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Background of a Geophysical Model of the Initiation of the Next Glaciation

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Evidence of (at least) five rapid hemispheric coolings of about 5°C during the last 10^5 yr has been found, each event spread over not more than about a century, as examples of a global-scale climatic intransitivity. Only some of them lead to a complete glaciation at the northern continents, others ended after a few centuries by a sudden warming ("abortive glaciation"). Starting from a modified version of Wilson's hypothesis of Antarctic ice surges, an air-sea interaction model with realistic geophysical parameters is outlined to interpret the sudden initiation of the North American ice sheet. Special attention is given to the Atlantic section, where the climatic anomalies during the last glaciation appear to have been significantly larger than in other sections.

I. INTRODUCTION

In a recent symposium (Kukla *et al.*, 1972; Kukla and Matthews, 1972), several well-known Pleistocene specialists discussed the question: When will the present Interglacial end? A few results will be considered here from a meteorological viewpoint.

(a) Since about 1945 global cooling, on a scale of ~ 0.01°C/yr, has reversed the warming trend of the first decades of our century. The bulk of these changes is most probably not man-made, but of natural origin. Evidence exists for several short cool periods during the last 5000 yr, as well as for catastrophic dry periods in subtropical areas lasting a few decades. None of these variations are comparable in scale with the Allerød fluctuations (Chap. IV).

(b) The climatic optimum of the present interglacial was reached 6–7000 y.a. Evidence from Northern Germany and England shows that the last interglacial (Eem = Sangamon) lasted little more than 10,000 yr; it was slightly warmer and wetter than the present interglacial, with a quite similar climatic time sequence.

(c) Based on more than 800 measurements of the $^{18}O/^{16}O$ ratio from fossil foraminifera, it has been concluded (Emiliani, 1972) that the tropical ocean surface temperatures were as high as or higher than today for only 10% of the last 400,000 yr. Considering the length of a glacial–interglacial cycle to be nearly 10^5 yr (Broecker and van Donk, 1970) the average duration of a warm epoch cannot have been longer than 10^4 yr.

(d) A large majority of the participants of the symposium concluded that the present warm epoch has reached its final phase, and that—disregarding possible man-made effects—the natural end of this interglacial epoch is "undoubtedly near." The time-scale of this transition may be a few millenia, perhaps only centuries.

If this is correct, earth scientists are confronted with a hitherto neglected question: What are the initial stages of a glaciation? How can we imagine the triggering of the formation of ice sheets on the

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northern continents eventually covering about 17×10^6 km^2 in North America and nearly 11×10^6 km^2 in Eurasia, with ice domes up to 3 or 4 km in height, thus reducing the ocean volume by about 4%, with a eustatic sinking of the sea level to —100 m, sometimes to —130 m? This question is not only of academic interest: its immediacy will be demonstrated in the following chapters.

II. STABILITY OR INSTABILITY OF CLIMATE?

E. Lorenz (1968) has recently raised a quite deep-rooted question: How stable is our climate? Considering a complete set of basic equations with fixed external parameters—such as the solar "constant," the rotation rate and radius of the Earth, and the chemical composition of the atmosphere—as a base for simulating the climate defined as a time-averaged state of the atmosphere, he discusses the number of possible solutions. Climate is then defined as *transitive* if only one solution exists. One may also perceive several more or less quasistationary different solutions, which may transform from one state into another by a sort of flip-flop mechanism ("vacillation"): this situation is defined as *intransitive*. Without discussing at length evidence for and against, Lorenz considers our climate as *semiintransitive*.

If this is true, we cannot expect to obtain from mathematical modeling unambiguous forecasts of climatic patterns. Simplified models with a crude parameterization of synoptic-scale meridional exchange processes—such as the models developed by Budyko (1969) and Sellers (1969, 1973)—showed either a great sensitivity to comparatively small changes of external conditions or (worse than that) distinct intransitivity under exactly the same conditions. In contrast to this, Washington (1972) demonstrated on the base of the much more advanced NCAR circulation model that the response of the model to different and even contrasting externally induced disturbances was nearly identical. This state of affairs—incomplete as it stands now—is seriously disquieting. Therefore, one of the most urgent tasks is a careful and critical search for evidence of climatic instability on a hemispheric or, better, global scale.

Examples of partial (regional) instability have been given elsewhere (Flohn, 1973). The best example is known from the equatorial Pacific, where the oceanic Ekman drift causes either equatorial upwelling or downwelling, depending on the surface wind distribution, and causing in the atmosphere either a stable, cloudless and dry equatorial zone or instability near the equator with high convective activity. Apart from short transition periods, any intermediate state cannot remain stable. Because of the large attendant differences in oceanic evaporation and precipitation, these contrasting patterns are correlated with many teleconnections over wide areas of the globe (Bjerknes, 1969; Flohn, 1972; Rowntree, 1972).

III. THE ROLE OF SURFACE ALBEDO

Within the heat budget of the Earth's surface, the high albedo of ice and snow (0.70–0.80)—in contrast to all other surfaces, except clouds, (0.05–0.35)—dominates most other terms. During winter the tropospheric baroclinic zones have a tendency to follow the margins of the seasonal continental snow-cover; this has been experienced by the author during the European winters between 1938 and 1948. In spite of all vagaries of weather, such a pattern remains superimposed in a statistical sense.

There exists (Manabe and Wetherald, 1967) a direct relation between surface albedo and an equilibrium temperature (Table 1), assuming constant relative humidity and an average cloud distribution. This relation can be checked against data on the varying extent of the Arctic and

TABLE 1

SURFACE ALBEDO (a_s) AND EQUILIBRIUM TEMPERATURE (T^*) DEVIATIONS (AREAS IN 10^6 KM^2)

	Oceans		Continents					
Albedo	Open 0.05	Ice 0.70	Open 0.12	Ice 0.75	Snow 0.30	Average albedo a_s	Deviation ΔT^* (°K)	Remarks
N. Hemisphere	145	10	70	3	27	0.1294	—	Actual (1901–50)
S. Hemisphere	190	16	33	13	3	0.1384	—	Actual (1901–50)
Earth (E)	335	26	103	16	30	0.1339	—	Actual (1901–50)
Model NH 0	142	13	70	3	27	0.1373	−0.95	N. Hemis. ca. 1890
Model SH 0	188	18	33	13	3	0.1434	−0.60	S. Hemis. ca. 1850
Model E 4	317	30	108	49	6	0.1731	−4.6	Ice age (sea level −100 m)
Model E 5	307	40	108	49	6	0.1860	−6.2	Ice age (sea level −100 m)
Model SH 3	165	41	33	13	3	0.2022	−7.6	Wilson surge

Antarctic sea-ice during the 19th and 20th centuries (Flohn, 1973). The observed decrease of the Arctic sea-ice from about 1880 to 1940 and the estimated increase of the Antarctic sea-ice during the 19th century (Lamb, 1967), both of the order of nearly 2 or 3×10^6 km^2, should correlate with changes in the hemispheric equilibrium temperature of the order of 0.6–0.9°C, in good agreement with the observed data (Table 1, Models SH 0 and NH 0).

A further check can be derived from the last glaciation, with a glaciated continental area of 49×10^6 km^2, accompanied by a eustatic drop of the sea level to —100 m, increasing the land area of the Earth from 149 to about 163×10^6 km^2. In this case (Model E 4) the equilibrium temperature of the whole Earth should be 4–5°C lower than today, once more in agreement with the observed data. There is sufficient evidence that the Atlantic sea-ice reached, during the maximum of the last glaciation, an average latitude of about 43° N (McIntyre *et al.*, 1972). Its extension into the Bay of Biscay must also be assumed when interpreting the exceptional cooling of the adjacent territories (from northern Spain to southern Ireland) by about 12°C, compared with only 5°C at the same latitudes at the Pacific coast of North America (Flohn, 1969). In this case the area covered by sea-ice increases to about 40×10^6 km^2 with a simultaneous global temperature drop of 6°C (Model E 5).

The good agreement between the predicted and observed equilibrium temperatures convinces us that the role of the albedo in the long-term climatic oscillations during the Pleistocene is certainly greater than that of variations of solar radiation due to the Earth's orbital elements (Hoinkes, 1971). It should be remembered that this point was raised as early as 1938 (Wundt, 1938); numerical model computations of the climatic effects of the Milankovich mechanism neglecting the positive feedback effect of albedo changes are incomplete. The role of surface albedo has also been demonstrated (Kukla and Kukla, 1972) from seasonal and interannual changes of the snow-cover. The existence of a quasiequilibrium between area-averaged surface temperature and area-averaged surface albedo (Flohn, 1969) leads to a serious consequence: if a climate-independent mechanism producing variations of the extension of ice exists—as suggested by A. T. Wilson (1964)—the usual chain of cause and effect may be reversed: large-scale Antarctic surges will produce immediate hemispheric cooling (Model SH 3). This necessitates a critical investigation of the real time-scale of Pleistocene coolings, which appears to be inconsistent with the Milankovich time-scale.

IV. TIME-SCALE OF GLOBAL COOLINGS

Based on investigations of ice cores, ocean bottom cores, and fossil peat bogs, several drastic coolings during the last 10^5 yr have recently been revealed. Some of these, with multiple evidence, will be reviewed in stratigraphic sequence, ignoring some inevitable minor differences of time and time sequence interpretation in the literature.

(1) During the recession of the last glaciation, the well-known sequence Bølling interstadial (warm)–Older Dryas (cold)–Allerød interstadial (warm)–Younger Dryas (cold) covered less than 2000 yr, with variation in the annual temperature of up to 6°C (Mercer, 1969). In the Mediterranean and at other subtropical and tropical sites only the second half of the sequence was marked, and the Older Dryas period was insignificant (van der Hammen *et al.*, 1971, Fig. 2). The Allerød warming period coincides with the abrupt global environmental change after the Würm–Wisconsin Glaciation, occurring at about 11,000 BP in the space of a few centuries, while the melting of the ice domes—reflected in the global eustatic sea level rise—lasted some 8000 yr.

This time sequence has been derived mainly from palynological evidence, hampered by the limited migration speed of biotopes. On the other hand, the isotopic changes preserved in the Greenland ice cap represent largely—disregarding here some systematic sources of error (Johnson *et al.*, 1972; Dansgaard *et al.*, 1971)—the temperature of formation of precipitation particles in clouds, i.e., the regional climate. Here (Johnson *et al.*, 1972, Fig. 6) the cooling prior to the younger Dryas lasted less than 350 yr; the following warming, 300 yr. However, simultaneity of the climatic changes on both sides of the Atlantic has been doubted (Mercer, 1969).

(2) Before the last long warm interstadial within the Würm–Wisconsin Glaciation (Fliri, 1970)—known as Stillfried B or Plum Point—a marked cold period of not more than about 2000 yr duration occurred at about 38,000 BP. It has been found in the Greenland ice core (Johnson *et al.*, 1972) as well as in the Indian Ocean off the Somali Coast (4–8° N) (Olaussen *et al.*, 1971); here the time between the beginning of the event and the temperature minimum is estimated to be not more than about 500 yr.

(3) Another cold period of this magnitude is also found in the Somali Current Area, interpreted as the beginning of the Würm I Glaciation at about 55,000 BP. It coincides well with the marked cooling in the Greenland core after the Odderade Interstadial and in Macedonia (van der Hammen, 1971) around 59,000 BP.

(4) The short cooling between the Brørup and Odderade Interstadials, near 70,000 BP, is quite dramatic: In Macedonia the vegetation changed from oak forest into steppe in much less than 1000 yr. It coincides with a marked cold period in the Greenland area, while in the Indian Ocean (Olaussen *et al.*, 1971) only a hint has been found. At the same time, a short intense cooling has been observed (Sancetta *et al.*, 1972) in an Atlantic deep-sea core at 52° N 22° W.

(5) The most dramatic short-lived cooling event was observed (Fig. 1) in the Greenland ice at about 89,000 BP (all Greenland dates before 12,000 BP are slightly uncertain). Here the climate changed within 100 yr ("almost instantaneously") from warmer than today into full glacial severity (Dansgaard *et al.*, 1972). This event has also been found in a stalagmite in a French cave (Duplessy *et al.*, 1971) at 97,000 BP with a cooling of the cave (!) by 3°C in a few centuries and an extremely rapid cooling (in less than 350 yr) has been described in many cores from the Gulf of Mexico at 90,000 BP (Kennett and Huddleston, 1972.) At the same time the first strong cooling after the Eem Interglacial was observed in Macedonia and in the Netherlands (van der Hammen, 1971);

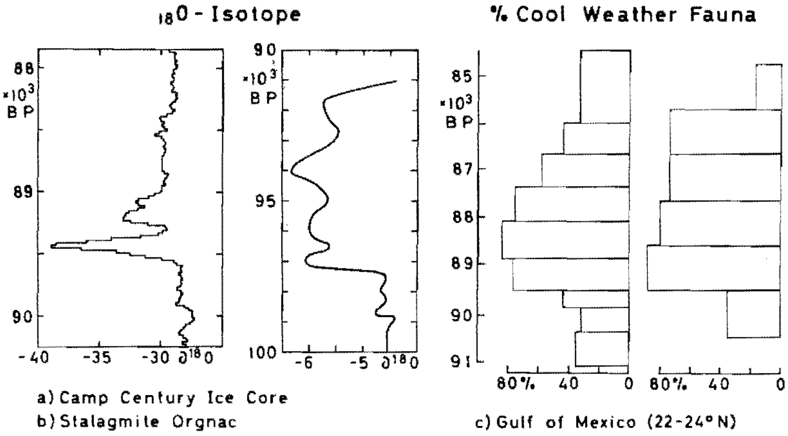

FIG. 1. Temperature variations at about 90,000 BP (see text). (a) $^{18}O/^{16}O$ ratio in Greenland ice core (76° N). (b) $^{18}O/^{16}O$ in a cave near Orgnac (S. France). (c) Percentage of cool weather foraminifera from the Gulf of Mexico.

and multiple evidence exists for a sudden sea level rise at the eastern coast of North America and at Bermuda, possibly caused by an Antarctic surge (Hollin, 1972).

These five events show coolings of the order of up to 5°C/century in contrast to not more than 1°C/century in recent fluctuations. This rate is in fact a minimum value because of the smoothing role of molecular diffusion processes (Johnson *et al.*, 1972). Of particular interest are the events in the area of the Somali Current (Olaussen *et al.*, 1971) which are far too short-lived to be interpreted as caused by orbital changes. Some other peaks, especially in the Greenland ice core (Fig. 2), may be added to this list, but (up to now) without supporting evidence from other sites.

Only one of these selected cases initiated, in the northern hemisphere, a continental glaciation: this is the beginning of Würm I (case 3). If we assume a maximum glaciation of the northern continents (with a volume of 47.4×10^6 km^3 and an area of 29.5×10^6 km^2, i.e., with an average thickness of 1610 m) as resulting from an average annual accumulation of 40 cm, the growth period lasts about 4000 yr, and the minimum duration of a full glaciation is still of the order of 10,000 yr (Lamb and Woodroffe, 1970). During that time the local increase of albedo favors the persistence of glaciogenic conditions. The other four cases represent only short-lived events, with a glaciogenic anomaly of the atmospheric and/or oceanic circulation lasting "only" a few centuries (case 5), certainly less than 2–3000 yr. From the viewpoint of a meteorologist, these incomplete or "abortive glaciations" are by no means less interesting: they reveal a very remarkable instability of the atmosphere–ocean–ice system repeating nonperiodically over a time-scale of the order of 2×10^4 yr.

In view of the rapidity of development the initial stages must have lasted less than a century, probably only a few decades. What kind of atmosphere/ocean circulation anomalies are able to produce such catastrophic events? Any answer to this question can only be more or less speculative; however, it should never be unsound from the viewpoint of an experienced meteorologist specializing in physical and theoretical climatology.

V. CLIMATIC CONDITIONS OF THE INITIAL STAGES OF A GLACIATION

The evidence of such dramatic global coolings with a quite short time-scale is of high importance when discussing the initi-

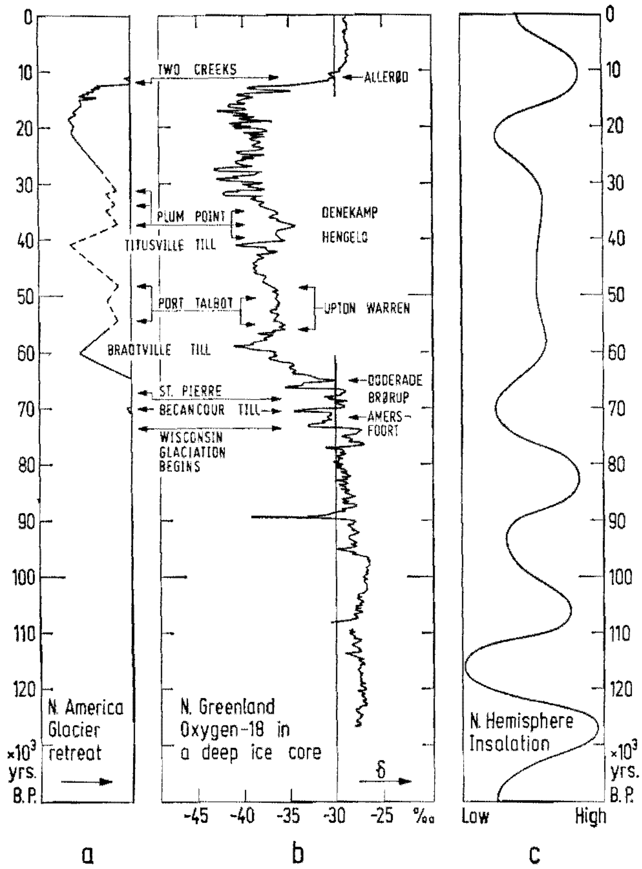

FIG. 2. Greenland ice core vs corrected time scale (Dansgaard *et al.*, 1971), plotted in 200-yr intervals, with tentative interpretation in European (right) and American (left) terminology.

ation of the glaciations of the northern continents, i.e., of the Laurentide and Scandinavian ice sheets. This initiation is in any case a problem: while in Scandinavia a spreading of the existing mountain glaciers (Svartisen, Jötunheimen) could be caused by a temperature drop of 5–6°C, without any necessity for a substantial increase of precipitation, this is not the case in North America. Here the Rocky Mountain Ice, which expanded from the (at present heavily glaciated) high mountains in western Canada and southern Alaska, remained of medium size and never did extend far to the east. The major glaciation of North America was due to the Laurentide ice sheet, which formed on the presently unglaciated Ungava Plateau of Labrador and Quebec (now at alt of 600–800 m, between 53° and 60° N lat) and even on the low-lying Keewatin country west of Hudson Bay, between 58° and 65° N. Only during the maximum phase of the last glaciation were the two ice sheets joined. Loewe (1971) has discussed the climatic conditions in Ungava and Keewatin, and concluded that a 6°C temp drop alone would be insufficient to cause a permanent snow-cover without an increase in total precipitation (snowfall). Because of the large extent of the Laurentian ice dome (which contained more than 62% of the total increase in ice volume during a glaciation), its formation must have a key position in the sequence of events.

The recent climate in Labrador–Ungava

and Keewatin is characterized by summer temperatures (June–August) of 11–12°C and by annual precipitation near 75–80 cm in Ungava but only about 35 cm in Keewatin.

Barry *et al.* (1959) and Brinkmann and Barry (1972) have investigated, by the methods of synoptic climatology, the meteorological conditions associated with high precipitation in the Labrador–Ungava area as well as in the Keewatin area, with different results. The formation of a nucleus of the ice dome on the Labrador highlands, with a temperature drop of about 6°C together with an increase of precipitation, is only possible for a semipermanent upper cold cyclone centered near 55° N 72° W or a deep upper trough extending from Baffin Bay or Ellesmere Island into the area near Boston. In this situation low-level flow from N or NW would permanently cool the area, between about long. 85° and 70° W, on the southern flank of low-level cyclones which themselves extended even farther east. Above 700 mbar, however, relatively warm and moist air from the south would flow northward above the Labrador Peninsula, forced to ascend along a more or less stationary frontal surface and causing abundant precipitation, most frequently as snow.

In a boreal subpolar climate, the snow-melting process is not finished before late May or early June, and the first snowfall may occur as early as the end of August or early September. Each glaciation must start with a permanent snow-cover lasting during the summer; its high albedo (even in a half-melted stage) prevents the soil from storing heat. If a snow-cover can survive one summer with its high sunshine duration, the probability of a much higher snow-cover in the next year rises substantially; this is the beginning of a positive feedback process. A few consecutive years of this type would be sufficient to build up a snow-cover of several meters over the whole Ungava plateau with an area of about 60,000 km² above 600 m: then the high surface albedo during summer will prevent easy destruction, even if the large-scale flow-type changes.

Let us assume a 20% increase of precipitation to 90 cm/yr, a snowfall fraction of 80% of precipitation and a (high) snow-cover density of 0.3, we obtain an annual snow accumulation of 240 cm (or 72 g/cm²). In a cloudy summer, the incoming global radiation should be slightly reduced to about 350 Ly/d (1 Langley = 1 gcal cm^{-2}); with a surface albedo of 55% (50) and an effective terrestrial radiation of 120 Ly/d (latitudinal average) we obtain a net radiation of 38 (55) Ly/d. During a melting season of 80 days, this would melt 38 (55) g/cm² snow; the remaining snow-cover would reach, with a density of 0.5, a height of 68 (34) cm at the beginning of next winter. With an average (!) albedo of 45% or less, snow-cover would completely melt. Under such assumptions, a 15 m tall forest could be completely covered by snow after some 22 yr (45), with a further rise of surface albedo. Cold air will then be permanently produced near the surface, due to the combined effects of high albedo and long-wave radiation from the snow-cover. This will lead, once a synoptic-scale diameter (300–500 km) is reached, to the formation of a superimposed cold low, which will force the upper flow to curve cyclonically and will further enhance snowfall. Such a powerful positive feedback mechanism is well known to the experienced meterologist; it was discussed half a century ago by C. E. P. Brooks (1926). It can be qualitatively interpreted with the aid of the heat balance equation for an atmospheric column:

$$H + LP - \operatorname{div} Q\!\downarrow - \operatorname{div} \vec{A}_h - \Delta T = 0,$$

where (H = flux of sensible heat into air, LP = release of latent heat by precipitation, $\operatorname{div} Q\!\downarrow$ = divergence of radiative fluxes, $\operatorname{div} \vec{A}_h$ = divergence of advective heat transport, and ΔT = heat storage in the column). In high latitudes LP is not as pre-

dominant as in the tropics; above a snowcover H is usually negative (from air to surface) and div $Q\downarrow$ is strongly negative, thus ΔT is likely to represent a heat sink.

The geophysical causes of this initial anomaly of the atmospheric circulation will be discussed in Section VI. Here it may be useful to outline the large-scale pattern connected with a deep semipermanent trough along 70° W, with a cold cyclonic center above western Labrador. This causes a meridionalization of the upper tropospheric flow, with anticyclonic ridges (and frequent blocking highs) near 125° W (Canadian NW-Territories) and 20° W (Iceland) (Flohn, 1969; Lamb and Woodroffe, 1970). A secondary trough over Scandinavia and Central Europe will develop near 15° E, together with a warm ridge in 50–70° E, including the mountains of Central Asia. The occurrence of a blocking high just east of Alaska leads to southerly flow over Alaska itself, locally reducing the rate of cooling. The frequent occurrence of a blocking high between Iceland and Scotland causes northerly flow over Scandinavia and Central Europe, increasing snowfall and cooling. This in turn causes a quasistationary pattern above Eastern Europe (25–40° E) corresponding to that above Labrador, with similar consequences, starting on the eastern flank of the Scandinavian mountains and in Finland. Such a pattern is nowadays frequent in cold winters and springs; here it is visualized—quite differently from today—as existing during the climax of the warm season.

VI. ANTARCTIC SURGES AND THEIR GEOPHYSICAL CONSEQUENCES

Since the observed short time-scale of global cooling events (cf. Chap. IV) is inconsistent with orbital effects, we ought to consider quite seriously the unorthodox Antarctic Surge hypothesis of A. T. Wilson (1964, 1966, 1969), based on the idea of a large-scale instability of the Antarctic ice dome. Since many recent examples of mountain glacier surges are known, particularly in the Alaskan mountains, and since some physical properties of glacier ice are subject to marked changes in the vicinity of the melting point, this hypothesis appears to be generally consistent with glaciological knowledge. Budd *et al.* (1970) have developed a geophysical model of the Antarctic ice, mapping such quantities as ice cap streamlines, balance flow velocities, strain and basal heating rates, temperatures and melt rates. One of the prerequisites of a surge is basal melting, which has been found at Byrd Station, 80° S 120° W at a depth of 2164 m (Gow *et al.*, 1968). However, according to this model less than 10% of the recent Antarctic ice cap is now subject to melting processes near the ground. According to Oswald and Robin (1973) 17 subice lakes have been discovered in East Antarctica by radioecho sounding flights, in regions of high ice thickness (2800–4200 m) and minimum velocity. Because of their small size (diameter along flight path between 2 and 15 km only), they cover, in the area of maximum frequency (around 75° S 125° E), only 0.5% of the surface.

Wilson's hypothesis of simultaneous circumpolar Antarctic surges in its original form is hardly consistent with the roughness of the subglacial topography and with the results of this model. We should expect surges—not necessarily simultaneous—concentrated around the present ice shelves: the Weddell and Ross Ice shelves, and, to a lesser degree, the Amery Ice shelf (near 70° E). In analogy to present conditions the Weddell area should always have been the most productive. From the viewpoint of the heat budget and of the weather conditions in southern oceans, Wilson's assumption of a continuous quasipermanent ice shelf around Antarctica with a size of $20\text{–}30 \times 10^6$ km² is unrealistic and unnecessary.

However, a surge spreading one-fourth or one-third of the present mass of the Antarctic ice dome—that means a volume of

$6–10 \times 10^6$ km^3—more or less disintegrated from the existing shelf zones into the ocean, during a time-span of a few decades or even centuries, does not seem too unrealistic. Assuming an average thickness of 200 m for tabular icebergs, a nearly simultaneous outbreak of 6×10^6 km^3 would produce an ice-covered ocean area of 30×10^6 km^2 (Table 1, Model SH 3). One of the prerequisites of such an event should be that the height of the ice dome approaches the highest mark on the ice-free mountains (Hoinkes, 1961), which indicates a further growth of 2–300 m is required before the next surge. (If the average positive mass budget is assumed to be—at a maximum!—4 cm/yr (Schwerdtfeger, 1970) a growth of 200 m would need another 5000 yr; both figures, however, are crude estimates.)

Independently of the duration of the surge, each outbreak of continental ice into the ocean must lead to a significant rise of the sea level; some evidence for such glacioeustatic rises has been found (Hollin, 1972). Assuming a mean density of 0.88 g cm^{-3}, each surging volume of 10^6 km^3 should produce a eustatic rise of 2.44 m. Then the first serious consequence of a surge of the size assumed by Wilson (1964) and Hollin (1972) would be a sea level rise of the order of 15–25 m, most probably spread over several decades; at any rate, it would be catastrophic for the densely populated coastal area, including all seaports. Denton *et al.* (1971) have given a critical survey of the climatic and glaciological history of the Antarctic ice-sheets. Evidence for and against the former occurrence of large-scale surges is presented; no really conclusive proof exists at this time. Several height fluctuations of the ice of East Antarctica are quite conspicuous; they do not coincide with the northern hemisphere glaciation (cf. Hughes as quoted in the Addendum, p. 401), in spite of the nearly parallel trend of temperature in both hemispheres. If this is real, it can be taken as a suggestion that short-lived large surges generally produce brief coolings, while the largescale climatic fluctuations are controlled by the long-lasting glaciations of the northern hemisphere and their role in the atmosphere-ocean heat budget.

The effect of an Antarctic surge of this size on the oceanic heat budget depends mainly on ice volume and temperature. Even more important is the effect on surface albedo and thus on the atmospheric heat budget: this depends on the albedo of and the area covered by ice, regardless of the degree of its disintegration. From the Mamabe–Wetherald Model (1967) it can be concluded that a surge of the size indicated by Wilson's hypothesis, with a sea-ice area increase of 30×10^6 km^2, is equivalent to a southern hemisphere temperature drop of 7–8°C (Model SH 3).

South of the oceanic Antarctic Convergence—now situated at lat. 49–50° S in the Atlantic and in the adjacent Indian Ocean, and near 60° S in the Pacific—melting of the surged ice will be quite slow, due to the low surface temperatures of the subantarctic ocean (between 0 and +3°C). Here the latent heat of melting plays only an insignificant role; antarctic cold water is permanently sinking and disappearing at the Antarctic Convergence, feeding the subantarctic intermediate water at a depth of 500–2000 m (Fig. 3). Let us assume that 40% of the injected ice (10^{22} g water equiv) melts in this belt, consuming 32×10^{22} gcal in latent heat distributed over a 1000 m deep ocean layer ($\sim 3 \times 10^{22}$ cm^3) during a period of 100 yr. If all other terms of the heat budget remain constant (which is certainly unrealistic), this melting would produce an annual cooling of about 0.1°C. Thus the regional effect of an ice surge on the heat budget of the subantarctic ocean is only small and short-lived.

The other 60% of the injected ice is assumed to be driven across the Antarctic Convergence into the warmer water on its northern flank. Here it will be disintegrated and melted much faster than before. Because of the position of the Weddell

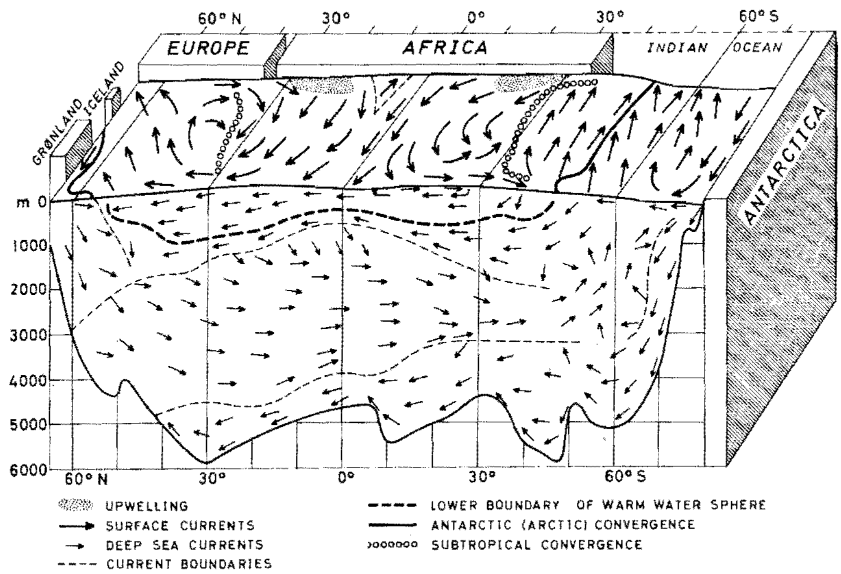

FIG. 3. Meridional cross section of abyssal circulation and surface currents of the Atlantic Ocean. Legend: Sk—subtropical convergence; P—polar front and Antarctic Convergence; dashed—boundary between warm and cold layers; dotted—layer of zero current (simplified after G. Wüst).

Sea—and in agreement with observations during the 19th century (Lamb, 1967; Schott, 1942) we may assume that 35% of the total ice volume penetrates into the narrow Atlantic sector (Fig. 4) (i.e., between long. 20° E and the Drake Passage), and the remaining 25% into the vast areas of the Indian Ocean and of the Pacific, covering three-fourths of the Earth's circumference.

Over the Atlantic sector the drop of the equilibrium temperature according to the Manabe-Wetherald model (1967) will then extend much farther north than over the Pacific and Indian sectors. One may therefore expect a broad, more or less permanent upper tropospheric trough extending to (and partly across) the equatorial region, thus disturbing the subtropical anticyclonic ridge and displacing the ITC region even more to the northern hemisphere than at present, especially during the southern summer. Under present conditions a similar (but weaker) pattern is frequent only during the southern winter. The annual distribution of winds and water temperature anomalies in the equatorial region will then resemble the present northern summer situation: e.g., a southerly flow across the equator and, consequently, prevalence of equatorial upwelling (cf. the results obtained by Henning for July—September, see Flohn, 1972, Figs. 3–4).

While the ice floating into the Indo-Pacific section is of lesser, only regional importance—e.g., in the narrow belt along the eastern coast of Africa (Olaussen *et al.*, 1971)—the ice transported into the Atlantic sector plays a key role. As a basis for discussion we may use the heat balance equation for an upper mixed oceanic layer with constant depth, above the thermocline, in the following form:

$$Q_{sf} - (H_a + LE) - H_m - \operatorname{div} \vec{A_t} - \Delta T = 0.$$

Here Q_{sf} is the net radiation at the surface, $H_a + LE$ is the turbulent flux of sensible and latent heat from sea into air, H_m is the heat used for melting of ice, and

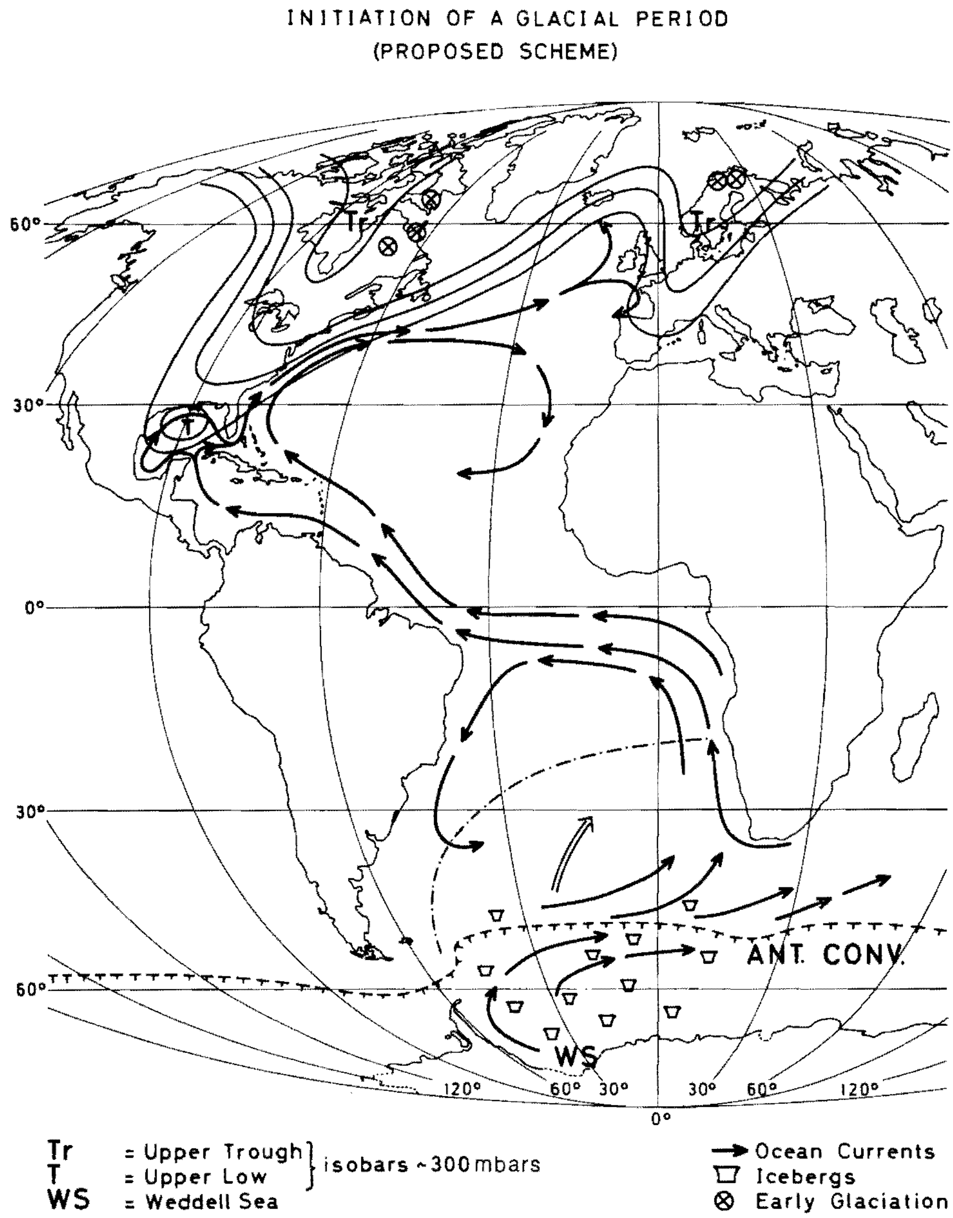

FIG. 4. Initiation of a glacial period (proposed scheme). Southern Hemisphere—Wilson surge (double arrow) across Antarctic Convergence (actual position) up to about lat. 25° S (dash-dot line). Ocean currents—actual situation. Northern Hemisphere during melting period—cold pool in Gulf of Mexico, upper troughs along long. 80° W and W Europe, centers of early continental glaciation.

div $\vec{A_t}$ = divergence of advective heat flux of the ocean layer.

For a first-order estimate of the heat budget changes within the Atlantic current system (Table 2), let us assume that the ice floating across the Antarctic Convergence Zone melts in a short period (of a few decades or centuries) in the south Atlantic between lat. 25° and 50° S ($\sim 18.6 \times 10^{16}$ cm^2). In this area, just in

TABLE 2

ESTIMATED HEAT BUDGET: MIXING LAYER DURING SURGE MELTING
$Q_{sf} - (H_a + LE) - H_m + \text{div } A_t - \Delta T = 0$ (in 10^{20} cal/a)

Atlantic zone	Area 10^6 km²	Transport 10^6 m³/s	Q_{sf}	$H_a + LE$	H_m	div A_t	ΔT 10^{20} cal/a	T_{exit} °C
A. 25–50°S, 50 yr	18.6	8	16.7	0	56	−34	+5.4	0
25–50°S, 100 yr	18.6	8	16.7	0	28	−8.6	+2.7	0
B. 5–25°S (E only) Benguela Current	6.0	8 →	66	38	0	−28 →	+28	11.2
C. 5°S—Carib Isl. Guayana Current	4.3	8 17↑ } 25	51	31	0	−20 −22↑ }	+42	15.8
D. Caribbean + Gulf of Mexico	4.2	25	50	31	0	−28	+28	18.2

front of the main surge region of the Weddell Sea, the ice coverage must be assumed to rise to 0.50; then, with an ice albedo of 0.75, the average albedo is 0.40. With a global radiation of about 118 kLy/yr (Sellers, 1965) and an atmospheric counterradiation of 63 kLy/yr (after Brunt's formula with $T = 273$ K), Q_{sf} will be drastically reduced from its present value of 74 kLy/yr to 9 kLy/yr for a total energy loss of 16.7×10^{20} gcal/yr. The average water temperature of about 14°C should drop to 0°C: this is equivalent—assuming a mixing layer of 100 m—to $\Delta T = 260 \times 10^{20}$ gcal. The melting of 35×10^{20} g ice (water equivalent) needs 28×10^{22} gcal or the equivalent of 168 yr net radiation. During the melting period, the fluxes of H_a and LE will be small and are thus neglected, as a first-order approximation. Assuming durations of the melting period in this zone A of 50 or 100 yr, the estimated annual heat budgets of the mixing layer are given in Table 2 (T_{exit} is the temperature of the resulting mixing layer at the northern boundary of the zone).

In the following and in Table 2 we consider only this oceanic current system: Benguela current–South Equatorial current and its branch north of the South American continent (Guayana Current)–Gulf Stream. It must be assumed that during the melting period many icebergs reach lat. 25° S—even in the 19th century some reached lat. 35° S; the intensity of the cold Benguela Current is therefore assumed to increase from 6 to 8×10^6 m³/s (disregarding the contribution of layers below 100 m). In the zones B, C, and D (Table 2) it is assumed that the extremely cold water increases atmospheric stability, and that the turbulent fluxes $H_a + LE$ are reduced by 40%. If we select a reduction rate of 30% or less (Flohn, 1969) the water temperature (T_{exit}) remains too low; this supports the idea of a drastic reduction of the oceanic evaporation and therefore of tropical rainfall (cf. Chap. VII).

A crucial point is the crossing of the equator: here we may assume that under nearly constant southerly winds (triggered by the difference of ocean temperatures in the two hemispheres) constant upwelling occurs. Its order of magnitude may be estimated by the use of Stommel's (1964) figures: he obtains, averaged between 5° N and 5° S, an upward velocity of 4×10^{-4} cm/s (about 35 cm/d). According to Henning's data (Flohn, 1972) we restrict upwelling to the latitudinal belt 5° S to 1° N; then the upward flow reaches, above an area of 4.3×10^6 km², a value of about 17×10^6 m³/s. This value is of the same order as the mass transport of the powerful

South Equatorial Current of the Atlantic. In calculating the heat budget, we assume a temperature of 14°C for the ascending flow—i.e., the permanent upwelling is assumed to have the minimum temperature observed near the Galapagos. The assumption of a higher ascending mass transport would lead to unreasonably low temperatures (which should have destroyed the coral reefs in the Carribbean). It should also be mentioned that, due to the semipermanent trough situation above the South Atlantic, the splitting of the South Equatorial Current should be displaced to the north, leading to a higher mass transport along the northern coast of South America and reducing the southward Falkland current.

A final value of T_{exit} as low as 18°C in the Gulf of Mexico is not presently supported by any evidence. However, if we take into account that the branch of the North Equatorial Current entering the area as the Caribbean Current, with a similar transport of 26×10^6 km^3/s and a temperature of 26°C, flows in a baroclinic pattern parallel to the course of the chilled waters from the Southern hemisphere, we come to the result that the observed surface temperature of 21–22°C during the ice-ages (Emiliani, 1970) (i.e., averaged over a much longer period of some 10^4 yr) are consistent with an area-averaged value near 22°C.

It would go beyond the purpose of this article to estimate the heat budget of the Gulf Stream during the melting period. It is sufficient to mention that its temperature at Florida Strait should be perhaps 5°C lower than today.

Such an—admittedly crude—consideration of the oceanic heat budget is necessary since the heat capacity of a 3 m water column is the same as that of the whole atmospheric column. If the ocean maintained its temperature and the cooling of the atmosphere above the greatest part of the globe occurred only due to advective processes in the air, the global climatic effect of an Antarctic ice outbreak would be rather insignificant. In this case it would be impossible to understand in what way the albedo-produced cooling of the subantarctic atmosphere could expand into the northern hemisphere, across the vast area of the warm tropical oceans covering—between lat. 30° S and 15° N—about 78% of the surface.

It has been shown in Chap. V that the key to the initiation of a northern glaciation (either complete or incomplete) is the summer climate of Labrador–Ungava, including Keewatin and probably Baffin Island. A quasipermanent pattern of low level polar air advection together with upgliding warm moist air can only be conceived together with a quasipermanent east coast trough reaching from Labrador to Florida (Fig. 4). Such a situation must be produced, sustained, and fixed by a pool of cool water in the Gulf of Mexico and Caribbean, with surface temperatures around 22 instead of 27–28°C. The effect of this situation in summer (with the strongly heated continent in the north) is much stronger than in winter; during the warm season a permanent high-tropospheric vortex will be maintained, and an upper trough will be fixed along the east coast by a slightly cooler Gulf Stream.

The geophysical model which has been outlined here has one advantage over others: it needs only a quite short timespan of 50 or 100 yr to produce a nucleus for glaciation growing by positive feedback, and is therefore apparently consistent with the evidence presented in Chap. IV. A crucial test of this model, however, lies in the magnitude of the unavoidable sea level rise (Hollin, 1972). The evidence presented to date is hardly sufficiently conclusive; however, it is certainly difficult to find convincing traces of a marine transgression with a lifetime of the order of only 10^3 yr. Immediately after the establishment of this circulation anomaly, the storage of wa-

ter in the form of ice above the northern continents begins and increases rapidly by the abovementioned feedback process.

VII. TIME AND SPACE CORRELATIONS

Interpreting global coolings as caused by large-scale surges of the Antarctic Ice, concentrated mainly in the Weddell Sea–Atlantic section, we may also comprehend some correlations with other evidence which remained hitherto hardly understandable.

(a) In all tropical continents, definite signs of a marked desiccation simultaneously with the northern glaciations have been found (Fairbridge, 1972). This is even true for the equatorial rain forests in Central Africa and South America (Vuilleumier, 1972; van Zinderen Bakker and Coetzee, 1972); in both areas semiarid dry forest prevailed, with only a few islands of humid forest. If under the present radiation regime the oceans are advectively cooled, the evaporation from the tropical oceans as the main source of the global hydrological cycle (68%) must have been substantially lower; a rough estimate based on advective processes alone (Flohn, 1969) gave a decrease of about 30%. Under such conditions the intensity and extension of the tropical Hadley cell must have decreased in comparison with present conditions.

(b) In several parts of the Atlantic sector the lowering of the snowline and vegetation limits during glacials was much greater than usual (equatorial zone some 800 m, midlatitudes about 1200 m). Here we mention the Itatiaya near Rio de Janeiro (Mortensen, 1957), the Costa Rica Volcanoes (Weyl, 1956) and, at least in some times, the Sabana of Bogota (van der Hammen *et al.*, 1971), each with a glacial cooling of up to 8°C (in earlier glaciations even 11°C) instead of 5°C in other areas (East Africa, Indonesia, New Guinea). Similar data have been collected from the humid Andes Mountains of Colombia and Venezuela (Wilhelmy, 1957). Of special interest is the somewhat controversial evidence for a widespread low-land glaciation of Eastern Patagonia (Czajka, 1957).

(c) According to new data collected in a interdisciplinary German–Mexican Project (Heine, 1973a, 1973b), the ^{14}C timescale of climatic events in the last 40,000 yr in the high volcanoes in Central Mexico deviates in a characteristic way from the usual sequence; the greatest glacial advances occurred at about 32,000 and (during 4–500 yr only) 12,000 BP, i.e., within the Allerød oscillation.

(c) The striking contrast between the glacial temperature anomalies in the area around the Bay of Biscay, from Ireland to Northern Spain (—12°C or even more), and at the same latitude at the Pacific coast of North and South America (about —5°C) has been stressed earlier (Flohn, 1969); it is consistent with abrupt progressions of polar water masses to lat. 42° N in the Atlantic (McIntyre *et al.*, 1972) i.e., more than 20° south of their present position. From a comparison between micropaleontological evidence and isotopic temperatures, Emiliani (1971) concludes that glacial surface temperatures were about 7–8°C lower in the Caribbean, 5–6°C in the equatorial Atlantic, but only 3–4°C in the equatorial Pacific. The contrast of the large extension of ice sheets on both sides of the Atlantic with the comparatively small glaciation on both sides of the Pacific requires also a geophysical interpretation (Flohn, 1969). According to this version of the Antarctic Surge model, cooling of the Pacific may have occurred only as a secondary effect.

After the revised calculation of the solar radiation fluctuations due to orbital variations (Vernekar, 1972) the equal severity of the most recent and earlier glaciations is difficult to understand. In contrast to the solar variations, the glacial and climatic history of at least the last 20,000 yr shows a clear coincidence at both hemispheres, instead of a time-lag of the order of 10,000 yr. Together with the discrepancies in the

time-scale involved, this seems to be one of the strongest arguments against a primary role of the extraterrestrial "Milankovich effect," which appears to have mesmerized nearly two generations of earth scientists, in spite of many sober and critical voices.

According to heat transport considerations (Newell, 1973) the initiation of a glaciation needs a heat deficit of 1 to 3×10^{19} gcal/d. A Wilson surge (Model SH 3) yields, with an average global radiation of 350 Ly/d, a heat deficit of about 9×10^{19} gcal/d or about 4500 TW lasting about a century. If the melting process of the ice lasts the same period, an additional loss of nearly 2×10^{19} gcal/d is to be added.

In contrast to this, the peak-to-peak variation of summer insolation due to the Milankovich effect—i.e., spread over a period of several 10^4 yr!—is 35×10^{20} gcal (Broecker, 1968) or little less than 2×10^{19} gcal/d, equivalent to 3.8 W/m^2 or 970 TW.

The mass of new evidence, consisting of quantitative determinations of temperature and age (certainly not without sources of systematic error!) has shown that the time-scale of the glacial–interglacial sequence is much more complex than the classical one (Emiliani, 1972). Obviously a series of hemispheric-scale coolings occurred, some followed by a glaciation on the northern continents, others not. Such "abortive" coolings, with a time-scale of a few centuries, are of vital interest to the meteorologist: in the human time-scale they are "irreversible," i.e., from the viewpoint of living mankind, of the economist, and the politician. They indicate that Lorenz's (1968) unorthodox suggestion of a potential instability of our climate is quite realistic and must be taken as a serious challenge of utmost significance.

The powerful feedback between the strong albedo gradient on the outer boundary of the snow- and ice-covered region and the baroclinic frontal zones (acting as cyclone tracks) contributes strongly to the development and maintenance of continental glaciations in the northern hemisphere. It is much more difficult to understand the total interruption of this process, which leads to disintegration and finally to deglaciation. In other words: what physical causes are responsible for the transition from an ice-accumulating to an ice-destroying pattern of the atmospheric circulation, right at the culmination of each glacial? It has been argued (Hoinkes, 1961; Bloch, 1964), that the aridity of northern continents causes loess dust from the barren fluvioglacial deposits around the ice (with most particles well below 2 μm) to be blown onto the ice, resulting in a lowering of the surface albedo. However, the role of a dust-laden atmosphere—as it can be observed now during summer above Pakistan, Turkestan, and Sinkiang—is more complex: Low-level dust absorbs solar and long-wave radiation and heats the atmosphere and the surface substantially. A most remarkable example of this effect has been observed recently in the Martian atmosphere (Gierash and Goody, 1972). Direct heating and atmospheric infrared radiation (at high temperatures) are more powerful melting agents than the decrease of the albedo alone. Time variations of the Ca content of the Greenland ice (Hamilton and Seligo, 1972) are apparently consistent with this hypothesis: the occurrence of Ca maxima after the beginning of cooling supports our view.

VIII. SUMMARY

In a time-scale of 10^4–10^6 yr the climate of the Earth–atmosphere–hydrosphere–cryosphere system is in fact unstable. This complex, self-regulating system—with nearly constant energy input—is in a delicate state of equilibrium, with its energy budget depending on the variable area of its subsystems. During the last 10^5 yr, evidence of at least five rapid hemispheric or global coolings has been found with temperature changes of about 5°C (i.e., the full difference between present and

iceage climate), occurring during a time-span of the order of a century. Only some of them led to a complete glaciation of the northern continents, others ended after a few centuries with a sudden warming. These facts are not compatible with the widely accepted orbital variations with a time-scale of some 10^4 yr ("Milankovich effect").

Starting from a modified version of A. T. Wilson's hypothesis of large-scale surges of the Antarctic ice dome, a purely geophysical model of such rapid coolings can be outlined:

(1) After a period of slow accumulation, the Antarctic ice dome surges—not necessarily simultaneously—mainly in the existing shelf areas, especially in the Weddell Sea. The amount of ice calving into the ocean is estimated to be $6–10 \times 10^6$ km^3, causing a eustatic sea level rise of 15–25 m, spread over several decades or centuries.

(2) The disintegrated ice spreads, in enormous tabular icebergs, over an area of $20–30 \times 10^6$ km^2; due to the increase of the surface albedo the average temperature of the southern hemisphere drops about 7°C. Since the tropical zone is, at the very beginning, only weakly affected, the midlatitude circulation intensifies significantly.

(3) During the melting period, a considerable part of the ice crosses the oceanic Antarctic Convergence, notably in the Weddell Sea–Atlantic sector. According to an estimate of the heat budget of the upper mixed layer, the melting process causes advective cooling of the system Benguela Current–South Equatorial Current–Gulf Stream of the order of at least 6–8°C during a time-span of about 50–100 yr, in addition to the albedo-produced cooling. This cooling should be accompanied by a broad upper trough in the South Atlantic and a marked cold-arid phase in the neighboring continents.

(4) Advective cooling of the surface of the Caribbean and the Gulf of Mexico, together with the Gulf Stream, causes, during the warm season, a permanent high-tropospheric vortex together with a deep trough along the eastern coast of North America. This circulation anomaly produces cooling and increased snowfall in the Labrador–Ungava region and enables a survival of the snow-cover during summer, as a potential nucleus of a continental glaciation.

(5) When the permanent snow-cover has reached a diameter of several hundred kilometers, a positive feedback mechanism (snow surface with high albedo–tropospheric cooling–cold upper vortex–enhanced snowfall) leads to a fast-growing ice sheet. This localized heat sink maintains a quasi-permanent trough–ridge pattern with blocking highs east of Alaska and east of Iceland, the latter producing a second ice center in the mountains of Fennoscandia.

(6) Further development of these ice centers either to a complete glaciation or to a reversal may perhaps be controlled by the actual state of the orbital elements; further studies are needed. From the meteorological point of view, an incomplete ("abortive") glaciation with a duration of only a few centuries is as important as a complete glaciation with a period of 10,000 yr.

(7) The final disintegration of the ice domes of the northern continents—in spite of the powerful feedback referred to in no. 5—can be understood as caused by frequent dust storms, lowering the surface albedo of the glacier, and heating the lower atmosphere through absorption of solar and long-wave radiation.

Such a geophysical model may serve as a background for a complete physicomathematical model of the complex multiphase system designed to simulate these most dramatic events in the climatic history of the Earth. Admittedly, our knowledge of the complex interaction of processes in our geophysical system is at this time rather inadequate. Several highly interesting model experiments try to simulate the climate of a fully developed glacial epoch (Alyea, 1972; Williams *et al.*, 1973). Here we propose a much more difficult, but also more

rewarding future task: the simulation of the initiation of a new glaciation. If—as stated in the introduction—a new glaciation should be expected to begin during the next, say, 5000 yr, one should expect, perhaps, a probability near 1%, that this may happen during the next 50 yr.

Man can hardly interfere with the mass budget and the intensity of the Antarctic ice. Denton *et al.* (1971) have pointed out that a minor increase of temperature—as we expect as a result of man-made effects—may lead to disintegration of the smaller West Antarctic ice sheet; both effects should rise the accumulation rate in East Antarctica. At any rate, this problem is not only of pure academic interest: it refers to our own planet Earth, to our habitat now and in the future. Within the lifetime of our generation, it deserves a much higher priority.

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The paper shall be devoted to the memory of the late F. Loewe (deceased March 1974). The careful revision of the original manuscript by Prof. Sue Bowling (Fairbanks, Alaska) is gratefully acknowledged, as well as some comments from the reviewers.

ADDENDUM

In a very thoughtful paper, T. Hughes (*J. Geophys. Res.* **78,** 1973, 7884–7916) evaluated a nonequilibrium profile along a flow-line of Ross Ice Shelf suggesting an instability of the smaller West Antarctic Ice Sheet. The slow retreat of the grounding line of the Ross Ice Shelf is greater than that caused by the postglacial eustatic sea level rise (which was, however, of the order of 100 m/8000 yr or 12 mm/yr); further retreat means therefore an increase of floating shelf ice and of the present eustatic rise near 1 mm/yr. Since the Weddell Sea drainage area is much more unknown than the Ross Sea area, similar investigations there are badly needed.

The physical background of the ice core data from Greenland and Antarctica has been recently discussed by W. Dansgaard, S. J. Johnson, H. B. Clausen, and N. Gundestrup in a monograph (Meddel. om Grønland **197,** No. 2, 1973, 53 pp). The use of the hypothetical cycles of constant length as a base for a time-scale has been seriously doubted by N. A. Mörner (*Boreas* **2,** 1973, 33–53) and (*Geol. Magaz.* **111,** 1974, in press) as well as the use of a logarithmic scale based on a flow model with constant velocity. Certainly any substantiated improvement of the chronology would be highly welcome; due to the relatively high degree of coincidence with many independent results (cf. Chap. IV) it is not expected that the results presented here would be significantly changed by an improved chronology.

In addition to Chap. VII paragraph (b) reference should be made to the isolated and short glacier advance in southern Chile (J. H. Mercer and C. A. Laugenie, *Science* **182,** 1973, 1017–1019) around 36,000 BP and to the evidence of frost weathering at the South African Cape Coast indicating a winter temperature depression of about 10°C (K. W. Butzer, *Boreas* **2,** 1973, 1–11). Such examples of short-living episodes deserve more attention; the highlands of Angola and Southwest Africa should be of special interest.

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