ICEBERG CALVING AND ICE SHEET MARGIN DYNAMICS,

WEST GREENLAND.

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Doctor of Philosophy,

University of Edinburgh, 1990.

i

I declare that this thesis is all my own work.

CHARLES WARREN,

DECEMBER 14th, 1990.

Fir'd at first sight with what the Muse imparts
In fearless Youth we tempt the Heights af Arts,
While from the bounded Level of our Mind
Short Views we take nor see the Length behind,
But more advanc'd, behold with strange Surprize
New, distant Scenes of endless Science rise!
So pleas'd at first the towering Alps we try,
Mount o'er the Vales, and seem to tread the Sky;
Th'Eternal Snows appear already past,
And the first Clouds and Mountains seem the last:
But those attain'd, we tremble to survey
The growing Labours of the Lengthen'd Way,
Th'increasing Prospect tires our wandring Eyes,
Hills peep o'er Hills, and Alps on Alps arise!

Alexander Pope, 1711.
'An Essay on Criticism',
lines 219 - 232.

O LORD, our LORD,
    how majestic is your name
        in all the earth!
You have set your glory above the heavens...

When I consider your heavens,
    the work of your fingers,
  the moon and the stars,
    which you have set in place,
  what is man that you are mindful of him,
    the son of man that you care for him? ...

O LORD, our LORD,
    how majestic is your name
        in all the earth!

Psalm 8: vv. 1,3,4 & 9.

ii

A C K N O W L E D G E M E N T S .

UNIVERSITY OF EDINBURGH
ABSTRACT OF THESIS (Regulation 3.5.10)

Ice sheets are integral to the earth's climate system, both modulating and responding to climatic change. Iceberg calving fronts are the only dynamic interface at which the atmosphere, oceans and ice sheets directly interact. Calving introduces mechanical instability to glacier systems such that the response of calving glaciers to climatic forcing is commonly non-linear. The interaction between calving dynamics and the ice-marginal environment, notably the topographic geometry of glacier troughs, can partially or totally decouple glacier fluctuations from climate for periods of several centuries.

In West Greenland these instability mechanisms appear to have been important both during deglaciation and recently. In the Late Glacial/early Holocene, trough geometry controlled the retreat stages of the ice sheet margin in the Ilulissat (Jakobshavn) area of central West Greenland. During the second half of the twentieth century, the oscillations of 72 outlet glaciers between 61°N and 72°N show that land-terminating glaciers respond directly to climate change (albeit with variable time lags) but that calving glaciers behave non-linearly. Freshwater calving glaciers have lower calving fluxes and calving rates than tidewater glaciers, and may be the first to respond to climatic cooling.

It is not clear whether ice sheet outlet glaciers oscillate cyclically as do calving mountain glaciers, but the instabilities introduced by calving cause many glaciers to respond more directly to topographic than climatic factors. It is therefore hazardous to attach palaeoclimatic significance to the glacial geomorphological record of the fluctuations of former calving margins, or to regard the behaviour of contemporary calving outlets as indicative of climatic trends. Factors affecting the stability of ice margins have a fundamental impact on the dynamics of ice sheets, and are important controls on the timing and patterns of ice sheet response to climate change.

C O N T E N T S .

C H A P T E R S .

1. ICE AND CLIMATE - UNDERSTANDING THE EARTH CLIMATE SYSTEM.

2. CRYOSPHERE - HYDROSPHERE INTERACTION: CALVING THEORY AND IMPLICATIONS.

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xi

xii

L I S T   O F   T A B L E S .

CHAPTER ONE.

ICE AND CLIMATE - UNDERSTANDING THE EARTH CLIMATE SYSTEM.

1.1. AIMS AND OBJECTIVES.

The overall aim of this research is to analyse the dynamics introduced to glacier systems by the presence of iceberg calving, and the impact that these have on the interaction between ice sheets and climate.

Specifically, the aim is to examine the evidence of fluctuations of calving glaciers in Greenland, and their relationship with regional climatic trends. This examination will consist of:

i) Description of recent oscillations of the western outlet glaciers of the Greenland Ice Sheet, and analysis of the controlling variables.

ii) Examination of the mode and mechanisms of Late Glacial and Holocene ice sheet retreat in central West Greenland as inferred from the glacial geological record.

This research permits a critical assessment of the validity of palaeoclimatic inferences drawn from the Holocene and recent patterns of ice marginal behaviour in Greenland. It has implications for the interpretation of the glacial geological record of the oscillations of former ice sheets in other regions. These in turn have relevance for the prediction of the likely response of ice sheets to future climatic change.

1

1.2. SCIENTIFIC BACKGROUND:

Ice Sheets and Climatic Change.

1.2.i. The Earth Climate System.

History is rich in examples of the impact of climatic change on human populations, but only recently has it become apparent that the activities of mankind have the potential to change the climate. As a result, the vicissitudes of climate have begun to loom large in world affairs, and research concerning the patterns and mechanisms of climatic change has escaped the confines of academic journals to become the stuff of headlines. Since the past is the key to the range of future possibilities, a fuller understanding of the controls and interactions which operated on and within the earth's climatic machine in recent geological time is a desirable global objective. Without such understanding, there can be no realistic extrapolation into the future.

The climatic system of the earth is very complicated indeed, functioning with inumerable interactions, feedbacks, and sub-systems (Fig. 1.1). Today, the system comprises four main sub-systems, the oceans, the atmosphere, the biosphere and the ice sheets, but this has not always been so. During much of geological time, large polar ice masses were not a feature of the global environment. However, the history of the earth has been punctuated by periodic ice ages, and we live in the midst of one such period; large ice sheets and extensive sea ice dominate the polar regions and play a significant role in the fluctuations of global climate.
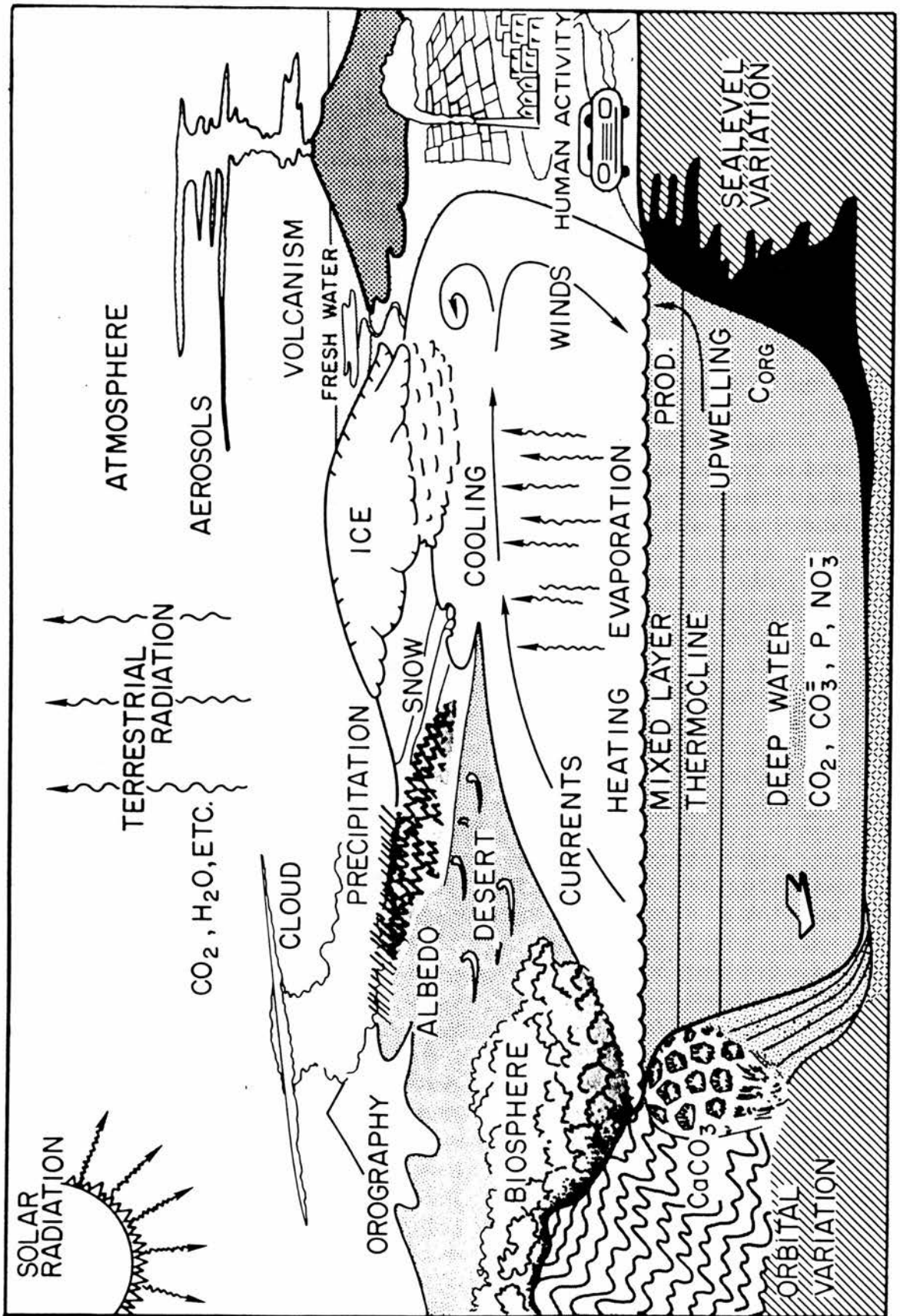
FIGURE 1.1. Major components of the climate system. The main sources of instability are: ice masses, volcanism, variable deep-water production, upwelling, and human activities. Positive feedbacks include: albedo, CO_2, and trace gases. (After Berger and Labeyrie, 1987.)

3

Global climatic change is driven externally by cyclical variation in the earth's orbital patterns, as initially suggested by Croll in 1864 and formalised by Milutin Milankovitch in 1941. Combinations of changes in eccentricity (the shape of the earth's orbit), obliquity (the tilt of the Equator on the earth's elliptical orbit around the sun), and climatic precession (a measure of the earth - sun distance at the summer solstice), cause mathematically predictable variations in the amounts of solar radiation received in certain important latitudinal bands at the top of the atmosphere (Berger and others, 1989). The dominant forcing periods of these three orbital parameters are around 100,000, 41,000, and 23,000/19,000 years respectively, with the latter two having the greatest effect on solar radiation.

The first convincing geochronological evidence to support the Croll-Milankovitch hypothesis was provided by fossil coral terraces (Mesolella and others, 1969), and, subsequently, oxygen-isotope evidence from the foraminifera of ocean-floor sediments clinched the argument (Hays and others, 1976). The changing ratios of oxygen-isotopes are largely controlled by global ice volumes, and these ratios closely reflect the periodicities of the orbital forcing frequencies (Fig. 1.2). Since the acceptance of the oxygen-isotope chronology in the mid-1970s, 'Milankovitch frequencies' have been found in records of environmental change preserved in many environmental records - deep ice cores (eg. Dansgaard and Oeschger, 1989; Lorius and others, 1989,1990; Petit and others, 1990), thick loess deposits (eg. Kukla, 1987; Begét and Hawkins, 1989), changing

4

levels of closed lake basins (eg. Gasse and others, 1989) and in continental pollen-stratigraphic sequencies (eg. Guiot and others, 1989). Recent results from ancient sediments show that the pre-Quaternary climate was also dominated by components with periods of these same frequencies (Berger and others, 1989).
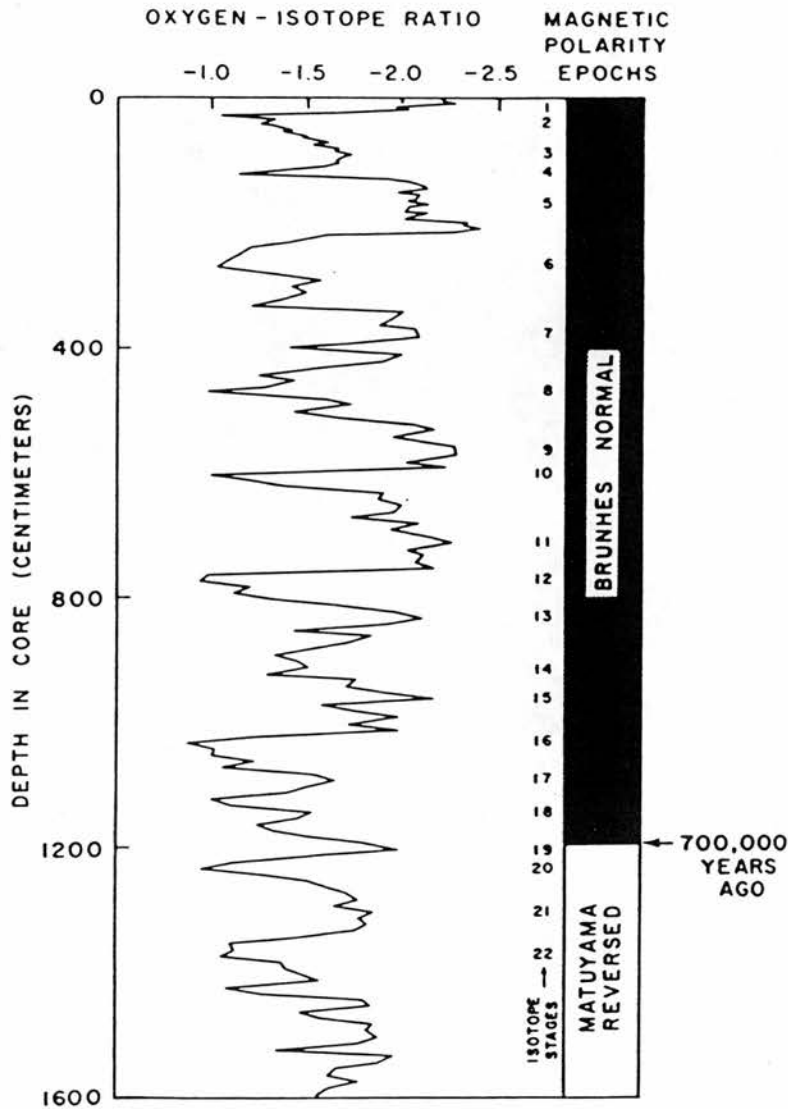
FIGURE 1.2. Record of global temperature oscillations as revealed by deep-ocean oxygen-isotope ratios in Pacific core V28-238. These isotopic and magnetic observations provided the first accurate chronology of late Pleistocene climate. (After Imbrie and Imbrie, 1979; data from Shackleton and Opdyke, 1973.)

5

The Quaternary has been a period of great, repetitive, and often rapid environmental change, consisting of the growth and decay of large ice sheets, changes in sea levels, alterations in the patterns of ocean currents and weather systems, and migration of climatic zones and their associated biota (CLIMAP members, 1976; COHMAP members, 1988). The climatic deterioration which culminated in the present ice age began around 3.15 Ma (million years ago) in the northern hemisphere, but ice sheets large enough to send quantities of ice-rafted debris into the northern oceans did not exist before 2.4 Ma (Ruddiman and Raymo, 1988). Between 2.4 and 0.9 Ma, there were a sequence of about 40 climatic cycles at the 41,000-year rhythm of orbital obliquity, but during the mid-Pleistocene the global climate began to respond progressively and in a broadly synchronous manner to the 100,000-year eccentricity cycle, and this came to dominate the climatic response after 0.45 Ma; numerous signals at 41,000 and 23,000/19,000 are superimposed on this dominant frequency (Fig. 1.2).

The Earth Climate System is a non-linear dynamical system in which variations can occur without direct forcing (Geller, 1989; James and James, 1989), as shown by the fact that gradual, monotonic astronomical forcing has produced abrupt, non-monotonic climatic changes (Overpeck and others, 1989). As such, it is a prime candidate for the application of Chaos Theory (Gleick, 1988), because stability thresholds, strong feedbacks, and great sensitivity to initial conditions are its important dynamic characteristics. If, therefore, we are to gain a fuller understanding of the responses of the

system, there is a need to investigate these dynamic interactions, feedbacks, and lags that operate to produce environmental change. This need is urgent because of the extent to which human activities are affecting the earth's surface, and the great uncertainties concerning the likely nature, speed, and magnitude of the response of the earth's climate to these changes.

There are three broad methods of investigating the functioning of the Earth Climate System: studies of today's systems, palaeo-environmental research, and numerical modelling.

i). Contemporary Empirical Studies.

Measurements of the current characteristics of the climate system, and the monitoring of change (eg. Zwally, 1990), provide a baseline against which to measure future change and to assess past fluctuations (eg. Raval and Ramanathan, 1989; Meier, 1990). Research into the empirical relationships between various parts of the system furnishes us with the only available analogue for dynamic interaction in the past, and is essential in order to provide accurate information for climatic modelling. Uncertainties about the physical nature of these relationships undermine the confidence that can be placed on model results (Slingo, 1989). The fundamental drawback of empirical studies of the Earth Climate System is that the scale of investigation is necessarily much smaller than the scale of operation of the climatic components, both temporally and spatially. For instance, ice sheets cover millions of square kilometres and operate with lag times and feedbacks over centuries and

millenia; satellite monitoring has now partially solved the spatial issue (eg. Zwally and others, 1983; Zwally, 1990), but there can be no empirical solution to the temporal problem. In order to make research feasible, it is necessary to focus on isolated parts of the whole, and inevitably this can only produce partial answers in the context of a strongly coupled system. Nevertheless, detailed understanding of the mechanisms effecting that coupling is a prerequisite for comprehension of the wider system.

ii). Global records of past environmental change.

Taken together with contemporary empirical studies, records of the past fluctuations of the earth's climate are the only basis on which to predict the likely response of the system to future natural and anthropogenic forcing. Direct evidence of repeated ice sheet oscillations is rare because ice sheets advance over the evidence of preceding cycles, destroying much of it in the process. The most complete records of Quaternary ice sheet volumes and global sea surface temperatures are the proxy records obtainable from the changing oxygen-isotope ratios and species composition of foraminifera in deep ocean sediments. In addition, a wealth of information about the most recent glacial period can be obtained from the glacial geomorphological record, glacio-lacustrine and glacio-marine sediment sequences, and from the variable abundance of iceberg dropstones in the deeper oceans. Detailed reconstructions of palaeoclimatic conditions in a particular region can be made from pollen sequenc es, loess deposits, and from fossil beetle

8

assemblages (eg. Pons and others, 1987; Kukla, 1987; Coope, 1987).

Palaeo-environmental information has also been preserved within the existing ice sheets themselves (Oeschger and Langway, 1989; Dansgaard and others, 1982; Lorius and others, 1989,1990; Chappellaz and others, 1990), the basal ice of which may be as much as 500,000 years old (Dansgaard, 1987). Isotopic and chemical analysis of the layered cores of ice yields information about changing patterns of temperature, precipitation, and storminess (Dansgaard and Oeschger, 1989; Koerner and Fisher, 1990a). For instance, recent analysis of the Dye 3 ice core in South Greenland has shown that, at 10,700 years ago, a temperature rise of 7°C, which represents half the total Pleistocene - Holocene warming in this region, took place in less than 50 years and perhaps in as little as 20 years (Dansgaard and others, 1989). From such data, inferences can be drawn concerning environmental change, the overall geometry of former ice sheets, the timings of their maxima and readvance stages, the dynamics of different sectors of the margin, and the mechanisms of deglaciation.

All these data sources are afflicted by uncertainties of interpretation. For example, the interpretation of ice cores is problematic because the various parameters are linked to climate in different ways, and because it is difficult to separate the direct isotopic temperature signal from the indirect effects of source water variability (Koerner and Fisher, 1985; Koerner, 1988; Fairbanks, 1989). Major differences of opinion exist over

9

even some of the most recent climatic changes. Thus, for instance, the explanation of the onset of the Younger Dryas cold episode around 11,000 - 10,000 BP is still a vexed question (Street-Perrott and Perrott, 1990; Rooth, 1990). Broecker and others (1989) and Broecker and Denton (1989) have developed an elegant mechanism for triggering this temperature reversal, namely the shutting down of North Atlantic Deep Water formation by massive influxes of Laurentide meltwater diverted from the Mississippi down the St. Lawrence by ice retreat. But Fairbanks (1989) and Shackleton (1989) dispute the effectiveness of this thermohaline mechanism, showing that the Younger Dryas was a time of minimum meltwater input and was straddled by low salinity events.

A further problem which dogs attempts to understand the linkages within Quaternary climatic change is the dating issue. Lowell and others (1990) have shown that inaccuracies within the radiocarbon method can lead to a 'radiocarbon stratigraphy' more complex than the ice margin history, because the error ranges of radiocarbon dates are often greater than the events they are meant to 'date'. Similarly, Bard and others (1990) have calibrated ^{14}C ages over the past 30,000 years using U-Th ages obtained from Barbados corals and have revealed substantial variability in the radiocarbon chronometer; their adjusted timescale shifts the date of the Last Glacial Maximum back from the accepted 18 ka BP to between 21 and 22 ka BP. Without detailed and reliable chronologies, it is impossible to resolve important questions concerning the mechanisms driving climatic change and which

factors led or lagged each other.

However, despite such conflicts of interpretation and dating, the broad cycles of Quaternary change are now known, even if the mechanisms that operated to regulate environmental change are still a matter for lively debate.

iii). Numerical Modelling.

Computer simulation of the operation of these large, complex natural systems over long time periods has several advantages. Models can deal with the system in question as an integrated whole, rather than having to break it down into its constituent parts. Long time periods can be modelled rapidly (eg. Reeh, 1984). Some of the infinite complexity of the natural world can be rationalised into limited numbers of variables, making it theoretically possible to isolate the important interactions. Experiments of the 'what if?' kind can be carried out on a computer which would be impossible in reality (eg. Huybrechts, 1990; Hulton, unpublished). In recent years, atmospheric General Circulation Models (GCMs), coupled to simplified models of the oceans and the ice sheets, have been used to conduct experiments simulating and predicting climatic change, with a particular, topical focus on the likely results of a doubling in levels of atmospheric CO_2 (eg. Manabe and Broccoli, 1985; Broccoli and Manabe, 1987; Muszynski and Birchfield, 1987; Sperber and others, 1987; Roeckner and others, 1987; Mitchell and others, 1989; Saltzman and Maasch, 1990), and the results have greatly furthered our understanding of many aspects of the climate system and how its components interact. For instance, Pollard's model experiments (1982,1983,1984) concerning the

11

growth and decay of ice sheets have shown that, if a good approximation to the oxygen-isotope record of ice sheet volume is to be obtained, expressions for delayed isostatic sinking and iceberg calving need to be explicitly contained in the model.

However, if modelling experiments are to further our understanding rather than our confusion, they must be informed by and tested against empirical reality. Model outputs are only as good as model inputs, and the ways in which variables are modelled and parameterized inevitably depends on our perception of the importance and functioning of those factors in the real world. The extent to which modelled reality represents the real world is sometimes difficult to assess. Furthermore, there is the issue of resolution. Thresholds and triggers are of central importance within the dynamics of chaotic systems, and in order to model these, both temporal and spatial resolution needs to be of an order appropriate to the scale of investigation. This is especially true, for instance, of the behaviour of the margins of ice sheets where topographic factors can provide thresholds of instability (Payne and others, 1989; Payne and Sugden, 1990b). Physical and practical constraints on modelling procedures, however, mean that there are necessarily many trade-offs between what would be ideal and what is practically feasible. Furthermore, although the main physical mechanisms in the climate system are moderately well understood, the presence of many feedback loops means that small errors in a deterministic model system are likely to grow in unpredictable ways, such that even

12

greatly improved models will never provide entirely reliable predictions (Berger and Labeyrie, 1987). In summary, then, modelling serves to highlight areas of potential significance in empirical research, while the latter feeds back into and constrains the former.

The emerging picture of global environmental change during the Quaternary has become more sharply focused as information from these three sources has accumulated, but many uncertainties remain. The Earth Climate System is thus analagous to a 'black box' system: the external controls (astronomical forcing) have been identified, and the 'output' - global environmental change - is broadly known, but the dynamic interactions within the system that produce the latter from the former are only understood to a limited degree.

1.2.ii. Glacier Response to Climatic Change.

1.2.ii. a) Mass Balance Relations.

The most informative measure of the current 'health' of a glacier is its mass balance. A positive net mass balance indicates potential for growth, while a sustained negative net mass balance results in shrinkage. All ice masses are essentially hydrological conveyor belts, receiving moisture in various forms and discharging it back into the hydrological cycle after delays ranging from months to hundreds of thousands of years. Glaciers can gain and lose mass at any point, but are conventionally divided into areas of annual net accumulation and ablation, divided at the position where accumulation and ablation balance - the Equilibrium Line Altitude (ELA). This input/output concept is theoretically simple but complex in reality (Letréguilly and Reynaud, 1989). For instance, the ELA concept is hard to apply to glaciers with a small elevation range, or to ice sheet outlets due to the problems of delineating the accumulation area (Braithwaite, 1984). A glacier's state of health can also be expressed in terms of the Accumulation Area Ratio (AAR), which is a measure of the relative surface areas on a glacier which lie above and below the ELA. For a glacier in equilibrium, the AAR is commonly about 0.7 (Paterson, 1981). The AAR is a useful measure, despite the wide variation in the values of equilibrium AAR's, and the difficulty of applying it in situations where catchments are not well defined, such as in ice-bounded catchments.

Measures of mass balance can be obtained directly, or through the use of repeat photogrammetry, hydrological

methods (Tangborn and others, 1975), reconnaissance aerial photography (Meier and Post, 1962) or satellite imagery (Hall and Ormsby, 1983). In such ways it is possible to gain a moderately accurate picture of well-defined and relatively small glaciers and ice caps, but the sheer size and remoteness of the world's big ice masses currently defy such accurate diagnosis. At present, we do not even know whether the Antarctic and Greenland Ice Sheets are growing or shrinking (Zwally, 1990; Reeh, 1985,1989; Reeh and Gundestrup, 1985; Doake, 1985), although it is thought that the former is slowly growing (Meier, 1990) and that the latter is in net equilibrium (Kosteka and Whillans, 1988; Ambach, 1989).

The main problem in understanding the behaviour of glaciers is the detailing of the linkages between climate and glacier response. If these can be accurately defined, then, in theory, it should be possible to deduce the inverse from the glacial geological record - namely, the climatic input which produced a certain glacial response. Clearly, there must be a strong relationship between mass balance and local climate, since the climate in the accumulation area determines the amount of snow or ice available, and surface melting on the glacier is controlled by the prevailing meteorological conditions. Recent work by Letréguilly and Reynaud (1989), for instance, has shown that, although different glaciers show very different mass balances for a given year, the variation with time, centered around the mean, is similar for all glaciers in a region. They conclude that 'the mass balances of mountain glaciers can really be

used as an annual climatological indicator' (p.168).

The broad fluctuations of glaciers do indeed reflect climatic change, both on geological and contemporary timescales, as shown recently, for example, by the global pattern of glacier advance during the Little Ice Age (Grove, 1988), and it is widely held that glacier fluctuations are good indicators of such change (Porter, 1981; Pelto and others, 1990; Johannesson and others, 1989). However, response of a glacier to a change in mass balance is not simple (Nye, 1960; Pelto, 1987). The ways that such changes are expressed at the terminus therefore vary considerably both in space and time. For instance, in their extensive survey of 960 North American glaciers, Meier and Post (1962) found that there were, in all regions, examples of advancing, stationary, and retreating glacier fronts. Some of this difference can be explained by local climatic contrasts, and some of it by dissimilarities between glacier characteristics such as size (Röthlisberger and Lang, 1987), steepness, velocity, and area-height distribution (Tangborn and others, 1990).

A further complicating factor is variable mass balance history. Glaciers and ice sheets may adopt multiple steady states under a given climatic regime (Alley, 1990; Oerlemans, 1989; Letréguilly and Oerlemans, 1990); this introduces the possiblity of bifurcation, whereby small climatic perturbation may lead to a large ice sheet response during the switch between states (Payne and Sugden, 1990a). Such factors mean that glaciers react in varying ways and with contrasting response times to specific changes in the

16

climate. 'The precise relationship between climatic change and glacier flow response is still incompletely understood' (Brugman, 1990, p.332).

In investigating the relationship of existing glaciers with climate, some of the apparently conflicting evidence of the frontal oscillations can be resolved through studies of the mass balances of those glaciers. However, if it is past climates that are the focus of interest, then the only available glacial evidence is depositional. The drawing of inferences about former climates from the apparent palaeoclimatic implications of glacial deposits is a process fraught with difficulty because, in addition to climatic controls, there are several specific non-climatic influences on glaciers which may result in contrasting magnitudes and timings of glacier extent (Fig. 1.3). These may operate singly or in combination to obfuscate the truth, and they must all be taken into account if the inferred former ice extent is to be used as a palaeoclimatic tool. Tangborn and others (1990) comment dryly:

> Once we understand the response of glaciers to influences other than climatic change, perhaps we can determine how glaciers respond to changes in climate. (p.278)

While in reality there is a continuum linking all these cause-and-effect relationships, such that boundaries drawn between them are necessarily artificial, the classification into primarily glacio-dynamic and primarily topographic controls shown in Fig. 1.3 is provided as a helpful visualisation and is followed below.
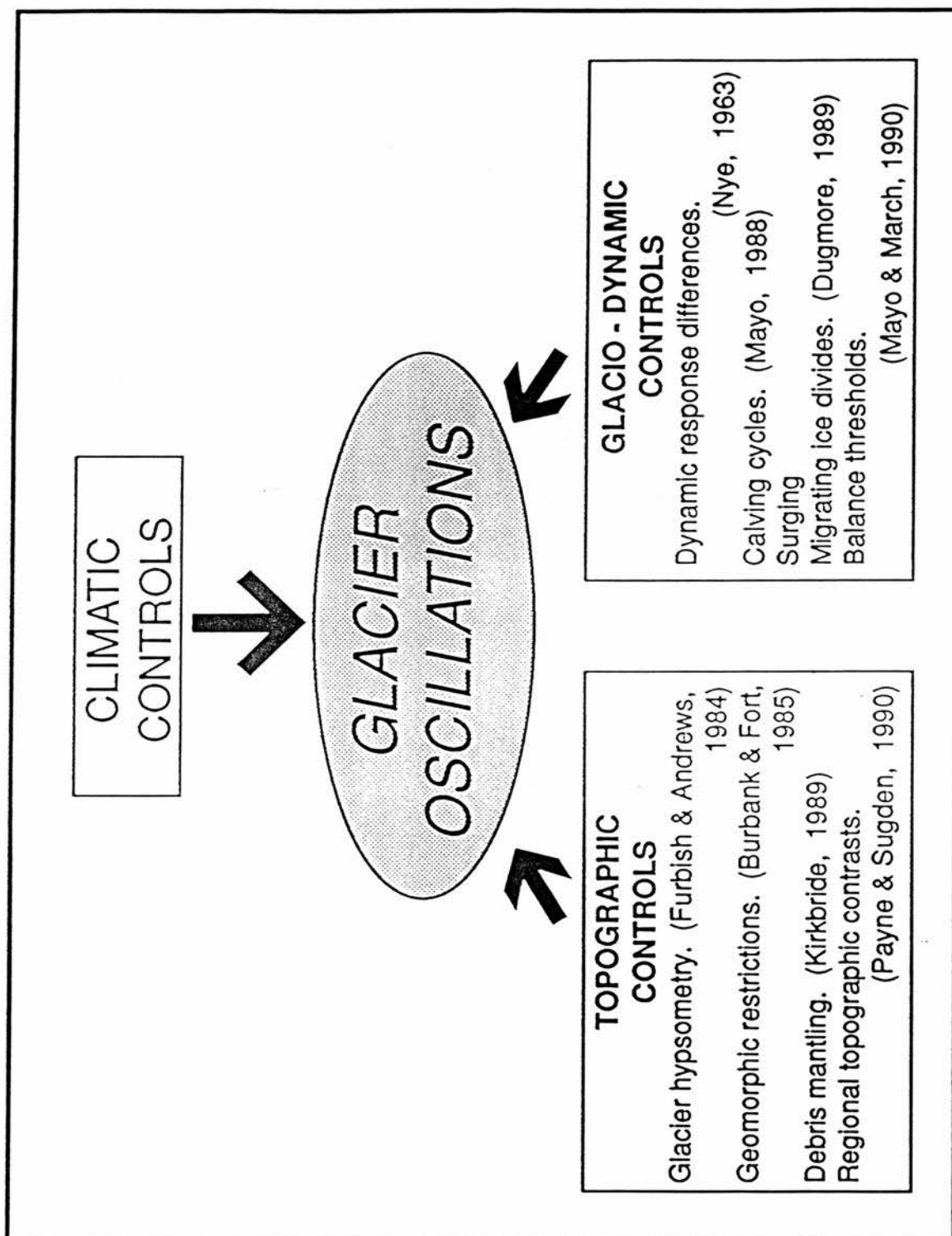
FIGURE 1.3. Some causes of asynchronous glacier behaviour.

1.2.ii. b) Causes of Asynchronous Glacier Behaviour.

i) Glacio-dynamic controls.

- Dynamic response differences: Starting with the premise that 'glaciers are extremely sensitive indicators of climate, for a very slight climatic variation is sufficient to cause a considerable advance or retreat of the ice' (Nye, 1960, p.559), Nye applied mathematical kinematic wave theory to glaciers. A small change in accumulation rate can lead to large thickening of the lower parts of glaciers through the propagation of kinematic waves travelling at 2 - 5 times the ice velocity. Great contrasts in kinematic wave generation, wave propagation, and wave diffusion rates exist between glaciers (Nye, 1963). As a result, the frontal expressions of a particular change in accumulation can vary significantly, both in timing and magnitude.

- Surging behaviour: Surging glaciers seem to fluctuate in response to internal glaciodynamic mechanisms rather than climatic inputs. Periodic surging of glaciers has excited interest for many years (eg. Tarr and Martin, 1914; Meier and Post, 1969). Work on Variegated Glacier in Alaska, which surged in 1982/1983, (Kamb and others, 1985; Raymond and others, 1987) demonstrated that almost all the motion was achieved by sliding, and that restructuring of the subglacial hydrological system accounted for the dramatic increases in velocity (Raymond, 1987). However, many questions remain unanswered, such as the nature of the initial triggering mechanism (Sharp, 1988a). Recognition of the geomorphic results of surges (Sharp, 1988b) is important because of the danger of making false palaeoclimatic inferences from the

19

glacial geological record; large parts of the late Pleistocene ice sheets may have exhibited surging behaviour (Clayton and others, 1985). Rapid and substantial glacial fluctuations of this sort can occur for reasons unrelated to climate.

- **Migrating ice divides:** When the drainage basins of glaciers are within an ice cap or an ice sheet, it is possible for relative change to take place in the size of neighbouring catchments (Waddington and Marriott, 1986; Clarke, 1987). Conceivably, glaciers may 'compete' for ice supply such that relative discharge rates determine the size and shape of drainage basins, rather than vice versa (Bindschadler, 1984). Ice divide migration may result in advance and retreat behaviour of outlet glaciers which is anomalous relative to the climatic forcing. For example, the Holocene fluctuations of Solheimajökull in Iceland have been out of phase with other outlets of the Myrdalsjökull icecap because the ice divide has been migrating across a subglacial topographic divide, changing the catchment volumes (Dugmore, 1987,1989).

- **Balance thresholds:** A much-debated question is whether glaciers advance becuse of cooling and reduced melting, or as a result of warming and increased precipitation. Mayo and Trabant (1984) gathered climatic and glaciological statistics at Wolverine Glacier between 1966 and 1982, and developed a balance model which uses temperature as the only independent variable (Fig. 1.4). It suggests that the response of glaciers to climate can be considerably more complex than previously imagined, and that the apparent antithesis in the

20

above question may in fact be misleading; an increase of temperature can lead either to a positive-balance response or to a negative-balance response, depending on the glacier's current position on the 'balance response curve'. Similarly, stability can occur under three widely different mean annual temperature regimes. They conclude:

> The analysis suggests that even large ice sheets such as Greenland and Antarctica may respond in surprisingly complex ways to rises in global air temperature. We suggest that general global warming could produce a worldwide mixture of glacier growth and glacier thinning with non-synchronous advances and retreats rather than a simple worldwide recession. (p.123)
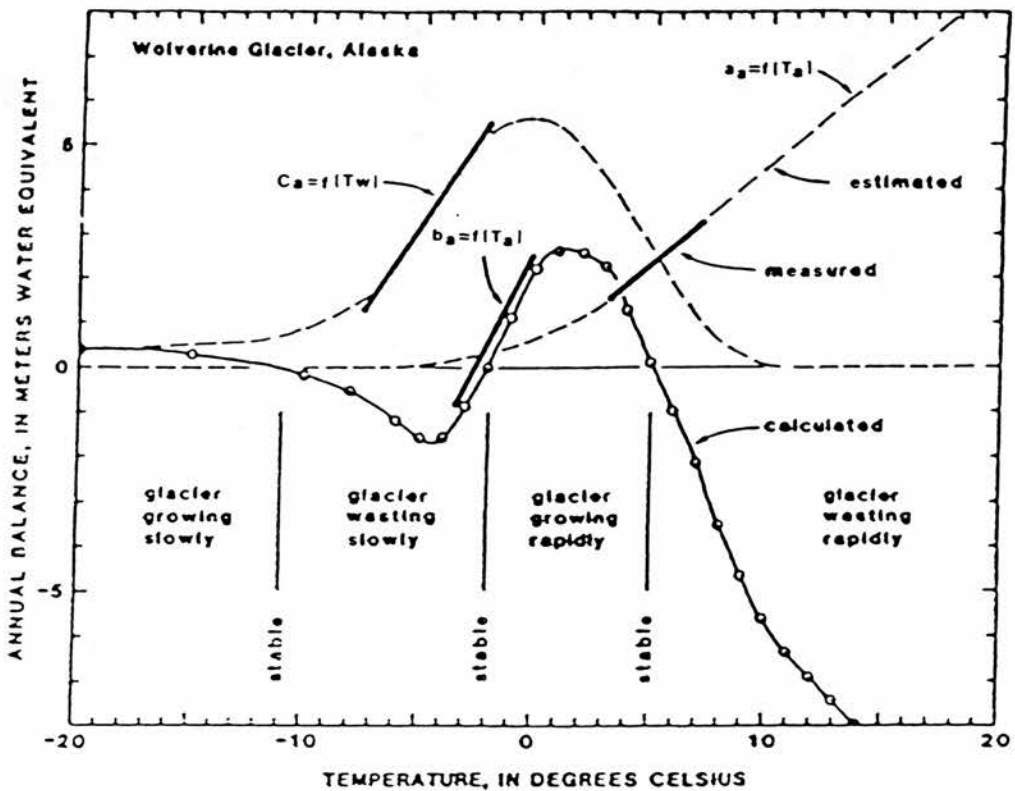
FIGURE 1.4. Response of Wolverine Glacier, Alaska, to changes in temperature and precipitation. Annual accumulation, c_a, as a function of winter temperature; annual ablation, a_a, as a function of summer temperature; and annual balance, b_a, as a function of annual temperature both measured and estimated for drastically different climatic conditions. (After Mayo and Trabant, 1984.)

21

Subsequent work suggests that the primary response of the ice sheets to global warming will be glacier growth (Mayo and March, 1990; Warren and Frankenstein, 1990; Koerner and Fisher, 1990b). Recent work by Dyke (1990) in the Canadian Arctic has also highlighted the importance of precipitation-controlled periods of end moraine formation at times of high temperatures during the Holocene.

- Tidewater glaciers: When glaciers terminate in tidewater or in large lakes, so that iceberg calving becomes an important ablation mechanism, it is possible for glacier fronts to oscillate substantially without significant climatic change, or even in ways opposed to the prevailing climatic forcing (Meier and Post, 1987; Mayo, 1988; 2.3.ii.). The potential for such anomalous behaviour was first suggested theoretically by Mercer (1961) and has since been confirmed empirically by the extensive work of the U.S. Geological Survey in Alaska (eg. Meier and others, 1980). As a result of iceberg calving dynamics, the stillstand locations adopted by calving glaciers as they advance and retreat within a fjord system are therefore often determined more by topographic than climatic controls.

ii) Topographic and Geological Controls.

- Differences of glacier hypsometry: Furbish and Andrews (1984) examine the way in which long-term glacier response is affected by its altitudinal distribution (Fig. 1.5). Glacier geometry determines the effect of a shift in the ELA by controlling the relative area of the glacier which is affected (Kick, 1989; Oerlemans, 1989). Thus if the surface area over which the ELA moves is large, the terminus response
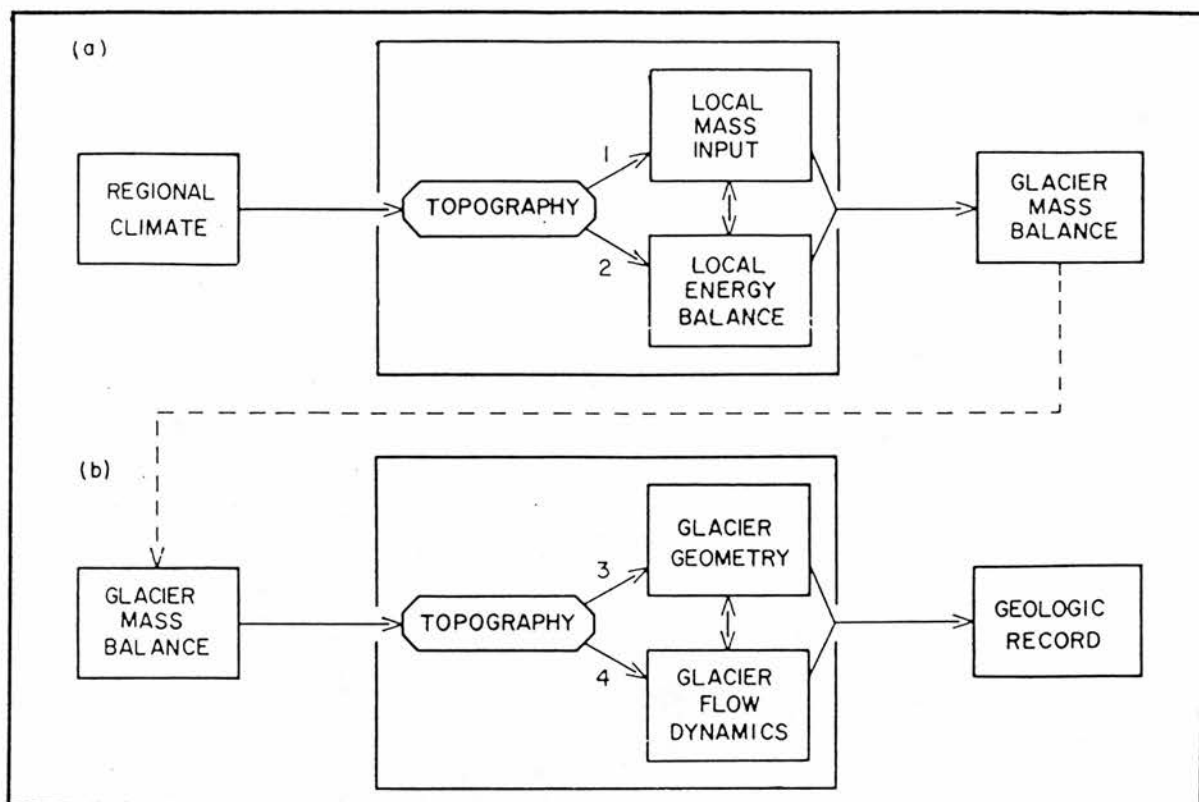
FIGURE 1.5. (a) Generalized linkage between regional climate and glacier mass balance. (b) Linkage between mass balance and glacier response. (After Furbish and Andrews, 1984.)

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will be great, and <u>vice versa.</u> Striking exemplification of this is provided by Miller (unpublished) in his work on the 'foot' and 'potbelly' glaciers of Baffin Island; glaciers with extensive surface area in their mid-sections exhibited a much greater terminal response to a particular shift in the ELA than those with narrow mid-sections, the latter having nested moraines while the former have widely spaced Late Glacial and Neoglacial moraines. The topographic setting is therefore of great importance (Tangborn and others, 1990; Dugmore, 1989; Pelto, 1990) since it modifies local climate (and hence mass balance), limits glacier geometry, and affects ice-flow dynamics.

23

- **Geomorphic restrictions on glacial advance:** Physical barriers to ice advance may lead to contrasting magnitudes of glacier response. For instance, Luckman and Osborn (1979) found that exceptions to the general regional pattern of post-Wisconsinan glacier fluctuations in the Canadian Rockies exist in places where rock glaciers or large moraines from earlier advances blocked the ice, suggesting that the relative size and number of moraines is a function of the geomorphic, geological, and topographic setting rather than of differing glacier behaviour. Similarly, Burbank and Fort (1985) found that the downvalley limit of glaciation in the North-Western Himalaya had been decoupled from palaeoclimate by impenetrable buttresses of vertically dipping bedrock strata. Thus the interpretation of former glacier front positions in terrain with complex topography is hard, particularly in places with reverse slopes where hysteresis can occur (Oerlemans, 1989).

- **Debris-mantling:** Recent work on the snout of the Tasman Glacier in South Island, New Zealand, has demonstrated that the insulating effects of a thick mantle of supraglacial debris can reverse the ablation gradients that would exist on an unmantled glacier, and that this can result in anomalous behaviour relative to debris-free glaciers nearby (Kirkbride, 1989). The snout of the Tasman Glacier has not moved this century, while neighbouring glaciers have retreated, and it is currently about 5 km more advanced than it would be if it had had no debris cover. Kick (1989) appeals to debris-mantling as one means of explaining contrasting timings of neighbouring glacier extensions in High Asia.

24

<u>- **Regional topographic contrasts:**</u> Contrasting magnitudes and timings of glacial activity between regions can result from topographic differences, both in the height and arrangement of mountains. In the Andes, for example, mountain heights south of 40° South average 2000 m, whereas north of that parallel, mean heights rise to around 4000 m. It is therefore likely that ice masses would survive in the north during interstadials, but might disappear completely in the south, and thus a return to cooler conditions could be reflected glacially in the north much earlier than in the south where glaciers would have to form from bare ground (Clapperton, 1988). The spatial arrangement of a mountain chain can determine the overall scale of glacial response to particular climatic forcing. For instance, in the North West Highlands of Scotland, during the Loch Lomond advance, ice flowed down either side of the main moutain chain and ablated at lower altitudes; in the Rannoch Moor area, however, the ice flowed down into a basin, filling it up and setting up a positive feedback loop which led to a substantially greater advance than further north (Payne and Sugden, 1990b).

The theoretically simple response of glaciers to climatic change is thus complicated by a number of almost infinitely variable topographic, geological, historical and dynamic contrasts, and this substantially obfuscates our ability to make simple inferences about past climates from the evidences of former glacier oscillations. Problems of reliable dating further complicate the issue (Porter, 1981). For palaeoclimatic reconstruction, therefore, the temporal resolution of the glacial geological record is important

25

(Mann, 1986) because some of the complexity that is observed in the responses of contemporary glaciers is damped by coarse resolution so that the broad patterns of palaeoclimate can be determined. Clearly this becomes increasingly true as the resolution becomes progressively coarser further back in time.

All of these complicating influences have been shown to be important in the interpretation of the geological records of valley glaciers and small icecaps, but the extent to which they affect ice sheet outlet glaciers remains largely unknown. Arguably, the role of iceberg calving in affecting the response of such outlet glaciers is the most significant of these unknowns to address because it has the greatest potential, both temporally and spatially, for decoupling glaciers from climate and is known to have affected large parts of the margins of the retreating Pleistocene ice sheets.

1.2.iii. The importance of iceberg calving.

Iceberg calving margins are the only place on the earth's surface where there is the potential for specific site interaction between the three main elements of the Earth Climate System (Fig. 1.6). The dynamics of calving fronts can be an important catalyst of change in the cryosphere and the hydrosphere, and indirectly are of great significance in effecting atmospheric change. For instance, rapid break-up of a marine ice sheet would flood the neighbouring oceans with icebergs, thereby lowering the sea surface temperatures and promoting the formation of sea ice, which in turn might significantly change the regional albedo; such changes would affect the temperature and humidity of the atmosphere, altering pressure and precipitation patterns, and this would then feed back into the ice sheet system.

Ice sheets have a rich spectrum of feedback loops (Oerlemans, 1989) and consequently are prime candidates for explaining some of the remaining puzzles about the non-linear response of the global climate to orbital forcing (Muszynski and Birchfield, 1987). There are five main reasons for this:

1. They obey a non-linear flow law.

2. They interact with the lithosphere on long timescales.

3. They have long lag times, their volume-response times being very slow.

4. They exist in meta-stable equilibrium (cf. Hughes, 1987), able to remain dynamically stable within certain limits, but exhibiting a 'runaway' response once certain key thresholds are exceeded (Alley, 1990).

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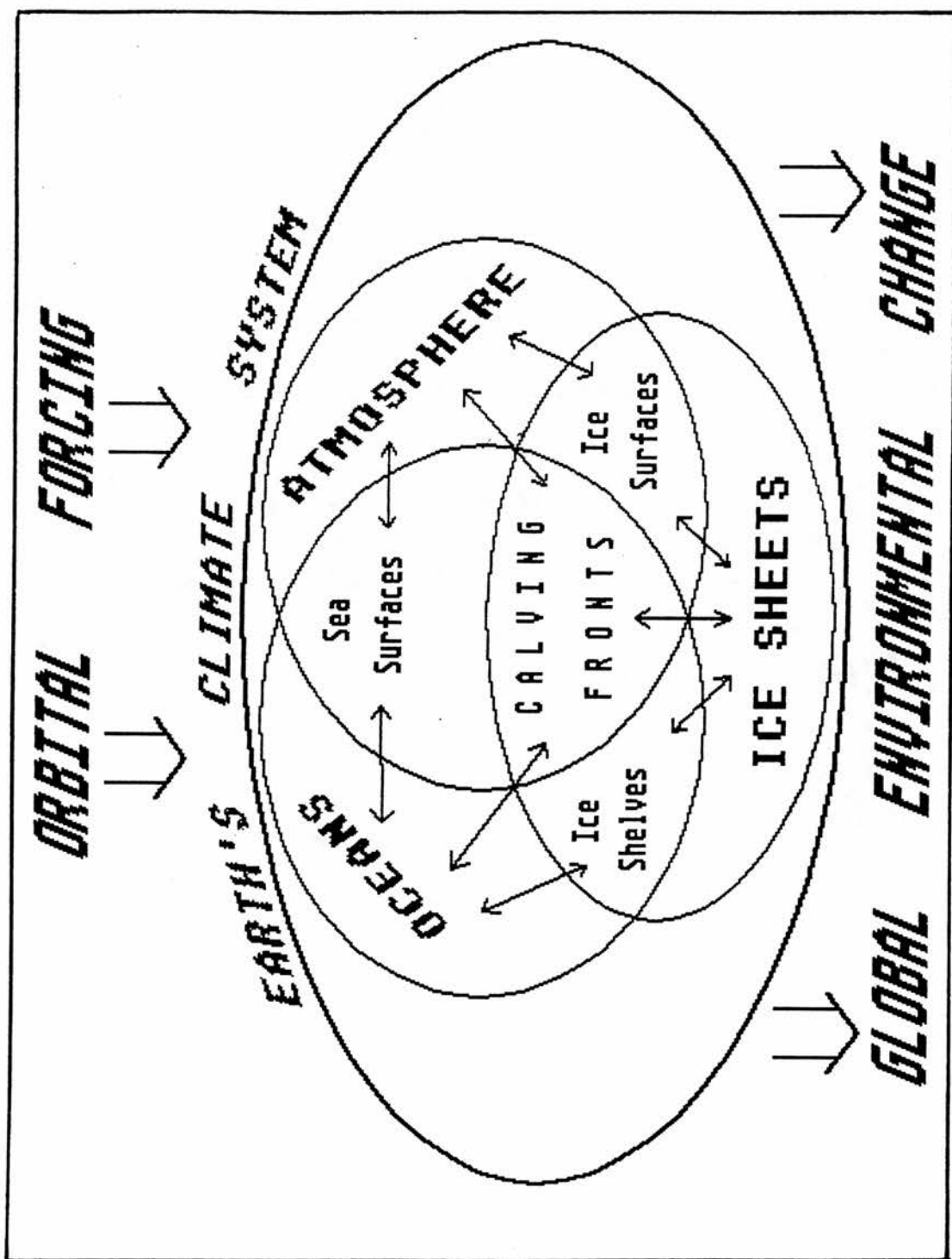
FIGURE 1.6. Locations of potential interaction between components of the Earth Climate System.

28

5. There is inherent asymmetry in the timescales of growth and decay: the build-up of continental ice sheets takes tens of millennia, but iceberg calving permits very rapid collapse at rates an order of magnitude faster.

This last is of considerable importance. The record of global ice sheet volume shows a consistent pattern of slow, oscillatory build-up to a short-lived maximum, followed by irreversible, catastrophic collapse or 'termination' (Fig. 1.2). Some mechanism which only comes into operation with maximum effectiveness at the times of greatest glacial expansion needs to be invoked to explain this pattern; iceberg calving conforms to this requirement. At their maxima, ice sheets in both hemispheres were uniquely vulnerable to this process because they had extensive marine and lacustrine margins and were situated over isostatically depressed beds, such that retreat occurred down a reverse slope, increasing water depths at the margin (Pollard, 1984). It is unlikely that iceberg calving on its own provides a full explanation; other important ablationary factors such as glacio-isostasy (Peltier, 1987) and fluctuations of the concentrations of greenhouse gases in the atmosphere (Lorius and others, 1990) probably worked in tandem as well. However, it is likely that a combination of marine calving and calving into large proglacial lakes (Andrews, 1973,1984,1987; Teller, 1987) were key components in the collapse of the large Quaternary ice sheets (Mix and Ruddiman, 1985; Ruddiman, 1987b).

In addition, the sensitivity of calving margins to relative sea levels (Thomas 1977,1979; Brown and others,

29

1982) may partly explain the curious fact that ice age events have been globally synchronous, despite asynchronous orbital forcing. Summer melting of Northern Hemisphere ice sheets due to favourable orbital configuration would raise sea levels eustatically, and this in turn would destabilise the marine components of ice sheets worldwide (Denton and Hughes, 1981,1983; Denton and others, 1986). Positive feedbacks involving dynamic responses and albedo changes could rapidly amplify the initial marginal melting, leading to rapid collapse of the ice sheets.

This ice sheet - climate linkage has recently been challenged by Broecker and Denton (1989) who propose that only major reorganisations of the ocean - atmosphere system can account for the global synchroneity and rapidity of the last termination around 14 ka BP. Accepted concepts of ice sheet dynamics and atmosphere - ice sheet coupling have also been challenged by Boulton and Clark (1990) who postulate a highly mobile Laurentide Ice Sheet with rapidly shifting central domes.

Nevertheless, despite the emergence of these new theoretical concepts, iceberg calving remains a critically sensitive mechanism, the operation of which represents an important non-linear climate-system response to the linear astronomical forcing. Thus there is 'considerable evidence that marine ice sheets may be critical cryospheric components on short as well as long timescales of climatic variability' (Muzsynski and Birchfield, 1987, p.5).

Recent concern about anthropogenic global warming provides a further, more immediate incentive for

understanding the dynamics of iceberg calving. Warming may, over a period of centuries, lead to greater runoff from the ice sheets due to increased melting, raising world sea levels and flooding cities. However, there are large and unknown lag times involved in this linkage (Meier, 1990), and in the shorter term it is expected that increased precipitation will cause a thickening of the ice sheets (Warren and Frankenstein, 1990). Iceberg calving is an important uncertainty in these equations. The Greenland and Antarctic Ice Sheets lose 50% and 90% of their _annual_ mass _loss_ respectively through calving. Total iceberg fluxes to the oceans are poorly known, and the response of the calving margins to a warmer, wetter climate is hard to predict. Warming may not lead to greater mass loss because increased melting in the calving sectors may be offset by reduced ice supply to the calving fronts (Braithwaite and Olesen, 1990); the iceberg discharge may therefore be perturbed by the changing rheological regime (Ambach, 1989).

From geological, glaciological, palaeoclimatic, and environmental points of view, therefore, iceberg calving is a significant problem that merits considerable attention.

1.3. APPROACH.

Modelling has highlighted the importance of iceberg calving within the climatic system, and contemporary studies show that calving glaciers have a unique relationship with climate. This research aims to contribute to our understanding of calving by adopting a dual approach, a combination of a survey of recent glacier fluctuations throughout West Greenland (Chapter 3), and a case study of early Holocene retreat behaviour in central West Greenland (Chapter 4).

1.3.i. Greenland - A Unique Analogue.

Greenland is an ideal area in which to test ideas about the role of calving dynamics in controlling the ways in which ice sheets interact with the oceans and the atmosphere. There are at least five reasons for this:

1. It is the only Northern Hemisphere ice sheet today, and, as such, it represents the best analogue for the former ice sheets around the North Atlantic region. Its continued existence is an anomaly, not only because other ice sheets in this hemisphere vanished at the end of the last Glacial around 10,000 years ago, but also because, if it melted away, it would not reform under present climatic conditions.

2. Iceberg calving is of great significance within the overall dynamics of the ice sheet, accounting for about 50% of total ablation; this consists of an estimated annual calving flux of 310 km^3 (Reeh, 1985), although estimates vary widely (Oerlemans, 1990, in press).

3. Unlike the Antarctic Ice Sheet which has an almost exclusively marine margin, the outlet glaciers of the Greenland Ice Sheet terminate in a wide range of environments spanning 22° of latitude. There are large numbers of tidewater-calving, lake-calving, and land-terminating outlet glaciers, together with extensive sectors of marine and terrestrial margin. This offers great comparative advantage for research into the contemporary dynamics of the ice sheet margin.

4. The fourth attribute of Greenland is the nature of the Late Glacial and early Holocene retreat of the ice sheet. At the peak of the most recent glaciation, the ice sheet had an exclusively marine margin and terminated many tens of kilometres beyond the present coastline. As it retreated, it passed through a series of intermediate conditions between fully marine and (in central West Greenland) fully terrestrial. Subsequent falls in relative sea level due to isostatic recovery, and generally good preservation of sediments, has left a relatively complete glacial geomorphological record of this transition phase.

5. The last advantage of Greenland for this kind of research is simply the very practical one of accessibility.

1.3.ii. Previous Glacial Research in Greenland.

Eirik the Red landed in Greenland in the year 874 AD, and purposefully gave this ice-dominated land its misleadingly attractive name in order to encourage other settlers. His ploy succeeded, and for more than 500 years a Norse community coexisted with the Greenlandic Inuit population along the shores of the south-western fjords. They then abandoned their settlements, either due to climatic deterioration, political change in Europe, or a combination (Sugden, 1982). Europeans appeared off the coasts of Greenland once again in the seventeenth century when whalers, travellers, and missionaries such as Hans Egede began to explore the country. Many of these pioneers such as Davis, Rink, Nordenskiold, and Rasmussen gave their names to places and regions, so that European and Inuit names jostle with each other on today's official maps.

One of the chief causes of early curiosity was the vast, unknown 'Inland Ice' - the ice sheet itself (Fig. 1.7) - and many of the early travellers ventured to the edges of the ice, recording in dramatic etchings, vivid prose, and sketch maps what they saw. Though sometimes of doubtful validity, these early records provide a fascinating, evocative, and invaluable backdrop against which more recent glacier fluctuations can be assessed, especially when the observations were made by a trained geologist such as the far-travelling J.A.D.Jensen. Weidick (1959) draws on such sources extensively and quotes from many of the earliest travel writings in his account of historic glacial variations in West Greenland.
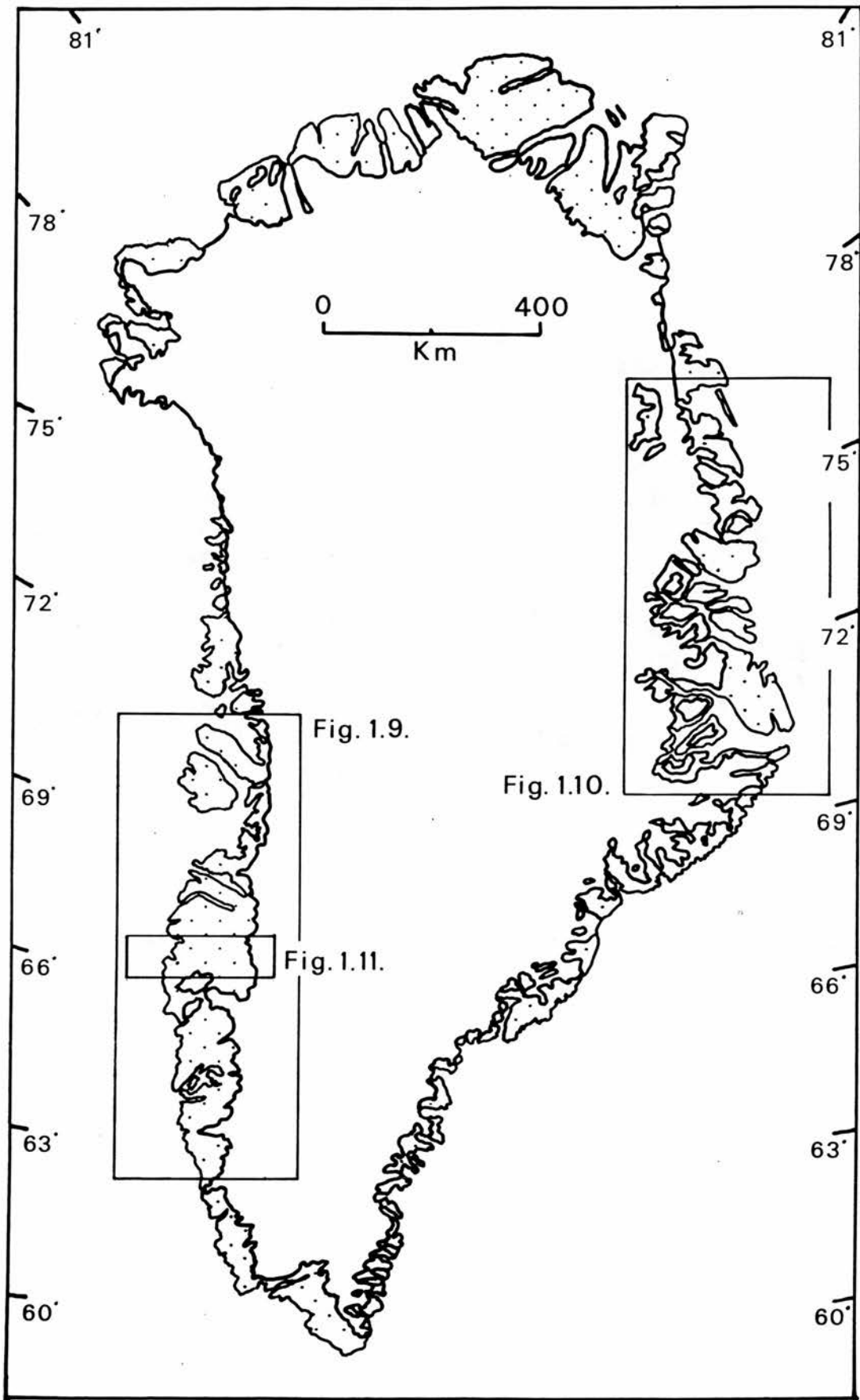
FIGURE 1.7. Present extent of the Greenland Ice Sheet.

35

Systematic geological studies, including work on the
Quaternary, were carried out from the middle of the last
century, and have been mainly associated with the mapping
projects of the Greenland Geological Survey. The results of
these projects have been compiled on three Quaternary map
sheets at 1:500,000 which cover West Greenland between Kap
Farvel (59°N) and 71°N (GGU, 1974,1978,1987). Scientific
expeditions from a variety of countries, but notably France,
Britain and the USA, have also contributed to our knowledge
of the contemporary ice sheet and its past oscillations.
Extensive and thorough reviews of the Quaternary geology of
Greenland have been published recently by Kelly (1985) and
Funder (1989), building on and updating the earlier reviews
by Weidick (1975b,1976,1984). They carefully discuss the
nature, validity, and variable interpretations of the
stratigraphic data. In addition, a thorough review of the
dynamic and climatic history of the ice sheet has just been
compiled by Reeh (1989). Various estimates of the mass
balance of the ice sheet are given in Table 1.1, and some
general statistics in Table 1.2. The wide range of values in
Table 1.1 demonstrates that much uncertainty still surrounds
our current understanding of the ice sheet.

ACCUMULATION.	ABLATION.	CALVING.	BALANCE.	SOURCE.
+ 630	-120 to -270	- 240	+270 to +120	Bader,1961.
+ 500	- 272	- 215	+ 13	Benson,1962.
+ 500	- 330	- 280	- 110	Bauer,1967.
+ 500	- 295	- 205	0(assumed)	Weidick,1985.
+ 487	- 169	- 318	0(assumed)	Reeh,1985.
		- 308		Reeh,1989.
+ 535				Ohmura & Reeh, in press.

TABLE 1.1: Estimates of the mass balance of the Greenland Ice
Sheet. All figures in $km^3 a^{-1}$ water equivalent.

36

AREAS:

LAND SURFACE:	2,175,600 km^2.		
GREENLAND ICE SHEET:	1,701,300 km^2	= 78%	of total area.
LOCAL GLACIERS:	65,500 km^2	= 3%	of total area.
ICE FREE:	408,800 km^2	= 19%	of total area.

Accumulation Area of ice sheet: 1,429,092 km^2 = 84%.
Ablation Area of ice sheet: 272,208 km^2 = 16%.

VOLUME: 2.74 x 10^6 km^3 water equivalent.
                (= 7.4 m rise in world sea levels if melted.)

MAXIMUM ELEVATION: NORTH DOME: 3205 m.
                                SOUTH DOME: 2830 m.

MAXIMUM THICKNESS: 3420 m at 71°42'N, 38°48'W.

MASS BALANCE: BROADLY IN EQUILIBRIUM.

EQUILIBRIUM LINE ALTITUDE: c.1000 m in the north.
                                1600 - 1800 m in the south.
        (annual variation $\leq$ 300 m in central West Greenland.)

MAXIMUM AGE OF BASAL ICE: ? > 500,000 years. (Dansgaard, 1987)

BASAL TEMPERATURE RANGE: -13°C - -2.6°C (= Pressure Melting
                                Point for 3000 m thick ice.)

ICEBERG CALVING DISCHARGE:

TOTAL: 308 km^{3}a^{-1} water equivalent. (178 km^3 of this total
        is based on estimates from a variety of sources.)

REGIONAL:

North:	20 km^{3}a^{-1}	water equivalent.
North East (70° -80°N):	35 km^{3}a^{-1}	water equivalent.
South East (59° -70°N):	80 km^{3}a^{-1}	water equivalent.
South West (59° -72°N):	113 km^{3}a^{-1}	water equivalent.
North West (72° -76°N):	60 km^{3}a^{-1}	water equivalent.

TABLE 1.2: Greenland Ice Sheet statistics.   All information
        from Reeh (1989) except where indicated.

1.3.ii. a) Greenland Ice Sheet history.

The majority of the Quaternary deposits and landforms in Greenland belong to the most recent glacial episode and its deglaciation phases, and to the associated subsequent glacio-isostatic marine transgression and regression. Information about previous glacial and interglacial periods has also been obtained from the few locations where earlier deposits have been preserved. Evidence from shelf areas indicates, for instance, that an early glaciation of Greenland, which was more extensive than any succeeding one, occurred near the end of the Pliocene about 2.4 Ma (Funder, 1989). The stratigraphic terminology is still evolving as research progresses, but Fig. 1.8 shows the recent scheme proposed by Funder (1989), which largely follows that of Kelly (1985). Only one extensive glaciation can be recognized during the most recent glacial period (Funder and Simonarson, 1984); the Sisimiut, Flakkerhuk, and Independence Fjord glaciations of West, East, and North Greenland respectively occurred after 40 ka, and are equivalent to the North American Wisconsinan Glacial and the European Devensian - Weichselian. These glaciations culminated around 14 ka and large-scale oscillations were apparently in phase in all parts of the ice sheet. At its maximum extent, this ice was grounded as much as 30 - 50 km beyond the present coasts of West Greenland (Fig. 1.9; Brett and Zarudzki, 1979). In central East Greenland it advanced tens of kilometres beyond the coast (Fig. 1.10), and locally, off South-East Greenland, it may have extended as much as 200 km (Funder, 1989). This view is based on the

CHRONOSTRATI GRAPHY		ISOTOPIC STAGES	AGE, (ka)	WEST GREENLAND	EAST GREENLAND	NORTH GREENLAND
Holocene		1				
Wisconsinan	Late	2	10	Taserqat stade	Milne Land stade	
			25	Sisimiut glaciation	Flakkerhuk glaciation	Independence Fjord glaciation
	Middle	3	65	Svartenhuk marine episode	Jameson Land marine episode	
	Early	4 to 5d				
			115			
Sangamonian		5e		Kaffehavn interglaciation	Langelandselv interglaciation	
			130			
Illinoian		6		Fiskebanke glaciation	Scoresby Sund glaciation	? Bliss Bugt glaciation

?

Pátorfik marine episode

?
↓

Early Quaternary		1800			Lodin Elv Formation	Zone II	Kap København Formation	member B
Pliocene						Zone I		member A

FIGURE 1.8. Stratigraphy of pre-Holocene glacial and marine events in Greenland. (After Funder, 1989.)

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existence of large, undated linear ridges on the continental shelf which are assumed to be terminal moraines of the last glaciation. Despite this much increased geographical extent, it is conceivable that the ice sheet at this stage was very little thicker than it is today, due to decreased accumulation rates, changed ice temperatures, and predominantly softer ice (Reeh, 1989).

Deglaciation was broadly contemporaneous with other northern hemisphere ice sheets. The evidence of the marine limit in West Greenland suggests that the deglaciation of the outer coasts took place between 14 ka and 10 ka (Kelly, 1985), but in most areas the ice margin was located near the current coastline at 10 - 11 ka (Funder, 1989).

FIGURE 1.9. West Greenland ice margins during (1) Sisimiut glaciation (ca. 14 ka), (2) Taserqat stadial (ca. 10 ka), (3) and Fjord stadial (8 ka). Dotted line shows present ice margin. (After Funder, 1989.)

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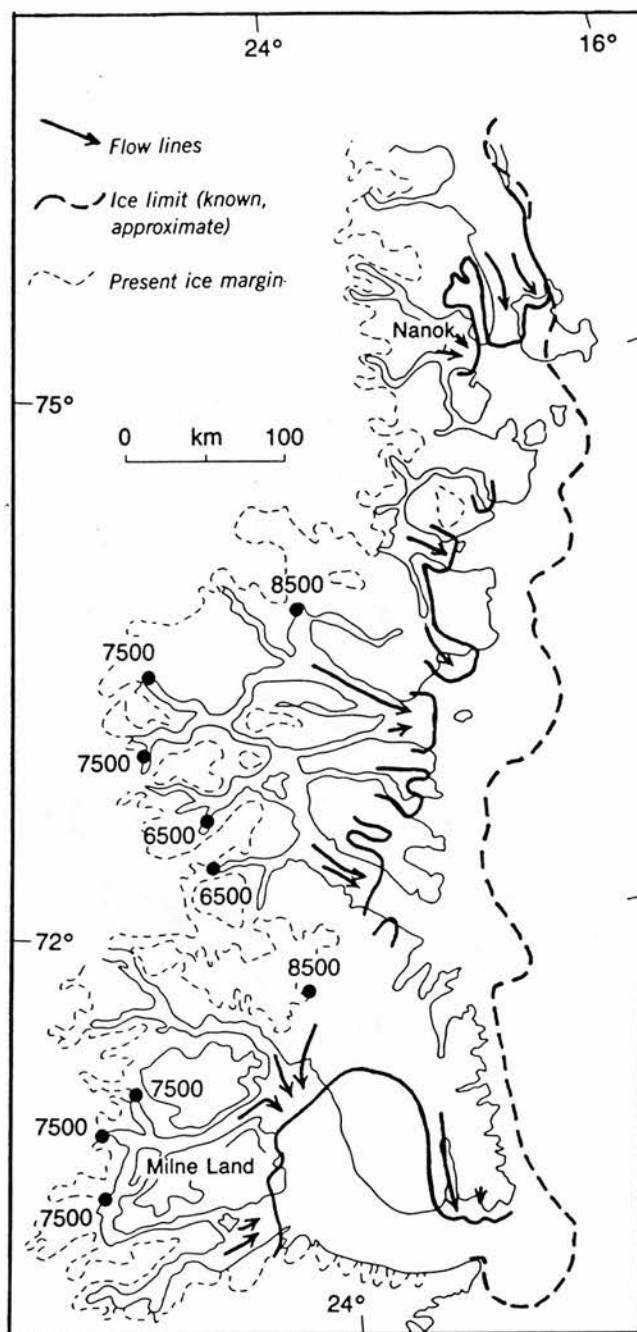
FIGURE 1.10. East Greenland ice margin at ca. 10,000 BP, and maximum extent of the Flakkerhuk Glaciation, ca. 14,000 ka. Dates are minimum ages of attainment of present margin position during retreat. (After Funder, 1989.)

The readvances of the Taserqat and Milne Land stadials in West and East Greenland (Figs. 1.9 and 1.10) are similar in age to Younger Dryas advances elsewhere, occurring at about 10 ka. By around 7 ka the ice sheet had retreated to its

41

present limits, and the contemporary flora and fauna show that at that time temperatures were slightly warmer than now (Fredskild, 1985; Funder, 1978). Between then and 3 ka the ice sheet is estimated to have retreated 10 km or more behind today's ice margin (Weidick, 1985). Readvance began after this time (Kelly, 1980), culminating in the Little Ice Age build-up of the late nineteenth and early twentieth centuries which saw glaciers almost everywhere adopting more advanced positions than at any time since the last glaciation.

1.3.ii. b) Glacio-climatic or glacio-dynamic control? A current debate.

The deglaciation phases left an abundance of geomorphological evidence. The central parts of West Greenland have been the seminal area for the study of these phases (eg. Weidick, 1968,1972a,1972b; Ten Brink, 1975; Ten Brink and Weidick, 1974) but although most areas have now been mapped in some detail, the interpretation of their glacio-climatic and glacio-dynamic implications remains a matter for debate. Funder (1989) regards the issue as unresolved at the moment:

> In West Greenland, the Taserqat stade left moraines along a considerable part of the coastline; however, whether they mark a climatically conditioned readvance or a glacio-dynamically conditioned stillstand of the ice sheet margin is uncertain. (Funder, 1989, p.786)

Kelly (1985) summarises the differences of opinion:

> A major uncertainty exists about the origins of these moraine systems. They may represent either significant readvances of the ice sheet margin in response to regional positive changes in the ice sheet mass balance, albeit during a period of overall deglaciation, or alternatively, local stillstands or readvances caused by topographic influences or by shifts in ice sheet streamline patterns during deglaciation. (Kelly, 1985, p.477)

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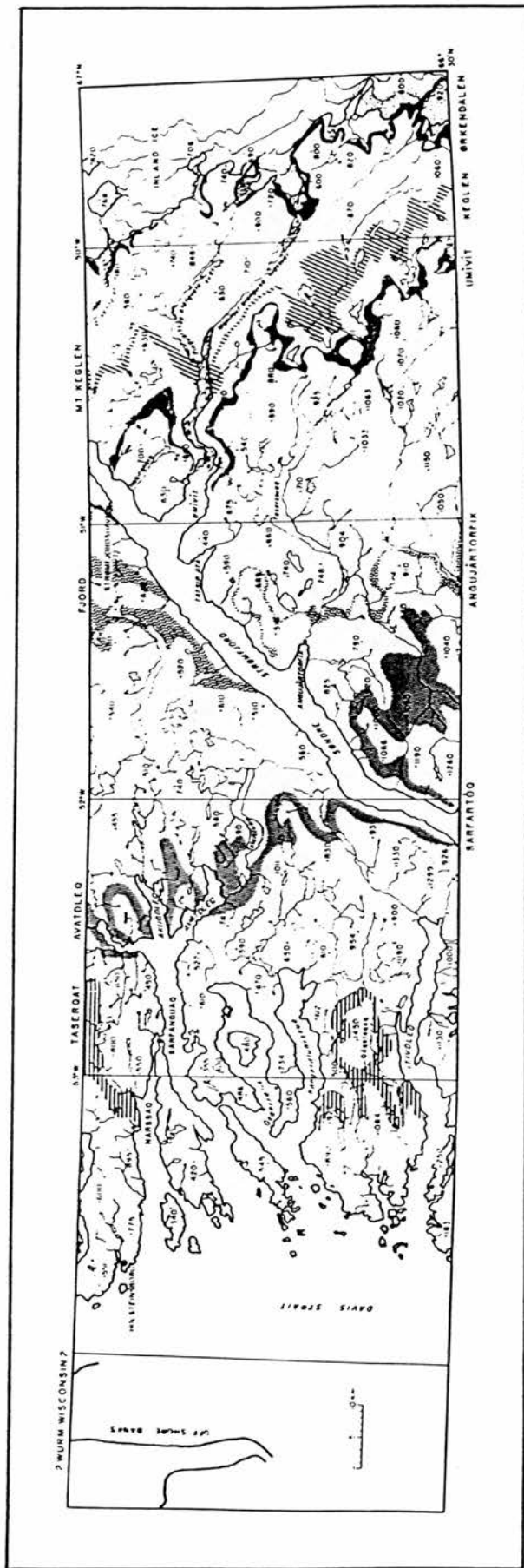
FIGURE 1.11. The glacial geomorphology of the Sondre
        Strømfjord area, showing the major moraine systems.
        (After Ten Brink and Weidick, 1974.)

43

Ten Brink and Weidick (1974), Ten Brink (1975), and Weidick (1968,1972a,1972b,1975a,1975b,1976,1985,in press) have taken the former view, identifying a series of 'moraine systems' and interpreting them as glacio-climatic phenomena - readvances caused by small but regional decreases in mean ablation season temperatures maintained over decades or a few centuries. Ten Brink and Weidick's (1974) map of the 'classic' central area around Sondre Strømfjord is reproduced here as Fig. 1.11. As evidence for their climatic hypothesis, they have pointed to the nature of the moraines - their continuity, their continuation over uplands between the major troughs, and to the overall similarity in the ages of moraine systems in different areas. Oscillations of the ice sheet margin today seem to be largely controlled by summer temperature (Weidick, 1968; Braithwaite and Olesen, 1984), and so, using such evidence, Weidick (1976,1985) has attempted to correlate the moraine systems with cold phases in the oxygen-isotope records in the ice cores (eg. Dansgaard, 1987). This hypothesis therefore suggests that:

> ...construction of the major Holocene moraine systems in West Greenland was caused by slight temperature decreases, which decreased rates of ablation and thereby produced practically immediate advances of the ice sheet margin, but did not necessarily affect the long-term equilibrium of the ice sheet. (Ten Brink and Weidick, 1974, p.439)

In contrast to this interpretation, it has been claimed that the moraine system continuity is 'more apparent than real' (Kelly, 1985; Funder, pers. comm. 1988) and that they could be explained as metachronous local stillstands during a period of slower retreat. The recent reviewers of the evidence (Kelly, 1985; Funder, 1989) both suggest that the

correlations between the much-studied central area of West Greenland and areas to the north and south is not particularly good, even though, in general terms, it seems that there were periods of moraine formation at 8.5 - 8.0 ka and around 7.5 ka which can be widely recognized.

This lack of correlation suggests that the local topography may have played an important part in moraine formation during the retreat of the ice sheet. Weidick (1968,1985) has noted this and specifically discussed the possible impact of Mercer's (1961) principle, giving recent examples of calving glaciers halting at topographic constrictions in the fjords (Weidick, 1968, pp.36 - 37). Funder (1970,1971,1972), in a series of papers concerning the Holocene retreat stages in inner Scoresby Sund in East Greenland, has stressed the role of topography in controlling the stillstand locations during retreat. Fig. 1.12, reproduced from his 1972 paper, clearly shows the relationship of stillstands to topographic characteristics within the fjord system. More recently Funder (1985,1989) has specifically questioned the climatic interpretations of previous workers. He discusses the role of topography in conditioning different ablation mechanisms (deep-water calving, grounded calving, melting) which in turn result in contrasting regional retreat chronologies, such that 'the sensitivity of the Inland Ice margin to climatic change in the Holocene was strongly controlled by ablation mechanism' (Funder, 1989, p.751). Areas where calving was dominant saw recession several millenia earlier than those where ablation was predominantly by melting (Weidick, 1984).
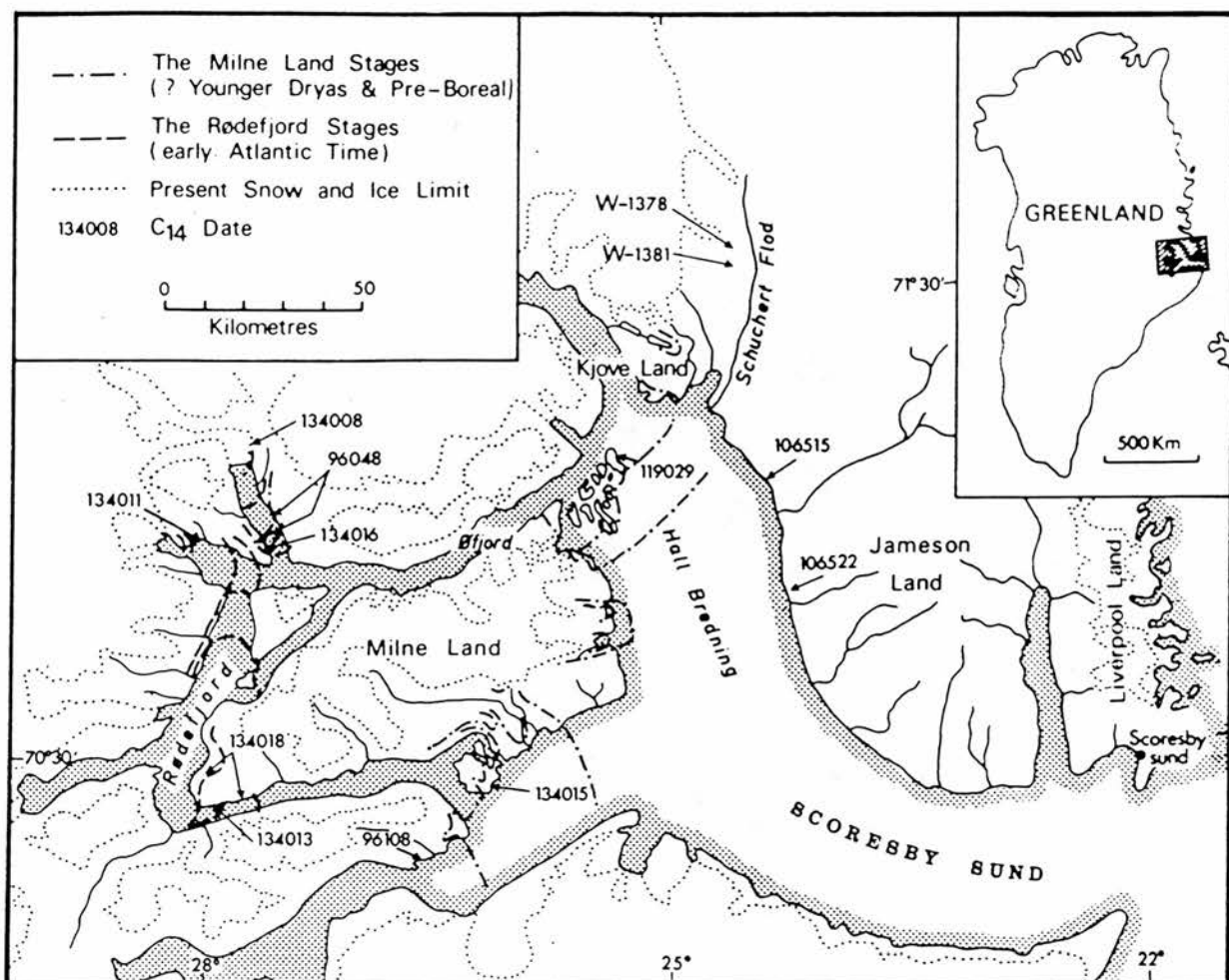
FIGURE 1.12. Retreat stages during the Milne Land stadial, inner Scoresby Sund. (After Funder, 1972.)

Funder's own view is as follows:

> The location of moraines along the fjords ... indicate that the moraines were formed as a product of interaction between the glacier and the topography of its bed, rather than in response to climatic change. (Funder, 1989, p.750)

The importance of the evolving sequence of dominant ablation mechanisms is also apparent from the depositional evidence at locations where the ice margin retreated out of the sea. Almost everywhere, such locations are marked by stillstands

46

or readvances. This has also been noted by Weidick (1985). Throughout the region, the ice margin was at or near the present coastline at 10 - 11 ka. In the opinion of Funder (1989), this synchroneity can best be explained by the assumption that the break-up of the ice margin was triggered by a eustatic sea level rise which rendered the shelf-based portions unstable along the entire margin, and that the ice sheet adjusted to this change by retreating until the margin was again land-based and stable, near the present coastline.

It seems then that 'the sea/land transition formed a major glacio-dynamic obstacle in the deglaciation process' (Funder, 1985, p.140). By determining a change in ablation mechanism, this marine/terrestrial transition causes substantial reductions in ice losses; it therefore provides stability, acts as a 'pinning zone', and apparently pre-determines a cessation of retreat during the overall recession, irrespective of the contemporary climate. In some locations this sea/land transition clearly provided great stability. For instance, Sermilik Fjord at 63°N was deglaciated prior to 13,380 ± 175 BP but the ice front then remained pinned at the present coastline for the ensuing 5000 years (Weidick, 1975a) during a period of substantial climatic change. In the cool, moist climate along the Melville Bugt coast of north-west Greenland, the ice margin has been at the sea/land transition for the last eight or nine thousand years.

A recent study in central West Greenland (Warren and Hulton, 1989,1990; Chapter 4), in an area mapped in detail by Weidick (1968), has provided further evidence that, while

47

climate was the dominant influence on the overall ice sheet recession, the local topography controlled the detailed pattern of ice sheet retreat. Kelly (pers. comm., 1989), working in North Greenland, has also noted the difficulty of differentiating between climatic and topographic effects. Unless these can be clearly separated, any palaeoclimatic inferences drawn from the glacial geomorphological record must be open to question.

Substantial differences of opinion exist over these issues. If the dynamics of the retreat of the Greenland Ice Sheet since the end of the recent glaciation, and its relationship to climate, are to be understood, there is therefore a pressing need to re-evaluate the established interpretations of parts of the Quaternary record and to examine the role of topography in controlling recent behaviour of the ice margin. This will aid our understanding of past glacio-dynamics, and improve the precision with which we can predict the likely response of ice sheet margins to future climatic change. It seems clear that one of the important uncertainties concerning the dynamic interaction between ice sheets and climatic change is the role within ice sheet dynamics of iceberg calving.

CHAPTER TWO.

THEORY: CRYOSPHERE - HYDROSPHERE INTERACTION.

2.1. ICEBERG CALVING.

2.1.i. Calving Glaciers.

Iceberg calving is the most important ablation mechanism today, accounting for about 90% of the ablation loss from Antarctica and about half of the mass loss from the Greenland Ice Sheet. At the height of the last glacial period, its relative importance was even greater because of the extensive marine and lacustrine margins of the Pleistocene ice sheets. With the exception of glaciers during surges, calving glaciers are the most dynamic on earth. For example, Jakobshavns Isbrae in central West Greenland has the highest sustained velocities recorded, reaching rates of over 20 m d^{-1} and discharging between 30 and 40 km^3 a^{-1} of ice into Jakobshavns Ice Fjord, the equivalent to up to 7.6% of the total net accumulation of the ice sheet (Pelto and others, 1989; Bindschadler, 1984). These figures are exceptional, but active calving glaciers such as Rinks Isbrae in West Greenland, Daugard-Jensen Gletscher in East Greenland and Columbia Glacier in Alaska, all sustain high velocities of many metres per day and calving fluxes measured in cubic kilometres each year (Olesen and Reeh, 1969; Reeh and Olesen, 1986; Meier and others, 1985).

Iceberg calving is important glaciologically for four reasons:

i) <u>Mass efficiency.</u> Calving is the most efficient means of

mass loss, permitting much larger volumes of ice to be lost over a given time than melting. Tabular icebergs with surface areas of thousands of square kilometres break off from Antartic ice shelves, and very rapid mass loss has been observed from tidewater glaciers (eg. Epprecht, 1987; Hughes, 1986). In 1982, catastrophic calving at the terminus of Jakobshavns Isbrae caused a discharge of 10.5 km^3 of ice (equivalent to 5% of Greenland's annual iceberg production) over an 18 day period at a peak rate of more than 500 million tonnes per day (Williams and Birnie, unpublished; Birnie and Williams, 1985; M.I.S.R./R.A.E., 1983).

ii) <u>Energy efficiency.</u> For other means of ablation, the necessary energy to remove the ice has to be imported to the ice mass. Iceberg calving is uniquely efficient in that the ice is exported to the energy source (Denton and Hughes, 1981), the icebergs breaking up and melting in the oceans (Robe and others, 1977; Dunbar, 1978; Hotzel and Miller, 1983).

iii) <u>Instability.</u> The presence of iceberg calving renders a glacier front inherently unstable (Hodge, 1979; Meier, 1979) because of the mass efficiency of the process and the sensitive relationship between calving speed and water depth (Brown and others, 1982; 2.1.iii). These glacio-dynamics can partially decouple calving glaciers from climatic forcing (Meier and Post, 1987) and set them apart from land-terminating glaciers; the latter can stagnate <u>in situ</u> whereas calving glaciers cannot. Tidewater glaciers have no stable response to long-term balance deficits, as land glaciers have; reducing ice flow to the front causes glacier

50

thinning, resulting in a less firmly grounded glacier sole, and an increase in bottom-sliding rate (Clarke, 1987). These decoupling influences are important today, but are likely to have been of much greater importance during the disintegration of the Pleistocene ice sheets when large sections of the ice margins lay in the sea or in large proglacial lakes.

iv) <u>Climatic information.</u> Observations of the changes in distribution and characteristics of icebergs in today's oceans can provide information about ice sheet mass balance and sea surface conditions (Orheim, 1987,1990). From the extent of iceberg-rafted debris in the ocean floor sediments, it is possible to make inferences about past sea surface conditions, and hence palaeoclimate (eg. Ruddiman and McIntyre, 1981).

2.1.ii. Calving Mechanisms.

Calving is controlled by the dynamics of the terminal section of the glacier. Stress regimes in the terminal parts of calving glaciers are extremely complex (Reeh, 1968; Lingle and others, 1981), and experimental research in the unpredictable and inaccessible environs of active calving fronts is inherently dangerous and fraught with problems. Consequently, the primary stress mechanisms responsible for iceberg calving are poorly understood. Explanations of observed calving phenomena have generally been proposed within discrete theoretical boundaries, whereas in reality many of the contributing forces will interact within the same ice front.

Rapid flow is typical near the termini of calving glaciers. Many calving glaciers flow rapidly because the water pressure at the bed approaches the ice pressure (Meier, 1989). Water pressure must be high in order to drive the water out of the glacier against the water pressure at the terminus; ice velocities at the terminal cliff vary directly with tidal cycles (Walters, 1989). Rapid flow is controlled mainly by the longitudinal gradient in normal stress (gradient stress) as well as by the shear stress (slope stress). The former is controlled by the geometry at or near the terminus, and this is related to the calving rate. At the terminal cliff of grounded glaciers (or at the grounding zone of glaciers that have a floating portion), as much of 95% of the fast motion may be effected through sliding (Jania, 1988), whereas higher up the glacier, deformation under large stresses may account for much of the

52

movement (Echelmeyer and Harrison, 1989,1990).

Iceberg calving occurs when the stresses set up within a part of the ice front by external forces or internal deformation exceed the strength of the weakest structural link. Failure then takes place along that weak link and a mass of ice is discharged into the lake or the sea. However, fundamentally different calving mechanisms operate at the termini of floating and grounded glaciers, primarily because of the new stress fields that are set up when flotation occurs (Theakstone, 1989). Calving at floating fronts is the product of both internal and external forces, whereas grounded calving cliffs calve primarily in response to stresses set up within the glacier.

2.1.ii. a) Calving mechanisms at grounded ice fronts.

A unique natural laboratory for the investigation of the stress fields in calving ice fronts was created by the 1970 volcanic eruption on Deception Island which removed part of the snout of a glacier and left an ice wall which is calving away in slabs (Fig. 2.1; Hughes, 1989). The analogy developed from analysis of this ice wall compares the slip in shear bands which are bending towards the crater with the slip between the pages of a book bent around its binding (Hughes and Nakagawa, 1989).

Shear bands rise almost vertically and curve forward, with shear offset increasing forwards, achieving maximum offset where the shear bands intersect ring fault crevasses on the up-slope side of the ice wall. Bending creep (resulting from horizontal pulling forces) causes the ice wall to overhang at angles of up to 20°; slabs calve off when
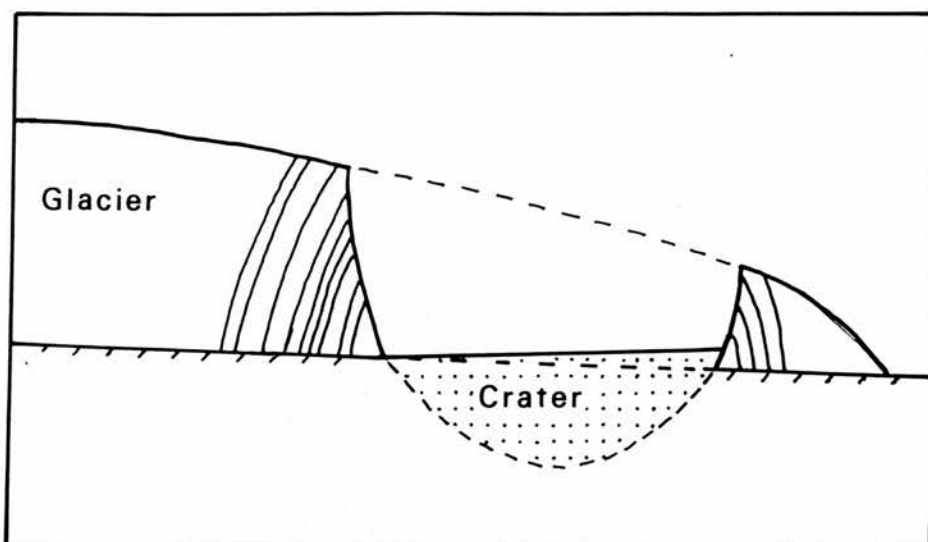
FIGURE 2.1. A schematic illustration of the Deception Island
ice wall, produced by the volcanic eruption of 1970.
(After Hughes and Nakagawa, 1989). It shows the
original glacier surface profile and the shear
surfaces developed subsequent to the eruption.

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bending shear opens crevasses to a depth where they allow the
ice slab to shear along one of the slip lines (ring faults).
Crystallographic analysis has revealed easy-glide fabrics
favouring shear deformation along lines of maximum shear
stress aligned at 45° to the calving wall. Shear rupture
along these 45° planes near the base of the slabs between the
ice wall and the shear band leads to slab-calving (Fig.
2.2.a).

Thus calving occurs when easy-glide fabrics in the
shear bands drastically reduce the coupling across the shear
bands, and drastically enhance the slip along the shear
bands. Since these shear bands are nearly vertical, up-
glacier uncoupling means that almost the entire weight of the
ice slabs between the shear bands must be borne by the ice at

54

the base. Shear rupture following the development of easy-glide fabrics can only occur once bending creep has produced these fabrics in the shear bands. Consequently, bending creep is the mechanism controlling the rate of slab-calving.

Hughes (1989), and Hughes and Nakagawa (1989), generalize from these observed calving dynamics of an ice wall on land, to apply their concepts to ice grounded in water and ice walls standing on beach shorelines. In the latter case, calving is promoted by the existence of a continuous shoreline groove eroded at the base by wave action in the intertidal zone (Fig. 2.2.b). Slab-calving is caused by near-vertical shear rupture along transverse crevasses immediately behind the wall. This type of calving is an important ablation mechanism today along the northern parts of the Antarctic Peninsula, and may have been important along ice margins on the shorelines of proglacial lakes during the last deglaciation.

The horizontal pulling force which bends the ice forwards increases as the square of the height of the wall. This force should therefore calve arcuate bights in the ice wall near the mid point along it where the height is usually greatest. However, water depth is also greatest at this point, and as the water depth increases, so the horizontal pulling force decreases. Similarly, basal shear stress decreases progressively to become zero at the point of flotation. This reduction of basal stress and pulling force should prevent shear bands from forming and thus suppress the slab-calving mechanism. If this mechanism were the only one operating, calving should decrease with increasing water
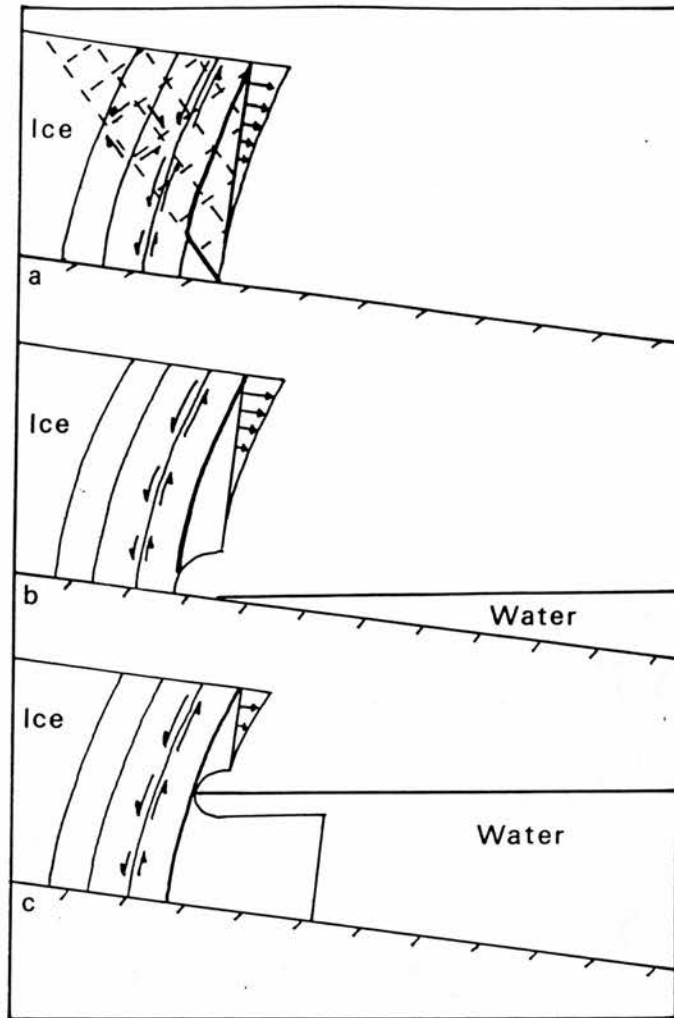
FIGURE 2.2. The bending creep mechanisms for shear rupture in calving ice walls. (After Hughes and Nakagawa, 1989). a) On land. b) At the shoreline of a beach. c) Grounded in deep water. Straight dashed lines intersecting the ice wall at 45° are the slip lines of maximum shear stress for homogenous creep. Curving solid lines rising vertically from the bed are shear bands produced by bending creep. Heavy lines are calving surfaces for shear rupture.

depth. That the reverse is the case (Brown and others, 1982) shows that other mechanisms take over as this subaerial slab-calving mechanism is suppressed by increasing water depths.

However, Hughes and Nakagawa (1989) suggest that underwater calving may also proceed by shear rupture along the shear bands produced by bending creep. At the waterline,

calving is again promoted by the presence of an eroded groove (Fig. 2.2.c). Above the waterline, rupture proceeds downwards due to gravity force; below the waterline it proceeds upwards as the result of upward buoyancy forces. If subaerial and submarine calving caused by shear rupture occurs at the same calving stress, then underwater slabs should be ten times larger than subaerial slabs because the density of ice is about ten times the density difference between ice and water.

Hughes and Nakagawa (1989) assert that:

i) In all the above environments, the mechanism controlling the rate of calving is bending creep leading to shear bands leading in turn to shear rupture.

ii) The calving rate is accurately predicted by a calving formula based on this mechanism.

iii) Calving follows shear rupture at a shear stress of about 1 bar.

However, no empirical observations have ever revealed the existence of the projecting submarine ice foot envisaged by Hughes and Nakagawa (1989; Fig. 2.2.c). On the contrary, submersible observations of the submerged parts of ice cliffs in Glacier Bay, Alaska, showed that the faces were either nearly vertical or slightly overhanging (Powell, in Hooke, unpublished). Submarine calving due to buoyancy forces is known to be important because large ice blocks are commonly seen rising to the surface at calving fronts, but, so far, observational evidence does not support the theoretical physics of submarine calving proposed by Hughes and Nakagawa (1989). Their generalisations from the special case of the

57

Deception Island ice wall to calving cliffs in water must therefore be regarded as questionable until further data become available.

Similar work has been undertaken by Iken (1977) and by Funk and Müller (unpublished) who have monitored and then modelled calving events into shallow lakes at the snouts, respectively, of the grounded Grubengletscher and Unteraargletscher in Switzerland. Crack progagation commences in the zone of maximum tensile stress behind the cliff, and was found to progress in stepwise fashion, as stress was alternately built up and then released. The build up of stress at the base of the crevasse resulted both from ice flow, which increased the overhang of the cliff, and, less importantly, undercutting of the cliff by lake currents. Iken (1977) found that the initial opening of the crack occurred perpendicular to the direction of tensile principal stress at a critical stress of about 1 bar.

However, in contrast to the mechanism proposed by Hughes and Nakagawa (1989), the Swiss work shows that the planes of shear fracture dip up-glacier at about 30° to the direction of stress. Similarly, on the basis of their work on Svalbard tidewater glaciers, Jania (1988) and Jania and Kolondra (1985; unpublished) liken the calving process to landslide slumping along slip-failure planes which are present due to the tensional flow in the basally-slipping tidewater terminus. The mechanism is enhanced by the assemblage of positive feedbacks which have been termed 'the Jakobshavns Effect' by Hughes (1986; 2.1.iii.c).

Which of these contrasting calving mechanisms operates will depend largely on the degree of basal coupling, and on the basal topography.

2.1.ii. b) Calving mechanisms at floating ice fronts.

The first analysis of the calving stresses in floating glaciers and ice shelves was presented by Reeh (1968). He likened the problem to a viscous beam subjected to torque produced by the imbalance of hydrostatic pressures at the front (Fig. 2.3).
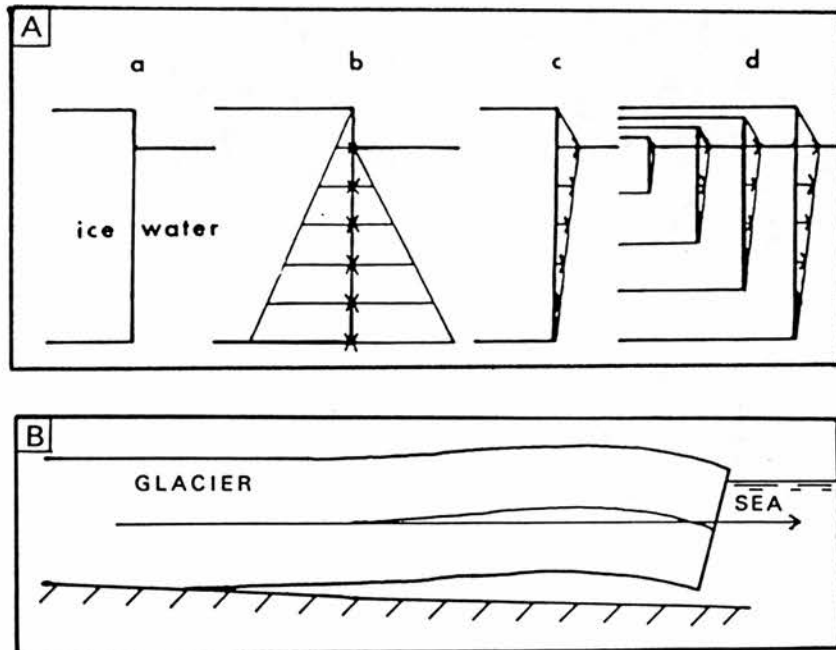
FIGURE 2.3. A) Pulling power in buoyant ice. (After Hughes, 1987). A floating ice margin is a vertical ice cliff (a), against which horizontal hydrostatic forces that increase linearly with depth point seaward in ice and landward in water (b), and can be subtracted from top to bottom to give a net horizontal hydrostatic force that pulls the ice seaward (c) and increases with the square of buoyant ice thickness (d). Pulling power is this Force multiplied by the ice velocity.
B) The effect of the imbalance of hydrostatic forces on a floating ice tongue, as modelled and observed by Reeh (1968).

This approach has been followed, amongst others, by Hughes (1987; Fig. 2.3.a). This imbalance will increase from bottom to top of the ice front so that the glacier will be subject to bending. The upper layers will consequently be stretched more than the lower layers, leading to ice front rotation and overhanging; the bending down of the ice front is compensated for by up-arching of the ice behind (Fig. 2.3.b). Tensile stresses and shear stresses reach a maximum at a distance from the front approximating to the ice thickness. The combination of shear and tensile stresses is potent, promoting fracturing from the upper surface of the glacier where the stresses are greatest. Reeh supported these ideas with observations at the snout of Rinks Isbrae in West Greenland; large calving episodes were preceded by frontal downwarping of 80 m behind which was a 20 m 'up-wave'. Reeh-type calving has also been observed at the proglacial Generator Lake on Baffin Island by Holdsworth (1973) and at the snout of Daugaard-Jensen Gletscher in East Greenland where the opening of a major crevasse characteristically preceded large calving events (Olesen and Reeh, 1969; Reeh and Olesen, 1986).

Fastook and Schmidt (1982) carried out a finite element analysis of calving, the results of which were qualitively very similar to those of Reeh (1968). They report the downwarping and up-arching of ice resulting from the horizontal imbalance of hydrostatic forces and consequent bending torque. In their model, the non-uniform tension distribution along the top surface peaked at a distance of about half an ice thickness behind the ice front with a

maximum tension of 3.44 bars. Such tension maxima lead to ice fracture. Although there is no adequate fracture criterion for ice, Fastook and Schmidt address the problem of subsequent crack propagation, and demonstrate that a crack needs to be nearly filled with water in order to prevent closure due to ice flow. Calving by this mechanism will necessarily be strongly seasonal because it depends on the presence of meltwater in abundance.

This analysis does not deal with the possibility of bottom crevasses opening up to join with top crevasses to cause calving (Hughes, 1983). The existence of bottom crevasses was proposed theoretically by Weertman (1973) and later confirmed empirically by Swithinbank (1978). The analysis of ice shelves by Weertman (1980) shows that, although in many situations the closing influence of water freezing onto the walls of cold ice (< -10 C) will overcome that of creep deformation tending towards opening, in ice shelves thicker than 400 m, bottom crevasses can extend through nearly 78% of the shelf thickness. The effect which the presence of bottom crevasses has on Reeh-type calving mechanisms and rates has not been analysed, but intuitively they must further weaken the ice front, promoting earlier calving of smaller icebergs.

In addressing the importance of the role of fracture in the disintegration of ice shelves, Hughes (1983) developed the concept of critical fracture strain. Strain energy accumulated at the grain boundaries in polycrystalline ice is released when a critical strain triggers viscoplastic instability. This release occurs rapidly by fracture and

slowly by recrystallisation. Ice moving from the grounding line to the calving front is moving through a constantly changing strain field as a result both of general flow and of deformation over or around bedrock obstacles. An ice fabric that is stable for a given strain field becomes metastable as the field changes and unstable at the critical strain. Further strain leads to fracture or recrystallisation. The critical strain is the strain of viscoplastic instability. If this strain requires a stress that exceeds the fracture stress, ice will crack instead of recrystallising. Crevasses open up on ice shelves when strain energy is reduced by fracture instead of by recrystallisation, and they open normal to the largest tensile principal stress.

Theoretical analyses such as these greatly aid our understanding of the forces and mechanisms involved in the calving process, but, in the real world, the process is complicated by:

i) the presence of crevasses formed higher up the glacier and carried to the front by glacier flow, and,

ii) by the interaction of floating ice fronts with tides, weather, and seasonal changes in the environment at the ice front.

Crevasses are weak links in the frontal zone that facilitate structural failure and calving. Intuitively, crevasses must form the boundaries of calving events, and an abundance of empirical evidence confirms this. For instance, the calving of a c. 0.45 km^2 segment of Jakobshavns Isbrae in 1982 followed the opening of a large transverse crevasse (Epprecht, 1987), and Dowdeswell (1989) describes how the

62

degree of crevassing of Spitsbergen glaciers determines the dimensions of the icebergs. Rapid calving can follow the arrival of a relict transverse crevasse field at the terminus (Powell, 1984). The characteristically chaotic nature of the surface of most active calving glaciers may increase the surface area by as much as three times; this triples surface melting and hence much increases the abundance of meltwater (Hughes, 1986). This meltwater may refreeze internally, releasing much latent heat, warming the ice and leading to faster flow and greater deformation. Alternatively, meltwater will flow to the glacier bed and promote basal uncoupling. Both possibilities produce powerful positive feedback.

Additional important stresses in floating ice fronts may be induced by **tides, changing lake levels, and extreme weather conditions.** Rink (1877) was the first to suggest that tidal flexure might contribute to iceberg formation, from his observations of floating glaciers in the ice fjords of West Greenland. Flexural stresses can lead to the breakup even of free-floating ice masses (Goodman and others, 1980; Kristensen and Squire, 1983a,b). Holdsworth (1969) made the point that the frontal bending stress mechanism for iceberg formation (Reeh 1968) produces prismatic icebergs, whereas most Antarctic icebergs are tabular; flexure stresses must therefore be influential. A geographical contrast confirms this. Ice tongues are found in the Ross Sea but not in the Weddell Sea; tidal ranges are three times greater in the latter, and thus hinge bending stresses would be much higher, preventing the formation of ice tongues.

63

Lingle and others (1981) analysed the stresses in the floating terminus of Jakobshavns Isbrae and concluded that the most important control on calving is the shear-strength of the ice tongue. The ice takes about a year to travel the 8 km from the grounding line to the calving front; during this time, about 700 tidal flexure cycles take place, resulting in fatigue failure along the marginal grounding lines as the ice moves downstream, and deep basal and surface cracking. Flexure thus contributes to frontal breakup by progressively decoupling the glacier from the fjord wall. The degree to which fatigue failure has progressed will determine how the terminus responds to other stimuli, such as rough seas and spring tides.

Williams and Birnie (unpublished) noted that the three big calving episodes at the front of Jakobshavns Isbrae in 1982 all coincided with spring tides and strong east winds, but that similar combinations at other times did not result in calving. Echelmeyer and Harrison (1989,1990) found that the velocity of the glacier varied by as much as 35% with the level of the tide. By contrast, velocities at grounded calving fronts only vary by about 10% with the state of the tide (Walters and Dunlap, 1987; Krimmel and Vaughn, 1987; Walters, 1989) and calving speeds are unaffected. Calving events are statistically uncorrelated with the state of the tide (Meier and others, 1980; Brown and others, 1982).

The role of tides and weather has been stressed also by Holdsworth (1974) in his description of the twentieth century behaviour of the Erebus Glacier Tongue in McMurdo Sound, Antarctica. Its recent history is suggestive of a quasi-

cyclic pattern, with steady advance being followed by a major calving event during extreme weather conditions when the stresses and flexure within the extended tongue exceed a critical value. Holdsworth and Glynn (1981), having studied a number of Antarctic ice tongues, show that tidal flexure is not the key mechanism, but that tidally-induced stresses are involved. They propose a vibration-calving mechanism for unconfined ice floating in shallow water. Ocean wave energy, (either direct or muted through sea ice), is filtered through the glacier. When the dominant frequency coincides with one of the natural glacier frequencies, resonant motion begins, causing sustained cyclic bending stresses, and these in turn result in crack propagation and fatigue failure. They suggest that this acts with other mechanisms which cause tensile stresses, but that the vibration mechanism is the trigger which raises the resultant stresses to the point at which fracture occurs.

In the case of lakes, the cyclic level changes are seasonal rather than diurnal, but the amplitudes of spring lake rises can be greater than tidal amplitudes. Holdsworth (1973) observed calving into Generator Lake on Baffin Island which was caused by a rise of 2 m in 17 days. Theakstone (1989) describes the complex and changing stress and velocity fields which were operating in 1986 in the terminal section of Austerdalsisen in Norway when a floating ice tongue extended out into Austerdalsvatnet from the base of an icefall. Buoyancy forces bent the tongue upwards at the distal end, but at the proximal end these were counteracted by downward flow from the icefall, so that the tongue had a

65

reverse slope. Warping caused by water level change or buoyancy change activated cracks associated with thinning and stretching of the ice due to acceleration; these cracks tended to work through the tongue (cf. Hughes, 1983). At the beginning of the summer of 1987, following the rapid spring rise of the lake, the entire tongue calved off in a single event and broke up. The glacier then ended at the base of the icefall.

Calving mechanisms are varied and complex, involving a wide range of cooperating and counteracting forces. The most important calving mechanisms and forces, however, can be summarised as follows:

Internal: Frontal bending stress - unequal hydrostatic pressures.

Slab-calving - horizontal pulling force, bending creep, shear rupture.

Critical fracture strain - crack propagation, crevasse fields, crevasse meetings.

Longitudinal strain.

External: Tidal flexure and buoyancy forces.

Extreme weather conditions.

2.1.iii. Controls on Calving Speed and Calving Flux.

2.1.iii. a) Grounded Calving Fronts.

At grounded calving fronts, the rate and magnitude of calving appears to be determined by combinations of the following:

- cross-sectional area of the channel (especially water depth).
- ice velocities.
- subglacial runoff.
- ice thickness.
- water temperatures.
- buoyancy forces.
- accumulated strain.

Water depth. A relationship between calving rate and water depth at the terminus has been noted for many years. Andrews (1903) observed that as soon as the terminus of Muir Glacier in Glacier Bay, Alaska, retreated from shallow water at the mouth of Muir Inlet, rapid calving ensued in the deeper water. These views were supported by, amongst others, Cooper (1923), Field (1937,1947), Goldthwait and others (1963) and by Post and LaChapelle (1971). Some quantification of this relationship has been provided by Post (1975,1980a - d), and by Meier and others (1980) from studies of 52 calving tidewater glaciers in Alaska. They found that:

i) All glaciers with stable, advancing, or slowly retreating termini ended in water less than 80 m deep.

ii) All rapidly retreating glaciers end in water deeper than 80 m, and that deeper water led to faster retreat.

iii) Retreat rate is influenced by channel shape. Shallows,

narrows, or sharp bends all reduce the rate of retreat, or causes a stillstand. However, if the water depth exceeds 80 m, even during the halt the iceberg discharge exceeds ice supply, and therefore the ice surface continues to lower until frontal retreat commences once more.

iv) There is no strong evidence in Alaska for any influence of accumulated strain, ice speed, water temperature, or tidal state on calving.

Similar relationships between terminal behaviour and water depth have been established elsewhere (eg. Dowdeswell, 1989, - Svalbard), suggesting that a depth of 70 - 80 m is a universally critical threshold for calving tidewater fronts.

Meier and others (1980) proposed a calving relation, refined by Brown and others (1982,1983), showing that one of the best fits to the data is one of the simplest possible calving relations:

$$v_c = c h_w \qquad\qquad \text{(Equation 1)}$$

where v_c is the calving speed, h_w is the water depth averaged over the width of the terminus, and c is a calving coefficient, the best estimate of which is $27.1 \pm 2\ a^{-1}$. The best-fit relation is very close to being linear. The two-variable linear relation shows that calving is approximately zero when water depth is zero, which further supports this simple one-coefficient, one-independent-variable calving relation.

Pelto (1987), Clapperton and others (1989), and Pelto and Warren (in press) have proposed an alternative, though still linear, relationship, based on annually averaged
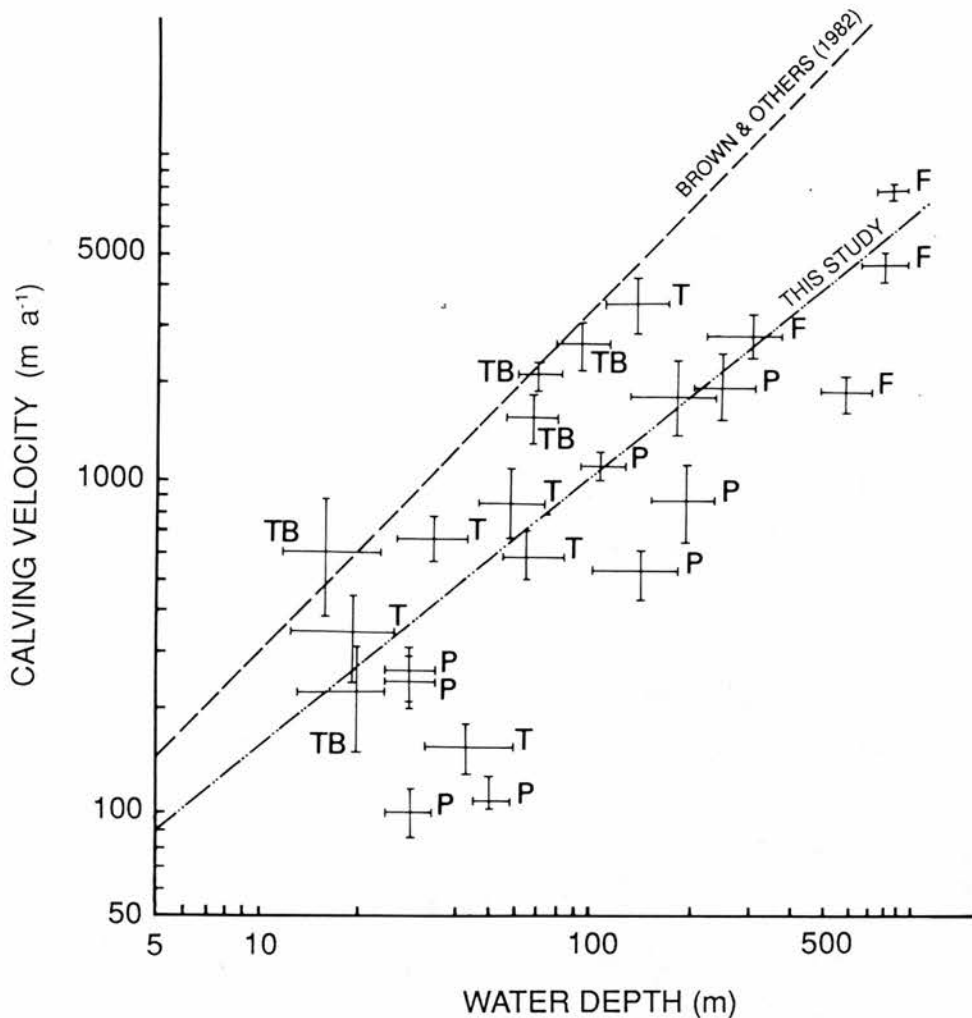
FIGURE 2.4. Relationship of calving speed to water depth for 23 tidewater glaciers in Alaska, West Greenland, Svalbard, and Antarctica. (After Pelto and Warren, in press). The relationship derived by Brown and others (1982) is based on summer-only calving speeds, that by Pelto and Warren on annual calving speeds. The vertical lines represent the error bars in velocity determination, the horizontal bars the error in water depth measurement. T = grounded temperate glacier, P = grounded polar glacier, F = floating polar glacier. B = glaciers used in the study by Brown and others (1982).

calving data from several regions of the world rather than simply the Alaskan ablation season data of the USGS. Calving is suppressed during the winter freeze-up and this reduces the annual calving fluxes relative to extrapolations from summer-only data. The two relationships are shown in Fig. 2.4. The v_c/h_w relation derived by Pelto and Warren (in

press) using linear regression analysis is:

$$v_c = 70 + 8.33 \; h_w \qquad \text{(Equation 2)}$$

with a correlation coefficient of 0.85, and a standard deviation of $\pm$ 22%.

Attention has recently been focused on freshwater calving by the construction of water storage dams in glaciated catchments in the European Alps (Funk and Röthlisberger, 1989; Funk and Müller, unpublished) and in Norway (Hooke and others, 1989). Calving rates in lakes are also linearly correlated with water depth. However, the results collated by Funk and Röthlisberger (1989) from Europe, Alaska, and Greenland show that, in any given water depth, calving rates in freshwater are an order of magnitude lower than in tidewater. Their v_c/h_w relation is:

$$v_c = 1.9 \; h_w + 12 \qquad \text{(Equation 3)}$$

Subglacial runoff. During observation periods of less than a year, there is a strong, apparently causal relationship between calving and subglacial water discharge, such that calving speed increases follow increased activity of the basal hydrological system, either through meltwater inputs or subglacial drainage of ice-dammed lakes (Theakstone and Knudsen, 1986; Meier and others, 1980). Additions to the subglacial meltwater system can lead to increased basal sliding on scales of hours to months (Meier and Post, 1987; Meier, 1989). The calving speed increases are partly the result of these increases in ice velocity following increases in pressures in the subglacial hydraulic system (Reeh and

70

Olesen, 1986), and partly the result of flotation forces. Flotation seems to play an important role in calving because the nearer to flotation the snout is, the more rapid the calving becomes; major subglacial drainage channels lead to local flotation and local catastrophic retreat, halted only by the stability of the surrounding ice (Bindschadler and Rasmussen, 1983). Sikonia and Post (1980) found that the location of seasonal calving embayments at the snout of Columbia Glacier, Alaska, related to the position and discharge of subglacial water outlets, and suggested that the massive calving event of August 1977 was caused by the draining of Kadin Lake, 15 km up-valley.

Observations of seasonal calving fluxes by Sikonia and Post (1980), Sikonia (1982), and Brown and others (1983) led them to propose the following calving relation for seasonal variation:

$$v_c = aD^b h_u{}^c \qquad\qquad \text{(Equation 4)}$$

where v_c is the calving speed, D is the meltwater discharge, h_u is the height of the unsupported ice column, and a, b, and c are constants. $h_u = h - h_w\, p_w/p_I$, where h = glacier thickness and p_w and p_I are water density and ice density respectively. Some empirical support for this relationship is provided by the retreat pattern of Columbia Glacier during the early 1980s; a finite element glacier dynamics model based on these relations (Sikonia, 1982) closely predicted the subsequent behaviour of the glacier (Meier and others, 1985). Also, the data gathered by Pelto and Warren (in press) confirms that there is a direct relationship,

71

albeit a weak one, between calving velocity and the percentage of the glacier that is buoyant at the calving front (Fig. 2.5).
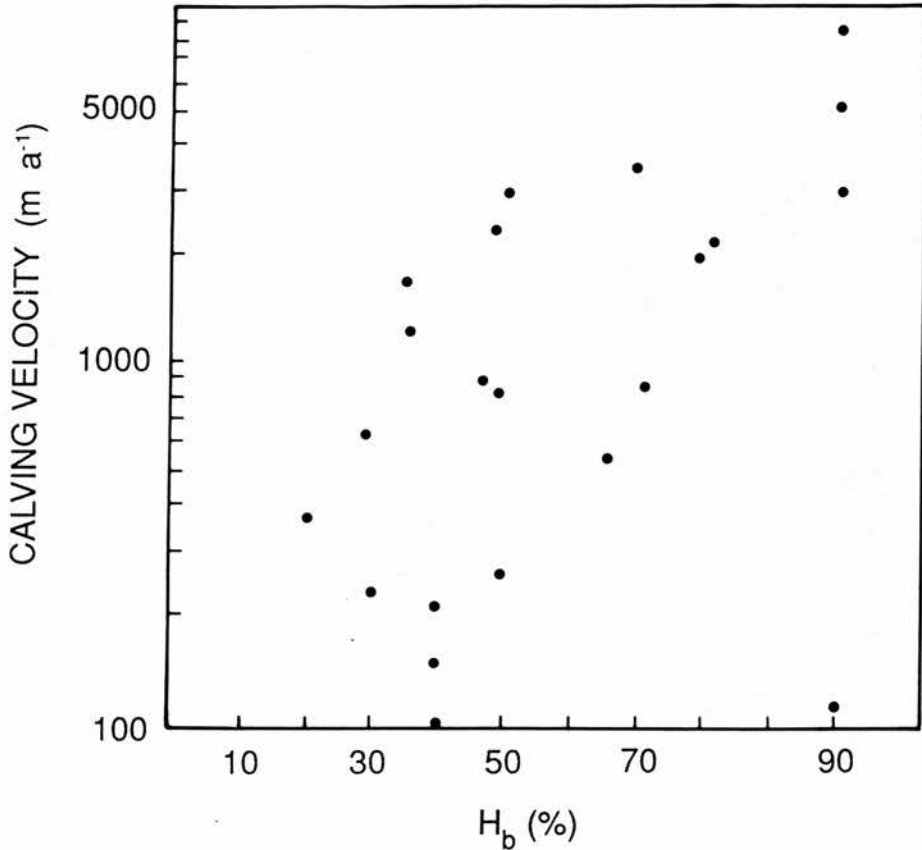
FIGURE 2.5. The relationship between calving velocity and the percentage of the glacier thickness that is buoyant (H_b) at the calving front, for the same glaciers as in Fig. 2.4. (After Pelto and Warren, in press). A weak direct relationship exists.

Subglacial meltwater discharges may also be important in indirect ways. The rising plume of water in front of the ice cliff will increase the turbulent transfer of heat in the proglacial water body, raising the temperature of water in the boundary layer near the ice cliff, and thus increasing melt rates. Also, subglacial water pressures drop in the

72

autumn with decreasing meltwater discharge; this will increase basal drag and promote differential flow within the glacier. This in turn may lead to a greater degree of oversteepening at the ice front, increased stresses and more rapid calving.

Equation 2 provides a good working relationship for annual or multi-annual data, while Equation 3 fits seasonal calving patterns well. There does not yet exist a general calving law which accurately fits both the detailed and general behaviour of grounded calving fronts. However, a broadly applicable calving flux relation is proposed by Pelto (1987) as follows:

$$Q_c = W_T \times h_T \times V_c \qquad\qquad \text{(Equation 5)}$$

where Q_c = calving flux,
W_T = the mean glacier width,
h_T = the mean glacier thickness, and
V_c = the mean calving velocity.

Ice velocities. Calving speed is strongly correlated with the width- and depth-averaged annual ice velocity for the simple reason that the equilibrium position of a calving front is where the calving speed equals the ice speed. The issue is whether high ice velocities promote rapid calving by some physical means, and this remains an unresolved question. Both ice velocities and calving rates are highest in autumn (Meier and others, 1985b), but each could be a cause or a consequence of the other.

Ice thickness/ ice cliff height. It is likely that calving rates might increase with increasing cliff height because

73

shear stresses along failure planes must increase with greater ice thicknesses, whether those planes are dipping down-glacier, as described by Hughes and Nakagawa (1989) and in Glacier Bay, Alaska, by Hooke (unpublished), or up-glacier, as proposed by Jania (1988), Iken (1977) and Funk and Müller (unpublished). Also, greater ice thicknesses will promote greater differential flow within the glacier, causing oversteepening and an increase in stresses in the terminus. No clear evidence linking ice thicknesses with calving rates has yet been published, but this may simply be because the majority of calving data sets describe Alaskan glaciers (notably the data gathered by Brown and others, 1982), and cliff heights are remarkably uniform in these data.

Water temperature. The highest temperatures in proglacial water bodies tend to occur in autumn. This is the season in which the most vigorous calving activity is regularly observed at Columbia Glacier (Meier and others, 1985b), and so there may be a dependence of calving on water temperature. However, the nature of this dependency is not clear, because melt rates at calving fronts are thought to be an order of magnitude lower than the calving speed (Hooke, unpublished). Modelling has shown that melt rates at ice fronts are sensitive to rates of meltwater upwelling, but has also confirmed that, even at high upwelling rates of 0.05 m s^{-1} with water temperatures above 3°C, ice losses due to melting are insignificant compared to those due to calving (Dowdeswell and Murray, 1990). Therefore, if this relationship is confirmed, there must be some as yet unknown enhancement mechanism whereby melting promotes calving. It

may be that this enhancement simply consists of an increase in the oversteepening of the calving face, thereby raising shear stresses to the critical point at which failure occurs. **Buoyancy forces.** These must be the principal forces responsible for submarine calving. However, the stresses involved are small relative to the subaerial forces acting on the cliff above the waterline, so buoyancy must operate in tandem with other forces to cause such calving events. In order for the observed vertical or even undercut ice cliffs to develop, therefore, the submerged part of the face must be weakened by bottom crevasses, melting, or the development of wave-carved notches.

Accumulated strain. Higher longitudinal strain rates or accumulated strain might be expected to weaken the ice and render it more susceptible to failure. High strain rates, and great variability in strain rates over short distances, have been observed at Columbia Glacier (Meier and others, 1985a). Meier (pers. comm., 1990) suspects that this may prove to be a more important factor in calving than has hitherto been realised, and calculations by Hooke (unpublished, Fig. 3) suggest that there is a correlation between the strain rate in the lower 500 m of the glacier and calving speed.

The mechanistic explanation for the relationship between calving rates and water depth remains unclear, and so it is also not yet apparent why there is a systematic difference between tidewater and freshwater rates. Hooke (unpublished) considers that the most promising working hypothesis that

75

might explain the v_c/h_w relationship is that calving rates are dependent principally on the degree of oversteepening of the ice face, and that this oversteepening is enhanced by increased differential flow within the ice due to increased cliff height or increased basal drag, and by increased melting on the submerged part of the cliff due to high water temperatures or increased circulation. In a recent analysis, Funk and Müller (unpublished) found that maximum tensile stress a short distance behind the calving front quadrupled with increasing inclination of the terminal cliff, so the degree of oversteepening does seem to be emerging as an important factor.

Retreat rates. A good approximation of calving speeds can be obtained from the water-depth relationships, but they cannot
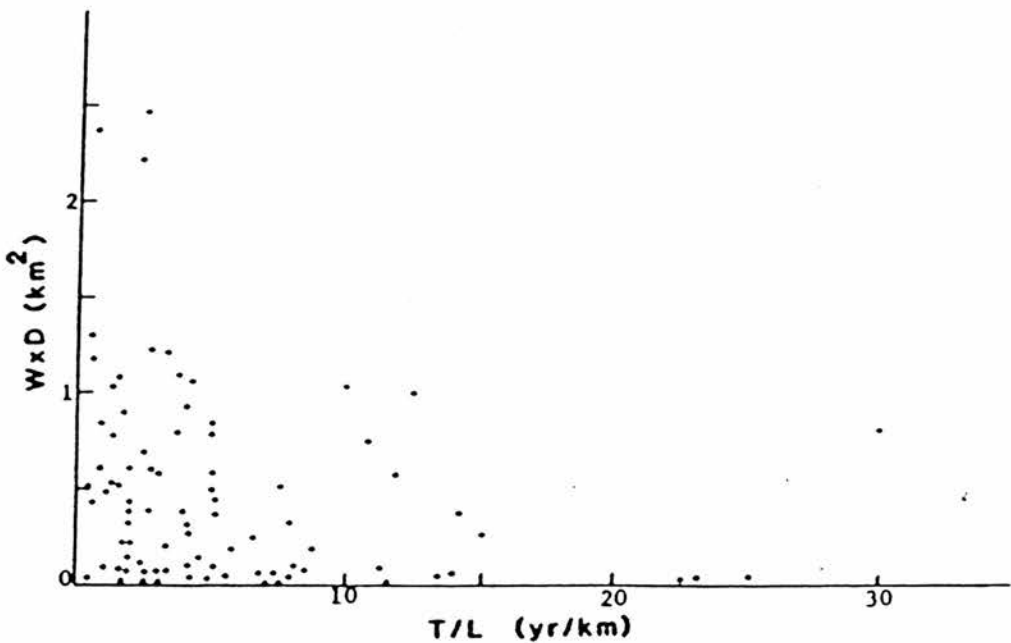
FIGURE 2.6. Relationship of retreat rates of Alaskan tidewater glaciers to the cross-sectional area of a channel. (After Powell, 1984). Average width (W) by depth (D) of fjord over a retreat period (T) averaged from seismic profiles versus the inverse of the retreat rate (T/L) during the period.

be used accurately to predict retreat rates because of the complexities introduced by changing velocity fields. A plot of retreat rates against channel cross-sectional area (Fig. 2.6) does not reflect the relationship of Equation 2 (Powell, 1984). Only some maximum and minimum rates of retreat show a relationship to the control of channel geometry. Mayo and others (1977) found, for instance, that the retreat rates of Portage Glacier in Alaska did not directly correspond to water depth at lake depths less than 70 m. The fact that the calving speed relationship of Equation 2 cannot be used accurately to predict retreat rates can be explained by consideration of the continuity equation for a tidewater terminus (Meier and others, 1980):

$$SX = Q - Q_c \qquad \text{(Equation 6)}$$

where S = the glacier's cross-sectional area,

X = the mean rate of change in time of the terminus position
    (positive when advancing)

Q = the glacier flux in m^3 a^{-1}, and

Q_c = the calving flux in m^3 a^{-1}.

Retreat rates are governed by the balance between glacier velocity and calving speed. As a terminus moves into deep water, glacier flow accelerates (Post and LaChapelle, 1971). Therefore, relatively, calving speed must increase dramatically to cause terminus retreat. If the terminus then encounters a floor threshold, it is a moot point whether variables affecting glacier velocity or those affecting calving speed are the most altered (Powell, 1984). Velocity may decrease with thickening because the overburden pressure

77

and hence the basal friction is increased, and <u>vice versa</u> (Krimmel, 1987). Ice velocities at the termini of calving glaciers vary on semi-diurnal, diurnal, and low frequency (days) timescales in accordance with the variable activity of the subglacial hydrological system and the tides (Krimmel and Vaughn, 1987; Walters and Dunlap, 1987; Walters, 1989; Meier, 1989) and this necessarily affects the short-term rates of frontal change.

It appears, therefore, that Equation 2 is a necessary but not a sufficient condition: if water depth changes, calving speed is likely to change, but the inverse is not necessarily the case. Furthermore, minor or short-term changes in water depth may not alter the calving speed. In detail, neither calving speeds nor retreat rates are directly controlled by cross-sectional area, even though in a general way, for major or long-term changes in water depth, they are.

2.1.iii. b) Floating Calving Fronts.

Theoretically, a v_c/h_w relation is uncorrectable for floating termini, since it would predict increasing calving rates with increasing depth of water below the ice front, and this is unreasonable (Brown and others, 1982). Once a glacier is afloat, that part of the calving rate which is due to buoyancy forces is 'fixed' at a maximum, and variations in calving rate and calving flux are controlled by factors other than water depth. However, the data on which Fig. 2.4 are based suggest that a v_c/h_w relationship <u>does</u> exist, although it cannot be a causal one. This may be a spurious result, due to limited data.

In addition to the factors discussed above, two other

78

controls which are of particular importance at floating ice fronts operate to regulate calving speed and calving flux, namely ice temperatures, and freezing of the proglacial water body. The importance of the ice temperature was noted by Reeh (1968). Fracture criteria vary with temperature, effective shear stress, mean normal stress and the time during which the stresses act. Generally, the thinner and colder a floating glacier is, the longer will be the intervals between calvings. The size of icebergs and the frequency of calving episodes depends on the thickness, temperature and density of the glacier ice; therefore the mass lost through calving will depend on these variables, together with the lesser influence of glacier width. Reeh came to these conclusions from theoretical considerations, but found observational support for his ideas from the contrasts between the nature of calving in Antartica and North Greenland on the one hand, and the warmer and thicker glaciers of the Disko and Umanak regions of central West Greenland, where calving is substantial and frequent. The analyses of Holdsworth (1969) suggest that the bending strength of cold glacier ice may exceed 15 bars.

Several influences combine to reduce calving in the winter. The lower atmospheric temperatures reduce the ice temperatures near the surface, and the glacial hydrologic system is usually shut off, reducing basal water pressures and hence ice velocities. Probably the most important restriction, however, is a very physical one - the freezing of the fjord or lake into which the glacier calves. This has been noted in many places. In Svalbard, for example, fast-

ice in the fjords both restricts winter calving by protecting the ice cliffs from undercutting and wave stress, and also curtails the movement of icebergs that are calved in the winter; these are then released as a spring pulse (Dowdeswell, 1989). Catastrophic spring calving of floating glacier tongues in West Greenland may be associated with this (Williams and Birnie, unpublished). The shear strength of the glacier tongue is probably lowest at the time when the sea ice breaks up, because it will have undergone several hundred tidal flexure cycles (cf. Lingle and others, 1981) during the winter while it is effectively confined by the sea ice. Furthermore, it is protected to a large extent from ocean swell by the presence of the sea ice. For these reasons the tongue may be highly susceptible to bending and heaving when the sea ice breaks up.

Such influences cause there to be broad regional contrasts in calving rates and mechanisms, but do not introduce the strong differences of ice front oscillation which topographic factors produce at neighbouring grounded ice fronts. Patterns of frontal change of floating calving fronts are governed less by topographic influences than are those of grounded calving glaciers. Discounting the seasonal variability, the calving speed at a floating front is at a 'constant maximum', and the glacier dynamics are not subject to the topographic thresholds of stability associated with depth changes which are so crucial at the fronts of grounded glaciers. Consequently, the frontal oscillations are more closely governed by changes in ice velocity and ice supply. Topography does remain influential, though, because rates of

frontal change are controlled in part by channel width through its control on calving flux.

Calving speeds at both grounded and floating ice fronts show much greater seasonal variation than do ice velocities. This causes an annual pattern of frontal advance and retreat to be superimposed on the longer-term trend. Columbia Glacier is experiencing seasonal advances of c. 0.5 km even during catastrophic retreat (Krimmel, 1987), and the magnitude of the seasonal oscillations have been increasing as retreat has progressed (Krimmel and Vaughn, 1987).

A distinctive pattern of frontal behaviour has been observed in the fjords of North and North-East Greenland. Ice velocities and ice temperatures are low, giving the floating ice tongues of these glaciers greater strength relative to the warm, fast-flowing glaciers further south. Calving events at such glaciers appear to be of low frequency but great magnitude (Dunbar, 1978; Higgins, 1988; Weidick, pers. comm., 1990). Semi-permanent sea ice in the fjords can maintain the integrity of the floating glacier fronts for one or two decades. A summer with substantial sea ice break-up will then lead to a sudden retreat of the front, but this is unrelated to mass balance status. Consequently:

> For floating glacier tongues, the position of the glacier terminus is not a reliable indicator of advance or retreat (Higgins, 1988, p.103).

2.1.iii. c) Feedbacks.

The mechanisms and rates of iceberg calving are further complicated by interactions and feedback loops. This is well exemplified by the case of Jakobshavns Isbrae. The

81

enormous 'pulling power' (cf. Weertman, 1957) of this glacier seems to depend on high rates of summer melting over the heavily crevassed surfaces, and Hughes (1986) coined the term 'the Jakobshavns Effect' to describe the relationship between surface melting and pulling power in a heavily crevassed ice stream. This effect consists of an association of positive feedbacks:

i) Surface roughness increases the surface area by a factor of about three, greatly raising its capacity for solar energy absorption.

ii) Surface melting is consequently much increased. Hughes (1986) calculates that the ice thins by 55 m a^{-1} over the 48 km^2 floating portion, and that this must either release 2 x 10^{18} cal a^{-1} if all the meltwater refreezes internally (cf. Echelmeyer and Harrison, 1989,1990), or discharge 20 km^3 of water subglacially.

iii) Extending flow: if all the surface meltwater refreezes internally, the vertical thinning requires a velocity increase of 550 m a^{-1} at the calving front. The velocity between the grounding line at the fjord headwall, and the calving front, increases from 15 m d^{-1} to 23 m d^{-1} (Lingle and others, 1981); this causes high longitudinal strain rates in extending flow.

iv) Basal uncoupling: the extending strain rate results in a thinning strain rate which has the effect of raising the basal ice at the grounding line. This combined with surface melting is a powerful means of basal uncoupling, consisting of bedrock lubrication, the elimination of pinning points, and the retreat of the grounding line. This last has

82

apparently been halted by the fjord headwall.

v) <u>Lateral uncoupling</u> from the relatively stagnant ice which lies between the shear zones and the fjord sidewalls is facilitated by thermal and strain softening within the shear zones. The floating portion is thus effectively unpinned, and acquires the enormous pulling power of an unpinned ice shelf (Weertman, 1957).

vi) <u>Rapid calving:</u> the bottom crevasses that are necessary to allow tabular icebergs to calve, can open from extending flow in the floating portion and from tidal flexure at the grounding line. Being filled with water, they can extend up to sea level and meet surface crevasses, causing rapid, large-scale calving.

All these feedbacks appear to be positive, which may account for the anomalously great glacial retreat up Jakobshavns Isfjord of 27 km between 1850 and 1964 (Weidick, 1966). The only stabilizing feature is the Ice Fjord itself; the sidewalls and headwalls prevent the 'Jakobshavns Effect' from spreading, which emphasizes the important influence of the topographic setting on the behaviour of active calving glaciers. Hughes (1986) concludes:

> Without these bedrock constraints, it is difficult to imagine what could prevent the Jakobshavns Effect...from collapsing most of the Greenland Ice Sheet. (p.48)

Calving speeds and calving fluxes are therefore controlled by the interaction of the calving glacier terminus with the topographic geometry of the trough, and also, though to a lesser degree, by the changing local environmental conditions over timescales ranging from days to centuries. Rates and patterns of frontal change are determined by the sum of calving speeds and ice velocities.

2.2. CALVING GLACIERS, TOPOGRAPHY, AND CLIMATIC CHANGE.

2.2.i. Topographic Control of Tidewater Glacier Fluctuations.

The uniquely important role of topography in controlling the oscillations of calving glaciers began to emerge in the early part of this century from observations in Alaska. While almost all the land-terminating glaciers were retreating in response to the warming climate, some of the tidewater glaciers were strongly advancing. Increasing amounts of data have revealed a wider pattern of anomalous behaviour. Of the 52 tidewater glaciers in Alaska, 7 have been advancing, 14 have been retreating and 31 have been stable in recent decades (Meier and others, 1980). These facts suggest that the oscillations of calving glaciers have little detailed relationship to climatic change. The evidence for this is fourfold:

1. Their oscillations this century have not been synchronous with each other (eg. Harvard and Yale glaciers - Field and others, unpublished) or with nearby land glaciers. Even glaciers fed from the same accumulation area have been behaving in contrasting ways. For instance, both the Mendenhall and Taku Glaciers near Juneau flow from the Juneau Icefield, but the former (Fig. 2.7) has been steadily retreating this century, while the latter (Fig. 2.8) has been advancing continuously since 1890 (Mayo, pers. comm.).

2. Advance or retreat trends have borne little or no relationship to accumulation trends, so that some glaciers are to be greatly out of equilibrium. For instance, the Hubbard Glacier near Yakutat (Fig. 2.9) currently has an AAR of 0.95 (Mayo, 1988).

FIGURE 2.7. The retreating Mendenhall Glacier, a lake-calving
outlet of the Juneau Icefield, Alaska, September 1989.

FIGURE 2.8.   The advancing Taku Glacier, Alaska, September
1989.   It used to calve into the tidal Taku River, but
proglacial sedimentation has now largely arrested
calving activity.

86

FIGURE 2.9. A calving event at the Hubbard Glacier, Alaska, September 1989.

3. Glacier area and volume change enormously between advanced and retracted positions.

4. Various calving termini have made different numbers of advances and retreats during Neoglacial time, from one to three or more.

Two important elements of topographic control can be distinguished, the effect of fjord configuration, and the role of varying water depth.

2.2.i. a) The Effect of Fjord Configuration.

It has long been recognized that the shape of the fjord influences the locations of stillstands through determining the total cross-sectional area of the glacier exposed to the sea (Reid, 1892).

> The rapid retreat of ice from one promontory or constriction in the channel to the next has been repeatedly noted. Headlands and channel constrictions provide anchor points where the glacier terminus often maintains a tenuous stability, sometimes for several years. Once the ice retreats from these points, breakup is catastrophic, and the glacier shrinks rapidly until the next constriction in the channel is reached. (Post and LaChapelle, 1971, p.77)

The first theoretical explanation of this kind of behaviour was presented by John Mercer in 1961. He highlighted the importance of fjord mouths, fjord bifurcations, widening points, and places where the gradient of the side slopes changed, and went on to show theoretically that:

> A glacier in an ideal fjord of constant width, when responding to a change in the firn limit [ELA], can only reach standstill in such positions, and will continue to advance or retreat between breaks in the width with no further change in the firn limit. Such a glacier may therefore be anomalous in its behaviour compared to glaciers on dry land in the vicinity. (Mercer, 1961, p.850)

88
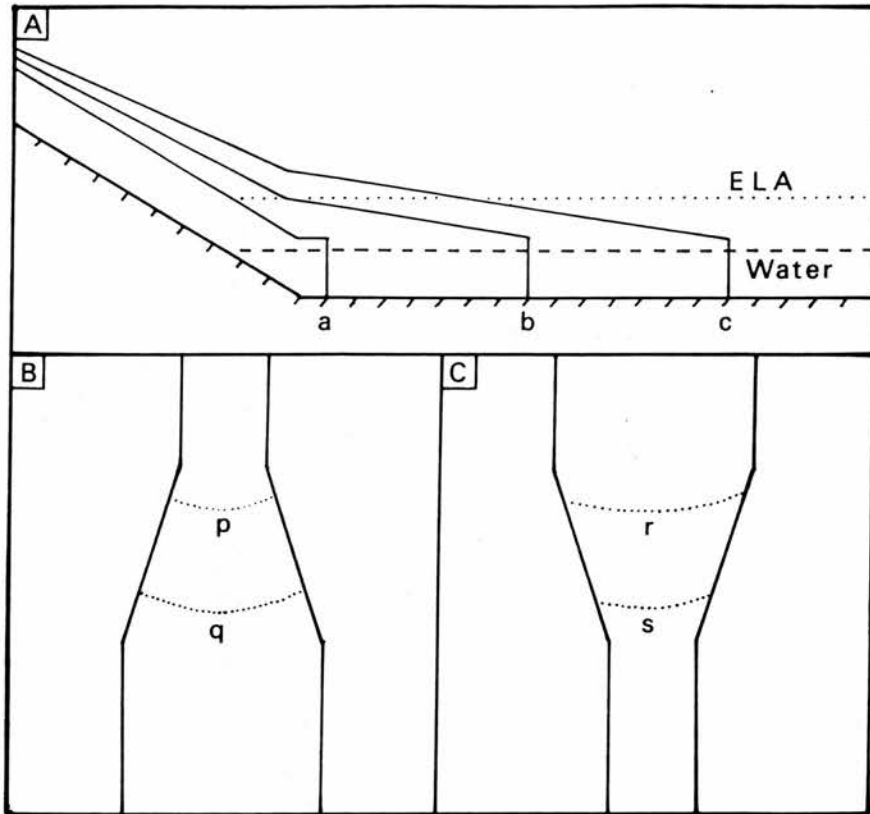
FIGURE 2.10. A) Response of an idealized tidewater glacier to a fall in the ELA. (After Mercer, 1961). Once an advancing tidewater glacier has passed point b), further advance will progressively increase the accumulation area while the ablation area remains unchanged. The advance will therefore become stronger unless or until the ablation area can expand.

B) Fjord configuration conducive to stability. In such a fjord, a glacier can reach equilibrium, since if it advances from p to q, it will increase its wastage, and if receding from q to p it will decrease its wastage.

C) Fjord configuration conducive to instability. In such a fjord, a glacier advancing from r to s will decrease its wastage and accelerate its advance; if receding from s, it will increase its wastage and accelerate its recession.

If a thickening of the glacier causes the gently sloping

fjord section to intersect the ELA, then further advance

along a fjord of equal width will simply increase the AAR and

strengthen the advance (Fig. 2.10.a). The only way in which

the glacier can increase its mass loss to counterbalance this

increasing ice supply is by advancing to a location where the calving front can increase; the four situations Mercer highlighted are such locations. Similarly, a glacier retreating up a fjord can only achieve a measure of stability at these places, and may in fact not be able to overcome the effects of thinning and mass loss until it reaches the head of the fjord, where the ice surface gradients increase and calving losses will decrease or cease.

Mercer also showed that the fjord geometry can either maximise or nullify the response of the terminus to a shift in the ELA (Fig. 2.10.b & c). Thus a small positive shift in mass balance can be greatly amplified if the glacier is advancing down a narrowing fjord. Conversely, a small rise in the ELA, which would only cause a small retreat of a land glacier, may result in catastrophic retreat of a tidewater glacier if the terminus begins to move back into a widening fjord, because of the greatly increased calving losses.

These mechanisms can all be triggered by climatic change, but the speed and magnitude of the glacial response varies greatly, and the response can continue after the climatic equilibrium has been restored or even despite a reverse change.

2.2.i. b) The role of varying water depth.

Mercer's theorizing was based on the concept of a fjord of constant width and constant depth. The extensive subsequent research on the tidewater glaciers of Alaska by the U.S. Geological Survey, especially the seminal work of Austin Post (1975,1980a - e), has provided empirical

confirmation of these ideas but has also revealed the linear water depth/calving speed relationship (Brown and others, 1982; 2.1.iii) which explains the catastrophic rates of recession which follow the retreat of a glacier into deep water.

The initiation of this kind of retreat has been extensively documented at Columbia Glacier near Valdez. Columbia Glacier was unique in the late 1970s because it was the only glacier in Alaska which was still extended to its Neoglacial maximum position, grounded in shallow water (2 - 30 m) on a morainic shoal. The glacial histories of other Alaskan glaciers suggested that catastrophic retreat was imminent (Post, 1975) and this was specifically predicted by Meier and others (1980). Behind the terminal shoal, the bed falls to 400 m below sea level, and predictions suggested rapid break-up of the glacier once the snout retreated back off the shoal (Bindschadler and Rasmussen, 1983; Sikonia, 1980). These have been largely borne out by the subsequent behaviour of the glacier (Meier and others, 1985b; Krimmel, 1987; Meier, 1989). Since a calving front cannot achieve stability in deep water, the terminus may well continue its retreat to the head of the fjord.

Because of the efficiency of mass loss of the calving process, calving glaciers can retreat extremely rapidly (at rates of 500 - 10,000 m a^{-1} - Mann, 1986) but, for the same reason, their advances are slow, typically 10 - 40 m a^{-1}. A calving snout cannot advance into deep water because the progressive increases in calving losses counteract the increasing ice supply. It is thus very unusual for a glacier

to advance beyond a fjord mouth onto the deepening, unconfined continental shelf, or, in any situation, to advance into water deeper than 150 m. Within a fjord, however:

> Glaciers...can advance asynchronously and over long distances due to the forward movement of protective submarine terminal moraines. (Post, 1975, p.6)

If the sediment supply is great enough, the glacier can build up a protective shoal which reduces the effective depth at the terminus, thus permitting advance (Cooper, 1937; Powell, 1981). The glacier moves the moraine down the fjord by eroding material from the back and depositing on the distal slope. The rate of advance is therefore controlled primarily by the rate of sediment supply and also by the ice velocities.

2.2.ii. Tidewater Glacier Cycles.

The Holocene record in Alaska clearly shows that there have been large scale advances and retreats of tidewater glaciers that have often covered the full extent of the fjords between the sea and the mountains, well exemplified by the Holocene oscillations of Hubbard Glacier in Yakutat Bay (Mayo, 1988). Such oscillations have been asynchronous both in relation to other calving glaciers and to land glaciers, and often appear not to relate to climatic trends. It has therefore been proposed that tidewater glaciers exhibit cyclical behaviour that is determined by glacio-dynamic mechanisms (Mann, 1986; Mayo, 1988). The trigger initiating the cycle may be climatic, such as a rise in the ELA, or non-climatic, such as an outburst flood, a rise in relative sea level, or even seismic activity (Tarr and Martin, 1912), but once the cycle is underway, internal mechanisms override external forcing.

This cycle is illustrated in Fig. 2.11. Most fjords have a concave long-profile resulting from efficient glacial excavation (Crary, 1966; Barnes, 1987). This means that once a glacier has been destabilized from its terminal moraine bank, it will be subject to deep-water calving until it has retreated to a location near the head of the fjord where the depth begins to decrease. The strength of the water-depth - calving relationship is such that the terminus will continue to retreat despite the increasing AAR. Furthermore, during the retreat, the terminus is insensitive to climate because the rapid calving overwhelms any potential response of the terminus to changes in the ELA.
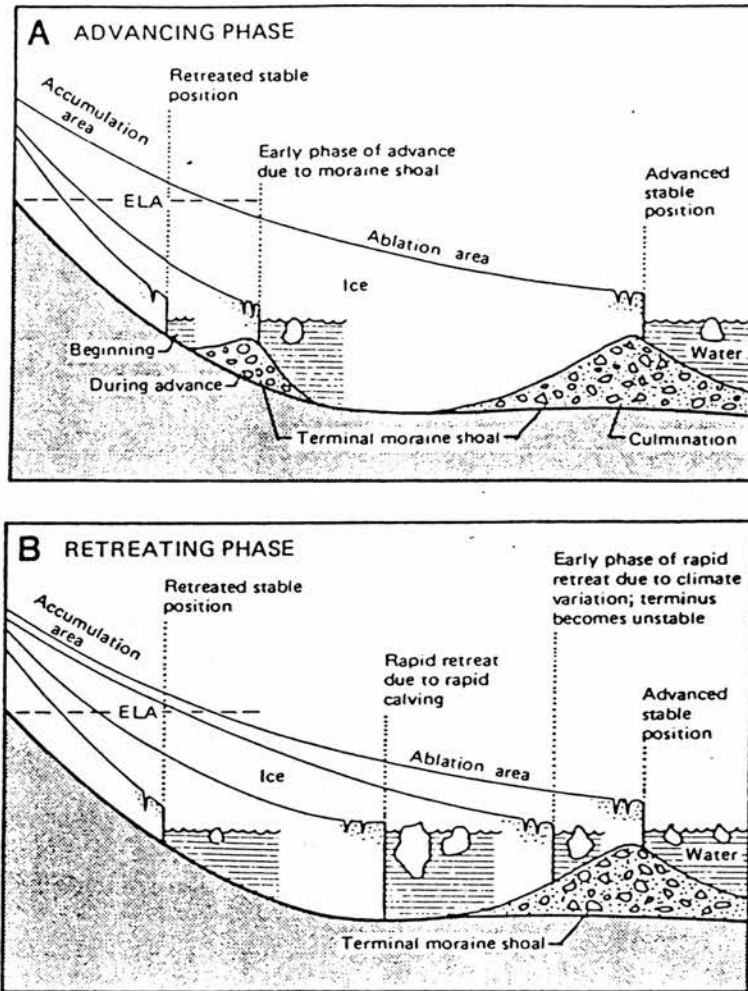
FIGURE 2.11. Tidewater glacier cycles. (After Mayo, 1988).
    For explanation, see text.

When the glacier snout once again achieves stability in
shallow water, the glacier system will be out of equilibrium
with the climate, and may have an extremely high AAR.
(Hubbard Glacier has an AAR of 0.95 at present - Mayo, 1988).
Compensatory advance on a morainic shoal therefore begins.
During this advance the glacier is also insensitive to
climate because of the highly positive mass balance which is
capable of damping the effects of most rises of the ELA, and
of masking the effects of a lowered ELA.   Even when the
terminus has regained a stable position, the AAR may remain

above an equilibrium value if the terminus faces a steeply increasing ablation gradient, such as at the mouth of a fjord out into the open ocean. Here, increasing rates of ice supply will increase the calving flux but will not alter the terminus position significantly.

For the duration of the cycle the glacier's behaviour does not reflect the prevailing climate, and the glacier is insensitive to climatic change. Mann (1986) has described the length of this cycle as 'the insensitive period'. In the Lituya glacier system in Alaska, for example, the cycle takes about 1000 years if the advance continues to the fjord mouth. From a palaeoclimatic point of view, the significance of the insensitive period depends on its duration relative to:

    i)   the frequency of major shifts of the ELA, and

    ii) the precision of the geological record.

The work of Mayo and others (1977) shows that 'tidewater glacier cycles' are not limited to glaciers terminating in tidewater, but can also apply to glaciers calving into lakes. Portage Glacier in Alaska used to extend to tidewater, but its own sedimentation has separated it from the sea. It now calves into a lake up to 200 m deep. Many of the features of the cycle have been observed, including slow, moraine-shoal advance, amplified response to climate, and deep water recession at rates of 140 - 160 m a^{-1}. There are hundreds of freshwater calving glaciers in Alaska (Krimmel and Meier, 1989), between 20 and 30 of which calve into large lakes (Post and Mayo, 1971), but the fluctuations of most of these remain unknown. The Portage Glacier example, and the case of the Crillon Glacier in Glacier Bay (Mayo, pers.comm., 1989),

suggest that the cycle may ultimately be self-arresting in areas of high sediment budget, such as Alaska, as sedimentation progressively fills in the fjords and lakes. It is therefore probable that some glaciers which now terminate on land have been subject to calving cycles in the past, prior to the infilling of their troughs.

In summary, it is possible to make some general statements about the important controls on the fluctuations of calving glaciers:

i)   The **calving speed** is controlled by water depth.

ii)   The **calving flux** is a product of the calving speed and the cross-sectional area of the glacier terminus, which is itself controlled by topography, (unless the snout is floating in which case only the width is controlled by the topography).

iii)   The **trigger for advance and retreat** is most usually climatic.   However, any event which has the potential to destabilise a calving front from its protective moraine shoal or from a rock bar may trigger a rapid or even catastrophic retreat.   Such events include earthquakes, tsunamis, jökulhlaups, and, in the longer term, sea level rise. Advance may be compensatory, following a rapid, deep-water retreat, or it may follow the exceeding of critical thresholds of sediment accumulation.

iv)   The **rate of advance** is controlled primarily by the sediment flux, but also by the mass flux of the glacier.

v) The **rate of retreat** is a function of the balance between calving speed and mass flux.

vi) The **distance of advance and retreat** and the **location of stillstands** are controlled by the geometry of the trough in which the glacier is moving and that of the drainage basin.

vii) **Dynamic adjustment** following the cessation of calving (either due to retreat onto land, or to proglacial sedimentation) will result in stillstands or small readvances unrelated to climate or mass balance status.

2.2.iii. Wider Geographical Applicability.

Very few detailed case studies of calving glacier oscillations have been carried out since the development of the 'calving cycle' theories. The examinations of the Lituya system (Mann, 1986) and the behaviour of Portage Glacier (Mayo and others, 1977) are exceptions. Outside Alaska, these ideas have been specifically applied in only a few places. There is, nevertheless, empirical evidence from many parts of the world to suggest that the relationships established in Alaska hold for calving glaciers in most, if not all, situations. Evidence comes from three sources:

a) The historic fluctuations of calving icecap outlets.

b) The inferred behaviour of the margins of the shrinking Pleistocene ice sheets.

c) Modelling studies.

2.2.iii. a) Historic Behaviour of Calving Icecap Outlets.

- **NORWAY:** The glacier dynamics of Austerdalsisen, a lake-calving outlet glacier of the Svartisen icecap, seem to be driven by calving (Theakstone, 1989,1990; Knudsen and Theakstone, 1981; Theakstone and Knudsen, 1986). The water-contact parts of the margin have retreated much more rapidly than land-based parts; calving retreat rates of 5200 m a^{-1} were observed for a short period in 1982 following the disconnection of part of the terminus from the lake shore. In the last part of the 19th Century, steady calving retreat of the south-east lobe in the 1870s and 1880s was followed by steady land-based retreat from the late 1890s. However, for at least seven years, between 1890 and 1897, the terminus

halted at the lake shore, seemingly a glacio-dynamic response to the change of ablation mechanism. Similarly, the retreat of the western margin halted for almost a decade following grounding in 1959. The general behaviour of the Svartisen glaciers is controlled climatically through changes in the mass balance, but, in detail, the pattern of response is moderated by the interaction of calving fronts with trough geometry.

- ICELAND: In southern Iceland, neighbouring outlet glaciers on the northern side of the Eyjafjallajökull icecap have behaved asynchronously, partly because of the contrast between lake- calving and non-calving snouts (Dugmore, 1987).

- SVALBARD: Polish and Russian research projects on the tidewater glaciers of Spitsbergen (Jania, 1988; Jania and Kolondra, 1985; Glazovsky and others, 1990), especially concerning Hans Glacier, have shown that recession rates depend sensitively on water depth at the terminus.

- PATAGONIA: From the patterns of historic fluctuations in Patagonia (Aniya and Enomoto, 1986; Aniya, 1988; Naruse and Aniya, 1990; Clapperton, 1983), it is '...evident that some water-terminating glaciers fluctuate out of phase with adjacent glaciers' (Clapperton and Sugden, 1988, p.189). The twentieth century norm has been steady retreat (Rabassa and Clapperton, 1990). However, some lake-calving glaciers have advanced (eg. Moreno Glacier in Argentina - Nichols and Miller, 1952), the tidewater Glaciar Brüggen has behaved unusually (Mercer, 1964), advancing 10 km between 1945 and 1983 to reach its Neoglacial maximum (Clapperton and Sugden, 1988), and other calving glaciers such as the San

Rafael and Cachet glaciers have shown anomalously great and rapid retreats (Nakajima, 1987). The reasons for such anomalies are not yet understood, '...but part of the explanation may rest on the particular dynamics of these calving glaciers and the topography of their basins' (Clapperton and Sugden, 1988, p.189). As in Norway, there is evidence that breaks in frontal trend follow changes in the dominant ablation mechanism; thus the Arco Glacier on the east side of the Northern Patagonian Icefield was retreating in a lake in the first part of this century, but has been stationary since 1944 when the Laguna Arco drained (Aniya and Enomoto, 1986).

- **SOUTH GEORGIA:** Clapperton and others (1989) addressed the anomaly posed by the 6.5 km Neoglacial advance of the Moraine Fjord glacier system when local glaciers advanced only a few hundred metres. They suggest that the advance in Moraine Fjord may not have climatic implications since it could have been triggered by the passing of a critical threshold in sediment accumulation which allowed the terminus to advance into deep water. Advances determined partly by sediment flux will be substantially lagged relative to land advances. They conclude that the anomalous advance can be explained in terms of a particular blend of fjord topography, sediment availability, and climatic deterioration; the extent of the advance was determined by the fjord topography.

- **NEW ZEALAND:** A most interesting example of a retreat controlled by calving dynamics is currently underway in the Southern Alps of New Zealand (Kirkbride, 1989). The terminus of the Godley Glacier was stagnant, stationary, and wasting

under a thick debris mantle throughout this century until the mid-1970s, when the coalescence of surface lakes, followed by sub-lacustrine calving, transformed the gently-sloping land-based glacier snout into a lake-calving cliff. Since then, rapid retreat totalling almost 2 km has taken place. It is very likely that the debris-laden snout of the nearby Tasman Glacier will undergo a similar transformation and retreat within the next few decades.

2.2.iii. b) Holocene Behaviour of Calving Ice Sheet Margins.

At the end of the last glacial period, the significance of calving, and the sensitivity of calving glaciers to small changes in boundary conditions, were greatly heightened in two ways:

i) Many of the outlet glaciers had unusually low gradients with ice thicknesses one fifth that of typical values for modern ice sheets (Mathews, 1974; cf. Buckley, 1969). The margins of glaciers with gentle surface slopes, whether they are calving glaciers or not, are greatly influenced by bed topography (Clayton and others, 1985).

ii) The isostatic response of the earth's crust to the loading of the ice sheets led to increased relative sea levels, and also to the ponding of large proglacial lakes behind isostatic forebulges (Peltier, 1987; Oerlemans, pers. comm., 1988; Oldale, 1988). Thus, for instance, no less than 56% of the area covered by the Laurentide Ice Sheet of ca. 8,500 BP is or was occupied by seas and lakes (Andrews, 1973).

There have been very few attempts explicitly to apply or test calving theories in relation to ice sheets, but glacial

geologic evidence which highlights the significance of cryosphere - hydrosphere interactions during the last deglaciation has been presented for the marginal areas of all the big Northern Hemisphere ice sheets:

- **THE LAURENTIDE ICE SHEET:** Many **Canadian** moraines occur just to landward or just within proglacial lake or marine sediments, suggesting an origin related to the dynamic adjustment following the cessation of calving (Andrews, 1973,1984,1987; 2.3.iii). The geomorphological records of **Baffin Island**, the **Ungava Peninsula**, and the **Boothia Peninsula** highlight the importance of calving dynamics and the influence of topographic 'pinning points' on calving termini (Lauriol and Gray, 1987; Dyke, 1983,1984). The Clyde Moraines and the Cockburn Moraines of Baffin Island, for instance, are both massive systems stretching for about 1000 km; the former loop around headlands between fjord mouths, and the latter are found almost exclusively at fjord heads, fjord narrowings, and at the narrowing of Frobisher Bay (Andrews and Ives, 1978). In **the Great Lakes region**, Mickelson and others (1981) explain large-scale, non-climatic oscillations of the Lake Michigan Lobe (totalling 850 - 1000 km of combined advance and retreat between 13,000 and 11,800 BP) as an effect of the interaction of a lake-calving front with the geometry of the Lake Michigan basin. Rapid retreat up **the St.Lawrence** occurred prior to 13,000 BP, suggesting that retreat was facilitated either by lubricating bands of marginal soft ice, or by calving at the grounding line and rapid seaward transport of icebergs, or a combination (Thomas, 1977). Kite and Lowell (1988) use geomorphological

evidence to show that the interplay between ice stream dynamics and topography were major controls of the deglaciation in **Maine and adjacent Canada**, and advocate downdraw into ice streams as one of the most important phenomena in the break-up of the Laurentide Ice Sheet.

- THE SCANDINAVIAN ICE SHEET: Lundqvist (1987), in his review of the abundant field evidence in **Scandinavia**, discusses the probable glacio-dynamics of the Younger Dryas marginal zone. He relates specific sediment suites and morphologies to glacio-dynamic conditions and changes, and highlights the key significance of water depths and isostasy in relation to the behaviour of calving fronts. Strömberg (1981) showed that the presence of deep calving bays (cf. Hoppe, 1959) during the recession in central **Sweden** can explain confusing striae patterns, while Eronen and Vesajoki (1988), supporting Ignatius and others (1980) and Punkari (1980,1984), state that the behaviour of the ice during the retreat in **Finland** was largely governed by the distribution of water masses in the marginal areas. In the fjord and island zone of **Norway,** most of the deglaciation moraines were deposited in or near the sea (Andersen, 1979), and although they can be traced for long distances, they are time-transgressive, suggesting a topographic rather than a climatic genesis (Corner, 1980). Nesje and Sejrup (1988) stress the important dynamic role of calving along the **North Sea** margins of the ice sheet, and Greene (unpublished) found evidence of topographic control of glacier behaviour during the Loch Lomond Readvance in **Western Scotland.**

- THE GREENLAND ICE SHEET: Downdraw mechanisms (2.3.i) appear to be operating today at least in central West Greenland (Hughes, 1986; Denton and others, 1986), and the interaction of calving fronts with topography was of crucial importance in controlling the retreat stages of the Greenland Ice Sheet during the late Wisconsinan and early Holocene (1.3.ii.b). For example, Funder (1985) reviews evidence that implies that sea level, rather than climatic change, was the driving force behind the early (pre-10,000 BP) deglaciation in central **West Greenland,** and describes the sea/land transition as 'a major glacio-dynamic obstacle in the deglaciation process' (p.140). In a case study of the Holocene of the inner parts of Scoresby Sund in **East Greenland,** Funder (1970,1971,1972; Fig. 1.12) specifically applied Mercer's principles and accounted for differential retreat rates up Øfjord and Fønfjord in terms of the contrasting fjord configuration. Øfjord closely approximates to Mercer's 'ideal' fjord. The fronts of the fjord glaciers during all the major halt or advance stages were located at topographic breaks - at fjord junctions and bends, at valley mouths, and places where topography changes from steep to gentle. This indicates topographic, rather than climatic, control in their location (Funder, 1989). He concludes:

> It appears that within the general framework of climatic change, ... fjord morphology was a most decisive factor during the deglaciation of the Scoresby Sund region. (Funder, 1972, p.41)

- THE NORTH ATLANTIC: Ruddiman and McIntyre's (1981) review of oceanic evidence throws much light on the mode and mechanism of the last deglaciation in regions surrounding the North Atlantic. Their data suggested that over 50% of the

total deglacial flux of meltwater and icebergs was delivered between 16,000 and 13,000 BP, implying substantial volumetric loss during a time of little areal change; this seemed to demand the operation of downdraw mechanisms. Subsequent evidence concerning mid-latitude sea surface temperatures, ice rafting, and changes in foraminifera productivity have led to a reassessment of this view (Ruddiman, 1987a), but it is still most probable that downdraw mechanisms operated effectively and that iceberg calving was the dominant ablation process during Termination Ia (14 - 12 ka BP; Mix and Ruddiman, 1985).

2.2.iii. c) Modelling Studies.

Most modeling studies have addressed global Quaternary issues, rather than the specific problem of ice sheet calving dynamics, but it has become apparent that calving is an important parameter to specify accurately if the model simulation is to be useful. Thus, for instance, ice age simulation experiments by Pollard (1982,1983,1984) have highlighted the shortcomings of previous, similar, models by Birchfield and others (1981) and Oerlemans (1980) which showed too little spectral power at the key 100,000 year interval, and did not lead to the complete disappearance of ice sheets in some interglacials. These limitations were eliminated by adding a crude parameterization of calving in proglacial and marine situations. The simulations then agreed well with the deep ocean records and their power spectra, especially over the last 400,000 years. The sensitivity of marine-based ice to sea level change and

downdraw mechanisms has also been used by many as a means of explaining the longstanding puzzle of the synchronous behaviour of ice sheets in both hemispheres, despite hemispherical contrasts in Milankovitch forcing functions (eg. Hollin, 1962; Hughes, 1975; Denton and Hughes, 1981,1983; Ruddiman and McIntyre, 1981; Denton and others, 1986; Muszynski and Birchfield, 1987).

Theoretical, empirical, and modelling results all show that calving glaciers have a unique relationship with climate. Only on the longest timescales ($> 10^3$ years) can climate be said to be of paramount significance in their fluctuations. On timescales of centuries or less, the advance and retreat behaviour of calving glaciers is a product of the interaction of glaciodynamics with topography.

2.3. ICEBERG CALVING, ICE SHEETS, AND PALAEOCLIMATE.

It is clear that the fluctuations of calving outlets of icecaps are critically influenced by topography, such that their oscillations may not reflect climatic change. Despite the evidence reviewed in 2.2.iii.b, the extent to which the climatic-decoupling effects of calving mechanisms were of significance during the collapse of the Pleistocene ice sheets remains an open question (Ruddiman, 1987b). Intuitively, it would seem that the critical role of topography, and the finely balanced mechanisms of Mercer (1961), would be much less important for ice sheets, for several reasons:

i) Catchments in ice sheets are ice-bounded, introducing the possibility of ice divide migration and drainage capture (Bindschadler, 1984; Clarke, 1987; Whillans and others, 1987).

ii) Drainage basins are much larger, which might tend to damp the sensitive ELA relationship.

iii) Terminus oscillations would have to be very great indeed in order to have the same relative impact on AARs as they do in Alaska.

iv) The abundant debris supply which controls part of the tidewater glacier cycle tends to be absent from most parts of contemporary ice sheets, ice temperatures are lower, and the ice sheets are simply much larger relative to the scale of the underlying terrain than are ice caps and mountain glacier systems.

However, during the last two decades, modelling studies and empirical research efforts have increasingly showed that

107

water-contact margins are important dynamic elements in contemporary ice sheets, and were of crucial significance to the dynamics of the last ice sheets. They suggest that the interaction between calving ice margins and topography is very important in controlling the retreat stages of ice sheets, and that advances can be caused and amplified by factors other than climate.

2.3.i. Marine Ice Sheets and Ice Streams.

Marine ice sheets (Mercer, 1968) differ from terrestrial ice sheets in having a substantial proportion of their area grounded below sea level at depths of 500 m or more (Hughes, 1987). Terrestrial ice sheets respond primarily to atmospheric forcing, whereas marine ice sheets respond also to changes in ocean temperatures, ocean circulation, and sea levels (Muszynski and Birchfield, 1987). The dynamics of marine ice sheets are fundamentally linked to relative sea levels (Thomas, 1979; McIntyre, 1985; Payne and others, 1989; Huybrechts, 1990). This link is mechanistic, operating through the impact that relative sea level has on the flow mechanics, calving rates, and grounding lines of ice streams and ice shelves. A change in relative sea level forces the grounding line to migrate, with a lowering of sea level leading to grounding line advance, and vice versa. Furthermore, ice streams have an inherent tendency to surge when relative sea levels exceed a critical value (Weertman, 1974; Hughes, 1977). Since the area of a marine ice sheet is determined chiefly by the elevation of the grounding line, (a fact first recognized by Hollin in 1962), the ice sheet area is controlled by sea level, the ELA, and calving rates (Hughes, 1987).

This linkage makes marine ice sheets uniquely vulnerable to changing environmental conditions, and also potentially unstable. Mercer (1968) was the first to perceive this unique sensitivity in the case of the West Antarctic Ice Sheet, the only existing marine ice sheet, and he later suggested (Mercer, 1978) that it might collapse in

catastrophic fashion, raising world sea levels by 6 m, with disastrous human consequences. An understanding of the mechanisms and controls which operate within and upon such ice masses is essential if we are to understand their fluctuations and the way that they both force and respond to change in the earth's climate system.

All empirical work concerning the dynamics of marine ice sheets has necessarily been concerned with the West Antarctic Ice Sheet. Marine ice sheets are composed of four important dynamic elements (Denton and Hughes, 1981):

- ice domes.
- ice streams.
- ice shelves.
- calving bays.

An **ice stream** is a morphologically and dynamically distinct river of grounded ice flowing within the main body of the ice sheet, first described by Rink (1877) in Greenland. Ice streams are the most 'energetic' elements within an ice sheet, discharging almost 90% of the inland precipitation of Antarctica (McIntyre, 1985), playing a pivotal role in the dynamics of the West Antarctic Ice Sheet (Blankenship and others, 1987), and therefore exercising the major control over its stability (Hughes, 1977; Thomas, 1979). They begin at scarp-like surface inflections or 'nick points' in the ice sheet which often, but not always, overlie bedrock sills several hundred metres in height (McIntyre, 1985). Downstream of this increase in surface slope, the mean gradients are less, there is a sharp fall in the driving stresses, and there is widespread

110

crevassing, especially at the boundaries which are marked by chaotic marginal shear zones (Shabtaie and Bentley, 1987).

The transition between sheet flow and stream flow is irregular but is marked by a change from slow or zero sliding to fast sliding, analagous to the non-surging - surging transition. The basal temperature regime must be very significant in this change in the flow regime (McIntyre, 1985), but also highly important are stress softening (Whillans and others, 1987) and basal lubrication provided by saturated, deformable sediments (Kamb, 1990; Alley and others, 1987; Rooney and others, 1987; Blankenship and others, 1987). Ice Stream B flowing into the Ross Ice Shelf is underlain by a deformable layer of porous, saturated sediment 8 m thick, and the discovery of this seems to solve what had been a great puzzle, namely, how fast sliding could be consistent with the very low driving stresses typical of ice streams of the Ross Group.

Ice streams can be as much as 500 km long from head to grounding line, and can be more than 100 km wide. Typically they flow at speeds only otherwise associated with surging glaciers. For instance, Ice Stream B is flowing at 827 m yr^{-1} (Clarke, 1987) and Byrd Glacier at 875 m yr^{-1} (McIntyre, 1985). However, puzzling exceptions exist, notably Ice Stream C which has all the features of an active ice stream but is only flowing at 5 m yr^{-1}. It has no surface crevasses but has a strong zone of crevasses buried at depths of 30 - 40 m, showing that it was active until about 250 years ago, and suggesting that ice streams may behave in a cyclical manner reminiscent of surging behaviour (Clarke,

111

1987; Whillans and others, 1987). Mass fluxes for the active streams near their grounding lines range from 9 km^3 yr^{-1} (Ice Stream F) to 37 km^3 yr^{-1} (Ice Stream B; Shabtaie and Bentley, 1987). The ice is discharged either into an ice shelf, which in turn calves tabular icebergs into the ocean, or directly into the sea at a calving bay if no ice shelf is present. There remain many important unknowns about ice streams, and there are great differences between the ice streams in East and West Antarctica, so that 'in neither the geographical nor the physical sense can an ice stream yet be defined precisely' (Bentley, 1987,p.8854).

An **ice shelf** is a body of glacier ice, dynamically continuous with the ice sheet, floating beyond the grounding line (Alley and others, 1987). Ice shelves help to stabilise the grounding line (Hughes, 1982) or 'coupling line', as it should perhaps more accurately be described (Hughes, 1977). They grow by simultaneous retreat of the grounding line and advance of the calving front, and shrink in the reverse manner. An ice shelf has a damming effect on ice stream discharge (Lingle, 1984), the extent of which varies with the constraints on the ice shelf. The magnitude of this back-pressure depends on:

i) the geometry of the embayment,

ii) the length of the ice shelf,

iii) the relative rates of grounding line movement and iceberg calving, since a retreat of the calving front may reduce the back-pressure (Muszynski and Birchfield, 1987),

iv) the presence or absence of ice rises and islands, and,

v) shelf thickness: for every 200 m by which the grounding

112

line thickness exceeds 400 m, the backpressure increases by c. 1 bar (Thomas, 1985). Thus, for example, the backpressure exerted by the 1400 m thick Ronne Ice Shelf is c. 5 bars. The distance up-stream that back-pressure is felt is determined by the ease of bed sliding; it is greatest for wide, fast glaciers with low basal shear stresses.

Ice shelves probably exist in metastable equilibrium (Hughes, 1983,1987). That is, they have a reversible response to small environmental perturbations, but an irreversible response once a critical threshold has been exceeded. The calving barrier at the edge of an ice shelf is located at a place at which ice supply is matched by calving losses, and is commonly coincident with ice rises (places where the ice is locally grounded) and islands. Shelves disintegrate when calving front retreat rates exceed grounding line retreat rates. This disintegration is aided by creep-thinning (Hughes, 1982), fracture (Hughes, 1983) and tidal flexure (Lingle and others, 1981). The critical destabilising forces affecting ice shelves are:

i). Climatic warming. Ice shelves are nearly flat, and therefore both the surface and basal equilibrium lines can migrate great distances with only small changes in the boundary conditions. A slight warming beyond a critical temperature may lift the ELA above the calving barrier and cause it to move back to somewhere near the grounding zone, radically and rapidly changing the mass balance.

ii). Oceanic warming. This can cause a major retreat of the bottom surface equilibrium line in similar fashion.

Combinations of i) and ii) cause shelf thinning and more calving (Thomas, 1985) which reduces the back-pressure, permits accelerated creep in the upstream parts of shelves, and therefore accelerated discharge into shelves.

iii). Sea level rise. Rising sea levels can float ice shelves off their pinning points, force grounding lines to retreat, and allow ice streams to surge (Thomas and Bentley, 1978).

iv). Delayed isostatic sinking. This lowers the ice surface, leading to a retreat of the ELA and consequently to a more negative mass balance due to surface melting. Furthermore, the pulling force of ice increases exponentially if the grounding line reteats downhill into an isostatically depressed subglacial basin, so this can be an important destabilising factor (Hughes, 1987).

v). Ice stream breakthrough. This can be a cause or effect of ice shelf break-up. Fistly, it may cause break-up if an ice shelf is not confined and pinned to the bedrock. If this is the case, the shelf may not be able to withstand the 'punch' of the streaming flow, and so the ice stream breaks through. (Even if it _is_ pinned, the links may be weak due to tidal flexure and shear rupture.) The shelf is then deprived of the stream input, causing an immediate negative shift in mass balance, and frontal retreat.

Secondly, as a result of combinations of i) - iv), the original steady-state calving front can be transformed into a catastrophic calving bay (Thomas, 1977) that migrates landward, bringing the calving barrier of the shelf close to the grounding line. Eventually, the stream will be able to break through, as a result of reduced back pressure and the

114

lateral uncoupling resulting from tidal flexure. The stream is then no longer contained, and can increase its velocity manyfold, rapidly drawing down the interior of the ice sheet. The calving bay follows the ice stream grounding line into the heart of the ice sheet.

Hughes (1977) dramatically concludes that:

... a relatively minor climatic fluctuation can unleash glacial dynamic processes independent of climate that cause calving bays to remorselessly carve out the living heart of a marine ice sheet. (p.43)

A disintegration scenario of this nature may recently have occurred in Pine Island Bay, Antarctica, where Thwaites and Pine Island Glaciers have punched through a confined shelf and are now surging (Hughes, 1983).

The dynamic responses of the **dome-stream-shelf system** to environmental change have been extensively considered in a series of papers by Hughes (1975,1977,1982,1983,1987,1988, unpublished) and also in many models by, for instance, Andrews and Peltier (1976), Thomas and Bentley (1978), Pollard (1982,1983,1984), Lingle (1984), Van der Veen (1985), Thomas (1977,1985), Muszynski and Birchfield (1987) and Payne and others (1989). With the single exception of Van der Veen (1985), there is a remarkable concensus about the theoretical interactions and the sequences of events involved. The Van der Veen model suggests that marine ice sheets are much more stable than is widely believed. However, part of this disparity can be explained in terms of the modelling process he used, which produces ice sheets which are inherently rather stable, and partly through the simplifications that he made; no account is taken either of calving processes or of subglacial topography, both of which have been found to be

important variables in other modelling studies.

Marine ice sheets are generally held to be inherently unstable, incorporating critical thresholds, the surpassing of which leads to dramatic advances or retreats. In particular, marine ice sheets are thought to be able to collapse very rapidly, by downdraw of ice from the interior, rapid discharge through ice streams, and high rates of iceberg calving. The concept known as downdraw, in which ice streams and ice shelves apparently 'pull' ice out of ice domes and discharge it into the oceans, has been developed by Ruddiman and McIntyre (1981), Denton and Hughes (1981,1983), Denton and others (1986), and Hughes (unpublished). The most important mechanism operating in this process is 'pulling power' (Hughes, 1987,unpublished). This is the result of two pulling forces, arising from:

i) the surface slope of grounded ice, and,

ii) the buoyancy of floating ice.

The pulling force caused by i) is zero at ice divides, is at a maximum at the inflection line where stream flow begins, and at a minimum at the grounding line where the surface slope is near zero. Pulling force caused by ii) is at a maximum at the grounding line where a non-zero surface slope coincides with zero basal traction. Pulling force increases as the square of the ice thickness at the grounding line (Weertman, 1957).

These two pulling forces combine to produce maximum pulling power at the head of the ice stream, drawing ice into the ice stream and creating converging flow. This maximises the surface slope, increasing basal traction and erosion, and

116

producing the characteristic foredeepening of ice stream channels. High erosion rates at the head of ice streams extend their channels into the ice sheet. The pulling power of an ice stream is thus able to propagate itself and reach far into an ice sheet (Hughes, 1987), enlarging the drainage basin and forcing ice divides to migrate. It may well be the primary force leading to the disintegration of marine ice sheets (Denton and Hughes, 1981; Hughes, unpublished). Downdraw mechanisms can lead to substantial volumetric decrease of ice sheets without significant areal shrinkage (Ruddiman and McIntyre, 1981; Ruddiman, 1987a). Through non-linear positive feedback mechanisms, small perturbations in the system may thus lead to a runaway process of irreversible collapse (Muszynski and Birchfield, 1987). The only stabilising influences which may halt such a 'total downdraw' scenario are bedrock sills anchoring the grounding line, or the presence of ice shelves (Thomas and Bentley, 1978).

Until recently, it was thought that these mechanisms of drawdown and disintegration operated exclusively through marine ice streams with retreating grounding lines. However, work on the Jakobshavns Isbrae ice stream in West Greenland, which is a terrestrial ice stream for over 100 km with a floating terminus only 8 km long, suggests that downdraw is not confined to marine ice streams (Denton and others, 1986). This glacier and 21 other ice streams between 69° N and 72° N, discharge 22% by volume of Greenland's accumulation, and move at speeds characteristic of glacial surges. The pulling power of Jakobshavns Isbrae itself is very large, due to its 7 km ice front, 800 m thick calving front and 900 m thick

grounding line; it 'literally pulls ice out of the Greenland Ice Sheet to produce the downdrawn interior drainage basin' (Denton and others, 1986,p.14). This is the downdraw - marine linkage termed 'the Jakobshavns Effect' (Hughes, 1986; 2.1.iii.c).

If the pulling power of ice streams is not only important for marine ice streams draining marine subglacial basins, but can be maintained even when grounding lines become anchored at fjord headwalls, the significance is considerable. This effect could have worked along with sea level rise on numerous marine ice streams draining the Northern Hemisphere ice sheets at the last termination. The Jakobshavns Effect allows the downdraw of terrestrial ice by marine mechanisms, as long as the the ice stream terminus remains afloat and thick so that the marine link is not broken. Furthermore, calving bays can occur in proglacial lakes as well as marine situations, and so elements of the downdraw process can operate if deep lakes exist at the margins of terrestrial ice sheet sectors (Denton and Hughes, 1981).

2.3.ii. Topographic Control of Ice Sheet Retreat Stages.

Implicit in this discussion is the fact that the topography over which the marine ice is flowing has a controlling impact on the way that ice responds to other environmental and dynamic influences.

> 'The areal extent of an ice sheet is determined by environmental conditions that can stabilise the ice sheet margins' (Hughes, 1987, p.188)

One of the prime stabilising influences at aquatic margins of ice sheets is the topography. Thus ice shelves provide effective buttressing for ice streams if they are pinned at islands and ice rises and confined within embayments, while the back pressure from ice shelves is much less if there are no such constrictions. The stabilising influence of such factors is reduced as shear rupture, crevassing, and the development of 'easy glide' ice fabrics progressively decouple the ice from the bedrock constraints and transform such locations into 'weak links' (Denton and Hughes, 1981).

The subglacial topography is crucial to the behaviour of the grounding line. A grounding line oscillating on a forward slope is in stable equilibrium; that on the crest of a bedrock sill is in metastable equilibrium; but once a grounding line is retreating down a reverse slope, it is in unstable equilibrium. A retreating grounding line can probably only be stabilised at bedrock sills. The bedslope is thus a critical element for determining stability (Weertman, 1974; Thomas and Bentley, 1978; Muszynski and Birchfield, 1987). Once an ice sheet margin has retreated to a shoreline location or has begun to retreat up fjord systems, the topographic control of the behaviour of the ice

119

front becomes even more marked (2.2.i).

The relative importance of topographic stabilising factors, and their persistence, will clearly be particular to the spatial context, and will depend on the stability of other elements of the environmental system. For instance, climatic change, sea level change, changes in oceanic temperatures, and ongoing sedimentation and erosion, will alter the specific relationship of the stability of the margin to topographic factors. Their importance also varies with the timescale of examination. On long timescales, for example, the rheology of the earth's crust will become significant, because glacio-isostatic sinking compels the grounding line to retreat, and vice versa, through the influence of relative sea levels.

Topography is an important 'filter' through which environmental influences are mediated to the ice sheet, determining the dynamic response of the ice sheet to those influences. For instance, time-dependent modelling of the Greenland Ice Sheet (Hulton, unpublished) suggests that different topographies can lead to contrasting responses to identical external forcing. One conclusion drawn from this is that:

> ...retreat behaviour and present day stability are strongly influenced by topography, particularly in calving locations (Hulton, unpublished).

Andrews (1973,1984,1987) has considered the dynamic implications of the change in ablation rate following the cessation of calving as a glacier retreats out of water. The response of the glacial system to a rapid calving retreat is high velocity flow. The retreat of the margin onto land

120

causes a rapid reduction in mass loss, but the response to this change is lagged as the rest of the system adjusts to the frontal transformation. This lag will therefore cause a stillstand or a readvance at this point. The location at which this occurs depends on relative shoreline altitudes and topography, and is a glacio-dynamic effect which has no causal relationship with climatic change.

In a similar vein, Hillaire-Marcel and others (1981) develop the theory of 're-equilibration moraines' from their studies in eastern Canada. Using the examples of the 500 km Sakami moraine in Quebec, and the St. Narcisse and Roulier moraines, they show how stabilisation of a retreating ice front can occur when the ice margin, previously floating, is suddenly grounded because of topographic blocking or an abrupt drop in water level following catastrophic drainage. Retreat then pauses as an equilibrium profile is restored. They conclude:

> Most of the moraines...in Quebec cannot be interpreted as being related to a...period of cooling in Late Glacial time. ... It is probable that most glacial stillstands reflect dynamic mechanisms associated with ice retreat. (p.213)

In a comprehensive review of the extent, evolution, and dynamic implications of proglacial lakes along the southern margins of the Laurentide Ice Sheet, Teller (1987) shows that, in the later stages of deglaciation, the majority of the margin must have been calving into large lakes. Large-scale, asynchronous oscillations of large lobate sectors of the ice were widespread (cf. Clayton and others, 1985), and Teller concludes with the assertion that, while climate was the overall driving force, an increasing weight of

121

chronostratigraphic evidence indicates that:

> ... most ice-marginal fluctuations during deglaciation were non-climatic in origin (Teller, 1987,p.61).

In modelling the growth and decay of the Antarctic Peninusula Ice Sheet, Payne and others (1989) found that although the forcing functions in the model were smooth, the output was stepped, representing a series of steady states separated by periods of rapid transition. This implies that there are 'critical thresholds, probably related to topography' (Payne and others, 1989,p.126) and they conclude that:

> Topography affects the dynamics of ice sheet behaviour by introducing thresholds of stability. (p.133)

Hughes (1987,1988) proposes a theoretical mechanistic explanation of these thresholds during ice sheet disintegration. His explanation is based on the concept of an evolving series of dominant ablation mechanisms. During deglaciation, the floating margins and termini of ice sheets and glaciers have two retreating margins, the calving front and the grounding line, both of which can have stable and unstable conditions largely conditioned by topography (Fig. 2.12). Hughes (1988) considers the retreat of an ice sheet from a glacial maximum position on the outer continental shelf, and envisages a series of stepwise retreat stages:

Step 1: The ice front calves back to a line of pinning points, such as islands or shoals, on the continental shelf.

Step 2: Compressive flow in the ice floating over the inner continental shelf reduces the calving rate, and therefore the
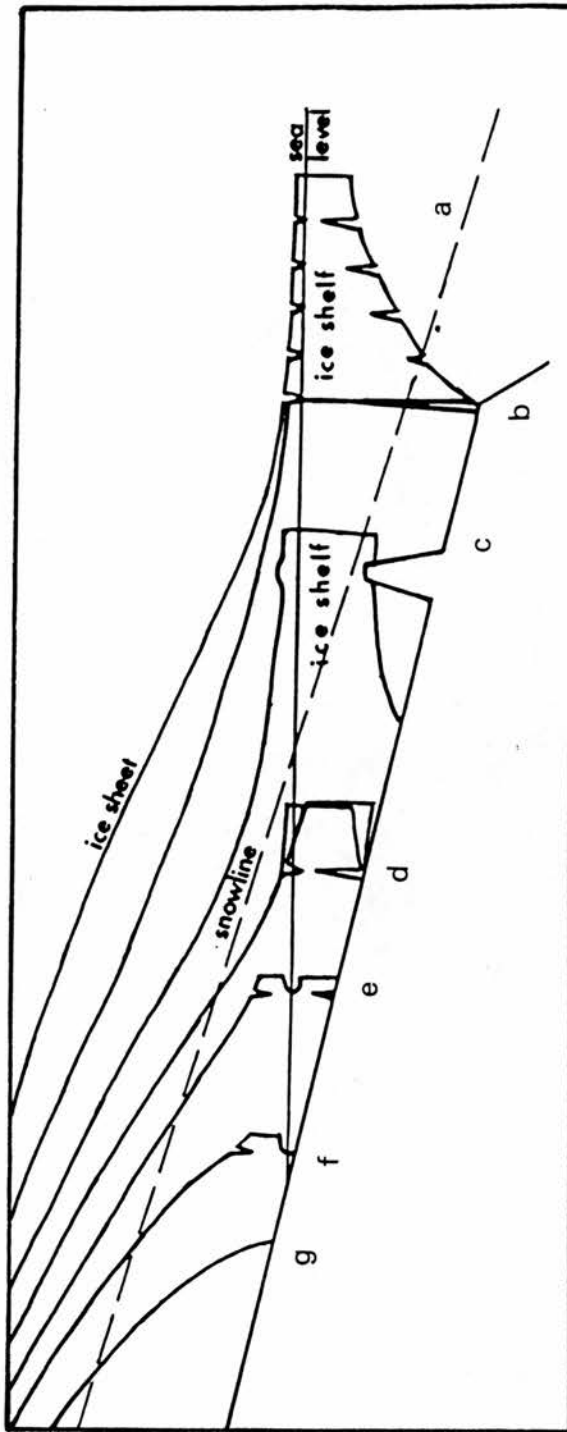
FIGURE 2.12. Stable locations for an ice sheet margin retreating out of the sea onto land. (After Hughes, 1987). Stable positions exist at (a) calving front of an unconfined ice shelf, (b) grounding line of an unconfined ice shelf, (c) calving front of a confined ice shelf, (d) grounding line of a confined ice shelf, (e) tidewater margin, (f) intertidal margin, and (g) terrestrial melting margin.

next semi-stable calving front is the tidal bending zone of   floating ice.

Step 3: Further retreat will result in grounding and the formation of a tidewater groove at sea level, with subaerial and submarine calving of slabs above and below the groove (cf. 2.1.ii; Fig. 2.2.c).

Step 4: The calving margin retreats until the groove stabilises temporarily in the intertidal zone on a beach, with only subaerial calving above the groove.

Step 5: Further retreat by slab calving and melting transforms the vertical ice wall into a ramp which ablates entirely by surface melting.

Distinctive calving mechanisms exist at all the temporary calving margins; the margin retreats, therefore, in steps, with each step characterized by the dominance of one calving mechanism, and the calving mechanism determined by the topographic location. The places that the margin halts, in this scenario, are therefore controlled by the bedrock geometry of the continental shelf and in-shore areas.

Theoretical considerations, empirical evidence, and modelling results all show that, during the collapse of ice sheets, the dominant external forcing brought to bear on ice sheets by atmospheric and hydrospheric changes, operates through interacting topographic and glacio-dynamic controls to produce a history of stepwise retreat.

2.3.iii. Palaeoclimatic Conclusions.

It is therefore clear that all calving ice masses have a very different relationship with climate from those with non-calving margins. At timescales of 10^4 and 10^5 years, insolation variations seem to determine the growth and decay of ice sheets in broadly the fashion proposed by Milankovitch (Hays and others, 1976). At all shorter timescales, and with progressively greater importance with greater temporal resolution, interactions of the cryosphere with the hydrosphere partially or totally decouple the behaviour of ice masses from climatic forcing.

The Quaternary geological record of glacial deposition, and terminal moraines in particular, have often been used to delimit major glacio-climatic events. This rests on the assumption that these deposits represent a glacial response to climate, whereas they may represent a change in the mass balance determined by changing ablation mechanisms and the glacio-dynamic response to such changes. If an ice margin has been in contact with large water bodies, there is therefore

> ... no simple transfer function between climate,
> tending to ice sheet mass balance, tending to response.
> (Andrews, 1973,p.197)

The implications are clear:

> If climatic change has little control over [calving]
> glacier fluctuations, then geologic records from such
> glaciers are of dubious value in palaeoclimatic
> reconstructions. (Mann, 1986, p.10).

The extent to which calving glacier behaviour reduces or obliterates the value of glacier fluctuations for palaeoclimatic reconstruction, and frustrates their use for correlation with other sources of palaeoclimatic information, depends very much on:

i)   the characteristics of the glacier systems in question, and,

ii) on the temporal resolutions of the records concerned.

The mechanisms outlined by Mercer (1961) are not applicable to all calving glaciers; not all glaciers terminating at fjord mouths are sensitive to ELA shifts. Furthermore, if the ELA intersects a glacier in a steep section, the mechanism shown in Fig. 2.10 does not operate. In Mann's (1986) case study of the Lituya glacier system, for instance, he found that the record of response to Holocene climatic change has been free from interference from internal, glacio-dynamic forcing, partly because the resolution of the geological record is coarse. Nevertheless, the historical fluctuation of the glacier has been valueless as a climatic indicator because the glacier is in the advancing stage of a 'tidewater glacier cycle'. Therefore a knowledge of the length of the insensitive period is very important for the interpretation of historic records, but may not be significant for older records.

Mann (1986) shows that the fjord glaciers most susceptible to anomalous responses to climate are those:

i)   Occupying complex fjord systems.

ii) In areas of low ELA.

iii) Terminating at a fjord mouth adjoining the open sea.

Other studies (eg. Funder, 1972,1985; Warren and Hulton, 1989,1990) bear these conclusions out.

It is not true to say that calving glaciers are without value for palaeoclimatic purposes, as suggested by England (pers.comm., 1989), but there are many potential sources of

126

'non-climatic noise' which must be considered in any interpretation of the geological records of such glaciers. Meier and Post (1987) close their review of 'Fast Tidewater Glaciers' by stating that:

> Assigning climatic significance to grounded tidewater glacier fluctuations is dubious at best. (p.9058)

In conclusion, therefore:

i). Valid palaeoclimatic inferences drawn from the geological record of calving ice masses are the exception rather than the rule.

ii). Temporal coincidence of retreat stages of calving margins are, in most cases, purely coincidental.

iii). Non-linear feedbacks inherent in marine ice sheets may greatly amplify the glacial response to small changes in any of the boundary conditions.

iv). Topographic configurations play a much greater role in controlling the oscillations of calving margins than terrestrial margins.

v). Correlations between calving and non-calving margins are unlikely to be valid unless the two are linked by sea level change.

iv). The only glacial deposits from which reliable and detailed inferences concerning climate can be drawn are those formed at terrestrial ice margins.

C H A P T E R   T H R E E .

RECENT GLACIER FLUCTUATIONS IN WEST GREENLAND.

3.1. CLIMATE AND GLACIER RESPONSE IN GREENLAND.

This chapter presents a study of the contrasting 20th Century behaviour of tidewater, lake-calving, and land-terminating glaciers in West Greenland, and an assessment of the relative importance of climatic and non-climatic controls on their fluctuations. West Greenland is an ideal area to address such questions because it has a large number of ice sheet outlets terminating in a range of environments, and because many decades of historic and climatic records exist. Iceberg calving is and has been an extremely important ablation mechanism in Greenland, and calving outlets are of great significance within the overall dynamics of the Greenland Ice Sheet (Reeh, 1985,1989).

The relationships between climate, mass balance, glacio-dynamic response, and frontal change of Greenland's outlet glaciers are still incompletely understood (Knudsen, 1986). As in many parts of the world, the Little Ice Age advance peaked in Greenland during the late nineteenth century (Weidick, 1968). A subsequent strong warming trend peaked in the 1940s, since when slight cooling has been recorded (Fig. 3.1). This trend has affected West Greenland between 59° N and 73° N (Frydendahl, 1989). Ablation conditions are fairly uniform between 60° N and 73° N, but summers are warmer and drier in the continental central area, than in the maritime north and south (Weidick, 1984).

128
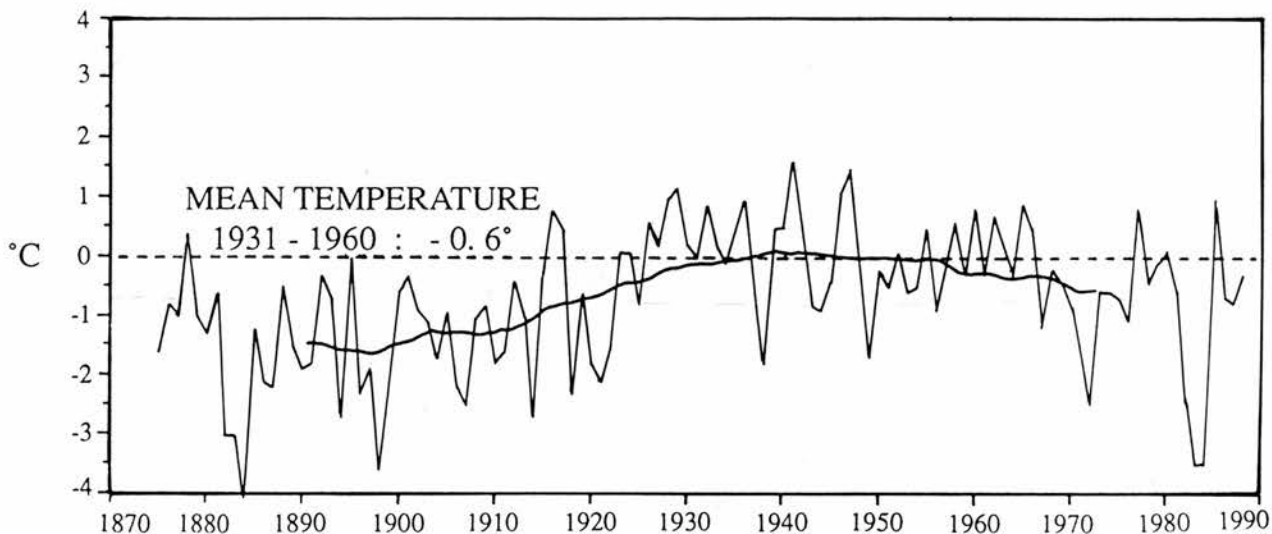
FIGURE 3.1. Temperature fluctuations at Nuuk (Godthaab), 1870
    - 1988. The trend line is a 30-year moving mean. Data
    from the Danish Meteorological Institute, 'Danmarks
    klima 1988'.

Warming caused general retreat, greatest between 1920 and
1940 and slower thereafter. The historic records of the last
two centuries show that oscillations of West Greenland outlet
glaciers are broadly controlled by fluctuations in summer
temperatures, rather than by precipitation (Weidick,
1959,1968 p.45), and recent glaciological studies have
confirmed that ablation rates are related in a nearly linear
way to temperature due to the strong dependence of the
sensible heat flux on temperature (Braithwaite and Olesen,
1984,1990).

      However, the behaviour of the calving glaciers does not
seem to relate to climatic changes in a straightforward way.
As early as 1959, their anomalous behaviour led Weidick to
state that the use of such glaciers 'as indicators of

129

climatic changes must be regarded with some scepticism' (Weidick, 1959, p.184). Weidick (1968, pp.36 - 37) used 'Mercer's principle' (Mercer, 1961) to explain some of these anomalous patterns, and noted that some of the longer stillstands of the glacier fronts reflected topographic rather than climatic control. Correspondence of stillstands with fjord shallowings, narrowings and bifurcations has been noted in East Greenland (Funder, 1972,1989), West Greenland (Funder, 1985) and North Greenland (Higgins, 1988).

3.2. METHODOLOGY.

3.2.i. Approach.

Many factors affect the frontal change of West Greenland's outlet glaciers; Weidick (pers.comm., 1988) has rightly stressed that to address the problem fully would require very costly, long-term monitoring projects on many outlet glaciers to gather detailed data concerning mass balance, hydraulic characteristics, bed topography, fjord bathymetry, ice fluxes, iceberg fluxes, and terminus fluctuations. With the exception of the data on fluxes and terminus fluctuations gathered by Bauer (1968) and Kollmeyer (1980) in central West Greenland, such information does not exist. It is not even known, in many cases, whether the calving glaciers are floating or grounded.

For practical reasons, therefore, an exclusively field-based approach was rejected. Remote sensing, though it limits the range of variables which can be measured, is the only approach giving adequate spatial and temporal coverage capable of providing meaningful answers to these issues. Satellite remote sensing will be the tool for future monitoring. Changes in glacier morphology and frontal positions have been monitored in many parts of the world using LANDSAT MSS & TM, NOAA AVHRR and SEASAT SAR imagery (eg. Williams, 1987; Fujii and Yamanouchi, 1987; Hall and Ormsby, 1983), and it has also proved possible to extract mass balance trends from satellite images (Ventura, 1987). In Greenland, SEASAT altimeter data have been used to measure changes in the surface altitude of the ice sheet (Zwally and others, 1983; Bindschadler and others, 1989;

131

Zwally, 1990), for detailed mapping in the ablation zone (Thomsen, 1983,1986), for monitoring iceberg discharge (Birnie and Williams, 1985; MISR/RAE, 1983) and for measuring ice sheet retreat since the Little Ice Age using trimlines (Knight and others, 1987).

However, despite the great potential which satellite remote sensing has for glaciological studies, at present, both the data span and spatial resolution are poorer than for aerial photographs. Since the research required detailed information about ice marginal and surface morphological changes over as long a period as possible, the former was rejected in favour of the latter. Consequently, the data was gathered from 170 stereo pairs of vertical aerial photographs, used in conjunction with the best available topographic maps (1:250,000). This was supplemented by two summer field seasons, and historical records collated by Weidick (1959,1968).

3.2.ii. Sampling, and Choice of Variables.

Seventy-two outlet glaciers in West Greenland were studied. Of these, 23 are tidewater glaciers, 25 are land-terminating glaciers, and 24 calve into lakes (Table 3.1). Glaciers 9 - 12, 15 - 17, 43 and 44 have been visited in the field. The choice of glaciers, the temporal coverage, and the choice of measured variables were largely constrained by data availability. A few Greenland glaciers have been photographed numerous times, but most just two or three times, and some only once. Data restrictons of this kind ruled out the original intention of including East Greenland glaciers in the research. The study focuses on the time period between the 1940s/1950s, when the majority of West Greenland was photographed for the first time, and the 1980s when another full coverage was achieved. This enables direct comparisons to be made between the glaciers, and assesses glacier response during a time of changing climate. The latitudinal range of the study (61^{0}N - 72^{0}N) covers almost the whole area of Greenland in which both calving and non-calving outlet glaciers co-exist; elsewhere, north and east, almost all outlets terminate in the ocean.

Fig. 3.2 illustrates schematically the parameters recorded for each glacier. Although the detailed understanding of the relationship of a particular glacier to climatic forcing requires much specific data, on a regional scale the important controls on glacier fluctuations are broadly climatic, glaciological, and topographic. The aim, therefore, was to correlate glacier fluctuations with factors in these categories known to control them.

TABLE 3.1. The Study Glaciers, Data, and Sources.

Glacier Number and Name	Latitude (°N)	Terminal Environment	Frontal Change (Central) (m)	Width (km)	Height a.s.l. (m)	Gradient (terminal 4 km)	Photograph Coverage
1. Sermeq Kangigdleq.	60°56'	Land.	− 275	1.1	c. 200	1:13	1953, 1981, 1985.
2. Jespersen Brae (south).	60°57'	Land.	− 145	4.8	c. 90	1:25	1953, 1981, 1985.
3. Søndre Qipisarqo Brae.	61°01'	Land.	− 1280	1.4	c. 30	1:13	1943, 1985.
4. Nordre Qipisarqo Brae.	61°01'	Land.	− 420	4.5	c. 30	1:12	1943, 1985.
5. Jespersen Brae (N.E. lobe).	61°03'	Lake.	− 3900	3.8	600	1:20	1953, 1981, 1985.
6. Jespersen Brae (north).	61°04'	Land.	− 30	2.7	c. 180	1:18	1953, 1981, 1985.
7. Unnamed.	61°05'	Land.	− 364	2.7	c. 180	1:20	1953, 1981.
8. Unnamed.	61°05'	Lake.	+ 800	0.8	600	1:9	1953, 1981.
9. 'Motfeldt Sø Sermia'.	61°10'	Lake.	+ 400	1.3	130	1:33	1953, 1981.
10. Qôrqup Sermia.	61°11'	Tidewater.	+ 500	1.8	−	1:12	1942, 1953, 1981.
11. 'Qôrqup Lake lobe.	61°12'	Lake.	+ 950	1.9	c. 130	1:24	1953, 1981.
12. Kiagtût Sermia.	61°13'	Lake.	+ 550	0.8	c. 70	1:25	1942, 1953, 1981.
13. Eqalorutsit Kitdlit Sermiat (West branch).							
14. Eqalorutsit Kitdlit Sermiat (East branch).	61°14'	Tidewater.	+ 2140	2.0	−	1:12	1943, 1981, 1985.
15. Eqalorutsit Kangigdlit Sermiat.	61°16'	Tidewater.	+ 1190	2.0	−	1:7	1943, 1981, 1985.
16. Sydgletscher.	61°20'	Tidewater.	+ 750	3.3	−	1:18	1951, 1981.
17. Nordbogletscher.	61°23'	Lake.	0	2.3	600	1:35	1953, 1981.
18. Nordgletscher.	61°24'	Land.	+ 640	1.4	660	1:21	1951, 1981.
19. Østgletscher.	61°26'	Land.	+ 400	0.6	650	1:15	1951, 1981.
20. 'Kangersôp Sermia'.	61°26'	Land.	0	0.7	c. 650	1:9	1951, 1981.
21. Frederikshabs Isblink (lobe in Kangersôp taserssua).	62°29'	Lake.	0	1.0	c. 210	1:5	1943, 1985.
22. 'Taserssuaq Sermia'.	62°31'	Lake.	− 1680	2.6	c. 30	1:13	1943, 1985.
23. 'Sorraitsup Sermia'.	62°40'	Lake.	− 630	1.8	c. 125	1:25	1943, 1985.
24. Sermeq.	62°44'	Lake.	+ 450	1.5	c. 20	1:250	1943, 1985.
25. 'Isortuarssôp Sermia'.	63°32'	Lake.	− 1600	3.5	c. 450	1:25	1949 (map), 1985.
26. Kangaussarssôp Sermia.	63°50'	Lake.	− 1560	2.8	175	1:22	1936 (map), 1985.
27. Kangiata Nûnata Sermia.	64°06'	Tidewater.	− 1560	0.6	−	1:10	1936 (map), 1985.
28. Akugdlerssôp Sermia.	64°19'	Tidewater.	0	4.1	−	1:13	1936 (map), 1985.
29. Qamanarssôp Sermia.	64°21'	Land.	0	3.5	c. 40	1:20	1943, 1985.
30. Narssap Sermia.	64°28'	Tidewater.	+ 600	1.1	−	1:16	1943, 1985.
31. Kangilinguata Sermia.	64°39'	Land.	+ 80	4.3	c. 5	1:17	1952, 1985.
32. 'Taserssiap Sermia'.	64°50'	Lake.	+ 240	4.4	c. 675	1:15	1952, 1985.
33. Unnamed.	66°17'	Lake.	+ 380	2.4	810	1:18	1952, 1985.
34. Unnamed.	66°20'	Lake.	+ 100	1.8	c. 800	1:17	1952, 1985.
35. Unnamed.	66°20'	Lake.	+ 240	4.2	826	1:35	1952, 1985.
36. Unnamed.	66°21'	Lake.	+ 1125	2.0	450	1:9	1944, 1985.
37. Unnamed.	66°23'	Lake.	+ 200	4.7	c. 80	1:11	1944, 1985.
38. Unnamed.	66°39'	Land.	+ 255	6.4	c. 80	1:12	1953, 1985.
39. Nordenskiolds Gletscher.	66°57'	Tidewater.	+ 240	5.4	−	1:18	1953, 1985.
40. Unnamed.	66°59'	Lake.	+ 255	2.6	c. 210	1:18	1953, 1985.
41. Unnamed.	68°20'	Lake.	+ 125	3.8	77	1:20	1953, 1985.
42. Unnamed.	68°20'	Land.	− 1320	2.1	190	1:10	1953, 1985.
43. Sarqardliup Sermia.	68°32'	Tidewater.	+ 880	4.8	−	1:17	1953, 1985.
44. Alangordliup Sermia.	68°42'	Tidewater.	+ 800	2.5	−	1:66	1953, 1985.
45. Jakobshavns Isbrae.	68°54'	Tidewater.	+ 320	6.3	−	1:20	1953, 1985.
46. Unnamed.	68°58'	Land.	+ 360	0.4	c. 100	1:18	1953, 1985.
47. Unnamed.	69°12'	Land.	+ 400	1.2	150	1:16	1953, 1985.
48. Sermeq Avangnardleq.	69°19'	Tidewater.	+ 640	2.4	−	1:16	1953, 1985.
49. Unnamed.	69°20'	Land.	+ 160	1.6	c. 90	1:13	1953, 1985.
50. Unnamed.	69°21'	Land.	+ 1575	0.5	c. 100	1:12	1953, 1985.
51. Unnamed.	69°25'	Land.	+ 1080	1.0	c. 275	1:13	1953, 1985.
52. Eqip Sermia.	69°34'	Tidewater.	0	4.8	c. 280	1:37	1953, 1985.
53. Unnamed.	69°38'	Lake.	+ 385	0.9	−	1:22	1953, 1985.
54. Kangilerngata Sermia.	69°47'	Tidewater.	− 45	4.1	c. 330	1:16	1953, 1985.
55. Sermeq Kujatdleq.	69°50'	Tidewater.	+ 750	5.0	−	1:24	1953, 1985.
56. Unnamed.	69°53'	Lake.	+ 1000	1.1	−	1:17	1953, 1985.
57. Store Gletscher.	70°00'	Tidewater.	− 350	4.9	c. 470	1:7	1953, 1985.
58. Sermilik.	70°12'	Tidewater.	+ 990	3.0	−	1:14	1953, 1985.
59. Unnamed.	70°24'	Land.	+ 450	0.1	−	1:23	1953, 1985.
60. Sermeq Silardleq.	70°38'	Tidewater.	+ 225	3.2	c. 650	1:6	1953, 1985.
61. Perdlerfiup.	70°41'	Tidewater.	+ 180	2.6	c. 100	1:17	1953, 1959, 1985.
62. Unnamed.	70°48'	Land.	+ 1700	2.1	−	1:4	1953, 1959, 1985.
63. Unnamed.	71°00'	Land.	+ 490	0.2	c. 730	1:17	1953, 1959, 1985.
64. Kangerluarssôp Sermia.	71°06'	Tidewater.	+ 1850	3.2	740	1:97	1953, 1959, 1985.
65. Unnamed.	71°09'	Land.	+ 1750	5.0	−	1:27	1953, 1985.
66. Unnamed.	71°14'	Lake.	+ 730	1.4	c. 1	1:16	1953, 1985.
67. Kangerdlugssuaq Sermia.	71°22'	Tidewater.	− 240	6.0	−	1:24	1953, 1959, 1985.
68. Johannes Brae.	71°23'	Land.		1.2	−	1:23	1953, 1985.
69. Umiamako Isbrae.	71°27'	Tidewater.		3.0	200	1:17	1953, 1959, 1985.
70. Rinks Isbrae.	71°34'	Tidewater.		4.2	190	1:12	1953, 1959, 1985.
71. 'Ilulialigssuaq Sermia'.	72°28'	Lake.		2.8			1953, 1985.
72. 'Qejartoriaq Sermia'.	72°41'	Lake.		1.2			1953, 1985.

Names in inverted commas are unofficial.

Pre-1981 scenes @ 1:40,000.
1980's scenes @ 1:150,000.

134
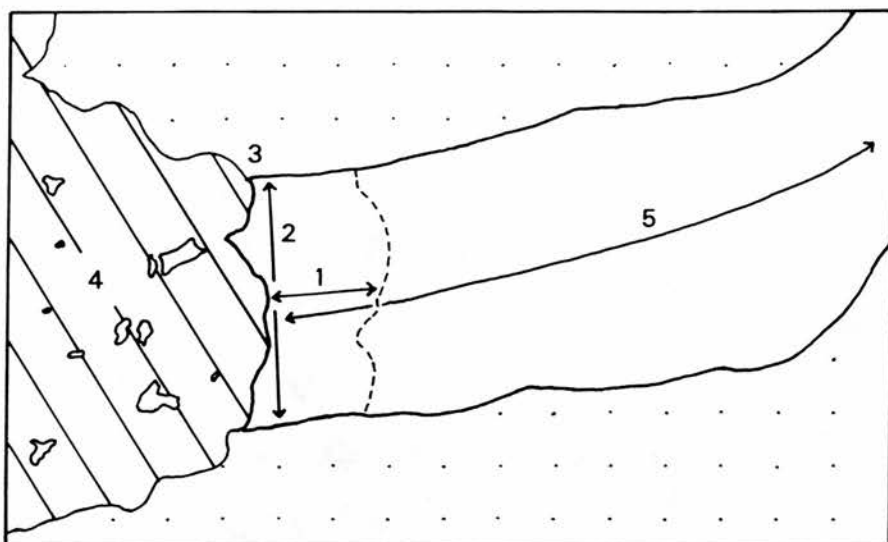
FIGURE 3.2. Schematic diagram of a 'typical' glacier showing
the parameters recorded: 1. Change in terminal
location: measured to the nearest 0.5mm (= 20m @
1:40,000 and 75m @ 1:150,000). Several measurements
were made in cases of non-uniform frontal change.
2. Terminus width, or approximate length of ice front.
3. Topographic characteristics: whether the glacier
fronts coincide with fjord mouths, narrowings, bends or
bifurcations. 4. Terminal environment. 5. Gradient of
the last 4 km.

Twentieth Century summer temperature trends, the dominant

climatic   variable   (Weidick, 1968),   are   broadly   known

(Fig. 3.1). In the absence of ice depth data, glacier width

was used as a proxy for glacier volume, which could control

response times.   Glacier gradient also affects response times

in Greenland (Weidick, 1984) and these were simply obtained

from 1:250,000 maps.

Topographic control of glacier oscillations operates

primarily   by   determining   the   nature   of   the   terminal

environment (and thereby the presence or absence of calving), and secondarily through the cross-sectional area of the glacier terminus (Powell, 1984), water depth being the most important single variable (Brown and others, 1982). In the absence of bathymetric data, the morphology of the glacial troughs was used as an indication of varying depth, particularly noting fjord bifurcations, bends, and junctions, and sharp variations in width. These have been highlighted by Mercer (1961), Mann (1986) and Funder (1989) as topographic determinants of stillstand locations for calving glaciers.

In addition to these parameters, qualitative glaciological data concerning relative rates of iceberg calving, morainic cover, degree of crevassing, and changes in the character of the terminus were also noted. In some cases, for instance, no change in terminus location is measurable at the resolution of the photographs, but the morphology and character of the snout had changed considerably, becoming steeper and cleaner, indicating advance.

3.2.iii. Data Reliability.

The photographic coverage is not ideal or complete; the photographs are of different scales and date from various years. The simple figures derived from the sparse coverage probably mask more complex patterns. This is especially true for the actively calving termini for which frequent repeat coverage would be necessary for a full record of frontal oscillation (Meier and others, 1985a). Glaciers such as Rinks Isbrae and Jakobshavns Isbrae have extremely high discharges (for the latter, an ice volume of 30.8 - 48.7 $km^3 yr^{-1}$ -Bindschadler, 1984) and their annual oscillations are often great enough to mask the long term trend. In July 1985, for instance, the calving front of Jakobshavns Isbrae retreated 2 km in 45 minutes (Hughes, 1986). Annual observation has shown, however, that Jakobshavns Isbrae is in quasi-equilibrium (Pelto and others, 1989). The figures given here for the actively calving lobes, based on aerial photographs in widely separated years, must therefore be treated with caution. It is also unfortunate that the grounded and floating calving glaciers have to be treated in an undifferentiated way due to lack of information. Controls on the dynamics of floating and grounded outlet glaciers are different, though poorly understood (Pelto and Warren, in press), and so separate analysis would be preferable. Given the scale of the problem and the nature of the data, however, there is no practical way of avoiding these methodological drawbacks.

Neither glacier width nor changes in fjord configuration are ideal proxy measures respectively of

137

glacier volume and fjord depth, but they are the only parameters accessible with this approach, and, again, there is no practical alternative. It is probable that, even if accurate ice volumes and the detailed bathymetry of the fjords were available, the general patterns emerging from the data would be the same.

3.3. RESULTS.

Table 3.1 shows the full data set. Fig. 3.3 shows the location of the glaciers, the spatial pattern of their advance and retreat behaviour, and their terminal environments. Table 3.2 presents a summary of the advance and retreat patterns, and Fig. 3.4 displays the data graphically. The data span for each glacier is that between the first and last date shown for it on Table 3.1; the figures have been averaged to produce mean net rates of frontal change for each glacier in order to facilitate comparison over the different time periods.

		SIGN OF TERMINUS MOVEMENT.		
		–	0	+
TERMINAL ENVIRONMENT:				
LAND. (25)	No.	17	4	4
	%	68%	16%	16%
TIDEWATER. (23)	No.	6	3	14
	%	26%	13%	61%
LAKE. (24)	No.	7	5	12
	%	29%	21%	50%

TABLE 3.2. Net terminus change of 72 West Greenland outlet glaciers between the 1940s/1950s and the 1980s.
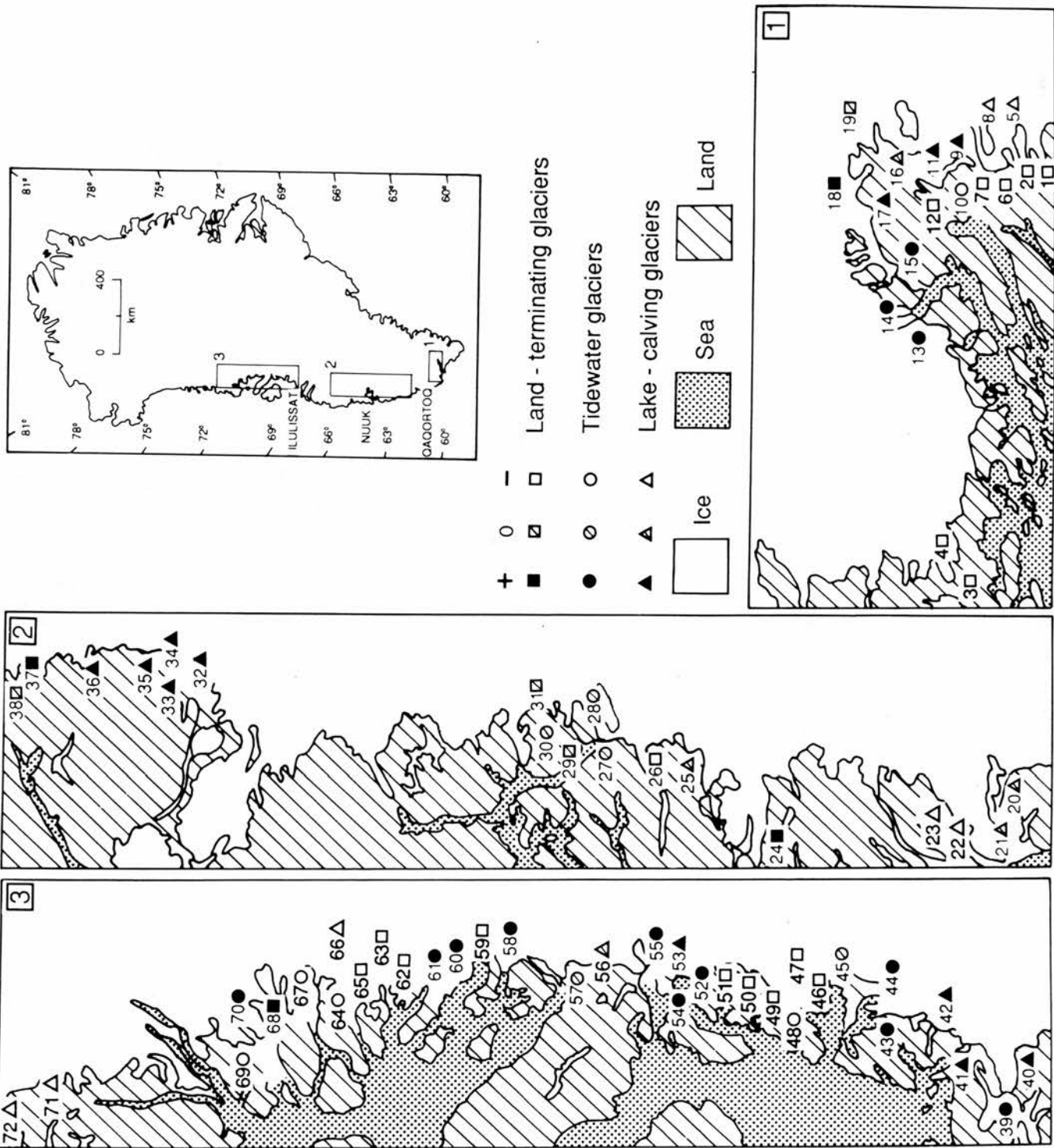
FIGURE 3.3. The location of the 72 study glaciers in West Greenland, their trends of frontal change during the last 40 years, and their terminal environments.
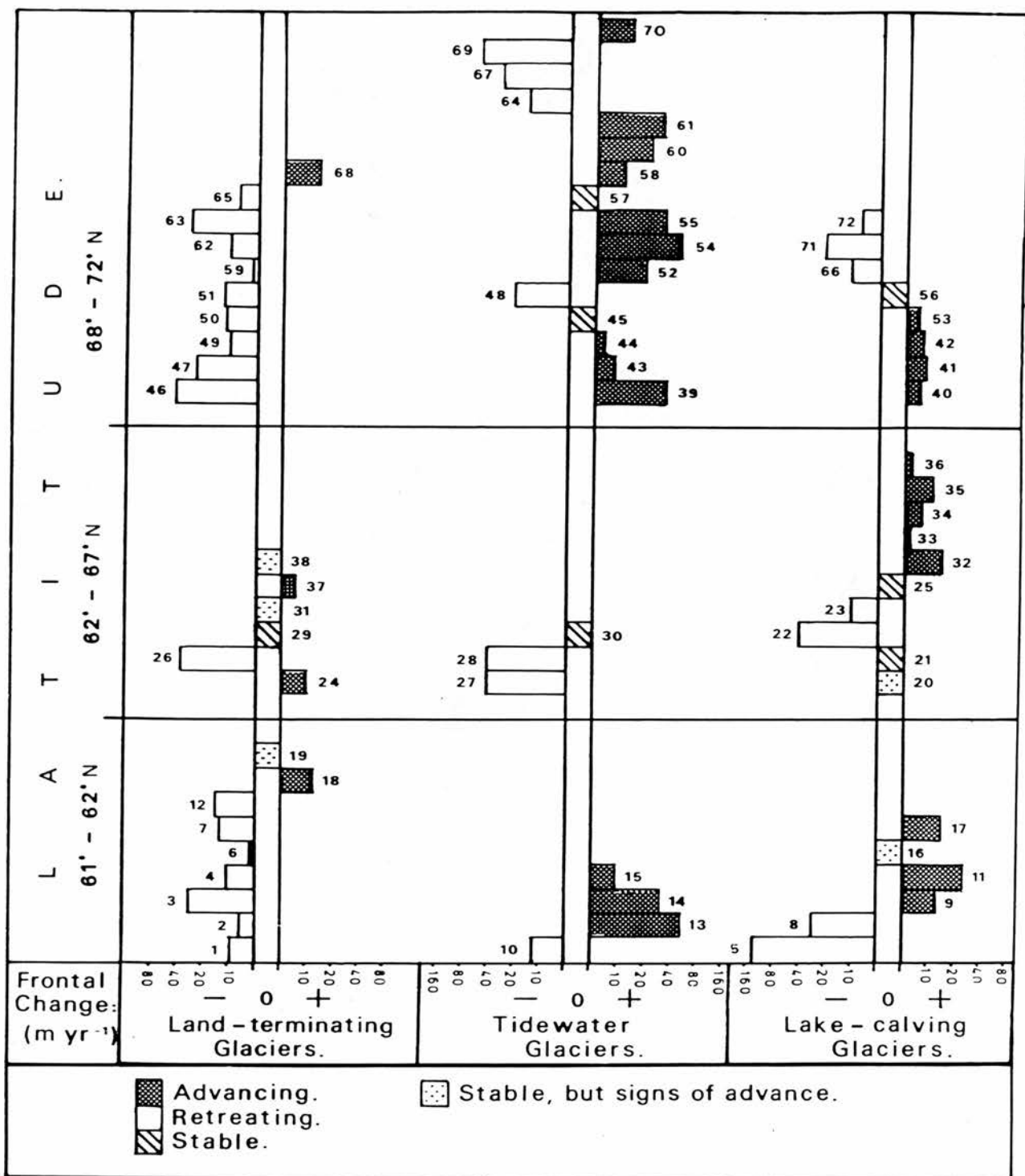
FIGURE 3.4.   Mean rates of terminus fluctuation for 72 West
Greenland outlet glaciers, mid-century - 1985.

3.3.i. Correlations.

3.3.i. a). Frontal change v. terminal environment.

Table 3.2 shows that, over the last 40 years, the calving and land-terminating glaciers have exhibited opposite trends of frontal behaviour; 84% of the land-terminating glaciers have been retreating or stable, while 50 - 61% of the calving glaciers have been advancing. This is particularly clear in the north where all but one of the land-terminating glaciers have been strongly retreating while there are more than twice as many calving glaciers advancing as retreating. The amplitude of oscillation is greater amongst calving than non-calving glaciers, (frequently thousands of metres as against tens or hundreds of metres), and is greatest amongst the tidewater glaciers.

Five of the advancing lake-calving glaciers (Glaciers 9, 11, 33, 34 and 36) were calving more profusely in the 1980s than in the 1940s/1950s. The nature of the calving often differs between tidewater and lake lobes. Active calving into the heads of fjords tends to result in a chaotic profusion of icebergs and brash ice which may extend up to several kilometres down the fjord. Calving into the lakes, however, normally produces a few large icebergs and very little brash ice. For example, one of the icebergs in Motzfeldt Sø in front of Glacier 11 in 1981 had a surface area of c. 135,000 m^2.

3.3.i. b). Frontal change v. glacier characteristics.

There are no strong correlations between recent frontal change of the glaciers and morainic cover, glacier width, or surface gradient. Abundant morainic cover

142

is relatively rare on the West Greenland glaciers, and only 15 of the study glaciers have a substantial mantling of supraglacial debris. More of these 'dirty' glaciers are retreating than advancing, but 3 are stable and 4 are advancing, so no dominant trend is apparent. The same is true for glacier width (Fig. 3.5) and for surface gradient (Fig. 3.6); neither appear to have had any significant influence on terminal behaviour.
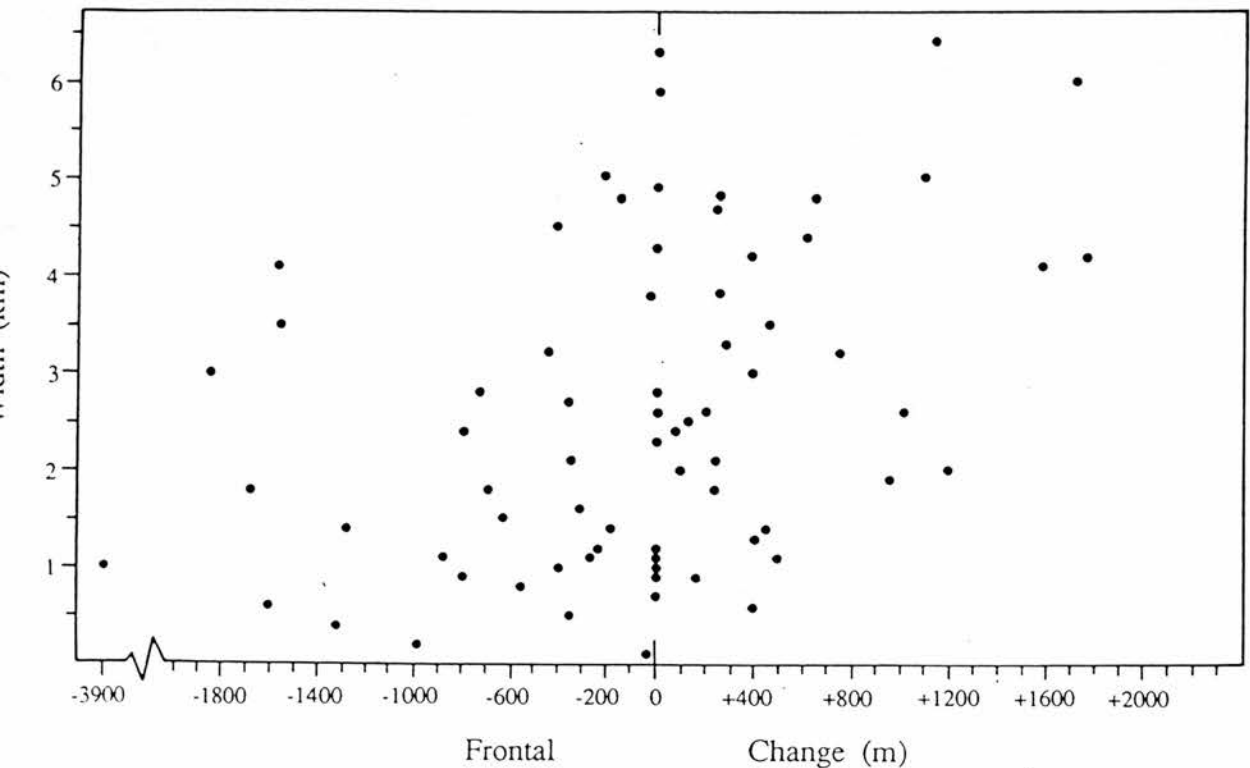
FIGURE 3.5. Relationship between frontal change and glacier width.
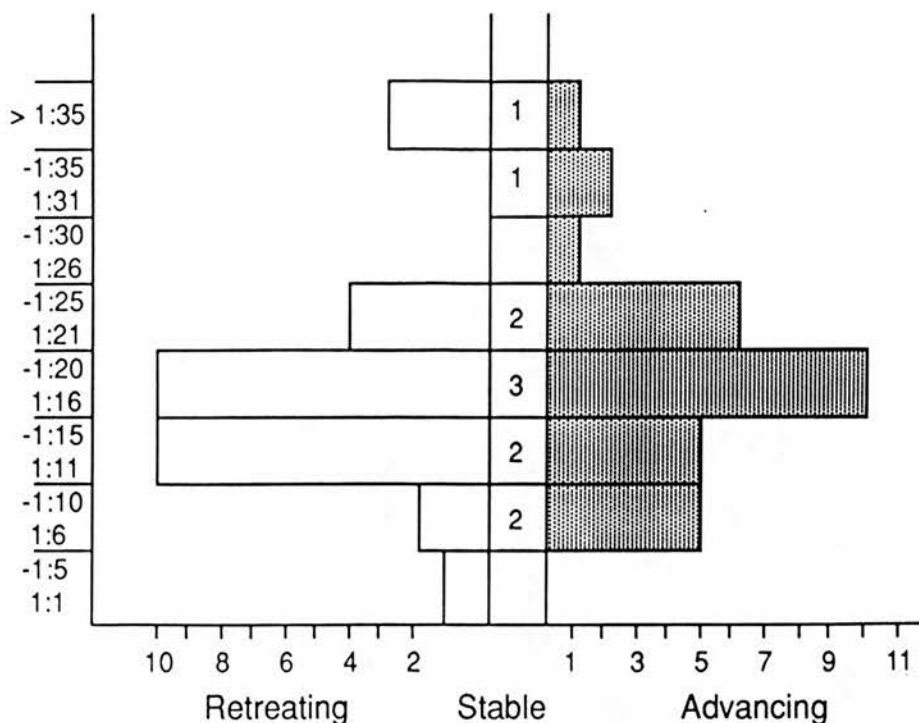
FIGURE 3.6. Relationship between frontal change and surface gradient of the last 4 km.

3.3.i. c) Frontal change v. latitude.

No clear relationship between terminus behaviour and latitude exists (Fig. 3.4), although there are some minor latitudinal contrasts. The glaciers in the south and north show a much greater degree of variability than those in the central zone. The highest rates of both retreat and advance were recorded south of 63°N and north of 68°N. While retreat predominates amongst the non-calving glaciers in the south and north, in the central belt there seem to be signs of a general advance. Glaciers 24 and 37 have advanced, and Glaciers 26, 31 and 38 have cleaner and steeper fronts now than in the middle of the century. Five of the lake lobes in this area are advancing, and advance appears to be imminent at the snout of Glacier 20. Only three glacier systems (the land-terminating Glacier 26, the 'hammerhead' lake lobe of Glaciers 22 and 23, and the tidewater termini of

144

Glaciers 27 and 28 (Fig. 3.7.C) which formed one terminus in 1936) have retreated.

3.3.i. d). Frontal change v. topography.

The behaviour of the land-terminating glaciers shows no clear relationship to topography, neither to valley width nor breaks of slope, with the possible exception of Kangilinguata Sermia (No. 31) which is stable at a slight narrowing of its valley. Frontal change of the calving glaciers, however, seems to be closely related to topography (Fig. 3.7). The calving fronts of five lake-calving glaciers (16, 20 [Fig. 3.7.J], 21, 25 and 56) and three tidewater glaciers (30, 45 and 57) have remained essentially unchanged. (Narssap Sermia [Fig. 3.7.D] has been at its current location since at least 1855 - Weidick, 1959). All of these glaciers terminate at fjord narrows, junctions, sharp bends or bifurcations. In addition, Glaciers 27 and 28 (Fig. 3.7.C) have stabilised at such points after a retreat which began ca. 1800 (Weidick 1959), and Eqalorutsit Kangigdlit Sermiat (Glacier 15, Fig. 3.7.B) is at a marked widening.

In the case of the floating terminus of Jakobshavns Isbrae (No.45), post-Little Ice Age retreat of the grounding line halted at a pronounced bedrock step (Hughes, 1986). Locations such as these can therefore be regarded as topographic 'pinning points', preventing or moderating the glacial behaviour which would occur in fjords of equal width and depth.

There are several examples of active calving glaciers which have been consistently advancing in their central sections but have moved little or remained 'pinned'
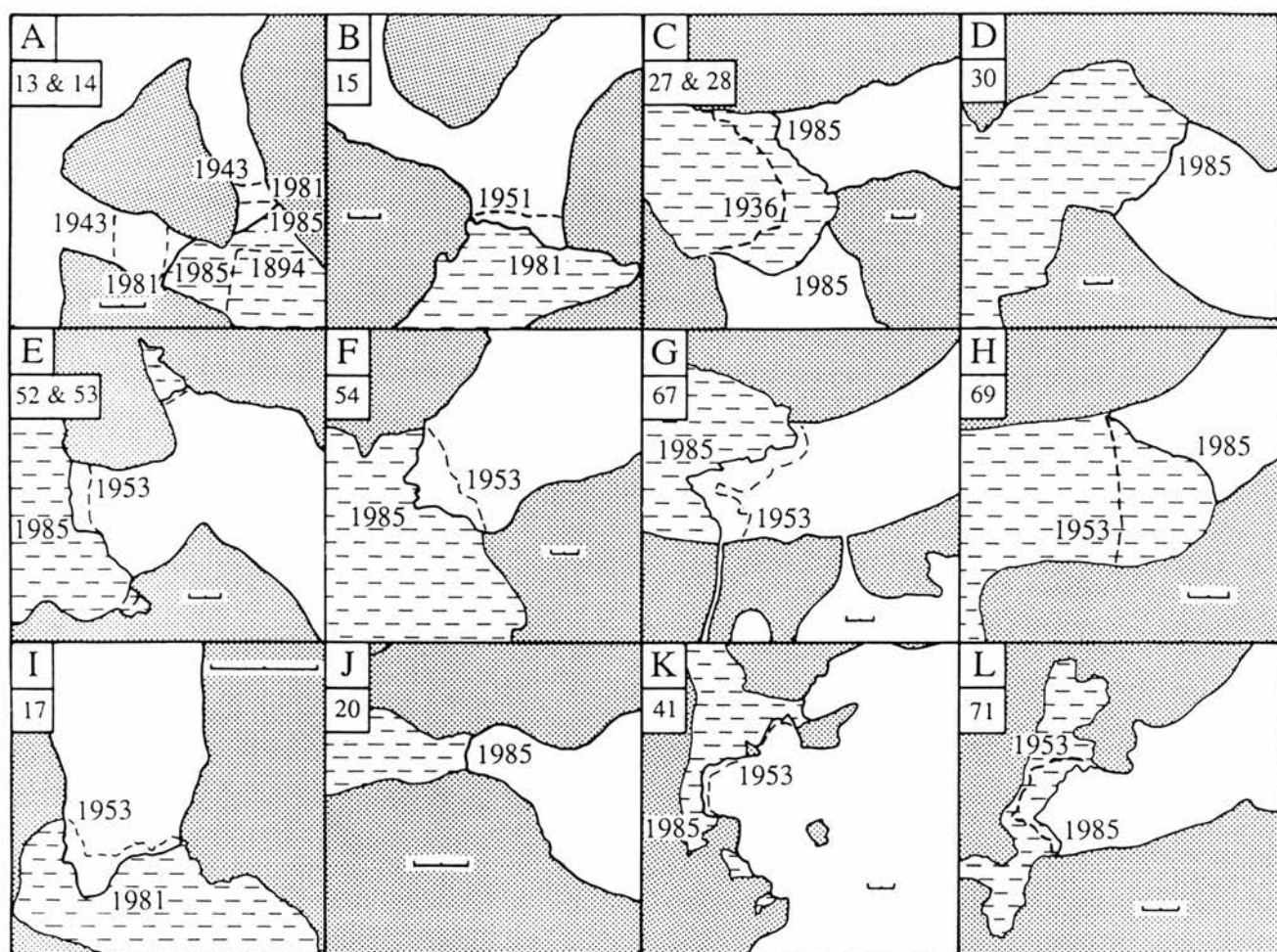
FIGURE 3.7. Examples of topographic control of calving
        fronts. A - H are tidewater glaciers and I - L are
        lake-calving glaciers. All sketches are orientated
        with north upwards, and the scale bars represent 1 km.
        Where only one ice front position is shown, the glacier
        has been stable throughout the period.

at stable lateral locations.   Kangerdlugssuaq   Sermia

(Fig. 3.7.G),   Kangilerngata   Sermia   (FIG.3.8F)   and

Nordbogletscher (Fig. 3.7.I), are all advancing in this

manner.   Other tidewater fronts, such as Qôrqup Sermia

(No. 10) and both branches of Eqalorutsit Kitdlit Sermiat

(Fig. 3.7.A), have changed in a pivotal fashion, with one

146

side of the front remaining fixed while the other moves a substantial distance. Umiamako Isbrae (Fig. 3.7.H) is a striking example of this; its northern side has shown no measurable change since 1953 while its southern side has retreated 3.0 km. (This kind of behaviour has been observed elsewhere in Greenland as well. Between the early 1950s and 1975, Petermann Gletscher in North Greenland advanced 10 km in its central section while its south-west side remained at the same point - Dunbar, 1978).

3.3.ii. Calving retreats.

Some of the retreats appear to have been amplified by glacio-dynamic factors. The highest rate of net retreat ($- 139.3$ m yr^{-1}) is that recorded for the lake-calving, unnamed, Glacier 5. At the other end of this lake, Glacier 8 retreated 800 m to a lake-shoreline position between 1953 and 1981. In the 1953 scene, the two termini are just 480 m apart, and the lake is completely choked with icebergs and brash ice. By 1981 they are separated by a lake 5.25 km long. It seems reasonable to presume that the two glaciers had flowed into each other until shortly before 1953. Similarly, Glacier 71 (Fig. 3.7.L) had just retreated from the far shore of the lake Ilulialigssuaq in 1953, and by 1985 it had retreated a further 700 m, accompanied by active calving. Detachment from the rock wall of the lake would have greatly increased the length of the calving front. In 1953 the calving front was 4.5 km long; in contrast, the glacier was only 2.8 km wide where it flowed into the lake. Although the glacier front retreated over 700 m during the period, the marked reduction in surface meltwater activity (streams and moulins), and the clean marginal ice, indicates that a readvance may begin as soon as the calving flux is reduced.

3.4. A CASE STUDY IN SOUTH GREENLAND.

A detailed case study of the outlet glaciers and local glaciers of the Narsarsuaq region in South West Greenland serves to illustrate and amplify some of these points.

3.4.i. South Greenland: The study area, and data acquisition.

In South Greenland, the Greenland Ice Sheet discharges through high (> 2000 m) alpine mountains in numerous outlet glaciers which terminate variously in the sea, lakes, or on land. It is therefore an ideal area for testing the hypothesis that an important variable controlling the response of outlet glaciers to climatic change is the nature of the terminal environment, and in particular, the presence or absence of iceberg calving. In a survey of the oscillations of 72 West Greenland glaciers over the last 40 years, Warren (1990, in press) highlighted the Narsarsuaq area as one of great interest because the glacier changes mirror the patterns found throughout West Greenland.

A unique glacier configuration exists in the Narsarsuaq area: two large outlet glaciers reach tidewater at active calving fronts, and each has large distributary glaciers terminating on land and in lakes (Fig. 3.8). Regional topography is strongly affected by bedrock geology; in the east of the study area the Nepheline Syenite gives rise to dramatic, barren scenery with high cliffs and mobile screes, whereas the gneissose granodiorite of the Julianhaab Formation which underlies much of the rest of the area forms lower, more rounded and vegetated plateaux and valleys; fresh bedrock and features of glacial erosion are well preserved in this area (Glasser and Warren, in press).

149
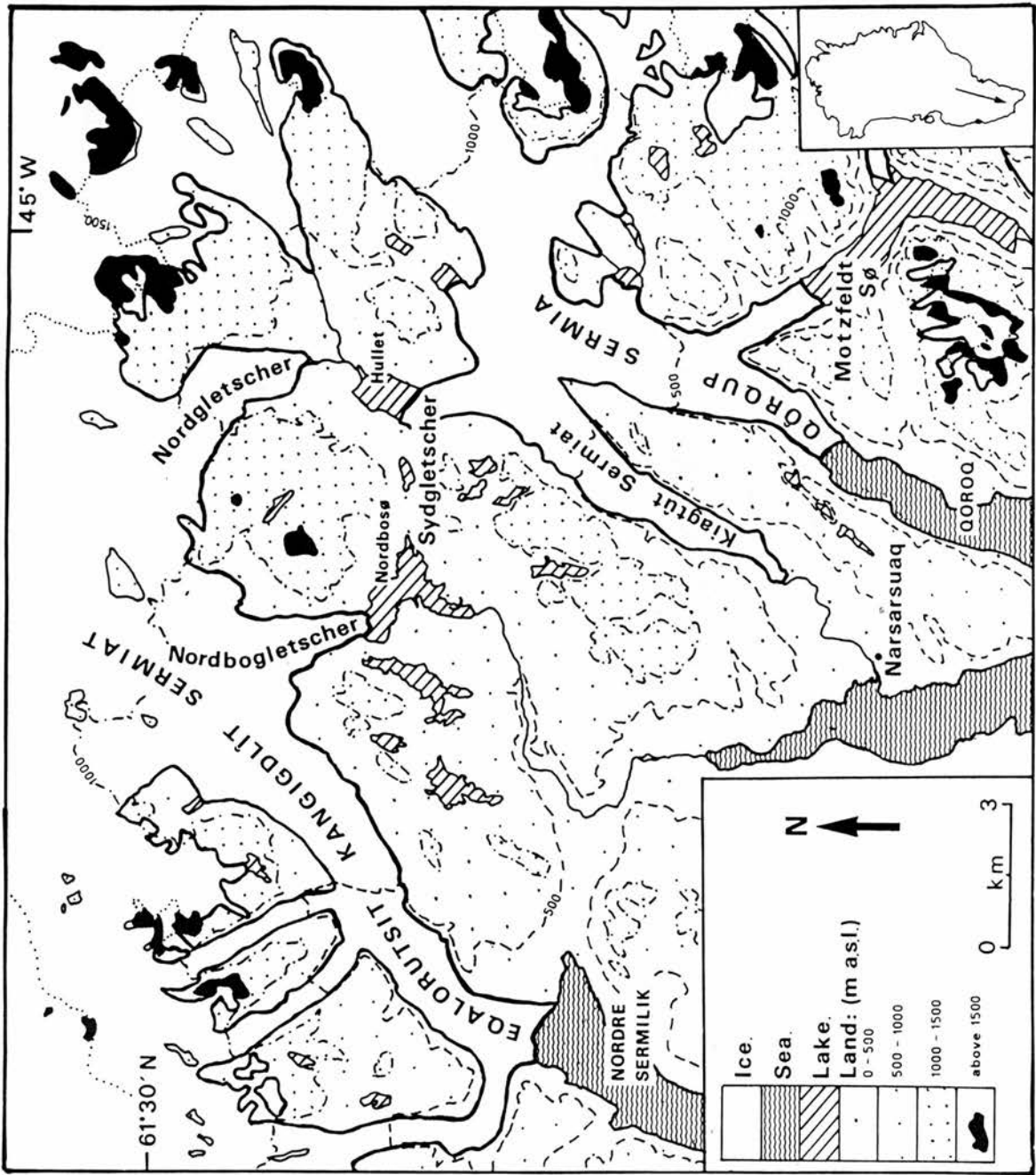
FIGURE 3.8. Location map of the Narsarsuaq area, based on the Danish Geodetic Institute's 1:250,000 series, Sheet 61 V3. Glaciers are shown near their 1953 positions.
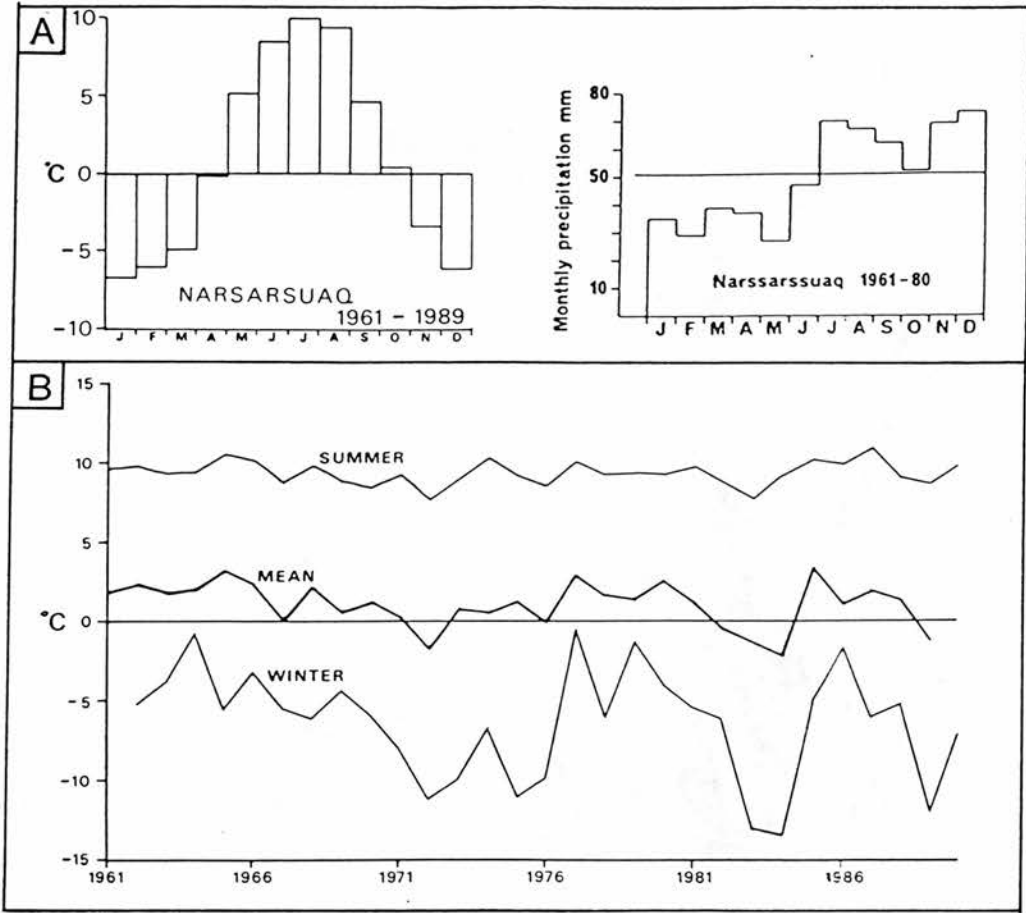
FIGURE 3.9. A. Recent climatic statistics from Narsarsuaq:
Mean monthly temperatures 1961 - 1989; mean monthly
precipitation 1961 - 1980,(after Clement, 1983b).
B. Temperature trends at Narsarsuaq, 1961 - 1989.
All data from the Danish Meteorological Institute.

Dwarf birch (Betula nana) over 2 m high grows in the
Narsarsuaq valley, and willow and birch scrub grows densely
up to 200 m a.s.l. in most areas.

Recent precipitation and temperature data from
Narsarsuaq, are shown in Fig. 3.9. Although temperature
statistics for Narsarsuaq only go back to 1961, the
similarity of the 1961 - 1989 trend with the same period of
the longer record at Nuuk shown in Fig. 3.1 permits a

151

reasonable degree of confidence that temperatures here have followed the same pattern.

Data concerning terminus changes were obtained from three sources:

i). Historical information collated from early travellers' and scientific reports by Weidick (1959).

ii). Stereo sets of aerial photographs taken in 1942/1953 (1:40,000) and 1981/1985 (1:150,000) (Table 3.3).

iii). Fieldwork reconnaissance, and updating of the aerial photograph results during June and July 1989. All seven major outlets were visited on foot; the terminal positions, surface characteristics, and the geomorphology of the immediate marginal zone were recorded, and the moraines profiled. Two of the three local glaciers above the lake Motzfeldt Sø were also examined.

GLACIER.	DATE.		REFERENCE NUMBER OF KEY SCENE.	
Qôrqup Sermia.	?	1942.	B34A	(56).
	August,	1953.	201K	(12498).
	July,	1981.	880C	(1570).
Glacier in	August,	1953.	201M	(12590).
Motzfeldt Sø.	July,	1981.	880C	(1570).
	July,	1985.	887L	(4308).
Kiagtût Sermiat.	?	1942.	B34A	(58).
	August,	1953.	201K	(12497).
	July,	1981.	880C	(1570).
Sydgletscher.	August,	1953.	201N	(12624).
	July,	1981.	880C	(1574).
Eqalorutsit Kangigdlit	August,	1953.	201J	(12469).
Sermiat.	October,	1981.	880A	(1620).
Nordbogletscher.	August,	1953.	201M	(12581).
	July,	1981.	880B	(1590).
Nordgletscher.	August,	1953.	202C	(12648).
	July,	1981.	880C	(1574).
Middle Glacier.	August,	1953.	201L	(12547).
	August,	1985.	887L	(4308).
West Glacier.	August,	1953.	201L	(12547).
	August,	1985.	887L	(4308).

TABLE 3.3. The aerial photographs used in this study, showing the dates of photography, and the Danish Geodetic Institute's reference number for the central scenes used for each glacier.

3.4.ii. Results.

Fig. 3.10 shows the frontal behaviour of the six outlet glaciers at which spatial change has occurred since 1942/1953.
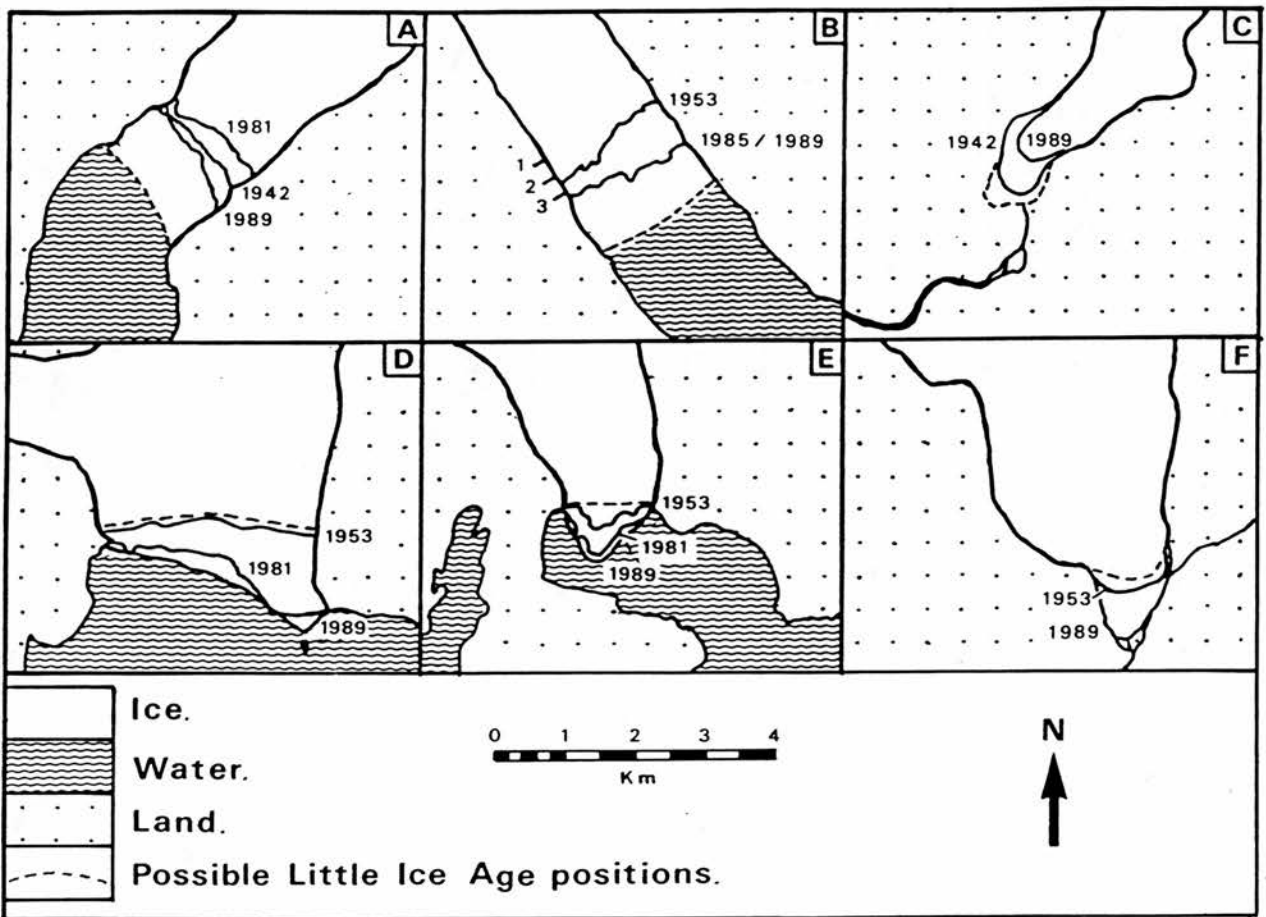
FIGURE 3.10. Sketch maps of frontal change since the Little Ice Age. A. Qôrqup Sermia. B. Motzfeldt Sø glacier. (1 - 3 mark the locations of the profiles in Fig. 3.16). C. Kiagtût Sermiat. D. Eqalorutsit Kangigdlit Sermiat. E. Nordbogletscher. F. Nordgletscher.

FIGURE 3.11. Aerial photograph of the Qôrqup Sermia system, July, 1981. (Number 880C - 1572; copyright Kort- og Matrikelstyrelsen, Denmark - A313-90).

3.4.ii. a). The Qôrqup Sermia System (Fig. 3.11).

1. Qôrqup   Sermia.

Description. The tidewater front (Fig. 3.12) has a chaotic,
crevassed surface, and terminates between steep cliffs at a
2 km long calving front in water of unknown depth.   The
terminal ice cliff is up to 25 m in height and calving events
occured every few minutes in early July 1989.   The fjord
Qôroq was choked with icebergs and brash ice for 3.5 km
beyond the calving front. No moraines are discernible in the
immediate ice marginal area because of the steepness of the
cliffs $(30^{o}$ $-50^{o})$ and the friable nature of the syenite.

Recent Behaviour.   Historical material is inconclusive but
suggests retreat during much of this century from a late
19th. Century maximum; presumed Little Ice Age (LIA)
trimlines preserved in the scree on the lower cliffs extend
1500 m beyond the present front to a narrowing of the fjord.
Between 1942 and 1981, net retreat of some 400 m occurred,
retreat occurring first on the west side (1942 - 1953) and
then on the east (1953 - 1981) in 'pendulum' fashion. Since
then, however, advance has begun, totalling 150 m on the west
side and 750 m on the east side.

2. Motzfeldt Sø glacier (Figs. 3.13 & 3.14).

Description. This glacier calves into the lake Motzfeldt Sø
which lies at 130 m a.s.l.   The 15 - 20 m high calving front
is 2.1 km long, heavily crevassed, and, in contrast to the
frequent, small-scale calving of Qôrqup Sermia, calves a few
large icebergs each year.   In 1981, one of the icebergs
measured 135,000 m^{2}, and similarly sized, tabular icebergs
were observed in 1989 together with some smaller icebergs.
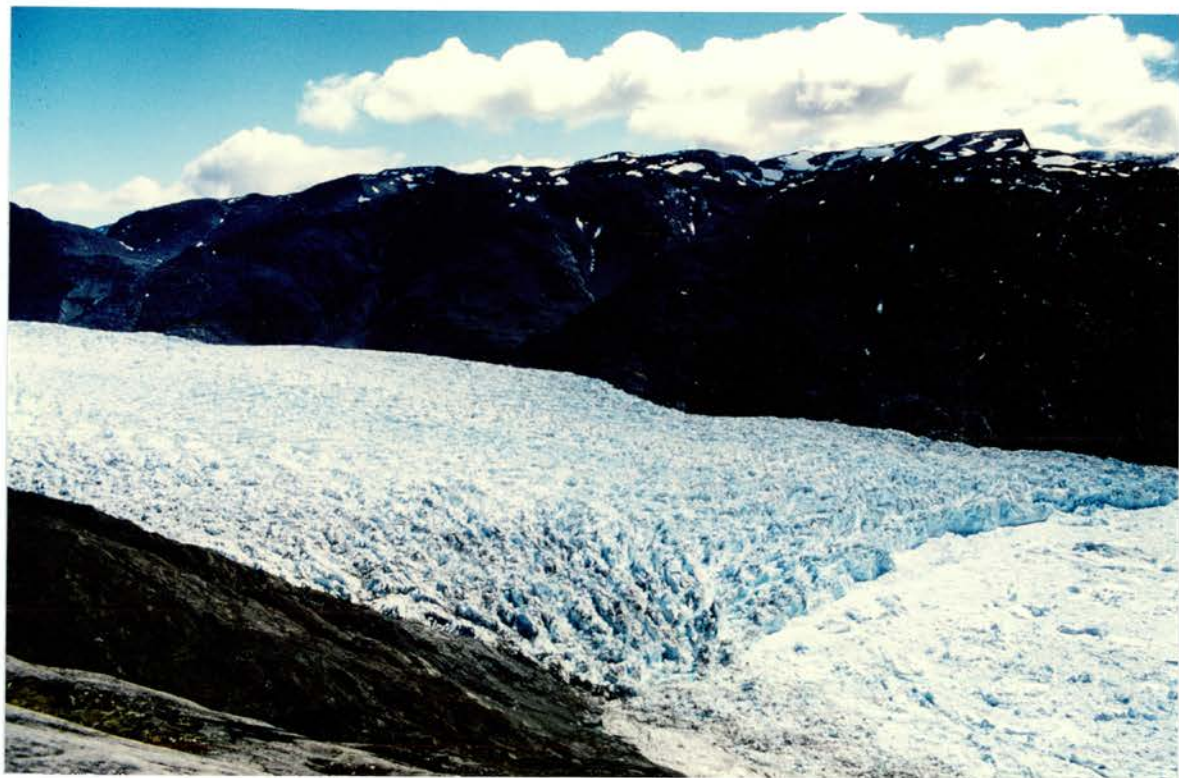
FIGURE 3.12.   Qôrqup Sermia from the west,   July, 1989.   The
     calving front is 2 km long.
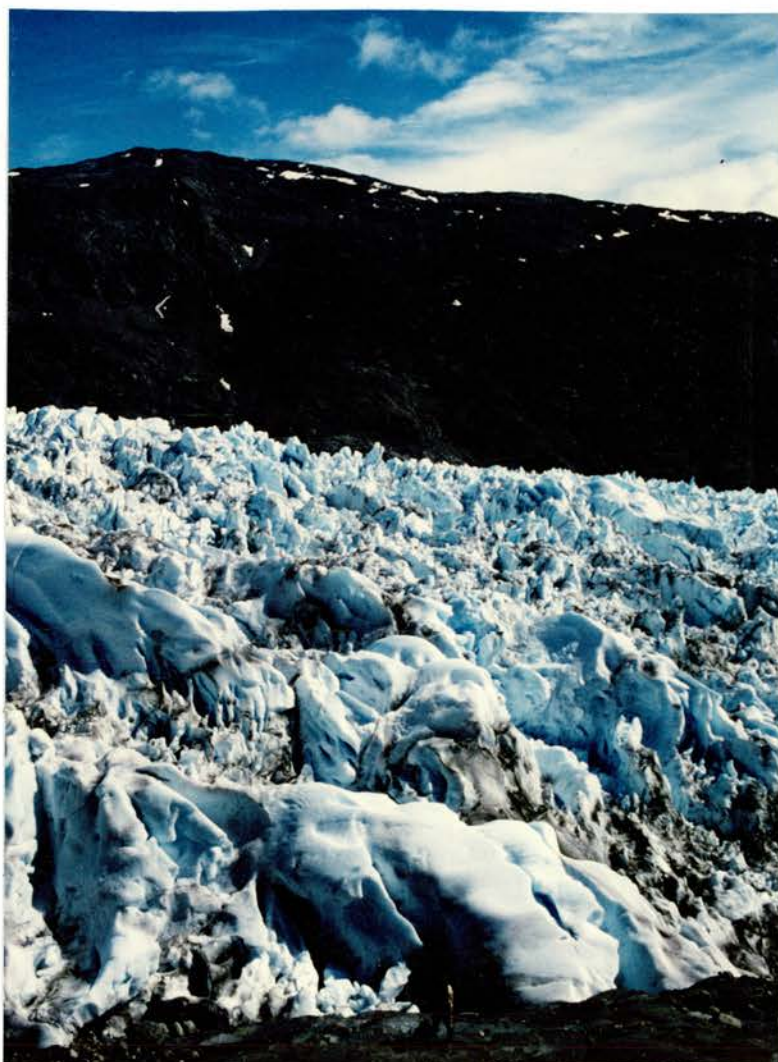
FIGURE 3.13.   Qôrqup   Sermia   from the west   side,   near   the
     calving front,   July,   1989.   Note figure lower centre
     for scale.

FIGURE 3.14. The Motzfeldt Sø glacier from the west, June, 1989. The glacier is 2.1 km wide and the cliffs rise about 1300 m from the lake.

FIGURE 3.15.   The Motzfeldt Sø glacier from the air, July,
     1989.

159

Recent Behaviour. No historical information exists for this inaccessible glacier, and on the eastern side, talus obscures any geomorphological record. On the western side, a large moraine, presumably dating from the LIA, parallels the ice margin and continues for 900 m beyond the present ice front. A suite of smaller lateral moraines run between the LIA moraine and the ice edge (Figs. 3.16 & 3.17), recording early 20th. Century shrinkage. No dateable material has been found, and there is no vegetation. Weidick (1963, p.75) noted that 2 km behind the front on the east side the glacier was thickening. Net advance 1953 - 1989 has totalled 360 m on the west side and 920 m on the east. Since 1985, the west side has advanced a further 30 m.

3. Kiagtût Sermiat (Fig. 3.18).

Description. This is a slow-moving (c. 60 m a^{-1}), land-terminating glacier separated from Qôrqup Sermia by a 70 m icefall at their junction above Sydgletscher. The surface is smooth, with large concentric arcs of debris-filled crevasses 3 - 5 m wide looping across the terminus zone, and the terminal 30 - 50 m is mantled by thick morainic debris. A shallow lake floods the immediate proglacial area and the last 15 - 20 m of the ice surface, before draining through a gorge down the Narsarsuaq river; fragments of sharp-crested moraine arcs marking about 10 former frontal positions break the water surface. The LIA trimline on the valley sides is clear, and the lake is dammed by a large (LIA?) former terminal moraine and associated outwash accumulations which form a 20 - 25 m wall across the main (southern) valley.
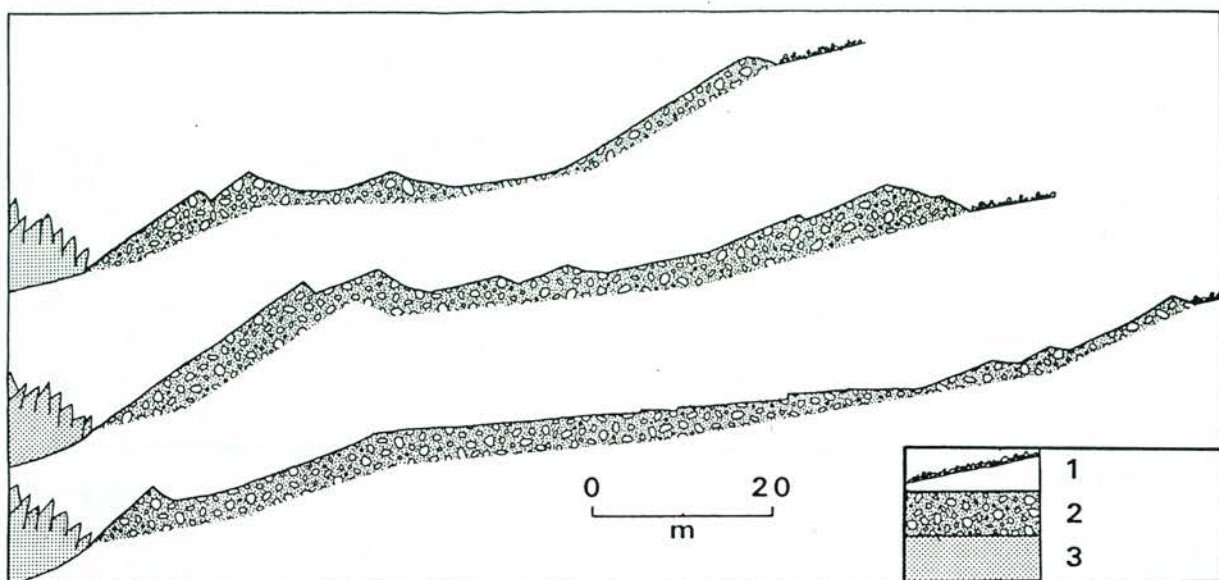
FIGURE 3.16. Cross-profiles of the lateral moraines beside the Motzfeldt Sø glacier. 1 = Amorphous boulder slope. 2 = Moraine. 3 = Ice margin. Profile locations are indicated in Fig. 3.10.B.

FIGURE 3.17. One of the lateral moraines beside the Motzfeldt Sø glacier, June, 1989.

FIGURE 3.18. The terminus of Kiagtût Sermiat from the east, July, 1989.

Recent Behaviour. The glacier has been slowly and consistently retreating since it was first reliably observed in 1876, when small-scale calving was occurring into a shallow proglacial lake (Steenstrup, in Weidick, 1959). Retreat continues today, with a net retreat of 550 m between 1942 and 1981. No discernible change in character or location is apparent since 1981.

4. Sydgletscher (Figs. 3.19 & 3.20).

Description. Sydgletscher flows almost due north into Lake Hullet with a centreline velocity of 365 m a^{-1} (Weidick, 1963). It descends into the lake in a staircase of pronounced, arcuate crevasses. A tongue of low-lying ice with large surface pools of water stretches north west across the lake and has pushed 20 m up the far shore forming two

162

FIGURE 3.19.   Sydgletscher and Lake Hullet from the north,
     July, 1989.   Note stranded icebergs top right marking
     pre-drainage lake level.   Push moraines profiled in
     Fig. 3.21 are in front of the dirty ice lower centre.

FIGURE 3.20. The western side of the front of Sydgletscher,
     where it is pushing up the lake shore to from the push
     moraines.

push moraines 1 - 3 m high (Fig. 3.21); this is very similar to the situation described by Weidick (1963). The calving front is 1 - 10 m high. The ice tongue is greatly affected by jökulhlaups; the lake drains under the glacier every one or two years. Lake levels fluctuate by as much as 120 m and this causes parts of the ice tongue to collapse and break up in the lake (Dawson, 1983; Clement, 1984a). The large flexure strains set up in the ice by the drainage and refilling episodes account for the broken nature of the terminal 2 - 3 km of the glacier. Following lake drainage in late June 1989, the lake had refilled to within 24 m of its
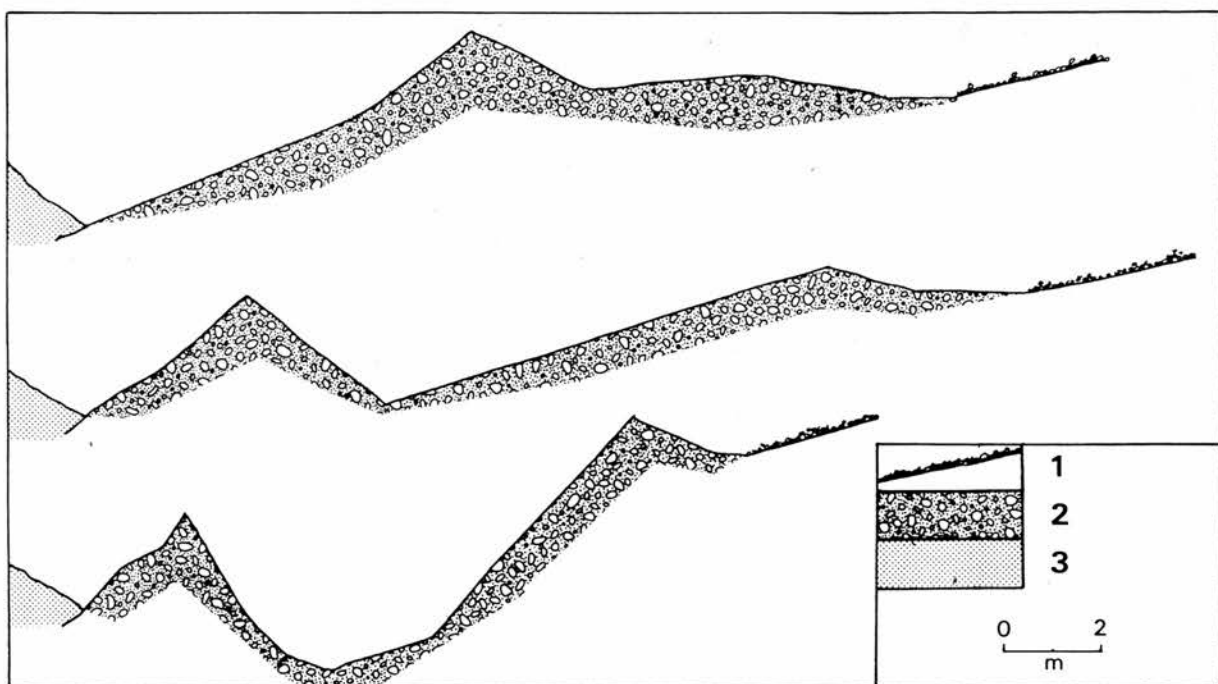
FIGURE 3.21. Cross-profiles of the push moraines at the snout of Sydgletscher. 1 = Amorphous boulder slope. 2 = Moraine. 3 = Ice front. Moraines consist of multigrade material, from clays and sands to large (1 - 2 m) angular boulders. Ridge nearest glacier is saturated and subject to small-scale mass movement.

pre-drainage level (marked by stranded icebergs and a clear shoreline) by July 21st. The southern half of Lake Hullet was covered by icebergs and brash ice (Fig. 3.19), but no calving was occurring, confirming that most of the mass loss occurs during and immediately after the jökulhlaups.

Recent Behaviour. The LIA maximum position, inferred from a large moraine and a trimline (Weidick, 1963), was some 1 - 2 km in advance of its present location. The nature and area of the snout in 1989 was similar to that on the aerial photographs of 1953 and 1981. While showing no marginal retreat in that interval, the tongue became more dissected. Transverse crevassing was characteristic and the tongue appeared stagnant in both years. However, between those years, ice thickened and encroached slightly onto the headland to the south around which the ice flows from Qôrqup Sermia. By 1989 , while still very dissected, the tongue had a continuous strain pattern of longitudinal, curving crevasses linking the arcuate crevasse area with the west shore.

3.4.ii. b). The Eqalorutsit Kangigdlit Sermiat System.

All three glaciers of this system are in their most advanced position in historical time (Clement, 1982).

1. Eqalorutsit Kangigdlit Sermiat (Figs. 3.22 & 23).

Description. The main glacier is fast-flowing (1500 m a^{-1} - Knudsen, 1983), profusely crevassed, and terminates at a 3.2 km long, 20 - 50 m high calving front producing an abundance of icebergs and brash ice which fills Nordre Sermilik Fjord for almost 14 km. The only depth sounding near the ice front (by A.Jessen in 1894, in Weidick, 1959)

165

FIGURE 3.22. Aerial photograph of Eqalorutsit Kangigdlit Sermiat, October, 1981. (Number 880A - 1620; copyright Kort- og Matrikel-styrelsen, Denmark - A313-90).

gave a depth of c.400 m. There are no trimlines above or beyond the ice, and the glacier has been in contact with marginal vegetation since at least 1955 (Weidick, 1959). Recent Behaviour. The location of the glacier front changed little between 1894 and the middle of this century, but by 1955 there were signs of strong advance (Weidick, 1959). Comparison of descriptions of 1894 and a photo of 1932 (both in Weidick, 1959), reveals that the character of the eastern side had evolved from a gently sloping ramp to a calving

166

FIGURE 3.23.   Eqalorutsit Kangigdlit Sermiat from the east,
        July, 1989.

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cliff; thus although the frontal location did not change
appreciably until the early 1950s, the build-up towards
advance began in the 1920s.   This is confirmed by the
changing nature of the junction with the western tributary
glacier near the ice front.   In 1894, this glacier was
described as 'pushing down over the main glacier' (A.Jessen's
diaries, quoted by Weidick, 1959, p.55).   The 1932 photo
shows the ice of the main glacier above that of the tributary
and in 1989 there was a substantial height differential
between the contorted surface of the main glacier and the
dirty, smooth ice of the tributary.   In 1953 the 'junction
moraine' between the two was almost straight, but by 1981 it
had been bowed back 600 m westward into a large arc,
distorting medial moraine behind (Fig. 3.22).   Like Qôrqup

167

Sermia, the ice front has changed in pendulum fashion, advancing just 100 m on the west side between 1953 and 1981, 750 m in the centre and over 1000 m on the east side, where in early July 1989 it projected beyond the headland to within 150 m of a small rocky island, a further advance of 150 m.

2. Nordbogletscher (Fig. 3.24).

Description. The clean ice of Nordbogletscher flows down into the lake Nordbosø (660 m a.s.l.) where it ends in a 2 km long calving cliff 7 - 15 m in height. Depth soundings in Nordbosø gave depths of 50 - 150 m but identified no end moraines (Knudsen, 1986; Clement, 1984b). Ice velocities are of the order of 90 m a^{-1} near the terminus, rising to 350 m a^{-1} at 1050 m a.s.l. (Clement, 1984b). In late July 1989, calving events were occurring once or twice per hour, producing small, prismatic icebergs which were rapidly removed from the ice front by a strong subglacial meltwater current and a katabatic wind.

Recent Behaviour. No historical records exist, but advance commenced between 1942 and 1953 (Clement, 1983b) and the central section advanced a total of 700 m between 1953 and 1981, thickening by 45 m above the 1953 snout location (Knudsen, 1986). Mass balance studies revealed that the glacier had a negative net mass balance in the years 1979 - 1982 (Clement, 1983a), and yet the advance continues; by 1989, the central point had advanced a further 150 m from its 1981 position, and advance has been recorded each summer during 1987, 1988 and 1989 (Funk, pers. comm.). However, the glacier margins at the entrance to the lake have remained essentially stable.

FIGURE 3.24.   Nordbogletscher from the east, July, 1989.
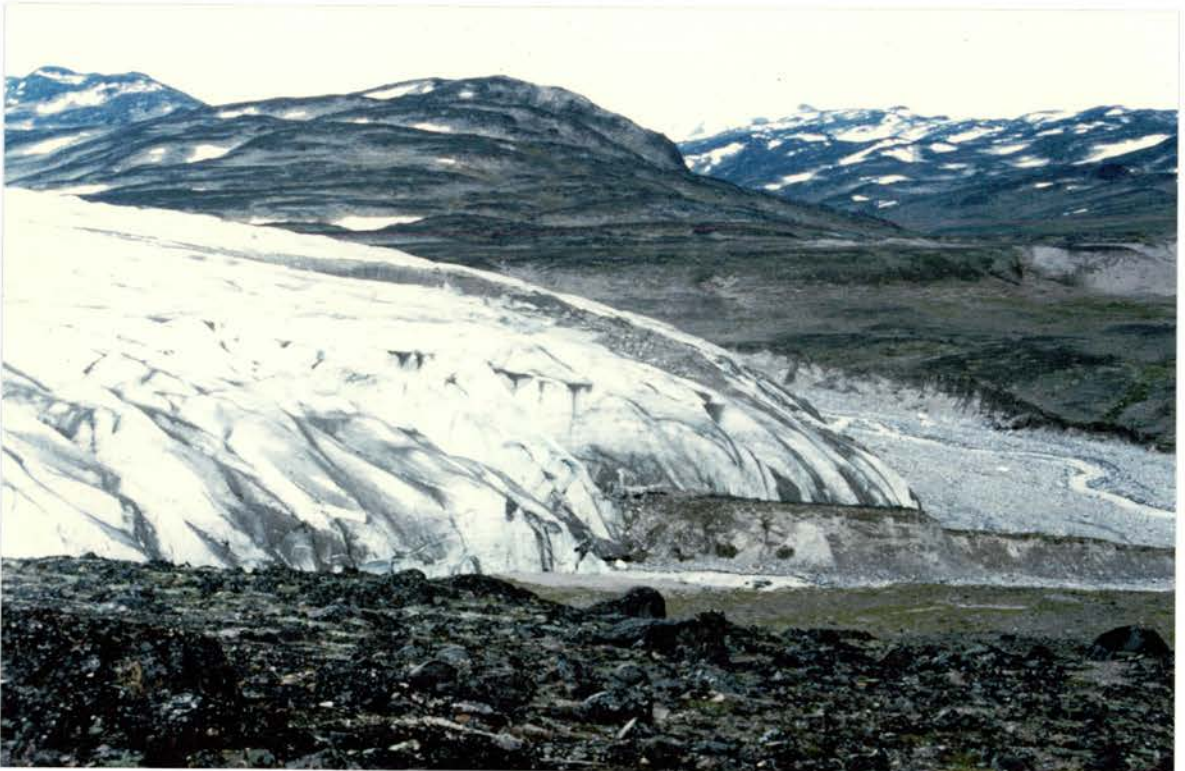
FIGURE 3.25.   Nordgletscher from the west, July, 1989.

FIGURE 3.26.   The advancing front of Nordgletscher, July,
        1989.

3. Nordgletscher (Fig. 3.25.)

Description. This is a land-terminating glacier 650 m a.s.l.
which terminates in a steep 25 - 30 m cliff.   The ice is
generally clean except for a large medial moraine.   The
western lateral margin for 500 m back from the snout also
consists of a steep cliff over 10 m high in places.

Recent Behaviour.   There is no information about the LIA
maximum nor any remaining geomorphological evidence.   The
glacier has been steadily advancing since at least 1953.   By
1981, net advance had totalled 750 m, with 120 m of
thickening above the 1953 terminus location (Knudsen, 1986).
The characteristics of the glacier show that advance has
continued since then.   Since 1960 (Weidick, 1963), the
terminal ice cliff has increased in height by c.10 m, the ice

170

surface behind the front is steeper, and there has been a decrease in supraglacial morainic cover. Unfortunately, quantification of the advance was not possible due to the limited resolution of the aerial photographs and lack of

FIGURE 3.27.   Middle Glacier from across the valley, June, 1989.

unambiguous landmarks. Advance for the period 1981 - 1989 was visually estimated at 100 m, which indicates a slowing of the advance from the rate of 50 m a^{-1} measured in 1980/1981 (Clement, 1982).

3.4.ii. c). Local glaciers above Motzfeldt Sø (Figs. 27 & 28).

Three local glaciers are nourished by the high snowfields above Motzfeldt Sø. The easternmost of these is inaccessible, its valley ending at a cliff high above the lake, but at the peak of the Little Ice Age, the western pair extended over 1 km down into the valley below where their maxima and a subsequent position are marked by clear terminal moraine arcs up to 3 m high (Fig. 3.28), as first described by Vorndran and Sommerhoff (1974). Adjacent to the glaciers, these LIA moraines are ice-cored ramparts 20 - 30 m high, with sideslope gradients of up to 32^{0}. Net retreat occurred between 1953 and 1981, totalling c.175 m for West Glacier and c.75 m for Middle Glacier. However, net advance of about 30 m has since occurred at West Glacier, and, although the resolution of the photographs does not allow measurement of advance of Middle Glacier, the ice surface is now cleaner than in 1953, and the ice fronts of both are now steep and clean (Fig. 3.29), terminating at gradients of 28^{0} and 25^{0} respectively.

FIGURE 3.28.   West Glacier, June, 1989.   Note the large
    Little Ice Age moraines running down into and across
    the valley.
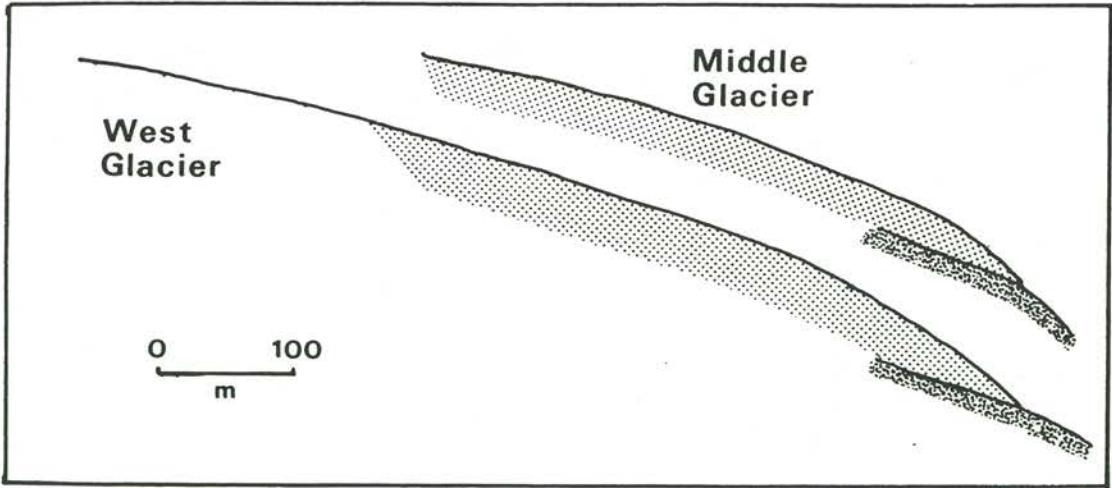
FIGURE 3.29. Long-profiles   of   the   terminal sections of
    Middle Glacier and West Glacier.

173

3.4.iii. Discussion.

Despite a very similar climatic setting, the glaciers of the Narsarsuaq region have exhibited strongly contrasting terminal behaviour since the Little Ice Age (Fig. 3.30). Clement (1981,1982,1983a, 1984b), Weidick (1963,1988), and Knudsen (1983,1986) have all commented on the complex glacio-dynamics of the area. Clement (1982) states that the reasons for these contrasts are unknown but that they 'seem to be dynamic rather than climatic' (p.95). To gain a full understanding of the combination of controls that explain such diverse glacial response to climatic forcing would require a large body of data for each glacier over a substantial time period. Mass balance data exists only for Nordbogletscher (Clement, 1983b). Subglacial topography, ice depths, calving fluxes, calving speeds, and the bathymetry of the fjords and lakes near the ice fronts remain largely or totally unknown. No rigorous tests of hypotheses are possible, therefore, but the information collated here is sufficient to indicate that:

i)   topography is important in controlling the frontal oscillations of calving fronts, and,

ii) calving and land-terminating outlet glaciers respond in different ways to the same climatic input.

3.4.iii. a). Topographic control of calving glaciers.

Topography can influence the response of land-terminating glaciers (eg. Kick, 1989), but can be an important control on the frontal change of calving outlet glaciers (Funder, 1989). Topography is of especial significance for glaciers where iceberg calving is the

174
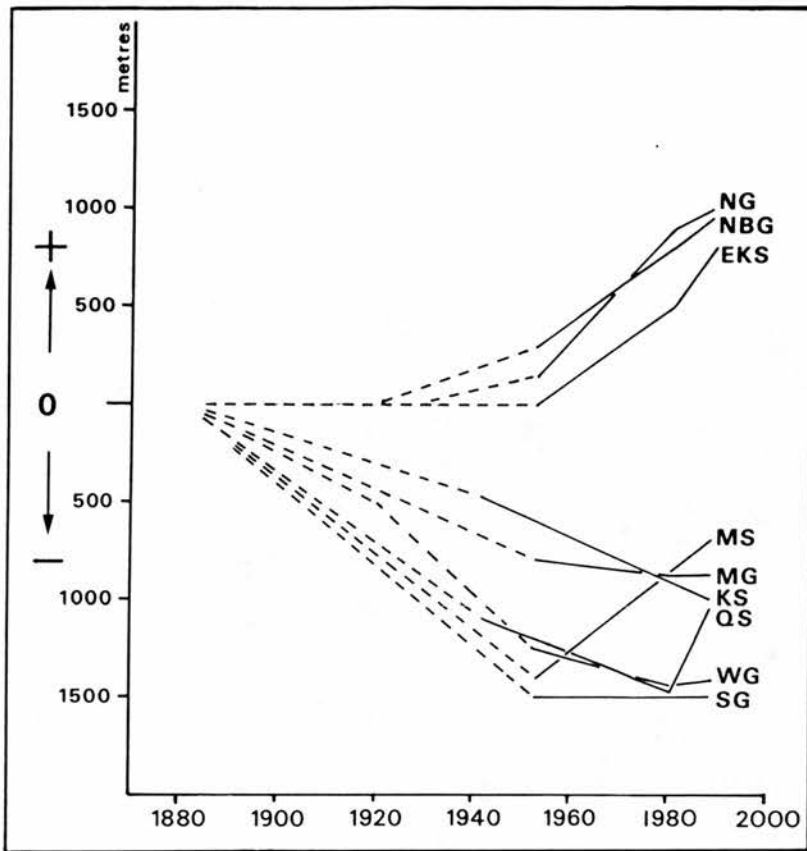
FIGURE 3.30. Known and inferred frontal change of the study
    glaciers since their positions at the peak of the
    Little Ice Age, taken to be c.1880.   NG =
    Nordgletscher. NBG = Nordbogletscher. EKS = Eqalorutsit
    Kangigdlit Sermiat. MS = Motzfeldt Sø glacier. MG =
    Middle Glacier. KS = Kiagtût Sermiat.   QS = Qôrqup
    Sermia. WG = West Glacier. SG = Sydgletscher.

dominant   ablation   mechanism   because,   by   affecting   the

calving   rate,   it   largely   controls   rates   of   mass   loss.

Narrows,   shallows,   bends   and   bifurcations   in   a   fjord

represent places of potential stability, or 'pinning points',

at which stillstands are to some degree preconditioned.   To

advance or retreat from such topographic pinning points, a

glacier   must   exceed   thresholds   of   stability,   and   thus   a

calving glacier will tend to exhibit a stepped, non-linear

terminal response to smooth climatic forcing.

175

Such topographic control is well exemplified by the behaviour of the two calving fronts of the Eqalorutsit Kangigdlit Sermiat system. Nordbogletscher has been consistently advancing in its central section but the sides have remained 'pinned' at the narrows at the north end of the lake (Fig. 3.10.E). Thus the terrestrial glacial geomorphological record will carry little trace of the 850 m advance but may, on the contrary, misleadingly suggest a substantial period of equilibrium.

It is also most likely that topography has damped the advance of the main tidewater front. The advancing trend has been strong, the build-up probably beginning in the late 1920s and the frontal expression of it commencing in the mid-1950s, and yet this large, active glacier has advanced just 100 m on its west side and 750 m in its central section. Even on the eastern, most active side, the advance of 1150 m represents an average of just 2% of its flow velocity each year. The proposed explanation for this lies in the topographic setting. The front is located at a striking pinning point at the head of Nordre Sermilik Fjord (Fig. 3.10.D); the width of the trough at this point is 3.3 km, but advance beyond this point would take the front out into an embayment over 6 km in width. The dramatic increase in calving rate that would result represents a prohibitively steep ablation gradient. Thus the glacier front is held at this constriction, a location at which it could probably remain throughout a considerable range of climatic conditions. Again, this shows that the combination of iceberg calving with particular topographic configurations

may render a glacier relatively insensitive to climatic change. In this context, it is of interest to note that the LIA trimline beyond Qôrqup Sermia shows that this glacier's historical limit also coincided with a 'pinning point', a fjord narrowing 2 km wide which precedes an embayment 2.5 km wide (Fig. 3.10.A).

3.4.iii. b). Contrasting frontal behaviour.

The Qôrqup Sermia system is a striking illustration of the contrast between calving and land-terminating glaciers. All four glaciers are fed by one ice stream but they have exhibited opposing trends of frontal behaviour during the last 40 years. The land-terminating, retreating Kiagtût Sermiat seems to be responding primarily to the rise in ablation-season temperatures following the Little Ice Age, and has not yet stabilised as a reaction to the cooling of the last 30 years. In contrast, the tidewater front and the lake-calving front in Motzfeldt Sø are both now advancing. Even Sydgletscher, which must be regarded as exceptional because of the jökulhlaup activity, shows by its thickening margin and changed strain patterns since 1981 that it too might be advancing but for the catastrophic effects of lake level fluctuations. The behaviour of the local glaciers of the region, those above Motzfeldt Sø, and also Valhaltindegletscher and Narssaq Brae nearby on which mass balance studies have been undertaken (Clement, 1984b), also appears to match the climatic signal, though with a greater time lag.

There are several possible explanations for the recent behaviour of these calving glaciers and the contrast with

their land-terminating neighbours:

1. The advancing trend may be independent of climate, part of an advance-retreat cycle analagous to tidewater glacier behaviour described in Alaska (Mayo, 1988). The historical behaviour of the nearby Eqalorutsit Kitdlit Sermiat glaciers which retreated 1.5 - 3 km between 1894 and 1953 but have since been advancing strongly (Fig. 3.7.A) could be an example of such cycles. Regular monitoring over decades or centuries, and/or detailed, finely dated chronologies of Holocene fluctuations, would be needed to test whether outlet glaciers are subject to such cyclical behaviour (cf. 5.2).

2. Climatic cooling since the 1940s may have indirectly affected the calving rate by reducing the subglacial meltwater activity. Calving rates can be positively correlated with meltwater events and the resulting increase in buoyancy forces (Sikonia, 1982). Data to test this idea do not exist.

3. Finally, and most probably, the explanation lies not in one simple cause-and-effect relationship but in a complex set of interactions between climatic inputs, variable response times, interactions between feeder glaciers and their distributaries, calving dynamics, and topographic controls. Weidick's (1963) extensive work in the area led him to the tentative conclusion that:

> ... the inland ice and Julianehaab ice cap in the Narsarsuaq region are swelling in their upper parts (p.78).

Knudsen (1986) showed that the long profiles of Nordgletscher and Nordbogletscher possibly reflect a sudden increase in mass balance which created a kinematic wave. On

the assumption that this is the case, and given that most calving glaciers in South-West Greenland are now advancing while most land-terminating glaciers are still retreating, the recent frontal changes of the glaciers in the region may be explained by a simple model of variable response times, calving dynamics, and topographic control.

The calving glaciers, having the highest velocities, reflect the growth trend first. Frontal expression of this is seen earlier at the lake-calving fronts than at the tidewater fronts because, for a combination of environmental reasons, lake-calving glaciers are probably more stable than tidewater fronts (3.5.ii). Thus Nordbogletscher begins to advance from the mid-1940s and the calving front in Motzfeldt Sø from the 1950s. Subsequently, the tidewater fronts begin to advance, accompanied by substantial calving. Eqalorutsit Kangigdlit Sermiat began to advance in about 1955 and Qôrqup Sermia from probably the 1970s. A small part of this difference in response times may be due to the geography. The lake-calving glaciers are slightly nearer the source of the kinematic wave which would therefore arrive a few years earlier. In this model, the land-terminating glaciers are the last to experience advance because of their much lower mass exchange rates relative to calving glaciers. Kiagtût Sermiat has yet to respond to the regional advancing trend.

The detailed changes of the calving fronts, and overall extent of advance, can be explained in terms of topographic control, as first proposed theoretically by Mercer (1961) and confirmed empirically for outlets of the Greenland Ice Sheet

by Funder (1972,1989). Topographic 'pinning points' (in these cases, headlands in the fjords), have provided the stable 'hinges' for the pendulum oscillations of the calving fronts, both during retreat and advance. Such pendulum behaviour has occurred at the tidewater fronts of the nearby Eqalorutsit Kitdlit Sermiat glaciers as well as in the fjords described here, and is typical in many fjords in West Greenland (Fig. 3.7). The limit of the advances, both during the Little Ice Age and recently, has been strongly influenced by trough geometry. With the exception of the glacier in Motzfeldt Sø, which is flowing in a valley whose width varies little, the stable marginal locations adopted by all the calving fronts coincide with topographic pinning points.

This model provides a conceptual framework within which the main trends can be explained, but it does not account for all the observations. The strong advance of Nordgletscher remains an unexplained anomaly, both within this area and in the wider regional context (3.3.i.a). Clement (1984b) accounted for it as an 'overspill effect' from the advance of the main glacier. Such an effect might also explain the advance of Nordbogletscher, but it could not apply to the Qôrqup Sermia system, even if all the glaciers of the latter were advancing; whereas the distributary glaciers of Eqalorutsit Kangigdlit Sermiat discharge laterally from high up the ice stream, those of Qôrqup Sermia diverge in digitate fashion, and, with the exception of Sydgletscher, terminate at or near sea level. The degree of dependence or independence of the individual distributary glaciers of such

180

diverging systems is a related, intriguing question.

The most important unexplained contrast in glacial behaviour, however, is that between the glaciers of the Eqalorutsit Kangigdlit Sermiat system, and all the other glaciers in the region. The former have shown signs of advance from the early part of the century and are now at their greatest historical extent, while the latter retreated rapidly from Little Ice Age maxima in the first half of this century (Fig. 3.30). Localised climatic contrasts are an unsatisfactory explanation for such a sustained, pronounced contrast. A plausible mechanism that could account for it is a migration of ice divides across subglacial bedrock thresholds, resulting in changing ice sheet streamline patterns and drainage basin areas. Such a mechanism was first suggested by Weidick (1968) for the Jakobshavn drainage basin, and was developed by Bindschadler (1984).

3.5. DISCUSSION.

3.5.i. Controls on glacier oscillations:
climate and topography.

The recent patterns of glacier oscillation throughout West Greenland emphasize the importance of the terminal environment and the geometry of glacial troughs in controlling the type and effectiveness of the dominant ablation mechanism, and, thereby, the nature and the extent of glacial response to climatic forcing. During this period of climatic transition from the end of the early 20th century warming through to a time of gradual cooling, many glacier characteristics that might have been expected to have an influence on frontal change correlate only very weakly with the observed oscillations, or show no relationship at all. The strongest pattern to emerge is the contrast between calving and non-calving glaciers. The non-calving glaciers are responding directly to the climatic forcing, though with variable lag times. Slowing retreat rates and the possibility of readvance were just discernible in the late 1960s (Weidick, 1968). Although these data reveal net retreat for most of these glaciers, the net advance of a few, and the transformed frontal character of several, confirms that a more general readvance of these glaciers is now in progress.

In contrast to this, the great variability in the behaviour of the calving glaciers demonstrates that controls other than climate influence their fluctuations. The only local influences which seem to be able to explain the variable patterns of frontal change are topographic factors. In Alaska it has been found that changes in the width of the

calving front have a fundamental impact on mass balance and may force advance, retreat or stability that is unrelated to climate (Meier and Post, 1987). Intuitively, it might be expected that the relationships developed from the Alaskan research would not hold for Greenland because of the very different glaciological context, which includes some floating termini, and exclusively ice-bounded catchments, permitting migration of ice divides and drainage capture. This was suggested as a means of explaining anomalous behaviour of tidewater glaciers by Weidick (1968, p.45). In ice-bounded catchments, it is possible that relative discharge rates determine the size and shape of the drainage basins rather than vice versa (Bindschadler, 1984).

However, the recent behaviour of calving glaciers in Greenland, as exemplified in Fig. 3.7, strongly suggests that the calving rate - 'pinning point' relationship at the terminus is strong enough to render these complexities insignificant, and therefore that the relationships developed in Alaska can be applied to ice sheet outlet glaciers. The retreats of Glaciers 5, 8 and 71, for instance, are examples of climatically-triggered calving retreats that have been glacio-dynamically amplified; the retreat was initiated by climatic warming, but the subsequent behaviour has owed more to the internal dynamics than the external forcing. Also, stability that is directly attributable to topographic configurations is evident. Narssap Sermia (Fig. 3.7.D) is the best example of this; the climatic fluctuations of the last 150 years have not caused the glacier front to move from its location at a narrow bend in the fjord.

It therefore seems clear from these results that, during the last forty years, the outlet glaciers have not been responding directly to the regional climatic forcing. Despite the wide variety in site characteristics and a latitudinal range of 11°, the degree to which the glacier oscillations have been a primarily climatic function has been determined by the terminal environment; the land-terminating glaciers appear to be reacting primarily to the climatic signal, while the behaviour of the calving glaciers has been much more variable. Once iceberg calving is an important ablation mechanism, the interaction of the glacio-dynamics with the local topographic geometry can largely decouple terminus change from climatic control.

3.5.ii.   Lake-calving glaciers: responsive to climate?

Half of the 24 lake-calving glaciers have advanced, and
another two are showing signs of advance.   Of the rest, 7
have been retreating and 3 have been stable.   These
proportions are similar to those for the tidewater outlets
(Table 3.2).   If calving outlets of ice sheets oscillate
cyclically as do other calving glaciers, at any one time a
higher percentage of all the calving lobes will be advancing
than retreating because the retreating phase of the cycle is
approximately an order of magnitude shorter than the
advancing phase (Mayo, 1988).   Interestingly, though, it
appears that there may be a contrast between the tidewater
and freshwater glaciers.   Eight of the latter should probably
be regarded as exceptional, for the following reasons.   Three
are those whose retreats may have been amplified glacio-
dynamically (Nos. 5, 8, and 71, and this last shows signs of
readvance); three are outlets of the Frederikshaabs Isblink
which is contiguous with the main ice sheet but dynamically
independent (Nos. 21, 22 and 23); and one (No. 66) terminates
in a shallow proglacial pond, showing no signs of calving,
which contrasts with the sizeable rock-cut troughs into which
the other glaciers in the sample calve.   Sydgletscher
(No.16), which is showing signs of advance, is an unusual
case, being affected by regular jökulhlaups which break up
the terminus (Dawson 1983).   If these eight glaciers are
removed from the sample, a strikingly positive trend emerges
for the lake-calving glaciers: 75% are advancing, another
(No. 20) shows signs of imminent advance, and only one
(No. 72) is retreating.

185

This positive trend, if real, is unexpected. Existing theory does not suggest that the glacio-dynamic response of tidewater and freshwater glaciers to climatic forcing is systematically different. However, Funk and Röthlisberger (1989) have collated empirical results from lake-calving glaciers in Alaska, Greenland and the European Alps which show that calving rates in any given water depth are more than an order of magnitude lower in freshwater than tidewater. Explanations for this contrast remain largely conjectural at present. Funk and Röthlisberger (1989) accounted for the difference in terms of the greater buoyancy of glacial meltwater in seawater, an hypothesis which requires a strong positive feedback through melting to calving, since melting at the calving front probably contributes only a small fraction of the total ice loss. The absence of tides, the relatively smaller influence of wave action, and the greater annual range of water temperature may all be contributary to the contrasting rates of calving.

For environmental reasons, lake lobes may therefore be more stable mechanically than tidewater fronts and thus able to advance in situations in which tidewater calving losses would be prohibitive. Contrasts such as these may partly explain the differing magnitude and frequency of calving events at lake and tidewater fronts. To compare directly the freshwater and tidewater glaciers of West Greenland may not be very realistic because their flow velocities vary widely and are poorly known, although, in general terms, the former are slower than the latter, and calving glaciers tend to be faster flowing than land-

terminating glaciers. Nevertheless, if other variables were held constant, the systematically lower calving rates in lakes might be expected to affect the relative rates and magnitude of frontal change of the two types of calving glacier. Oscillations of lake-calving glaciers are likely to be smaller, retreats slower, and advances more rapid. Fig. 3.4 shows that the oscillation distances are mostly smaller in lakes. Advance and retreat rates are not comparable here because of the absence of ice velocity data. Given that a general readvance of the West Greenland Ice Sheet is underway, it seems that some lake-calving glaciers are reacting most rapidly and consistently; therefore, of all the types of outlet glacier, they may be the first to reflect climatic deterioration.

3.6. CONCLUSIONS.

The dominant controls on the oscillations of ice sheet outlets are determined by the terminal environment. Non-calving glaciers respond primarily to changes in summer temperatures. Topography is an important control on the fluctuations of ice sheet margins because it can greatly amplify or damp the glacio-dynamic response to climatic forcing by introducing thresholds of stability to the system (Payne and others, 1989). Climate may be the forcing function which causes the threshold to be exceeded, because, unlike land-terminating glaciers, calving glaciers cannot stagnate in situ, making them especially sensitive to climatic warming. However, once such a threshold has been crossed, frontal change may become decoupled from climatic control such that the spatial behaviour of the glacier front is controlled by topography. Thus the response of calving glaciers to climatic change is mediated through the influential 'filter' of trough geometry.

C H A P T E R   F O U R .

PAST GLACIER FLUCTUATIONS IN WEST GREENLAND:

A Case Study in the Ilulissat (Jakobshavn) Area.

4.1. INTRODUCTION.

An important question arising from the conclusions of the previous chapter is the extent to which non-climatic influences affected the margins of the Greenland Ice Sheet during the retreat from its glacial maximum position. It is known that climatic change in Greenland at the beginning of the Holocene was at times very rapid indeed. Dansgaard and others (1989) have been able to show that, at the Dye 3 borehole site on the South Greenland ice dome 10,700 years ago, the $^{16}O/^{18}O$ ratio in the ice shifted by 0.5% in less than 50 years, representing a $7^{o}C$ warming. The deuterium excess and dust concentration signals shifted in less than twenty years. This means that half the total Pleistocene - Holocene warming in South Greenland took place in less than half a century, and perhaps in as little as two decades. This is not unrealistic in the light of the $4^{o}C$ warming experienced in North Greenland during the 1920s.

In the context of such dramatic climatic change, it would perhaps be expected that the more finely balanced topographic and glacio-dynamic controls that affect the ice sheet margin in its current state of near equilibrium, would not have been operative, or at least, not as effective. Clearly the retreat behaviour of an ice sheet margin is not controlled exclusively by either climate or topographic

189

influences, there being a constant interplay between the two. However, the significance attached to these two categories of control by different workers is widely divergent. The status of this debate is summarised in 1.3.ii.b, and it is a problematic conflict to resolve because of the generally imprecise, indirect dating of the moraines (Kelly, 1985). On the one hand, Weidick (1968,1985) and Ten Brink and Weidick (1974) have accounted for the moraine systems of the retreat stages in West Greenland in primarily climatic terms, while noting the influence of topography in certain instances. On the other, Kelly (1985) highlights the possibility that:

> ... the moraine system continuity is more apparent than real and [the moraines] could be explained as metachronous local stillstands during a period of slower retreat. (p.477)

Funder (1985,1989) has stated that most of the stillstand locations along fjords have been determined by variations in channel shape and do not necessarily carry climatic significance.

In order to test the extent to which fjord geometry affected the retreat stages of the ice sheet margin during the early Holocene, a case study of one fjord system in central West Greenland was undertaken during August and September 1988. Mann (1986) shows that the contexts in which calving glaciers are most susceptible to non-climatic controls are:

i) complex fjord systems.

ii) areas of low Equilibrium Line Altitude (ELA).

iii) fjord mouths adjoining the open sea.

The work of Brown and others (1982), Funk and ROthlisberger (1989), and Pelto and Warren (in press) has shown that

calving rates are linearly related to water depths in all situations studied to date. Hughes (1987), Hillaire-Marcel and others (1981), and Funder (1985,1989) have demonstrated the significance of the sea/land transition as a location at which stillstands almost always occur during ice sheet retreat.

Therefore, the best place to carry out a case study would be a complex fjord system for which bathymetric data exists, and in which, at some point, the ice front occupied a fjord-mouth position and experienced a sea/land transition. The Tasiussaq Fjord system south of Jakobshavn fulfils all these criteria.

4.2. THE STUDY AREA.

In central West Greenland, the coast swings east to form
the large bay of Disko Bugt which is dominated by the island
of Disko (Fig. 4.1). The ice-free mainland here consists of
a 15 - 40 km wide coastal strip of plateaux and hills
(≤ 550 m a.s.l.), separated by long, structurally aligned
glacial troughs in Precambrian gneiss. The major fjord is
Jakobshavn Ice Fjord which is constantly choked with large
icebergs and brash ice from Jakobshavns Isbrae, a 6 km wide
calving glacier which discharges almost 8% of the total
annual ice loss of the Greenland Ice Sheet (Bindschadler,
1984).
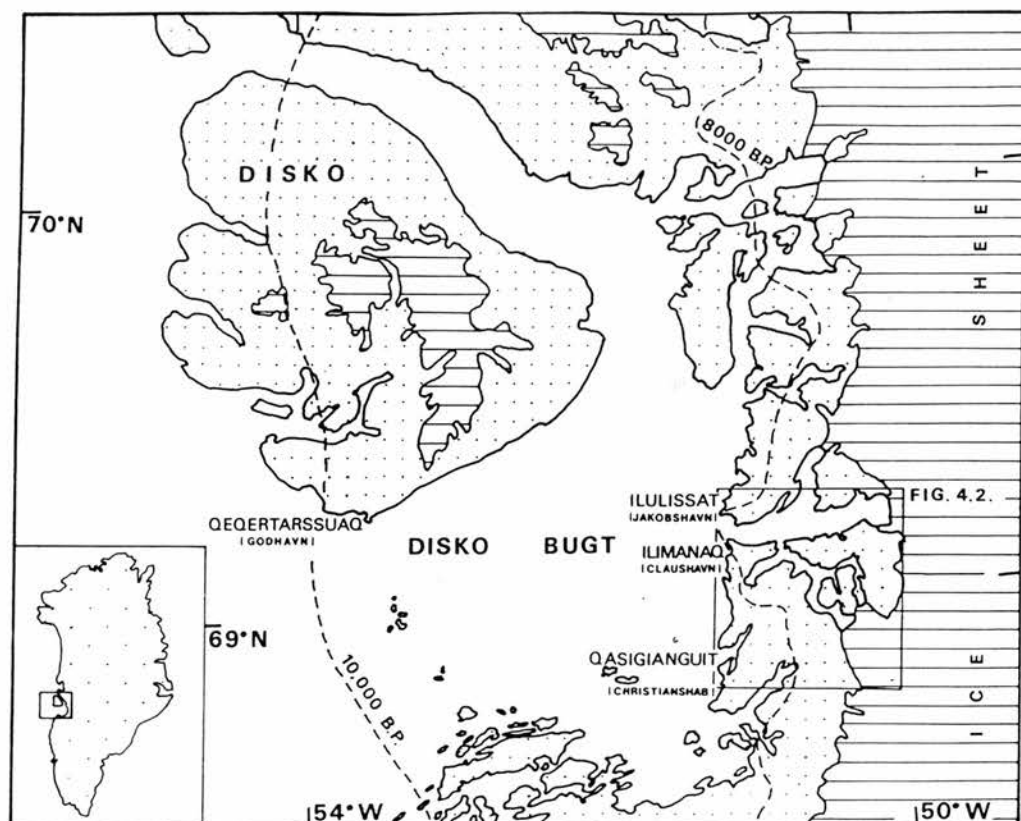
FIGURE 4.1. The study area in its regional context, showing
the supposed ice margin positions at 10,000 BP and
8000 BP suggested by Weidick (1985).
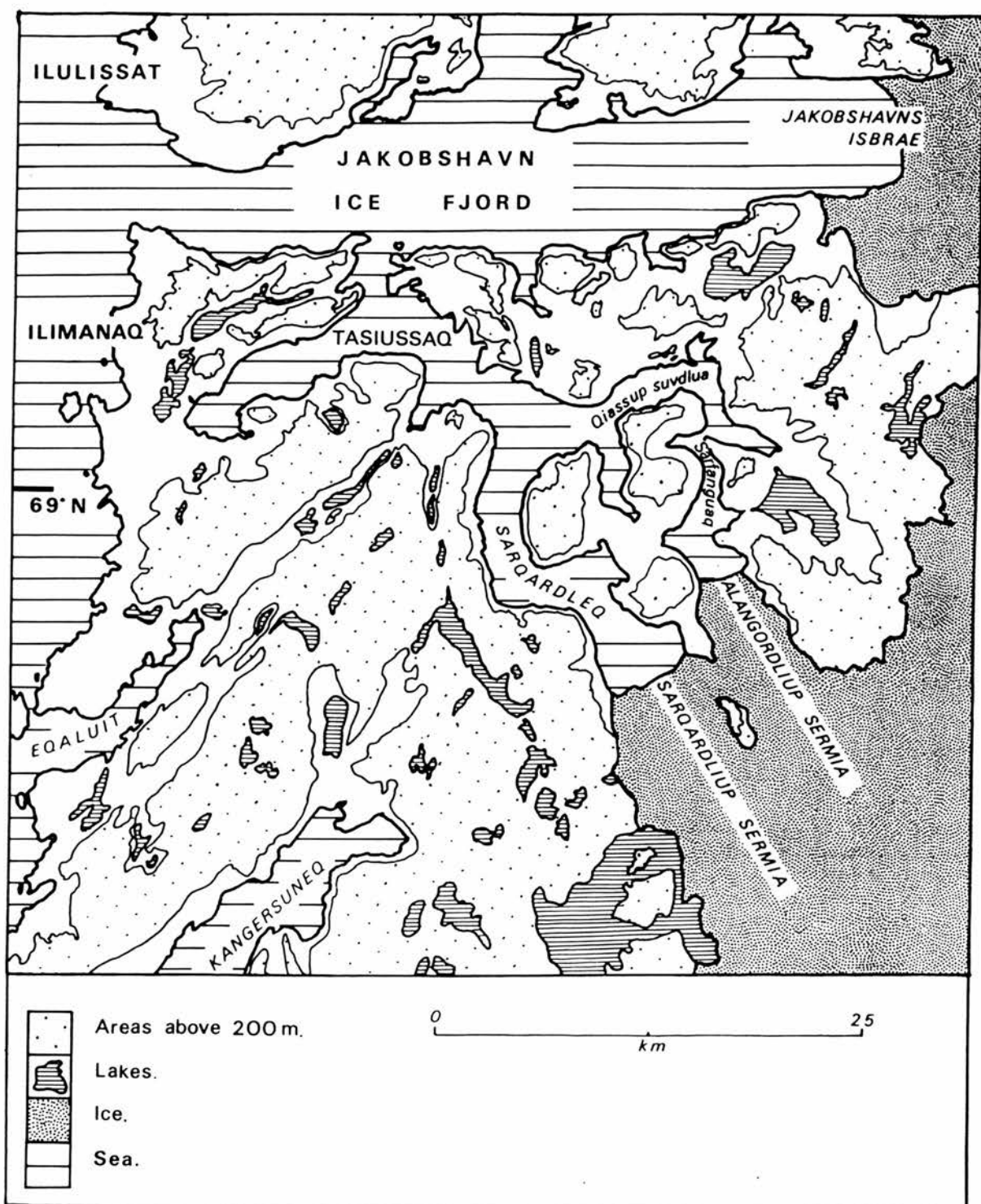
FIGURE 4.2. Jakobshavn Ice Fjord and the Tasiussaq Fjord system.

The Tasiussaq Fjord system (Figs. 4.2, 4.3, 4.4, & 4.5) is connected with Jakobshavn Ice Fjord via a 1.5 km channel through which icebergs are often blown; most of the icebergs in Tasiussaq derive from Jakobshavns Isbrae and not the two glaciers at its south-east extremity. These two glaciers, Sarqardliup Sermia and Alangordliup Sermia, are both calving tidewater outlets but their calving flux is low (Figs. 4.6 & 4.7). Little calving activity was observed during September 1988. Both have experienced net advance since the mid-twentieth century (Table 3.1), and recent moraines at the margin of Sarqardliup Sermia (Fig. 4.8) testify to continuing advance. The fjord is one of the few in Greenland for which any bathymetric data is available, and averages 100 - 200 m in depth (Engell, 1905; Fig. 4.9). Formation and/or preservation of glacial deposits was often impossible along the main glacial troughs, because of steep slopes and cliffs up to 500 m high (Figs. 4.10 & 4.11). It is an area of retreat-stage transition from the maximum, marine-based state of the Inland Ice to its current, predominantly terrestrial condition.
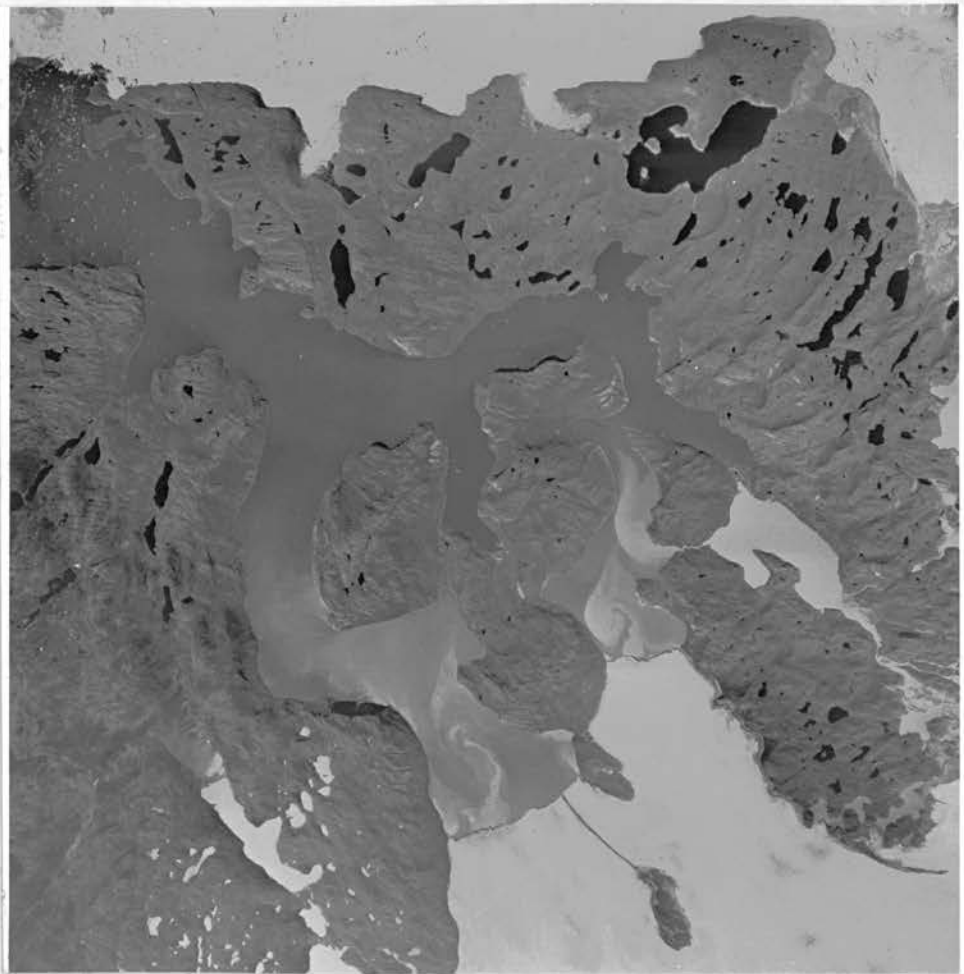
FIGURE 4.3. Aerial photograph (1:150,000) of the eastern part of the Tasiussaq Fjord system, July 1985.
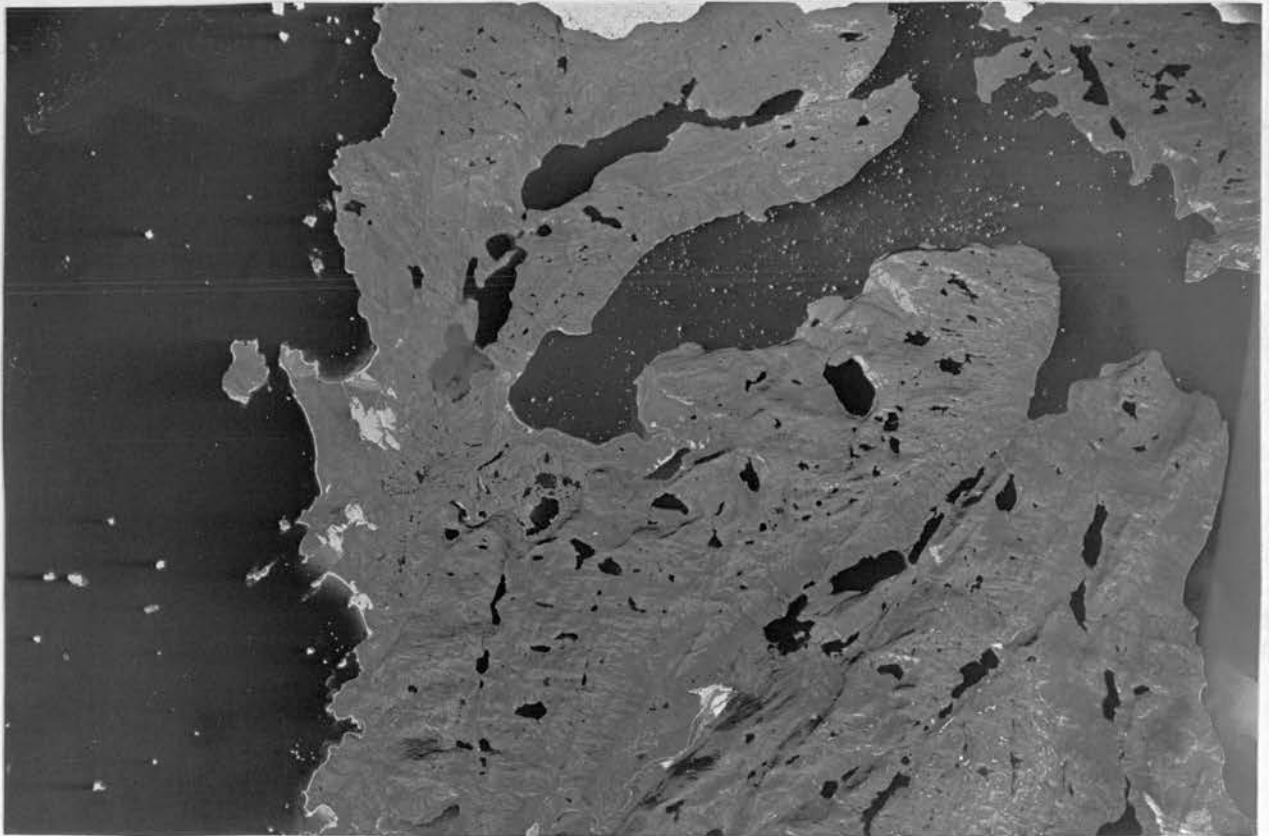
FIGURE 4.4. Aerial photograph (1:150,000) of the western part of the Tasiussaq Fjord system, July 1985.

FIGURE 4.5. View down Tasiussaq from the western end, August 1988.

FIGURE 4.6. The present ice front at Sarqardliup Sermia, September 1988.

FIGURE 4.7. The calving front of Sarqardliup Sermia, with Alangordliup Sermia beyond, September 1988.

FIGURE 4.8. Fresh lateral moraines 3 km behind the terminus of Sarqardliup Sermia.

197
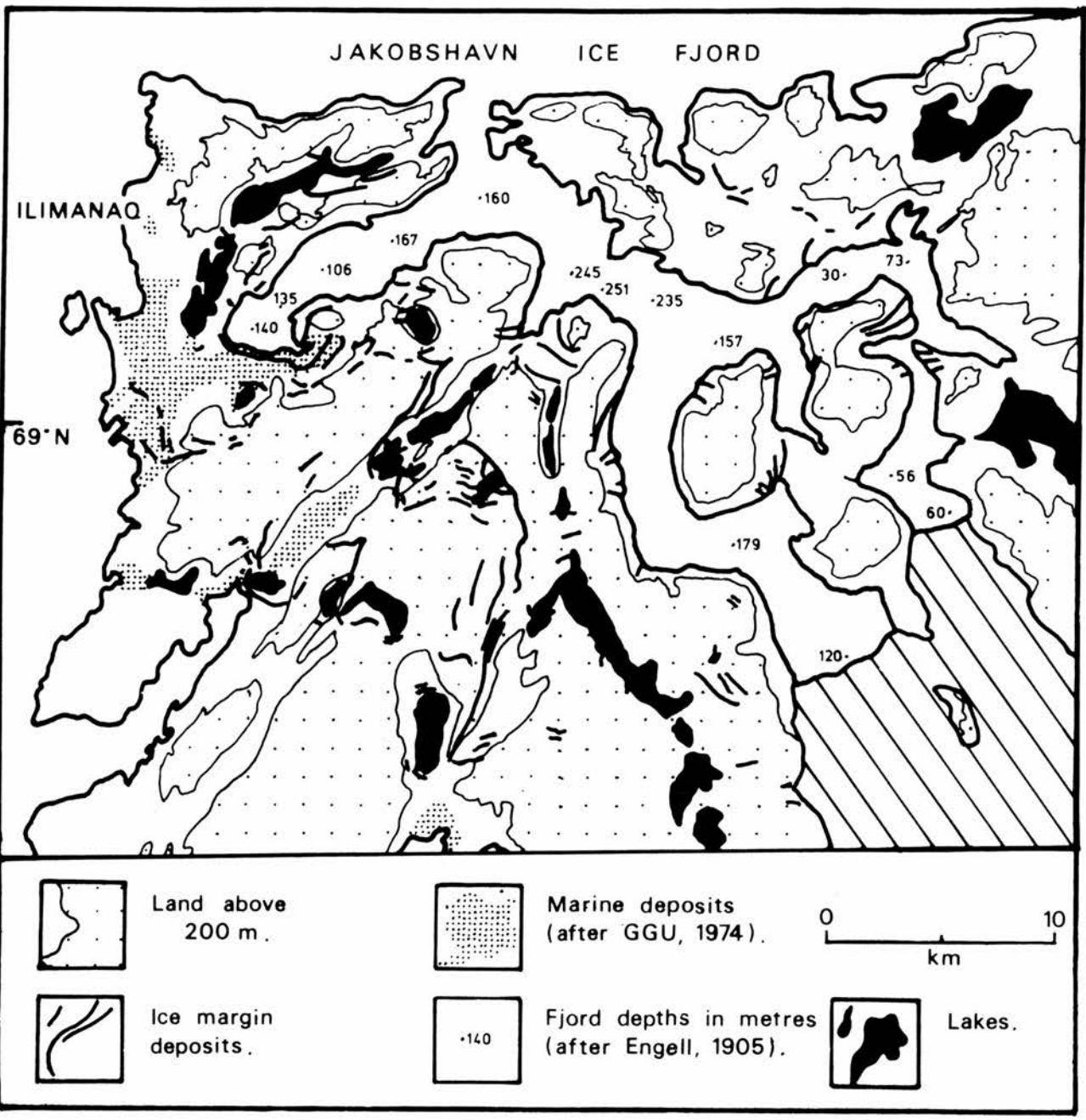
FIGURE 4.9. Tasiussaq Fjord system: Glacial geomorphology and bathymetry.

FIGURE 4.10. View across Tasiussaq from the south side towards the iceberg-choked Jakobshavn Ice Fjord, August 1988.

FIGURE 4.11. View north up the Sarqardleq arm of the fjord. The cliff on the right rises 500 m from sea level.

The Inland Ice retreated from a Sisimiut Glacial maximum position in Disko Bugt at sometime between 14,000 and 10,000 BP, reaching a position close to the present margin around 7000 BP (Weidick, 1985; Kelly, 1980,1985). Possible regional ice-margin positions at 10,000 and 8000 BP from Weidick (1985) are shown in Fig. 4.1. The 10,000 BP position is disputed by Funder (pers. comm., 1988) who believes that rising sea level triggered catastrophic calving retreat through Disko Bugt shortly after 11,000 BP such that the ice margin had reached a location roughly coincident with the present coastline by about 10,000 BP. This was probably followed by a stillstand of considerable duration at this point (Funder, 1985).

Two major phases of stillstand in the Tasiussaq area were originally proposed by Weidick (1968) who interpreted the various moraines as contiguous margin positions between the three valleys of Tasiussaq, Eqaluit and Kangersuneq (Fig. 4.2), the younger dating from 7500 to 8500 BP, the older phase antedating it by an unknown amount. Large depositional features forming the south-western end of Tasiussaq (Weidick, 1968) are the only West Greenland example of a definite Holocene readvance (Kelly, 1985).

4.3. RESULTS.

The main geomorphological elements of the landscape were mapped onto aerial photographs at a scale of 1:40,000, glacial landforms being identified on the basis of morphology and sedimentology, and used to delimit former ice margins (Fig. 4.9). The detailed glacial geomorphology of the central area is shown in Fig. 4.12.

4.3.i. Morphology and sedimentology.

Abundant ground moraine lies scattered on the scoured bedrock in all areas, but especially on the high plateaux further inland, where thick spreads of blocky moraine often bury the bedrock completely. There are many fragments of terminal and lateral moraines, and some of the larger ridges can be traced for several kilometres. Most of the ridges are 1 - 3 m high, of rounded form and vegetated with mosses and lichens, usually consisting of angular to sub-angular material of a wide range of grades. The ice-marginal forms commonly form part of an assemblage of features including kames, kame terraces, and outwash spreads, with the various features grading into one another. However, a few of the moraines are much larger features. Running across the plateau to the east of Qivdlertup Valley towards the lake Akiamiut Taserssua (Figs. 4.12 & 4.13) is a continuous unvegetated ridge 5 - 10 m high and up to 75 m wide (Fig. 4.14) which is clearly visible even on the 1:150,000 aerial photographs (Fig. 4.3). This is informally named 'the Akiamiut Moraine'.
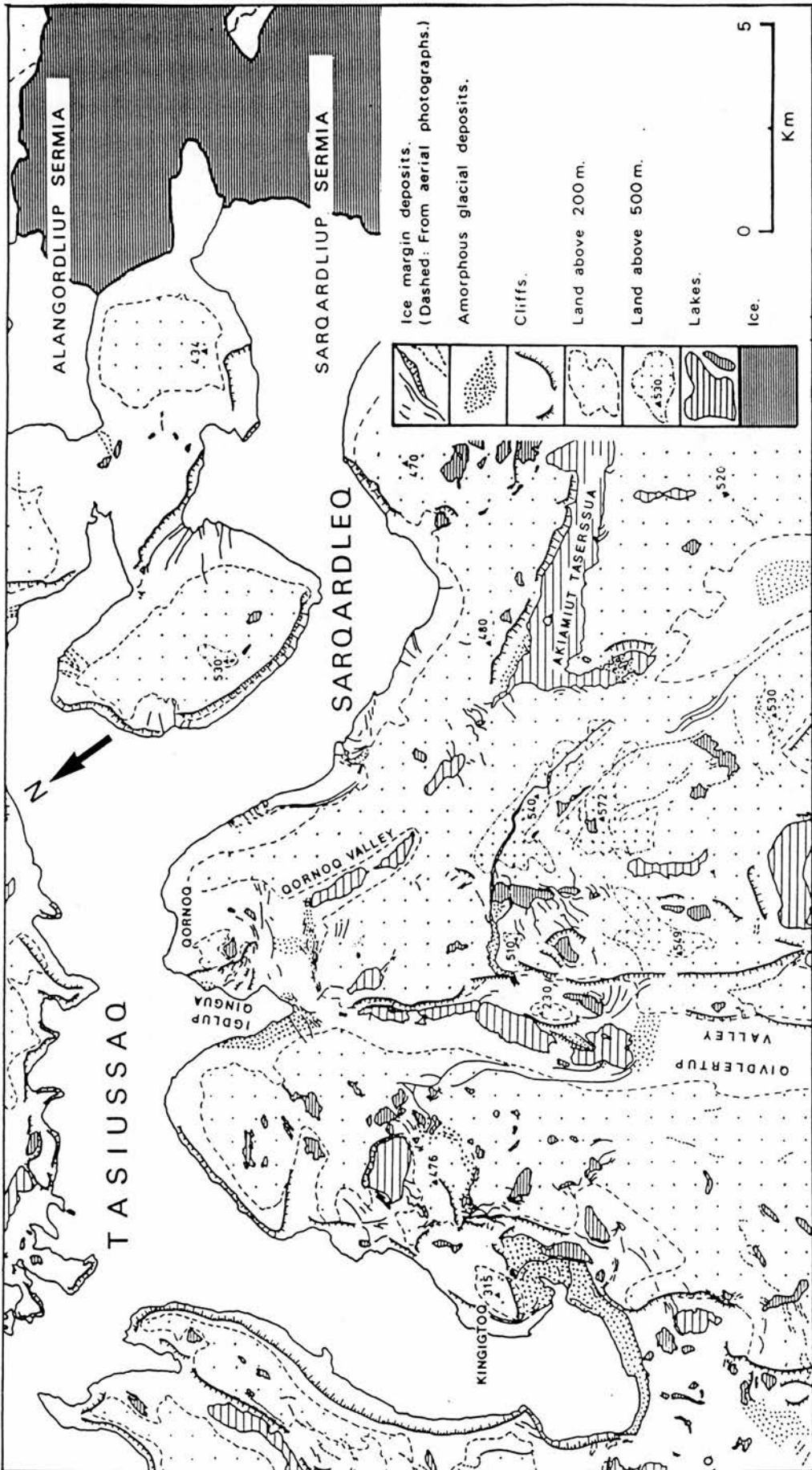
FIGURE 4.12. Detailed glacial geomorphology of the central area.
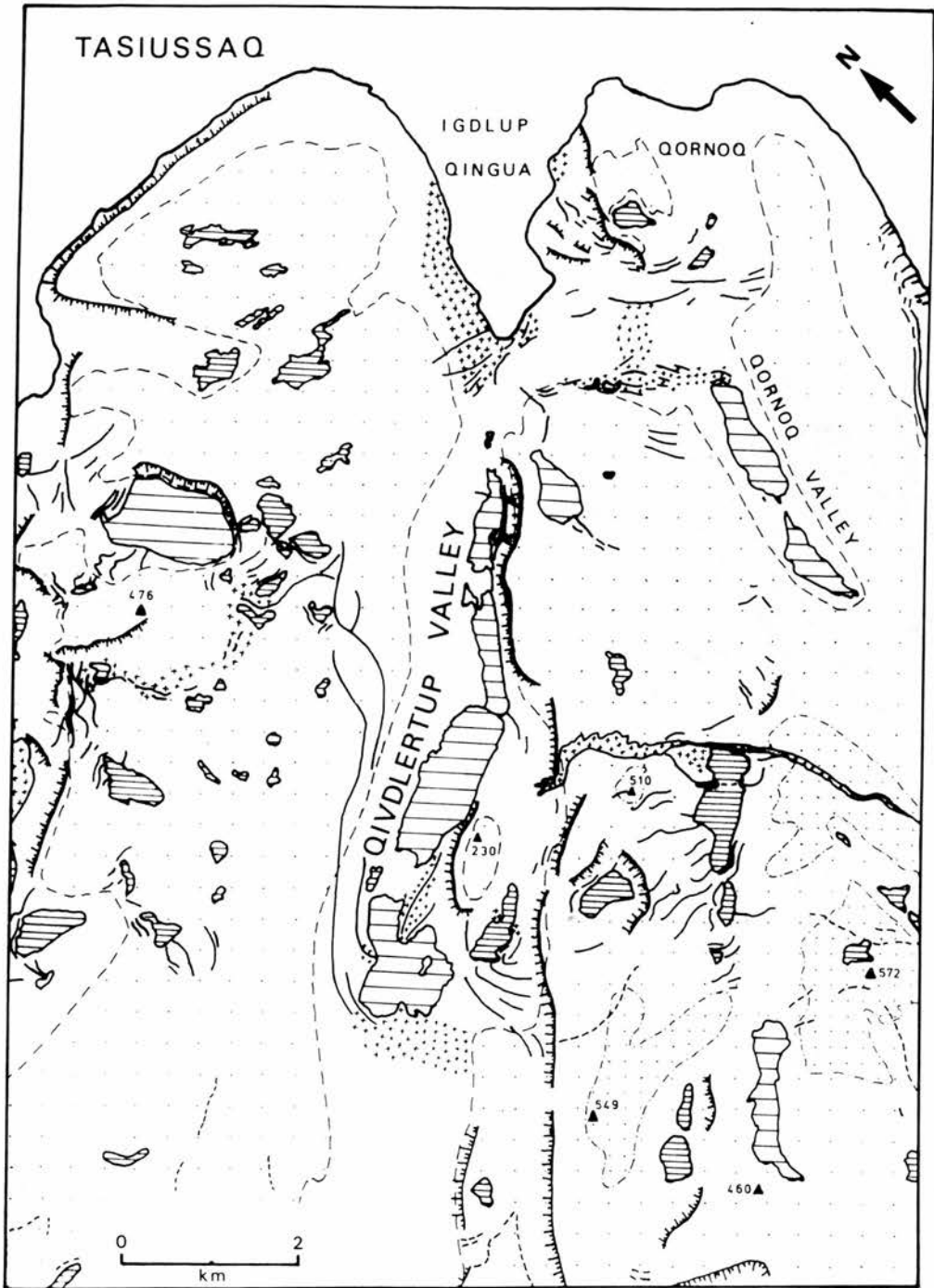
FIGURE 4.13. Glacial geomorphology of Qivdlertup Valley and the area around Igdlup Qingua and the Qornoq peninsula. Key as for Fig. 4.12.

FIGURE 4.14. The Akiamiut Moraine on the plateau east of
Qivdlertup Valley. Note figure for scale.

FIGURE 4.15. The Akiamiut Moraine where it divides and
crosses the lake on the plateau. One arm forms the
dam, and two others can be seen forming a promontory
into the lake centre left.

This moraine lies at the north east end of a steadily rising plateau; north east of it the ground falls off steeply into the continuation of Qornoq Valley. It dams a lake south east of point 510, and, where it crosses the lake, the main ridge divides into a series of sub-ridges, one of which can be clearly seen continuing underwater (Fig. 4.15). At the south side of Qivdlertup Valley it terminates at a steep cliff, down which run three large eskers; the main ridge has no clear continuation across the valley, but appears to divide into a series of end moraines around the hill (point 230) in the middle of the valley. A continuous though smaller lateral moraine runs up the north slope of the valley and it seems probable that this was formed contemporaneously.

The largest ice-marginal deposits anywhere in the area are those which form the south west end of Tasiussaq (Figs. 4.16, 4.17, & 4.18), informally named 'the Tasiussaq Moraine'. This is an assemblage of proglacial and ice-contact features 3.5 km long and up to 1 km wide. Fig. 4.16 schematically illustrates its general form in its western section where it rises from the beach at 26° to a height of 67 m a.s.l.; on its distal side it drops to a height of 50 - 55 m before merging with the marine clay plain which continues to the present outer coast. It is clearly related by its sandur to a sea level at 50 m. The proximal slope consists of a chaotic mass of sands, gravels, silts and clays, the last in nodular concentrations, with no clear stratigraphy; considerable slumping has occurred.
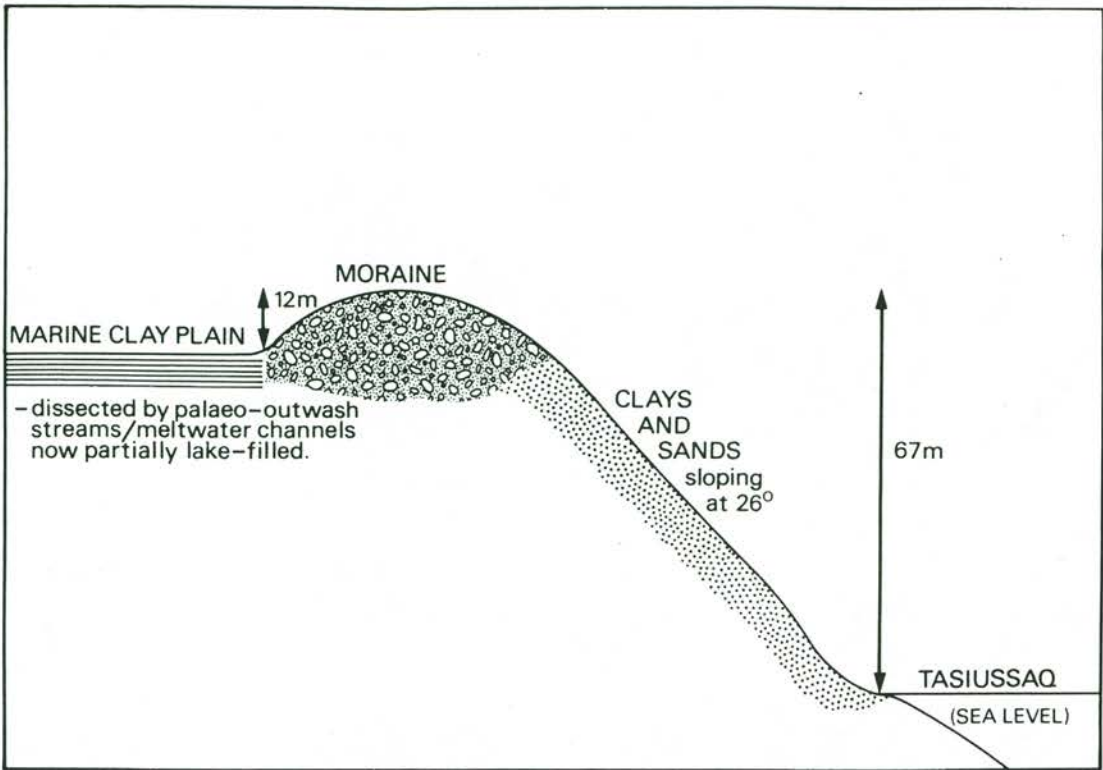
FIGURE 4.16. Schematic   cross-section   of   the   Tasiussaq
    Moraine.

FIGURE 4.17. View east along the Tasiussaq Moraine.

FIGURE 4.18. View west along the Tasiussaq Moraine.

At the top of the slope, the finer material grades rapidly into coarser morainic accumulations, and the deposit is crowned by a clear moraine ridge with angular boulders up to 1.5 m in diameter resting on the surface. Further east the deposit broadens and becomes considerably more complex. The proximal slope is lower and shallower, in places rising only 3 - 5 m above sea level, and numerous moraines of various sizes run parallel and normal to the shoreline. The easternmost ridge is over 200 m wide and of complex form with subsidiary ridges. Talus slopes on the steep cliffs which back the embayment are in several places truncated by the outermost parts of this morainic assemblage.

One of the most complex areas of ice-marginal features is that around the bay Igdlup Qingua and the Qornoq peninsula (Fig. 4.13). The area is thickly mantled with ground moraine and finer glacial deposits which are commonly amorphous. Fragments of moraine ridges, kame terraces, kettle holes and morainic mounds exist in abundance and grade into one another; different parts of the same feature consist of widely differing grades and angularities of material, from sands and fine gravels to large angular blocks. Most of the linear features are aligned longitudinally within the valley, rather than transversely. Subsequent stream action has washed out some areas, and there are also many meltwater channels. Just to the west of the head of Idglup Qingua, two large (2 m high) moraines run almost directly downslope, crosscutting the structural lineation, whereas the moraines the other side of the bay head are mostly parallel with the shore.

4.3.ii. Spatial configuration.

Large accumulations of glacial debris are commonly found at topographic narrowings or bifurcations. Throughout the inner parts of the fjord system, in the Sarqardleq, Qiassup suvdlua, and Sarfanguaq fjord arms, the locations of the moraines along the shorelines all coincide with narrowings, bends, bifurcations, and/or shallowings along the glacial troughs (Fig. 4.9). In the Sarqardleq trough, for example, the main lateral moraines indicate terminus positions at successive narrowing points up the fjord (Fig. 4.19). South of the narrowing point of the bay Igdlup Qingua, a continuous mantle of morainic debris buries the bedrock but is absent where the bay opens out into the main Tasiussaq trough. On the west side of the lake Akiamiut Tasserssua, moraines coincide with prominent bedrock ridges projecting into the lake. Similarly, a moraine 3 m high occurs on a narrow rock ridge which separates the two lakes at the north end of Qivdlertup Valley; in this valley the large glacial deposits are concentrated where the valley bifurcates around an isolated hill 230 m high (Fig. 4.13). Between such points, lateral moraines tend to be smaller, less continuous, and show a diverging spatial pattern; moraine ridges in close association diverge as they are traced down-slope, down-fjord, and down towards present or past water bodies.

The best example of this occurs on the western flanks of Sarqardleq, where the fjord is narrowing southwards. At the entrance to Sarqardleq from Tasiussaq (Point A, Fig. 4.19), several terminal moraines 3 - 4 m high
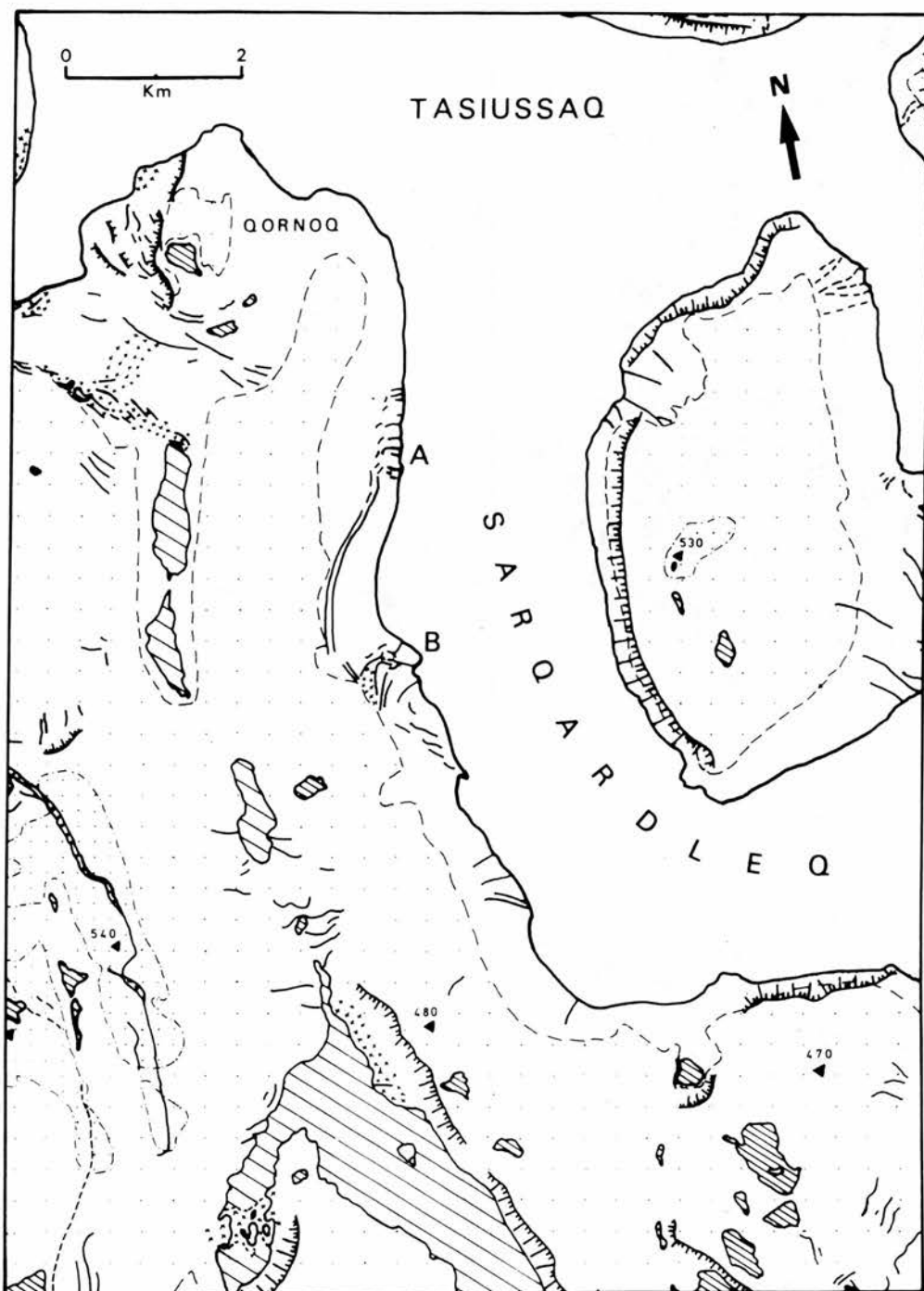
FIGURE 4.19. Glacial geomorphology of the Sarqardleq trough
and surrounding area. Key as for Fig. 4.12.

210

extend short distances onshore and upslope before swinging round and forming gently inclined, less distinct lateral moraines which run back along the sides of the fjord for 1.5 km. Here they meet a rocky spur which projects into the fjord (Point B). The spur is mantled with an abundance of ground moraine, and there are several moraines running steeply downslope to the shoreline. A rock island 30 m offshore (the only one in the fjord) and a bulge in the cliff opposite, indicate that this spur continues across the fjord; unfortunately, no bathymetic data exists at this point to confirm that this is a shallower part of the fjord. Thus the angles of the lateral moraines steepen progressively about the axis of the upper part of the spur. Just 10 m of elevation separates these two sets of lateral moraines where they are in close association on the spur, while the terminus positions to which they relate are 1.5 km apart.

4.4. DISCUSSION.

The spatial pattern of the glacial deposits indicates that the ice margin retreated in episodic fashion, alternating between rapid retreat and relative stability, and that parts of the margin were stable when other parts were unstable. Since variation in ablation is likely to have been the dominant control on ice margin configurations and behaviour during retreat, the episodic retreat was probably caused by variation in the factors controlling ablation rates.

4.4.i. Climatic control.

Weidick (1968,1985), and Ten Brink and Weidick (1974), proposed that the Boreal and early Atlantic moraines of central West Greenland were formed in response to periods of cooler summers during the early Holocene warming. According to them, some of the larger moraines extend over long distances, are of similar morphology, and although the continuity over the high ground is not always clear, the connection between the large deposits in the neighbouring troughs is obvious. On this basis Weidick (1972a,1976,1985) has attempted to correlate the retreat patterns with variations in the oxygen isotope ratios in the Camp Century and Dye 3 ice cores (Fig. 4.20).

However, there are problems in attempting to correlate detailed ice marginal behaviour with the temperature signal inferred from ice cores, because oxygen isotope records are integrations of environmental change over continental-scale areas (Koerner and Fisher, 1985). Detailed investigation of ice cores from Devon Island by Koerner and

212
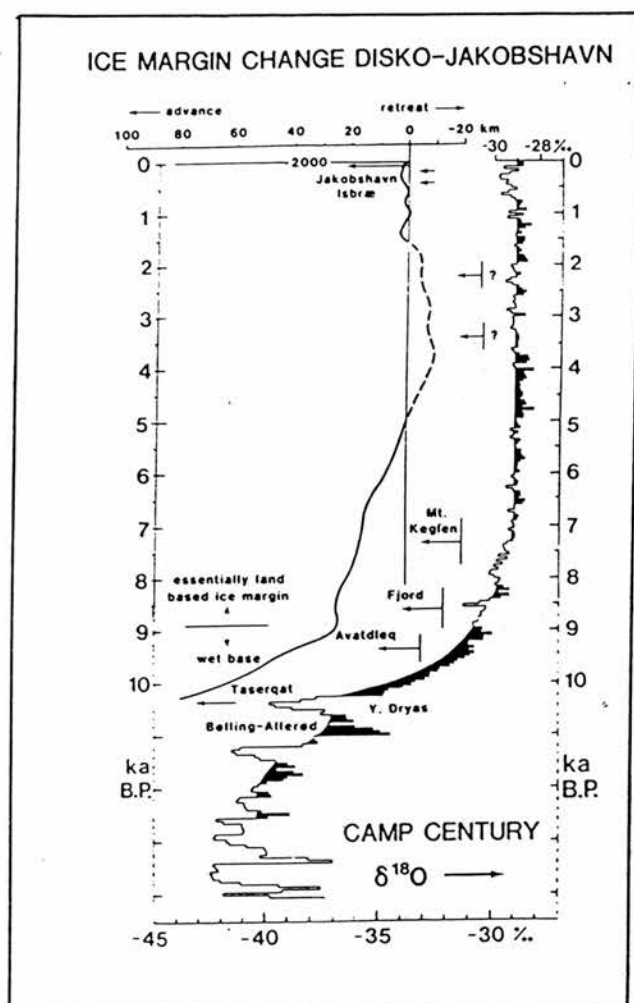
FIGURE 4.20. The recession of the ice margin between Disko and Jakobshavn, compared to the Camp Century climatic record (Dansgaard and others, 1984). Halts and readvances of the ice margin are marked by arrows. Note halt at 9000 BP, the time of transition from wet-based to dry-based retreat. After Weidick (1985).

Fisher (1985) showed that although there was broad agreement between the climatic record in the ice cores and that of the glacial geology, the ice cores failed to record the oscillations of the last few thousand years. They concluded that 'there is simply not a direct linear relationship between ^{18}O ... and mass balance' (p.315). Furthermore, Fairbanks (1989) and Koerner (1988) have shown that the isotopic temperature record in Northern Hemisphere ice cores

is distorted by meltwater feedback effects. The enormous influxes of glacial meltwater to the oceans around the ice sheets would have altered the isotopic composition of the evaporation source areas for precipitation, so that the snow falling on the ice sheets during the Late Glacial and early Holocene would have been anomalously negative isotopically. Thus the 'temperature signal' inferred from ice cores may in fact only partly reflect temperature changes, the rest of the isotopic fluctuation being due to source water variability.

If the moraines are the result of changes in regional mass balance controlled predominantly by climate, then the interpretation of Weidick and Ten Brink must be broadly correct. However, palaeontological and palynological evidence (Donner and Jungner, 1975; Donner, 1978; Fredskild, 1983,1985) suggests that the early Holocene (9000 - 8000 BP) in West Greenland was a period of uninterrupted warming and drying, reaching a peak warmer and drier than today sometime between 7500 and 6500 BP. The subarctic West Greenland Current was established from 9100 BP, and there was intrusion of water warmer than at present from 7300 BP until after 5000 BP (Kelly, 1985).

The contemporary ice sheet is in quasi-equilibrium with today's climate (Reeh and Gundestrup, 1985; Reeh, 1989). For the rapidly wasting early Holocene ice sheet to halt, or readvance slightly, in positions far in advance of today's margin, marked falls in summer temperatures for substantial periods of time would be necessary in order to counterbalance the rapid re-equilibration to the warming climate that was in progress. However, neither the fauna of the time (Fredskild,

214

1985) nor the temperature records inferred from the ice cores (Hammer and others, 1986; Dansgaard and others, 1982) indicate anything more than minor cooling episodes superimposed on the warming trend during this period. In historical times, even slight temperature decreases have led to minor advances of the Inland Ice because of the contrasts between the dynamics of the ablation zone and the rest of the ice sheet (Ten Brink and Weidick, 1974), but this effect is only known to have occurred recently, involving an ice sheet in quasi-equilibrium; it is not known whether such responses would occur in the context of rapid ice sheet retreat.

Therefore, while Fig. 4.20 demonstrates that the first order forcing on ice sheet retreat was indeed climatic, it is questionable whether the detail of the retreat stages of the ice sheet margin were controlled dominantly by climate. Indeed, over 20 years ago Weidick (1968) specifically noted the potential effectiveness of Mercer's (1961) principle and topographic control of the West Greenland ice margin. In a more recent paper, he expressed the opinion that the behaviour of the ice sheet margin during recession was:

> ... often locally determined by topography rather than by changes in mass balance. (Weidick, 1985, p.306).

4.4.ii. Topographic control.

The glaciers appear to have halted only at places where the topographic configuration gave them an increased degree of stability. Retreat beyond such points into wider and/or deeper parts of the fjord would have led to increased rates of iceberg calving, and continuous, rapid retreat until the next topographic 'pinning point' was reached.

The outermost moraine systems may be an example of the hypothetical sequence of evolving marginal dynamics envisaged by Hughes (1987; Fig. 2.12). The warming climatic conditions and rising sea levels of the early Holocene are thought to have caused catastrophic calving and dramatic retreat of this marine-based sector (Funder, pers. comm., 1988). The transition from a marine-based to a predominantly land-based ice sheet would have led to greatly reduced ablation rates due to the fall in iceberg production. This mechanistic cause of stillstand is clearly implied by Weidick (1985) in his attempt at correlating the retreat pattern with fluctuations in the oxygen isotope ratios in ice cores (Fig. 4.20). Furthermore, isostatic rebound was rapid at this time (Donner and Jungner, 1975) so that the length of the transition period would have been shortened and the dynamic impact consequently heightened. Just as glacio-isostatic sinking compels the grounding line to retreat (Hughes, 1988), so glacio-isostatic emergence enables it to advance. Under such circumstances a halt or a readvance would be expected.

Two distinct states of this transition can be recognised. The first was the time when the ice sheet margin initially reached a grounding position at or near the contemporary coastline, and the second when the rapidly calving outlet glaciers retreated either to stable, grounded calving locations or ceased calving altogether. The outermost moraines south-west of Tasiussaq show that the margin closely paralleled the coast, strongly suggesting that here, as in many parts of West Greenland:

> ... the sea/land transition formed a major glacio-dynamic obstacle in the deglaciation process. (Funder 1985, p.140).

The readvance of the Tasiussaq ice lobe relates to the second of these transition phases. Initially, rapid calving probably led to retreat to the narrows of the fjord at Kingigtoq hill, a very sudden topographic constriction relative to the unconfined plain to the west. Dynamic adjustment to the much reduced ablation area and continuing isostatic rebound then resulted in a readvance, followed by adoption of a stable, grounded position. The complex glaciomarine sedimentology of these large deposits suggests that this readvance occurred on a morainic shoal as observed in many fjords in Alaska (Meier and Post, 1987). The marine clay plain to the west represents a contemporary sea level at 50 - 55 m a.s.l., suggesting a date for this readvance of 7250 - 7500 BP from the emergence curve of Donner and Jungner (1975).

It is therefore proposed that the dominant outer moraines of the area relate to this period of adjusting ice dynamics following retreat out of the sea. The topography of the coastal zone and the relative altitudes of these parts of the glacial troughs (45 - 55 m a.s.l.) is such that this transition happened at almost the same time in each of the major valleys, as shown by Weidick's (1968) ^{14}C dates. Thus these ice margin deposits may be divided broadly into an outer and an inner phase, as originally proposed by Weidick (1968), but the primary causal link is a dynamic response to topography rather than climate.

4.4.iii. Reconstructing the retreat stages.

Fig. 4.21 presents a hypothetical reconstruction of the ice sheet retreat in the Tasiussaq Fjord system ca.9000 - 7000 BP.

4.4.iii. a) Dating control, and rates of retreat.

Almost no absolute or relative dating control exists for this scenario. The dating bracket of 7500 - 8500 BP presented by Weidick (1968) for his inner moraine phase (equivalent to 'the Tasiussaq Moraine' and the moraines in the central section of Qivdlertup Valley) are the only dates available; neither he nor subsequent workers have been able to find any other dateable material. On the evidence available, retreat rates during this period averaged just 20 m a^{-1}; if Funder (pers. comm. 1988) is right in thinking that the ice margin reached the outer coast by ca.10,000 BP, and Weidick (1985) is right that it had reached a position close to the present margin by 7000 BP, then the rate of retreat over the 40 km between the coast and today's ice sheet drops to only 13 m a^{-1}. Since calving margins have frequently been observed to retreat at rates of many hundreds of metres each year, such rates seem much too low, especially since the ice sheet was not in equilibrium with the warming climate. The steepness of some of the lateral moraines along the fjords strongly suggests that rapid retreats did occur, followed by periods of stillstand and readjustment of the glacial long-profiles to the new frontal position.
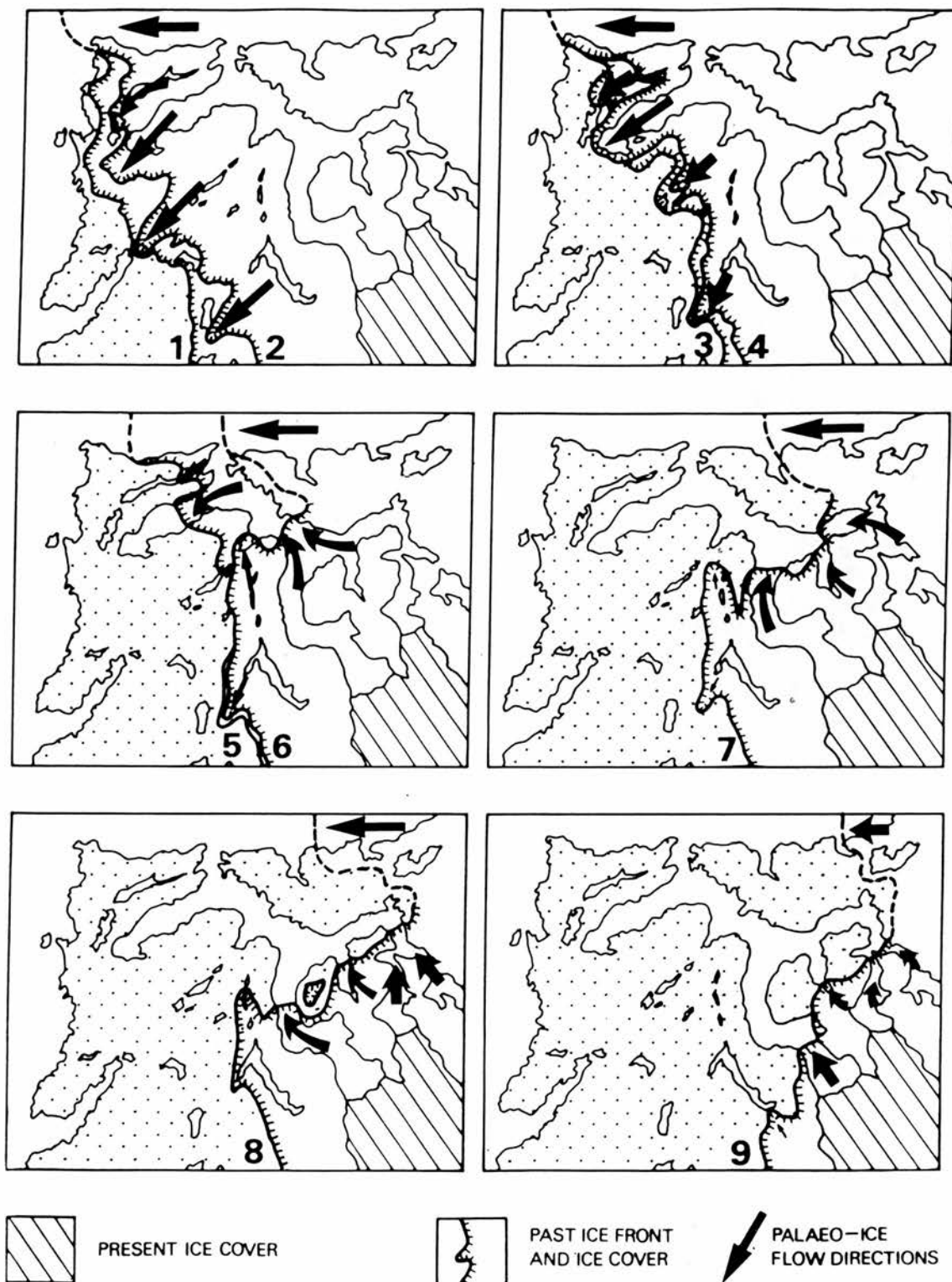
FIGURE 4.21. Reconstruction of ice-margin configurations during retreat, ca.9000 - 7000 BP. Only the present coastline is shown. The dashed lines show possible contemporary positions of Jakobshavns Isbrae. The reconstruction is based on the geomorphological evidence and the theoretical ideas of Mercer (1961) and Hughes (1987,1988).

Thus, in order to resolve the apparent discrepancy between the slow net retreat rates on the one hand, and the observational and geomorphic evidence on the other, one has to presume:

i) that the stillstands between these rapid retreat episodes were of the order of several centuries, or,

ii) that the timescale is too long.

The first option is possible in the outer, coastal areas during the transitional phases from a floating to a terrestrial margin; the Tasiussaq Moraine complex in particular is indicative of a long term halt. In the inner parts of the fjord system, however, the geomorphic evidence nowhere indicates extended halts, and they would be unlikely anyway within the overall regional context of a rapidly wasting ice sheet.

The second option is also a possibility. The date of 7000 BP for attainment of the present margin position is based simply on the age of the marine limit at the present ice margin and the age of reworked shell fragments within the present glacial sediments (Kelly, 1980) so it is wholly possible that an earlier date could apply. It is therefore probable that, within a reduced timeframe (perhaps 2500 years), the retreat consisted of rapid wastage interspersed with one or two long stillstands in the early Holocene and brief, topographically conditioned halts thereafter.

4.4.iii. b) Topography and retreat stage configurations.

Despite the lack of dating control, the glacial geomorphological evidence and the resulting reconstruction of retreat configurations shown in Fig. 4.21 clearly indicates the following:

i) The calving margins retreated faster than the adjacent terrestrial portions of the ice margin.

ii) The deeply excavated glacial troughs determined the changing ice flow directions during retreat.

iii) As a result of i) and ii), great changes of flow direction took place over short distances as the ice front retreated up the troughs. In the earlier parts of the retreat (Stages 1 - 4 in Fig. 4.21), the dominant flow direction was south-west. At Stage 4, the lobe in Tasiussaq is shown stable at the Tasiussaq Moraine, and the ice margin across the plateau south of Qivdlertup Valley is at the Akiamiut Moraine which is a demonstrably stable location at the highest side of a steadily rising plateau. Following this time, however, the Tasiussaq glacier retreats back into deep water and begins to retreat rapidly up the fjord. The ice margin on the plateau, however, continues to retreat only slowly, or remains stable, forming the numerous moraines, the abundant ground moraine, and the Akiamiut Moraine itself. Between Stage 4 and Stage 6, ice flow directions at Igdlup Qingua wheel round through almost 130^{0}; having been broadly south-west, they become north or even north-north-east. It is consequently small wonder that the resulting geomorphological record there is so complex.

Following Stage 6, the calving glaciers in the fjords continue to retreat rapidly. The lobes in the valleys of Qornoq and Akiamiut, by contrast, are progressively deprived of ice supply by the thinning of the ice over the topographic divides separating them from Sarqardleq, and become slowly wasting lobes. The thick spreads of moraine, evidence of abundant meltwater, and the relative rarity of distinct ice marginal forms in this inner area testify to this.

Thus the relative rates of retreat were determined by the contrasting terminal environments which conditioned which ablation mechanism (calving or melting) was dominant. The configuration of the ice margin during retreat, together with the specific location of the stillstands, were predominantly controlled by the topography.

4.5. CONCLUSIONS.

The nature of the episodic retreat here is governed by the influence of topographic configurations on marginal ice dynamics, particularly iceberg calving dynamics. The evidence indicates that stepwise retreat as envisaged by Hughes (1987,1988) occurred. The contemporary warm climate was the dominant forcing function causing ice sheet retreat, and led to destabilisation from pinning points, but the geomorphological evidence shows that the stillstands were topographically induced. An implication is that palaeoclimatic significance should not automatically be attached to the Late Glacial and Holocene stillstands of tidewater outlet glaciers of ice sheets, especially stillstands resulting from the sea/land transition.

C H A P T E R   F I V E .

CONCLUSIONS.

5.1. THE RESPONSE OF CALVING GLACIERS TO CLIMATIC CHANGE.

Glaciers and ice sheets are climatic phenomena. The geographical distribution of ice masses on earth can be explained in terms of present or past climatic regimes, and their macro-fluctuations in space and time are driven by climate system variability which is in turn driven by variations in the amounts of solar radiation received at the earth's surface. Over long timescales (longer than 10^3 years), therefore, the growth and decay of ice masses is primarily a function of climate change, and their oscillations closely follow changes in temperature and precipitation, albeit with variable time lags determined by glacier dynamics. It is especially clear that ice sheet collapse is initiated astronomically:

> Insolation forcing is the main impetus for deglaciation. (Ruddiman, 1987b,p.474).

The record of global ice volume during the Quaternary is consequently an accurate reflection of the gross patterns of climate change.

However, even at these long timescales, neither the patterns of build-up and decay nor the overall geographical extent of ice masses are free from the complicating influences of geography, environment, and their effect on ice sheet dynamics. Macro-geomorphological features such as mountain ranges, uplifted shields and, in particular, the edges of continental shelves, fundamentally affect ice sheet geometry. The contrasting rates of operation of accumulation

225

and ablation result in pronounced asymmetry of growth and decay which is not a direct result of climatic forcing. Thus only at timescales of 10^5 and longer can it be said that ice sheets exhibit a direct, linear response to climate. At all shorter timescales, and with progressively greater effect, the interplay between ice dynamics and enviromental factors results in a progressive decoupling of ice margin response from the climatic signal. The more precise the chronology, therefore, the less likely it is that the record of glacier fluctuation can be used as a detailed and reliable proxy for climate change.

Arguably the most powerful decoupling mechanism is iceberg calving. Iceberg calving introduces mechanical instability to the glacier system. For this reason, site factors which are conducive to stability or instability will have a proportionally greater effect on calving glaciers than on land-terminating glaciers. This, in turn, largely explains why topography exerts such a dominant influence on the oscillations of calving glaciers.

The net result of these non-climatic influences is that calving glaciers exhibit a unique relationship with climate. In some regards, they are much more sensitive to climatic change than land-terminating glaciers, whereas in others they are almost completely insensitive. For instance, they can be much more sensitive to climatic warming because, unlike land-terminating glaciers, they cannot stagnate in situ. This is especially true if the glacier is at an advanced position in its fjord such that a small rise of the ELA will greatly increase the ablation area.

On the other hand, calving glaciers can be insensitive to climatic change during the cyclical advances and retreats that occur in some fjord systems. During such periods, the spatial behaviour of the terminus is largely a function of the interaction between glacio-dynamics and the topographic geometry of the fjord. The strength of these effects is such that even fairly large climatic shifts will hardly affect the dominant trend. Climate may well be the forcing function which initiates an advance or retreat, but, once initiated, the subsequent behaviour may bear little relation to subsequent climatic change. Environmental factors introduce thresholds of stability to the system at all temporal and spatial scales and therefore ice masses can exhibit a stepped, non-linear response to smooth climatic forcing. The dynamics of the iceberg calving process can nullify, damp, or greatly amplify the glacial response to a particular climatic shift. The complexity that can be found in places with a range of terminal environments and fjord geometries has been illustrated from the Narsarsuaq area of South Greenland (3.4).

Climate will always be the first order control on the growth and decay of ice bodies large and small, but climatic inputs pass through a series of morphological and glaciological filters before they find expression at ice margins:

Filter 1. The terminal environment. This determines the dominant ablation mechanism - iceberg calving or melting - and hence a large part of the potential for dynamic decoupling from climate, because land-terminating glaciers

are dominantly controlled by fluctuations of summer temperature, whereas the response of calving glaciers is highly variable. On longer timescales (10^3 years or more) sea level change is important in this context, while, in the short term, rapid changes of lake levels due to jökulhlaups can radically transform the dynamic regime almost instantaneously.

Filter 2. Topography. This introduces thresholds of stability which determine the location of stillstands, the extent of advances, and the detail of ice front oscillations within a fjord system. At larger scales, topography is important in conditioning differing ablation mechanisms during the retreat of ice sheet margins from marine to terrestrial situations (Hughes, 1988; Fig. 2.12).

Filter 3. Sediment supply. By affecting the effective water depth at the calving front, this has some influence on rates of retreat and a controlling role on rates of advance in situations where advance takes place on a moraine bank.

Filter 4. The contrast between tidewater and freshwater calving environments. This affects calving dynamics, calving rates, and hence calving fluxes, and therefore may affect the timing and rate of frontal movement of calving glaciers.

Together, these filters render the response of ice sheet outlets to climate progressively non-linear and variably lagged. The importance of the terminal environment has been brought out in Chapter 3 by the strong correspondence of glacier behaviour with variable terminal environments in West Greenland. Topography seems to have operated with significant effect both in West and East Greenland even

228

during the strong climatic warming of the early Holocene (Chapter 4) and is controlling the frontal behaviour of ice fronts in West Greenland today (Chapter 3). An investigation of the role of sediment supply for ice sheet outlets has been beyond the scope of this research, but has been demonstrated for tidewater termini of mountain glaciers in Alaska (Meier and others, 1980) and South Georgia (Clapperton and others, 1989). The basis for believing that Filter 4 is important in determining the nature and rate of glacier response is largely conjectural at present; it is suggested by the patterns of twentieth century change of West Greenland glaciers described in Chapter 3. However, the empirical results of Funk and Röthlisberger (1989) suggest that further testing of this idea would lend it more support.

Some of these effects are illustrated schematically in Fig. 5.1 which shows how a single climatic input can give rise to a range of glacial responses in contrasting terminal environments. It illustrates how topography introduces stability thresholds to the glacier system, resulting in sensitive bifurcation points. The model charts the oscillation history of five ice sheet outlet glaciers over a notional 2000 year period during which mean annual temperatures fluctuate over a range of 8^0C. The land-terminating glacier (A.2) responds directly to the climatic forcing, with a short time lag. The result is a record of frontal change that approximates to a smoothed mirror image of the proxy climate record.

By contrast, calving glaciers in a fjord system with complex topography (A.3) exhibit a response to climate
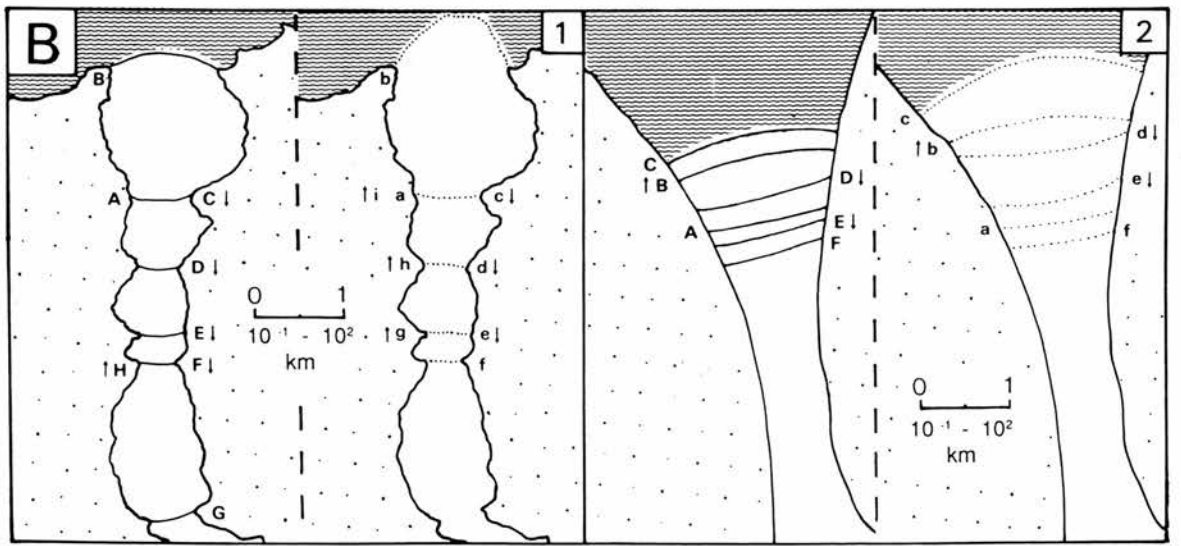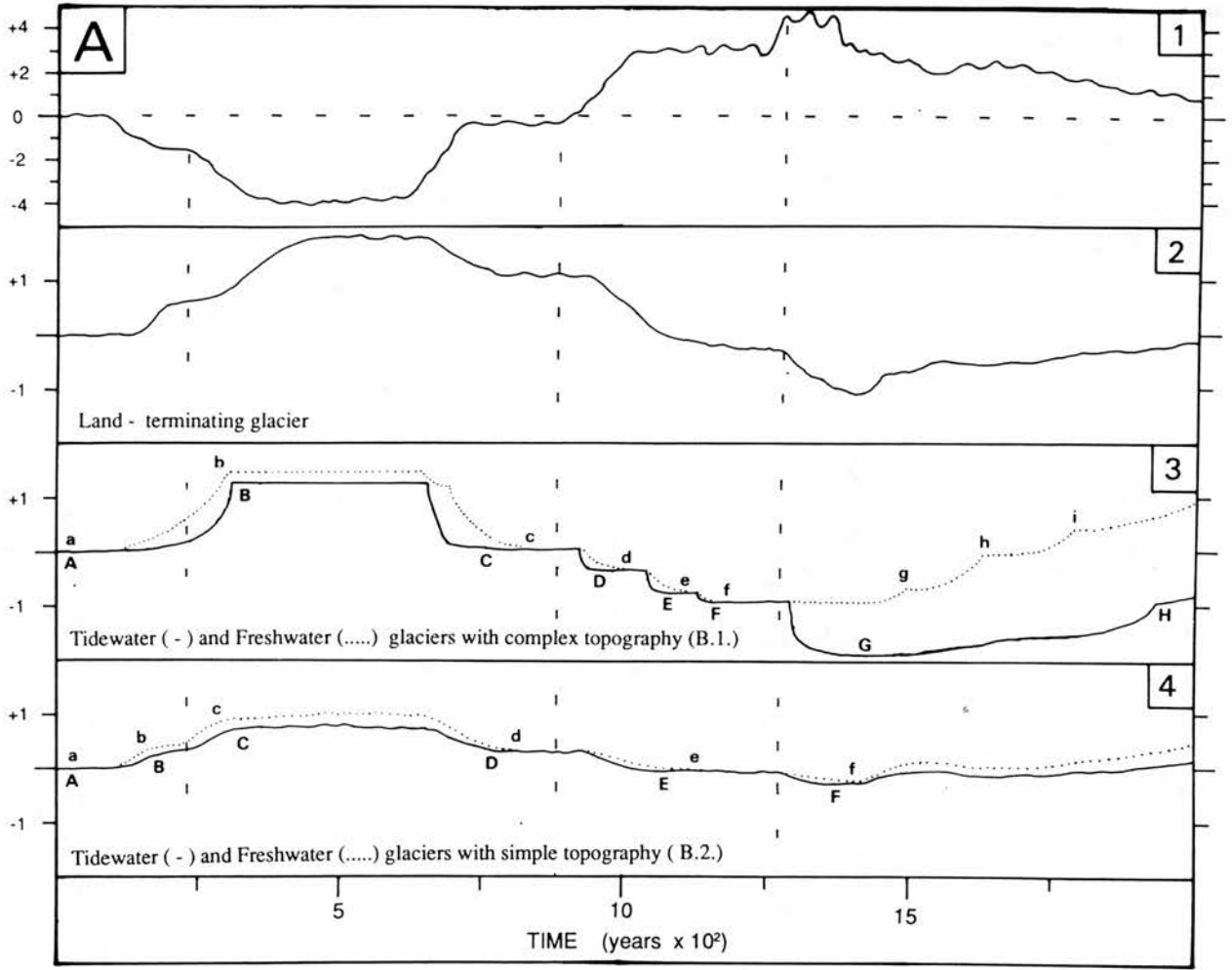
229

FIGURE 5.1. A conceptual model illustrating topographic
control of ice sheet outlet fluctuations in an
idealized thermal environment.
For full caption, see next page.

FIGURE 5.1.   A conceptual model illustrating topographic
     control of ice sheet outlet fluctuations in an
     idealized thermal environment.   Mean annual
     temperatures (A.1) are shown as being the primary
     climatic signal; precipitation variations are not
     explicitly considered.   In the idealized fjord
     illustrated in B.1, width and depth are assumed to be
     directly proportional.   The only variable in A.3 and
     A.4 is the contrast between tidewater and freshwater
     calving environments.   Note the variable length scales.
          A.2:   Land-terminating glaciers respond
     directly to the climatic forcing, with a slight time
     lag.
          A.3 & B.1: Calving glaciers in complex fjord
     systems oscillate in a stepped manner, the stable
     locations being determined by topography.   Topography
     produces the bifurcation between the tidewater and
     freshwater glacier records after 1300 years.
          A.4 & B.2: Calving glaciers in a 'bell mouth'
     fjord configuration respond almost directly to climate
     with only minor topographic influence; the oscillations
     are of smaller magnitude than those in complex fjords.

          For full description, see text.

forcing which is stepped and non-linear.   The overall trend

of advance or retreat is determined by climate, but the

timings and locations of stillstands are determined by

topographic pinning points within the fjord system.   After

point A, the cooling climate triggers the advance, but

despite the strong cooling, advance is initially slow because

of the strong fjord widening (B.1) and the resulting

progressive rise in calving losses.

However, once the ice front passes the widest point, the

rate of the advance increases exponentially because of the

falling calving rate.   As a result, the glacier does not

react to the step in the temperature record at - 1.75°C.

Moreover, once it reaches the fjord mouth it faces a

prohibitively steep ablation gradient and so no further

advance occurs, despite a further fall in temperature of

2.25°C; it is probable that calving fluxes would be high during this period, but frontal change minimal. Thus, during the initial two-step deterioration in climate, the calving glaciers can be seen, at different times, to be advancing during a period of stable climate, and stable during a period of cooling climate.

The commencement of warming at about 600 years leads to the destabilisation of the calving front from its fjord-mouth position, and subsequent catastrophic retreat back to point C; it remains here for a period of about 200 years due to a combination of climatic and mechanical stability. Continued warming thereafter leads to a stepped retreat back through points D, E, F, and back to point G, all of which consist of initial rapid retreat through wide parts of the fjord, and then slowing retreat into the narrows, producing the convex shape of the retreat steps.

The freshwater calving front in A.3 follows broadly the same pattern as the tidewater front, although advances tend to commence earlier and be more rapid, while retreats are slower. At point b, the freshwater tongue is shown projecting further into the deepening, widening open-water situation than is the tidewater front. This is primarily because the calving losses in the given circumstances would be lower. The most important divergence between the two records comes after 1300 years. A peak of warmth, reaching + 4.5°C, proves sufficient to destabilise the tidewater front from point F, causing rapid retreat back to point G, whereas the freshwater front is able to maintain itself at point f. Thereafter, the cooling climate is operating on glaciers

232

which are in different topographic situations; they therefore exhibit contrasting patterns of subsequent frontal change. The tidewater front advances slowly through the wide section of fjord between points G and H and is a little in advance of point H at year 2000 (frontal position not shown), whereas the freshwater front advances in stepped fashion through the series of pinning points in mid-fjord and is in advance of point i as the record comes to an end.

In the identical circumstances used in this model, the bifurcation is caused by the contrast between the dynamics of tidewater and freshwater calving fronts, but in the real world such divergence could result from a range of factors including slightly different fjord configurations, drainage basin topographies, ice supply rates and sediment supply rates. Whatever the cause, the potential for bifurcation introduced by the dynamics of the calving process is significant because a uniform climatic record can produce widely divergent patterns of glacier behaviour. Bifurcation can occur either spatially or temporally. Neighbouring glaciers may behave differently in response to a particular climate change, and one glacier system may respond in contrasting ways to similar climatic shifts at different times.

In the case of calving glaciers oscillating in simple topographic circumstances (A.4 & B.2), there are no stability thresholds that work to decouple frontal change from climate. As a result, the response of the tidewater and freshwater glaciers to the changing climate is qualitatively similar to that of the land-terminating outlet, with few non-

linearities. However, the 'bell mouth' topography does act to damp advances and amplify retreats through its effect on calving rates and fluxes.

As stressed by Mann (1986), questions concerning the relative roles of climatic, glaciological, topographic, and site factors in determining ice front behaviour can only be satisfactorily answered via case studies of individual glaciers because the range of possible influences and assemblages of controlling factors is so large. There are several sources of 'non-climatic noise' in the fluctuations of calving glaciers, and the relative importance of these factors varies greatly for each glacier and with the temporal resolution of the particular study.

Nevertheless, the conclusions of this research have broad implications. The geomorphological record of past ice sheet oscillations can provide reliable palaeoclimatic information, but only where the 'calving noise' and other forms of 'environmental interference' (Chapter 1) can be interpreted. Calving margins and outlet glaciers tend to achieve stability only at topographic pinning points, and the central part of a calving glacier may fluctuate while the margins remain stable at these points. Consequently, large moraines are often found at pinning points along glacial troughs and at places where transitions between ablation mechanisms occurred. Such geomorphological evidence not only masks the true history of glacier oscillations but can appear to carry palaeoclimatic significance which it does not have.

Iceberg calving was an important ablation process along many sectors of the retreating Pleistocene ice sheet margins.

Palaeoclimatic interpretations of the deposits formed along these margins may need to be critically reassessed if, as is probably the case, they do not represent climatic halts but, instead, glacio-dynamic thresholds such as the places where calving margins retreated to grounded or non-calving locations. Moraines at such locations are probably 're-equilibration moraines' (Hillaire-Marcel and others, 1981) not truly reflecting climatic events. On the basis of an increasing wealth of chronostratigraphic evidence around the margins of the former Laurentide Ice Sheet, Teller (1987) goes so far as to say that:

> ... most ice-marginal fluctuations during deglaciation were non-climatic in origin. (p.61)

Valid palaeoclimatic inferences can be drawn from ice sheet marginal deposits only where such glacio-dynamic explanations can be explicitly ruled out.

However, the very mechanisms which render calving glaciers unstable and thereby complicate their climatic response can, in some circumstances, provide unique insights into sequences of past climate change. The anomalously large advances and retreats sometimes lead to the burial of organic material which can provide dates at times when nearby land-terminating glaciers were fluctuating on a small scale across an unvegetated proglacial area. Thus some short-term climatic events may only be recorded in the depositional record formed by calving glaciers. In addition, the discrepancies in the timings of glacial maxima of adjacent glaciers may provide confirmatory evidence as to the timing of a particular climatic event if the glacier dynamics can be reconstructed. An example of the kind of positive insights

235

which can be gained from anomalous calving glacier oscillations is given in a case study in South Georgia by Clapperton and others (1989).

5.2. UNANSWERED QUESTIONS.

Much effort is currently being expended worldwide in attempts to gain a greater understanding of the earth's climate system. Iceberg calving represents the dominant mechanism through which the cryosphere interacts with the oceans and, indirectly, with the atmosphere. It has played this important role throughout much, if not all of the Quaternary. The past is the only key we have to the range of future possibilities, and it is therefore important to gain as full an understanding as possible of the significant interactions and sensitive thresholds within the system. Considering the prominence of iceberg calving within the earth's climate system both today and in the past, our level of knowledge of the processes involved is relatively limited. If we are to improve the validity of palaeoclimatic reconstructions and make useful predictions concerning the likely timing and nature of future responses of the climate to natural and anthropogenic impacts, answers to the following questions would aid us in our task.

5.2.i. What controls calving rates?

No single calving law has yet been devised which incorporates all the variables that are known to affect calving rates on various timescales. The best relationship for grounded tidewater glaciers is that between calving rates and water depth. The reason for this relationship is poorly understood, and it does not provide a good fit with data of a duration of less than one year. Over shorter periods calving rates bear a closer relation to subglacial water discharge. Therefore the unsatisfactory situation outlined by Brown and

237

others (1983), whereby different calving laws apply over different time periods, continues. The importance of longitudinal strain rate, tidal action, melting at the calving front, and extreme meteorological events (especially the effect of waves) are poorly defined.

No calving law exists for floating glaciers. It is known that tides and wave action play a mechanical role in promoting calving by increasing stress fatigue but it remains unclear whether variables affecting ice velocities or those directly involved in the calving process control the calving rate. Furthermore, a curious puzzle has been discussed by Pelto and Warren (in press) whose annual calving data for grounded and floating glaciers from various parts of the world seem to suggest that calving rates at floating glacier margins are also controlled by water depth. This relationship cannot be causal and remains unexplained.

The mechanics of calving are also a matter of current debate. Hughes and Nakagawa (1989) have advocated their idea of bending shear whereby the relative ice movement leading to fracture occurs towards the top of the ice wall, whereas Jania (1988), Iken (1977) and Funk and Müller (unpublished) see calving as being analogous to landslip slumping through basal rotational slip with the shear bands dipping up-glacier. Both ideas are based on empirical observation. It may be that both occur at different points along one calving front, or at one glacier at different times, depending on the geometry of the ice front and of the glacial trough. The interacting roles of ice temperature, water temperature, ice cliff height, the degree of oversteepening of the ice front,

238

and basal sliding in affecting subaerial and submarine calving mechanisms and rates are not well known either.

5.2.ii. What are the thresholds of instability?

It has been established that calving glaciers can behave in ways unrelated to climate. However, it is not known at what point along the spectrum from a land-terminating glacier to a floating glacier the decoupling mechanisms set in. Do they become operative progressively and smoothly as calving becomes progressively the more dominant ablation mechanism, or is there a threshold which has to be exceeded before the mechanical instabilities become important within overall glacier dynamics? What are the relative percentages which melting and calving contribute both to overall ablation and to ice losses at the calving front at glaciers in different parts of the world?

Such data are very sparse. What information there is suggests that the percentage of total ablation contributed by iceberg calving does not need to be very high in order for the resulting instabilities to have a significant effect on glacier dynamics. For instance, Jania (1988) and Jania and Kolondra (unpublished) found that mass loss by calving at grounded tidewater glaciers in Spitsbergen accounts for only 25 - 30% of total ablation, and Funk and Röthlisberger (1989) found melting to be of major importance at the freshwater calving front of Unteraargletscher in Switzerland. The most recent work in Spitsbergen, however, suggests that calving may constitute as much as 65% or more of overall ablation (Glazovsky and others, 1990).

5.2.iii. Do calving ice sheet outlet glaciers oscillate cyclically?

The body of theory about calving glaciers which has been developed from studies of icecap outlets of Alaskan ice caps appears to apply to ice sheet outlets in many regards. However, the concept of cyclical behaviour within fjord systems has not been specifically tested outside Alaska, although behaviour of an analagous kind was proposed by Mickelson and others (1981) and discussed further by Brown and Mickelson (1990) as an explanation for the rapid oscillations of the Laurentide Ice Sheet margin in Lake Michigan during the Late Glacial. Theoretically, there is no reason to think that calving ice sheet margins should not behave in this manner. However, several contrasts between the Alaskan context and ice sheets might substantially change the spatial and temporal impact of cyclical oscillations.

Firstly, the possibility of ice divide migration in the accumulation areas might in some way alter the relationships involved. Quite how this would affect the mechanism is hard to predict because of our limited understanding of the interaction between neighbouring ice sheet catchments and of the controls on their relative sizes.

Secondly, scale contrasts may reduce the likelihood of the occurrence of cyclical change. An oscillation of an icecap outlet glacier up and down the full length of its fjord effects a much greater relative change in the Accumulation Area Ratio than it will for most ice sheet outlets. This is because large parts of the accumulation and ablation areas of ice sheet outlets are up on the higher

240

parts of the ice sheet, largely free of topographic constraints. The climatic equilibrium of ice sheet outlets is unlikely, therefore, to be destroyed to the same extent as at glaciers fed from smaller, rock-bounded catchments, and the degree of decoupling from climatic forcing will be that much reduced.

Thirdly, contrasts in sediment load and sediment supply may be important in affecting retreat and advance rates of ice sheet outlets, by influencing both calving rates and ice velocities (Alley, 1990; Kamb, 1990). Intuitively it is to be expected that sediment is of less significance for ice sheets than for ice caps and smaller glaciers because sediment loads are frequently lower for the former than the latter. If ice sheet outlets are subject to the same kind of cyclical oscillation as ice cap outlets, this would render the advancing phase of the cycle longer due to lower sediment supply. This would in turn render the period of insensitivity to climatic change that much longer.

If it is found that such cycles do occur at ice sheet margins, it may force a re-evaluation of our interpretations of recent and former marginal oscillations. The duration of these cycles, dubbed the 'insensitive period' by Mann (1986), varies considerably between fjord systems, depending on factors such as topography, scale, ice velocities, and sediment supply, and so it is possible that some of the puzzling asynchroneity displayed by the Greenland Ice Sheet margin both recently and during the Late Glacial/Early Holocene may be partly explicable in these terms.

However, the balance of the evidence at the moment

241

suggests that, while the termini of calving ice sheet outlet glaciers are susceptible to the same kinds of instabilities as are calving glaciers fed from rock-bounded catchments, cyclical oscillations are not the norm. The trends of frontal behaviour in West Greenland in historical time (Weidick, 1968) show that the majority of all the outlet glaciers, calving and non-calving alike, reached extended positions during the Little Ice Age and have retreated during the present century. Certainly there are exceptions to this generality, and most of those exceptions are calving glaciers, but the broad pattern of agreement argues against cyclicity for the following reason.

The model of tidewater glacier oscillation developed from the Alaskan evidence suggests that the advancing phase of the cycle is about an order of magnitude longer than the (often catastrophic) retreat phase. Therefore, at any one time, one might expect to find about 90% of tidewater glaciers advancing, irrespective of the climatic trend. This is not the case. Indeed, in the early parts of this century, almost all the calving glaciers were in retreat. This would seem to suggest that the instabilities introduced to the glacier system by iceberg calving affect the magnitude, rates, patterns and timings of frontal change but do not fundamentally decouple these glaciers from the climatic signal in the way in which the cyclicity theory suggests.

Intuitively, the most natural way to account for this apparent difference between the calving outlets of ice sheets and smaller ice masses is the contrast between ice-bounded and rock-bounded catchments. Ice divide migration

242

may well represent a self-regulatory mechanism which damps the effectiveness of the disequilibrium forces introduced by calving.

However, it is interesting to note that even in the Alaskan 'type area' from which the model of cyclical behaviour has evolved, Meier and others (1980) found that only 7 (13%) of the 52 tidewater glaciers have been advancing in recent decades.

This raises two inter-related questions:

i). It would seem to call into question the universality of the model itself. Actual cyclical behaviour has been well demonstrated for only three drainage basins, the Lituya system (Mann, 1986), the Hubbard system (Mayo, 1988), and the Glacier Bay system (Post, 1975). Elements of the cycle have been observed in many other glaciated basins (eg. Mayo and others, 1977), but it remains unclear in most instances whether the instability mechanisms are merely amplifying/damping the climatic signal or whether genuine cyclical oscillations are occurring. Presumably the geometry of the fjord system will largely determine the extent to which calving instabilities result in non-climatic behaviour, as envisaged by Mercer (1961; Fig. 2.10). If so, then Mann's (1986) point about the preferability of case studies to generalities of dubious applicability is reinforced.

ii). The fact that a large majority of both the Alaskan and Greenland tidewater glaciers reached Little Ice Age maxima at some stage poses the question: How insensitive are glaciers to climate during the 'insensitive period'? It would seem that, during times of strong climatic change such as the

243

Little Ice Age, the coldest period of the Holocene, climate reasserts its dominance over glacial behaviour and the instability mechanisms serve only to modulate the pattern. Perhaps mechanical decoupling from climate, and the resulting cyclicity, only operates dominantly during times of minor climatic change. Alternatively, it may be that cyclicity is only triggered after episodes of strong, climatic advance which bring calving glaciers forward to 'vulnerable', meta-stable locations; retreat from such positions will tend to be catastrophic and of great magnitude.

Both these possibilities raise the question of how great the magnitude of climatic change needs to be in order to qualify, on the one hand, as 'minor', and on the other, as 'strong'. Again, it is unlikely that a general answer can be provided since what constitutes a strong climatic change in one drainage basin may not be strong enough in another to trigger cyclical oscillations.

In summary, then, there is no clear evidence yet that calving ice sheet outlet glaciers oscillate cyclically. It is highly unlikely that cyclical behaviour takes place in basins with floating calving fronts because of the contrasting dynamics and controls on grounded and floating fronts. Case studies of individual catchments and fjord systems with grounded outlet glaciers may yet reveal cyclicity. However, it is unlikely to occur in all catchments with grounded calving termini, and, even where it does operate, there are good reasons for believing that the nature and rates of cyclical oscillations will differ from those in smaller, rock-bounded catchments.

244

5.2.iv. Is there a systematic dynamic difference between tidewater calving and freshwater calving glaciers?

The work of Mayo and others (1977) in Alaska, Clement (1983a) in South Greenland, and Funk and Röthlisberger (1989) in the European Alps all show that calving rates at freshwater calving fronts are lower for any given water depth than at tidewater fronts. The results presented in Chapter 3 provide a first indication that the response times of freshwater glaciers and tidewater glaciers to climatic change may also be systematically different. This idea will remain conjectural until given a thorough empirical testing. The questions that remain, therefore, are as follows:

i) Are these calving rate contrasts universally applicable?

ii) What accounts for these differences?

iii) Do freshwater glaciers always respond to climatic change in ways that contrast in magnitude and timing with the response of tidewater glaciers?

These questions have considerable potential significance because, at many stages during the retreat of the last great ice sheets, there were extensive sectors calving into both oceans and large proglacial lakes. If there is a systematic difference between the climatic response of freshwater and tidewater calving margins, this should be apparent in the glacial geologic record and may contribute to our understanding of the dynamics of large ice masses and the ways that they interact with the other components of the earth's climate system.

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TOPOGRAPHIC AND GLACIOLOGICAL CONTROLS ON HOLOCENE ICE-SHEET MARGIN DYNAMICS, CENTRAL WEST GREENLAND

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ABSTRACT

The retreat of the West Greenland ice sheet from its Sisimiut (Wisconsinan) glacial maximum, was punctuated by a series of stillstands or small readvances that formed numerous moraines. These landforms have been interpreted in the past as the result of short-term, regional falls in ablation-season temperatures. However, mapping of the geomorphological evidence south of Ilulissat (Jakobshavn) suggests that retreat behaviour was not primarily governed by climate, and therefore that the former ice margins are not palaeoclimatically significant. During warm climate ice-sheet wastage, the successive quasi-stable positions adopted by the ice margin were largely governed by topography. The retreat of the inherently unstable calving glaciers was arrested only at topographically-determined locations where stability could be achieved.

INTRODUCTION

The West Greenland ice sheet retreated from a marine glacial maximum position to become predominantly terrestrial during the Holocene; this study examines the influence of topography and ice/water interactions on the nature of this transitional retreat. There is a close link between topographic influences, marginal environmental conditions, and total ice-sheet stability (Mercer, 1961; Brown and others 1982; Payne and others, 1989). The areal extent of an ice sheet is controlled primarily by changes in equilibrium line altitude (ELA), sea level, and calving rate, and the relative dominance of these depends on the terminal environment (Hughes, 1987). Iceberg calving appears to play a key role in the collapse of ice sheets (Denton and Hughes, 1981; Pollard, 1983) and during this process topography can exercise a powerful control on the ice margin (Mercer, 1961); narrowings, bends, bifurcations, and changes in the sideslope gradient are places where the terminus of a fjord glacier can achieve greater stability. Apparently anomalous behaviour can result from the intersection of the ELA by a thickening glacier, and thus terminus behaviour may be unrepresentative, or even contrary to climatic trend.

Work on the calving tidewater glaciers of Alaska (Meier and Post, 1987) has extended Mercer's original ideas. It has shown that retreat rates of grounded calving glaciers are controlled by channel shape and water depth, that the calving rate increases almost linearly with water depth, and consequently that the climatic significance of grounded tide-water glacier fluctuations is problematic. According to Mann (1986), the fjord glaciers most susceptible to anomalous responses to climate are those which occupy complex fjord systems, lie in areas of low (ELA), or terminate at fjord mouths adjoining the open sea. Existing work therefore suggests that the key variables determining the extent of topographic control of an ice-sheet margin are trough geometry, relative sea/lake level, and resulting water depth.

WEST GREENLAND: CONFLICTING HYPOTHESES

Weidick (1968, 1985) was the first to attempt an
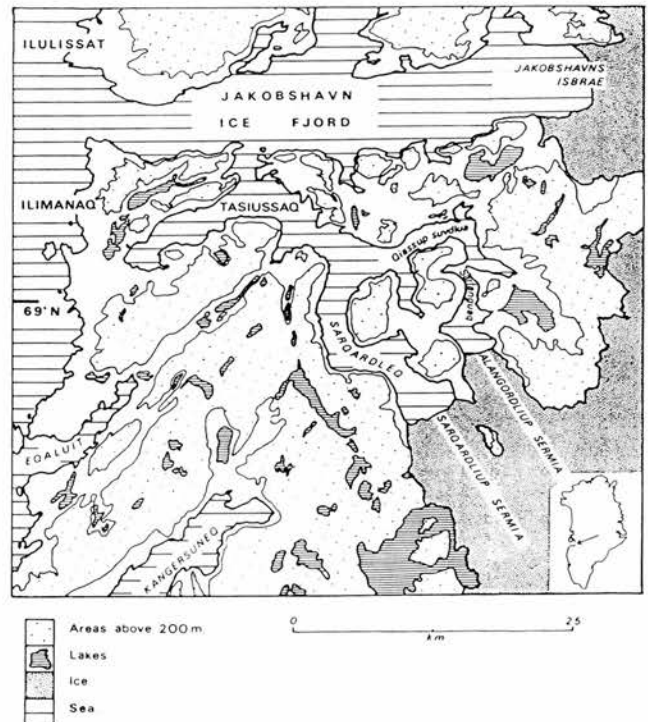

Fig. 1. Jakobshavns Ice Fjord and the Tasiussaq Fjord system.

overall chronological interpretation of the glacial deposits of West Greenland. Major uncertainty exists concerning their origins, and two hypotheses can be advanced to account for them:

" ...They may represent either significant readvance of the ice sheet margin in response to regional positive changes in ice mass balance, albeit during periods of overall deglaciation, or alternatively, local stillstands or readvances caused by topographic influences or by shifts in the ice-sheet streamline patterns during deglaciation" (Kelly, 1985, p.477).

Weidick proposed the former, interpreting the moraine systems as evidence of short periods of colder summer temperatures during the collapse of the ice sheet, consequently suggesting that the ice margin features are the product of climatically-controlled changes in mass balance. The alternative possibility, that the characteristics of the marginal environment may themselves influence the stability of the Greenland ice sheet, has seldom been considered. In this second scenario, halts or readvances during ice-sheet retreat are caused by a sequence of changing ablation mechanisms at the margin. Thus Funder (1989) has noted that the location of moraines along the fjords indicates that they were formed by interaction between the glaciers and bed topography, rather than by climatic change.

THE STUDY AREA

The research area (Fig. 1) lies in a 15–30 km wide coastal strip of plateaux and hills below about 550 m, separated by long, structurally aligned glacial troughs in Precambrian gneiss. The Tasiussaq Fjord system averages 100–200 m in depth (Engell, 1905). Formation and/or preservation of glacial deposits was often impossible along the main glacial troughs, because of steep slopes and cliffs up to 500 m high. It is an area of retreat-stage transition from the maximum, marine-based state of the inland ice to its current, predominantly terrestrial condition. The inland ice retreated from a Sisimiut glacial maximum position in Disko Bugt at some time between 14 000 and 10 000 B.P., reaching a position close to the present margin around 7000 B.P. (Weidick, 1985; Kelly, 1985). Two major phases of stillstand were identified by Weidick (1968) and proposed as contiguous margin positions between the three valleys of Tasiussaq, Eqaluit and Kangersuneq (Fig. 1), the younger dating from 7500 to 8500 B.P., the older phase antedating it by an unknown amount. Large depositional features forming the western shore of Tasiussaq (Weidick, 1968) are the only West Greenland example of a definite Holocene readvance (Kelly, 1985).

RESULTS

The main geomorphological elements of the landscape were mapped onto aerial photographs at a scale of 1 : 40 000, glacial landforms being identified on the basis of morphology and sedimentology, and used to delimit former ice margins (Fig. 2). Large accumulations of glacial debris are commonly found at topographic narrowings or bifurcations. In the Sarqardleq trough, for example, the main lateral moraines indicate terminus positions at successive narrowing points up the fjord (Fig. 3). South of the narrowing point of the bay Igdlup qingua, a continuous mantle of morainic debris, buries the bedrock but is absent where the bay opens out into the main Tasiussaq trough. On the west side of the valley of Akiamiut taserssua, moraines coincide with prominent bedrock ridges projecting into the lake. Similarly, a moraine 3 m high occurs on a narrow rock ridge which separates the two lakes at the north end of Qivdlertup valley, and in this valley the large glacial deposits are concentrated where the valley bifurcates around an isolated hill 230 m high. Between such points, lateral moraines tend to be smaller, less continuous, and show a diverging spatial pattern; moraine ridges in close
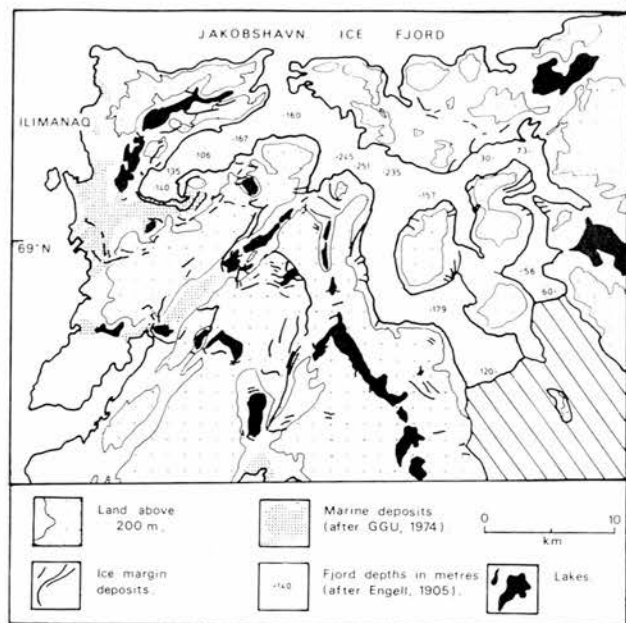

Fig. 2. Tasiussaq Fjord system: glacial geomorphology and bathymetry.

association diverge as they are traced down-slope, down-fjord, and down towards present or past water bodies. On the western flanks of Sarqardleq, for example, where the fjord is narrowing southwards, 10 m of elevation separates lateral moraines which relate to terminus positions 1.5 km apart.

DISCUSSION

The spatial pattern of the glacial deposits indicates that the ice margin retreated in episodic fashion, alternating between rapid retreat and relative stability, and that parts of the margin were stable when other parts were unstable. Since variation in ablation is likely to have controlled ice-margin oscillations, episodic retreat was probably caused by variation in the factors controlling ablation rates.
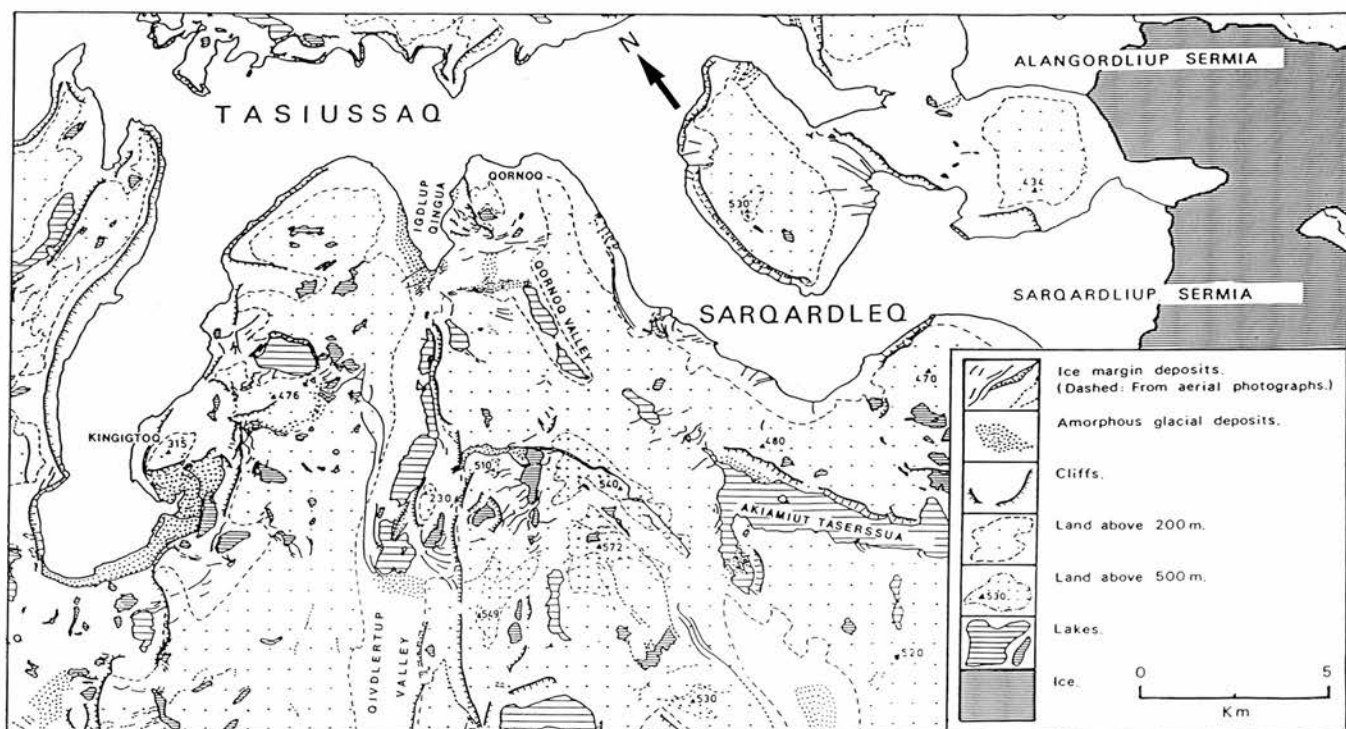

Fig. 3. Detailed glacial geomorphology of the central area.

Climatic control

Weidick (1968, 1985) proposes that cooler summers during the Early Holocene warming caused the stillstands. According to Weidick, some of the larger moraines extend over long distances, and are of similar morphology; although the continuity over the high ground is not always clear, the connection between the large deposits in the neighbouring troughs is obvious. If the moraines *are* the result of climatically controlled changes in regional mass balance, then this must be the case. However, palaeontological and palynological evidence (Donner and Jungner, 1975; Fredskild, 1985) suggests that the Early Holocene in West Greenland was a period of uninterrupted warming and drying, reaching a peak warmer and drier than today sometime between 7000 and 6500 B.P. The sub-Arctic West Greenland Current was established from 9100 B.P., and there was intrusion of water warmer than at present from 7300 B.P. until after 5000 B.P. (Kelly, 1985). For the rapidly wasting ice sheet to halt, or readvance slightly in positions far in advance of today's margin (which is in quasi-equilibrium with the existing climate), marked falls in summer temperatures for substantial periods of time would be necessary, yet neither the fauna of the time nor the ice core evidence (Hammer and others, 1986) indicate anything more than minor cooling episodes in this period. In historical times, even slight temperature decreases have led to minor advances of the Inland Ice, but this effect is only known to have occurred recently, involving an ice sheet in quasi-equilibrium; it is not known whether such responses would occur in the context of rapid ice-sheet retreat. It is questionable, therefore, whether the retreat stages of the ice-sheet margin were controlled climatically. Indeed, Weidick in his more recent paper states that the recession was "often locally determined by topography rather than by changes in mass balance" (Weidick, 1985, p.306).

Topographic control

The glaciers appear to have halted only at places where the topographic configuration gave them an increased degree of stability. Retreat beyond such points into wider and/or deeper parts of the fjord led to increased rates of iceberg calving, and continuous, rapid retreat until the next topographic "pinning point" was reached. Using the ideas of Mercer (1961), Figure 4 presents a hypothetical reconstruction of the ice-sheet retreat in the Tasiussaq Fjord

system c. 9000–7000 B.P., showing only the *present* coastline, and also possible contemporary positions of Jakobshavns Isbræ as dashed lines.

The outermost moraine systems may be an example of the hypothetical sequence of evolving marginal dynamics envisaged by Hughes (1987, fig. 9, p.189). The warming climatic conditions and rising sea levels of the Early Holocene are thought to have caused catastrophic calving and dramatic retreat of this marine-based sector (personal communication from S. Funder). The transition from a marine-based to a predominantly land-based ice sheet would have led to greatly reduced ablation rates due to the fall in iceberg production. Furthermore, isostatic rebound was rapid at this time (Donner and Jungner, 1975) so that the length of the transition period would have been shortened and the dynamic impact consequently heightened. Just as glacio-isostatic sinking compels the grounding line to retreat (Hughes, 1988), so glacio-isostatic emergence enables it to advance. Under such circumstances a halt or a readvance would be expected.

Two distinct states of this transition can be recognised. The first was the time when the ice-sheet margin initially reached a grounding position at or near the contemporary coastline, and the second when the rapidly-calving outlet glaciers retreated either to stable, grounded calving locations or ceased calving altogether. The outermost moraines south-west of Tasiussaq show that the margin closely paralleled the coast, strongly suggesting that here, as in many parts of West Greenland, "the sea/land transition formed a major glacio-dynamic obstacle in the deglaciation process" (Funder, 1985, p.140). The readvance of the Tasiussaq ice lobe relates to the second of these transition phases. Initially, rapid calving probably led to retreat to the narrows of the fjord at Kingigtoq hill, a very sudden topographic constriction relative to the unconfined plain to the west. Dynamic adjustment to the much reduced ablation area and continuing isostatic rebound then caused a readvance, followed by adoption of a stable, grounded position. The complex glaciomarine sedimentology of these large deposits suggests that this readvance occurred on a morainic shoal as observed in many fjords in Alaska (Meier and Post, 1987). The marine clay plain to the west represents a contemporary sea level at 50–55 m a.s.l., suggesting a date for this readvance of 7250–7500 B.P. from the emergence curve of Donner and Jungner (1975).

It is therefore proposed that the dominant outer moraines of the area relate to this period of adjusting ice dynamics following retreat out of the sea. The topography of the coastal zone and the relative altitudes of these parts of the glacial troughs (45–55 m a.s.l.) are such that this transition happened at almost the same time in each of the major valleys, as shown by Weidick's (1968) ^{14}C dates. Thus these ice-margin deposits may be divided broadly into an outer and an inner phase, as originally proposed by Weidick (1968), but the primary causal link is a dynamic response to topography rather than climate.

CONCLUSION

The nature of the episodic retreat here is governed by the influence of topographic configurations on marginal ice dynamics, particularly iceberg calving dynamics. The evidence indicates that stepwise retreat occurred as envisaged by Hughes (1987, 1988). The contemporary warm climate was the dominant forcing function causing ice-sheet retreat, and led to destabilisation from pinning points, but the geomorphological evidence shows that the stillstands were topographically induced. An implication is that palaeoclimatic significance should not automatically be attached to the Late Glacial and Holocene stillstands of tidewater outlet glaciers of ice sheets, especially stillstands resulting from the sea/land transition.

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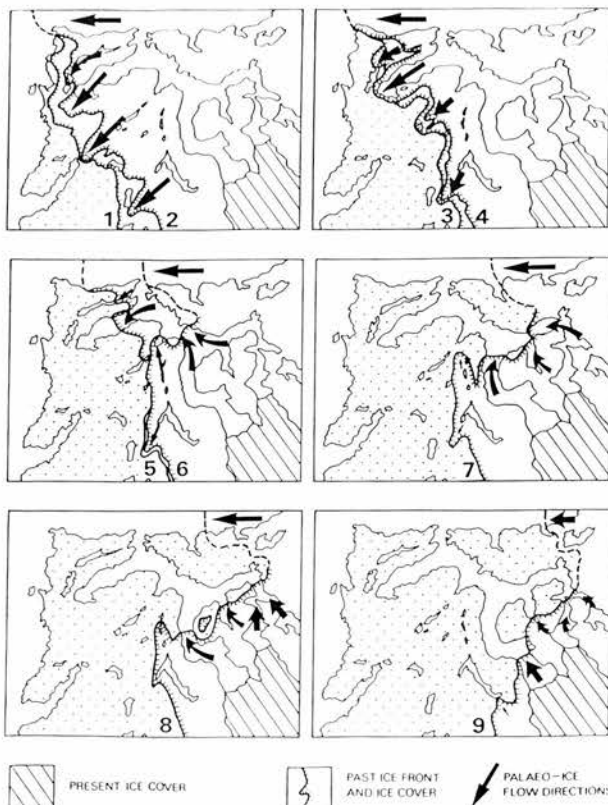|	
PRESENT ICE COVER	PAST ICE FRONT AND ICE COVER
	PALAEO-ICE FLOW DIRECTIONS

Fig. 4. Reconstruction of ice-margin configurations during retreat, c. 9000–7000 B.P.

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