the motions. While the intermediate reactions are not the same and do not involve the same molecules, the budget reaction for respiration is exactly the opposite one of the photosynthesis (*Saugier* [1996]). Respiration of the vegetation can be separated into two terms: the maintenance respiration and the growth respiration. The first one is proportional to the biomass and depends on temperature; the second is proportional to the growth and, therefore, to the carbon available from photosynthesis after consumption by the maintenance respiration. Decomposers also contribute to the total respiration of the ecosystem but their share depends on temperature, soil humidity, organic matter accessible for decomposition, etc. Finally, the animals also respire but release much less CO_2 in the atmosphere than vegetation.

Carbon dioxide is exchanged between the atmosphere and the biosphere but also with the ocean (see figure 1.10). At the sea surface, CO_2 dissolves in water inducing a carbon flux between the atmosphere and the ocean (about 90 GtC/yr in both directions, so that the net uptake of carbon dioxide by the ocean is almost zero if the human perturbation is not taken into account). This surface flux is influenced by the difference between the atmospheric and oceanic CO_2 concentrations, and on water temperature. Indeed, cold water can contain more dissolved carbon than warm water. Furthermore, the dissolution of carbon dioxide in sea water is achieved through the following reactions (see *Brasseur et al.* [1999]):

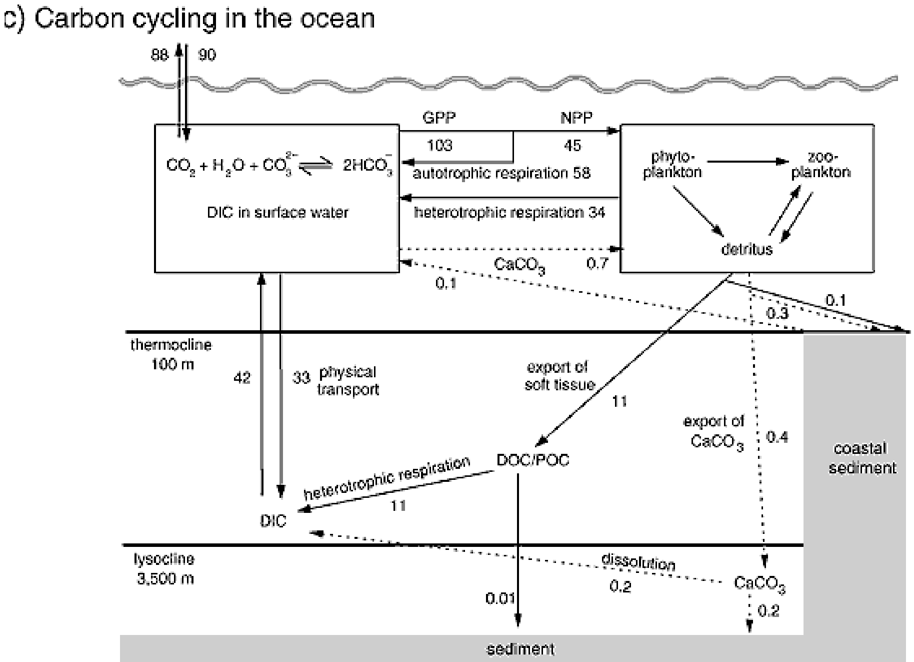

Figure 1.10: Carbon cycling in the ocean (*Prentice et al.* [2001]).

$$
\begin{aligned}
CO_{2(g)} &\rightarrow CO_{2(aq)}, \\
CO_{2(aq)} + H_2O &\rightleftharpoons H_2CO_3, \\
H_2CO_3 &\rightleftharpoons HCO_3^- + H^+, \\
HCO_3^- &\rightleftharpoons CO_3^{2-} + H^+,
\end{aligned}
$$

where (g) and (aq) stand for gas and aqueous phase, respectively. The relative abundance of the 3 forms of dissolved inorganic carbon (DIC), i.e., dissolved CO_2, HCO_3^-, and CO_3^{2-}, is thus determined by the water pH and the equilibrium constants of the above reactions. As the surface flux between atmosphere and ocean depends on the dissolved CO_2 concentration, all these parameters modify the air-ocean flux. The DIC is well mixed in the surface layer thanks to turbulence. It is transported by oceanic currents and sinks at greater depths in areas of deep convection (following the thermohaline circulation). Similarly to what happens on land, dissolved CO_2 is exchanged with the marine biota (103 GtC/yr by photosynthesis and 92 GtC/yr by autotrophic and heterotrophic respiration). Most of this occurs in the sunlit surface water. The organic productivity is greatly influenced by nutrient supply inducing strong regional discrepancies. Dissolved organic carbon compounds (DOC) are produced by the decay of dead cells or by direct release of organic molecules. It can penetrate more deeply inside the ocean, creating a net downward flux. Marine organisms also form shells of solid calcium carbonate ($CaCO_3$). Taking place mainly in sunlit surface water, this process depletes surface CO_3^{2-} and changes the surface water pH. This creates in out-gassing of CO_2 towards the atmosphere. When the shell-forming organisms die, they dive leading to sediments. Nonetheless, as they settle towards the bottom of the ocean, they often dissolve in deep waters and only a small fraction reaches the ocean floor.

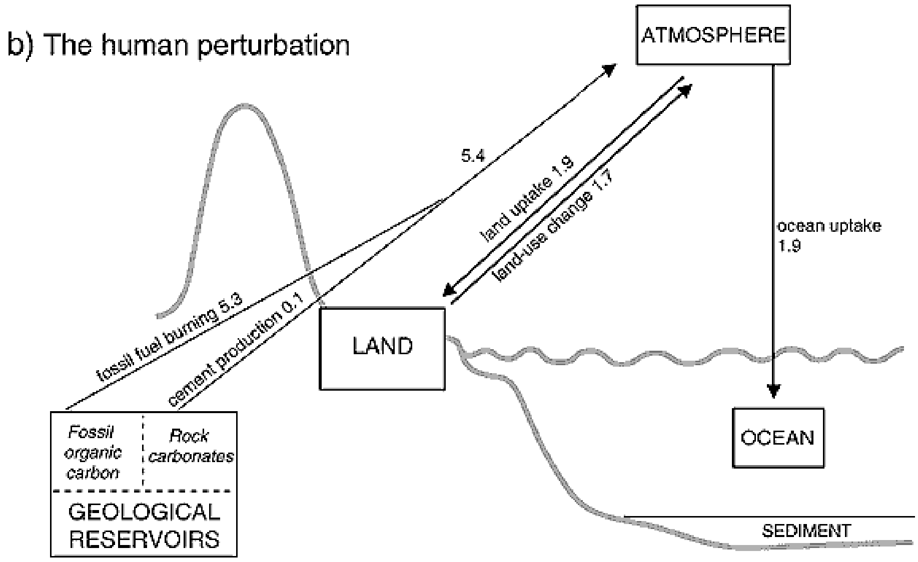

Figure 1.11: The human perturbation of the carbon cycle (*Prentice et al.* [2001]).

Let's draw some concluding remarks on the natural carbon cycle. On the one hand, many of these mechanisms depend on temperature, currents, etc., i.e., on climatic variables. On the other hand, these mechanisms determine the distribution of CO_2 in the atmosphere, which influences the Earth's climate. Thus, the carbon cycle introduces many positive and negative feedbacks in the climate system. Some of them are shortly explained hereafter (see further details in *Prentice et al.* [2001]).

- The atmospheric CO_2 affects the Earth's climate thanks to its radiative properties. Reciprocally, climate changes the uptake or release of CO_2 by the biosphere influencing its concentration in the atmosphere.
- The sea-surface temperature is partly governed by the radiative impact of CO_2 concentration in the atmosphere, but the net carbon uptake as dissolved inorganic carbon in the ocean is influenced by this temperature.
- Intensification of the greenhouse effect could alter the atmospheric near-surface temperature and the precipitation patterns, and the oceanic thermohaline circulation is very sensitive to these variables. Conversely, modifications of the oceanic circulation would change the distribution of dissolved inorganic carbon in the ocean and especially at the surface, leading to modifications in the atmosphere-ocean flux of CO_2.

For an ecosystem in equilibrium, the whole carbon uptake associated to photosynthesis is given back to the atmosphere through respiration of plants, animals, and decomposers. Similarly, the ocean CO_2 release and uptake should balance each other at equilibrium. It is often assumed that this would be the case without human intervention. Human activities modify this equilibrium (figure 1.11): first, by changing land use, and second by using fossil fuels as energy resource.

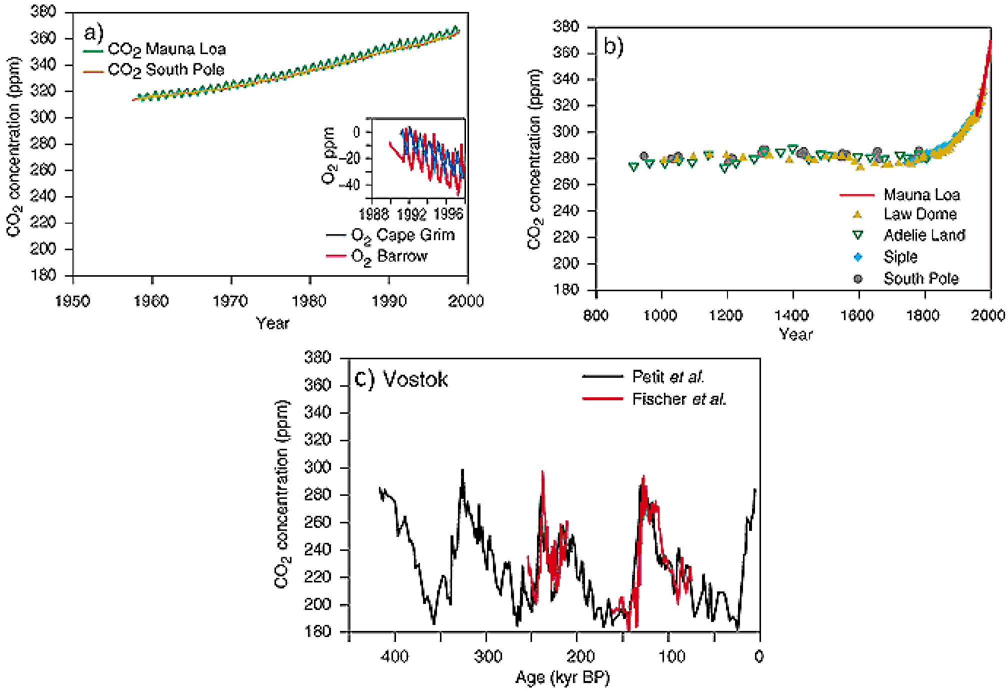

Figure 1.12: Variations in atmospheric CO_2 concentration on different time-scales from *Prentice et al.* [2001]. (a) Direct measurement of atmospheric CO_2 concentration. (b) CO_2 concentration in Antarctic ice cores for the past millennium. Recent atmospheric measurements at Mauna Loa are shown for comparison. (c) CO_2 concentration in the Vostok Antarctic ice core.

- For example, by replacing a forest by a field, we reduce the amount of carbon stored in the biomass and we also reduce the amount of carbon stored in the soil as we take away the harvest leaving only roots available for decomposition in the ground.

- The current combustion of fossil fuels as the main energy supply releases in the atmosphere a huge amount of carbon. This CO_2 was previously stored in the ground. During the 1980s, fossil fuel emissions amounted on average to 5.4 ± 0.3 GtC/yr, but these are still increasing and have reached 6.3 ± 0.4 GtC/yr during the 1990s. It took millions of years to store this carbon, while we are releasing it in a few centuries.

These are the main anthropogenic processes that release CO_2 in the atmosphere. Only part of this CO_2 (about 50%) stays in the atmosphere; the rest is taken up by the land (plants and soil) or by the ocean. This imbalance in the carbon cycle increases the atmospheric CO_2 concentration.

This theoretical point of view is confirmed by observations. The atmospheric concentration of carbon dioxide has increased from 280 ppmv in 1750 to 367 ppmv in 1999 (+31%), as shown in figure 1.12. The present CO_2 concentration has not been exceeded during the past 420,000 years and likely[17] not during the past 20 million years. Measurements of the

[17]In the IPCC 2001 report (*Houghton et al.* [2001]), virtually certain means greater than 99% chance

SOURCES OR SINKS	RANGE ($TgCH_4/yr$)
Natural sources	
Wetlands	92-237
Termites	20
Ocean	10-15
Hydrates	5-10
Anthropogenic sources	
Energy	75-110
Landfill	35-73
Ruminants	80-115
Waste treatment	14-25
Rice agriculture	25-100
Biomass burning	23-55
Other	15-20
TOTAL SOURCES	500-600
Sinks	
Soils	10-44
Tropospheric OH	450-510
Stratospheric loss	40-46
TOTAL SINKS	460-580
IMBALANCE (Trend)	+22

Table 1.3: Estimates of global methane budget (according to *Prather et al.* [2001]).

CO_2 concentration in air bubbles trapped in the Vostok ice core (Antarctica) show that it has been low (but ≥ 180 ppmv) during glacial periods and higher (but ≤ 300 ppmv) during the interglacial periods (figure 1.12c). Natural processes during the glacial-interglacial cycles have maintained CO_2 concentration within these bounds. The rate of increase in atmospheric CO_2 concentration has been about 1.5 ppmv per year over the past two decades. Again, this rate is without precedent, at least during the past 20,000 years. The isotopic composition of atmospheric CO_2 and the observed decrease in oxygen (O_2) demonstrates that the observed increase in CO_2 is predominantly due to the oxidation of organic carbon by fossil-fuel combustion and deforestation.

1.4.2 Methane

After water vapour and CO_2, CH_4 is the most abundant greenhouse gas in the troposphere. Its atmospheric concentration is linked to the carbon cycle but, whereas CO_2 is practically inert in the atmosphere, the main sink of CH_4 should be searched in the atmospheric chemistry.

that a result is true; very likely 90-99% chance; likely 66-90% chance; medium likelihood 33-66% chance; unlikely 10-33% chance; very unlikely 1-10% chance; exceptionally unlikely less than 1% chance.

CH_4 is the end product of a variety of reductive pathways during the decomposition of organic matter. Methane is therefore mainly produced in oxygen-deficient wetland habitats (see table 1.3). Methane is also released by anaerobic microbial activity in the stomachs of cattle, termites, etc. The current emissions from CH_4 hydrate deposits[18] appear small, about 10 TgCH$_4$/yr. But these deposits are enormous (about 10^7 TgC[19]) and are therefore a great potential source of methane.

In the troposphere, oxidation of methane by OH leads to the formation of formaldehyde, of carbon monoxide CO, and, in environments with sufficient NO_x, of ozone (see *Brasseur et al.* [1999] for the complete methane oxidation scheme). This results in a global loss rate of $\sim$ 507 Tg/yr. In the stratosphere, reactions with chlorine atoms, OH, and O (^{1}D) also destroy a small amount of CH_4 ($\sim$ 40 Tg/yr).

Attempts have been made to partly summarize this information in one value representing the strength of the sources (or sinks) compared to the total mass of CH_4 in the atmosphere. This quantity, called the lifetime, gives the order of magnitude of the time which would be needed by the sinks to remove the atmospheric methane, or by the sources to restore the atmospheric methane reservoir. The lifetime of methane in the atmosphere is approximately 8.4-12 years (*Prather et al.* [2001]).

As shown in table 1.3, there are anthropogenic sources of methane (e.g. agriculture, natural gas activities). Slightly more than half of current CH_4 emissions are anthropogenic. This additional release of methane increases the atmospheric CH_4 concentration. Direct measurements have been made since 1983 (figure 1.13b) and the record of atmospheric concentration has been extended to earlier times from air extracted from ice cores (figure 1.13a). Atmospheric methane concentrations have increased by about 150% since 1750 (from 700 ppbv to 1745 ppbv in 1998). The present concentration has not been exceeded during the past 420,000 years (figure 1.13c).

1.5 The cycle of other species

1.5.1 Trace gases

The energy cycle brings into light the impact of some trace gases called greenhouse gases (H_2O, CO_2, CH_4, N_2O, CFCs, and HCFCs) in the radiative budget of the climate system. The biogeochemical cycles of the first three have already been described in previous sections (1.3 and 1.4). Now the budget of the others will briefly be depicted.

As explained for methane, all greenhouse gases except CO_2 and H_2O are removed from the atmosphere primarily by chemical processes within the atmosphere. The Earth's atmosphere can be thought of as a low-temperature combustion system in which energy from the Sun is employed to drive a wide variety of oxidative reactions. The atmospheric composition is determined by a complex chemical mechanism that consists of many thousands of elementary reactions (the main ones are described in *Brasseur et al.* [1999]). Thanks to solar radiation, photolysis provides the system with excited species, which rapidly react

[18] Gas hydrates, also called gas clathrates, are natural solids composed of water molecules forming a rigid lattice of cages; each cage containing a molecule of gas, mainly methane. Gas clathrates are stable at high pressure, low temperature, and high gas concentration. (*Kvenvolden* [1993]; *Dickens* [1999])

[19] 1 TgC is one ton of carbon, 1 $TgCH_4$ is one ton of methane.

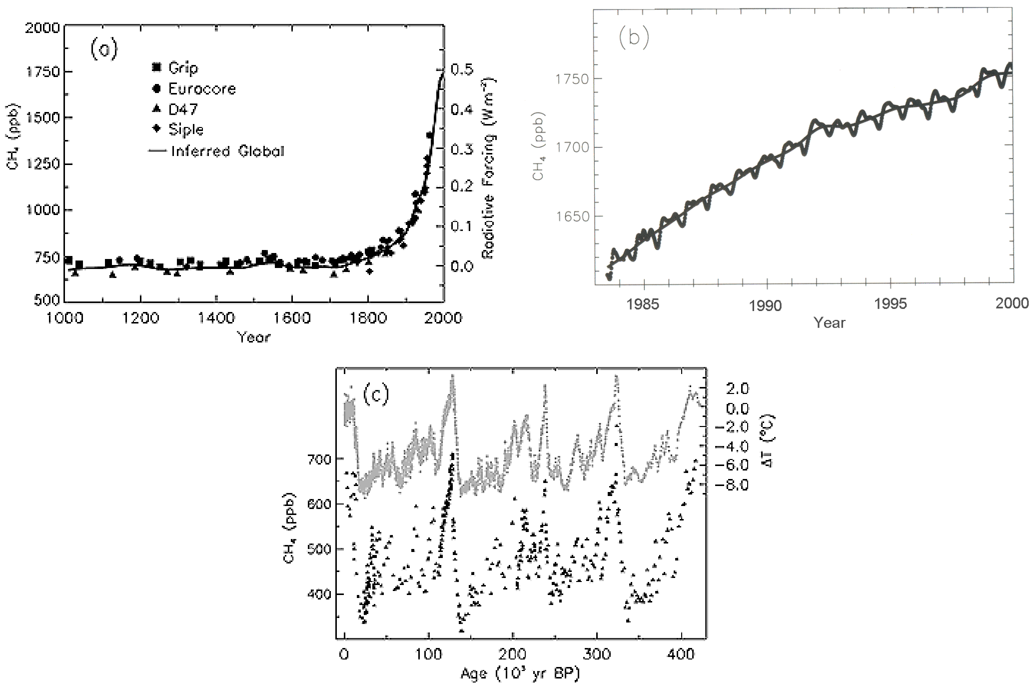

Figure 1.13: Variations in atmospheric CH_4 concentration for different time-scales from *Prather et al.* [2001]. (a) CH_4 concentration in ice cores for the past 1000 years. (b) Direct measurement of atmospheric CH_4 concentration. (c) Atmospheric CH_4 abundance determined for the past 420 kyr from the Vostok ice core.

with other trace gases and trigger chain reactions.

According to *Prather et al.* [2001], the most important sink of greenhouse gases containing one or more H atoms (e.g. CH_4, *HFCs* or *HCFCs*), as well as other pollutants, is their reaction with hydroxyl radicals (OH). This removal mainly takes place in the troposphere. The other greenhouse gases (e.g. N_2O, CFCs), that do not react with OH in the troposphere, are destroyed in the stratosphere or above by solar ultraviolet radiation.

On the contrary, greenhouse gases are mainly emitted at the Earth's surface, although some of them (like N_2O) are directly produced in the atmosphere through chemical reactions (like oxidation of NH_3). Further details about the known sources of greenhouse gases and an estimation of their strength can be found in *Prather et al.* [2001]. These are not described here as the main purpose of this work does not deal with chemistry.

The atmospheric concentration of N_2O has increased during the Industrial Era from 270 ppbv in 1750 to 314 ppbv in 1998. The present nitrous oxide concentration has not been exceeded during at least the past thousand years.

The atmospheric concentrations of many of those gases that are both ozone-depleting and greenhouse gases (CFCs) are either decreasing or increasing more slowly. But the observed atmospheric concentrations of the substitutes for the CFCs are increasing, and some of these compounds are greenhouse gases (HCFCs).

1.5.2 Aerosols

Every cubic centimeter of air contains up to thousands of suspended particles; most of these are only a fraction of a micrometer in diameter. Particulate matter and multi-phase mixture of solid or liquid particles dispersed in the atmosphere is commonly referred to as aerosol. Cloud drops are also in suspension in the atmosphere, but their diameter range from tens to hundreds of micrometers. Clouds and aerosols are not separate entities. As explained before in section 1.3, aerosols are cloud condensation nuclei. So clouds rely on the existence of aerosols for their formation, and aerosols are left behind if clouds re-evaporate.

Aerosol particles not only act as condensation nuclei but also affect the optical properties of the Earth's atmosphere. First, they alter the amount of solar energy that penetrates the atmosphere. Second, as condensation nuclei, they influence the cloud droplet size and therefore the cloud optical properties. There are called the direct and indirect radiative effects of the aerosols, respectively.

In other respects, aerosol particles have a significant effect on chemical reactions occurring in the atmosphere. In the troposphere, they are two main removal processes associated with aerosols (see *Brasseur et al.* [1999]). The first one is dry deposition. Aerosol particles are subject to gravity force. Their deposition velocity rapidly increases with particle diameter, so that dry deposition is dominated by large particles. Secondly, wet removal is an important sink of aerosols as the atmosphere contains water. Indeed, precipitation is an efficient process for cleaning the atmosphere of trace constituents. Water vapour condensates in aerosol particles and forms cloud droplets. This mechanism, called nucleation scavenging, transforms aerosols into cloud droplets. The cloud condensation nuclei are then carried away by rain drops towards the Earth's surface. Falling precipitation further scavenge aerosols through impaction, while this process is less efficient than nucleation scavenging.

SOURCES OR SINKS	RANGE (TgS/yr)
Sources	
Biomass burning	1-6
Fossil fuel and industry	60-100
Volcanos	6-20
Oceans	13-36
Land biota and soils	0.4-5.6
TOTAL SOURCES	80.4-167.6
Sinks	
Dry deposition	50-75
Wet deposition	50-75
TOTAL SINKS	100-150

Table 1.4: Global sulfur fluxes in TgS/yr (*Brasseur et al.* [1999] and *Penner et al.* [2001]).

Tropospheric aerosol particles have a relatively short lifetime according to this link between aerosols and the hydrological cycle. Therefore, their spatial and temporal distributions, as well as their physical or chemical properties, exhibit an important variability according to local emissions. On the contrary, stratospheric particles have longer lifetime. This is a consequence of the lack of wet removal processes.

On the other extremity of the aerosol cycle in the atmosphere are the sources. Mechanical processes (like wind blown dust, volcanos, sea spray) produce relatively large aerosol particles. But some fine particles (like soot or aromatic hydrocarbons) are emitted also near the surface. Nonetheless, the aerosol sources are not limited to the Earth's surface. They are also created in the atmosphere by physical or chemical transformations. For example, sulphate SO_4^{2-} forms by conversion of sulfur gases to condensible species. This kind of processes known as gas-to-particle conversion is an important sink for some tropospheric trace gases.

A more complete overview of aerosol sources, sinks, and properties can be found in *Brasseur et al.* [1999] or in *Charlson and Heintzenberg* [1995].

For example, sulphate particles are believed to be the major source of cloud condensation nuclei. The biogeochemical cycle of sulfur has therefore drawn attention of scientists. The main sources and sinks of sulfur are listed in table 1.4. Sulfur is a chemical element essential to life on Earth. Living organisms assimilate sulfur and release it in various forms as an end product of metabolism. Marine biota "produces" dimethyl sulfide (DMS), which is then released into the atmosphere. Vegetation and soils generate a flux of $0.4-5.6\,\mathrm{TgS/yr}$. Volcanos and biomass burning also bring sulfur into the atmosphere. Human activities create the most important source $(60-100\ < rmTgS/yr)$. Once in the atmosphere, all the sulfur components emitted at the Earth' surface are oxidized by reactions primarily with OH to form SO_2 and sulphates. These two species are removed from the atmosphere by dry and wet depositions.

The regional nature of aerosols makes tropospheric aerosol trends more difficult to observe than trends in long-lived trace gases. Ice cores from both Greenland and the

Alps display the strong anthropogenic influence of sulphate deposited during this century. Sulphate in Antarctic ice cores shows no such trend since its source in the Southern Hemisphere is primarily natural. Satellite observations are naturally suited to the global coverage demanded by regional variations in aerosols. This kind of studies are going on since the 1990s.

1.6 Climate variability and climate change

Everybody has already heard on the radio or on TV that climate is changing. Many international meetings have been devoted to this question: Rio, Bonn, Kyoto, Den Haag, Marrakech. These were brought to the general public's attention by journalists. Furthermore, after any natural disaster related to climate (e.g. unusual drought, flooding, storm, snowfall) the same question comes back: "Is this due to human activities ?".

From a scientific point of view, unusual conditions are not easy to determine. Indeed, a phenomenon is unusual for people because they do not remember a similar event, but precise human memory is short term, at most 30 years. For scientists, this is not significant. We have to take into account the natural climate variability prior to say that an event is unusual. Thus the detection of climate change is a statistical "signal-in-noise" problem. As defined by the IPCC[20] (*Houghton et al.* [2001]),

> "*detection* is the process of demonstrating that an observed change is significantly different (in a statistical sense) than it can be explained by natural internal variability".

Detection studies have been performed under the framework of the IPCC. Their main conclusions concerning the observed climate change are summarized hereafter. Let's start with temperature changes.

- The globally averaged surface temperature (the average of near surface air temperature over land and sea-surface temperature) has increased since 1861 by 0.6 ± 0.2°C (figure 1.14). Most of the warming occurred during two periods, 1910 to 1945 and 1976 to 2000. Globally, it is very likely[21] that the 1990s was the warmest decade in the instrumental record since 1861.

- Thanks to the sparseness of instrumental climate records prior to the 20th century, estimates of global climate variability during the past centuries must rely upon indirect "proxy"-indicators. These includes width and density measurements from tree rings; layer thickness from laminated sediments cores; isotopes, chemistry and accumulation from annually resolved ice cores; isotopes from coral; and sparse historical documentary evidence available over the globe during the past few centuries. The Northern Hemisphere average surface temperature evolution has been reconstructed

[20] Intergovernmental Panel on Climate Change

[21] In the IPCC report, *virtually certain* means greater than 99% chance that a result is true; *very likely* means 90-99% change that a result is true; *likely* 66-99% chance; *medium likelihood* 33-66% chance; *unlikely* 10-33% chance; *very unlikely* 1-10% chance; *exceptionally unlikely* less than 1% chance.

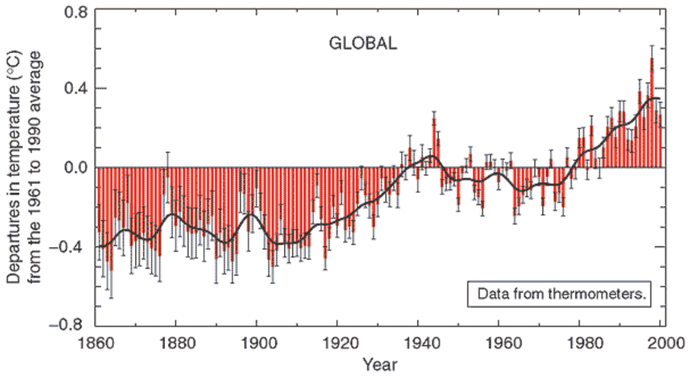

Figure 1.14: Variations of the Earth's surface temperature over the last 140 years. Year-by-year values (red bars), approximately decade by decade values (black line), and uncertainties (thin black whisker bars represent the 95% confidence range) (*Folland et al.* [2001]).

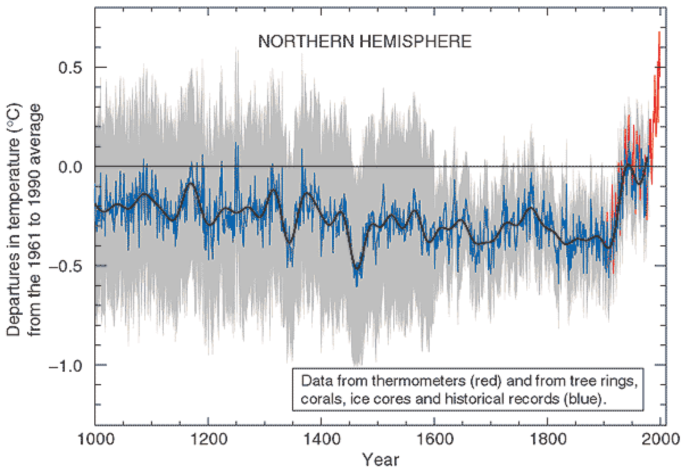

Figure 1.15: The year-by-year (blue curve) and 50 year average (black curve) variations of the average surface temperature of the Northern Hemisphere for the past 1000 years. The 95% confidence range in the annual data is represented by the grey region. The thermometer data since 1861 is also shown in red (*Folland et al.* [2001]).

for the past 1000 years (figure 1.15). From these data, it is likely that the rate and duration of the warming of the 20th century is larger than any other time during the last 1000 years. Furthermore, the 1990s is likely to have been the warmest decade of the millennium in the Northern Hemisphere.

- This warming is not uniform in space and time. The regional patterns of the warming are different in the early or latter parts of the 20th century. Some regions like Antarctica do not seem to undergo any temperature change. On average, minimum land-surface temperature for 1950 to 1993 are increasing at about twice the rate of maximum temperature (0.2 versus 0.1°C).

- Surface and lower troposphere (lowest 8 km) have warmed, but both satellite and weather balloon data (since 1979) show significantly less lower troposphere warming than observed at the surface.

Temperature is not the only changing climatic parameter. Many related aspects of climate are also modified. For example, the global ocean heat content has increased since the late 1950s and sea level rose between 0.1 and 0.2 m during the 20th century.

Decreasing snow cover and land-ice extent is positively correlated with increasing land-surface temperatures.

- The extent of snow cover since the late 1960s has very likely been decreased of about 10%. Over the past 100 to 150 years, a reduction of about two weeks in the annual duration of lake and river ice in the mid- to high latitudes of the Northern Hemisphere is very likely.

- *Houghton et al.* [2001] relates that late summer to early autumn Arctic sea-ice thickness has decreased by about 40% between the period of 1958 to 1976 and the mid-1990s. The observed thinning of sea ice in winter is however substantially smaller. Nevertheless, these conclusions must be taken with caution as they are drawn from local and partial datasets (see *Fichefet et al.* [2003a] for more details). No significant trend in Antarctic sea-ice extent is apparent over the period of systematic satellite measurements (since 1978).

The hydrological cycle has also been disturbed.

- Over most mid- and high latitudes of the Northern Hemisphere, annual land precipitation has very likely increased by 0.5 to 1%/decade in the 20th century. On the contrary, it is likely that land-surface rainfall has decreased on average by about -0.3%/decade over the Northern Hemisphere subtropics (10°N to 30°N). It is also likely that precipitation has increased by about 0.2 to 0.3%/decade over the tropical (10°N to 10°S) land areas. In contrast to the Northern Hemisphere, no significant trend has been detected in broad latitudinal averages over the Southern Hemisphere.

- Changes in total cloud amounts over Northern Hemisphere mid- and high latitude continental regions indicate a likely increase in cloud cover of about 2% since the beginning of the 20th century. In most of these regions, these trends have a strong

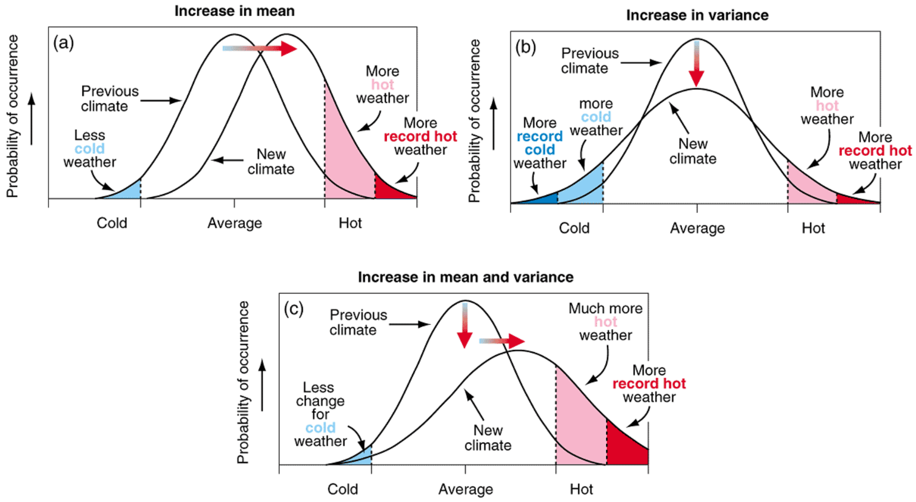

Figure 1.16: Schematic showing the effect on extremes temperatures when (a) the mean temperature increases, (b) the variance increases, and (c) when both the mean and variance increase for a normal distribution of temperature (*Folland et al.* [2001]).

correlation to precipitation increases. Furthermore, the cloud-cover trend is positively correlated with decreases in the diurnal temperature range (clouds avoid solar radiation to reach the surface during day-time cooling the maximum temperature while they trap infrared radiation during night-time increasing the minimum temperature).

The mean values are changing but also climate variability and climate extremes. Understanding changes in climate variability and climate extremes is made difficult by interactions between the changes in the mean and variability. Such interactions vary from variable to variable depending on their statistical distribution. Some examples are shown in figure 1.16 for a variable like temperature with a normal distribution. In figure 1.16a, the range between the hottest and coldest temperatures does not change. An increase in variability without a change in the mean (figure 1.16b) implies an increase in the probability of both hot and cold extremes as well as the absolute value of the extremes. Increase in both the mean and variability (figure 1.16c) are also possible, which affects the probability of hot and cold extremes and their absolute value. Other combinations of changes would of course lead to different results. Some recent modifications of climate variability has been detected:

- It is likely that there has been a widespread increase in heavy and extreme precipitation events in regions where total precipitation has increased, e.g., the mid- and high latitudes of the Northern Hemisphere.

- A significant reduction in the frequency of much below normal seasonal mean temperatures has been observed across much of the globe. A corresponding smaller increase in the frequency of much above normal temperatures has also been detected. Consequently, the global mean temperature has increased.

- New record high night-time minimum temperatures are lengthening the freeze-free season in many mid- and high latitudes regions.
- In some regions, such as parts of Asia and Africa, the frequency and intensity of drought have been observed to increase in recent decades.
- There is no compelling evidence to indicate that the characteristics of tropical and extra-tropical storms have changed. In general, trends in severe weather events are notoriously difficult to detect because of their relatively rare occurrence and large spatial variability.
- The behaviour of ENSO[22] has been unusual since the mid-1970s compared with the previous 100 years, with warm phase ENSO episodes being relatively more frequent, persistent, and intense than the opposite cold phase. This of course induces variations in precipitation and temperature over much of the global tropics and subtropics.

Studies have also been performed using estimations of climate internal variability simulated by climate models. This approach provides an other point of view and a tool to estimate the consistency between the observed changes. While models are not perfect and differ from each other, conclusions on the detection of an anthropogenic signal are insensitive to the model used to estimate internal variability. It is thus unlikely that natural internal variability alone can explain the changes in global climate over the 20th century.

However, the detection of a change in climate does not necessarily imply that its causes are understood. This attribution problem is defined as follows by the IPCC (*Houghton et al.* [2001]):

> "*attribution* is the process of establishing cause and effect with some defined level of confidence, including the assessment of competing hypotheses".

As experimentation with the climate system is clearly not possible, scientists use climate models to simulate the climate response to external forcing. Furthermore, more and more complex statistical methods are used to detect the fingerprint of the forcing in the observed changes, not only on global or hemispheric averages but also on horizontal, vertical, and temporal patterns of the response. Recent studies have tried to estimate the impact of either natural (solar and volcanic) forcing and anthropogenic (greenhouse gases and aerosols) forcing during the 20th century.

First, some attempts have been made to identify the contribution of natural forcing to climate variability and change (figure 1.17a). Fully coupled atmosphere-ocean models have used reconstructions of solar and volcanic forcings over the last one to three centuries.

[22] The El Niño-Southern Oscillation (ENSO) is the strongest natural fluctuation of climate on interannual time-scales. It is a coupled atmosphere-ocean phenomenon. During an El Niño event, the prevailing trade winds in the low equatorial atmosphere weaken. Consequently, the equatorial countercurrent strengthens in the ocean, causing warm surface waters in the Indonesian area to flow eastwards to overlie the cold waters of the Peru current. This event has a large impact on the wind, sea-surface temperature, and precipitation patterns in the tropical Pacific. It has climatic effects throughout the Pacific region and in many other parts of the world. These changes have a profound impact on humanity and society because of associated droughts, floods, and heat waves that can severely disrupt agriculture, fisheries, etc. The opposite of an El Niño event is called La Niña.

This brings the model variance closer to the one deducted from paleo-reconstructions. They show that natural external forcing may have contributed to the warming that occurred in the early 20th century, but the climate changes over the past 30 to 50 years are unlikely to be due by natural forcing. Indeed, the overall trend that they induce is likely to have been small or negative over the last two decades. Furthermore, the observed changes in vertical temperature structure of the atmosphere is inconsistent with natural forcing.

Secondly, simulations have been performed with models including the individual greenhouse gases, the direct and indirect effects of sulphate aerosols, and the influence of changes in stratospheric and tropospheric ozone. These produce estimates of the climate response to anthropogenic forcing (figure 1.17b). There is a wide range of evidence of qualitative consistencies between observed climate changes and model responses to anthropogenic forcing. Models and observations show increasing global temperature, increasing land-ocean temperature contrast, diminishing sea-ice extent, glacial retreat, and increases in precipitation at high latitudes in the Northern Hemisphere. But the models predict a faster rate of warming in the mid- to upper troposphere than is observed in data records. Over the last 50 years, the model estimated rate and magnitude of surface warming due to increasing concentrations of greenhouse gases alone are comparable with, or larger than, the observed warming. But most model simulations that take into account both greenhouse gases and sulphate aerosols are consistent with observations over this period, as the sulphate aerosol forcing, while uncertain, is negative.

Finally, the best agreement between model simulations and observations over the last 140 years has been found when all the above anthropogenic and natural forcing factors are combined, as shown in figure 1.17c.

All these arguments bring the IPCC experts to these important conclusions in the 2001 report (*Houghton et al.* [2001]):

> "In the light of new evidence and taking into account the remaining uncertainties, most of the observed warming over the last 50 years is likely to have been due to the increase in greenhouse gas concentrations".
>
> "It is very likely that the 20th century warming has contributed significantly to the observed sea-level rise, through thermal expansion of seawater, and widespread loss of land ice".

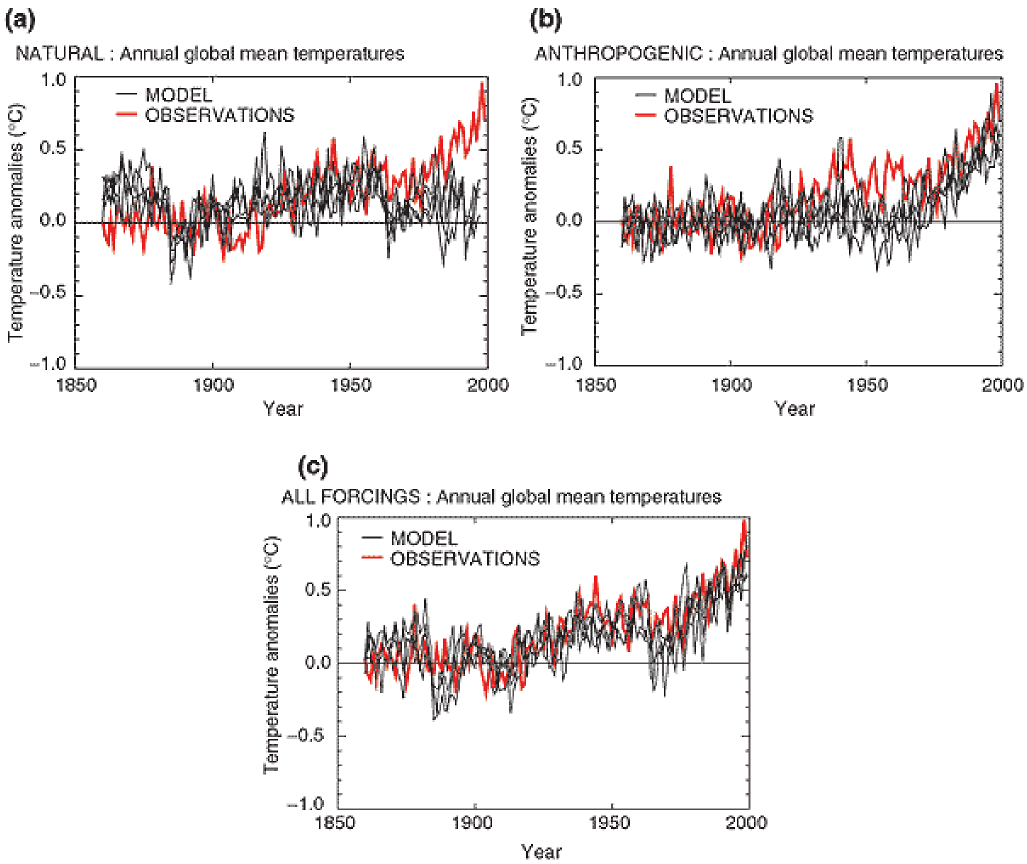

Figure 1.17: Global mean surface temperature anomalies relative to the 1880 to 1920 mean from the instrumental record compared with ensembles of four simulations with a coupled atmosphere-ocean climate model driven (a) by solar and volcanic forcings, (b) by anthropogenic forcing and (c) by all forcings, both natural and anthropogenic (*Mitchell et al.* [2001]).