Some Perspectives of the Major Biogeochemical Cycles
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CHAPTER 1

Factors Controlling Global Climate of the Past and the Future

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ABSTRACT

Viewpoints expressed in the literature on causes of climatic change are examined and critically assessed. It appears that global climate is a result of a series of factors that operate on different time scales. For this reason we distinguish between climate curves which show changes that have periodicities of millions, thousands, or hundreds of years. All three curves are superimposed and global climate at any one time in the geological past is an expression of this superposition. It is concluded that plate tectonics is the principal long-term regulating factor of global climate by controlling land-sea ratio and albedo. Orbital periodicities of the earth which fluctuate at medium-term time scales (10^4–10^6 years) control the solar radiation curve. The main contributing factor for climatic variation over the past few hundred to few thousand years is short-term periodic changes in the luminosity of the sun and in the volcanic dust concentrations of the atmosphere. In contrast, CO_2 in the atmosphere has not been a climate-controlling device in the past, but will become the principal anthropogenic agent over the next hundred years.

1.1 INTRODUCTION

Our planet Earth has undergone, from the time of its formation, a series of warm periods and ice ages. Since life appeared on land 500 million years ago, however, the global average temperature has fluctuated only by a few degrees Centigrade, although glaciations have occurred during the Permian [270–230 m.y. B.P. (million years before present)] and the Quaternary (2–0 m.y. B.P.). Closer examination of the geological record reveals that the earth was always warm enough to support life, and excursions into ice ages were only of short duration (10^6 yr). Yet, their impact on terrestrial fauna and flora was substantial due to a shift in the positions of climatic zones during the cold interludes.

At present we live in a moderately warm interglacial stadial of an ice age or, expressed differently, in an overall climatic situation which in a geological scale, has little chance of persisting in time. Actually, over the past few hundred thousand years, the climatic pendulum has swung several times almost over the full range from the extreme warm to the extreme cold. These events are registered in sediments in the form of distinct markers and our main objective in this paper will be to use such indicators to find out more about the reasons behind climatic changes throughout the history of the earth.

For the purpose of this discussion we shall distinguish climatic factors which operate on long-term (10^6–10^8 yr), medium-term (10^4–10^6 yr) and short-term (10–10^4 yr) basis. These time scales represent the length of galactic cycles, the period of planetary orbital cycles, and the time span of solar cycles. The climate at any one time is then the superposed effect of all the climatic factors, each operating at its respective time scale.

CO_2 in the atmosphere will be dealt with separately, since it has not been a climate-controlling factor in the past, but may become one over the next few hundred years.

1.2 LONG-TERM FACTORS

1.2.1 Galactic Model

The length of the present galactic year is estimated to be 274 m.y. (Innanen, 1966). This order of magnitude has led to the belief that there is a correlation between the last ice ages in the Quaternary, Permian, and Eocambrian, and the passage of the sun through compression lanes in the spiral arms of the galaxy. Hoyle and Lyttleton (1939) suggested that the additional accretion of interstellar gas would increase the luminosity of the sun which would in turn enhance precipitation and the accumulation of ice on earth. By studying the density of interstellar gas in the vicinity of the solar system, Dennison and Mansfield (1976) could not detect any accumulation of material dense and near enough in space to be responsible for the last glaciation. Steiner (1978) on the other hand presented six lead-isotope events (major geologic disturbances in the uranium/lead radioactive system), two of which he correlated with two dated Precambrian glaciations (940–950 m.y. and 2290 m.y. B.P.). He thereby predicted that there were at least four more Precambrian glaciations, corresponding to the four remaining lead-isotope events. By employing a rather flexible galactic model in which the sun gradually spirals inwards, following an excentric orbit, Steiner was able to place all known (and predicted) glaciations on to one model curve of the same galactic parameter. According to Steiner, glaciation occurs when the solar system reaches its apogalacticum (galactic parameter at a minimum). At the perigalacticum (near the centre of the galaxy), the earth experiences warm periods like those of the Jurassic and the Cretaceous.

1.2.2 Geotectonic Model

A Background

According to plate tectonics, the surface of the earth may be divided into a number of lithospheric plates which move relative to one another. Two plates moving away from each other generate an accreting boundary, along which material of the upper mantle upwells to form a mid-oceanic ridge system. Two converging plates, on the other hand, result in the subduction of one of them in the form of a deep-sea trench, or in the consumption of both of them via mountain-building. Two plates gliding by each other produce a transform fault, where little deformation is to be observed.

These plate motions, in particular place accretion and subduction of an oceanic ridge, have a profound effect on global climate. Firstly, by producing variations in the volume of the mid-oceanic ridge system and hence different rates in the rise and fall of sea-level, they give rise to world-wide transgressions and regressions. In turn, changes result in the ratio of land area to ocean area, the amount of cloud cover, the oceanic circulation pattern, and thereby the global albedo. Secondly, plate motions alter the latitudinal distribution of continents. The percentage of land concentrated within the tropics and in the polar regions therefore changes, causing albedo variations and changes in the position of the large cloud masses. Consequently, perturbations of the albedo and medium-term climatic changes are produced. We shall discuss each of these effects in turn.

B Transgressions, Regressions, and Global Climate

Eustatic sea-level changes have been attributed to various causes. The most effective of these are sudden, catastrophic events such as glaciation and deglaciation and dessication and flooding of small ocean basins. They are, however, ephemeral. Potentially, the fastest way to change sea-level on a long-term basis is to change the volume of the mid-oceanic ridge system (Hallam, 1963; Russel, 1968; Menard, 1969; Valentine and Moores, 1972). Other possibilities involving the production of juvenile water at active ridge areas, the continuing differentiation of the lithosphere (which alters the volume capacity of the ocean basins), variations in sedimentation, and crustal shortening through orogeny produce considerably smaller rates of sea-level change (Hays and Pitman, 1973; Pitman, 1978).

After hot upper mantle material is accreted to a plate, it moves away as part of the plate. In the process of doing so, the material cools and subsides. It is this subsidence that bestows a definite depth–age relationship upon the oceans of the world, regardless of whether the ocean has been generated at a fast spreading ridge (e.g. The Pacific), a slowly spreading ridge (e.g. the Atlantic), a ridge spreading with various rates at various times (e.g. the Indian Ocean), or at a ridge that has ceased spreading a long time ago, e.g. the Labrador Sea (Sclater *et al.*, 1971; Lister, 1972;

Sclater and Dietrick, 1973; Parker and Oldenburg, 1973; Oldenburg, 1975; Tréhu, 1975). By making certain reasonable assumptions about the physical and thermal properties of the lithosphere, it has been shown (Parsons and Sclater, 1977) that to a first approximation, the depth of the sea floor varies as the square root of its age, t, for the age range of 0 to 70 m.y.:

$$d(t) = 2500 + 350(t)^{1/2} \text{ m} \quad (t \text{ in m.y. B.P.}) \tag{1}$$

For an older ocean floor, the depth is given by:

$$d(t) = 6400 - 3200 \exp(-t/62.8) \text{ m} \tag{2}$$

Knowing the sea floor spreading history of the oceans, knowing the ocean depth as a function of age, and knowing the length of actively spreading ridge axes at any point in geological time, we can compute the volumetric change in the mid-oceanic ridge system as a function of time. By assuming that the volume of seawater has remained constant, the computed volumetric change can be interpreted as an inverse change in oceanic basin capacity. Thus, an oceanic ridge volume increase of ΔV would mean a decrease in oceanic basin capacity of ΔV.

To translate volumetric changes in oceanic basin capacity to sea-level fluctuations, two corrections must be made (Hays and Pitman, 1973). The first is the isostatic adjustment of the oceanic basins relative to the continents. When water depth increases by an amount h, the ocean floor subsides a distance d, whereby $h = 3.4\ d$, when the upper mantle density is assumed to be 3.4 g cm^{-3}. The change in continental freeboard is hence $h–d = 0.7\ h$. The second correction arises because as sea-level rises, more surface area becomes water-covered. About one sixth of the earth's surface (8.5×10^7 km^2) lies between 0 and 500 m (Sverdrup *et al.*, 1942). Assuming that as sea-level rises the additional area flooded by the sea increases linearly (1.7×10^5 km^2 per each metre rise in sea-level), the actual sea-level change $0.7\ h$ (in m) can be calculated from

$$\Delta V = hA_0 + 170(0.7\ h)^2/2 \tag{3}$$

where ΔV (km^3) is the change in volume of the mid-oceanic ridge system and A_0 (360×10^6 km^2) the present-day area of the oceans.

Using the method outlined above, Pitman (1978) computed the sea-level curve from the upper Cretaceous (85 m.y. B.P.) to mid-Miocene (15 m.y. B.P.) (Figure 1.1). From the mid-Miocene on, glaciation has dominated fluctuations in the sea-level curve, so that the present method is no longer applicable. Two points are of particular importance. Firstly, during the entire period 85–15 m.y. B.P., there was a steady fall in sea-level at a rate equal to or less than 0.7 cm/10^3 yr. Sea-level dropped slowly from 85 to 65 m.y. B.P., quite rapidly through Paleocene and early Eocene, less rapidly during late Eocene, and more rapidly again in the Oligocene. No sea-level rise could be deduced. Secondly, sea-level was about 350 m higher than today in late Cretaceous time. This level is consistent with paleogeographic data (Ronov, 1968) and data on the position of the late Cretaceous shoreline (Sleep, 1976). It

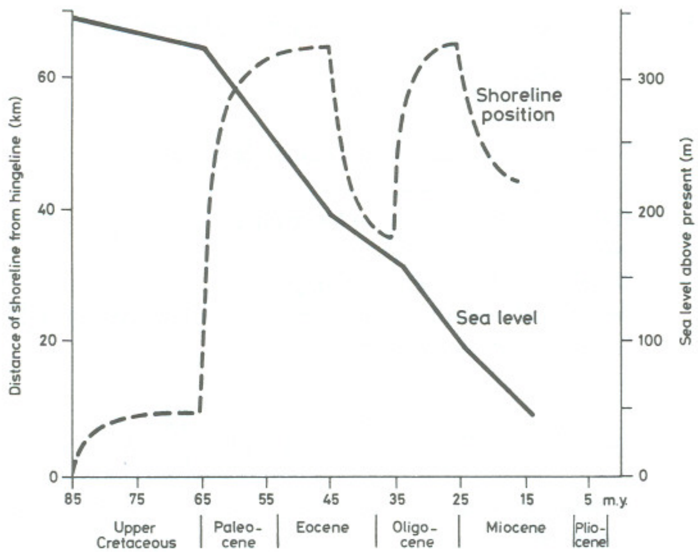

Figure 1.1 Change of sea-level due to volumetric changes in the mid-oceanic ridge system from 85 to 15 m.y. B.P. (million years before present) (solid line). From 15 m.y. B.P. onwards, glacial events determine sea-level curve. Broken line gives the distance of the shoreline from the hinge line (after Pitman, 1978)

also implies that approximately 35 per cent of the present land surface was covered by water. Since water has a much larger thermal capacity compared to rocks on land, this increase in area of sea relative to land must have had a tremendous moderating and stabilizing effect on world climate. In particular, as the extent of epicontinental seas rapidly expanded, convective transfer of heat between low and high latitudes was greatly facilitated, so that a more benign climate was enhanced.

For a long time, it has been believed that transgressions are due to a eustatic sea-level rise and regressions to a eustatic fall. Such a scheme, however, need not necessarily be correct, particularly because a eustatic fall due to glaciation between 85 and 15 m.y. B.P. is not likely. We observe that along Atlantic-type continental margins, kilometres of seaward-thickening, stratified sedimentary sequences overlie a subsided, faulted basement. These sequences are deposited in their entirety within several hundred metres of sea-level. Thus, the sedimentary environment is characterized by rapid subsidence, perhaps at a rate of 2 cm/10^3 yr, so that despite a continuous fall in sea-level, the shoreline remains confined to the shelf. To quantify this interplay of subsidence, sedimentation, and sea-level change, we assume that subsidence occurs about a fixed landward hinge line, and that its rate is constant. Furthermore we assume that sedimentation varies spatially in such a way that the slope of the coastal plain shelf is constant despite subsidence, i.e. sediment infill has kept pace with the subsiding margin. Following Pitman (1978), the rate of change of position of the shoreline ($\mathrm{d}x/\mathrm{d}t$) may now be expressed as:

$$S_{\mathrm{L}}\frac{\mathrm{d}x}{\mathrm{d}t}=R_{\mathrm{SL}}-\frac{xR_{\mathrm{ss}}}{D}+S \tag{4}$$

where S_{L} = slope of the surface of the coastal plain and shelf,
R_{SL} = rate of sea-level change,
x = distance of shoreline from hinge line,
R_{ss} = subsidence rate at shelf edge,
D = distance of shelf edge from hinge line,
and S = additional uniform sedimentation, to allow for possible sedimentation on the coastal plain.

Integrating gives:

$$x=\frac{D}{R_{\mathrm{ss}}}\left(R_{\mathrm{SL}}+S\right)-\left(\frac{R_{\mathrm{SL}}D}{R_{\mathrm{ss}}}+\frac{SD}{R_{\mathrm{ss}}}-x_0\right)\exp\left(-\frac{TR_{\mathrm{ss}}}{DS_{\mathrm{L}}}\right) \tag{5}$$

in which x = position of shoreline after time T and x_0 = position of shoreline at the beginning of interval T.

For large time intervals (10^7 yr), reasonable values of D (250 km), S_{L} (1/5000), and R_{ss} (2.5 cm/10^3 yr) yield

$$\exp\left(-TR_{\mathrm{ss}}/DS_{\mathrm{L}}\right)=1/148$$

so that

$$x\approx\frac{D}{R_{\mathrm{ss}}}\left(R_{\mathrm{SL}}+S\right)\text{ or }R_{\mathrm{SL}}\approx\frac{X}{D}R_{\mathrm{ss}}-S \tag{6}$$

i.e. the shoreline tends to stabilize at a point where the rate of change of sea-level equals the difference between the subsidence rate at this point and the sedimentation rate. From the late Mesozoic to the present, excepting glacial effects, the sea-level has been continuously falling. An increase in the rate of fall would require a seaward migration of the shoreline (equation 6) and hence a regression. Likewise, a decrease in sea-level fall would result in a transgression. Hence, transgressions and regressions need not correspond to maxima and minima in sea-level stand. A transgression will occur when sea-level is made to rise more rapidly or to fall more slowly; and if sea-level is made to rise more slowly or fall more rapidly, a regression will result.

In Figure 1.1 the position of the shoreline calculated by equation (5) using the sea-level curve for 85–15 m.y. B. P. is depicted as a broken line. Here the Eocene transgression and the Oligocene regression so often discussed (e.g. Hallam, 1963) are seen to be results of changes in the rate of fall of sea-level.

With the position of the shoreline known, the amount of land surface relative to sea surface can be calculated. Since the reflectivity and its variation with the angle of incident radiation changes from land to ocean, transgressions and regressions cause changes in the global albedo and therefore in the surface temperature of the earth. These changes are, of course, reflected in perturbations in the climate prevailing at the time.

C Changes in the Distribution of Continents and Global Climate

That plate motions alter the spatial distribution of continents and hence the percentage of land within a particular latitudinal zone is obvious (see e.g. Smith *et al.*, 1973; Briden *et al.*, 1974; Smith and Briden, 1977). Here, we shall take a brief look at the effect of these alterations on global climate.

The global albedo A may be expressed as:

$$A = A_C\alpha_C + A_L(1-\alpha_C)\alpha_L + A_W(1-\alpha_C)(1-\alpha_L) \qquad (7)$$

where A_C, A_L, and A_W are the average albedos of cloud, land, and water respectively, α_C the fractional cloud cover and α_L the subaerial fraction of the earth's surface. Of the three terms, it appears that the albedo of cloud is the most important (Schneider, 1972). Thus changes in the amount of cloud cover and in its distribution as a result of plate motions must be considered in any theory of climatic change. It has been suggested that the percentage of cloud cover for the earth has remained practically unchanged over evolutionary time periods (10^8–10^9 yr) (Henderson-Sellers, 1979). However, over shorter periods (10^4–10^7 yr), cloud masses could have undergone changes in their distribution pattern with different continental configurations. Areas of high cloud concentrations are not only determined by atmospheric circulation, they are also often found over open oceans. Computer model studies show that a 10 °K increase in surface temperature could lead to a 10 per cent decrease in cloud cover given the present continental configurations, and this decrease would in turn result in a six per cent decrease in the global albedo (Henderson-Sellers, 1979). A 10 °K increase in temperature appears extreme since a 5 °K change is believed typical of glacial–interglacial conditions (Bryson, 1974). However, the interesting point is that such a change produces an inherently stable system, i.e. one with negative feedback in terms of cloud formation. This dynamic stability could be disrupted by possible variations in cloud positions as a result of continental drift. Computations based on different climate models show that such variations tend to perturb the albedo, and though the effects are secondary, they could be important in medium-term global climatic changes.

Changes of continental configuration also effect the land and water terms of equation (7), and these effects are better understood. To a first approximation, we take the land albedo A_L to be constant at 15 per cent, and consider the water albedo A_W to be constant (five per cent) from the equator to 50°N or S, thereafter increasing linearly to 12 per cent at the poles. This change implies that the albedo contrast $(A_L - A_W)$ decreases with increase in latitude.

When plate motions concentrate land masses in the circumpolar regions, the tropics become cloudier and wetter. In the polar regions, replacement of A_W by the slightly higher A_L results in a slight cooling, but the larger albedo contrast within the tropics produces a substantial warming. On the whole, the steeper poleward temperature gradient is overshadowed by a globally warm climate (Cogley, 1979). When continents drift into the tropics, the effect of the albedo contrast produces a rapid cooling at low latitudes and a slight warming near the poles. Global cooling overshadows the reduced poleward temperature gradient.

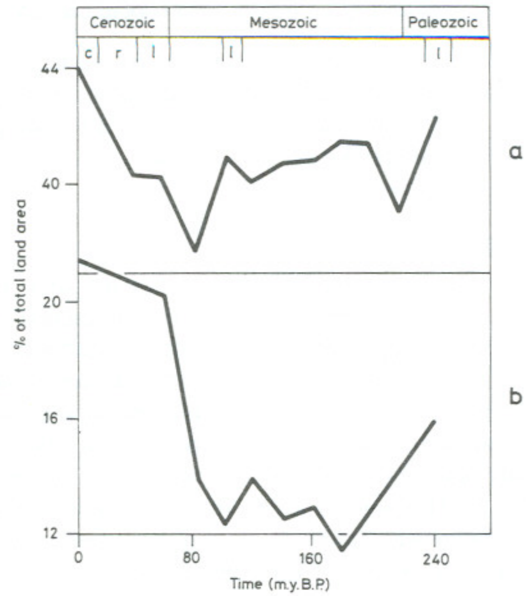

Figure 1.2 (a) Percentage of land area between 30°N and 30°S. (b) Percentage of land mass poleward of 60°N and 60°S; c, r, l stand for continental, regional, and local glaciation, respectively (after Cogley, 1979)

Sellers and Meadows (1975) have shown that land mass movements cause albedo changes, in the sense discussed above, and may be responsible for triggering or reinforcing glacial periods. They state that land mass concentrations near the poles are conducive to glaciations. The argument above (Cogley, 1979) suggests that perhaps concentrations of land both around the poles and in the tropics are necessary, as both represent climatic instabilities.

We plot in Figure 1.2 the percentage of land areas berween 30°N and 30°S (curve a), and polewards of 60°N and 60°S (curve b) respectively. The Cenozoic cooling trend, the glaciations since mid-Miocene, and the end of the extensive Paleozoic glaciations (which is as far back in time as these curves extend), all seem to be accompanied by high percentages of land both in the polar regions and in the tropics.

1.2.3 Effect of a Faster Earth's Rotation Rate in the Geological Past

Growth patterns of shells and corals suggest that in the late Precambrian about 1.5 $\times 10^9$ yr ago, the earth's rotation rate was 2–2.5 times greater than at present (Mohr, 1975). The effect of such a high rotation rate on climate, although speculative, has been evaluated by Hunt (1979) using a numerical model of atmospheric circulation. By assuming that cloud cover, surface albedo, and CO_2 content in the atmosphere were all the same as today, he deduced that a faster rotation of the earth would result in a reduction in the scale size and intensity of the zonal winds, a decrease in poleward heat transport, and a more equatorial location of the tropospheric westerly jet stream. In Precambrian times, therefore, the surface wind stress, being proportional to the square of the wind velocity, would have been drastically reduced. The oceanic gyre would have been weakened and wind-induced vertical mixing in the oceans would have been much less significant, so that the surface isothermal layer would be shallower and warmer. The implication is a diminished poleward transport of heat by the oceans, and hence warmer tropics and a colder polar region. Hunt speculated that for the Precambrian, the climatic changes deduced from a faster earth rotation alone could lead to glaciations. By 6×10^8 yr ago, the much slower rotation rate had improved the climate to such an extent that the maintenance of glacial conditions was no longer possible. The Precambrian glaciations thereby came to an end.

Whether Precambrian glaciations were indeed caused by a faster rotation of the earth may be debatable. The effect of a faster rotation on global climate, however, clearly deserves closer scrutiny.

1.3 MEDIUM-TERM FACTORS

1.3.1 Tectonic Pulses

The long-term geotectonic model was based on constant rates of tectonic subsidence, uplift, and spreading for periods of millions of years. Closer examination reveals that tectonic events occur in pulses which are followed by times of tectonic quiescence. The frequency of such pulses is of the order of ten thousand to a few hundred thousand years. At times they come in series, are globally distributed, and have consequently led to major mountain-building epochs, such as the Caledonian or the Variscian. At other times they are really confined and of shorter duration. Nevertheless, they are accompanied by the same phenomena as discussed earlier, i.e. subsidence, uplift, transgression, or regression, and their impact on regional climate can be substantial. The likelihood even exists that regional tectonics proceeding in 'climate-strategic' parts of the earth may influence global climate through a series of feedback systems. As an example we consider Southeast Europe during the Quaternary.

The Quarternary started with the Akčagylian marine transgression which transformed wide areas of southern Russia into a shallow sea by uniting the Black Sea, the Caspian Sea, and Lake Aral. The sea measured more than 2000 km in a north-south direction and almost 1000 km at its widest east-west opening (Figure 1.3). Shallow water conditions prevailed for almost two million years and the Danube drained into the Caspian Sea. About 400 000 years ago and continuing towards the

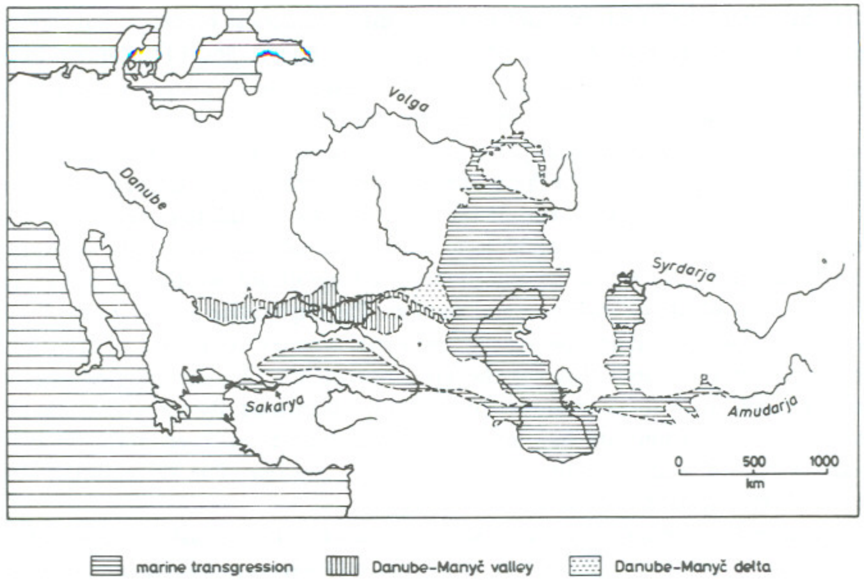

Figure 1.3 Akčagylian transgression in the Black Sea–Caspian Sea–Lake Aral region (after Degens and Paluska, 1979)

present, pulses of tectonic activity caused rapid subsidence of a basin chain extending from the Caspian Sea to the lowlands of the Po valley; rates of subsidence were as high as 1-2 cm yr^{-1} with a maximum about 200 000 years ago (Degens and Paluska, 1979; Paluska and Degens, 1979). Synchronous with the subsidence in the south was a general uplift in parts of Russia, Anatolia, Central Europe, and Scandinavia. For example, within a hundred thousand years Scandinavia became uplifted by one kilometre and more, while at the same time the Caspian and Black Sea basin floors sank by the same amount. This vertical motion provided the kinetic energy for massive erosion and formation of moraines in the aftermath of glacial events. The fact that the past three major ice ages coincide with a major tectonic pulse in Europe suggests that rapid tectonism leading to regional changes in land/water ratio, albedo, topography, orography, and bathymetry may contribute to global climatic alterations due to the 'weather-strategic' position of this area.

1.3.2 Orbital Periodicities of the Earth

From the periodicities in the tilt angle of the rotation axis (41 000 years), the precession of the earth's axis (21 000 years), and the eccentricity of its orbit (93 000 years) Milankovitch deduced a curve spanning the last 300 000 years for the radiation values averaged over the months March to September. Today the solar radiation at 50°N is 847 langley day^{-1} (or 35 447 kJ m^{-2} day^{-1}). For the past 300 000 years, the Milankovitch curve shows fluctuations between –35 and 50 ly day^{-1} with respect to the present day value. These differences in radiation are, according to the Weertman model (1976), sufficient to cause glaciation on northern high latitude continents. The Milankovitch curve suggests that the earth has recently finished a phase for which the solar radiation incident on its surface was very high. At present, this incident radiation is 30 ly day^{-1} less, so that cooling is expected. Weertman's model predicts the beginning of a new ice age within the next several hundred or several thousand years. It should last approximately 60 000 years, i.e. its duration should be similar to the Wisconsin Ice Age. However, Weertman also pointed out that his model requires much higher precipitation rates than those prevalent in the continental polar areas today. Furthermore, he also noted that the occurrence of ice ages in model calculations is largely a consequence of choosing the 'right' model parameters.

Orbital periodicities have recently been suggested by several scientific schools to be the principal cause for climatic change of the type recorded by the glacial-interglacial pattern of the Quaternary. For a comprehensive account see Imbrie and Imbrie (1979).

1.4 SHORT-TERM FACTORS

1.4.1 Volcanic Dust

Large volcanic eruptions eject immense amounts of volcanic dust into the stratosphere. These dust particles increase the reflected portion of the incoming solar radiation, thus effectively diminishing the proportion of solar energy reaching the earth's surface. Consequently, temperatures at ground level become lower, and colder weather is to be expected.

In 1815 Mount Tambora in Sumbawa in Indonesia exploded, ejecting over 100 km^3 of rocks and ash. Large amounts of ash were blasted into the stratosphere, whence it started to distribute around the globe. Throughout May, 1816, the weather in eastern North America and parts of Europe stayed cool to become what was the coldest summer on the temperature record for New Haven (Connecticut) kept at Yale College (Stommel and Stommel, 1979; Figure 1.4). Although whether this apparent cooling can be directly attributed to the Tambora eruption is disputable (Self and Rampino, 1979; Stommel, 1979), that the eruption did play a role in the weather change is perhaps not controversial.

A global cooling was evident in the temperature trend following the eruption of Krakatoa in 1883 (Mitchell, 1977). Likewise, the injection of aerosol into the atmosphere accompanying the burst of Mount Agung on Bali in March, 1963, probably caused low Pacific and high tropical stratospheric temperatures (Newell and Weare, 1976).

Schneider and Mass (1975) calculated temperature changes from a climatic model in which dust concentrations and solar activity were taken into consideration. Solar transmission measurements at the Mauna Loa Observatory carried out after the

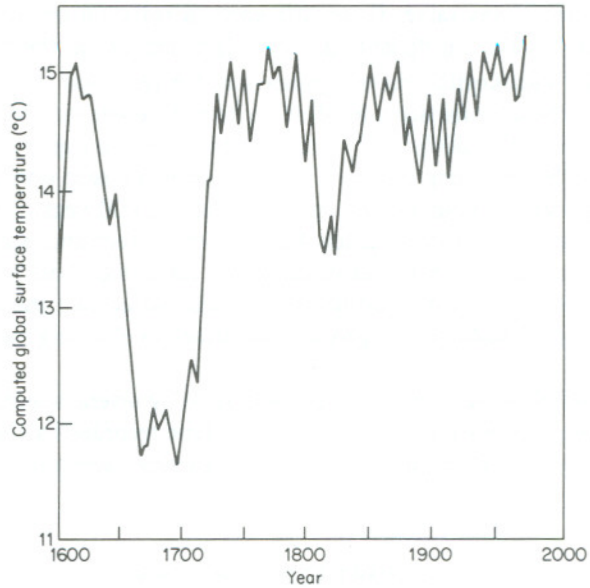

Figure 1.4 Mean June temperature records for New Haven, Connecticut, over a period of 70 years (after Stommel and Stommel, 1979)

Agung eruption showed a reduction in the solar energy directly incident on the earth by two per cent due to volcanic dust. By arguing that nevertheless, three quarters of this missing two per cent eventually reached the earth's surface via scattering, they calculated that the total solar parameter S is reduced by 0.5 per cent as a result of an Agung-scale eruption.

One input function to the climatic model of Schneider and Mass (1975) is the dust veil index obtained empirically from historic records of volcanic eruptions (Lamb, 1970). A second input function is that component of the solar parameter which is dependent on the sunspot number N:

$$S(N) = 1.903 + 0.011\,N^{0.5} - 0.0006\,N \text{ cal cm}^{-2} \text{ min}^{-1}$$

This functional dependence is based on solar constant measurements of Kondratyev and Nikolsky (1970), and its validity is not universally accepted. For the quiet ($N \approx 0$) and the very active sun ($N \approx 200$), it yields values for the sunspot component of the solar parameter, $S(N)$, about two per cent lower than those for a normal solar activity ($N \approx 80$).

Schneider and Mass expressed the solar parameter S as the sum of two factors:

$$S = S(N) + S(D)$$

where $S(D)$ is that component of the solar parameter which is dependent upon dust concentration in the atmosphere. This sum S constitutes the forcing input to their

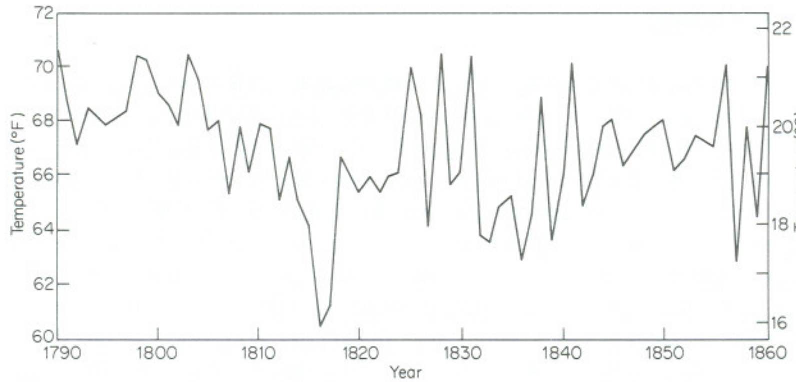

Figure 1.5 Mean global surface temperatures computed by the climate model of Schneider and Mass (1975)

climatic model from which the global surface temperature variation since 1600 can be calculated (Figure 1.5).

The resulting temperature curve is in fair agreement with well-known climatic features, such as the little ice age which is explained by the Mauder minimum in the solar activity, and the cold early eighteen hundred which are explained by both a high dust index and a low solar activity.

$\delta\ ^{18}O$ values (believed to be a measure of temperature) as a function of age have been determined using the Camp Century ice core from Greenland (Dansgaard *et al.*, 1971). This function has been shown to correlate significantly with the envelope of the lunar tidal stress curve at 60°N (Roosen *et al.*, 1976). Roosen *et al.* (1976) and Lamb suggested that this correlation implies a dependence of climate on lunar tides. Indeed, they argued that volcanic eruptions are statistically apt to occur when high tidal stresses prevail in the lithospheric plates. Such eruptions increase the stratospheric dust content, thereby leading to a cooling in the climatic trend. Coincidence of the 179-year lunar tidal period with the 180-year period of the $\delta\ ^{18}O$ maxima is strong evidence for such a dependence. However, it should be noted that vulcanism (and dust production) is not the only mechanism by which lunar tides might induce periodic climatic variations.

1.4.2 Tidal Forces and Planetary Alignments

Gribbin (1973) observed that the alignment of the outer planets also follows a periodicity of 179 years. He thereby speculated that the tidal forces due to these planets modulate the amplitudes of the individual 11-year sunspot cycles and that the periodicity of this modulation is 180 years. The nature and magnitude of this tidal influence on the sun have as yet not been discussed in a climatological context.

1.4.3 Solar Activity

The only record of solar activity which can be analysed for periodicities is that of sunspots (Schove, 1955). Fourier analysis of this record (Ekdahl and Keeling, 1973) reveals that well-known 11-year sunspot cycle, a cycle of about 80 years, and evidence for the 180-year cycle. Longer cycles can only be deduced from earthbound climatic records with the assumption that the changes found have in fact been produced by the sun. It has often been pointed out (e.g. Smith and Gottlieb, 1975) that the sunspots themselves cannot cause any significant change in the radiation balance of the sun. Even at high sunspot numbers only a 0.1 per cent change in the visible solar flux could result from direct shading by sunspots. The mechanism of climate influence must therefore be more complicated. Dicke (1979) suggested that sunspots are only the surface expressions of a 'solar chronometer' which modulates the luminosity of the sun. This magneto-fluid dynamic oscillator, deeply buried inside the sun, causes magnetic fields to float to the surface of the sun to create sunspots years after the luminosity change occurs. A full magnetic cycle of the sun contains two sunspot cycles, i.e. the magnetic cycle has a length of 22 years. From one 11-year to the next 11-year cycle, the magnetic sign of the sunspots changes: after a cycle with positively charged sunspots on the northern hemisphere of the sun, a cycle with negative spots follows. Epstein and Yapp (1976) found that the deuterium/tritium (D/T) isotope record of a particular bristle cone pine tree shows a 22-year periodicity. By comparing this D/T isotope curve with the sunspot cycle, Dicke (1979) showed that the curves for the latter lagged the former by about 13 years. This time lag may be identical with the time lag between the actual luminosity change of the sun and the appearance of sunspots. A climatic change on the earth reflects of course changes in the solar luminosity.

The evidence for an 11-year frequency in climatic data, however, is intriguing. Drought data have been found to correlate with sunspots (Roberts, 1975) as well as glacier recession, temperatures, and a wealth of other meteorological parameters (Bandeen and Maran, 1975). For Lake Van in eastern Turkey, lake level oscillations have been correlated with sunspot activity (Kempe, 1977). These oscillations not only show a time lag of one year (or following Dicke, 12 years) but evidence for a 10-year solar (?) cycle could be found in the early Holocene varved lake sediments.

Since concentration and generation of ^{14}C in the atmosphere is a function of the intensity of cosmic rays, it might be correct to call ^{14}C periodicities solar. Suess (1970) found periods of 2400 and 405 years. Dansgaard and co-workers (1971) originally found 2100- and 350-year cycles in the oxygen isotope data in the Camp Century ice core, which were standardized to give a time scale with the same periodicities as those of the ^{14}C data. A review of a variety of other cycles of climatic indicators is given by Schove (1978), while Flohn (1978) gives details on acyclic, abrupt events for comparison.

1.5 CO_2 AND CLIMATE

1.5.1 Introductory Remarks

The gigantic climatic experiment mankind is currently undertaking, i.e. the combustion of fossil fuels (which have been accumulated over much of the earth's history) within geologic-zero-time, has focussed our attention on this climatic thermostat. The temperature of the atmosphere is largely determined by the concentration of CO_2, because molecular CO_2 absorbs short wave radiation and releases it as infra-red. Earth is habitable only because of the 300 p.p.m. CO_2 in its atmosphere. If this concentration were substantially lower, then life would have been impossible on this planet because it would have been too cold.

Over geologic history an ingenious sytem has developed whereby CO_2 is redistributed between atmosphere, vegetation, oceans, and rocks by the carbon cycle (Bolin *et al.*, 1979). This cycle works fast for the atmosphere. If no reflux occurred the air would lose its CO_2 to vegetation and ocean within seven years; small surpluses are almost immediately stored away in the oceans. At present perhaps as much as 10 billion tons* of anthropogenic carbon are added annually to the atmosphere in the form of CO_2, five billion tons of it from industrial activities and the rest of it from vegetation and soil destruction (Bolin *et al.*, 1979). About 2.5 billion tons of carbon stay in the atmosphere, corresponding to an increase of about 1 p.p.m. per year (Keeling and Bacastow, 1977; Lowe *et al.*, 1979). The remainder is accounted for by the 'CO_2 buffer capacity' of the earth by way of the carbon cycle. The future atmospheric CO_2 trend is difficult to predict because of uncertainties in the pattern of industrial and agricultural CO_2 production.

1.5.2 Natural Processes

Carbon is stored in the earth's crust principally in the form of coal, oil, natural gas, kerogen (finely disseminated organic matter in sediments), and carbonates. Part of this carbon becomes reactivated by magmatic and volcanic processes and re-enters the atmosphere and hydrosphere mainly as dissolved or gaseous CO_2. Furthermore, erosion exposes fresh rock formations to the action of weathering and liberates additional fossil carbon. In the process of denudation, part of the vegetation cover and soils is removed and carried to the sea as particulate organic matter or in the dissolved form. A substantial fraction of the plant material, however, becomes oxidized to CO_2 and escapes into the air. In the following, we shall briefly examine to what extent natural processes can alter the atmospheric CO_2 level.

Volcanic activity presently produces some 0.05 billion tons of carbon per year, a value over one hundred times less than the estimated annual discharge rate of

*Here one billion $\equiv 10^9$

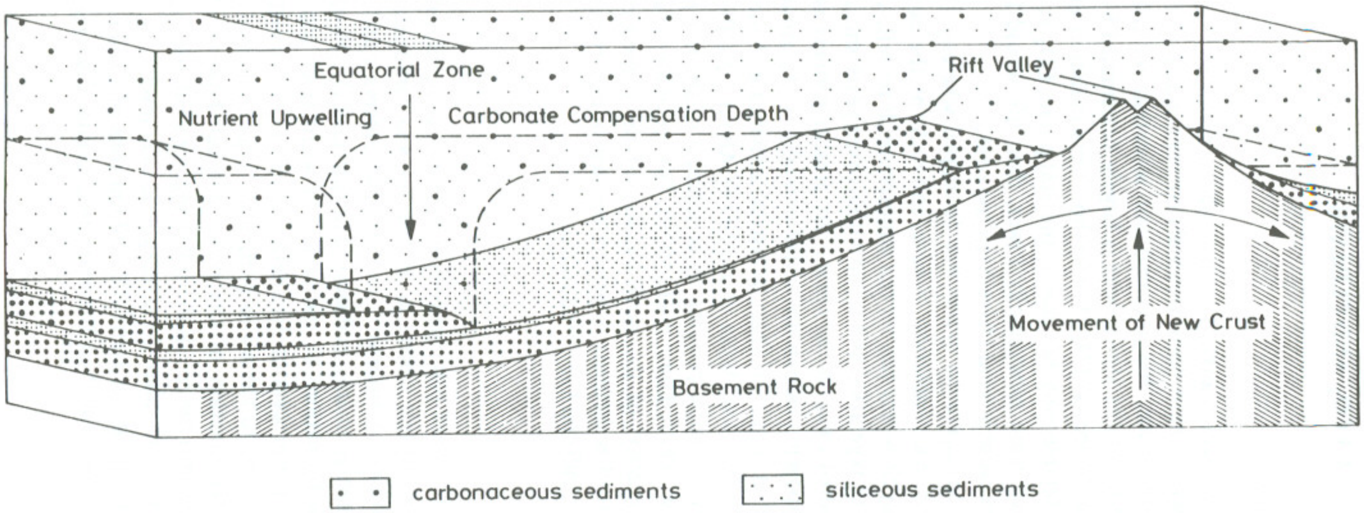

Figure 1.6 Dynamic model (schematic) showing the effect of the carbonate compensation depth (CCD) on the kinds of bottom sediments that accumulate (after Heezen and Macgregor, 1973). Note the movement of new crust which is indicated by a vertical striation pattern reflecting magnetic reversals. Sea floor spreading will move areas into and out of the influence of the CCD. If euxinic conditions develop, due to density stratification, the lysocline and CCD will gradually move up towards the pycnocline; under such conditions a hiatus can be created in abyssal regions that are far removed from land (Degens and Stoffers, 1976)

man-made CO_2. Tectonism and volcanic activities were episodic in the geological past, being linked to the rate of plate accretion and subduction. We live today in a tectonically active phase which could imply that the above volcanic production figure of 0.05 billion tons carbon is a high one on the scale of CO_2 emissions. An additional factor is worth mentioning. Magmatic CO_2 released from the ocean floor into the deep sea can be incorporated into the hydrosphere without substantial changes in the CO_2 partial pressure of the atmosphere. The carbonate compensation depth in the ocean simply adjusts itself in accordance with this development (Figure 1.6).

When the sea floor lies at a level beneath the lysocline, below which carbonates do not exist in the solid phase, carbonaceous sediments cannot be deposited. The depth of the lysocline therefore determines the amount of carbonate which can be removed from the water column. On the other hand, interactions at the air–sea interface control the release of CO_2 into the atmosphere and the biosphere. Things appear quite harmless at first sight. Changes in water density induced by temperature or salt content stratify the sea in mid-water. However, if this stratification is maintained for an extended period of time, mixing of water masses is restricted. The stratification boundary may become so stabilized that it only moves up and down in response to changes in water temperature, tectonic activities, or other environmental disturbances; but it may rarely break up entirely (Degens and Stoffers, 1976, 1977). In such a situation, molecular oxygen would remain abundant in the upper layer but would gradually drop to zero below the density boundary. In short, a euxinic environment (oxygen being absent) is created.

The oscillating pycnocline (stratification boundary due to a rapid density increase with depth) constitutes one of the most powerful means of keeping a status quo for the CO_2 system. We illustrate schematically in Figure 1.7 the operational principle, and its effect on carbonate and sapropel formation.

Oscillating lysoclines and pycnoclines maintain a kind of steady state and shield the CO_2 system from perturbations coming from the atmosphere (air temperature), the lithosphere (volcanic degassing), and the biosphere (primary productivity and decomposition). ^{13}C measurements on tree rings (Freyer, 1978; Stuiver, 1978) suggest that carbon^{-13} isotopic composition of atmospheric CO_2 has decreased by about 2 per mil (parts per thousand) over the past hundred years due to the alteration of the natural CO_2 isotopic composition by the addition of fossil fuel CO_2 (Suess, 1955). The pre-industrial atmosphere has a $\delta\ ^{13}C$ of –5 per mil which conforms with the $\delta\ ^{13}C$ of CO_2 presently evolving from the mantle (Taylor *et al.*, 1967). These data suggest that the 'master switch' for CO_2 control lies deep inside the earth. In view of the huge carbon reservoir in the crust and mantle (Bolin *et al.*, 1979), the dynamics of crustal movement (Engel *et al.*, 1974), and the buffer capacity of the ocean (Takahashi, 1975; Broecker and Takahashi, 1978), this interpretation appears to be realistic.

Thanks to the ocean, carbon dioxide levels in the atmosphere have probably remained fairly uniform, at least since life appeared on earth. Consequently, CO_2

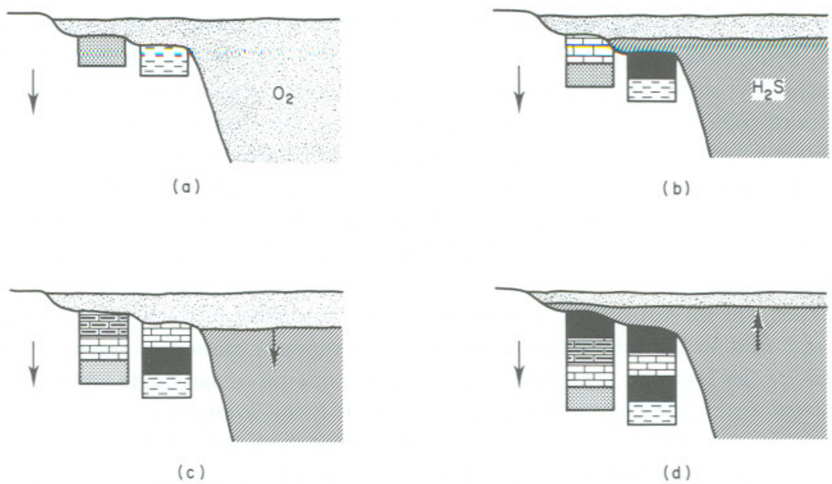

Figure 1.7 Formation and evolution of stratified waters (after Degens and Stoffers, 1976): (a) fully oxygenated sea and deposition of sand (dotted) and clay (dashed areas); (b) stratified sea (density boundary in mid-water); carbonates (brick-wall pattern) form above pycnocline and euxinic sediments (black) below pycnocline in anoxic water; (c) lowering of pycnocline will lead to spasmodic deposition of marl (brick-wall dashed pattern) and carbonates; (d) upward progression of pycnocline will extend euxinic conditions to shallower parts of the sea

is believed to have had only a slight impact on paleoclimate (Degens, in press). In contrast, the present man-made input of CO_2 into our atmosphere is so dramatic that serious consequences for climate and environment are to be expected in the near future.

1.6 SYNOPSIS

In the geologic past likely causes of climatic changes were:

(A) Extraterrestrial

1. variations in the properties of interstellar space or gravity due to galactic rotation (length of the galactic year, ca. 270×10^6 years)
2. variation in the rotational rate of the earth in the geologic past
3. periodic or aperiodic changes in the luminosity of the sun (various solar cycles of 11-, 22-, 80-, 180-year periods)
4. periodicities in the rotation of the earth around the sun (Milankovitch curve; 10^4–10^5 year periods)
5. planetary and lunar tidal effects (180-year period).

(B) Terrestrial (Modulating the Incoming Solar Energy)

1. global changes in the ocean to continent ration (10^6–10^7 year range)
2. redistribution of continents into different latitudes and elevations (10^6–10^7 year range)
3. changes in land/water ratio and elevation due to regional tectonic pulses (10^4–10^5 year range) in climate-strategic parts of the earth
4. episodic changes in concentration of volcanic dust in the atmosphere (1–10 year range).

Extraterrestrial causes are generally of a periodic nature, whereas terrestrial ones proceed within a certain time range.

Geological data indicate that the principal 'dictator' of terrestrial climate is geotectonics. Long-term changes can be attributed to global tectonic activities, mid-term changes to regional tectonic events, and short-term changes to volcanic dust injections into the upper atmosphere. Superimposed are modulating effects of extraterrestrial origin.

Our planet is habitable only because of the presence of carbon dioxide and water vapour in the atmosphere. Should these substances somehow be removed, the effective surface temperature would be reduced by about 25°C (Rasool and de Bergh, 1970). CO_2 must therefore have played a very important role in the stabilization of global temperatures. Over at least the past 500 million years, CO_2 levels in the atmosphere have stayed uniform due to the buffering effect of the carbonate system in the ocean. This situation may, however, change significantly in the near future. While natural processes will continue to influence climate in the ways described earlier, the role of man-made CO_2 is expected to increase, so that it may become the dominant factor of global climate in the next few hundred years.

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