THE CLIMATE SYSTEM

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Contents

Summary

The climate system is extensive, multi-faceted, and all-pervasive. It has been defined as encompassing components of the atmosphere, hydrosphere, biosphere, geosphere, and cryosphere. Its space scales range from points to the planet, and the timeframes of

interest to climate scientists extend from seconds to eons. Over the longest timescales, the Earth's climate has been remarkably stable: liquid water has been present on the planet's surface since at least 3.8 billion years ago. Glacial epochs have occurred throughout climate history prompting the build-up of masses of land-ice and hence the drawing down of sea levels globally. Disturbances from the mean, from catastrophes to hiccups, can be recognized in the climate record. These include periods when particulates, some caused by meteorite impacts and others from volcanos, dramatically cooled temperatures, and when a biophysically derived build-up of greenhouse gases warmed the Earth.

Future climates will be susceptible to the same internal and external forces as past climates. In addition, human activities are known to be modifying the atmospheric composition and the continental surface in ways that are likely to modify the pace and, perhaps, the direction of natural climatic change. Human impacts on climate differ from many natural processes because of the speed of the changes.

Climate models have been developed to try to characterize all the important features of the Earth's climate system. Different model types are intended to be better suited to specific aspects of climate change. Greenhouse climate prediction has been conducted with global climate models; integrated assessment models are being developed to assess the interactions of biospheric changes and climate shifts; and simpler energy budget models have been employed to explain the Milankovitch cycling of climate in response to orbital changes of the Earth around the Sun.

A hospitable climate is fundamental to sustainable development on Earth. Understanding climate sensitivities to internal and external changes and its reciprocal, biospheric, societal, and technological sensitivity to climate are crucially important to sustaining life systems on Earth.

1. Earth's Climate

1.1 Introduction

Climate was once defined, rather simply, as "average weather," but "the climate system" has come to be defined more completely over the last few decades. As far back as 1975, the Global Atmospheric Research Programme (GARP) of the World Meteorological Organization stated that the climate system is composed of the atmosphere, hydrosphere, cryosphere, land surface, and biosphere. The United Nations Framework Convention on Climate Change (FCCC) followed in 1992 with its definition of the climate system: the totality of the atmosphere, hydrosphere, biosphere, and geosphere, and their interactions. While these definitions are similar, the emphasis on interactions can be seen to have grown in the twenty-seven years separating them and indeed this emphasis continues to increase.

People's concept of the climate system is concerned with personal and societal issues of habitability and sustainability. Most of us evaluate climate in fairly simple terms such as temperature: is it too hot or too cold? Chemistry: is the atmosphere breathable? Sustenance: is there enough water for drinking and for growing crops? And ambient environment: will it feel comfortable, i.e. not too humid nor too cold? These characteristics are interdependent, together forming the climate system and posing a larger question: can this planet continue to sustain life? At the present time, we do not know how to answer this question but some fundamentals are clear. If the global mean temperature does not lie between 0^{o} C and 100^{o} C (at a pressure of about 1 atmosphere) water cannot exist in liquid form and the chemistry of the atmosphere will be greatly changed. At the same time, the most important greenhouse gas in the Earth's atmosphere is water vapor, so that its maintenance in a vapor phase becomes a sustaining mechanism for the climate conditions that humans consider to be habitable.

Today, the atmosphere is undergoing global changes unprecedented in human history and, although changes as large as today's have occurred in the geological past, relatively few have happened with the speed which also characterizes today's climate changes. Concentrations of greenhouse gases are increasing, stratospheric ozone is being depleted and the changing chemical composition of the atmosphere is reducing its ability to cleanse itself through oxidation. These global changes are threatening the balance of climatic conditions under which life evolved and is sustained. Temperatures are rising, ultraviolet radiation is increasing at the surface, and pollutant levels are increasing. Many of these changes can be traced to industrialization, deforestation, and other activities of a human population that is itself increasing at a very rapid rate.

Now, for the first time in the history of our planet, emissions of some trace gases from human activities equal, and perhaps even exceed, emissions from natural sources. This is important for the climate system because the atmosphere's primary constituents, molecular nitrogen and molecular oxygen, are transparent to infrared radiation, so that the greenhouse gases (mainly water vapor, carbon dioxide, ozone, methane, nitrous oxide, and the chlorofluorocarbons, or CFCs), present in much smaller amounts, play a major role in the Earth's energy budget and climate through the greenhouse effect. Trace gases also govern the production and destruction of ozone, affect many biospheric processes and play other important roles in the climate system. It is of great importance to determine where and how these constituents enter the atmosphere, and how they are distributed and transformed by the complex interactions of sunlight, air, land, sea, and living organisms; in sum, how they behave in the climate system. Trace gases are carried from their surface sources to the upper atmosphere and around the world by numerous, interdependent processes such as mixing in the atmospheric and oceanic boundary layers, vertical exchanges associated with weather systems and moist convection at mid latitudes, deep convection in tropical storms and below sea-ice, and the mean circulation of the atmosphere, oceans, and even the rocks.

Possibly the most important species in atmospheric terms (and hence for the chemistry of the climate system) is the hydroxyl radical (OH). Generated by interactions among ozone, water vapor, and ultraviolet radiation, OH is the atmosphere's primary agent of oxidation, which is the means by which many compounds are transformed into others more readily removed from the atmosphere. The atmospheric OH level may be changing, not only because of changes in ozone and ultraviolet light, but also because of changes in its rate of loss due to reaction with carbon monoxide (CO). Climate changes cannot be fully understood without improved determination of the net effect of these complex interactions on the abundance of this dominant oxidant.

Global climate system changes resulting from human influence have been described as "creeping crises." Their characteristics seem to be that they are slow to develop and, therefore, may not become apparent until their effects have become dangerously advanced. The iconic demonstration of this dates back to 1985, when British scientists discovered that the mean ozone column abundance for October over the Antarctic station of Halley Bay had been decreasing very rapidly since the late 1970s, forming the so-called Antarctic ozone hole. Stratospheric ozone concentrations at the South Pole in spring are now very much less than half of their values only 30 years ago. Worldwide, stratospheric ozone has declined noticeably (now by a few percent) and in the Northern Hemisphere, where the stratospheric circulation is more complicated, similar springtime depletions have developed over the last decade.

To the first order, the Earth's climate is controlled by the amount of incident solar radiation that is absorbed by the planet, and by the thermal absorptivity of the gases in the atmosphere, which controls the balancing emitted infrared radiation. Radiation from the Sun drives the climate of the Earth and, indeed, of the other planets in the solar system. Solar radiation is absorbed and, over the mean annual cycle, this absorption is balanced by radiation emitted from the Earth. This global radiative balance, which is a function of the surface and atmospheric characteristics, of the Earth's orbital geometry, and of solar radiation itself, controls the habitability of the Earth, mean temperatures, and the existence of water in its three phase states. These characteristics, together with the effects of the rotation of the Earth on its axis, determine the dynamics of the atmosphere and ocean, and the development of snow masses. Over very long time-scales, those commensurate with the lifetime of the Earth, astronomical, geological, and biological processes control persistence of ice caps and glaciers, the biota, rock structures, and global geochemical cycling.

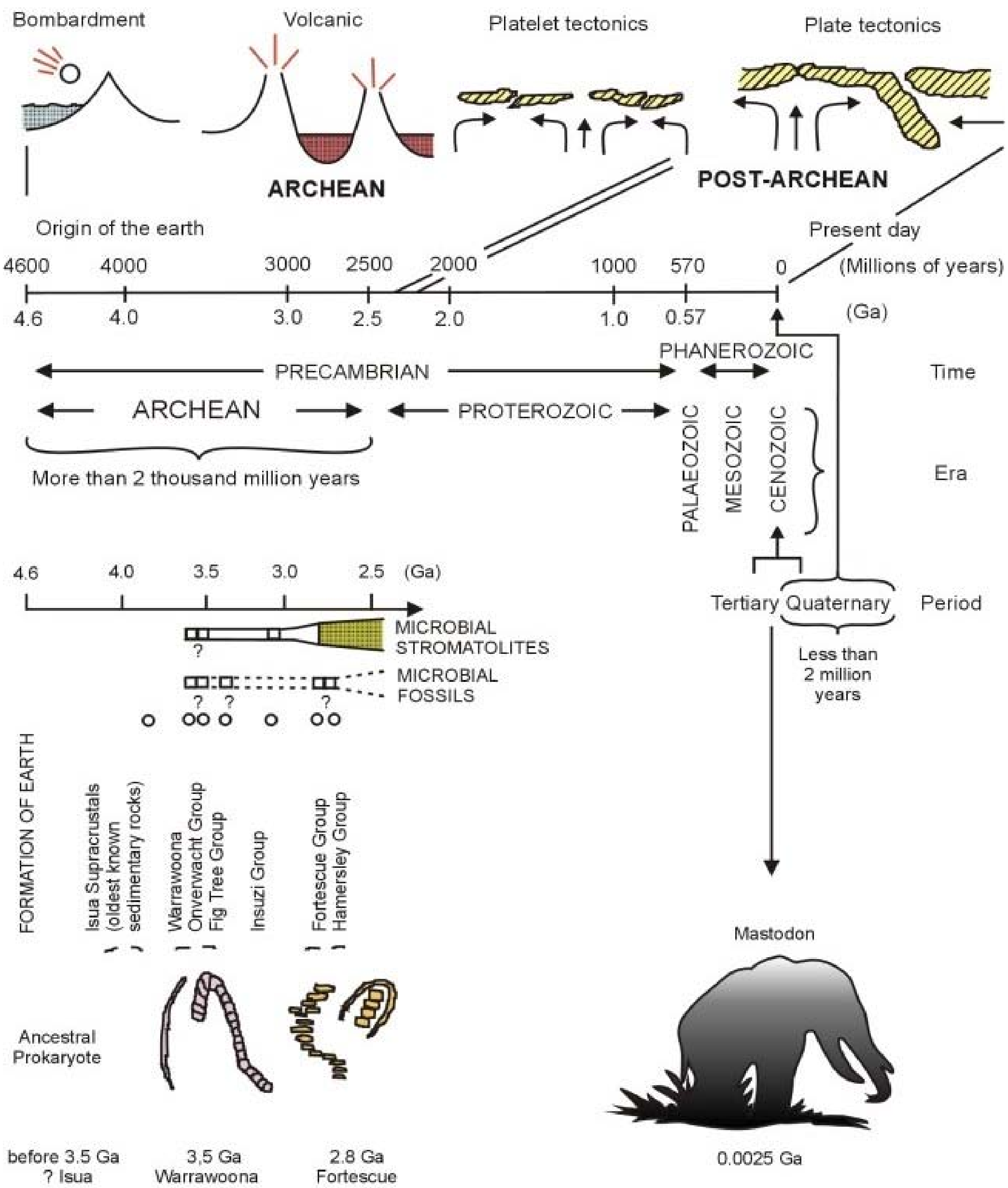

Figure 1 (a): Schematic representing the geophysical and astronomical forcings exerted on the climate system over the lifetime of the planet. Despite massive upheavals and large-scale changes in external and internal environments, life has developed and persisted throughout almost the full extent of Earth's history (after Henderson-Sellers *et al.*, 1991).

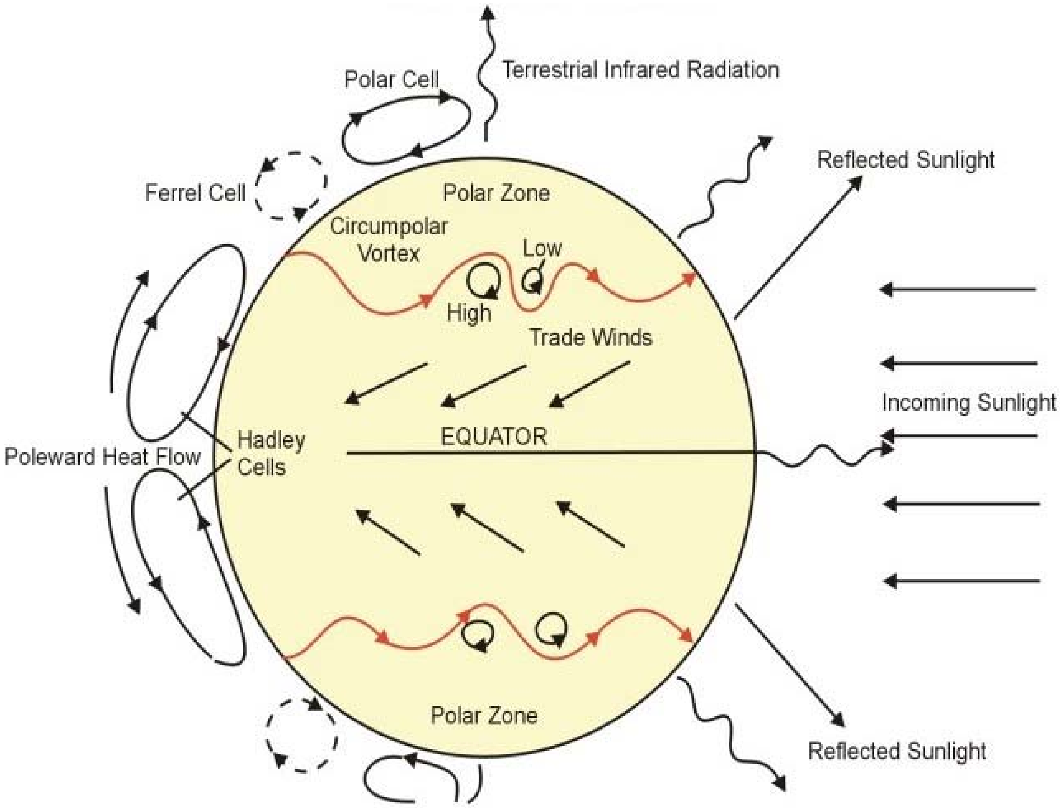

Figure 1(b): Astronomical controls on the climate system include the planetary radiation balance (absorbed solar equals emitted infrared) and the effect on atmospheric and oceanic circulation of the planet's spin around its axis (after Henderson-Sellers, 1995)

The Earth's climate system strongly couples the atmosphere, the land surface, the oceans, and surface water (the hydrosphere) to those parts of the Earth covered with ice and snow (the cryosphere) and the biosphere (the vegetation and other living systems on the land and in the ocean).

There are at least two different and complementary timeframes that need to be borne in mind in reviewing the climate system. The first is the evolutionary timescale, which controls the very long-term aspects of the climate components, and those factors which force it, such as the physics and chemistry of the planet itself and the luminosity of the Sun. Viewed in this timeframe, the Earth's climate is prey to the forces of astro- and geophysics. The Archaean period represents almost half of Earth's history from the formation of the Earth around 4.6 billion years ago until the beginning of the Proterozoic era about 2.5 billion years ago (Figure 1(a)). Geophysically, the Archaean differed from the post-Archaean period because there were no plate tectonics. Instead, the planetary surface was first bombarded by cometary and meteoritic material as the debris from the planetesimals was "swept up." This was followed by a period of surface vulcanism, and then by a platelet tectonic regime. Life must have become established around or before 3.5 billion years ago, because rocks from that era contain fossil evidence of viable and diverse microbial communities. The oldest known rocks, the Isua supercrustal, are dated at 3.83 billion years and show evidence of a global climate system not grossly different from that of the present. In particular, they contain sedimentary material that was waterborne. This suggests that the Earth's climate has been very stable when viewed in this evolutionary timeframe.

Within this very long timescale it is possible to take a "snapshot" view of the climate system (e.g. Figure 1(b)). In this "instantaneous" view, the shortest timescale processes are most evident. Of these, the most important are the latitudinal distribution of absorbed solar radiation (large at low latitudes and much less near the poles), as compared to the emitted thermal infrared radiation which varies much less with latitude. This latitudinal imbalance of net radiation for the surface-plus-atmosphere system as a whole (positive in low latitudes and negative in higher latitudes) combined with the effect of the Earth's rotation on its axis produces the dynamical circulation system of the atmosphere (Figure 1(b)).

The latitudinal radiative imbalance tends to warm air which rises in equatorial regions and would sink in polar regions were it not for the rotation of the Earth. The westerly waves in the upper troposphere in mid-latitudes, and the associated high and low pressure systems, are the product of planetary rotation affecting the thermally driven atmospheric circulation. The overall circulation pattern comprises thermally direct cells in low latitudes, strong waves in the mid-latitudes, and weak direct cells in polar regions. This circulation, combined with the vertical distribution of temperature, represents the major aspects of the atmospheric climate system.

Figure 1 illustrates two very fundamental characteristics of the Earth's climate system. The first is that over the lifetime of the planet and despite massive upheavals, the climate has remained remarkably stable. The second is that at any instant of time, the most visible characteristics of the climate are the direct result of two astronomical forces: the planetary radiative balance and its axial rotation.

The other timeframe that is of fundamental importance for the climate system is that for equilibration, i.e. the time needed to adjust to a new forcing. The equilibration times for different subsystems of the Earth's climate system differ very markedly (see Table 1). The longest equilibration times are those for the deep ocean, glaciers, and ice sheets (10^{10}-10^{12} seconds); the remaining elements of the climate system have equilibration times nearer 10^5–10^7 seconds.

Climatic Component	Seconds	Equivalent
Atmosphere		
Free	10^6	11 days
Boundary layer	10^5	24 hours
Hydrosphere		
Ocean Mixed layer	10^6 – 10^7	months – years
Deep ocean	10^{10}	300 years
Lakes and rivers	10^6	11 days
Cryosphere		
Snow and surface ice layer	10^5	24 hours

Sea-ice	$10^6 – 10^{10}$	Days – 100s of years
Mountain glaciers	10^{10}	300 years
Ice sheets	10^{12}	3000 years
Biosphere		
Soil/vegetation	$10^6 – 10^{10}$	11 days – 100s of years
Lithosphere	10^{15}	30 million years

Table 1: Equilibration times for components of the Earth's climate system (modified from McGuffie and Henderson-Sellers, 1997)

The result of these two timescales (evolution and equilibration), and of the complex interactions between these components of the climate system, is a rich spectrum of climatic change (Figure 2(a)). The largest peaks in this spectrum relate to astronomical forcings: the Earth's rotation, its revolution around the Sun, variations in this orbit, and the formation of the solar system. Coupling and feedbacks amongst processes within the climate system components, the atmosphere with the oceans, surface water with ice masses, and so on is responsible for the myriad of variations in this climate system spectrum.

1.2 Earth as a Planet

The planets in our solar system were formed about 4.6 billion years ago by accretion of the gases and particulates left over from the formation of the sun. The giant planets collected most of the lighter gases while the terrestrial planets accreted from material similar to that forming today's comets and meteors. While our sun possesses practically all the mass of the solar system, the planets' angular momenta sum to about 200 times that of their central star. The orbital periods and axial rotation rates of the nine planets vary enormously (see Table 2(a)). These differences combined with the differences in mass, density and character of their surfaces and atmospheres to give rise to planetary climates that range from frozen wastelands, through stormy skies to baked deserts and hot acidic rain.

It is possible to recognize similarities between the Earth and its nearer neighbors For example, the inner four planets (the terrestrial group) share somewhat similar sizes, densities, and atmospheric constituents as compared to the massive outer planets (the jovian group), which are larger, lighter, and composed mostly of hydrogen and helium (Table 2(a)). Despite these similarities, it is also apparent that the Earth presents some highly anomalous characteristics (e.g. Table 2(b)). Why, for example, is water so abundant on Earth and yet virtually absent from Mars and Venus? Why is carbon dioxide the dominant gas in the other terrestrial planetary atmospheres, but only a trace component of the Earth's atmosphere?

The anomalies in the Earth's climate system arise because the actual concentrations of very many atmospheric constituents are very different from calculated equilibrium concentrations that would be present if the Earth were devoid of life. Observed concentrations range from nine to twenty nine orders of magnitude different from those expected (Table 2(c)). At a planetary scale, the need to explain and fully understand this "climate anomaly" is critically important.

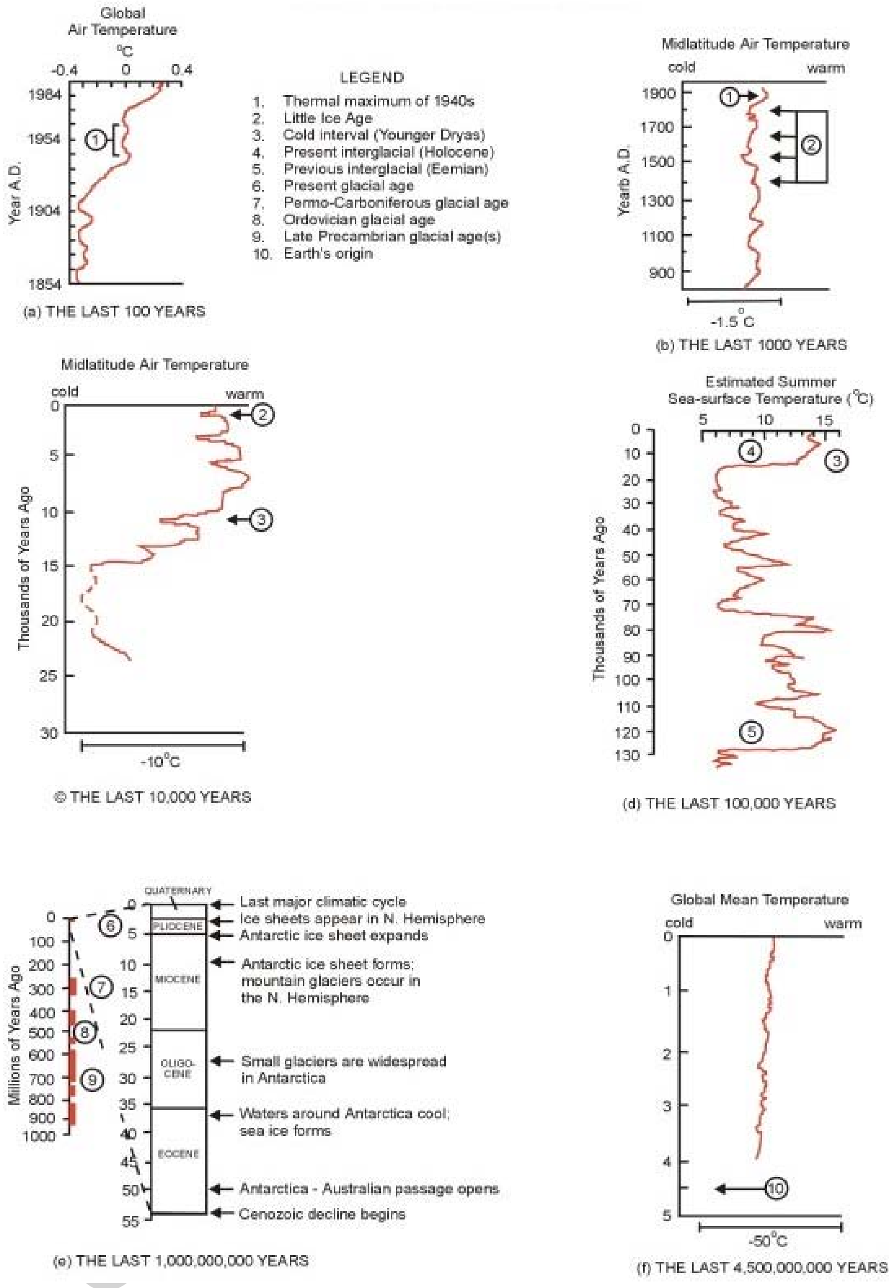

Figure 2: The climate of the Earth as represented by average air temperature records over periods (a) 100 years (data from Jones, 1995), (b) 1000 years (data from Hughes and Diaz, 1994), (c) 10 000 years (data from US National Academy of Sciences, 1975), (d) 100 000 years (data from Berger, 1995), (e) 1 000 000 000 years (data from Berger, 1988 and Barron, 1995) and (f) 4 500 000 000 years (data estimated from many sources including Kump and Lovelock, 1995 and Henderson-Sellers, 1989). Dashed lines indicate that the proxy records are inadequate for the reconstruction portrayed. *(after Henderson-Sellers, 1995)*

	Mercury	**Venus**	**Earth**	**Mars**	**Jupiter**	**Saturn**	**Uranus**	**Neptune**	**Pluto**
Mean distance from Sun (millions of kilometers)	57.9	108.2	149.6	227.9	778.3	1427	2869.6	4496.6	5900
Period of revolution	88 days	224.7 days	365.26 days	687 days	11.86 years	29.46 years	84.01 years	164.8 years	247.7 years
Rotation period	59 days	-243 days (retrograde)	23 hours 56 min 4 s	24 hours 37 min 23 s	9 hours 50 min 30 s	10 hours 14 min	-11 hours (retrograde)	16 hours	9 hours
Inclination of axis	<28°	3°	23° 27ft	23° 59ft	3° 05ft	26° 44ft	82° 5ft	28° 28ft	17° 10ft
Equatorial diameter (km)	4880	12104	12756	6787	142800	120000	51800	49500	6000 (?)
Mass (Earth = 1)	0.055	0.815	1	0.108	317.9	95.2	14.6	17.2	0.0017
Density (water = 1)	5.4	5.2	5.5	3.9	1.3	0.7	1.2	1.7	?
Atmosphere (main components)	None	Carbon dioxide	Nitrogen, oxygen	Carbon dioxide, argon (?)	Hydrogen, helium	Hydrogen, helium	Hydrogen, helium, methane	Hydrogen, helium, methane	Methane

Table 2(a). Astronomical and physical characteristics of the planets (modified from Henderson-Sellers, 1983)

	Venus	**Earth**	**Mars**
CO_2 (%)	98	0.03	95
N_2 (%)	1.7	79	2.7
O_2 (%)	Trace	21	0-.13
H_2O (meters)	0.003	3000	0.00001
Pressure (bars)	96	1	0.0064
Temperature (°C)	477	17	-47

Table 2(b). Anomalous atmosphere of Earth cf. other terrestrial planetary atmospheres (modified from Margulis and Kinkle, 1991)

Gas	**Abundance**	**Expected Equilibrium Concentration ***	**Discrepancy**	**Residence Time (yr)**	**Output (10^6 tons yr^{-1})**
Nitrogen	0.8	10^{-10}	10^{9}	3×10^{6}	1000
Methane	1.5×10^{-6}	$<10^{-35}$	10^{29}	7	2000
Nitrous oxide	3×10^{-7}	10^{-20}	10^{13}	10	600
Ammonia	1×10^{-8}	$<10^{-35}$	10^{27}	0.01	1500
Methyl iodide	1×10^{-12}	$<10^{-35}$	10^{23}	0.001	30
Hydrogen	5×10^{-7}	$<10^{-35}$	10^{23}	2	20

*Assumes 20% oxygen

Table 2(c). The climate anomaly (after Margulis and Kinkle, 1991)

The Earth’s climate is also remarkable, at least to human perception, both for its benign character and for its long-term stability. The oldest known record of the climatic conditions on Earth comes from the Isua, only about 670 million years younger than the Earth itself. These rocks indicate the presence of liquid water on the surface. From about 3.5 billion years ago, the fossil record includes increasingly abundant evidence of life, which is usually interpreted as indicating that globally-averaged climatic conditions, such as temperature and precipitation, were not too different from the present day. Figure 2(f) shows mean temperatures rising slightly over the whole of Earth's history because we know that solar luminosity has increased by about 30 percent since the Earth was formed. However, this curve is only an interpretation of the probable record of the Earth's global mean temperatures, based partly on observational evidence and partly on results from climate models which indicate that negative feedback effects can reduce large swings, while "recovery" of habitable conditions from either a total planetary freeze or a "runaway greenhouse" is virtually impossible.

The very long-term stability of the climate system has been attributed to the self-rectifying feedbacks that seem to exist in the fully-coupled geophysiological system that is the Earth. The very impressive impact of the biosphere on the climate system that results in the anomalous features of the Earth's climate system (Table 2(c)) have been described in terms of a planet-wide phenomenon: the "Gaia hypothesis."

The Gaia or geophysiology hypothesis of the Earth attempts to consider the responses to known external forcings in terms of a holistic view of the planet. In this system, the biota plays an active part in many feedbacks, tending to dampen extreme variations and causing the Earth to continue to be hospitable to life. A number of possible mechanisms of biospheric feedback on the components of the climate system have been proposed. For example, there is the proposal regarding the involvement of CO_2 and the biogenic weathering of rock in global temperature regulation (Figure 3(a)). The terrestrial and marine biota is sensitive to temperature change and reacts by changing rates of CO_2 uptake directly or indirectly from the atmosphere. This uptake, if combined with subsequent long-term burial, modifies the rate of CO_2 removed. Similarly, biologically-induced weathering of silicate rocks can prompt CO_2 removal rate changes. The atmospheric CO_2 concentration then determines the temperature via the global greenhouse and the loop is closed.

The most fully described Gaia hypothesis of biogenic influence on the environment involves the altering of cloud albedo over the open oceans by algae producing dimethylsulfide (DMS) (Figure 3(b)). The phytoplankton in this process "protects" them from excessive insolation by increasing the brightness of clouds, which form over them. In order for negative feedback to be established, warmer seas or greater irradiance below the clouds are assumed to stimulate phytoplankton to flourish and increase their production of DMS. Although the processes in the main feedback loop have not yet been measured with enough accuracy to determine if the climate system is actually functioning to produce a self-adjusting response, the questions are well posed. Major uncertainties have been identified, including whether the response of marine plankton to increased sea surface temperature or insolation causes an increase or decrease in the total DMS production, and whether the relative production of cloud droplets by cloud condensation nuclei from biologically-generated DMS is greater than by salt particles occurring without biogenic input in the marine atmosphere. Detailed observations have found a seasonal and latitudinal correlation between DMS fluxes and the numbers of cloud nuclei in the marine atmosphere. This and other testable propositions derived from the Gaia hypothesis have been instrumental in focusing attention on the potential importance of biogenic forcing of the climate system.

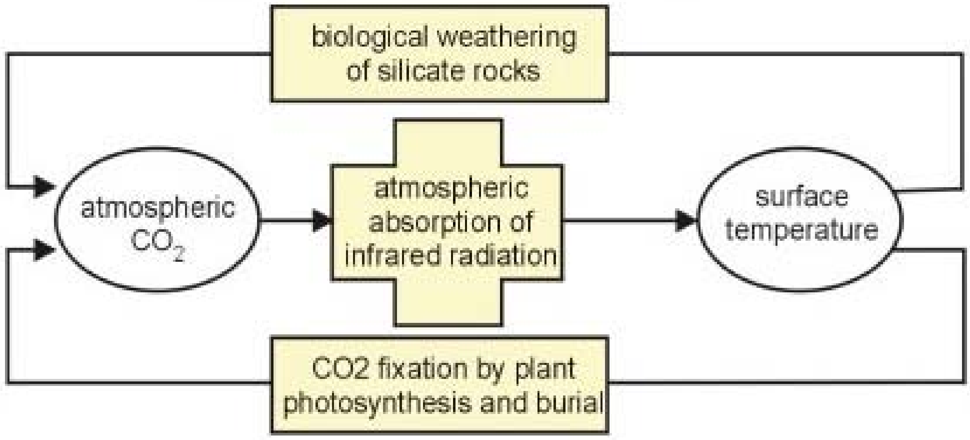

Figure 3 (a): Schematic showing possible feedback loops linking the Earth's surface temperature to biological controls of the atmospheric CO_2 concentration. Physical quantities are shown in ovals and processes that modify them in shaped and shaded boxes displaying the sign of the change caused in the quantity immediately downstream of the shaped box in response to an increase in the upstream oval quantity (modified from Shearer, 1991).

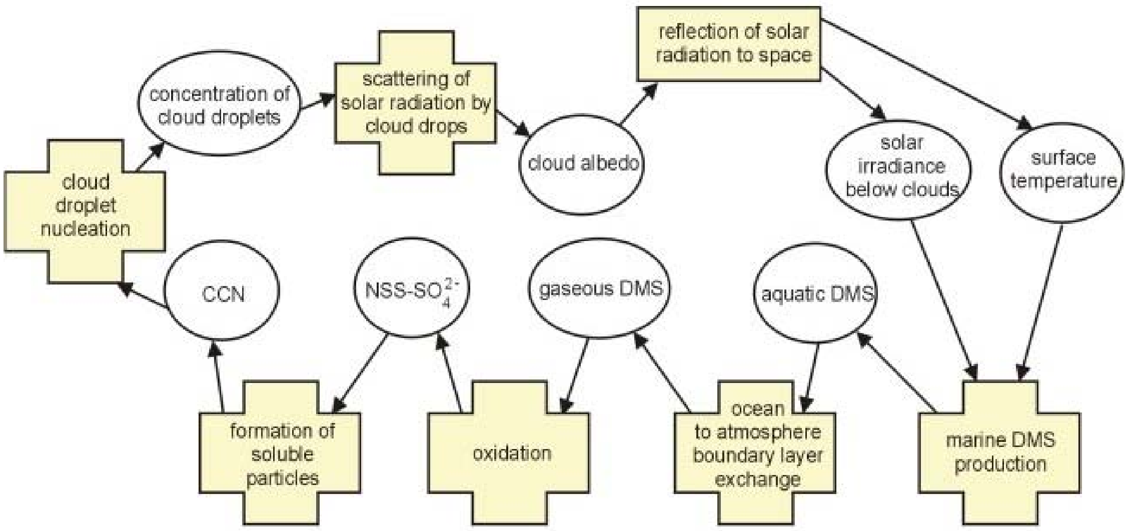

Figure 3(b): Proposed components of a biological feedback involving DMS production by marine phytoplankton to control oceanic surface temperatures and/or incident solar insolation (modified from Charlson et al., 1987 and Shearer, 1991). The same convention as in (a) is used for quantities, processes and their proposed Gaian impacts.

1.3 Forcing and Feedbacks

The state of the climate system at any time is determined by the forcings acting upon it, and the complex and interlocking internal feedbacks that these forcings prompt. In the broadest sense, a feedback occurs when a portion of the output from an action is added to the input, so that the output is further modified. The result of such a loop system can either be an amplification (a positive feedback) of the process, or a dampening (a negative feedback): positive feedbacks enhance a perturbation whereas negative feedbacks oppose the original disturbance.

1.3.1 Snow and Ice: A Surface Feedback

If some external perturbation (say an increase in solar luminosity) acts to increase the global surface temperature, then snow and ice will melt and their overall areas reduce in extent. These cryospheric elements are bright and white (i.e. their albedo, the ratio of reflected to incident radiation, is high), reflecting almost the entire solar radiation incident upon them. The surface albedo, and probably the planetary albedo (the reflectivity of the whole atmosphere plus surface system as seen from "outside" the planet), will decrease as the snow and ice areas reduce. As a consequence, a smaller amount of solar radiation will be reflected away from the planet and more absorbed so that temperatures will increase further. A further decrease in snow and ice results from this increased temperature and the process continues. This positive feedback mechanism is known as the ice-albedo feedback mechanism.

This feedback mechanism is also positive if the initial perturbation, say a reduction in solar luminosity, causes a decrease in global surface temperatures. With lower temperatures, the areas of snow and ice are likely to increase, thus increasing the albedo and leading to a further decrease in temperature because more solar radiation is reflected away and the cooling spiral continues.

1.3.2 Water Vapor: An Atmospheric Feedback

A similar example can illustrate another positive feedback mechanism in the climate system. Suppose that planetary temperatures begin to increase, perhaps because there is less snow one year. This prompts an increase of atmospheric water vapor resulting from increased evaporation as ocean and land temperatures are raised. Water vapor contributes to the greenhouse warming of the surface as a result of its absorption of infrared radiation emitted from the surface. The additional greenhouse effect of the extra water vapor enhances the temperature increase: a positive feedback effect. Once again it is also positive (i.e. enhancing the initial disturbance) in the other direction. If temperatures fall, water vapor will condense out of the atmosphere, thus reducing the atmospheric greenhouse effect and causing a further cooling.

1.3.3 Biogenic Feedbacks in the Climate System

As well as physical and chemical feedbacks in the climate system, there are biogenic processes, which can act both to enhance and to diminish disturbances. Negative feedback processes associated with biological changes in response to fluctuations in atmospheric CO_2 are illustrated in Figure 3(a). If CO_2 levels rise (or fall), the planet becomes warmer (or cooler), which increases (decreases) both the biological weathering of silicate rocks, and the fixation of CO_2 by plants and its subsequent burial. Both these effects reduce (increase) atmospheric CO_2, thus providing negative feedbacks, i.e. the feedbacks tend to dampen the initial disturbance.

Another negative biogenic feedback within the climate system is illustrated in Figure 3(b). This feedback loop evokes the production of more (or less) dimethylsulfate (DMS) by marine phytoplankton as a means by which the local sea surface temperatures and/or the incident solar irradiation at the ocean surface can be returned to an earlier state following a disturbance. The mechanism involves DMS oxidation to water-soluble particles, which become cloud condensation nuclei (CCN). The production of more (fewer) CNN by the marine phytoplankton results in higher (lower) cloud albedos, and hence increased (decreased) reflection of solar radiation, reducing (increasing) the initial disturbance to ocean surface conditions.

1.3.4 Combining Climate System Feedbacks

Since many feedback effects operate within the climate system at any point in time in response to a variety of perturbations, it is important to understand the way in which these feedbacks combine. For example, consider a system in which a change of surface temperature of magnitude δT is introduced. Given no internal feedbacks, this temperature increment will represent the whole change in the system temperature. If feedbacks occur then there will be an additional temperature change, and the feedback effects will augment the new value of the system's surface temperature change:

$$\delta T_{\text{system}} = \delta T + \delta T_{\text{feedbacks}} \tag{1}$$

where $\delta T_{\text{feedbacks}}$ can be either positive or negative. The value of δT_{system} can be related to the perturbation, which caused it to provide a measure of the sensitivity of the climate system to such a disturbance. A commonly-used climate sensitivity parameter, λ, can be defined in terms of a perturbation in the global surface temperature δT, which occurs in response to an externally-generated change in planetary net radiative flux, δQ:

$$C[\partial(\delta T)/\delta t] + \lambda\,\delta T = \delta Q \tag{2}$$

Here $\lambda\delta T$ is the climate system's net radiation change resulting from the internal dynamics of the system, t is time, and C represents the climate system's heat capacity. Although Equation (2) represents an extreme simplification of the system it is very valuable for interpreting and summarizing the sensitivity of the overall climate system, and also provides a means of calibrating and intercomparing sensitivities of numerical climate models. A convenient reference value for this climate sensitivity parameter λ is the value λ_B, which λ would have if the Earth were a simple black body with its present-day albedo, i.e. the value if the Earth is assumed to radiate perfectly and absorb incident solar radiation in an amount controlled only by its present day reflectivity:

$$\lambda_B = 4\sigma T_e^3 \tag{3}$$

where σ is the Stefan–Boltzmann constant and T_e is the Earth's effective temperature. This effective temperature, T_e, can be determined from establishing a simple radiative balance for the Earth. That is, taking S, the incident solar flux, to be 1370/4 W m^{-2}, and A, the Earth's albedo, to be 0.3, giving:

$$S\ (1 - A) = \sigma T_e^4 \tag{4}$$

This gives T_e = 255K and hence λ_B = 3.75 W m^{-2} K^{-1}. The overall climate system sensitivity parameter λ_{system} is based around this value, being composed of the summation of λ_B and all contributing feedback factors λ_i. Therefore, consideration of the feedbacks described in **1.3** from water vapor, ice reflectivity, and biogenic processes, gives:

$$\lambda_{system} = \lambda_B + \lambda_{water\ vapor} + \lambda_{ice\ albedo} + \lambda_{biogenic} \tag{5}$$

A positive value of the sum of all the feedback factors (λ_i) implies stability or negative feedback, while a negative value of feedback factors implies positive feedback and the possibility of growing climate instability. Climate models can be compared directly by comparing the values they generate for λ_{system}.

Alternatively, the climate system change resulting from a particular perturbation can be calculated by multiplying the inverse of the value of λ_{system} by the value of δQ. This relationship (derived directly from Equation (2) for the case of zero temperature change) where $\delta T = \delta Q/\lambda$ has given rise to the definition of the climate system feedback factor as $1/\lambda$ or λ'. It is this sensitivity parameter which has been used as a measure of the sensitivity of climate models in the assessment of human-induced greenhouse warming, and in some recent climate model intercomparisons including those by the IPCC.

For doubling atmospheric CO_2 it has been shown that δQ = 4.2 W m^{-2}. Using λ_B = 3.75 W m^{-2} K^{-1} and taking $\lambda_{water\ vapor}$ = -1.7 W m^{-2} K^{-1} and $\lambda_{ice\ albedo}$ = -0.6 W m^{-2} K^{-1}, λ_{system} = 1.45 W m^{-2} K^{-1} so that the globally-averaged temperature rise due to doubling atmospheric CO_2 would be about 2.9 K. If the ice–albedo feedback had been neglected, the predicted temperature increase would have been only about 2.0 K. Various estimates have been made of the feedback effects likely to be caused by biogenic feedbacks. The addition of such feedbacks to those considered above could raise the surface temperature increase due to doubling CO_2 or reduce it to effectively a zero response, depending on the anticipated sign (negative or positive) (cf. biogenically induced feedbacks).

From these examples, it can be seen that the importance of both forcings and feedback effects in the climate system depends upon the time-scale of behavior of the subcomponents each affects. This concept of time-scale of response, variously referred to as the equilibration time, the response time, the relaxation time, or the adjustment time, is crucially important to all aspects of the climate system (Table 1). It is a measure of the time each part of the climate system takes to re-equilibrate following a perturbation to it. A short equilibration time-scale indicates that the subsystem responds very quickly to perturbations and can therefore be viewed as being quasi-instantaneously equilibrated with an adjacent subsystem, which possesses a much longer equilibration time. These concepts are important both for interpreting past climates and in the construction of models for the prediction of future climates.

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Biographical Sketch

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