Abstract

In the first chapter some contributions to maritime climatology and air-sea interaction are given. They include time-series of fluxes of sensible heat and evaporation, comparisons of maritime wind observations and geostrophic winds, large-scale changes of Atlantic circulation patterns and the role of equatorial upwelling/downwelling for the exchange of water vapour and carbon dioxide.

These examples have been selected according to their role for changes of the CO₂ content under natural conditions and for a rational interpretation of the enigma of "abrupt" climatic changes. At the interannual scale much smaller counterparts ("jumps") do occur.

Introduction

Oceans cover 71 percent of the earth and control climate due to their storage capacity of heat and water. Physical understanding of recent climatic fluctuations depends, to a large extent, on our knowledge of air-sea interaction and its evolution since the beginning of instrumental observations. PALTRIDGE and WOODRUFF (1981) suggested that the global temperature trend of the sea surface temperatures (SST) is delayed with respect to air temperature (AT) indicated by northern continental stations. FOLLAND et al. (1984) analysed a global set of SST and maritime AT (MAT) data, applying some corrections. A comparison between MAT and AT (his Fig. 2d) at the northern hemisphere seems to indicate similarly a delay, at least between 1900 and 1940; no critical discussion can be given now. These marine observations now become increasingly important, because of the need for long homogeneous records at a global scale and for heat flux studies; the latter should be a key in a geophysical interpretation of climatic fluctuations.

Air-sea interaction plays also a key role in another newly arisen problem. Increasing evidence shows climatic fluctuations of the geological past did not happen at a $10^4$-$10^5$ year scale, as expected from the earth's orbital fluctuations, but at a much shorter scale in the order of a century (FLOHN 1979, 1984). In European physical laboratories - at Bern (OESCHGER et al. 1983a, 1983b) and Grenoble (LORIUS and RAYNAUD 1983) - a surprising discovery was made: such "abrupt" changes are accompanied by simultaneous variations of the CO₂ content of the atmosphere. A possible working hypothesis (FLOHN 1983) interprets this coincidence as caused by the air-sea exchange of CO₂ and H₂O; at any rate, it emphasizes the role of the greenhouse effect for climatic change, be it natural or man-made.

A) Air-Sea Interaction and Atmospheric Circulation

1) Sea surface temperature and air temperature

Maritime observations include wind, state of the sea, SST and less frequently humidity of the air. Techniques of measuring SST and MAT varied with time, and new techniques - measuring SST in buckets or at the intake of cooling water in the engine room, MAT in screens or with sling psychrometer - led to systematic errors which are difficult to quantify and to correct. While intake measurements of SST tend to be 0.4-0.6°C higher than bucket measurements - see, among many others, BARNETT (1984), RAMAGE (1984) -, recent evaluations seem to indicate that the correction is smaller, e.g. (FOLLAND et al. 1984) +0.3°C for data before 1940. German observations (WEBER 1984 and earlier results) show however, that the difference SST-MAT which is so important for the flux of sensible heat (and to some degree also of latent heat), varies also systematically before 1920. At a representative field in the tropics (5-10°N, 70-80°E) this difference drops from 0.25°C (1902-13) to 0.16°C (1922-37), but rises again to 0.55°C (1958-67), using all available observations. While the diurnal temperature variations of SST remain constant, that of MAT during the middle period is definitely higher (Fig. 1) than during the other periods. This reflects a change in measuring MAT: thermometers in screens gave higher daytime values due to the solar heating of the ship and bad ventilation. Together with the uncertainty of wind speed estimates this yields unrealistic long series of the flux of sensible heat H (RAMAGE 1984); no absolute calibration is possible. The same is true for evaporation LE. However, it is possible to evaluate spatial and time variability of these parameters from series after 1948, especially in well-documented areas along the main shipping routes.

2) Time series of sensible heat flux and evaporation

In most tropical and subtropical latitudes H is positive and relatively small, compared to evaporation; the Bowen ratio is usually of the order 0.05-0.10, in temperate latitudes of the order 0.2-0.3. In wintertime along east coasts of continents, where very cold arctic air is advected over warm water, very large fluxes of sensible heat occur. Negative values can occur during spring and summer along coasts with prevailing advection of warm continental air above cool water: this suppresses convection. The same is true for areas with upwelling cool water, including the equatorial Pacific and Atlantic. The effective difference is that between oceanic skin temperature and true air temperature just above the interface; SST and MAT give only a rough estimate which nevertheless proves to be quite sensitive. As an example the annual variation of air-sea exchange parameters are given for an area off the southern coast of the Arabian peninsula (Fig. 2), along the shipping route to the Persian Gulf, with great seasonal variations. In this area the reversal of SST-MAT occurs between May and September, with negative values of H; SST drops nearly 6°C between May and August. The time variability of seasonal values of H and LE at the ocean surface is surprisingly high (Table 1).

While at low latitude oceans only the sign of H is important, LE often provides the largest term in the surface energy budget. However, reliable ship measurements of relative humidity or dew-point are rather rare. In the southern North Sea, both E and H depend on season and wind direction; a long marine fetch is only given with wind directions between about 300° and 360°. Here E becomes negative between March and June, but limited to the land-
borne wind direction between $0^\circ$ and $270^\circ$ (Fig. 3); the field of $H$ is similar, but the area with negative (downward) flux of $H$ is larger, slightly displaced towards a later month. In this investigation (KATZSCHNER 1978) the bulk aerodynamic formula is used with a constant drag coefficient of $1.3 \times 10^{-3}$. Unfortunately the correlation between the thermal stability SST-MAT, which is available in the Historical Sea Surface Temperature Data since 1861, and the saturation deficit $q_s - q_a$ ($q = \text{specific humidity}$) is positive but not very high: at the equatorial Atlantic it reaches, under purely oceanic conditions, 0.62 resp. 0.72, in the eastern Pacific west of the Galapagos 0.49 and in the coastal upwelling region off western Africa only 0.29. Any attempt to estimate $q_s - q_a$ (and then LE) from the available data cannot lead to really reliable values. This is the more true since the correlation between wind speed and SST-MAT is small, partly even negative.

In equatorial upwelling regions LE is drastically reduced. At tropical oceans the daily amount of E varies around 4-5 mm/d; in the equatorial areas of the eastern Pacific and Atlantic it can drop, in years with undisturbed upwelling and SST $\sim 23^\circ$, to values near or below 1 mm/d (HENNING-FLOHN 1980); see also Chapter A 5.

Since upwelling is coupled with thermal stability (SST$<$MAT), the relative humidity at the boundary layer increases to values around 90 percent, to be compared with average values around 78 percent. This fact should not be interpreted as indicating higher evaporation: indeed E is low, but the water vapour is trapped in the boundary layer, while tropospheric subsidence leads to a minimum of precipitable water in the atmospheric column above (GRODY et al. 1980).

3) Ocean surface winds and geostrophic winds

Few studies are available comparing, above the oceans, observed surface winds and geostrophic winds derived from the sea-level pressure field. This gap can be filled at least over the North Atlantic, where fairly reliable daily pressure maps are available since 1881. Three comparable fields have been selected:

<table>
<thead>
<tr>
<th>Field</th>
<th>Boundaries</th>
</tr>
</thead>
<tbody>
<tr>
<td>34</td>
<td>46-52°N, 23-30°W</td>
</tr>
<tr>
<td>44</td>
<td>40-46°N, 20-30°W</td>
</tr>
<tr>
<td>62</td>
<td>27-32°N, 17-25°W</td>
</tr>
</tbody>
</table>

Here the boundaries of these fields are fairly similar to the $5^\circ \times 10^\circ$ grid of the pressure fields. From these examples, some tentative results shall be outlined. While field 34 is situated right at the North Atlantic main shipping route (British Channel to American east coast), the position of field 44 is just south of it. Field 62, along the shipping route from Europe to South America, lies off the coast of Morocco, i.e. in the N.E.-trade region.

Only a few (preliminary) results shall be given here (Table 2), limited to the annual values. Since during both world wars observations are lacking, we distinguish four different periods: 1885-1899, 1900-14 (or 13), 1920-38 and 1946-60, denoted as a, b, c and d; the number of pairs varies between 14 and 19. The resultant (marine) surface winds show a weak tendency to rotate clockwise from period a to period b and c; an even weaker tendency of an anticlockwise rotation can be observed towards period d. However, the standard deviation of the u- and v-components indicate that this rotation remains below the significant level. The resultant geostrophic wind
fails to coincide with these changes; here the permanent smoothing of the pressure field during the analysis renders it less sensitive against minor changes. The scalar speed of the (marine) wind tends to show highest values during period a, and decreases during periods b and c; during period d a weak increase is observed. But these variations could also be an artifact, caused by the different calibrations of the Beaufort scale (RAMAGE 1984).

Correlating the u- and v-component of both data sets, the correlation coefficient tends to increase towards period d. This could be interpreted as a gradual improvement of the reliability of the pressure analyses; the greatest deviations of individual seasons and years occur before 1900. 53 percent of these ccf's are 0.80 or higher; only in the field 62 (trades) statistically insignificant ccf's occur (35 percent). The surface resultant wind is distinctly weaker than the geostrophic resultant, in field 62 82 percent of the latter. The constancy (steadiness) of the winds increases distinctly around 1900; at field 62 it is fairly constant and high during summer, but quite variable during the remaining seasons (Fig. 4).

4) Variations of the position of the North Atlantic "centers of action"

Sea-level pressure data at 5° Lat x 10° Long. grid points, available at the North Atlantic between Lat. 20°N and 70°N since 1881, serve as base for an investigation of the variability of the latitude of the two main centers of action, the Icelandic Low (IL) and the Azores High (AH). Zonal averages are used for the section Long. 10°W-70°W; position and central pressure of both centers have been derived for each individual month using polynomials (for details see R. GLOWIENKA 1984) as well as the meridional pressure gradient. The ccf's between pressure and latitude of IL (AH) are negative (positive) and statistically significant for most months. Those between the position of both centers are positive (between 0.46 and 0.70), between the pressure negative, as found also by WALLACE and GUTZLER (1981).

Of high interest are the (linear) trends of the positions (Fig. 5). During summer both centers are shifted northward; the displacement rates for IL (AH) July are 0.046° (0.034°) Lat. per year. In contrast to this both centers are shifted southward during winter; the January displacement rates are resp. -0.044° (-0.022°) Lat. per year. A total displacement around 400 kms during 103 years is certainly not negligible. While the shifts are significant at the 99 percent level, the associated pressure changes are insignificant, except the IL pressure during December, with a trend of +0.053 mb/year.

The meridional pressure gradient varies strongly, especially during winter: in January the extreme 5-year-averages are 0.35 and 0.90 mb/Lat.

5) Parallel variations of H₂O and CO₂?

Recent studies in upwelling areas (HENNING-FLOHN 1980, WEBER-FLOHN 1984) have shown that saturation deficit and evaporation drop sharply with decreasing SST. Surprisingly enough, this has a global effect on tropospheric temperatures: PAN and OORT (1983) found a significant ccf up to +0.65 between SST at the equatorial Pacific and the temperature of the entire mass of the Northern Hemisphere atmosphere 3-7 months later. At least during northern winter, the H₂O content of most of the global atmosphere, especially in the tropics, was significantly higher in cases with "warm" SST at the key region than in the "cold" cases.
Regarding CO$_2$, several investigators (NEWELL et al. 1978, ANGELL 1981) found significant positive correlations between the growth rate of CO$_2$ after removal of the seasonal fluctuations, and the SST anomalies in the equatorial Pacific. BACASTOW and KEELING (1981) demonstrated a negative correlation between the CO$_2$ growth rate and the Southern Oscillation Index, which itself is highly negatively correlated with SST in the eastern equatorial Pacific. In five El Niño years (with warm water at the key region: 1958, 1963, 1965, 1969, 1972) the annual increase of CO$_2$ was 1.04 ppm, while in five "cold" years (1960, 1964, 1967, 1971, 1974) it dropped to 0.57 ppm. Selecting periods of 6 months length with "warm" or "cold" SST at Christmas Island (1.5°N, 157°W), WEBER and FLOHN (1984) found in 12 "warm" periods a CO$_2$ increase of 1.06 ppm/6 months, in 10 "cold" periods an increase of only 0.38 ppm; the difference is significant at the 95 percent level. Similar differences had been found when selecting "dry" cases (with suppressed rainfall during upwelling) and "wet" cases (with high rainfall during El Niño); the correlation between SST anomaly and rainfall index is 0.70. These interannual differences of CO$_2$ exchange through the air-sea interface reach ~1.4 ppm or ~3 Gt (= 10$^{12}$ kg) Carbon per year, i.e. they are of the same order as the annual storage of fossil CO$_2$ in the air (2.6 Gt).

While the H$_2$O exchange through the air-sea interface occurs with a phase change, following physical laws, the CO$_2$ exchange is also a biological process (NEWELL et al. 1978, BAES 1982): in the upwelling nutrient-rich water primitive algae consume CO$_2$ from the air and from the CO$_2$-super-saturated water, while the nearly sterile warm water of the tropical mixing layer is deficient of nutrients and thus releases CO$_2$ into the atmosphere. Recent investigations of the bioproductivity of the upwelling oceans indicate an annual CO$_2$ production by photosynthesis in the right order of magnitude (WEBER-FLOHN 1984). With a global model of the oceanic primary productivity VIECELLI (1984) estimates the increase of atmospheric CO$_2$ content with a 1 percent productivity loss to 0.5-2.5 percent (depending on turbulent mixing), which is also of the right order of magnitude (Gt/year). CHAVEZ et al. (1984) estimate the primary productivity under cool conditions to 2 Gt/a, under El Niño conditions to about one third.

These investigations indicate a parallel variation of the water vapour and carbon dioxide contents of the atmosphere. Both gases contribute to the greenhouse effect of the atmosphere – indeed H$_2$O can be more efficient than CO$_2$ (RAMANATHAN 1981), especially in the tropics. This parallel course enhances the CO$_2$ effect; however, it will be difficult to find convincing evidence for a gradual increase of the atmospheric H$_2$O content too. But this is likely if the global SST increase of about 0.6°C since the beginning of this century (FOLLAND et al. 1984) is correct. It would also be difficult to find convincing evidence for an increase of global precipitation. The problem of maintenance of the global water budget belongs to the most difficult, but also the most urgent questions: it is a real challenge for future climate research.
B) Abrupt Climatic Changes (scale 1-1000) Now and in the Past

1) Occurrence of climatic "jumps" in recent data series

Detailed studies of long climatic series have indicated that climatic changes - statistically described as a more or less linear trend - in some cases occur stepwise, with a short and rather rapid transition between two different modes, each of them including relatively minor fluctuations. Two quite dramatic examples have been observed in the runoff series of the Nile (1899) and of Lake Victoria (1962). The mean annual Nile runoff at Aswan fluctuated, during the period 1870-1898, around an average of 109.8 km$^3$/a (standard deviation 13.5 km$^3$/a), while during 1899-1928 the equivalent figures were 83.7 and 13.9 km$^3$/a respectively. This means a decrease of about 25 percent, twice the standard deviation; KRAUS (1955) has demonstrated some parallel trends, while RIEHL et al. (1979) and HASSAN (1981) discussed longer time-series. The Aswan record has been found to be homogeneous; thus this drop reflects a drastic change in the summer rainfall regime of Ethiopia, which contributes about 80 percent of the annual discharge. In recent decades the Nile runoff has been increasingly manipulated for irrigation purposes.

Even more striking is the abrupt change of the level of Lake Victoria, caused by the rainy season 1961/2 which brought 200-500 percent of the normal value in large areas of eastern Africa. Lake Victoria rose about 2.5 m and remained high: the runoff of the White Nile rose from 20.8 km$^3$/a for the period 1950-61 to 41.2 km$^3$/a for 1962-78, with standard deviations of 4.6 and 4.7 km$^3$/a respectively (FLOHN and BURKHARDT 1984). Possible physical interpretations have been discussed elsewhere (FLOHN 1983a).

Similar infrequent cases exist. We may mention the rainfall over the Sudan-Sahel belt of western Africa, where drought conditions last nearly uninterrupted since 1968. Using normalized anomalies from 20 stations (Lat. 11-18°N, west of 9°S, ca. 2 x 10°km$^2$), LAMB (1984) indicates, during the wet period 1950-67, only one year with weak negative anomalies, during the dry period 1968-83 only one year with weak positive anomalies. Multi-year droughts are well-known in this area, but the present duration (including 1984) seems to be exceptionally long.

Another example has been given by an evaluation of the temperature of the 500/1000 mb layer above the central Arctic, averaged over the whole area north of Lat. 75°N (about 8.8 x 10°km$^2$). These data (SCHAIRDEL-FLOHN 1983) indicate a rather abrupt cooling between 1961 and 1963, mainly during the summer months. The average deviation of the temperature of the lower troposphere is +0.48°C (1949-61) and -0.42°C (1962-76); the standard deviations of the two periods are 0.43°C and 0.35°C. Only two years of each period indicate a weak opposite deviation. Surface temperatures show a similar drop, while the sea-ice area apparently has increased by about 0.5 x 10° km$^2$; its correlation with tropospheric temperature is -0.54, significant at the 99 percent level.

These "jumps" have in common that the difference of the averages of consecutive periods (in the order of 15-30 years) reaches or exceeds twice the standard deviation of the individual periods. They occur at a regional (not local) scale and represent areas of several million km$^2$ (Lake Victoria only about 10° km$^2$). At present, speculations about their origin are not meaningful: spatial coherence and cohesiveness in time are hitherto unknown.
2) Occurrence of abrupt changes in the geological past

Recent investigations at continental sites have revealed that in a number of cases during the Pleistocene a change of annual or summer temperatures of 3 K or more evolved in a time-span of about 100 years. These abrupt changes are much shorter than expected as a consequence of orbital variations (10^4-10^5 years); they comprehend half or more of the difference glacial/interglacial. Table 3 gives a list of such events; for comments and references see FLOHN 1984.

Even more challenging cases have been found by physical investigations of a recent ice core (Dye 3 in southern Greenland, Lat. 65°N). DANSGAARD et al. (1982) have evaluated the δ¹⁸O record which represents very nearly surface temperature, while OESCHGER et al. (1983a) and OESCHGER and STAUFFER (1983b) have measured the CO₂ content entrapped in air bubbles with an improved technique. Here we refer to OESCHGER-STAUFFER's investigations comparing both records at a depth of 1860-1900 m, equivalent to an age of 27-40 ka (10^3 years) ago. In a 30 m section (equivalent to about 10 ka, Fig. 6) they found non-periodic simultaneous fluctuations of both δ¹⁸O and of CO₂ (between about 260 ppm and 190 ppm). The accuracy of the measurements has been discussed by BARNOLA et al. (1983), the age difference between the bubble air and ice by SCHWÄNDER and STAUFFER (1984). Most convincing is the high-resolution investigation of a 3 m section (OESCHGER-STAUFFER Fig. 8) equivalent to about 1 ka: here the transition from a "cold" level with an average CO₂ content of 190 ppm to a "warm" level (256 ppm) took not more than 100 years (32 cm). Fig. 7 gives a time-independent diagram of CO₂ data versus simultaneous δ¹⁸O values (ccf = 0.95). A linear regression equation (WEBER-FLOHN 1984) reads

$$\text{CO}_2 \text{ (ppm)} = 617 + 11.7 \delta^{18}O \text{ (per mil)}$$

The same procedure for the much longer time series of Fig. 6 gives

$$\text{CO}_2 \text{ (ppm)} = 537 + 9.47 \delta^{18}O \text{ (per mil)}$$

with a ccf of 0.82 (due to the lower resolution and the less accurate fixation of simultaneous points). The δ¹⁸O difference of the two different climatic periods in Fig. 7 (5.2 per mil) is equivalent - using DANSGAARD's (1964) data - to a local temperature difference of 8 K. With other techniques similar results have been reached for the well-documented cases a), b) and d) of Table 3.

Since at present the CO₂ fluctuations affected mostly the northern hemisphere, spreading towards the South Pole station in 1-2 years, such data stored nearly 2 km deep in the Greenland ice are indeed indicative for the global climate. The most intriguing facts are the occurrence of "abrupt" climatic changes in the "human" 1000-scale and the simultaneous participation of CO₂ fluctuations: both facts now force hesitating climatologists to revise some of their fundamental textbook ideas.

3) The "greenhouse effect" as a key for an abrupt climatic change?

What processes could have been responsible, under purely natural conditions, for CO₂ fluctuations of a comparable rate (60-80 ppm per 100 years) as those now produced by man? Based mainly on our present understanding of seasonal variations of the atmospheric/oceanic circulation, a working hypothesis has been proposed (FLOHN 1981, 1982, 1983c; WEBER-FLOHN 1984). It starts from the observed fact that the variable inter-
annual changes of the atmospheric CO$_2$ content are correlated with the sudden reversals of oceanic upwelling/downwelling (El Niño); see Chapter A 5. The coincidence between glacial and global dryness and between interglacials and global wetness (SARNTHEIN 1978) suggests that the (under present conditions) observed parallel course of CO$_2$ and H$_2$O may have been a quite general feature of climatic history at scales between 1 and 10$^3$ years. The interglacial/glacial difference between global annual precipitation/evaporation has been (conservatively) estimated to 20-25 percent of the present value (FLOHN 1983c).

These fluctuations during the last ice age, in earlier interglacials or glacial/interglacial transitions (Table 3) indicate that in rare cases (ca. 0.1-0.5 per ka) abrupt climate fluctuations as described above occur under purely natural conditions.

In Chapter a it has been shown that the sudden transition between upwelling and downwelling - occurring in a few weeks or months - are responsible for large-scale changes in the H$_2$O and CO$_2$ fluxes between sea and air. Both gases are mainly responsible for the greenhouse effect; this effect contributes also to the difference between glacial and interglacial (ÖSECHGER et al. 1983a) and to the stadial-interstadial fluctuations with a (non-periodic!) time-scale in the order of 2-3 ka. Fluctuations of this scale are also observed during the Holocene as glacier variations in the Alps (GAMPER and SUTER 1982), Scandinavia (KARLEN 1976, 1979) and other high mountains. They are also confirmed at deep-ocean cores; SCHNITKER (1982) has proposed a climatic feedback mechanism of this time-scale based on the slow deep-sea currents transporting bottom water from its source in the subarctic Atlantic to the other oceans.

Our proposed feedback (FLOHN 1983c) - see also the more general view of SIEGENTHALER and WENK (1984) - is based on the observed seasonal and/or interannual fluctuations of the El Niño-type; these experiences are only expanded in time. We start from a hypothetical cooling (or warming) event caused e.g. by a cluster or a lull of heavy volcanic eruptions; their thermal effect shall be concentrated - as a consequence of the geometry of solar rays at high latitudes and of a longer residence time of aerosol in the polar stratosphere - in polar regions. In equatorial regions temperature changes are distinctly smaller and (due to the extension of oceans) slower. This differential cooling (or heating) corresponds with an expansion (or contraction) of the circumpolar vortex (Ferrel circulation), in principle similar to the seasonal variations of each year. This is also equivalent to the equatorward (poleward) shift of the subtropical anticyclonic belt with a strengthening (weakening) of the tropical easterlies. This causes increasing upwelling, correlated (under natural conditions) with suppressed evaporation, precipitation and CO$_2$ release into the atmosphere and further radiative cooling, or in the case of warming, equatorial downwelling (Niño) with increased evaporation and CO$_2$ release, increased precipitation and tropospheric warming due to radiative processes as well as release of latent heat (PAN and OORT 1983).

Conclusions

The ideas expressed in Chapter B on the occurrence of abrupt climatic changes and their correlation with the air-sea exchange processes of greenhouse gases are based on a transfer of recent experiences at the interannual time-scale to a larger scale of 10$^2$ (and finally 10$^3$-10$^4$) years. Similarly the observed interannual climatic "jumps", at a regional scale, seem to be an analogue to these "abrupt" changes which most probably are
of a global scale. Thus we may emphasize two points: Understanding of climatic changes at the time-scale near 100 years is not possible without due account of the key role of the oceans and of the time variability of the composition of the earth's atmosphere. And: the CO2 climate problem is not only a recent environmental issue; it is of a quite general nature within the climatic evolution of our planet. This perception greatly underlines the need for further research. This is especially necessary at the field of paleoclimatology: only here is a possibility to verify model results which otherwise remain in a vacuum. Perhaps more essential are the fundamental questions, which nature puts at the disposal of far-sighted model-designers: we need their answers.

During the last years, several authors (e.g. BUDYKO 1982) suggested a key role of CO2 variations for the evolution of the global climate since primordial times. This follows early ideas pointed out by Sv. ARRHENIUS (1896) and Th. CHAMBERLIN (1897). Indeed the "acryogenic" climate of the Mesozoic and Early Cenozoic - both poles ice-free - can probably not be understood without the assumption of a higher atmospheric CO2 level. It may be possible that the gradual transition to a unipolar glaciated earth during the Late Cenozoic (existing between 14 and about 3 million years ago) and finally into the formation of a more or less permanent Arctic sea-ice cover, are both correlated with changes in the atmosphere's composition. But here it is impossible to enter into the geophysical and geochemical background of climatic evolution. Climatologists are now beginning to widen their time horizon from 300 years of instrumental observations into the geological past. In this view the anthropogenic growth of the CO2 level is not new - we have to look seriously into the very long history of the evolution of our climate and to learn its lesson.

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H. FLOHN (1979): Quaternary Research 12, 135-149


H. FLOHN, Th. BURKHARDT (1984): in press


M. GAMPER, J. SUTER (1982): Geographica Helvetica 37, 105-114


R. RIEHL et al. (1979): Monthly Weather Review 107, 1546-1553


Table 1: Interannual Variability (IAV) of evaporation (LE) and sensible heat flux (H), Watt/m²

<table>
<thead>
<tr>
<th>Weather Ship C</th>
<th>LE</th>
<th>Max.</th>
<th>Min.</th>
<th>IAV</th>
</tr>
</thead>
<tbody>
<tr>
<td>(53°N, 35°W, 22 years)</td>
<td>151.0</td>
<td>192.3</td>
<td>118.1</td>
<td>23.1</td>
</tr>
<tr>
<td>October-March</td>
<td>H</td>
<td>22.3</td>
<td>38.8</td>
<td>5.1</td>
</tr>
<tr>
<td>LE+H</td>
<td>173.3</td>
<td>213.1</td>
<td>134.2</td>
<td>23.0</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Field 0-4°S, 28-34°W</th>
<th>LE</th>
<th>Max.</th>
<th>Min.</th>
<th>IAV</th>
</tr>
</thead>
<tbody>
<tr>
<td>(13 years)</td>
<td>93.4</td>
<td>114.0</td>
<td>81.0</td>
<td>8.8</td>
</tr>
<tr>
<td>June-October</td>
<td>H</td>
<td>3.1</td>
<td>6.0</td>
<td>1.0</td>
</tr>
<tr>
<td>LE+H</td>
<td>96.5</td>
<td>117.0</td>
<td>82.0</td>
<td>9.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Light Ship Elbe 1</th>
<th>LE</th>
<th>Max.</th>
<th>Min.</th>
<th>IAV</th>
</tr>
</thead>
<tbody>
<tr>
<td>(54°N, 8°E, 25 years)</td>
<td>75.7</td>
<td>92.7</td>
<td>65.1</td>
<td>7.7</td>
</tr>
<tr>
<td>September-December</td>
<td>H</td>
<td>19.0</td>
<td>26.0</td>
<td>11.0</td>
</tr>
<tr>
<td>LE+H</td>
<td>94.6</td>
<td>117.5</td>
<td>76.5</td>
<td>11.4</td>
</tr>
</tbody>
</table>
Table 2: Comparative annual results of maritime surface winds (m) and geostrophic winds (g) at the northern Atlantic Ocean. Average positions given in degrees latitude and longitude; periods and sea state. \( q \) = constancy \((v_{\text{scal}}/v_{\text{result}})\) in percent; \( dd \) = direction of resultant winds \((v)\), velocity in m/s.

<table>
<thead>
<tr>
<th>Period and central position</th>
<th>Marine observations ( v_{\text{scal}} )</th>
<th>( q )</th>
<th>( dd_{m} )</th>
<th>geostrophic data ( dd_{g} )</th>
<th>( v_{\text{result}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field 34 a</td>
<td>9.48 34</td>
<td>241.6°</td>
<td>255.2</td>
<td>5.51</td>
<td></td>
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<td>m: 48.3°N, 26.1°W b</td>
<td>9.04 43</td>
<td>257.8</td>
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<td>g: 47.5°N, 25.0°W c</td>
<td>9.18 40</td>
<td>259.2</td>
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<td>d</td>
<td>9.57 41</td>
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<td>Field 44 a</td>
<td>8.69 26</td>
<td>257.5</td>
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<td>m: 42.5°N, 24.7°W b</td>
<td>8.17 28</td>
<td>268.2</td>
<td>261.9</td>
<td>5.05</td>
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<tr>
<td>g: 42.5°N, 25.0°W c</td>
<td>7.81 31</td>
<td>255.6</td>
<td>262.6</td>
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<tr>
<td>d</td>
<td>8.01 33</td>
<td>263.6</td>
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<td>Field 62 a</td>
<td>6.67 48</td>
<td>31.5</td>
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<tr>
<td>m: 29.5°N, 19.5°W b</td>
<td>5.74 55</td>
<td>37.1</td>
<td>40.0</td>
<td>4.67</td>
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<td>g: 30.0°N, 20.0°W c</td>
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<td>37.2</td>
<td>44.1</td>
<td>4.30</td>
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<tr>
<td>d</td>
<td>6.26 58</td>
<td>31.3</td>
<td>42.4</td>
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Table 3: Abrupt Paleoclimatic Events
(Time Scale: 100k)

<table>
<thead>
<tr>
<th>( \Delta T )</th>
<th>Time BP</th>
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<tr>
<td>W Transition Y. Oryas - Preboreal</td>
<td>-10.2 ka</td>
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<tr>
<td>C Alleröd - Y. Oryas</td>
<td>10.8</td>
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<tr>
<td>? Onset Moist Period (Lat. 15 - 35°N)</td>
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<tr>
<td>W Transition O. Oryas - Belling</td>
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<tr>
<td>C Stages 5c/4</td>
<td>73</td>
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<tr>
<td>C Stages 5c/5b</td>
<td>95</td>
</tr>
<tr>
<td>C Stages 5e/5d</td>
<td>115</td>
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<td>C Cold Episode Holstein Interglacial</td>
<td>150.7</td>
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<tr>
<td>(Stage 7, 8, 9)</td>
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<tr>
<td>C Stage 19</td>
<td>700</td>
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<tr>
<td>(Matuyama - Brunhes)</td>
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</table>
Fig. 1 SST and MAT in the field 5-10°N, 70-80°E (W of Sri Lanka). Above: average difference SST-MAT; below: diurnal variation of SST and MAT.

Fig. 2 Time series of LE, wind speed u, SST, SST-MAT and q_e-q in off the Somali coast (10-12°N, 51-55°E), summer 1959-1976 (WEBER). Note the anomalies of the Niño year 1972.
Fig. 3 Evaporation in mm/d at Lightship Elbe 1 in the German Bight (54°N, 6°E) as a function of season and wind direction (270° = W);
1 mm/d equivalent to LE = 28.5 W/m².

Fig. 4 Steadiness of winds $v_{RES}/v_{scal}$ in percent at extreme seasons (summer = JJA, winter = DJF) off the NW coast of Africa (Lat. 27-32°N, Long. 17-25°W). Data source: Historical Sea Surface Temperature Data (HSSTD) (K.-H. WEBER 1984).
Fig. 5 Position of Icelandic Low (left) and Azores High (right), individual years (dashes) and 5-year averages (solid lines) for the period 1881-1983, averaged over Long. 10-70°W, January (above) and July (below). Note, the opposite trend (R. GŁOJNIŃKA 1984).
Fig. 6 $\mathrm{CO}_2$ concentration and $\delta^{18} \mathrm{O}$ ratio at 1860-1890 m depth (approximately 29-38 ka BP), Dye 3, Lat. 65°N, Greenland (OESCHGER-STAUFFER 1983). Circles = individual $\mathrm{CO}_2$ measurements, distance between averaged values $\pm 300a$.

Fig. 7 $\mathrm{CO}_2$ content versus $\delta^{18} \mathrm{O}$ ratio at 1896-1899 m depth (about 40 ka BP) at the Dye 3 core, evaluated from the data given by OESCHGER-STAUFFER 1983, Fig. 8.