Interactions of ice sheets: instability and self-organization

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ABSTRACT The entire dynamic systems consisting of ice sheets and their environment are considered in order to understand the mechanisms of instability. We propose that the Late Cenozoic behavior of the global climatic system ice-ocean-atmosphere-lithosphere (including Pleistocene glacial-interglacial cycles) is triggered by a series of glacial catastrophes.

INTRODUCTION

Instability of glaciers and ice sheets occurs under various geographical conditions. Surges of mountain and outlet glaciers are, perhaps, most notorious (Clarke et al., 1977; Kamb et al., 1985). There is a variety of jump-like processes within glaciers and ice sheets and these processes may be described uniformly in terms of catastrophe theory (Gilmore, 1981). Namely, the curve of equilibrium states of glaciers and ice sheets is a cubic one.

The instability may be caused by the internal properties of glaciers and ice sheets. But they are not isolated geographical objects as they interact with atmosphere, ocean, and lithosphere. The interactions lead to a variety of instabilities in both ice masses and their environment. So we should consider the entire dynamic systems comprised of ice masses and the other bodies in order to describe the instabilities.

The instability of ice mass/environment systems results in a number of geographical consequences. The instability caused by interaction with atmosphere puts limits on ice sheet dimensions. In particular, purely terrestrial polar ice sheets cannot reach a stable state—they are transformed into marine ice sheets. The instability caused by interaction with the ocean naturally explains fast collapses of marine ice sheets. Both instabilities lead to oscillations of ice sheets.

However, the instability of ice mass/environment systems is not only "destructive". It may be "creative" when instability results in synergetic effects and leads to self-organization of regular spatial structures (Haken, ...
1978) within ice masses and their environment. For example, the instability caused by the interaction of ice sheets with their beds gives a natural mechanism for self-organization of ice streams within ice sheets and large-scale linear landforms (coastal fjords and submarine troughs) within their beds. So glaciology accepts the challenge by Professor H. Haken, one of the "fathers" of synergetics. Prof. Haken wrote (1985) to geographers: "We will present a few further examples from chemistry and biology, and hope that synergetics may find application in geography and planning. However, it is for the specialists in these fields to decide whether this approach can be used in their studies".

The instabilities of ice sheet/environment systems discussed here are a convenient tool to explain Late Cenozoic variations of ice sheets and climate.

![Cubic curve of equilibrium dimensions of a terrestrial ice sheet against snow-line parallel translation. Solid lines of the cubic curve represent stable equilibria, the dashed line represents unstable equilibria.](image)

**FIG. 1** Cubic curve of equilibrium dimensions of a terrestrial ice sheet against snow-line parallel translation. Solid lines of the cubic curve represent stable equilibria, the dashed line represents unstable equilibria.

**INTERACTION WITH THE ATMOSPHERE**

As regards terrestrial ice sheets, whose dynamics is governed by their interaction with atmosphere, both responses - stable and unstable - occur. The kind of response depends on the geometry of a snow surface showing climatic conditions for ice sheets and separating their accumulation and ablation zones. The snow-line elevation decreases as latitude and proximity to ocean increase. There are plural equilibria of the ice sheet under given climatic conditions. Fig. 1 shows a full set of equilibrium states of a terrestrial ice sheet as the snow line varies while its slope remains constant. The curve of equilibrium states looks like a reduced cubic curve. Hence all generic properties of cubic curves are applicable to the equilibrium curve of a terrestrial ice sheet. In
particular, there are three critical sizes of an equilibrium terrestrial ice sheet: \( L_0 = 0 < L_1 < L_2 \). When an ice sheet is unstable and its dimensions are smaller than the smaller critical size \( L_1 \), these dimensions are "prohibited" (the first prohibition principle). Moreover, although an equilibrium ice sheet is stable its dimensions may be between the smaller critical size \( L_1 \) and the larger critical size \( L_2 \), then these dimensions may occur only while retreating and are "prohibited" while advancing (the second prohibition principle). The dimension of advanced ice sheet jumps from zero right to the larger critical size \( L_2 \) - the dimensions smaller than the larger critical size \( L_2 \) are "unattainable" for an advanced ice sheet. The "attainable" dimensions of an advancing ice sheet are larger than the larger critical size \( L_2 \), but they are only potential dimensions: in reality, the spread of a terrestrial ice sheet may be stabilized at a smaller size level by switching on new ablation mechanisms.

Calculations in a perfect plasticity approximation supposedly (Mazo, 1989) yield the following estimates: \( L_1 = \text{const} \alpha^{-2} \) for all snow-line slopes, and \( L_2 \approx 8000 \text{ km} \) for a typical slope \( \alpha = 10^{-3} \). Hence the larger critical size \( L_2 \approx 8000 \text{ km} \) is of a "continental scale". This estimate confirms Grosswald's (1989) picture of the Pleistocene Eurasian ice sheet system. In East-West direction the snow-line slope is smaller than the typical slope \( 10^{-3} \), as the larger critical size is larger than \( 8000 \text{ km} \). Hence it was not difficult, at least potentially, for the Eurasian ice sheet system, as a single continuous system, to extend from Scandinavia eastward right to Chukotka. In the equator-pole direction the size of the Eurasian ice sheet system was not as large as \( 8000 \text{ km} \), although the typical snow-line slope \( 10^{-3} \) was reliable. The only reason for the size reduction was that the northern fronts of the Eurasian ice sheet system were not stabilized at the land/ocean boundary but extended northwards over the continental shelves. Therefore, the terrestrial ice sheet system was transformed into a marine ice sheet, and a new - ice/ocean - kind of interaction was involved. However, the state of the marine ice system was controversial: the ice/ocean interaction stabilized the system advance, yet at the same time led to a new - marine - kind of instability.

As regards high-plateau ice sheets, e.g. a former Tibetan ice sheet (Kuhle, 1988), the same arguments may be applied to them as well: a high-plateau ice sheet is stabilized as far as at the edges of the plateau covering it completely.

INTERACTION WITH THE OCEAN

Ice-atmosphere interaction cannot stabilize the advance of polar terrestrial ice sheets. Terrestrial margins transform into marine margins and terrestrial ice sheets
transform into marine ice sheet/ice shelf system. Ice calving stabilizes the ice advance. However, ice-ocean interaction creates a new mechanism of ice instability.

Instability of marine ice sheet/ice shelf systems is a consequence of the pulling force at terminal ice walls (Weertman, 1957). The pulling force increases as the square of ice thickness and is the dominant force within a floating ice shelf from an ice wall nearly to a grounding line because of total uncoupling of ice from the bed. Moreover, the pulling force reaches far into a marine ice sheet along ice streams. The latter are concave transitional zones where the pulling and shear forces are comparable. Concave transitional zones link a convex ice-sheet interior where the shear force is dominant and nearly flat ice shelves where the pulling force is dominant.

Fjords serve as a bed for ice streams. Typically, a fjord is terminated by a bedrock sill at the edge of the continental shelf and a headwall in coastal mountains with subglacial basin between them (Fig. 2). So the longitudinal profile of the ice-stream bed is a cubic curve. There are three slopes: two uphill slopes (the first the continental slope beyond the bedrock sill and the second is the landward slope of the subglacial basin from the bedrock sill to the floor of the basin). If a grounded line happens to be on the uphill slope, the ice thickness increases while advancing and decreases while retreating. With an increase in ice thickness, the pulling
force increases. Since ice discharge increases with an increase of pulling force, the uphill slopes are stable states of the grounding line. On the contrary, the downhill slopes are unstable states of the grounding line. Therefore, the cubic curve of the ice-stream bed is a set of stable and unstable equilibria of the grounding line against its location.

All properties of the generic cubic curve may be applied to the stability problem of marine ice sheet/ice shelf systems. Particularly, there are critical positions of the grounding line. The first is the crest of the bedrock sill. Once the grounding line leaves the slope of the continental shelf and retreats over the bedrock sill, the pulling force causes the marine ice sheet/ice shelf system to collapse. While collapsing the grounding line jumps from the crest of the bedrock sill to the uphill slope of the subglacial basin. The grounding line is stabilized at an elevation where the collapse began. This manner of the collapse of marine ice sheet/ice shelf systems is a common place in marine glaciology.

Another critical position is the floor of the subglacial basin. While advancing a grounding line moves slowly down the landward slope of the subglacial basin to its floor. When the grounding line reaches the floor, it jumps seaward to the slope of the continental shelf at the elevation of the basin floor. An example of the catastrophic advance of a marine ice system is surges of tidewater glaciers without climatic forcing (Mercer, 1961).

Qualitative description of the instability of marine ice systems should be supported by theoretical and computational models. However, some difficulties arise. The main is that the model has to integrate the dynamics of convex terrestrial, concave transitional, and nearly flat floating components of the systems. There are reliable models of the terrestrial (see, e.g. (Hutter, 1983)) and floating (Morland, 1987) components which are driven by only one kind of a dominant force: the shear force within terrestrial components and the pulling force within floating components. Transitional components - marine ice streams - are much more difficult for theorizing and computing because both the forces govern their dynamics. A number of theories have been proposed (Lingle, 1984; Lingle & Clark, 1985; Van der Veen, 1985; Fastook, 1987; Hughes, 1987; Lingle & Brown, 1987; Muszynski & Birchfield, 1987; MacAyeal & Barlicon, 1988; MacAyeal & Lange, 1988; Mazo, 1989). The theories are partly compatible, partly controversial. A "final" experiment is needed to choose a theory-to-be.

The dynamics of marine ice streams is not the only problem to be solved. A variety of problems associated with the whole marine ice systems are listed by Hughes (1987).
INTERACTION WITH THE BED

Ice streams are associated with large-scale longitudinal troughs (coastal fjords and submarine channels) dissecting the ice-sheet bed and being the deepest landforms within the bed. Since the pulling force increases with an increase of ice thickness, ice streams are most unstable members of marine ice sheet/ice shelf systems.

It is glacial erosion that causes large-scale longitudinal troughs, and the rate of glacial erosion is sufficiently high to create the above landforms (Grosswald & Glazovsky, 1983). But the next question arises: why glacial erosion is not homogeneous within the ice-sheet beds, why does it localize in particular zones and thus generates trough valleys?

According to a widespread opinion, heterogeneities of glacial erosion occur because of a variety of different non-glacial and ad hoc reasons: preglacial bed configuration, lithological differences of bedrock, and others (see the critical review by Grosswald & Glazovsky, 1983).

An alternative approach is to connect the heterogeneities of glacial erosion to heterogeneities of ice-sheet flow. The bed does not only vary under the impact of glacial erosion: ice-sheet flow, in turn, varies according to changes in the bed. Therefore, we encounter a purely glaciological self-supported process for the generation of heterogeneities of ice-sheet flow and bed configuration. It appears adequate to connect the dynamics of the eroding ice sheet and eroded bed to describe the genesis process.

According to the synergetic ideology (Haken, 1978), we should investigate the dynamics of random small-amplitude perturbations against the background of a homogeneous flow and a homogeneous bed. The perturbations may develop into finite-amplitude spatial structures like trough valleys.

We postulate that initially there are small random transverse perturbations, i.e., small random longitudinal channels on the otherwise homogeneous bed. These bed perturbations cause small transverse perturbations of the otherwise homogeneous ice flow, i.e. incipient ice streams. If the flow perturbation cannot overcome the dissipative effect of ice viscosity, the incipient ice streams as well as the incipient channels die out, and the ice flow and bed configuration tend to a homogeneous state. However, if the flow perturbation overcomes the dissipative effect of ice viscosity, the small channels deepen because of the high rate of erosion caused by rapid ice streams. This results in further acceleration of the ice streams and in further deepening of the corresponding channels in an unstable fashion.

The analysis (Mazo, 1987, 1989) shows that, indeed, both phenomena may take place. The spacing of ice streams and channels should correspond to scales at which instability occurs. The wider is spacing, the more
unstable are the perturbations on the bed and in the ice flow. Then the characteristic spacing of both heterogeneities in the ice flow and corresponding "peaks" and "troughs" in the bed should be much wider than the average ice thickness (Fig. 3).

![Diagram of an ice sheet](image)

**FIG. 3** Transverse section of an ice sheet.

To summarize, we may say that glaciated and deglaciated surfaces should be smooth, but dissected by widely spaced channels along present and former ice flow directions. The field data confirm this postulate (Flint, 1971; Grosswald & Glazovsky, 1983).

The analysis above shows that the genesis of glacial relief may be explained by purely glaciological causes. The critical feature is the dependence of the bed-erosion rate on the parameters of the eroding ice masses. This leads to coupling between the ice flow and bed topography.

**ICE-SHEET INSTABILITY AND LONG-TERM CLIMATE VARIATION**

Marine ice sheet/ice shelf systems are the most aggressive and unstable members of the global ice-ocean-atmosphere-lithosphere climatic system in its long-term variations (Lockwood, 1095; Ruddiman & Duplessy, 1985). Let us consider long-term climate variations on three time scales.

Figure 4a shows climate variations on a longer - Cenozoic - time scale (Willett, 1950). There are three
singular points on the Cenozoic temperature curve. Initially, climate variations were reduced to a slow monotonous cooling. Changes in geography of the continents and the ocean are supposed to cause the cooling. The interval ended with a series of abrupt falls of temperature. The first singular point divides the gradual and steep branches of the Willett’s curve. The second marks transition in temperature changes from a monotonous cooling to oscillations (20–30 Myr B.P.). The third – an abrupt increase in the oscillation magnitude (2–3 Myr B.P.). These three transitions were synchronous with three great glaciological events.

![Cenozoic annual air temperature in Central Europe](image1)

**FIG. 4** (a) Cenozoic annual air temperature in Central Europe (Willett, 1950) (numbers show singular points); (b) spectral analysis of the marine oxygen-isotope record for the last 650 kyr (Hays *et al.*, 1976; Imbrie, 1985); (c) a smoothed version of the marine oxygen-isotope record for the last 150 kyr (Broecker & van Donk, 1970).

The Late Cenozoic temperature falls happened when the climate conditions reached critical levels corresponding to both/either water-ice transition on the ocean surface and/or the transition from "black" to "white" snowy
land–surface (Berger et al., 1981). The positive feedback of the surface to atmosphere led to the cubic curve and cusp surface as plots of climate states (see, e.g., (Oerlemans & Van der Veen, 1984). The temperature jumps are a consequence of these generic plots.

The other two abrupt climatic changes happened when large ice sheets arose. Antarctic Ice Sheet originated in 20–30 Myr B.P. and ice sheets in northern polar regions in 2–3 Myr B.P. Therefore we are forced by these coincidence to incorporate ice sheets in the global climatic system when the system oscillates. To clarify the role of ice sheets in climatic oscillations, let us consider the oscillations on shorter time scales.

Figure 4b shows spectral analysis of the climatic oscillations on an intermediate-Pleistocene-time scale. The evidence of the Pleistocene Earth’s climate comes primarily from oxygen-isotope deep-sea sediment records (Hayes et al., 1976; Imbrie, 1985) and from coal reef elevation, sea surface and global ice records (reviewed recently by Broecker & Denton (1989)). The marine oxygen-isotope and other records show long-term climatic variations modulated by 19/23, 41 and 100 kyr periodicities. In accord with the orbital theory by Milankovitch (1941) and its modern versions (Hayes et al., 1976; Imbrie et al., 1984), the global climatic system responds to variations in summer insolation caused by variations in the Earth’s orbital geometry of the same periodicities. In fact, 19/23 kyr and 41-kyr climatic oscillations vary in time-lag harmony with 19/23-kyr and 41-kyr orbital oscillations. However, a puzzle is that the dominant periodicity in climate oscillations is 100 kyr rather than 19/23 kyr and 41 kyr, while orbital forcing has a greater power at the 19/23 and 41-kyr frequencies.

Figure 4c shows a "fine structure" of the 100-kyr cycles. Another puzzle is the asymmetrical shape of the 100-kyr cycles. Long intervals of gradual decrease in about 100 kyr ended abruptly with rapid increase in less than 10 kyr referred to as terminators (Broecker & Van Donk, 1970). Additional problem arises because during gradual increase the Earth’s climate system and ice sheets never respond irreversibly to the Millankovitch’s warm peaks except one – a 100-kyr cycle. Only the last warm peak within a 100-kyr cycle causes irreversible response of the Earth’s climate and ice sheets.

Over the last decades, much effort has gone to the modelling of a non-linear link of orbital forcing to the Earth’s climate and global ice (see discussion by Hughes (1987) and Broecker & Denton (1989)). The most difficult challenge is to produce 100-kyr cycles ending with sharp terminations. Most models address only terrestrial ice sheets which are supposed to be stable in their response to insolation forcing. Denton & Hughes (1983) and Denton et al. (1986) suggested a conceptual model that incorporated marine ice systems along with terrestrial ice
sheets. Full-bodied marine ice systems are usually unstable and collapse.

Below we suggest to integrate the peculiarities of the Late Cenozoic climate dynamics on a single basis of ice instability presenting them as a trigger for long-term climate changes.

Late Cenozoic sharp falls of temperature followed by its gradual decrease lowered snow surface significantly. The snow-surface depression was sufficient to switch on the mechanism of ice-sheet instability caused by ice-atmosphere interaction. Large ice sheets "instantaneously" occupied polar plains and high plateaus. Terrestrial polar ice sheets advanced into ocean until they reached the edges of the continental shelves and transformed into marine ice sheet/ice shelf systems.

When full-bodied, ice sheets interacted with the bed. This erosional interaction resulted in self-organization of ice streams within ice sheets and large-scale linear landforms within the bed associated with ice streams.

Development of ice streams, along with isostatic sinking, made marine ice systems unstable in relation to ice-ocean interaction. While ice domes of marine ice systems remained relatively stable ice streams were the most unstable members of these systems. With a maximum insolation, ice-stream instability caused by sea-level rising and other reasons (listed by Hughes (1987)) was released and led to the propagation of calving bays up ice streams and downwards in ice-stream drainage basins. As a result, the whole marine ice systems disintegrated into isolated ice domes and collapsed in a short time (less than 10 kyr).

Unstable response of an ice sheet was not possible to every insolation maximum, but once during a 100-kyr cycle. An ice sheet had to be "ripe" for instability. The conditions of the ripeness depend on ice-ocean interaction. A collapsed ice sheet was mostly terrestrial and responded reversibly to Milankovitch’s warm peaks. It required a long time (approximately 100 kyr) for the ice sheet to become again an unstable marine ice sheet/ice shelf system.

Theoretical and computational models of marine ice streams and the whole marine ice sheet/ice shelf systems could have confirmed by this speculation. However, proposed models are not yet reliable because they do not describe tutti the terrestrial and marine conditions for ice.

We would like to emphasize that ice instabilities are a trigger only for the global climatic system. Ice instabilities cannot explain all peculiarities of the long-term climate dynamics of their own and we need to incorporate ocean instability (Broecker & Denton, 1989). But it is ice sheet instability that switches on the ocean instability.
CONCLUSION

The point of this paper is to prove that ice sheets are inherently unstable. All kinds of ice sheet interaction with their environment cause instability of ice sheets and, correspondingly, of the global climatic system.

The Pleistocene dynamics of the global climatic system is very much non-equilibrium, it represents a cyclic series of instabilities. A 100-kyr cycle is divided into two phases: a long interval of unstable buildup of ice sheets controlled by ice-atmosphere interaction and a short interval of unstable collapse of ice sheets controlled by ice-ocean interaction. Ice-bed interaction prepares the conditions for unstable collapsing.

REFERENCES


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