

is natural to call such species *species-repairers* (see Section 6.7). Evidently, species-repairers should achieve high population numbers and high population density immediately after disturbances and exist in lower numbers in the absence of disturbances. External disturbances do not occur at regular intervals. If a time interval between two consecutive disturbances (e.g. fires) becomes too long, the 'unemployed' species-repairers may lose their genetic stability, as soon as the process of stabilising selection is most efficient in dense populations. Thus, in order to 'train' species-repairers, natural biota has to support *species-disturbers* that disturb the ecosystem at regular intervals and, by doing so, periodically activate species-repairers, preventing them from genetic decay. Primitive man was among such species-disturbers together with many other large animals (e.g. elephants making large lawns amidst tropical forests). This issue has been previously discussed in Sections 4.7 and 6.7.

Homo sapiens has preserved the strategy of behaviour based on the genetic programme of positive and negative emotions inherited from large animals. Natural ecological role of man in the biosphere was to use fire and other means (e.g. stubbing up trees to get arable land) in order to introduce periodical disturbances of local ecosystems and activate species-repairers. Thus, the genetically-based programme of environmental impact of humans is aimed towards transformation of the natural environmental conditions (e.g. forest) into more favourable artificial ones (e.g. field), i.e. creation of an artificial 'internal' environment forcing out a less favourable external environment. This can be regarded as a further advancement of the principle of large body size (i.e. when more and more space is occupied by the internal milieu of the organism), which, as explained above, reduces the ability of *Homo sapiens* as a species to regulate the external environment. Cultural information accumulated by humankind in the course of historical development of civilisation has enabled people to impose a huge impact on the biosphere. But, as ever, the strategy of this impact is confined to the genetic programme of behaviour of *Homo sapiens*. Thus, most human–environment interactions result in large-scale perturbations of the natural biota and the forcing out of the natural 'wild' environment. It is this unprecedented scale of these processes that is now threatening the future existence of humanity.

8

Unique Nature of Climate Stability on Earth

In this chapter evidence is presented in favour of the statement that Earth's modern climate is physically unstable. In the absence of natural biota it should change spontaneously to one of the two possible physically stable states—that of complete glaciation of the Earth's surface and that of complete evaporation of the Earth's hydrosphere, both of which are life-incompatible. The existing climate has remained suitable for life during the last four billion years. It means that natural biota of Earth is the only mechanism that keeps the Earth's climate in a state suitable for life.

8.1 MAJOR CLIMATIC CHARACTERISTICS OF EARTH

All biogeochemical cycles in the environment represent steady ordered processes maintained at the expense of solar energy. A steady state of all substances cycling in the environment means that masses and concentrations of all substances remain, on average, constant in all local areas of the environment despite continuous exchange of substances between different areas. It means that for all substances sources are equal to sinks in all areas of the environment. If this equality is broken, masses and concentrations of corresponding substances begin to change. Rates of such changes are determined by properties of the corresponding sources and sinks.

Atmospheric oxygen and carbon dioxide are in physical thermodynamic equilibrium with oxygen and carbon dioxide dissolved in rivers, lakes, seas and oceans. On average, water solutions of these two substances are saturated. Solubility of these two gases increases with decreasing temperature. Thus, in polar regions there is a net flux of these gases from the atmosphere to the ocean, while in equatorial regions there is a reverse flux of gases from the ocean to the atmosphere. Physical cycles of oxygen and carbon dioxide depend mainly on the average temperature of the Earth's surface and on the difference between average temperatures of polar and equatorial regions. These two parameters are determined by the total amount of energy received from the Sun and in practice do not depend on informational characteristics of solar energy that are determined by the difference between temperatures of Earth and Sun

(see Chapter 7). Thus, on a global scale, physical cycling of oxygen and carbon dioxide can be described by traditional thermodynamics.

Water is present in the biosphere in all the three possible phases—solid, liquid and gaseous. Conditions for thermodynamic equilibrium between different water phases are met only in very small local areas. On a global scale, the three phases of water are not in physical thermodynamic equilibrium. If this were the case, water vapour would be present in the atmosphere in a saturated concentration. In reality, the observed global mean value of relative humidity, which is equal to the relative difference between the saturated and real concentration of water vapour, is less than 100%.

As noted in Section 6.6, atmospheric water vapour concentration changes hundreds of times in different regions of the planet. The observed non-random unevenness of distribution of atmospheric water vapour over the Earth's surface represents a departure from the chaotic state of thermodynamic equilibrium. Its ultimate cause is the orderliness of the solar radiation. Unevenness of water vapour distribution is a prerequisite for the alternating processes of evaporation and precipitation that are vitally important for the terrestrial biota.

Thus, for some substances (e.g. oxygen and CO₂) the steady state of cycling is close to physical thermodynamic equilibrium and is totally determined by Earth's average temperature and atmospheric pressure. For other substances (e.g. water) the steady state of cycling is far from physical thermodynamic equilibrium. Cycling of such substances represents a highly-ordered process and depends on the degree of orderliness of the solar energy, that is, on the difference between the Sun's and Earth's temperatures (Chapter 7).

Some of the most vital questions of modern fundamental ecology are as follows. What factors determine the average temperature of the Earth's surface? Are concentrations and masses of globally cycling substances constant and to what accuracy? Is the observed state of global environment a stable one? Are there any other stable states of Earth's environment? Are they suitable for life? Can the present-day global environment of Earth transform into one of the other possible stable states and how much time will it take? We address below some of these issues.

The orbital position occupied by the Earth in the solar system determines the flux of the solar energy incident upon the planet outside the atmosphere. It is called the solar constant I (Table 8.1). About one third of this flux is reflected back into space, predominantly by clouds in the atmosphere and by ice cover at the Earth's surface. The relative part of solar radiation reflected back into space is called the planetary albedo (Table 8.1). Due to non-zero values of albedo, Earth and other planets of the solar system are visible in space. Solar radiation absorbed by the Earth's surface generates cycling processes in the biosphere and finally converts into heat, that is, thermal radiation. In the atmosphere thermal radiation of the Earth's surface is trapped by the so-called greenhouse gases. Molecules of the greenhouse gases emit thermal photons in all possible directions, so that about half of this radiation returns to the Earth's surface where it is re-emitted to the atmosphere and so forth. This phenomenon is known as the greenhouse effect and can be quantified as the relative part of thermal radiation of the Earth's surface that is effectively

Table 8.1. Planetary energy and temperature characteristics (after Mitchell, 1989).

Planet	Solar constant I , W/m ²	$A = B = 0$	$A > 0, B = 0$	$A > 0, B > 0$		
		(Orbital temperature t_o , °C)	(Effective planetary temperature t_e , °C)	(Average values at planetary surface) B , % t , °C		
Mars	589	-48	15	-56	7	-53
Venus	2613	+58	75	-41	99	+460
Modern Earth (2)	1367	+5	30	-18	40	+15
Earth, if total ice cover (1)	1367	+5	80	-90	7	-85
Earth, if total evaporation of oceans (3)	1367	+5	75	-80	99	+400

I is the solar constant, i.e. the power of solar radiation incident upon unit of perpendicular area outside the planet's atmosphere. $I/4$ is the average power of solar radiation incident upon unit of Earth's surface area, see Section 8.5. Orbital temperature t_o (degrees Celsius) is uniquely determined by I , i.e. by the orbital position occupied by the planet in the solar system. Under the approximation of the black body radiation, I and t_o are connected by the expression $T_o = (I/4\sigma)^{1/4}$, where $T_o = t_o + 273$ is the absolute orbital temperature in degrees Kelvin, $\sigma = 5.67 \cdot 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$ is the Stephan-Boltzmann constant.

A is the planetary albedo, i.e. fraction of solar radiation reflected by the Earth back to space. The remaining fraction of solar radiation, $(1 - A)I/4$, is absorbed by the planet and is equal to the planet's heat emission to space. This flux can be directly measured from outer space by artificial satellites. It determines the effective planetary temperature t_e as $T_e = [(1 - A)I/4\sigma]^{1/4} = T_o(1 - A)^{1/4}$, where $T_e = t_e + 273$. B is the relative greenhouse effect, i.e. the fraction of the planet's heat radiation that is reflected by the atmosphere back to the planet's surface. The heat flux at the Earth surface q is equal to $q = (1 - A)I/4 + Bq$, that is $q = (1 - A)I/4(1 - B)$. Accordingly, temperature at the Earth's surface t (degrees Celsius) is determined by values of planetary albedo and greenhouse effect: $T = (q/4\sigma)^{1/4} = T_o[(1 - A)/(1 - B)]^{1/4}$, where $T = t + 273$.

In the state of total ice cover of Earth, the value of planetary albedo is taken as equal to the albedo of ice and snow sheets, the value of greenhouse effect was taken as on Mars. In the state of total evaporation of oceans on Earth, the values of albedo and greenhouse effect are taken as on Venus. Note that changes of A and B lead to drastic changes of temperature on the planet's surface.

reflected back to the Earth's surface by the atmosphere. Modern atmosphere of Earth effectively reflects about 40 percent of the thermal radiation, while on Venus this effect amounts to 99 percent (Table 8.1).

Temperature of the Earth's surface is uniquely determined by the cumulative flux of thermal radiation from the Earth's surface, which is proportional to the fourth power of absolute surface temperature. Irrespective of what orbital position is occupied by the planet, this flux is totally determined by the planetary albedo and greenhouse effect (see notes to Table 8.1). Thus, the planet's surface temperature can assume almost arbitrary values depending on values of albedo and greenhouse effect, the latter being completely determined by inherent environmental characteristics of the planet. In other words, the orbital position does not guarantee a suitable-for-life environment on the planet.

Physical density and gaseous composition of the atmosphere determine atmospheric absorption of less than one-third of the incoming solar radiation. Thus, the majority of solar radiation reaches the Earth's surface making possible photosynthesis and, therefore, life of the whole biota of Earth. If atmospheric density were several times higher than at present, sunlight could not practically reach the Earth's surface which would remain dark, similar to the present situation at oceanic depths. Under such conditions, no modern life would then be possible on the Earth's surface. If, on the contrary, atmospheric density were several times lower than it is now, the atmosphere would feature some life-incompatible characteristics. In particular, concentration of oxygen would be insufficient for breathing, while concentration of CO₂ would be insufficient for photosynthesis (Brody, 1945; Larcher, 1980).

The gaseous composition of the atmosphere is no less important for life than the atmospheric density. Modern atmosphere is relatively transparent to the incident sunlight, but opaque with regard to thermal radiation of the Earth's surface. Absorption of thermal radiation is performed by greenhouse gases. Gases absorb and emit radiation at frequencies that correspond to transitions between discrete energy levels of their molecules. Transitions between some energy levels of molecules of greenhouse gases are in resonance with characteristic frequencies of the thermal radiation of Earth. The cumulative mass of all greenhouse gases is less than two thousandths of the total mass of the atmosphere. The major greenhouse gas is water vapour. Its volume content in the atmosphere is less than 0.3 percent. The second important greenhouse gas is carbon dioxide with a volume content of about 0.03 percent. The remaining part of the greenhouse warming is determined by methane, nitrous oxide and ozone. Their cumulative volume content does not exceed 0.0003 percent (Mitchell, 1989). Note also, that ozone which is present in the atmosphere in negligibly small concentrations, ensures an almost complete absorption of ultraviolet radiation harmful to terrestrial life.

Thus, we have seen that the properties of the modern atmosphere are in a unique concert with requirements of life. Minor changes in atmospheric properties can be fatal to any life on Earth.

In a stationary state—that is, when the average energy content of the planet remains constant over time—the power of the thermal radiation released by Earth into space is naturally equal to the power of the solar radiation absorbed by the planet. The latter is totally determined by the solar constant and the planetary albedo (see notes to Table 8.1). Thus, thermal radiation released by Earth into space does not depend on the greenhouse effect and is characterised by the so-called effective planetary temperature (Table 8.1). This temperature characterises the upper radiative layers of the atmosphere from which the thermal radiation is emitted directly into space. Thus, the effective planetary temperature can be measured from outer space. One cannot judge about the temperature of the planet's surface from the effective planetary temperature.

The greenhouse effect increases temperature of the planet's surface as compared with its effective planetary temperature. It leads to formation of a temperature gradient between the planet's surface and the upper radiative layers of the

atmosphere. A well-known manifestation of this gradient is the existence of eternal snow caps on tops of high mountains. This gradient, so-called lapse rate, constitutes 5.5°C km⁻¹ and remains nearly constant up to about 6 km above the sea level. At that altitude the temperature difference between the Earth's surface and the atmosphere equals 33°C, which gives the magnitude of the greenhouse effect measured on a temperature scale. The atmospheric layer at height of about 6 km represents an effective radiating surface that determines thermal radiation of the planet emitted directly into space.

In the stratosphere the density of atmospheric gases rapidly decreases with height, while the temperature of these rarefied gases increases and exceeds not only the temperature of the effective radiating tropospheric layer, but even that of the Earth's surface. As soon as the thermal radiation of Earth measured from outer space is actually emitted by molecules of the upper tropospheric radioactive layers, the increased temperature of the stratosphere means that molecules of the rarefied stratospheric gases are not in thermal equilibrium with long-wave (thermal) radiation of the planet. This means that the stratosphere and processes taking place there impose no considerable impact on the heat balance of the planet and cannot influence the climate of Earth. The fact that the thin ozone layer absorbs ultraviolet radiation is essential for life on the planet, but does not seriously influence the heat balance of the planet.

In the absence of greenhouse gases the average temperature of the Earth's surface would be equal to the effective planetary temperature of the planet, $t_e = -18^\circ\text{C}$ (Table 8.1). There would be no temperature gradient present in the atmosphere. The snow cover of high-latitude territories would gradually spread down over the whole Earth's surface. Thus, the planetary albedo would increase up to the value characteristic of snow and ice, that reflect about 80% of the solar radiation (Mitchell, 1989). As a result, the average global temperature would drop to about -90°C (Table 8.1), a state where no life is possible.

Hydrosphere of Earth is sufficient for total glaciation of the Earth's surface. At the same time, Earth's hydrosphere is the main source of atmospheric water vapour, the major greenhouse gas. Absolute concentration of water vapour in the atmosphere, as well as the relative humidity, vary greatly both in space and time, changing several times as compared with their average values. However, the average concentration of the water vapour changes proportionally to its saturated concentration (the maximum possible concentration under given conditions). Saturated concentration of water vapour grows exponentially with increasing temperature in accordance with Boltzmann distribution and the Clausius-Clapeyron equation (Landau *et al.*, 1965). The concentration of water vapour is negligibly small at low temperatures. With increasing temperature, concentration of water vapour grows exponentially and causes an increase of the greenhouse effect that leads to further increase of temperature and so forth. At 100°C water vapour pressure becomes equal to the atmospheric pressure and water begins to boil. At higher temperatures the atmospheric pressure is totally determined by the pressure of water vapour, which continues exponential growth with increasing temperature (Landau *et al.*, 1965). The process ends in a state of complete evaporation of the hydrosphere, when processes

of evaporation and condensation of clouds occur in the atmosphere and do not pertain to the Earth's surface. As a result, the equilibrium temperature of the Earth's surface jumps up to several hundred degrees Celsius due to the catastrophic increase of greenhouse warming, as on Venus (Table 8.1). In this state life is not possible either.

The existing positive feedback between the surface temperature and atmospheric water vapour concentration points to the physical instability of the current climate of Earth, where hydrosphere is maintained predominantly in the liquid phase (water of oceans, rivers, lakes). The solid phase (ice of glaciers) constitutes less than 14% and the gaseous phase (atmospheric water vapour) less than two millionth parts of the total amount of liquid water which is mostly concentrated in the oceans (Lvovitch, 1974), see Table 6.1.

When values of albedo and greenhouse effect are temperature-independent, the surface temperature of the planet $T = T_s$ is stable. It is determined by the balance of the absorbed solar radiation and the emitted thermal radiation. The stability of stationary surface temperature T_s is ensured by the following negative feedback. When the surface temperature accidentally increases, $T > T_s$, the emitted flux of thermal radiation increases as well, according to the Stephan-Boltzmann law. As a result, the planet loses more energy than it receives and cools down back to T_s . When the surface temperature decreases, $T < T_s$, the planet loses less energy than it receives. As a result, temperature increases back to T_s .

There are two physically stable states where values of albedo and greenhouse effect remain constant in a wide temperature interval. These are the state of total glaciation of the Earth's surface at temperatures about -90°C and the state of total evaporation of the Earth's oceans at temperatures close to 400°C (Table 8.1). In both states life is impossible. Constancy of albedo and greenhouse effect in these states is determined by the fact that in both states water exists predominantly in only one phase—solid at low and gaseous at high temperatures.

Under modern climatic conditions water exists in all three phases. Values of albedo and greenhouse effect depend on temperature. For example, with increasing temperature the ice shields melt and planetary albedo decreases, while atmospheric water vapour concentration grows and the greenhouse effect increases. Stable existence of life during the last four billion years gives unambiguous evidence that the modern, suitable-for-life state of the Earth's climate is stable and spontaneous transitions to both lifeless states are forbidden. The degree of stability of the modern climate depends on the rates of changes of albedo and greenhouse effect with temperature. This is discussed in sections to follow.

8.2 SPECTRAL CHARACTERISTICS OF THERMAL RADIATION

Spectral distribution of energy over radiation wave frequencies (or inverse wave numbers) is the main characteristic of any flux of radiation. Those parts of the Earth's surface that are not covered by ice or snow absorb almost all incident

radiation. It means that in a stationary state, when the surface temperature remains constant, the energy spectre of radiation emitted by the Earth's surface does not depend on properties of its molecules, but is determined by temperature alone. Such radiation is known as the black-body radiation and is described by the well-known Planck function.

Unlike the Earth's surface, atmosphere only absorbs thermal radiation at certain spectral intervals defined by the presence of absorption bands of the greenhouse gases. Thus, thermal radiation of the atmosphere cannot be reduced to black-body radiation depending on the gas temperature alone. Rather, thermal radiation of the atmosphere depends on concentrations of greenhouse gases and their height distribution. In order to evaluate temperature dependence of the greenhouse effect, it is necessary to understand how radiating properties of the atmosphere depend on concentrations of the greenhouse gases.

Energy spectre of thermal radiation of the Earth's surface $q(T)$ over radiation frequencies ω can be written as follows:

$$q(T) = \sum_i \Delta\omega_i I(\omega_i, T) = \Delta\Omega \bar{I}(\bar{\omega}, T) = \sigma T^4 \quad (8.2.1)$$

Here $I(\omega, T)$ is the observed spectral density of thermal radiation emitted from the Earth's surface, which is close to the Planck function; T is the observed global mean absolute temperature of the Earth's surface. According to the properties of the Planck function, the effective width of the spectre $\Delta\Omega$ grows linearly with increasing T , while the maximum and the mean values of $I(\omega, T)$ increase proportionally to the third power of T . Thus, $q(T)$ grows proportionally to T^4 . Proportionality coefficient σ is close to the Stephan-Boltzmann constant characterising black-body radiation. The average frequency $\bar{\omega}$ grows linearly with T .

Unlike the Earth's surface which absorbs the most part of the incoming radiation, atmospheric gases absorb radiation at some frequencies but allow radiation at other frequencies to pass through unimpeded. Greenhouse gases absorb thermal radiation in those parts of the spectre (8.2.1) that correspond to their molecular absorption bands. The latter are determined by molecular properties of the gases and only weakly depend on atmospheric temperature and density. In some parts of the spectre (8.2.1) there are no absorbers at all or their concentration is very low, so that the absorption of thermal radiation in the atmosphere in such parts of the spectre is not complete. Such parts of the spectre are called spectral windows. Thermal radiation of the Earth's surface emitting into space through spectral windows is characterised by spectral intensity $I(\omega, T)$ determined by the surface temperature T . Thermal radiation of the planet measured in the outer space, q_e , can be thus represented as a sum of the spectral window radiation and radiation emitted from the upper radiating layer of the atmosphere with temperature T_e :

$$q_e(T_e) = \sum_{i \neq k} \Delta\omega_i I_e(\omega_i, T_e) + \sum_k \Delta\omega_k I(\omega_k, T) \approx \sigma_e T_e^4, \quad (8.2.2)$$

where frequency intervals $\Delta\omega_k$ stand for spectral windows. Here $I_e(\omega, T_e)$ is the observed spectral intensity of the upper radiating layer of the atmosphere. Note that $I_e(\omega, T_e)$ can be substantially different from the Planck function $I(\omega, T)$, because the atmosphere is not a black-body radiator. In the modern atmosphere only a minor part of the terrestrial radiation escapes directly into space through the spectral windows (Barry and Chorley, 1987). This makes it possible to consider q_e approximately as a function of T_e only, neglecting the minor contribution of the second sum in (8.2.2). The constant σ_e in (8.2.2) may differ from the Stephan-Boltzmann constant by several tens of percent.

8.3 TRADITIONAL ESTIMATES OF THE CONTRIBUTIONS FROM DIFFERENT GREENHOUSE GASES TO THE GREENHOUSE EFFECT

The observed global mean fluxes of terrestrial radiation $q(T)$, atmospheric radiation $q_e(T_e)$ and their difference $f \equiv q_e - q$ are as follows:

$$\begin{aligned} q &= \sigma T^4 = 390 \text{ W m}^{-2}, & T &= 288 \text{ K } (15^\circ\text{C}), \\ q_e &= 240 \text{ W m}^{-2} \approx \sigma_e T_e^4, & T_e &\approx 255 \text{ K } (-18^\circ\text{C}), \\ f &= q - q_e = 150 \text{ W m}^{-2}. \end{aligned} \quad (8.3.1)$$

Note that f has a meaning of absolute greenhouse effect in energy units (greenhouse forcing).

In a stationary state the energy lost by the Earth back into space, q_e , is balanced by the energy received by the Earth from the Sun, i.e. q_e is completely determined by values of I and A , $q_e = (I/4)(1 - A)$. The atmosphere is relatively transparent to the solar radiation, which is transformed into heat predominantly at the Earth's surface. It is therefore q_e that serves as the primer radiation flux. It interacts with the greenhouse gases in the atmosphere and initialises a cascade of absorption-emission processes, the result being the observed greenhouse effect. Introducing relative greenhouse effect $B \equiv f/q$ (Raval and Ramanathan, 1989) we may write:

$$q = q_e + Bq \quad \text{or} \quad q = q_e/(1 - B) \quad (8.3.2)$$

The relative greenhouse effect B can be interpreted as the fraction of terrestrial radiation that is reflected by the atmosphere back to the planet's surface. Note that both absolute, f , and relative, B , values of the greenhouse effect depend on two temperatures, T and T_e .

The absolute greenhouse effect f can be represented as a sum of spectral components, similar to q (8.2.1) and q_e (8.2.2). Using (8.2.1), (8.2.2) and the definition of f (8.3.1) we have:

$$f(T, T_e) = \sum_{i \neq k} \Delta\omega_i [I(\omega_i, T) - I(\omega_i, T_e)] \quad (8.3.3)$$

Here spectral intervals $\Delta\omega_i$ are the same as in (8.2.2). Naturally, spectral windows $\Delta\omega_k$ (8.2.2) do not make any contribution to (8.3.3).

Modern greenhouse effect on Earth is by more than 90 percent determined by atmospheric water vapour and carbon dioxide (Mitchell, 1989). These two gases absorb thermal radiation over a wide range of frequencies spanning almost the whole spectre of terrestrial radiation. Thus, terms $\Delta\omega_{\text{H}_2\text{O}}$ and $\Delta\omega_{\text{CO}_2}$ are the major contributors to (8.3.3):

$$\begin{aligned} f(T, T_e) &\approx \Delta\omega_{\text{H}_2\text{O}} \Delta I(\bar{\omega}_{\text{H}_2\text{O}}) + \Delta\omega_{\text{CO}_2} \Delta I(\bar{\omega}_{\text{CO}_2}), \\ \Delta I(\omega) &\equiv I(\omega, T) - I(\omega, T_e) \end{aligned} \quad (8.3.4)$$

Using the Planck function $I(\omega, T)$ and corresponding spectral widths $\Delta\omega_{\text{H}_2\text{O}}$ and $\Delta\omega_{\text{CO}_2}$ known from spectroscopic measurements and taking into account a few minor correction coefficients we obtain the following estimates for the greenhouse contributions of water vapour and carbon dioxide (Mitchell, 1989):

$$\begin{aligned} f_{\Delta\omega_{\text{H}_2\text{O}}} &= \Delta\omega_{\text{H}_2\text{O}} \Delta I(\bar{\omega}_{\text{H}_2\text{O}}) \approx 100 \text{ W m}^{-2}, \\ f_{\Delta\omega_{\text{CO}_2}} &= \Delta\omega_{\text{CO}_2} \Delta I(\bar{\omega}_{\text{CO}_2}) \approx 50 \text{ W m}^{-2}. \end{aligned} \quad (8.3.5)$$

The cumulative contribution of the two gases is approximately equal to the total value of f (8.3.1), which justifies use of the Planck function in (8.3.5) instead of the empirically observed values of spectral intensity $I(\omega)$. According to (8.3.5) the contribution of carbon dioxide is two times less than that of water vapour.

Note that values of concentrations of the two gases did not enter the calculations of the result (8.3.5). The only fact used implicitly was that these gases are present in the atmosphere in sufficiently large (saturated) concentrations to ensure almost complete absorption of terrestrial radiation in the respective parts of the spectre, which makes it possible to use the Planck function. It was assumed that after absorption has been saturated, further increase of the gas concentration does not impose any major impact on the greenhouse effect.

Concentration dependence of the greenhouse effect is only observed for the minor greenhouse gases which are present in very small quantities and have absorption bands in spectral windows where there are no other absorbers. Greenhouse contributions of such gases (chlorofluorocarbons mostly) grow linearly with concentration until the latter is high enough to ensure almost complete absorption in the middle of the absorption bands. Further increase of concentration will result in a very slow (square root or logarithmic) broadening of the absorption interval $\Delta\omega$ and a correspondingly slow increase in the greenhouse effect due to saturation of the wings of the absorption lines, provided that there are no other saturated gases absorbing at the same frequencies (Mitchell, 1989).

Such a situation is realised for the atmospheric CO_2 . Thus, the widely discussed potential global warming is traditionally calculated on the basis of logarithmic growth of the CO_2 greenhouse contribution (8.3.5) with concentration (Ramanathan *et al.*, 1987; Mitchell, 1989; IPCC, 1994).

However, the results (8.3.5) obtained with use of (8.3.3) and (8.3.4) do not actually represent contributions of real *concentrations* of the greenhouse gases to the greenhouse effect but, rather, contributions of those spectral *intervals* that correspond to absorption bands of CO₂ and H₂O. The calculations were based on the known values of temperature T_e of the upper radiating layer of the atmosphere and temperature T of the Earth's surface. Values of T_e and T entered the calculations as empirical parameters. In reality, however, the difference $T_e - T$ is completely determined by and strongly dependent upon atmospheric concentrations of the greenhouse gases.

True greenhouse contribution of a gas can be measured as the flux of energy re-emitted by this gas back to the Earth's surface. True greenhouse contributions of gases are not additive. It means that a mixture of greenhouse gases does not produce a greenhouse effect equal to the sum of greenhouse effects caused by each gas alone. Energy absorbed by certain gas A in a given spectral interval is partly re-emitted back to the Earth's surface heating it. The Earth's surface emits then more radiation in all spectral intervals due to re-distribution of the absorbed energy over the whole black-body spectre. In the presence of gas A other greenhouse gases have therefore an opportunity of absorbing and, consequently, re-emitting more thermal radiation in their own spectral intervals.

As far as energy absorbed by one gas in a given spectral interval can be transmitted to another gas and thus re-distributed over other spectral intervals, greenhouse contributions from different spectral intervals calculated as in (8.3.5) do not give information about true greenhouse contributions of gases. The real dependence of the greenhouse effect on concentrations of the greenhouse gases, which is critical to evaluating stability of the Earth's climatic system, in such consideration appears to be hidden.

8.4 DEPENDENCE OF THE GREENHOUSE EFFECT ON CONCENTRATIONS OF THE GREENHOUSE GASES

Let us consider a one-dimensional atmosphere consisting of N greenhouse gases with non-overlapping absorption intervals $\Delta\omega_l (l = 1, 2, \dots, N)$ covering the whole range $\Delta\Omega$ of terrestrial radiation frequencies, $\sum_{l=1}^N \Delta\omega_l = \Delta\Omega$ (8.2.1). Let δ_l be the relative portion of energy of thermal radiation of the Earth's surface corresponding to the absorption interval of the l -th greenhouse gas. We obtain from (8.2.1):

$$\delta_l = \frac{\Delta\omega_l I(\omega_l, T)}{\sigma T^4}; \quad \sum_{l=1}^N \delta_l = 1. \quad (8.4.1)$$

Let n_l be the number of optically dense layers of the l -th gas present in the atmosphere, i.e. the number of such layers that with sufficient accuracy ensure complete absorption of the thermal radiation in the respective part of the spectre. The number of optically dense layers is proportional to the so-called optical

thickness. Thickness of a single optically dense layer can be approximated as the average length of free path of thermal photons between two successive collisions with molecules of the gas. The total number of optically dense layers is thus proportional to the gas concentration given the constant height of the atmosphere. Each optically dense layer absorbs thermal radiation in the corresponding spectral interval and reradiates in all possible directions, i.e. up and down in the one-dimensional atmosphere.

Let $x_{l,k}$ be the flux of thermal radiation emitted by the k -th layer of the l -th gas in either upward or downward direction (we consider these two fluxes equal). $x_{l,1}$ stands for radiation of the upper radiative layer emitted directly into space.

Energy balance equations for Earth as a whole, (8.4.2), each layer of each greenhouse gas, (8.4.3), and the Earth's surface, (8.4.4), can be written in the following form, see Figure 8.1:

$$q_e = \sum_{l=1}^N x_{l,1}, \quad (8.4.2)$$

$$2x_{l,1} = x_{l,2}, \quad 2x_{l,2} = x_{l,1} + x_{l,3}, \dots, \quad (8.4.3)$$

$$2x_{l,k} = x_{l,k-1} + x_{l,k+1}, \dots, \quad 2x_{l,n_l} = x_{l,n_l-1} + q\delta_l, \quad (8.4.4)$$

$$q = q_e + \sum_{l=1}^N x_{l,n_l}. \quad (8.4.4)$$

Here q and q_e are fluxes of radiation emitted from the Earth's surface and the upper radiating layers, respectively. Note that q_e is equal to the flux of solar radiation absorbed by the Earth's surface. Equation (8.4.4) can be deduced from all the other equations. Figure 8.1 gives an idea of the described situation for $N = 1$ and $\delta_l = 1$ (i.e. when only one gas is present).

The recurrent equations (8.4.3) reflect the fact that any layer absorbs radiation emitted from the two neighbouring layers only. Radiation that is emitted from more distant layers does not reach this particular layer being completely absorbed by the intermediate layers. Solving the system (8.4.3) we obtain that $x_{l,k} = kx_{l,1}$, ($k = 1, 2, \dots, n_l$), which can be easily tested. Using this expression in the last equation of (8.4.3) and solving the latter together with (8.4.2), we arrive at the following expression for q :

$$q = \frac{q_e}{\frac{\delta_1}{n_1+1} + \frac{\delta_2}{n_2+1} + \dots + \frac{\delta_N}{n_N+1}} \equiv \frac{q_e}{b}, \quad \text{where} \quad b = \sum_{l=1}^N \frac{\delta_l}{n_l+1} \quad (8.4.5)$$

$$\text{and} \quad \sum_{l=1}^N \delta_l = 1$$

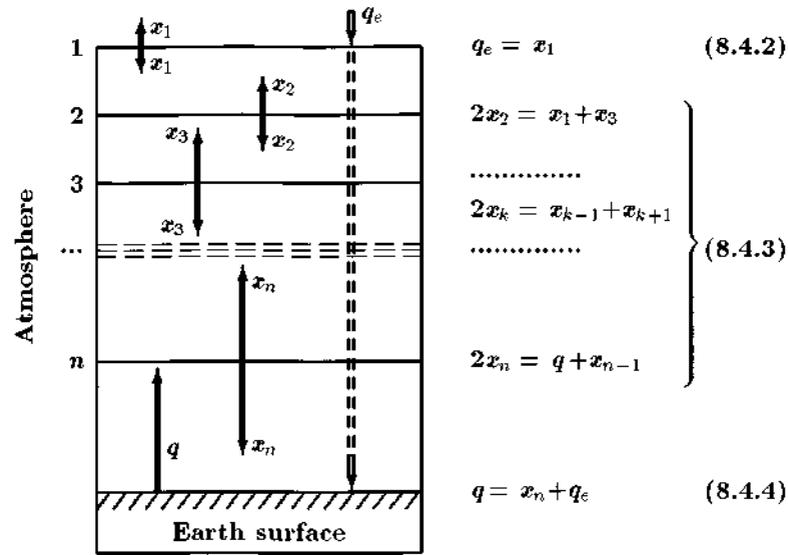


Figure 8.1. Dependence of the greenhouse effect on the number of optically dense layers n . x_k ($k = 1, 2, 3, \dots, n$) is the intensity of thermal radiation emitted by the k -th layer upwards and downwards; q is the intensity of thermal emission by the Earth's surface; q_e is the intensity of solar radiation absorbed by the Earth's surface, which is equal to the intensity of radiation of the upper radiative layer, x_1 . The atmosphere is considered to be completely transparent to solar radiation. It is assumed that the considered gas ensures complete absorption of thermal radiation over the whole terrestrial spectre, $\delta_l = 1$.

Comparing (8.4.5) and the definition of the relative greenhouse effect B (8.3.2) we obtain for the relative and absolute values of the greenhouse effect B and f :

$$B = 1 - b = 1 - \sum_{l=1}^N \frac{\delta_l}{n_l + 1}; \quad f = Bq \quad (8.4.6)$$

If the number of optically dense layers n_l is of the same order of magnitude for all gases, the flux of terrestrial radiation q and the absolute greenhouse effect f increase infinitely with growing n_l , while b tends monotonously to zero and B tends to unity. This is especially clear in the case of one greenhouse gas ($N = 1$) with n layers. Then $B = 1 - 1/(n + 1)$ and $f = q_e n$. Expressions (8.4.5) and (8.4.6) make it clear that contributions of different greenhouse gases into the greenhouse effect are not additive, as opposed to the results of (8.3.5).

Let us now consider a situation when one of the N gases is absent, $n_k = 0$. Then there is a spectral window in the frequency interval $\Delta\omega_k$. Terrestrial radiation in this part of the spectre passes unimpeded through the atmosphere directly into space and is equal to $\delta_k q$. In such a case an increase in the number of optically dense layers n_l and, consequently, in concentrations of the remaining greenhouse gases cannot lead to an infinite increase of the terrestrial radiation q . Increase in q is saturated when all

the $(N - 1)$ items with non-zero n_l in the denominator of fraction (8.4.5) become infinitely small as compared to the finite k -th item corresponding to the spectral window (at $n_k = 0$ it is simply equal to δ_k). The saturated values of q , B and f are then as follows:

$$q = q_e / \delta_k, \quad B = 1 - \delta_k, \quad f = q_e (1 - \delta_k) / \delta_k \quad (8.4.7)$$

The effect of saturation has a very clear explanation. Higher concentrations enable the greenhouse gases to re-emit more radiation, heating the Earth's surface and increasing the surface temperature and terrestrial radiation q . As a result, more and more radiation is released into space through the spectral window while less and less radiation is emitted from the upper radiating atmospheric layers ($x_{l,1} \rightarrow 0$), the total amount of the released radiation being limited by the absorbed solar energy q_e . Saturation corresponds to the case when practically all the radiation passes through the spectral window, $q_e = \delta_k q$.

The role of the spectral windows in the modern atmosphere can be estimated comparing the share of window radiation with the amount of absorbed solar energy q_e . Observations show that radiation escaping the Earth's surface through the spectral windows constitutes less than 10% of the total released radiation (Barry and Chorley, 1987), $\delta_k q \leq 0.1 q_e$. Thus, the situation is that in the modern atmosphere there is very far from saturation of the greenhouse effect owing to the spectral windows.

In the absence of clouds the limiting contribution to the greenhouse effect comes from the l -th gas that is characterised by the maximum value of $\delta_l / (n_l + 1)$ in (8.4.5). If the concentration of that gas remains constant, changes in concentrations of the other gases do not have any considerable effect on q and f . When the concentration of this gas and, hence, n_l , increases, the term $\delta_l / (n_l + 1)$ decreases, and values of q and f grow practically linearly with n_l until the diminishing value of $\delta_l / (n_l + 1)$ becomes equal to that of some other gas, $\delta_m / (n_m + 1)$. After that the concentration of the m -th gas becomes the limiting factor for q and f and so on.

Clouds absorb terrestrial radiation rather evenly over the whole thermal spectre contributing to all spectral intervals $\Delta\omega_l$. Thus, the number of optically dense layers of clouds n_0 should be added to the number of layers of every greenhouse gas giving the following expression for b :

$$b = \sum_{l=1}^N \frac{\delta_l}{n_l + n_0 + 1} \quad (8.4.8)$$

In a case when the major contribution to the greenhouse effect comes from clouds ($n_0 + 1 > n_l$), b can be expanded into a Taylor series in terms of small quantities $n_l / (n_0 + 1)$. Then we arrive at the following expression for the absolute greenhouse effect $f \equiv q_e B / b$:

$$f = q_e \left\{ n_0 + \sum_{l=1}^N \delta_l n_l \right\}, \quad n_l < n_0 + 1 \quad (8.4.9)$$

Note that contributions from different greenhouse gases are additive only in such limiting case.

For the modern atmosphere where the greenhouse gases (CO₂ and H₂O mostly) absorb thermal radiation practically over almost the whole terrestrial spectre, see (8.3.5), expression (8.4.9) can be written as follows:

$$f \equiv q_e B/b = q_e \{n_0 + \delta_{\text{H}_2\text{O}} n_{\text{H}_2\text{O}} + \delta_{\text{CO}_2} n_{\text{CO}_2}\} \quad (8.4.10)$$

The available estimates of the contribution of clouds to the thermal radiation of the modern atmosphere (Raval and Ramanathan, 1989; Kondratyev, 1999) coincides by the order of magnitude with contributions from CO₂ and water vapour. Thus, for the modern atmosphere the expression (8.4.10) is true to the accuracy of the order of magnitude only. However, with increasing cloudiness its contribution into greenhouse heating becomes dominant and the accuracy of (8.4.10) should increase. The absolute greenhouse effect f grows then proportionally to the mass of clouds, while contributions from concentrations of CO₂ and water vapour become less and less important.

In the Sections that follow we discuss the influence of the obtained temperature dependence of the greenhouse effect on the stability of the Earth's climate.

Note that the existence of the observed vertical temperature gradient in the atmosphere means that the energy of excitation of greenhouse gases' molecules owing to absorption of thermal radiation is fairly evenly distributed over all energetic degrees of freedom of air molecules, including degrees of freedom of chaotic (thermal) movement. This is achieved via rapid energy exchange during molecular collisions. Were there no such even distribution (e.g. if the lifetime of the excited states of greenhouse molecules were much shorter than the time interval between two successive molecular collisions), no vertical temperature gradient could form in the atmosphere. In such a case the air temperature in the upper troposphere would coincide with the temperature of the Earth's surface, although the greenhouse effect could remain absolutely the same. That is, the upper troposphere would be as warm as the Earth's surface is now, i.e. with a temperature much higher than the effective planetary temperature T_e (Table 8.1). (To stress the difference: in the absence of the greenhouse effect the Earth's surface would be as *cold* as the upper troposphere is now.)

This suggests that the greenhouse effect (i.e. heating of the Earth's surface above the effective planetary temperature) is not necessarily coupled to the presence of the vertical temperature gradient. Hence, the dependence of the greenhouse effect on the concentrations of the greenhouse gases cannot be calculated on the basis of the observed temperature gradient alone.

With account made for distribution of the excitation energy of greenhouse molecules over all molecular energetic degrees of freedom, the observed constant vertical temperature gradient, and, consequently, the linear dependence of the air temperature on height, follows easily from the obtained solution of Eqs. (8.4.2-3). $x_{i,k} = kx_{i,1}$, if one recalls the small ratio of the difference between the surface and upper tropospheric temperature, $T - T_e$, to the absolute effective surface temperature T .

8.5 POSSIBLE CLIMATES ON EARTH AND THEIR STABILITY

Energy balance for a unit of the Earth's surface area consists in the fact the rate of energy content change per unit area is equal to the difference between the average flux of solar radiation absorbed by the Earth and the average flux of thermal radiation emitted by the Earth to space.

Energy content is equal to cT , where c is the average heat capacity per unit area of the Earth's surface and T is its absolute temperature. Due to rotation of the Earth, the solar flux I incident upon the Earth's cross section area πr_E^2 , where r_E is the Earth's radius, is distributed over the whole planet's surface area, $4\pi r_E^2$. As a result, the average flux of solar radiation per unit of the Earth's surface area is equal to $I/4$. The absorbed solar radiation is equal to $aI/4$, where $a \equiv 1 - A$, A is the planetary albedo, that is, the fraction of solar radiation reflected by the planet back to space. The net flux of heat from the Earth to space is equal to the flux of heat from the Earth's surface σT^4 (Stephan-Boltzmann law) multiplied by the coefficient $b \equiv 1 - B$, where B is the relative greenhouse effect and describes the part of terrestrial radiation effectively re-emitted by the atmosphere back to the Earth's surface.

The energy balance equation for a unit area of the Earth's surface can be written as follows:

$$c \frac{dT}{dt} = \frac{I}{4} a - \sigma T^4 b \equiv -\frac{dU}{dT} \quad (8.5.1)$$

We introduced in (8.5.1) the potential Lyapunov function U characterising stability of the energy balance equation. The only variable in (8.5.1) is the temperature T . Coefficients a and b are also temperature-dependent. The potential function U is also temperature-dependent. It is chosen so that the negative value of its first temperature derivative is equal to the rate of energy content increment. Equation (8.5.1) is based on the law of energy conservation. Its accuracy is determined by the accuracy of characterisation of the whole Earth's surface by a global mean temperature. Minor deviations of the thermal radiation of Earth from blackbody radiation in some parts of thermal spectre are taken into account in coefficient b . Processes of convection, water evaporation and condensation are ordered processes that are generated by highly-ordered solar energy (see Chapter 7). These processes, as well as absorption of solar radiation by the lower atmosphere, represent intermediate stages of dissipation of the solar energy into the thermal radiation of the Earth's surface.

In a stationary state, when the energy content does not change, $c \frac{dT}{dt} = 0$, the derivative of U turns to zero, and, consequently, U has an extreme—maximum or minimum. The right-hand part of the equality in (8.5.1) also turns to zero, and this equality determines a stationary temperature $T = T_S$:

$$\frac{I}{4} a - \sigma T_S^4 b = 0, \quad \text{or} \quad T_S = T_o \left(\frac{a}{b} \right)^{1/4},$$

$$\text{where} \quad T_o \equiv \left(\frac{I}{4\sigma} \right)^{1/4} = 278 \text{ K} \quad (8.5.2)$$

(See notes to Table 8.1 for numerical values of σ and I .) If $a = b = 1$, which means that both the planetary albedo A and greenhouse effect B are equal to zero, the stationary temperature of the Earth's surface would be equal to $T_0 = 278 \text{ K}$ (5°C). This temperature, which is totally determined by I , that is, by the planet's location in the solar system, can be called the planet's orbital temperature (Table 8.1).

Note that here and below we imply that the notion of stationary state describes a state where the average energy content cT does not change with time. Here all oscillatory processes after averaging over time periods longer than periods of oscillation are included, as well as chaotic fluctuations that do not change the average energy content.

When the first derivative of U is equal to zero, the sign of its second derivative $U'' \equiv \frac{d^2U}{dT^2}$ determines the character of the extreme. It is a minimum, when U'' is positive, and maximum, when U'' is negative. The sign of U'' allows one to judge about stability of stationary solutions of (8.5.1). This can be shown as follows.

In the neighbourhood of the stationary point $T = T_S$ the first derivative $U' \equiv \frac{dU}{dT}$ can be expressed as a Taylor power series in terms of a small deviation $x \equiv T - T_S$:

$$U'(T) = U'|_{T=T_S} + U''|_{T=T_S}(T - T_S) = U''|_{T=T_S}(T - T_S) = U''|_{T=T_S}x$$

Equation (8.5.1) can be then written as follows:

$$\frac{dx}{dt} = -kx, \quad \text{where} \quad k \equiv \frac{1}{c} U''|_{T=T_S}$$

Heat capacity c is positive, so that the sign of coefficient k coincides with that of U'' . Solution of the above equation looks like $x = x_0 e^{-kt}$, where x_0 is an arbitrary constant, t stands for time.¹ Thus, when $k > 0$ (U has a minimum), any initial deviation of temperature T from the stationary value T_S exponentially damps out, which means that the stationary state is stable. In such case k can be interpreted as coefficient of relaxation, while the reciprocal value k^{-1} characterises the time of recovery of the stationary state after a perturbation. On the contrary, when $k < 0$ and U has a maximum at $T = T_S$, any deviation exponentially grows with time. In such case the stationary state $T = T_S$ is unstable.

We have seen therefore that stability of stationary states of (8.5.1) can be readily illustrated by the character of function U in very much the same manner as the gravitational potential of Earth can be described by relief of the surface. Stable stationary states correspond to minima (pits) of function U . The degree of stability depends on a pit's depth. Unstable states correspond to maxima (hills) of function U . Any deviation from the stationary state leads to sliding down the hill to one of the two nearest pits located to the left and to the right of the hill.

¹ The value $\lambda \equiv U''|_{T=T_S} = kc$ has the dimension of the flux of entropy, $\text{W m}^{-2} \text{K}^{-1}$, and is usually called climate sensitivity (Kondratyev, 1999).

Values and signs of U'' and relaxation coefficient k are unambiguously determined by temperature dependence of functions a and b in (8.5.1) and, consequently, temperature dependencies of albedo A and greenhouse effect B . Expanding functions a and b in Taylor power series with respect to small deviations $x \equiv T - T_S$ and using relation (8.5.2) we obtain the following expression for U'' in the stationary point T_S using temperature derivatives of a and b , a' and b' :

$$U''|_{T=T_S} = \frac{I}{4} \frac{a}{T_S} (4 + \beta - \alpha), \quad \text{where} \quad \alpha \equiv \left(\frac{a'T}{a} \right)_{T=T_S},$$

$$\beta \equiv \left(\frac{b'T}{b} \right)_{T=T_S}. \quad (8.5.3)$$

From (3) it is evident that a stationary state is stable when $\alpha - \beta < 4$ and unstable when $\alpha - \beta > 4$.

It is convenient to seek solution of (8.5.2) in a graphical form drawing the curve $y_1(T) = T_0 \left(\frac{a(T)}{b(T)} \right)^{1/4}$ and looking for intersections of this curve with the line $y_2(T) = T$.

8.6 PHYSICAL STABILITY OF THE EARTH'S CLIMATE

Let us now consider the particular physical behaviour of the greenhouse effect and albedo and functions $a(T)$ and $b(T)$ when the temperature T changes from the state of total glaciation 1 to the state of complete evaporation of the hydrosphere 3, Table 8.1.

The stationary steady state of an ice-covered Earth lies in the interval of temperatures lower than -80°C . In this state all the major components of the environment, including atmospheric CO_2 , are present in the solid phase. As soon as the solid phases of most components persist over a broad interval of low temperatures, one can assume that in the vicinity of the stationary state 1 neither the greenhouse effect nor albedo depend on temperature. Albedo of the ice-covered Earth should be equal to that of snow cover, i.e. to 80% (North *et al.*, 1981; Mitchell, 1989) making the value of $a(T)$ equal to 0.2. The relative greenhouse effect on Mars where the mean surface temperature is higher than -80°C and gaseous carbon dioxide is relatively abundant, is about 7% (Pollack, 1979; Kasting *et al.*, 1988; Mitchell, 1989). Thus, for the ice-covered Earth the relative greenhouse effect $B(T)$ does not presumably exceed 1–5%, and the value of $b(T)$ is not more than 0.95. We use below $b(T) = 0.95$ for the state of total glaciation.

The stationary steady state of complete evaporation of the hydrosphere corresponds to global mean surface temperature higher than 400°C (Table 8.1) and atmospheric pressure exceeding the present one by a factor of several hundred. As soon as all the hydrosphere is evaporated, the atmospheric concentration of water does not further change with temperature. Thus, it is reasonable to assume that in

state 3 neither albedo nor greenhouse effect change considerably with temperature, similarly to the situation in the stationary state 1. On Venus the surface temperature is about 460°C, while the planetary albedo (which is completely due to the cloudiness) is equal to that of the ice-covered Earth (Table 8.1). Thus we may take the value of albedo in state 3 equal to 80% and value of $a(T) = 0.2$, as in state 1. The relative greenhouse effect on Venus can be calculated from the difference between the temperature T on the planet's surface and the effective temperature T_e of thermal radiation of the planet measured from the outer space. The effective thermal radiation of Venus is equal to $q_e = \sigma T_e^4 = Ia/4 = 163 \text{ W m}^{-2}$, $T = 232 \text{ K}$. Thermal radiation of the planet's surface is calculated as $q = \sigma T^4 = (Ia/b)/4 = 16\,000 \text{ W m}^{-2}$, $T = 730 \text{ K}$ (Table 8.1). From these values and (8.3.2) we obtain

$$b \equiv 1 - B = (T_e/T_s)^4 = 163 \text{ W m}^{-2}/16\,000 \text{ W m}^{-2} = 0.010$$

It is reasonable to accept the obtained value of $b(T) = 0.010$ for the stationary state of the totally evaporated hydrosphere on Earth. The atmosphere of Venus consists of carbon dioxide to the extent of 96% and has a pressure of about 93 bars² (Kasting *et al.*, 1988). On Earth the cumulative mass of the oceans exceeds mass of the modern atmosphere by 300 times (Allen, 1955). In the state of total evaporation of the hydrosphere the atmospheric pressure would be about 300 bars. In such a case, the terrestrial H₂O will find itself, similarly to CO₂ on Venus, above the critical point where the differences between gases and liquids vanishes (Landau *et al.*, 1965). Thus it is natural to assume, with allowance made for the cloudiness, see Section 8.4, that in the state of total evaporation of the hydrosphere the relative greenhouse effect B on Earth would be at least not less than it is on Venus, making the value of $b \equiv 1 - B$ not greater than 0.010.

The chosen values of $a(T)$ and $b(T)$ in states 1 and 3 completely determine the mean global temperature of the Earth's surface in these two states, see (8.5.2) and Table 8.1.

Let us now consider possible physical mechanisms of transition from the stable state of the ice-covered Earth 1 to the stable state of total evaporation of the hydrosphere 3.

The major part of the modern greenhouse effect on Earth is due to the atmospheric water vapour. Atmospheric concentration of water vapour varies greatly in space and time. However, the average concentration of the water vapour changes proportionally to its saturated concentration (Ramanathan *et al.*, 1987; Raval and Ramanathan, 1989). Saturated concentration of water vapour, as well as the saturated partial pressure, $p_{\text{H}_2\text{O}}$, grows exponentially with increasing temperature in accordance with the Clausius–Clapeyron equation (Landau *et al.*, 1965):

$$p_{\text{H}_2\text{O}}(T) = Ce^{-\frac{Q_{\text{H}_2\text{O}}}{T}} \equiv e^{-\frac{Q_{\text{H}_2\text{O}}}{T} + \epsilon}, \quad T_{\text{H}_2\text{O}} \equiv \frac{Q_{\text{H}_2\text{O}}}{R} = 4871 \text{ K} \quad (8.6.2)$$

Here $Q_{\text{H}_2\text{O}}$ is the latent heat of evaporation of 1 mole of water vapour, R is the gas

²The global mean atmospheric pressure on Earth is equal to 1.013 bar, 1 bar = 10 N m⁻² (Allen, 1955).

constant, $T_{\text{H}_2\text{O}}$ is the effective temperature characterising the energetic of evaporation process, $C \equiv e^\epsilon$ is a temperature-independent constant.

As is shown in Section 8.4, the greenhouse effect is predominantly determined by the greenhouse gas that leads to formation of clouds and creates an absorption interval comparable with the total width of the thermal spectre. The absolute greenhouse effect grows linearly with concentration of clouds, (8.4.9). On Earth the most important greenhouse component responsible for cloud formation is the water vapour. Atmospheric concentration of water vapour changes with temperature according to (8.6.1).

In a stationary state we may write for the thermal radiation of the Earth's surface $q(T)$:

$$q(T) = \frac{I}{4}a + Bq(T) \quad (8.6.2)$$

Here I is the solar constant, $(I/4)a$ defines the amount of the absorbed solar energy ($a \equiv 1 - A$, where A is the planetary albedo), and is equal to the amount of thermal energy released by the planet into space, $(I/4)a = q_e$. Term $Bq(T) \equiv (1 - b)q(T) = f$ describes the additional radiation of the Earth's surface due to the greenhouse effect.

Using (8.6.1) and the results of Section 8.4, according to which the absolute greenhouse effect should grow proportionally to the concentration of atmospheric water, we may write

$$f \equiv Bq(T) = \frac{I}{4}ae^{-\frac{T_{\text{H}_2\text{O}}}{T} + \epsilon} \quad (8.6.3)$$

Using (8.6.2) and (8.6.3) we obtain the following expression for $b(T) \equiv 1 - B$:

$$b(T) = \frac{1}{1 + \exp\left(-\frac{T_{\text{H}_2\text{O}}}{T} + \epsilon\right)} \quad (8.6.4)$$

Expression (8.6.4) does not take into account the fact that the Earth's hydrosphere has a finite mass. At sufficiently large values of ϵ the value of b (8.6.4) diminishes almost infinitely. In reality, however, the decrease of b is stopped when all the hydrosphere is evaporated and there is no further increase in water vapour concentration and change of function $b(T)$ with increasing temperature. Thus, it is reasonable to specify the limiting value of b as $b_{\text{min}} = 0.010$, making it equal to the value of b observed on Venus, as discussed above. This can be done by adding the term 0.010 to the whole fraction (8.6.4). Parameter ϵ can be determined from the condition that at the modern mean global temperature $T = 288 \text{ K}$ the value of $b(T)$ given by (8.6.4) is equal to the observed value, $b(288) = 0.60$. The obtained value of ϵ is equal to 16.5. Noting that $T_{\text{H}_2\text{O}} = 4871 \text{ K}$ we finally obtain for $b(T)$:

$$b(T) = \frac{1}{1 + \exp\left(-\frac{4871}{T} + 16.5\right)} + 0.01 \quad (8.6.5)$$

The character of dependence of the planetary albedo on temperature remains to a large extent unknown. Its basic features, however, can be taken into account if the

unknown function $a(T)$ is approximated by a Gaussian curve. The limiting value of $a_{\min} = 0.2$ at high and low temperatures corresponds to the states of the total glaciation and total evaporation of the hydrosphere, where the planetary albedo $A \equiv 1 - a$ is at its maximum due to the high reflectivity of ice cover and cloudss⁴, respectively. The observed modern value of a is 0.7 ($A = 0.3$). We assume that the maximum value $a_{\max} = 0.8$. It corresponds to the state of already melted ice-cover but still predominantly liquid hydrosphere, which is chosen at a global mean temperature $T_m = 298$ K (25°C). We thus obtain for $a(T)$:

$$a(T) = 0.20 + 0.54 \exp \left[-\frac{(T - T_m)^2}{(\Delta T)^2} \right]; \quad T_m = 298 \text{ K (25°C);}$$

$$\Delta T = 40 \text{ K (40°C)} \quad (8.6.6)$$

Proportionality coefficient corresponding to the exponential term and the effective width of the curve ΔT are retrieved from the condition that at the modern mean global temperature $T = 288$ K the value of $a(T)$ given by the formula (8.6.6) is equal to the observed value, $a(288) = 0.70$, and the condition that Earth is totally covered with ice at -15°C . Variations of all the chosen parameters within physically reasonable limits do not change the results to be obtained.

Note that following temperature changes accompanying transition from the state of complete glaciation of the Earth's surface to the state of complete evaporation of the hydrosphere, the absolute greenhouse effect changes a hundredfold as compared to the two- to threefold change in the planetary albedo at maximum. Hence, the temperature dependence of the albedo has practically no impact on the solutions of (8.5.2) and their stability.

Figure 8.2-Ia shows the graphical solution of equation (8.5.2) with $a(T)$ and $b(T)$ specified by (8.6.6) and (8.6.5). Potential function U obtained by integrating (8.5.1) gives information on stability of the obtained three stationary states (Figure 8.2-Ib, c) (the integration constant was chosen such that the potential function U turns to zero in the stationary state of total evaporation of the hydrosphere). Stationary states of total glaciation and total evaporation of the hydrosphere are stable, as could be expected, while the stationary state corresponding to the modern climate proves to be unstable.

The obtained results suggest that the physical mechanisms alone cannot account for the observed substantial stability of the Earth's climate with regard to the global mean temperature during the four billion years of life existence. In the following section we address the problem of the biotic nature of climate stability on Earth.

8.7 BIOTIC STABILITY OF THE MODERN CLIMATE OF EARTH

According to the available paleodata global mean temperature of the Earth's surface never went beyond the interval from 5°C to 25°C (Savin, 1977; Watts, 1982;

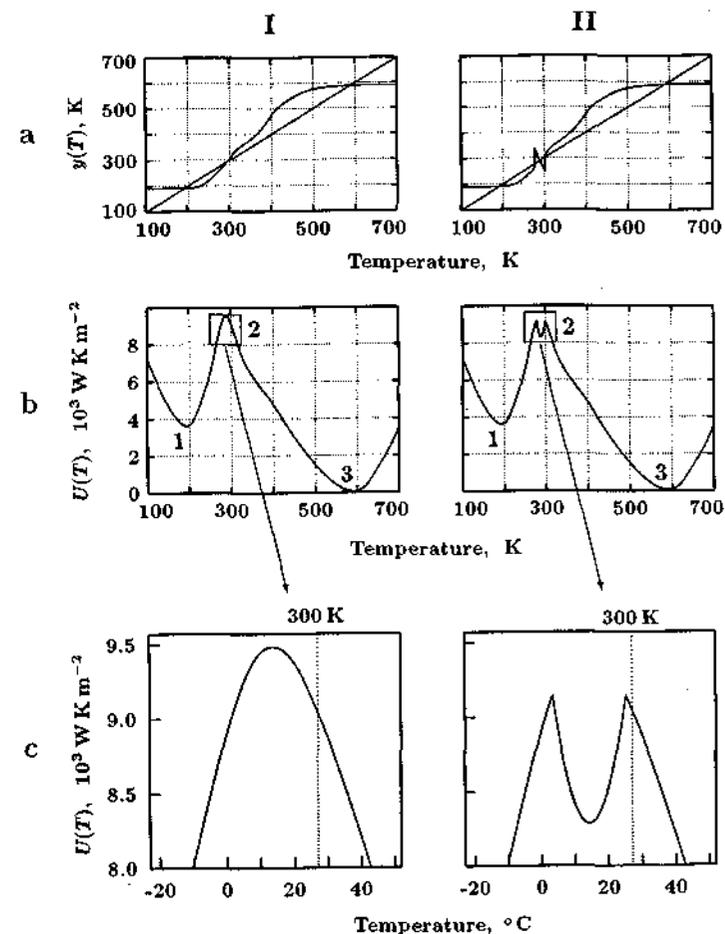


Figure 8.2. Physical (I) and biotic (II) stability of the global mean surface temperature on Earth. a – graphical solution for determination of stationary states 1, 2, 3 (8.5.2). Intercepts of the two lines are the stationary states corresponding to the extreme of potential function U (8.5.1). b, c – Potential (Lyapunov) function describing stability of the Earth's climate. The minima correspond to stable states. The maxima correspond to unstable states.

Berggren and Van Couvering, 1984; see also Figure 8.3), fluctuating around a stable mean value of about 15°C . It means that there exists a sufficiently strong negative feedback between changes in temperature and albedo and greenhouse effect, e.g. when an increase in temperature causes a reduced greenhouse effect that lowers the temperature back to the initial value.

The maximum possible stability of the stationary state 2 can be retrieved from the condition of the maximum physically possible negative feedback in the temperature interval $5^\circ\text{C} < T < 25^\circ\text{C}$. The maximum physically possible feedback can be illustrated by a hypothetical situation when the relative greenhouse effect B increases with decreasing temperature up to the maximum possible value, which is

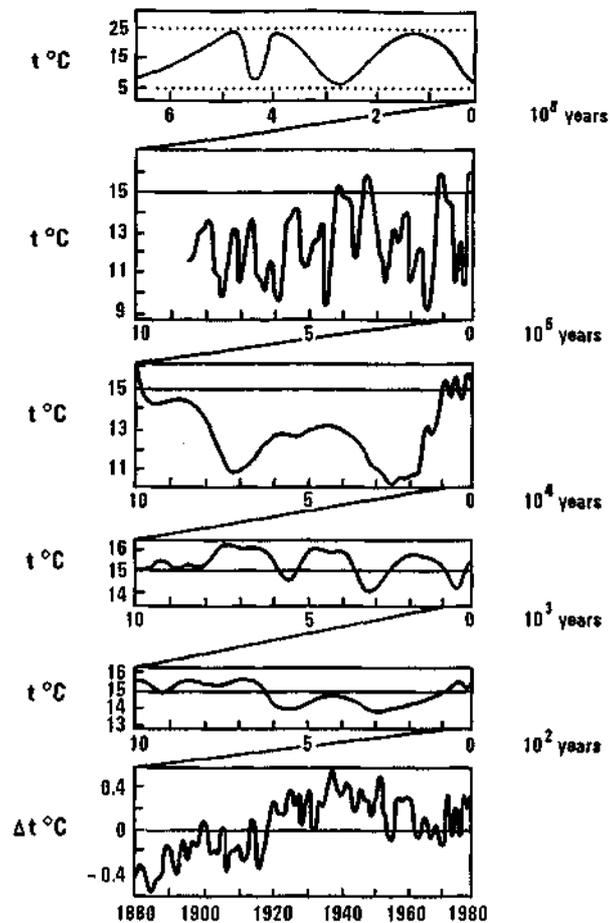


Figure 8.3. Time variations of the mean global surface temperature on Earth. (After Savin, 1977; Watts, 1982; Berggren and Van Couvering, 1984.)

realised at the lower boundary of the considered temperature interval, while the albedo A decreases with decreasing temperature and attains there the minimum possible value, $B(5^\circ\text{C}) = 1$, $A(5^\circ\text{C}) = 0$. The same situation is realised when the limiting values of albedo and greenhouse effect correspond to the upper boundary of the considered temperature interval, $B(25^\circ\text{C}) = 0$, $A(25^\circ\text{C}) = 1$.

Figures 8.2-IIa, b, c show the graphical solution of equation (8.5.2) and the potential function U (8.5.1) for a model example of weaker negative feedback:

$$A(T) = 0.3 + (t - t_S) 0.02; \quad B(T) = 0.4 - (t - t_S) 0.02;$$

$$5^\circ\text{C} < t < 25^\circ\text{C}; \quad t_S = 15^\circ\text{C}$$

In this example the boundary values of greenhouse effect and albedo are not the

maximum (minimum) possible ones: $A(5^\circ\text{C}) = 0.1$, $B(5^\circ\text{C}) = 0.6$, $A(25^\circ\text{C}) = 0.5$, $B(25^\circ\text{C}) = 0.2$. We assumed here that the global mean surface temperature in the stable stationary state chosen by life coincides with temperature of unstable physical stationary state. In this case the potential biotic pit (Figure 8.2-IIb, c) is symmetrical.

The real behaviour of albedo and greenhouse effect in the temperature interval $5^\circ\text{C} < T < 25^\circ\text{C}$ can be presumably deduced only empirically on the basis of detailed studies of paleodata, which is not our goal here. The natural biota of Earth uses highly ordered solar energy to ensure negative feedback of climatic perturbations in accordance with the genetic information of biological species. This is performed on the basis of non-equilibrium processes that cannot be accounted for in models based on physicochemical properties of the atmosphere alone. If the biosphere is introduced into the model as a physicochemical system, this critical property of the biota appears to be lost irrespective of whatever large number of empirical parameters that we are able to process in the model calculations. In other words, if we compare the biota to a programmed machine which is able to use the external flux of ordered energy in different regimes, we cannot predict this machine's functioning under different conditions without knowing the machine's program and judging by the characteristics of its current regime. As shown in Chapter 7, functioning of the biota is characterised by a degree of complexity that cannot in principle be modelled, while the biotic climatic impact (manifested in biotic power of control of the water cycle, carbon cycle, etc.) is huge. This calls for caution when interpreting predictions of mathematical climate models.

We have shown that the observed stability of the Earth's climate with regard to the mean global surface temperature points to the complex nature of the temperature dependencies of albedo and greenhouse effect within the life-compatible temperature interval. We have shown that, on the basis of the known physical atmospheric properties alone, there are no grounds to expect such anomalies (Figure 8.2-IIa) in behaviour of either albedo or greenhouse effect. This, in its turn, points to existence of a biotic mechanism of climate control within the interval of life-compatible temperatures. Such control can be based only on highly-ordered processes that are generated by solar radiation due to large differences between temperatures of solar radiation (short-wave radiation of the Sun) and terrestrial thermal radiation (long-wave radiation of the Earth).

The main quantitative characteristic of orderliness of processes generated in the course of any type of energy transformation is the difference between temperatures of the initial and final states, see Section 7.2. The final state of all processes taking place in the environment corresponds to chaotic thermal energy of the Earth's surface, which is characterised by global mean surface temperature $T_E \sim 300\text{K}$. Solar energy is characterised by temperature $T_S \sim 6000\text{K}$. The relative difference between these temperatures constitutes

$$\eta_S = \frac{T_S - T_E}{T_S} \sim 0.95 \quad (8.7.1)$$

In thermodynamics the variable η_S has the meaning of the efficiency of energy

transformation (Landau *et al.*, 1965). This is the maximum possible work efficiency of a reversible heat machine based on a succession of equilibrium thermodynamic processes, see also Section 7.2. The same variable characterises the efficiency of highly-ordered non-equilibrium processes generated by solar energy in the Earth's environment.

All processes of global circulation in the ocean and atmosphere are due to the temperature difference between the equator and the poles, which is of the order of $\Delta T_E \sim 30^\circ\text{C}$. The efficiency of ordered processes generated by global circulation are characterised by

$$\eta_E = \frac{\Delta T_E}{T_E} \sim 0.1, \quad (8.7.2)$$

which is an order of magnitude lower than the value of η_S (8.7.1). It means that a unit energy of global circulation generates in the Earth's environment 10 times less energy of ordered processes than the same unit of solar energy. The amounts of solar energy consumed by the global biota and by the global circulation processes are of the same order of magnitude (see Section 7.3, Table 7.1).

Hence, stability of the modern climate can be explained taking into account highly-ordered biotic processes generated by solar energy. Ordered processes that are due to global circulation make an order of magnitude lower contribution and may be neglected. There are no grounds to expect that detailed elaboration of global circulation models yields reliable predictions of climate change.

Until now, studies devoted to the climate stability problem have been based on quantitative characteristics of energy fluxes, while the degree of orderliness of corresponding processes has never been taken into account (North *et al.*, 1981; Ramanathan *et al.*, 1987; Tang and Weaver, 1995; Li *et al.*, 1997). This is predominantly due to the fact that most processes generated at the Earth's surface by solar energy go through stages that are very far from thermodynamic equilibrium and cannot be formally described within traditional thermodynamics. However, the parameter η , (8.7.1) and (8.7.2), is valid for all processes, including non-equilibrium ones, and can be used when estimating their orderliness. Note also that all models concerned about climate stability are based on empirical characteristics of the modern climate that enter the models as parameters. It is not surprising that in all such models the modern climate proves to be stable. The question *why* the observed characteristics of climate ensure its stability is rarely addressed.

Highly-ordered processes generated by solar energy with efficiency close to (8.7.1) are of biological nature and take place only in the biota. That means that the climate stability on Earth can be due only to the regulatory function of the natural biota. Ordered processes aimed at maintenance of climate stability should be necessarily characterised by strictly defined succession, direction and duration, i.e. fluxes and stores of information. Such information is contained in the genetic programme of biological species combined into ecological communities of the global biota. Natural ecological communities are characterised by strictly specified population densities of all species that undergo directional changes in response to environmental perturbations. Anthropogenic perturbation of natural ecological communities is manifested

in artificial change of the natural population density distribution of species and violation of genetic programmes of natural species due to creating new sorts of plants and breeds of animals. This leads to degradation of the stabilising environmental potential of the biota.

During large-scale natural perturbations of the environment of the past (e.g. glaciations, volcanic eruptions, meteorite fall, etc.) the regulatory potential of the biota decreased in proportion to reduction of areas occupied by the natural biota, but remained substantial. For example, were the area occupied by the global biota to reduce by half, the global regulatory potential would halve as well. That means that the rate of biotic compensation of external perturbations would be reduced twofold. Estimates of relaxation coefficients of the nonperturbed biota presented in Section 6.4 show that the characteristic time of complete recovery of the initial stable state of the environment after natural external perturbations is of the order of 10 years for the nonperturbed biota. Were the global biota reduced by half, that time would only double, which may not have any drastic consequences to the environmental stability.

In contrast, anthropogenic cultivation of territories occupied by natural biota has a catastrophic impact on environmental stability. The perturbed biota loses information necessary for correct regulation of the environment, but retains the absolute huge value of environmental impact. The destabilising impact of the perturbed biota may be larger than stabilising impact of the remaining intact territories, see Figure 6.6a. Hence, we again come to the conclusion that in order to ensure long-term stability of climate on Earth natural biota should be restored and protected on the majority of the continental surface. As one of the immediate measures, one can name considerable reduction and ultimate abandonment of forest use, see Section 6.5.

9

Genetic Bases of Biotic Regulation and Life Stability: Theoretical Consideration

Biotic regulation of the environment is performed on the basis of complex processes pertaining to huge fluxes and stores of information (Chapter 7). The information needed to ensure a negative feedback response to environmental perturbations may only be stored in DNA molecules of species genomes. Mechanisms preventing the genetic information of species from spontaneous decay are considered in this chapter.

9.1 ORGANISATION OF GENETIC INFORMATION OF SPECIES

Genetic information of most species of the biosphere is written in polymer double-strand molecules of DNA, which consist of the two mutually-complementary spirals. There are four different monomer units composing the double-helix DNA molecule. They are known as nucleotide pairs or base pairs (bp). In other words, the genetic text of the species-specific properties is written with the help of the four-letter alphabet. The complete genetic text of all the properties of the individual is called its genome. Genomes of higher organisms are very large and consist of separate chromosomes (separate books, to continue the analogy).

Replication of DNA is performed by unwinding the double helix and building a complementary daughter strand of nucleotide pairs at each of the single parent strand, so that two copies of the parent DNA appear. Random errors that arise in the course of such copying and other cell activities are known as mutations. They include random change, insertion or deletion of nucleotide pairs (these mutations are referred to as *point mutations*) and deletion, insertion or transposition of whole fragments of DNA (these mutations are known as chromosome mutations or *macromutations*) (Ayala and Kiger, 1984; Lewin, 1987).

The probability of random mutation depends on the quantum characteristics of the polymer molecule and does not correlate with information written into that molecule (the number of misprints in the book does not depend on the book's content). That probability is reduced by approximately a millionfold by a

DNA-synthesis catalytic system such as proofreading and mismatch repair (Loeb and Kunkel, 1982; Goodman and Fygenon, 1998), which is itself encoded into the DNA macromolecule. The proofreading and other repairing catalysts operate in one and the same way upon any section of the DNA molecule, independent of information written into them. One may reasonably assume that the probability of random mutation, ν , per nucleotide site per division is a universal value typical for the whole genome.¹

Synthesis of proteins begins with transcription of the genetic information of the DNA to single-chain polymer molecules of the RNA, which is followed by translation of the RNA information to proteins, which are polymer molecules built of 20 aminoacids (protein letters). Genomes of certain viruses are built directly of RNA molecules. Replication of RNA molecules is not controlled by the proofreading and mismatch repair systems. Thus the probability of random mutations in RNA viruses is on average a million times higher than that in organisms having their genetic information written into the DNA molecules (Holland *et al.*, 1982; Domingo *et al.*, 1996).

Rates of all processes, including those of mutagenesis and cell division, are proportional to the metabolic rate of the organism. If we assume that the system of catalytic repair is also universal for any DNA genome (Radman and Wagner, 1988), one may expect that the probability of random mutation, ν , per base pair (bp) per division (d) is also a universal value for most DNA-based organisms. Such an assumption agrees satisfactorily with the experimental data available (Drake, 1969, 1974; Radman and Wagner, 1988; Orr, 1995), according to which ν is of the order of

$$\nu \sim 10^{-10} (\text{bp})^{-1} \text{d}^{-1} \quad (9.1.1)$$

Deviations from (9.1.1) in either direction, observed in various species, do not usually exceed an order of magnitude (Drake, 1974).

Genes consist on average of about 10^3 nucleotide pairs. Thus the probability ν_g of a random mutation per gene per division is of the order of

$$\nu_g \sim 10^{-7} (\text{gene})^{-1} \text{d}^{-1} \quad (9.1.2)$$

Probabilities of mutual mutational substitutions between the four nucleotide pairs differ slightly from pair to pair (Kimura, 1983; Jukes, 1987). The cumulative probability of various types of macromutations pertaining to a given site does not exceed the average probability of point mutation, (9.1.1), by more than a factor of five (Ayala and Kiger, 1984). Thus, recalling that the value of (9.1.1) is found to the accuracy of an order of magnitude only, we may further use that value as a representative estimate of the probability of a single random mutation of any type in a given site.

¹ The probability of random mutation in certain small parts of the genome may increase in cases where special 'mutator' genes are present, that is, coding enzymes that identify these particular parts of the genome and increase the probability of mutations in them, for example blocking the system of proofreading (Ayala and Kiger, 1984). Decay of mutator genes results in the failure of synthesis of the respective enzymes and in the evening out of the rate of mutations in the genome as a whole.