Chapter 14

MODERN GLACIOMARINE ENVIRONMENTS

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14.1. INTRODUCTION

Glaciomarine sediment contains the most detailed and continuous direct record of Late Cenozoic glacial fluctuations and often the first direct indications of a glacial period. In addition, many of the Earth's pre-Cenozoic glacigenic sedimentary rocks are interpreted as glaciomarine (Hambrey and Harland, 1981; Anderson, 1983). Knowledge of glaciomarine environments and processes that occur within them is, therefore, invaluable.

Glaciomarine (glacimarine, glacial marine) environments may be defined as marine environments in sufficient proximity to glacial ice that a glacial signature can be detected within the sediments. Some or all of the sediment is released from grounded or floating glacial or sea ice (cf. Table 14.1). The environment has a plethora of processes and sediment sources due to the combined action of ice (glacial and sea), water (fresh and sea), wind and biological activity.

Knowledge of glaciomarine processes has advanced in the past decade. However, large gaps still remain in understanding these environments and many interesting aspects remain unsolved. Sedimentation processes in cool-temperate, boreal and, to a lesser degree, polar-tundra climatic regions are perhaps best documented whilst processes in polar climatic regions are probably least documented.

This chapter will review present knowledge of sedimentological processes in modern glaciomarine environments. To do so, the glaciological and climatological conditions pertinent to glaciomarine environments will be discussed from the glacier to its margin in seawater and beyond to those related environments distal from the glacier. Non-glacial processes that contribute sediment to glaciomarine environments and modify previously deposited sediment will be considered. An evaluation will be made concerning rates of sedimentation and fluxes of sediment through glaciomarine depositional systems. These aspects are crucial in considering the nature of the sediment record left by glaciers, that ultimately are controlled by advances and retreats of marine-ending glaciers. Finally, glacier fluctuations are discussed in relation to the packages of sediment within a glacial record.

14.1.1. Glaciation and Climate

Glaciation at sea level depends on latitudinal and regional climatic factors such as winter and summer temperatures, insolation, snowfall and cloudiness, and a wide range of climatic conditions which, when

TABLE 14.1.

Climate Region	Summer Temperatures	Glacial Temperature	Equilibrium Line Elevations
Polar Ice Cap	<0°C	<0°C, except at base under special conditions	At calving line or seaward*
Polar Tundra (Subpolar)	<10°C	<0°C	At calving line or 100s m above s.l.
Boreal	>10°C (1 to 3 mo.)	At and below pressure melting	100s m above s.l.
Temperate Oceanic	>10°C (4 to 7 mo.)	At pressure melting throughout	100s to 1,000s m above s.l.

*The potential equilibrium line elevation below present sea level can be calculated, for instance see Robin (1988).
Climate regions and summer temperature characteristics are taken from: Trewartha, G.T., 1968, An introduction to weather climate, McGraw-Hill.
After Anderson and Domack (1991).

combined with relief, are capable of maintaining glaciers at sea level (Table 14.2) (Chapter 4). Most glaciers of coastal Antarctica, for example, are in the Polar Ice Cap category where mean summer temperatures are less than 0°C and ice temperature is subzero. Surface ablation is limited mainly to sublimation; therefore, even though accumulation averages <2 cm year^{-1} on the surface interior of the ice sheet, glaciers can be maintained at sea level. Meltwater and terrestrial vegetation are limited or lacking under these conditions. Unglaciated areas are cold, polar deserts with little atmospheric moisture and rapid evaporation. Thicker portions of the ice sheet and its major drainage systems are melting or melting/freezing at their base. Sea ice is extensive and can be strongly seasonal such as surrounding Antarctica, or persistent as in the Arctic Ocean.

The transition from Polar Ice Cap to Polar-Tundra or Subpolar climatic conditions occurs where mean summer temperatures are >0°C. These conditions occur along the western side of the Antarctic Peninsula, sub-Antarctic islands, coastal Greenland, arctic Canada, Iceland, Svalbard, Nova Semlya and Severnaya Zemla. Some summer melting occurs and glacier equilibrium lines are at or just above sea level, but terrestrial vegetation is limited. Glaciers, under these climatic conditions, experience surface melting during the summer but freezing temperatures remain throughout most of the glacier.

Boreal climates have one to three months with mean temperatures >10°C and, except for summers, are rather dry. These climatic conditions support large areas of mountain glaciers but only a few reach sea level, for example, along the Kenai Peninsula, north-central Gulf of Alaska. Glaciers experience surface melting, but vegetation is shrubs and lowland forest.

The mildest climate supporting glaciation at sea level is Temperate Oceanic, occurring along the eastern Pacific mountainous coasts of North and South America as far as latitude 57°N in southeastern Alaska and northern British Columbia and 47°S in western Patagonia. Mean temperatures for the four- to seven-month summers are >10°C, temperature minima are 0–2°C, and locally, rainfall occurs year-round. Extremes in ablation are balanced by very high snow accumulation, internal ice is temperate, glaciers

TABLE 14.2. Sedimentation rates of recent Antarctic glacio-marine deposits of the continental shelf

Location	Sediment Type	Rate (mm/yr)	Method
East Antarctica			
Mertz-Ninnis Trough, 1 (George V Coast)	diatom mud, ooze	3.1-3.4	^{14}C
		3.0	^{210}Pb
Prydz Bay, 2	diatom ooze	1.5-0.67	^{14}C
	silty clay (pebbly mud)	2.1	^{14}C
Ross Sea			
Terra Nova Bay 3	diatom mud	0.05-0.20	^{14}C
Antarctic Peninsula			
Andvard Bay, 7	diatom, pebbly mud	2.0-5.0	^{14}C
Hughes Bay, 7	diatom, pebbly mud	2.8	^{210}Pb
Brialmont Cove, 7	sandy mud	1.7-49.8	^{14}C
Marguerite Bay, 5	diatom mud	0.3	^{14}C
		2.0	^{210}Pb
Gerlache Strait, 5,6	diatom mud	2.7	^{210}Pb
		2.7	^{14}C
Bransfield Strait, 5	diatom mud	1.8	^{210}Pb
South Shetland Islands			
Admiralty Bay 7	sandy, pebbly mud	4.7-33.2	^{14}C

References:

1. Domack *et al.* (1989).
2. Domack *et al.* (1991).
3. Krissick (1988).
4. Nittrouer (1988). personal communication.
5. DeMaster *et al.* (1988).
6. Venkatesan and Kaplan (1987).
7. Domack (unpublished NSF report)

are melting or melting/freezing at their base and equilibrium lines are well above sea level. It is the presence of coastal mountains that allows glaciers to thrive under temperate oceanic conditions. These coastlines are biotically productive with dense ground vegetation and even forested glacier termini (e.g. Malaspina Glacier).

14.2. SUBGLACIAL PROCESSES

Generally, transportation and deposition of englacial and subglacial debris by marine-ending glaciers is similar to terrestrial glaciers (Chapter 8; Menzies, 1995b, Chapter 2). However, some differences may occur due to the fast and extensional flow behind grounding lines of major marine-ending glaciers. Knowledge on this question is poor and conjectural. Lodgement of basal debris may occur in the form of till sheets that may be discontinuous. However, because of fast basal ice flow and high basal water pressures, deforming beds may be common. Where subglacial water is in excess, most subglacial sediment appears to be flushed out by meltwater streams (Powell and Alley, *in preparation*). The presence of deforming beds and large volumes of subglacial water may produce different types of depositional processes at grounding lines as will be discussed in Section 14.5.

14.3. MARINE-ENDING TERMINI

Glaciers that terminate in seas or oceans may, in the final few 100 m or kilometres, enter the body of water as (1) a floating terminus, or (2) a grounding tidewater cliff (Plate 14.1). These different modes of terminating in water result in: variations in how sediment arrives at the point of delivery from the ice to the water body; the effect of marine conditions such as temperature, currents, and salinity on sedimentological and glaciological processes; the transient interaction between marine and glacial environments; and the specific sedimentological processes ongoing within the proximal zone of the ice, the immediate proglacial subaquatic zone and in more distal aquatic areas.

14.3.1. Floating Termini

14.3.1.1. Formation and maintenance

Floating termini form when glacial flow lines converge and flow is unconfined due to very low frictional forces between glacier and sea. Commonly the annual 0°C isotherm is at sea level, but even if not, the ice mass is cold (polar). The termini form a continuum from confined floating glacier-tongues, to confined ice shelves, to unconfined floating glacier-tongues, to fringing ice shelves, to large embayment ice shelves (Anderson and Domack, 1991; Fig. 14.1). At present 57% of the Antarctic coastline can be delineated as floating termini, 38% is tidewater termini and 5% is sediment/rock coast (Drewry *et al.*, 1982). Most marine-ending glaciers elsewhere have tidewater termini, although some in the arctic terminate as fringing ice shelves or floating glacier-tongues.

Having a large horizontal surface area at low elevation (<100 m) means that floating termini are susceptible to rapid surface ablation should equilibrium lines rise (cf. Mercer, 1970). This condition probably places a climatic limit for ice shelves at about a mean annual temperature of −8°C and mean summer temperature of −2°C (Swithinbank *et al.*, 1988). Because the temperature of ocean water is often at or above 0°C under floating ice, basal ablation of these ice masses may be just as important as surface ablation (Jacobs, 1989). Nearly all of the mass loss from the George VI ice shelf (about 4.5 to 1.8×10^3 kg m^{-2} year^{-1}) is by basal melting aided by warm ocean waters (Bishop and Walton, 1981; Potter *et al.*, 1984).

14.3.1.2. Debris transport and release

Sediment deposited under floating termini comes from glacial sources and marine currents. Basal debris is melted out from floating termini within a few kilometres of a grounding line (Drewry and Cooper, 1981; Jacobs *et al.*, 1979). However, englacial debris entrained by converging outlet glaciers along the coast and then buried by snowfall, can be released from the base by bottom melting in the distal zones of a floating terminus (Powell, 1984). These influences of debris release can be seen in the confined George VI and fringing Larsen Ice Shelves in Antarctica (Domack, 1990). Debris is probably contributed in similar styles to the Ross Ice Shelf, Antarctica, where several large outlet glaciers have grounding lines located up-valley from where they join the ice shelf (cf. Swithinbank *et al.*, 1988).

Release of debris from beneath floating termini depends on oceanic temperature and pressure gradients which are controlled by water depth and circulation (Doake, 1976). For example, water flowing toward a grounding line under a floating terminus flows down a pressure gradient and can potentially melt the basal ice (Foldvik and Kvinge, 1974), whereas outflowing water moving up the pressure gradient will probably freeze to the base

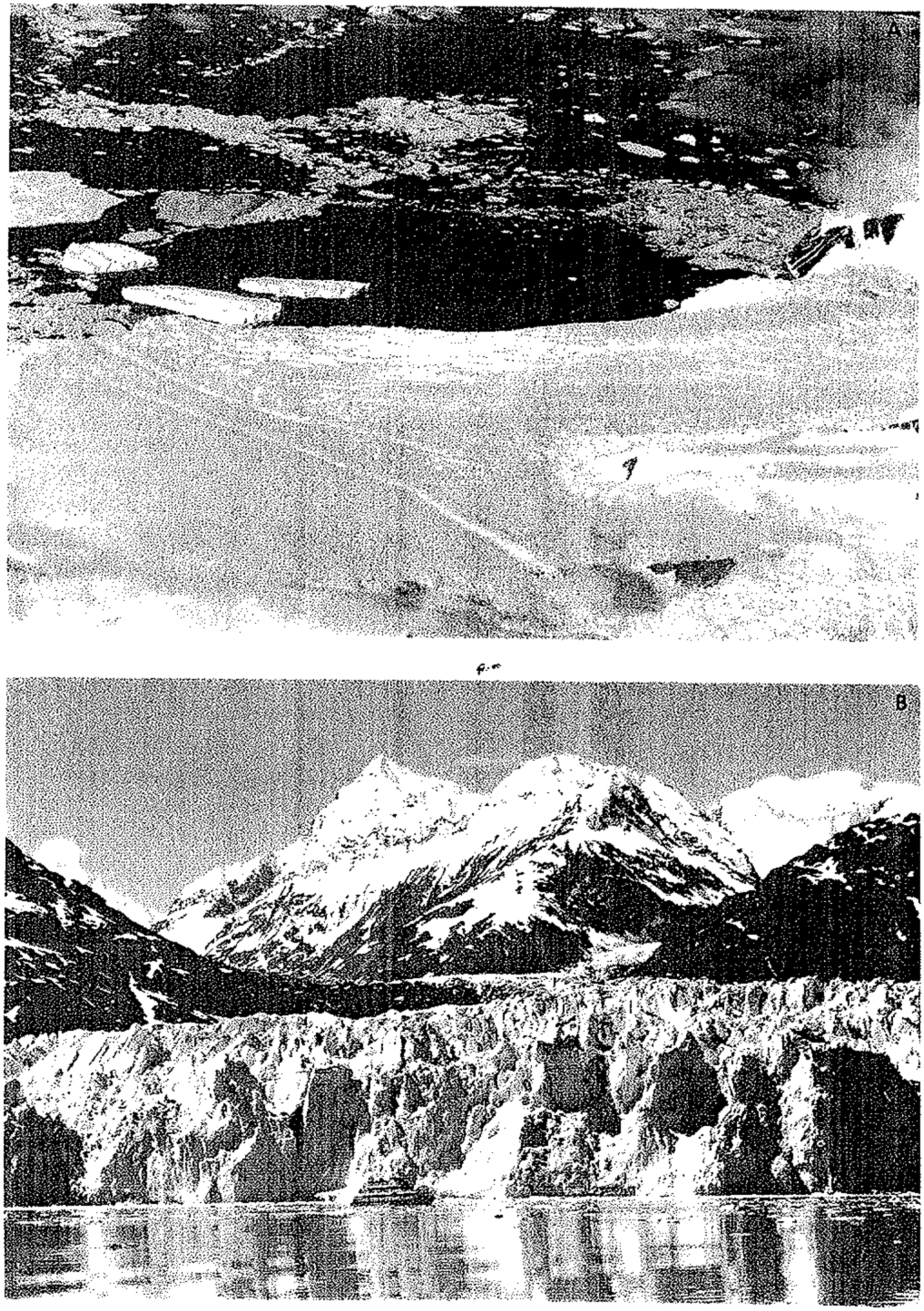


PLATE 14.1. Examples of two types of glacier termini ending in the sea. (A) A floating glacier-tongue of an outlet glacier from the East Antarctic Ice Sheet producing icebergs in the Ross Sea, in which there are also remnants of seasonal sea ice. (B) A grounded vertical cliff of the Margerie Glacier, Glacier Bay, Alaska, with its bergs and a tourist craft in Tarr Inlet.

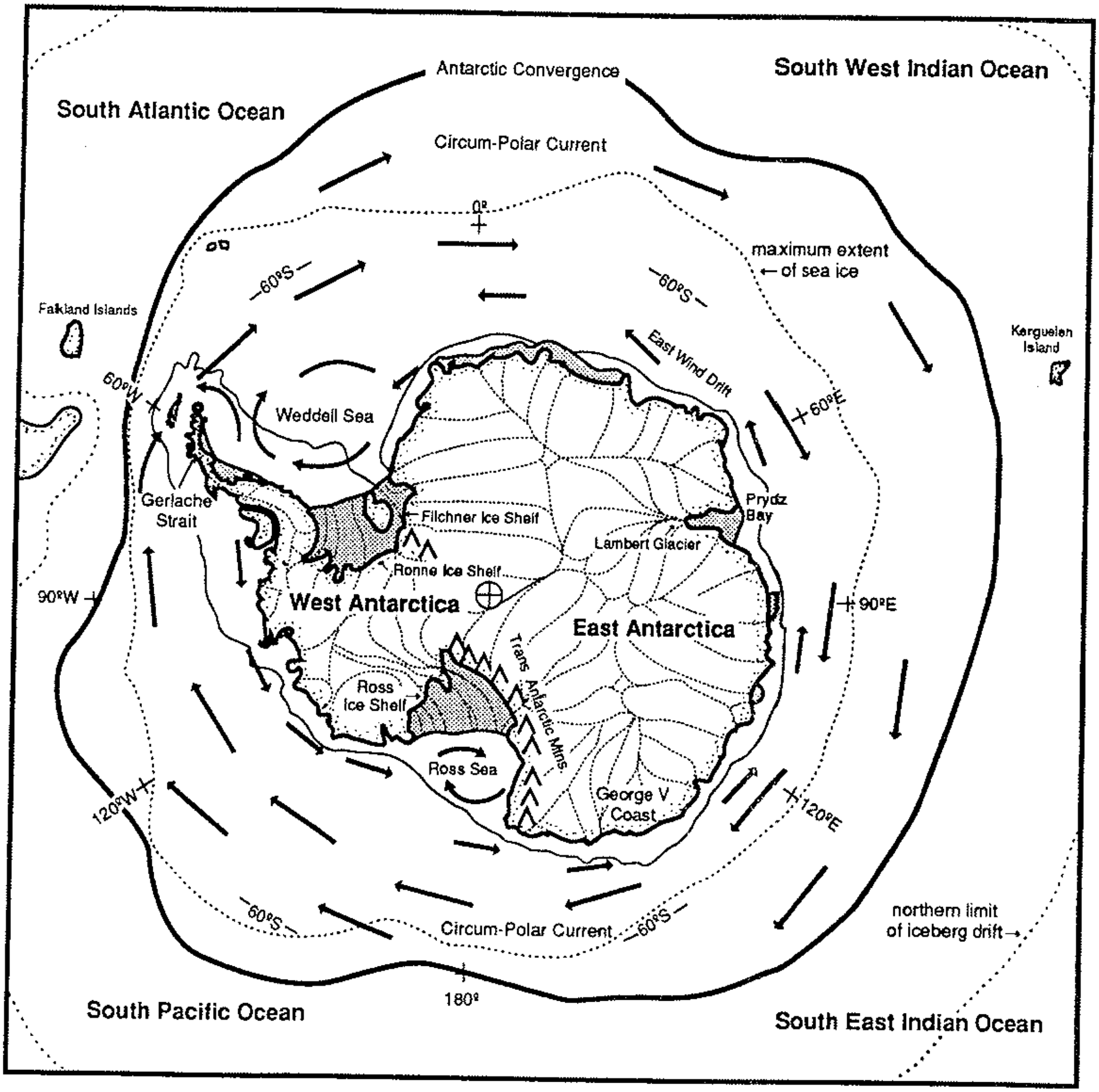


FIG. 14.1. Map of Antarctica and surrounding oceans showing coastal regions mentioned in text and illustrations. Dotted lines on continent are ice flow directions (after Drewry, 1983) and arrows in ocean are current directions. Dashed line in ocean is maximum limit of seasonal pack ice while dotted line is northern extent of drifting icebergs. Solid line in ocean is the approximate location of the southern polar front or Antarctic Convergence. Major ice shelves are indicated by the shaded areas.

(Robin, 1979). The ensuing pattern of debris release may vary amongst floating termini, especially embayment-type ice shelves, and may unpredictably change through the history of a floating terminus depending on oceanic circulation conditions.

There is growing indirect evidence from Antarctica indicating that the grounding lines of large ice shelf systems are dynamic settings with regard to sediment transport and delivery to the marine environment. An early intuitive model by Drewry and Cooper (1981) recognized the role that grounding-line cavities should play in the circulation of basal melt-out and suspended debris. Recent geophysical investigations along the Siple Coast of West

Antarctica have demonstrated that grounding lines of ice streams (fast flowing currents in an ice shelf) may occupy distinct zones of subglacial debris transport in the form of deformable, water saturated diamicton (Alley *et al.*, 1986). 'Till deltas' are one possible product of ice stream/ice shelf interaction (Alley *et al.*, 1989b, Fig. 5) (Section 14.4). The presence of thin subglacial water films is also recognized within the system but the transport of suspended particulates out into the sub-ice shelf or floating glacier-tongue environment remains a major point of future investigation (Barrett, P.J. *et al.*, 1988). Results from the first studies of these processes are described in Section 14.6.4.

Shallow water environments, especially those bordering ice cliff or piedmont settings, are greatly affected by currents of wind or tidal origin. Low siliciclastic sediment supply at the calving line allows such currents to effectively sort and transport sediment into broad aprons of sandy sediments rich in carbonate bioclastic detritus.

14.3.1.3. Ice shelf pumping

Studies of ice shelf grounding-line environments, including the Ronne Ice Shelf, have demonstrated significant vertical movement of the shelf indicating that the grounding-line may in fact undergo periodic uplift and set-down under the influence of tides (Stephenson and Doake, 1982; Kobarg, 1988). The effect within any grounding-line cavities would be similar to a bellows with the expulsion of water and suspended sediment out into the marine setting during falling tide or ice shelf let down (Talbot and Von Brunn, 1987). Unfortunately this pulsing process is difficult to observe directly but when combined with debris rain-out from undermelt, it should produce laminae and deformation structures within sediment such as diamicton. In many cases a disrupted diamicton mélange is produced (Talbot and Von Brunn, 1987).

Indications of how tidal pumping may influence sediment transport are provided by studies of partially-floating termini along the Antarctic Peninsula. Oceanographic data demonstrate that, within the water column, mid-water fine-structure consists of distinct interflows of turbid water adjacent to the glacial termini in deep water (Fig. 14.2). A mechanism similar to tidal pumping is implicated in the formation of these features (Domack and Williams, 1990; Fig. 14.3). Suspended particulates sampled from these features include quartz silt grains with a fresh appearance and even very fine-grained sand (Plate 14.2).

14.3.1.4. Calving styles and processes

Calving (sometimes termed berging) is a rather infrequent occurrence for ice shelves; the mean annual iceberg production in Antarctica is about 2800 to 3040 km^3 (Ørheim, 1985), 60 to 80% of which comes from embayment ice shelves and 15 to 25% from ice streams and outlet glaciers (Drewry and Cooper, 1981). Calving mechanisms are not fully understood but possible causes can be grouped into those related to: glacial flow (creep-failure with extensional flow, rift zone development, surface and bottom crevasses); a combination of glacial and oceanic processes (Reeh-type calving from force imbalances on the ice face (Reeh, 1968)); bending and flexure by tides and cyclic vibration (Holdsworth and Glynn, 1978); impacts of large icebergs; and oceanic processes (storm and tsunami wave energies and ocean currents) (Kristensen, 1983).

Calving processes from floating termini produce the largest icebergs which can survive long transport distances. For instance, one recent calving event from the Ross Ice Shelf, Antarctica, produced a tabular iceberg which was 154 x 35 km^2 and it drifted across 2000 km of the Ross Sea during a 22 month period (Keys *et al.*, 1990). It contained 1200 km^3 of ice which is nearly 50% of the net annual accumulation of the Antarctic Ice Sheet (Giovinetti and Bentley, 1985). The actual calving event is relatively passive and has no influence on sea floor sediment that can be several hundred meters below (Plate 14.3).

14.3.2. Tidewater Termini

14.3.2.1. Formation and maintenance

Glaciers that end as tidewater termini occur in every climatic setting (a cool-temperate example is shown in Fig. 14.4). Glaciers of temperate ice do not

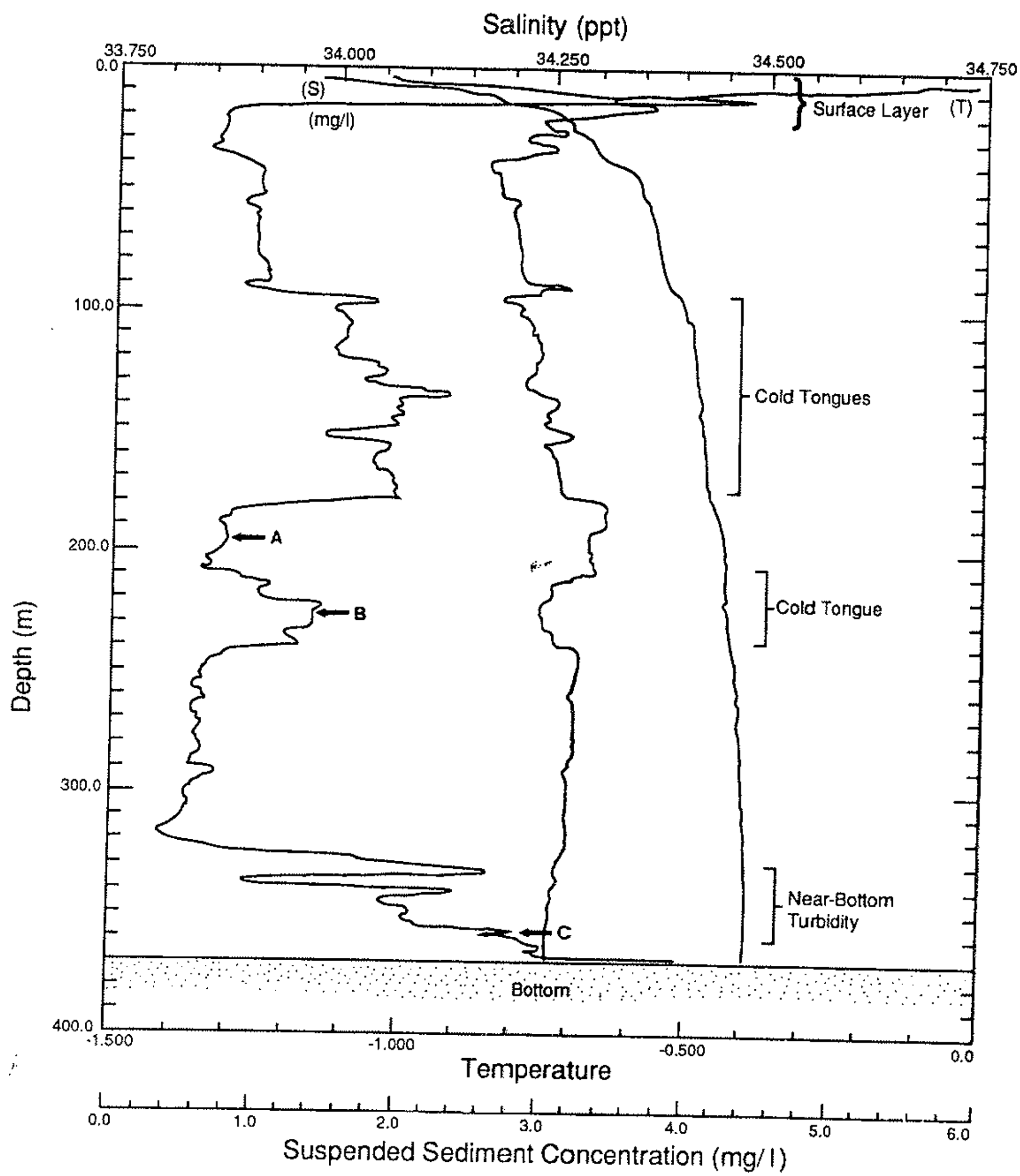


FIG. 14.2. Vertical profile of temperature (T), salinity (S) and suspended particulate matter (mg l^{-1}) within an Antarctic (subpolar) fjord. Location of SEM images (Plate 14.2) of suspended particulate matter are also shown (A,B,C). Note significant suspended particulate matter within mid-water layers (cold tongues) and comparison to surface layer and near bottom turbidity (modified after Domack *et al.*, *in press*).

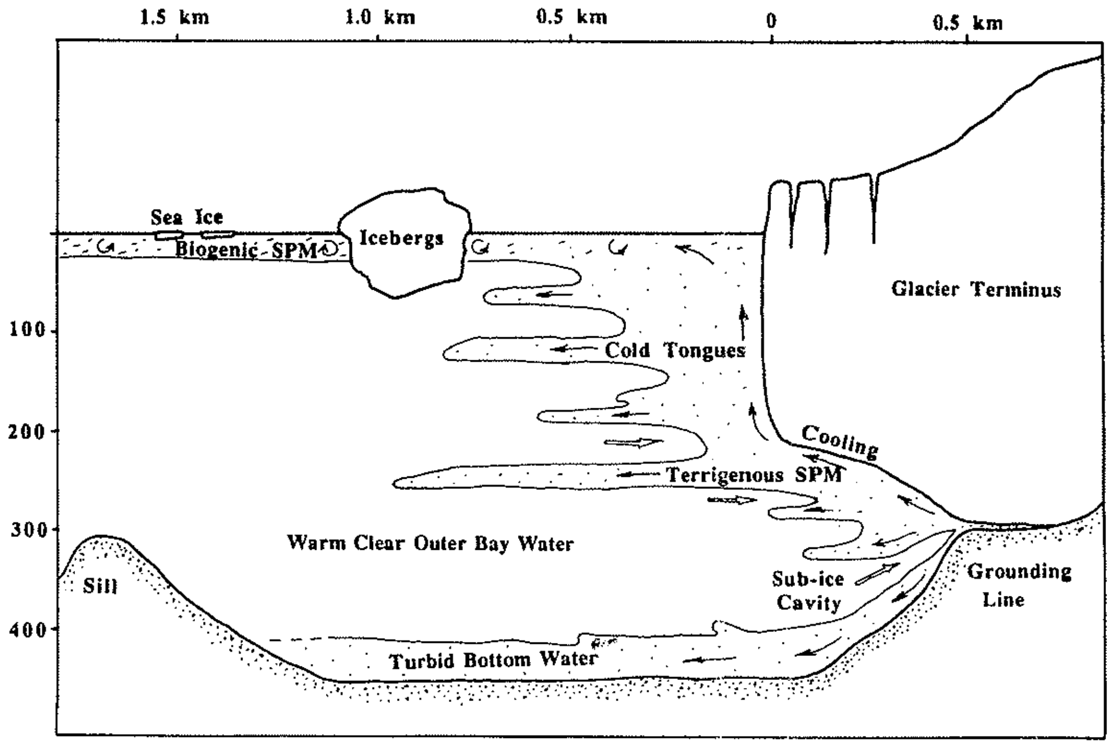


FIG. 14.3. Model for cold tongue (interflow) formation in polar fjord settings as based upon observations in Antarctic fjords (Domack and Williams, 1990; reprinted with permission of the authors and the American Geophysical Union). Circulation is driven by storm or tidal surges within the basal cavities of tidewater termini or ice shelves.

have the tensile strength to float as intact 'slabs' (Powell, 1980), while those of cold ice apparently do not have sufficient flow velocities to maintain a floating terminus (Plate 14.4).

Glaciers form a vertical cliff once they reach sea level because the glacier calves faster than it melts. As the ratio of melting to calving increases a more parabolic surface profile is produced. Under these later circumstances a cliff will remain grounded unless the conditions required for floating are met.

Maintaining a tidewater terminus appears to be primarily a function of the water depth in which the grounded cliff ends (Brown *et al.*, 1982). The ELA of glaciers with these termini is not as sensitive as those with floating termini. If the 0° isotherm is at sea level, then a cliff is maintained relatively easily in the sea. Where the ELA is at high altitude a large supply of winter snow is required to create sufficient ice flux to maintain a tidewater terminus (Section 14.9).

14.3.2.2. Debris transport and release

Downwasting of a glacier surface produces supraglacial debris by melt-out, especially on valley and outlet glaciers. Most of that debris is from supraglacial sources and, for marine termini, it often falls into the many crevasses caused by related extensional flow. Little englacial and no basal debris is moved to the surface by ice flow at the terminus, because flow lines, under these terminus conditions,

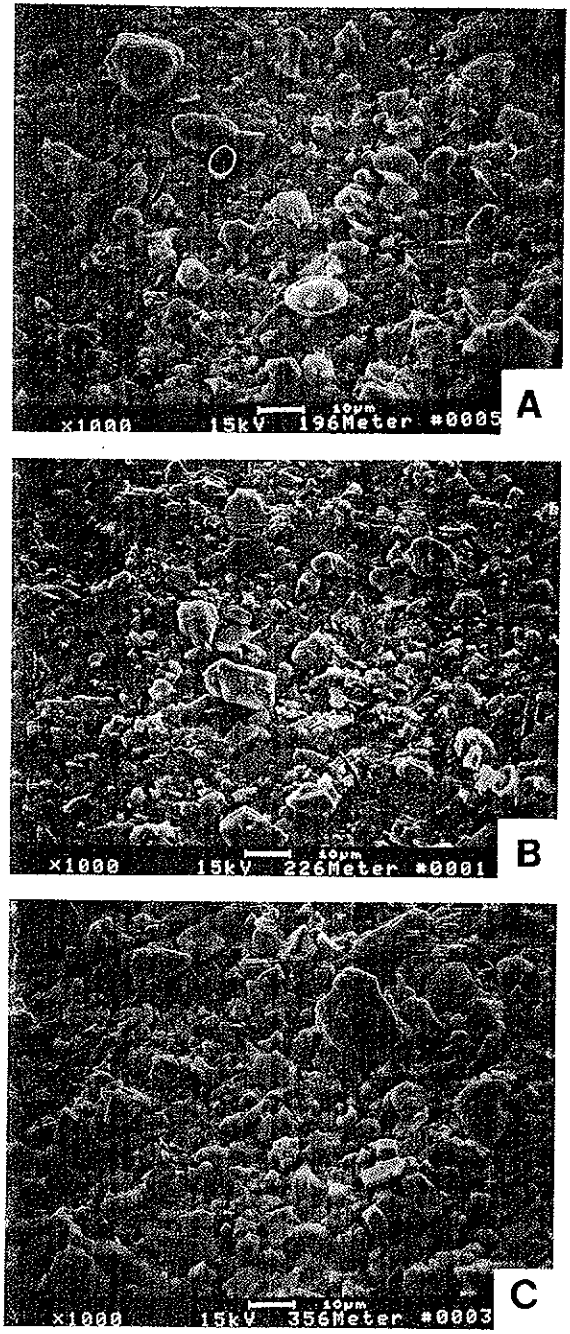


PLATE 14.2. Scanning Electron Microscope photographs of suspended particulate matter within oceanographic features shown in Fig. 14.2. Note dominance of siliciclastic particles in all samples especially sample B from mid-water cold tongue. Greater biogenic (diatom) constituents are observed in sample from less turbid water at 196 m (A).

remain near-horizontal. Englacial debris derived from up-glacier tributaries is often exposed in cross-section on the cliff face in near-vertical layers which may be deformed by local ice flow. All basal debris being horizontal remains near the sea floor and is only exposed above sea-level in shallow water depths.

Melt-out/Fallout. A tidewater terminus is melted by solar heat, stream discharges and sea. In general, melting by air is much slower than calving and other melting rates (Powell and Molnia, 1989). Meltwater streams rising up the cliff can melt small vertical runnels or larger channels that produce 'caves' at sea level (Powell, 1980; Gilbert, 1983; Syvitski, 1989). This process can enhance melt-out of debris and assists in creating calving embayments around the efflux (Sikonia and Post, 1980).

Melt-out from the face produces rockfall and debris grainfall. Dumps of supraglacial debris are produced from the surface and crevasse-fills during calving. Debris flows and slurries slide down the face but disaggregate due to the steep angle of the face before entering the sea (Powell, 1983a). All supraglacial contributions to grounding-line systems are minor except perhaps locally where glaciers have abundant supraglacial debris. Melt-out till similar to that from terrestrial glaciers is difficult to produce in these situations due to fast flow and non-stagnant conditions. Some melt-out deposits may form on the up-glacier side of morainal banks or where glacial 'toes' become buried by glaciomarine sediment.

Squeeze/Push. Sediment that is pushed by marine-ending glaciers is generally preserved by seasonal fluctuations of termini that are undergoing net retreat (e.g. Hoppe, 1959; Smith, 1982; Boulton, 1986b). Because marine-ending glaciers advance across sediment with a high water content, the probability of ejection out from a grounding line by squeeze and liquefaction processes is likely (e.g. Andrews and Smithson, 1966; Powell, 1984; Smith, 1990). Squeeze-up into basal crevasses is also strongly implied by the distinctive sea floor pattern observed after surges (Solheim and Pfirman, 1985) (Menzies, 1995b, Chapter 4).

Mass Movements. Sliding, slumping and sediment gravity flowage are common events at tidewater termini of temperate and subpolar glaciers. These processes are common because of high sedimentation rates, iceberg calving impacts and glacier-push of

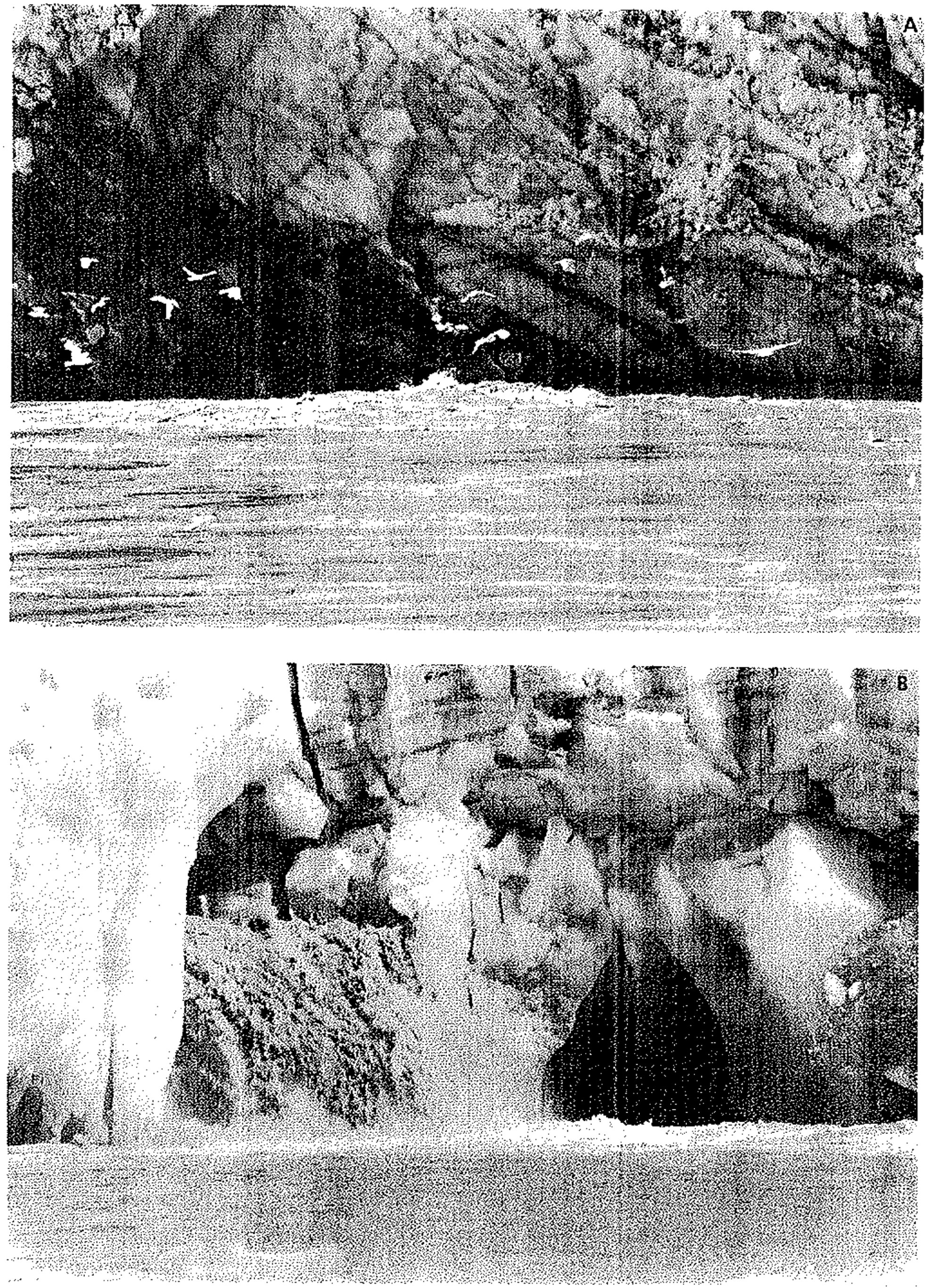


PLATE 14.3. Fluvial discharges from temperate tidewater termini in Glacier Bay, Alaska, which of all the glacial sources, contribute the largest volume of sediment to the sea. An upwelling at the sea surface is an expression of the meltwater jet discharging from under Grand Pacific Glacier at about 75 m depth. Gulls are feeding on shrimp that live at the ice face and which are being forced up to the surface by the upwelling (A). An example of a smaller, englacial stream discharging at sea level from Muir Glacier. The diameter of the conduit is about 4 m (B).

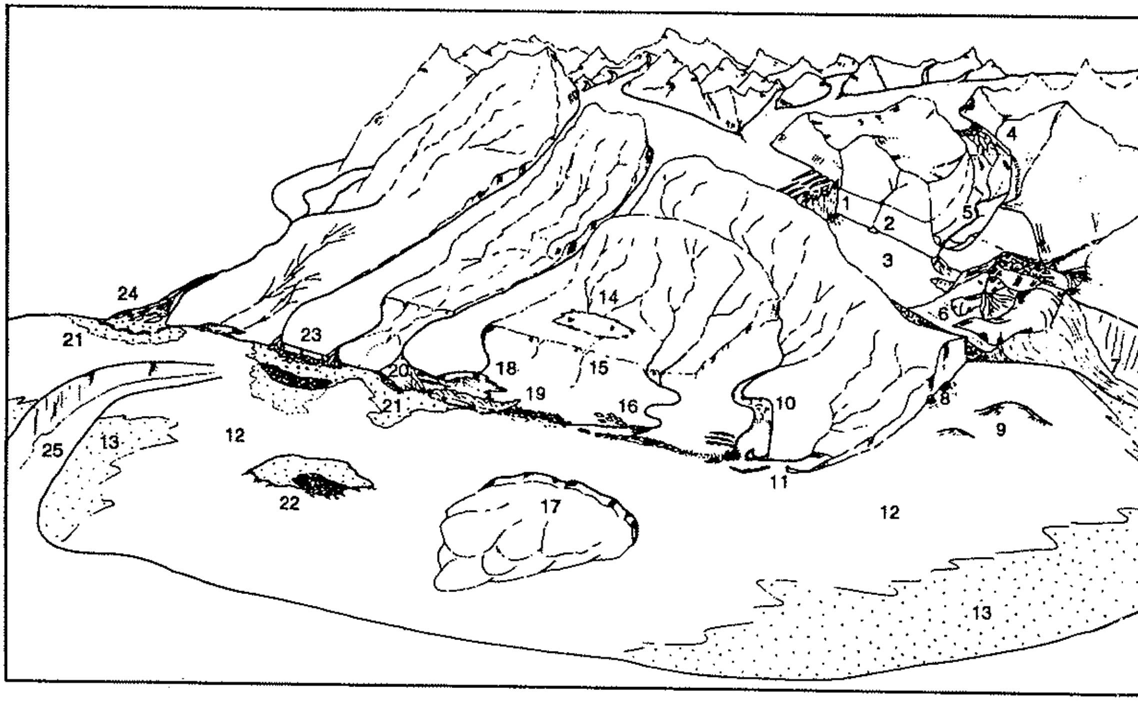


FIG. 14.4. Three-dimensional sketch showing the major environments for glaciomarine sedimentation in a cool-temperate climate, based on the present north-northwest Gulf of Alaska (not true orthographic scale). Numbered features are as follows: 1 — Tidewater terminus (see Figs 14.7a and b for details), 2 — Side drainage with gravel pocket beach and talus fan on fjord floor, 3 — Bergstone mud on glacial fjord floor, 4 — Sandur plain in marine outwash fjord, 5 — Sediment gravity flow channels in marine outwash mud, 6 — Entrance sill capped with morainal banks and grounding-line fan, 7 — Continuation of outwash mud, 8 — Rocky shore with gravel pocket beaches, 9 — Tectonically uplifted banks of older rocks exposed by winnowing on the shelf, 10 — Extensive tidal mudflat in estuary, 11 — Recurved and cuspate spits from alongshore transport, 12 — Marine outwash mud on the continental shelf, 13 — Relict sand deposits, 14 — Thick temperate rain forest of spruce/hemlock with muskeg swamps, 15 — Raised marine terrace, 16 — Eolian dunes fed from spit at wave-dominated delta, 17 — Large slide/slump areas on the continental shelf, 18 — Ice-dammed lake with sublacustrine grounding-line fan and lake laminites with iceberg-rafted debris, 19 — Terminal moraine and gravelly beach, 20 — Sandur/delta system with small moraines and gravelly beach, 21 — Modern littoral sand, 22 — Bank on continental shelf of relict glacier-marginal sediment, currently being winnowed, 23 — Tidewater terminus just at sea level with short-headed stream, fan deltas, 24 — River-dominated estuary with stunted barrier island system offshore, 25 — Sea valley on continental shelf being infilled with marine outwash mud (after Powell and Molnia, 1989; reprinted by permission of Elsevier Science Publishers BV, Amsterdam).

grounding-line sediment accumulations (grounding-line depositional systems) such as morainal banks, grounding-line fans and ice-contact deltas (Powell, 1981a, 1990). Small events with short (within 1–2 km) runout distances are common and may occur at intervals of days, hours or even continuously (Powell and Molnia, 1989; Powell, 1990; Phillips *et al.*, 1991). Larger events occur more sporadically but may remove large volumes of sediment from a grounding-line and transport it tens of kilometres away from the glacier (Powell, 1991, *unpublished data*).

Sediment accumulated against a grounding line can collapse back toward the glacier in small events as basal icebergs calve, as well as in quite large volumes when the grounding line retreats from the site (Powell, 1981a, 1991).

Hydrographic Processes. Waves, tides and ocean currents may affect melting of tidewater termini producing distinctive sediments. Melting of ice cliffs by currents flowing toward them and then upwelling to the surface is governed by the relationship:

$$M_i = 6.74 \quad 10^{-6} \quad U_w^{0.8} \quad \Delta T \quad L_{ic}^{-0.2} \qquad (14.1)$$

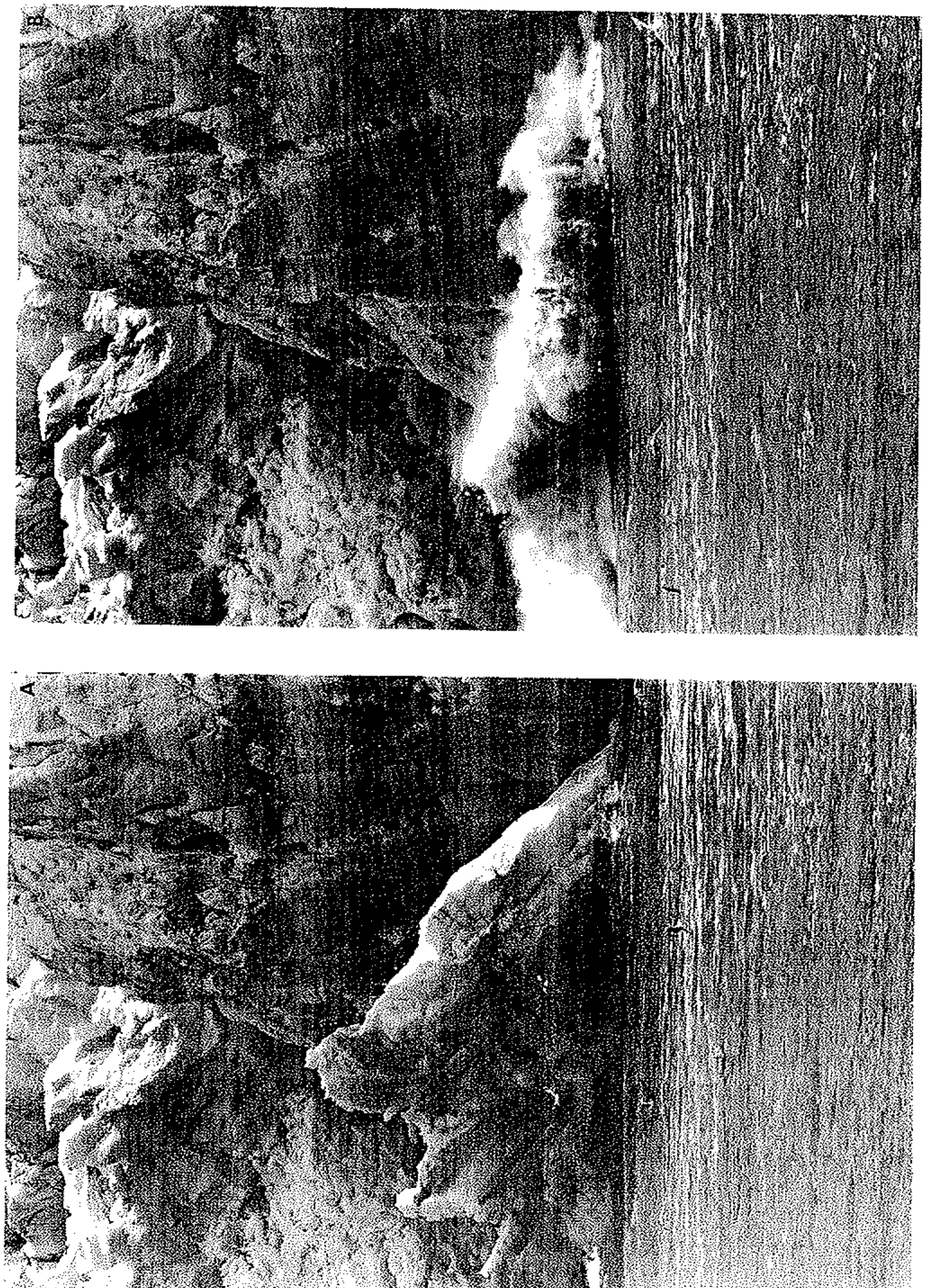


PLATE 14.4. A calving event from the grounded, 55 m high tidewater cliff of Lamplugh Glacier, Glacier Bay, Alaska. The slab is photographed about half way through its rotating fall (A) and then a large splash and wave is produced on impact (B).

where M_1 is the melting rate, U_w is water velocity, ΔT is temperature difference between ice and water and L_{ic} is the characteristic length of the face (Weeks and Campbell, 1973). When currents flow parallel with the face, mean melting rate (M_2) is:

$$M_2 = \frac{1}{h_{sw}} \int_0^{h_w} 3.7 \;\; 10^{-2} \;\; (T_\infty - T_s)^{1.5} \left(\frac{x}{k}\right)^{0.25} dx \quad (14.2)$$

where h_{sw} is height of submerged tidewater cliff wall, T_∞ is far field ambient water temperature, T_s is surface water temperature, x is the coordinate in water flow direction and k is a scaling factor (Huppert, 1980; Huppert and Joseberger, 1980). In general, $M_2 < M_1$.

Commonly, waves, tides and ocean currents are second-order controls on grounding-line sedimentation except where large storm events occur on the continental shelf, or where the volume of sediment supplied from the glacier is low as in polar settings, or in shallow water where the processes can rework much of the sediment.

Fluvial Discharges. Fluvial discharges from tidewater termini are greatest from temperate glaciers. Such discharges dominate sediment production in temperate, and to a lesser extent subpolar, glaciers. Streams enter the sea from englacial and subglacial conduits; the latter position providing most of the sediment. Discharge is so high in the melt season that it forms turbulent jets where momentum forces in the flow are stronger than the buoyancy forces of the freshwater entering denser saltwater (Plate 14.5)(Syvitski, 1989).

For slower flows such as during early rising and late falling discharges or from colder glaciers, the effluent is a forced plume where buoyancy dominates momentum forces. That is, when

$$R_n = Q \, B_f^{0.5} \, M_f^{-1.25} \leq 0.56 \text{ flow is a jet,} \quad (14.3)$$

$$> 0.56 \text{ flow is a plume;} \quad (14.4)$$

where R_n is the Richardson number, Q is discharge or volume flux, B_f is buoyancy flux equivalent to $g'Q$ and M_f is momentum flux or $\rho Q U_{mo}$; where $g' = g(\rho_{ad} - \rho_o)/\rho_o$, ρ_{ad} is ambient fluid density, ρ_o the initial discharge density and U_{mo} the initial near-centreline velocity (List, 1982; Powell, 1990). The angle at which the jet issues into the sea can be determined by the densimetric Froude number (F)

$$F = U_m^2 (g' D_c)^{-1} \quad (14.5)$$

where U_m is mean centreline velocity at the conduit mouth and D_c is conduit diameter (Abraham, 1965). As $|F| \rightarrow 0$ the jet is vertical, as $|F| \rightarrow \infty$ the jet follows the axis of the conduit and as $F < 0$ when $\rho_{ad} < \rho_o$, a continuous underflow may form (Syvitski, 1989). Underflows have been inferred (e.g. Mackiewicz *et al.*, 1984) and may be enhanced by hyperconcentrating sediment as particles rapidly fall out of transport when the flow detaches from the sea floor (Powell, 1990), but these remain unconfirmed.

Eventually the jet transforms to a plume as it rises to sea level. Occasionally the flow may cause 'boiling' or, at slightly higher discharges, 'fountaining' at sea level above the upwelling (Plate 14.3). The plume spreads as a barotropic flow away from the discharge at a level of neutral buoyancy, which in sea water is commonly at sea level, that is, it forms an overflow or hypopycnal flow. It is from this jet/plume system that laminated sediments (cyclopsams and cyclopels) can form (Mackiewicz *et al.*, 1984; Cowan and Powell, 1990).

14.3.2.3. Calving styles and processes

Tidewater termini differ from floating termini in that the calving line coincides with the grounding-line. Calving from ice cliffs occurs by three processes: (1) spalling of seracs by fracturing above sea level and the iceberg falls into the sea, (2) large sheets shear off the glacier face and the iceberg sinks vertically or topples forward into the sea (Plate 14.4), and (3) detachment below sea level and then the iceberg rises to the surface. While Muir Glacier, Glacier Bay, Alaska was retreating at about 2 km year^{-1}, about 10^6 m^3 day^{-1} of ice was calved (Powell, 1983b). For a fast flowing (3 km year^{-1}) glacier with a stationary tidewater terminus 2 km wide ending in about 100 m of water, about 1.7×10^6 m^3 day^{-1} of ice must be calved.

Subaerial calving occurs by fracture propagation (Iken, 1977; Hughes, 1992b) and if icebergs do not shatter on impact they can travel down through the seawater column. Some blocks can either impact

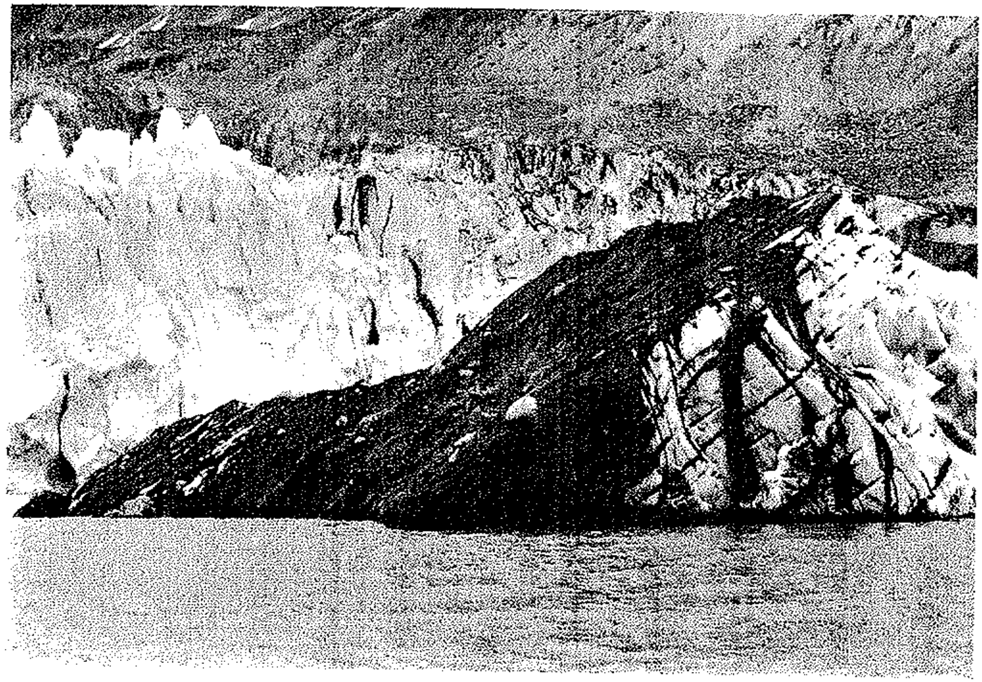


PLATE 14.5. A berg produced by a submarine calving from the McBride Glacier tidewater cliff in the background, Glacier Bay, Alaska. These types of bergs provide excellent exposures of sections of glaciers otherwise unobtainable. Here, the sole of the glacier with subglacial till is exposed on the left side of the berg and all the basal debris layers to the right are in a stratigraphic-up sense. All of this debris will produce iceberg-rafted debris.

upon the sea floor, or come sufficiently close, so that their preceding pressure wave redistributes sediment (Powell, 1985). Surface waves caused by the impact generally can not influence the sea floor initially because it is too deep (Powell, 1980); however, at shorelines the waves can cause erosion and transport icebergs above the high tide line (Syvitski, 1984).

Icebergs from submarine calving are often termed bergy bits and commonly contain englacial and basal debris (Powell and Molnia, 1989). These icebergs can disturb bottom sediment as they detach, and some are seen with sea floor sediment on top of them. They are also major sources of iceberg-rafted debris (IBRD) from tidewater termini (Gottler and Powell, 1990) (Plate 14.5).

14.4. TYPES OF GROUNDING-LINES

Grounding-line systems are defined as sedimentary depositional systems (*sensu* Fisher and Brown, 1972) formed at grounding-lines of glaciers ending in large bodies of water, such as large lakes or seas (Powell, 1988). Each system is made from one or several sedimentary environments and can be distinguished by its processes of formation, geometry, facies associations and internal architecture.

Recent studies of glaciomarine sediments have shown there are several different varieties of grounding-line systems that include: (1) grounding-line fans (Powell, 1990) and glacier-contact deltas (Thomas, 1984a; McCabe and Eyles, 1988), (2) morainal banks (Powell, 1981), and (3) grounding-line wedges ('till deltas' of Alley *et al.*, 1987b) (Figs 14.5, 14.6, 14.7 and Plate 14.6). Understanding how to recognize each type by lithofacies analysis and being able to infer processes of formation are important for inferring the potential stability of grounding-lines (cf. Powell, 1991) and for interpreting paleoglaciological and paleoclimatological conditions from a sedimentary record.

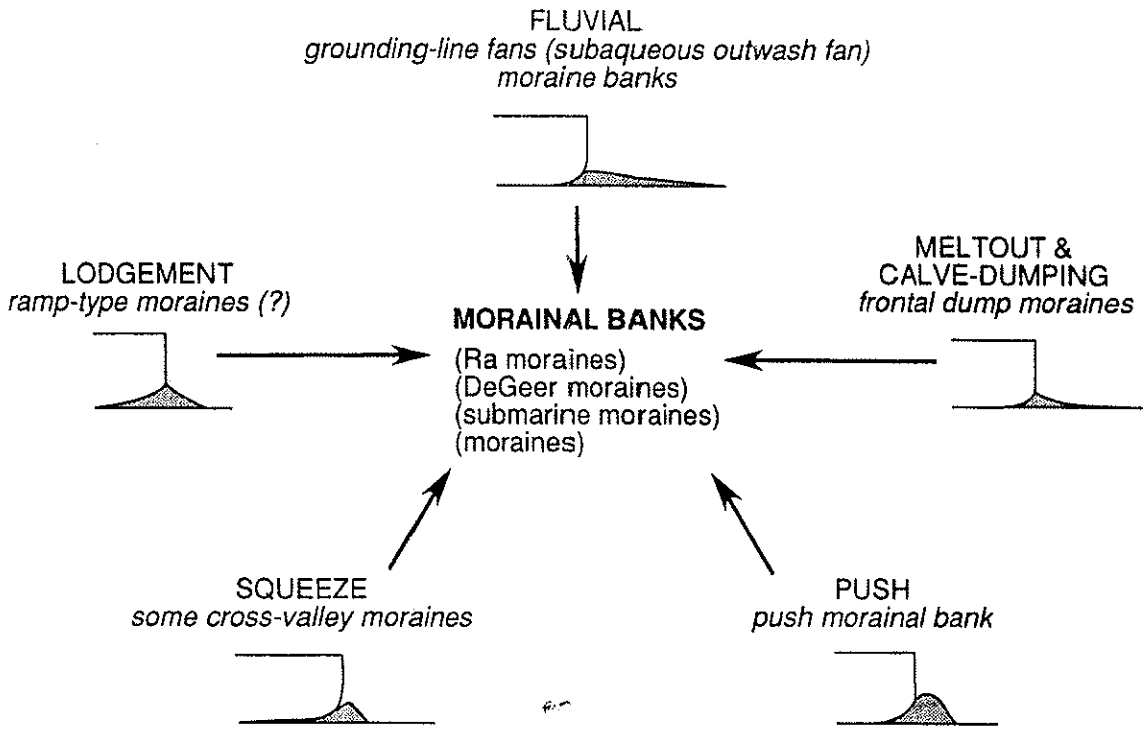


FIG. 14.5. End-member processes that contribute to forming grounding-line systems in the form of morainal banks (not to scale). Beneath each process are terms that have been used to describe the deposit in the literature; here it is recognized that morainal banks can form by the end-member process or any combination of those processes. Alternative terms for morainal banks in the literature are also listed.

Grounding-line fans are point-source depocentres with a fan geometry (*sensu* Mitchum *et al.*, 1977), formed at conduit mouths where meltwater streams discharge subaquatically (Fig. 14.6). They are composed of subaquatic outwash (deposited from the discharge while in contact with the floor of the water body) and a variety of sediment gravity flow and suspension settling deposits. They have also been named transverse eskers and deltas (Glückert, 1975, 1977), beaded eskers (Banerjee and McDonald, 1975), subaqueous outwash fans (Rust and Romanelli, 1975), subaqueous esker deltas (Thomas, 1984b) and glacier-contact fans (Boulton, 1986b) (Menzies, 1995b, Chapter 2).

Glacier-contact deltas are commonly formed by short-headed streams, (cf. Flores, 1975) and are typified by processes, facies associations and the architectural arrangements of fan deltas with glacial influences such as ice-rafting and pushing (cf. Syvitski and Skei, 1983; McCabe and Eyles, 1988; Nemec and Steel, 1988). These fans often develop from grounding-line fans building up to sea level (Glückert, 1975, 1977; Powell, 1990).

Delta processes are grouped into bedload dumping, hemipelagic sedimentation, by-pass and diffusion of sediment (Syvitski *et al.*, 1988). As a river enters the sea its competence is dramatically reduced and bedload is deposited and dumped rapidly near the river mouth. The suspended load can continue into the sea in a hypopycnal flow of fresher, less dense, water flowing over saline water. Particles suspended in this plume eventually settle out and deposit hemipelagic sediment (Section 3.5.1). Meanwhile, some processes on the sea floor act to remove sediment totally from the nearshore delta area. Such processes as low density turbidity currents and large-scale slides behave in this manner. Other bottom processes act to 'smear' bottom sediment

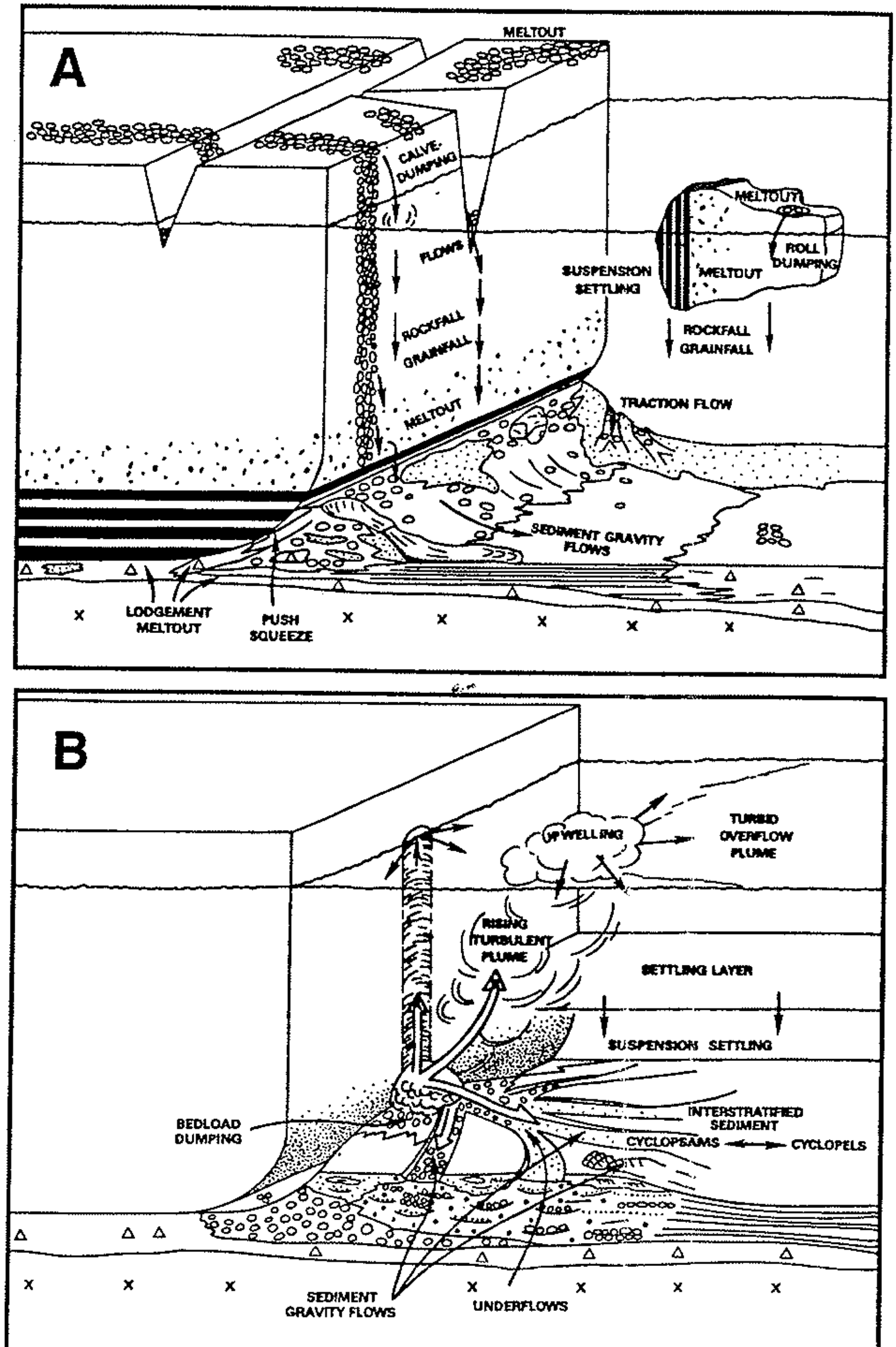


FIG. 14.6. Processes and lithofacies associations contributing to grounding-line systems at tidewater termini of temperate glaciers (not to scale). (A) A morainal bank, (B) submarine outwash of a grounding-line fan and generation of cyclopsams and cyclopels (after Powell and Molnia, 1989; reprinted by permission of Elsevier Science Publishers BV, Amsterdam).

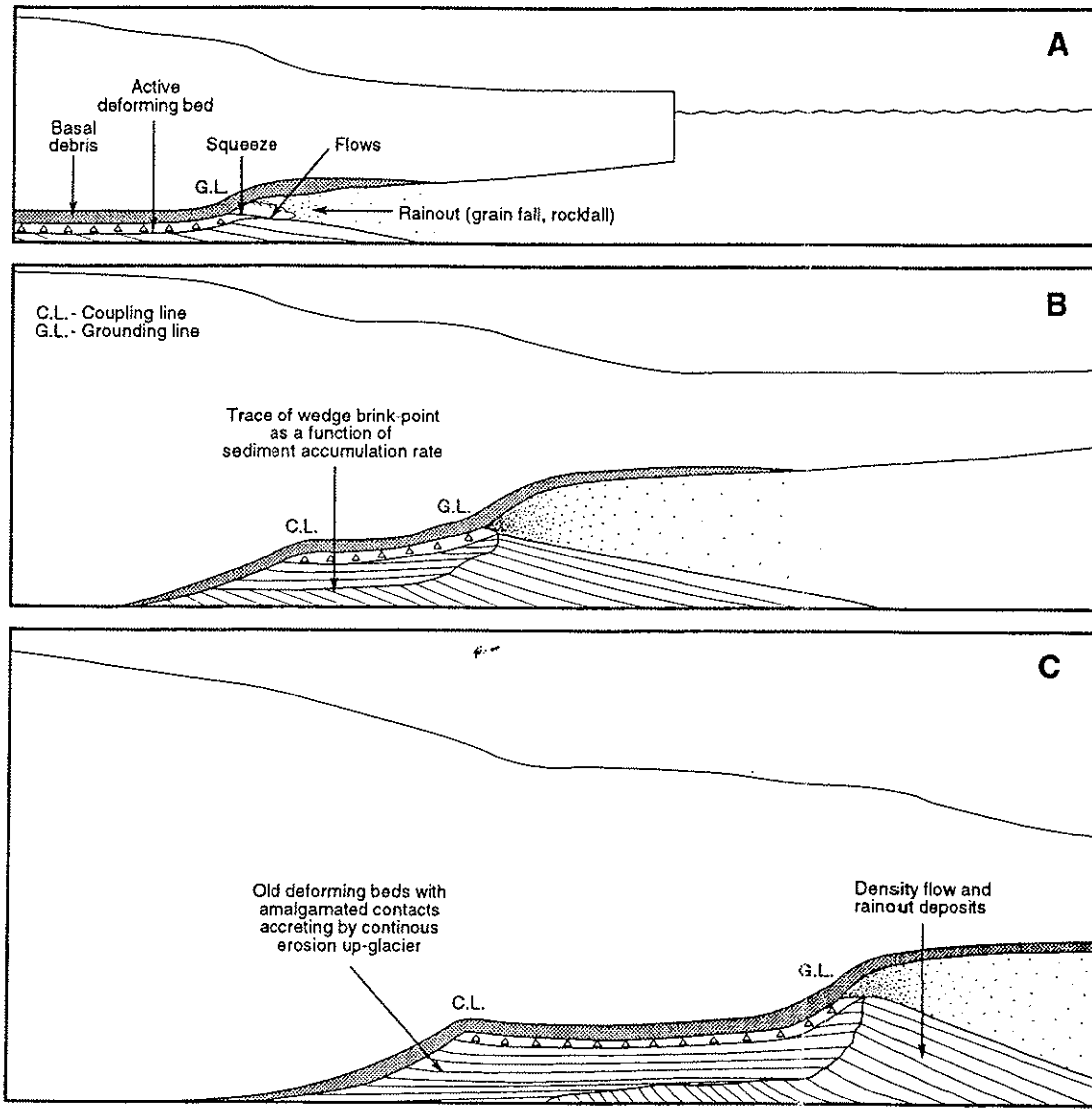


FIG. 14.7. A possible mechanism for formation of grounding-line wedges and how they can expand with glacial advance to form sediment sheets that terminate as wedges. An ice stream feeding an ice shelf has basal debris that is feeding a subglacial deforming bed (A). At the grounding line the deforming bed is squeezed out to form sediment gravity flows on an inclined surface. The wedge of sediment produced is also contributed to by grain fall and rock fall from under-melt of the ice shelf. The volume of that rain-out sediment decreases with distance from the grounding line and also contributes to the thinning of sediment distally. As the sediment builds to meet the base of the ice shelf, the grounding line advances as does the coupling line (B). Upglacier from the coupling line, erosion can occur. Between the coupling line and the grounding line, the deforming bed aggrades by being added to from basal debris, but because transmitted sheer decreases with depth lower sediment in the deforming bed is deposited. With continued advance, the distance between the grounding line and coupling line increases producing a deposit with a sheet-wedge geometry (C). Predictably, the deposit is composed of sediment gravity flow and rain-out facies topped by deformed-bed deposit with amalgamated contacts (not to scale).

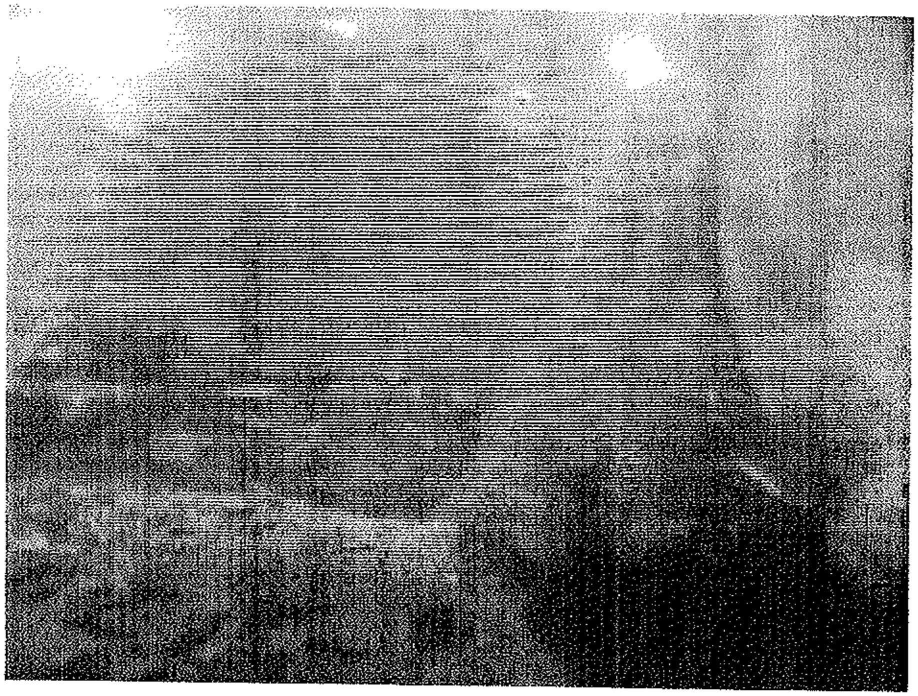


PLATE 14.6. View of a grounding line at a tidewater terminus of a temperate Alaskan glacier. The bottom left-hand segment is sorted glaciomarine sediment of gravels up to boulder size draped with a coating of mud from meltwater streams and pebbles from meltout of icebergs and the glacier cliff. The rest of the photograph is a wall of clear glacier ice with bubbles and pebbles within. This was the first observation made of a grounding line, which in the photograph is about 15 cm long.

away into deeper water, such as tidal or wave action, creeps or slumps and perhaps hyperconcentrated sediment underflows from the river. By using simple density contrasts it is easily shown that underflows directly from rivers can occur only if their sediment concentrations are higher than about 30 g l^{-1} (Gilbert, 1982; Powell, 1983).

Morainal banks are made of various glacier-contact lithofacies (Figs 14.5, 14.6 and 14.7) that include chaotic mixtures of diamicton, gravel, rubble, sand and mud facies formed by rock and grain fall processes due to calve-dumping and melt-out (Powell, 1981; Powell and Molnia, 1989), lodgement (Holtedahl, 1960; Glückert, 1975, 1977; Smith, 1982) and squeeze/push processes from under the glacier (e.g. Andrews, 1963; Smith, 1990).

The deposits have a bank geometry (following the terminology of Mitchum *et al.*, 1977) and are analogous to terrestrial end moraines, but are formed subaquatically. Banks created mainly by pushing are push-morainal banks (cf. Powell, 1981; Smith, 1982; Eyles and Eyles, 1984; Boulton, 1986b). Morainal banks also include the ramp-type sublacustrine moraines that form more by subglacial lodgement processes (Holdsworth, 1973; Barnett and Holdsworth, 1974), frontal-dump moraines from melt-out at the terminus and dumping of supraglacial debris during the calving process (Powell, 1981; Syvitski and Praeg, 1989). Other banks are thought to form by squeezing out of subglacial sediment beyond the grounding-line to produce landforms like the cross-valley moraines (Andrews and Smithson, 1966) (cf. Chapter 10). Other forms of morainal banks may have their entire lengths made of facies similar to grounding-line fans (moraine banks) (Leavitt and Perkins, 1935), indicating subglacial discharge of meltwater in either sheet flows or rapidly migrating conduits across a grounding-line. These types of banks confirm a continuum from lone grounding-line fans (that can form beaded eskers) (Banerjee and

McDonald, 1975), to interstratified grounding-line fan and glacier-contact facies to solely glacier-contact facies.

Morainal bank forms are documented in modern process studies in lakes and the sea (e.g. Holdsworth, 1973; Barnett and Holdsworth, 1974; Powell, 1981; 1990; Powell and Molnia, 1989). They are described from older deposits in fjords (Vorren, 1973; Elverhøi *et al.*, 1983; Gilbert, 1985; Syvitski, 1989; Holtedahl, 1989) and continental shelves (Andersen, 1968; Blake, 1977; Matishov, 1978; Wong and Christoffel, 1981; Knebel and Scanlon, 1985; Oldale, 1985; Barnes, 1987; Josenhans *et al.*, 1988; Anderson and Bartek, 1991) and from Pleistocene deposits presently exposed on land (e.g. McCabe, 1986; Ingólfsson, 1987). The latter types have been variously termed moraine banks (Leavitt and Perkins, 1935), DeGeer moraines (e.g. Hoppe, 1959; Smith, 1982), washboard moraines (Elson, 1969), some of the cross-valley moraines of Andrews and Smithson (1966), Ra moraines (e.g. Holtedahl, 1960), submarine moraines or simply, moraines (terminal, end, lateral) (e.g. McCabe *et al.*, 1984; Oldale, 1985; Holtedahl, 1989).

'Till deltas' were defined as wedges of sediment made up of subglacial till overlying dipping strata inferred to be mainly sediment gravity flow deposits (Bentley *et al.*, 1988; Alley *et al.*, 1989b; Blankenship *et al.*, 1989; Kamb and Engelhardt, 1989; Scherer, 1991). They are thought to be forming today beneath the Ross Ice Shelf based on seismic evidence and theoretical considerations. Here the term 'grounding-line wedge' is used rather than 'till delta' to avoid the problem of defining all of the sediment as till, and to avoid an association with sea level that could be inferred from the term 'delta' (Powell and Alley, *in preparation*).

Grounding-line wedges are thought to form mainly from deforming subglacial till that reaches the grounding-line (Fig. 14.7). There, sediment redistribution by gravity-flow processes produces deposits with dips as low as 1° or less, because the deforming till should contain abundant fine-grained matrix (cf. Mitchum *et al.*, 1977; Sangree and Widmier, 1977; Lawson, 1979a, p. 41; Hampton *et al.*, 1987). Grounding-line advance across these deposits will cause deforming subglacial till to overlie them disconformably. The up-glacier limit of this disconformity is called the coupling line and coincides with a change in surface slope of the glacier, up-flow from the grounding-line. The deforming layer may thicken downglacier across the coupling line, causing the coupling line to be dynamically significant to the glacier. Grounding-line wedges are less well known than the morainal bank varieties because the one modern example so far postulated (Alley *et al.*, 1987b) is still under investigation (Blankenship *et al.*, 1989; Engelhardt *et al.*, 1990). However, an older deposit on the Antarctic continental shelf has the possibility of being a relict Pleistocene example (cf. Karl *et al.*, 1987, Fig.4).

Dreimanis (1987) identified an older, smaller-scale deposit that may be equivalent to grounding-line wedges. This feature, which he termed a "subaquatic till tongue", occurs in Wisconsinan deposits at the base of a cliff bordering Lake Erie, Ontario. It is described as "a diamicton tongue interpreted as...till...which was deposited, partly by lodgement, under the edge of the Erie lobe, which terminated in a deep proglacial lake, and partly by extrusion and gravity flows of dilated lodgement till..." (Dreimanis, 1987, p. 23).

Powell and Alley (*in preparation*) suggest that larger-scale deposits independently termed 'till tongues' by King and Fader (1986) also may be older deposits equivalent to the 'till delta' of Alley *et al.* (1989b) and the 'subaquatic till tongue' of Dreimanis (1987). The 'till tongues' of King and Fader (1986) were defined as wedge-shaped deposits of till interbedded with stratified glacimarine sediment (King *et al.*, 1987). They were mainly described using high-resolution seismic stratigraphy and were attributed to deposition at a fluctuating grounding-line of an ice shelf, by different processes than attributed here to grounding-line wedges.

The apparent major control over the two end-member types of grounding-line systems, morainal banks and grounding-line wedges, is subglacial meltwater. With abundant, free-flowing water morainal banks appear to form; in contrast, grounding-line wedges appear to form with smaller volumes of more confined water and deforming beds.

Using these simplistic limitations, Powell and Alley (*in preparation*) suggest that deposits such as 'till tongues' may be formed mainly by polar or subpolar glaciers flowing on deforming beds; the glaciers may or may not end as floating termini. Alternatively, morainal banks are most likely formed by temperate or subpolar glaciers where supraglacial water contributes to subglacial water in channelized or sheet flows, and the glacial termini are most commonly tidewater cliffs.

14.5. PROGLACIAL AND PARAGLACIAL ENVIRONMENTS

Proglacial and paraglacial environments lie beyond the glacier proximal settings of grounding-line systems and subglacial ice shelf systems. The term paraglacial was used by Church and Ryder (1972) to define non-glacial processes that are directly conditioned by glaciation and it is used here to include marine systems that receive glacigenic sediment but are not in direct contact with glaciers nor icebergs (Gilbert, 1983; Powell and Molnia, 1989). In many instances processes in proglacial and paraglacial settings are similar. However, the presence of glacial ice sometimes creates distinctive processes in proglacial settings.

14.5.1. Hemipelagic Suspension Settling

Sediment plumes from subglacial and marginal streams are the transportation agents for most glaciomarine sediment in front of temperate and subpolar marine-ending glaciers. The overflow originating from fresh water sources at tidewater termini has been modelled as a barotropic flow originating as a buoyant jet (Calabrese and Syvitski, 1987, Figs 1 and 2). Velocities in different zones are:

for plug flow:

$$U_{x,y} = (Q_o b^{-1}_o H^{-1}_o) \exp[-(y + 0.1x - 0.5 b_o)^2 (2\sigma^2)^{-1}] \quad \text{for } x \leq 5.2 b_o,\ y > b_o - 0.2x \qquad (14.6)$$

for established flow:

$$U_{x,y} = 2.28 (Q_o b^{-1}_o H^{-1}_o)(b_o x^{-1})^{0.5} \exp[-y^2 (2\sigma^2)^{-1}] \quad \text{for } 5.2 b_o < x \leq x_b \qquad (14.7)$$

for constrained flow:

$$U_{x,y} = 2.28 (Q_o b^{-1}_o H^{-1}_o)(b_o x^{-1}_o)_{0.5} \exp[-y^2 (2\sigma^2)^{-1}] \quad \text{for } x > x_b \qquad (14.8)$$

where $U_{x,y}$ is longitudinal velocity, b_o is stream channel width, H_o is channel depth at the stream mouth, Q_o is stream discharge, x is down-fjord distance, y is fjord width distance and $\sigma = 0.108x$ is the standard deviation of the Gaussian distribution for velocity across the plume width.

Sedimentation rates ($Z_h(x,y_o)$) beneath the respective parts of the plume are predicted by the relationships (Syvitski *et al.*, 1988):

$$Z_h(x,y_o) = Z_o \exp[-(\lambda_{rs} U^{-1}_o)x] \quad \text{for } x \leq 5.2 b_o \qquad (14.9)$$

$$Z_h(x,y_o) = Z_o \exp\{-\lambda_{rs} [(1.76 b_o U^{-1}_o) + 0.29 x^{1.5} (U_o b_o 0.5)^{-1}]\} \quad \text{for } 5.2 b_o < x \leq x_b \qquad (14.10)$$

$$Z_h(x,y_o) = Z_o \exp\{-\lambda_{rs} [(1.76 b_o U^{-1}_o) - (0.15 x_b^{1.5} U^{-1}_o b_o^{-0.5})]\} \quad \text{for } x > x_b \qquad (14.11)$$

where $Z_o = \lambda_{rs} Q_s (U_o b_o)^{-1}$ is the rate of sedimentation at the stream mouth, λ_{rs} is the removal rate constant for the suspended particles, x_b is the distance from the stream mouth where the plume becomes constrained by basin geometry, U_o is the longitudinal velocity component at the stream mouth and Q_s is the suspended load carried by the stream. Under these conditions Calabrese and Syvitski (1987) predicted an exponential decrease in settling rates down flow from a submarine discharge as was later documented by Cowan and Powell (1991) in Alaska.

Proximal marine outwash occurs as a traction facies deposited along the run-out distance of the jet from the submarine efflux of a subglacial stream (Fig. 14.8). Sometimes the jet detachment will be at the glacier face but with larger discharges the zone may be away from the face. At the detachment zone small pebbles to sand sizes appear to fall out and are virtually 'dumped' (Powell, 1990). Particles as coarse as medium sand can remain in the plume as it rises to the surface and sand can be transported by the plume as much as 1 km from a face (Cowan *et al.*, 1988).

Sediment is released from overflow plumes episodically and the areal position of a plume changes due to tides and surface wind sheer (Fig. 14.6). Cowan and Powell (1990) suggest that particles are

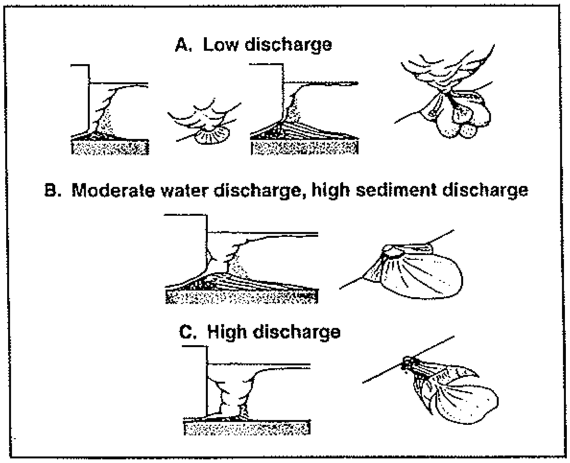


FIG. 14.8. Alternative fan types with different discharges (not to scale). (A) At low discharges the fan grows progressively with coarse traction deposits dumped at the efflux and most deposits from settling or slumping/sliding and sediment gravity flows. (B) At higher sediment discharges the horizontal jet remains in traction to produce rapid sedimentation deposits in sheets or crude scour-and-fill structures. Settling and sediment gravity flow deposits make up more distal deposits. (C) At high discharge the horizontal jet can include a migrating barchanoid bar at the detachment zone (after Powell, 1990; reprinted with permission of the Geological Society of London).

released from a plume during ebb tidal stage when internal turbulence within the plume is smallest. The released particles then fall as a turbid layer through the water column. Coarser particles of sand/silt fall faster than flocs of mud and the result is laminites. Another control on the paths of plumes is the coriolis effect (Gilbert, 1983) that can lead to an increase in sediment accumulation rates at particular sites by 2 to 3 times in high latitude fjords (Syvitski, 1989). Oceanic currents such as those moving along continental shelves are important for mud transport (Carlson *et al.*, 1975), as are eddies both in distal areas and near a glacier face.

Settling of finer particles from the plume can be enhanced by three processes: flocculation, agglomeration and pelletization (Syvitski *et al.*, 1987). Flocculation occurs by electrostatic binding of particles in brackish solutions (salinity > 3‰); agglomeration occurs due to attachment of mineral grains to organic detritus; and pelletization is the result of zooplankton egestation of particles as fecal pellets. These particle accumulations apparently fall more slowly than sand/silt grains once released from an overflow to allow coarse and fine grained laminae to be produced (Cowan, 1988).

The rising plumes from submarine discharges can have sufficient sediment concentration that they reach a level of neutral buoyancy within the brackish water column and form an interflow (Cowan *et al.*, 1988). Increased sediment concentration may originate from rainstorm events or sudden release of sediment stored subglacially by channel migration if there is a change of head or gradient such as during submarine iceberg calving. Interflows have also been documented from delta fronts in subpolar regimes (Gilbert, 1982). Cold tongues as described from three Antarctic fjords may represent a type of interflow which forms adjacent to subglacial cavities (Domack and Williams, 1990). These processes transport significant quantities of siliciclastic particulates in suspension which may contribute to laminites within inner fjord basins (Domack, 1990) (Plate 14.6).

During winter, these discharges can shut down completely, especially from subpolar glaciers. In cool-temperate climates subglacial and some marginal streams can keep flowing, though with decreased discharge and sediment concentrations. The resulting fjord water is 45 to 170% more saline in winter and has about 96% lower sediment concentrations (Cowan, 1988).

14.5.2. Biological Production

Biological activity adds primary sedimentary particles to glaciomarine environments and they can be a major component of glaciomarine sediments where siliciclastic production is low. Organisms also speed settling rates of siliciclastic particles by pelletization and mix bottom sediment by bioturbation.

An important aspect of polar glaciomarine environments is the abundance of biogenic facies. The primary reason for this facies, particularly in the ice-proximal environment, is the limited amount of siliciclastic sediment supply to the sea. High supply rates of siliciclastic sediment dilutes any biogenic sediment produced in warm glaciomarine settings; but by contrast, bioclastic carbonate and siliceous ooze are

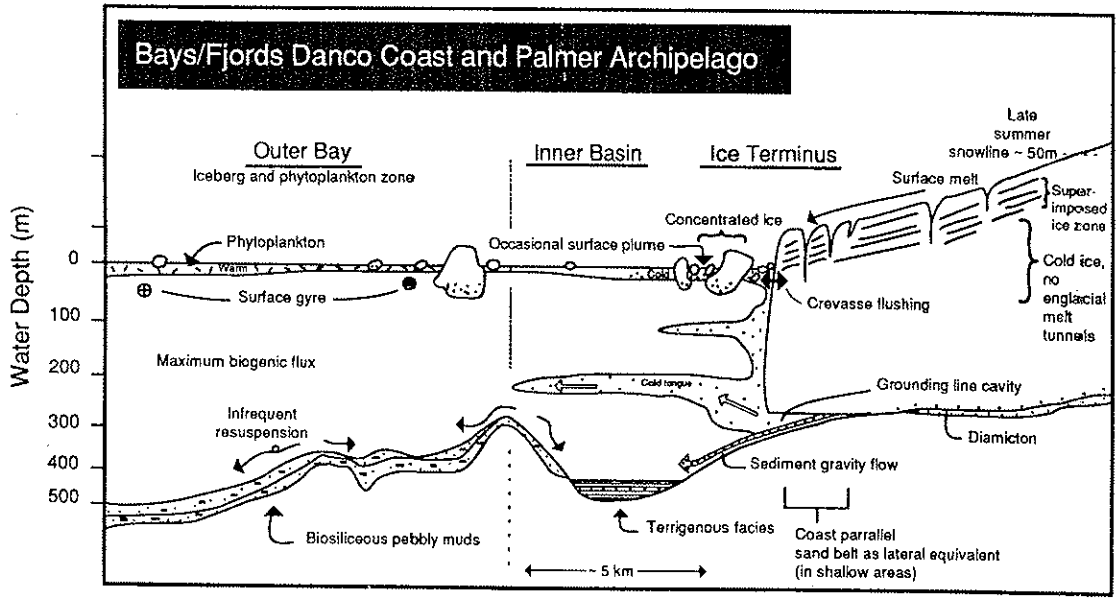


FIG. 14.9. Facies relationships representative of fjords found along the northwestern side of the Antarctic Peninsula, Gerlache Strait. The dominant source of siliciclastic sediment includes, ice-rafted debris, cold tongues (<–0.5°C), and sediment gravity flows. All of these processes are most active in ice-proximal settings near the terminus and within the inner basins. The absence of abundant meltwater prevents the formation of surface overflows and hence, restricts siliciclastic sedimentation to within the inner basins. Phytoplankton (diatoms) are most productive within the warm (>0°C) surface waters of the outer bays where iceberg and sea-ice rafting also takes place. These two processes lead to biosiliceous pebbly muds within the outer basins.

the dominant facies found in Antarctica today. Bioclastic carbonates occur as shell coquina which fringes tidewater cliff margins of the East Antarctic Ice Sheet and as localized deposits of coquina on shelf banks, such as those found in the Ross Sea (Tavianni *et al.*, 1993) and near Prydz Bay (Quilty, 1985). These sediments contain bryozoans barnacle plates, echinoids, a variety of molluscs and abundant and diverse calcareous foraminifera fauna. In more restricted settings, such as fjords along the Ingrid Christianson Coast, carbonate bioherms are actually forming at depths of 10 to 50 m. Here the largest tube worm (*Surpula narconensis*) reefs in the world are forming (Kirkwood and Burton, 1988). Absence of siliciclastic loading and a high rate of biogenic input into the fjord system were considered the key variables in allowing these carbonate reefs to form less than 10 km from the world's largest ice sheet. Other favourable conditions are a restricted entrance and strong tidal currents in Ellis Fjord that prevent icebergs and anchor ice from scouring the sea floor.

Biogenic siliceous muds and oozes are the dominant deposit within deep water basins along the Antarctic continental shelf, including deep fjords along the Peninsula (Fig. 14.9). At depths >500 m, icebergs are prevented from scouring biogenic accumulations and mixing the sea floor sediment and, as a result, diatomaceous particulates accumulate from vertical settling and dilute, sediment gravity-flow deposition. Both processes occur but the primary one is settling of phytoplankton from the pulse of productivity during austral summers. The production of organic carbon and opaline silica by diatoms in Antarctic waters has been studied for a number of years (Smith and Sakshaug, 1990). The present consensus is that productivity is highly variable both spatially and temporally. Factors which lead to higher productivity, however, are not thought to be linked to nutrient supply. Instead, physical stability of the surface layer which is formed as sea ice melts in the austral spring, is the leading variable in controlling production. Where winds and waves are at a minimum as the sea ice is melting, high rates of primary production occur (962 mgC m^{-2} day^{-1}) (Smith and Sakshaug, 1990). Hence, production is usually greatest in nearshore settings, particularly fjord or coastal embayments along the Antarctic Peninsula (3.62 gC m^{-2} day^{-1}) (Burkholder and Mandelli, 1965). The transport of this material to the bottom ultimately depends upon a number of steps as discussed by Dunbar *et al.* (1989). Oxidation and bacterial consumption takes place within the water column and reduces the C/Si ratio of suspended particulate matter (SPM) (Dunbar *et al.*, 1989). Usually resuspension of SPM takes place following deposition and this may be stimulated by periodic increases in bottom current intensity, such as occur during the winter in McMurdo Sound and the Ross Sea. Direct settling to the bottom is also important as shown by the preservation of depositional laminae of seasonal character from basins surrounding the East Antarctic margin (Domack, 1988, Fig. 4a).

In Polar-Tundra, Boreal and Cool-Temperate climates, aquatic biological activity is high, but successions are dominated by siliciclastic sediment due to higher sediment yields from glaciers. Primary productivity is not much different than in Antarctica (Smith and Sakshaug, 1990) and it is thought to be high due to high nutrient levels and well-oxygenated waters at tidewater termini caused by upwelling of deep water (cf. Apollonio, 1973). Annual or biannual phytoplankton blooms occur in fjords in Spitsbergen and Alaska and may produce annual layers in siliciclastic sequences (Sharma, 1979; Elverhøi *et al.*, 1980). In the Arctic and Boreal fjords of Baffin Island and Spitsbergen, a significant proportion of sediment transported to the sea is biogenic; and there is also biological productivity within the fjords. This organic detritus is a major energy source for benthic fauna which, in turn, play an important role in mixing, sorting and binding sediment particles, and in determining the sedimentary chemical environment (Syvitski *et al.*, 1987). Fjords open to the ocean can receive influxes of planktic forms (Elverhøi *et al.*, 1980). However, faunal diversity is commonly low (e.g. Elverhøi *et al.*, 1980) and total productivity is generally controlled by phytoplankton production and subsequent grazing and predation (Syvitski *et al.*, 1987). Phytoplankton blooms commonly occur during spring, but they are delayed until later summer at higher latitudes.

Phytoplankton are generally the largest source of particulate organic carbon in these fjords. Phytoplankton commonly fall to the sea floor in aggregates called 'marine snow' or as fecal pellets from grazing zooplankton (Syvitski *et al.*, 1987). Much of this organic carbon is consumed on its way to the sea floor, but organic carbon commonly comprises most of the carbon in bottom sediment (Syvitski *et al.*, 1990). However, due to high siliciclastic sedimentation, organic carbon is less than 0.1% of bottom sediment near tidewater termini and increases in proportion down-fjord (Syvitski *et al.*, 1990).

Although productivity is high in the Cool, Temperate Gulf of Alaska, population size is controlled by grazing of macro-zooplankton (Sambrotto and Lorenzen, 1987). Zooplankton, in biannual blooms, are uniformly distributed along the Gulf of Alaska due to strong coastal currents and the population is dominated by copepods (Cooney, 1987). Although biological diversity is low, close to Temperate tidewater termini, biological activity is high even though siliclastic sediment accumulation rates are >10 m $year^{-1}$ (Simenstad and Powell, 1990). However, predicting a fossil record from modern settings for biofacies analysis is still not possible because of lack of data. The foraminiferal record is not well known but studies indicate relationships of benthic ecology of foraminifera and molluscs to substrate character and water depth in Alaska (Echols and Armentrout, 1980; Hickman and Nesbitt, 1980; Quinterno *et al.*, 1980; Lagoe *et al.*, 1989).

Macrofaunal communities in cool-temperate to polar-tundra climates are dominated by infaunal populations in soup- and soft-grounds and epifaunal communities on firm- and hard-grounds (e.g. Dale *et al.*, 1989; Aitken, 1990). Common fauna include polychaetes, pelecypods, gastropods, pectens, octocorals, bryozoans, hydrozoans, tunicates, porifera, echinoderms and fucoid algae (Hoskin and Nelson, 1969, 1971; Sharma, 1979; Dale *et al.*, 1989; Aitken, 1990).

14.5.3. Ice Rafting

Ice-rafted debris (IRD) originates from floating termini such as ice shelves as well as icebergs and sea ice; ice-shelf rafted (ISRD), iceberg rafted (IBRD) and sea-ice rafted debris (SIRD) will be used to specifically identify the debris source (cf. Powell, 1984). The glacial character of IRD is important to recognize because sea ice can raft debris in areas where glaciers are absent on land (e.g. Gilbert, 1990). Generally, coarser particles such as gravel and sand sizes are most often recognized as being ice rafted, although contributions of finer particles can be significant (Gilbert, 1990; Powell, 1991). Resulting sediments include discrete lonestones that may be recognized as dropstones if they deform strata (Thomas and Connell, 1985), or as outsized-clasts when their diameter is larger than the thickness of strata in which they are embedded, or in the form of clusters of particles. Other features of IRD are till pellets and mud pellets that were frozen debris encased in glacial ice (Ovenshine, 1970; Powell, 1983b).

14.5.4. Ice-Shelf Rafting

Ice-shelf rafting occurs when basal and englacial sediment is transported beyond a grounding-line by ice deforming into a floating glacier-tongue or an ice-shelf. Direct sampling and observation of this process is very difficult since the floating ice is commonly too thick to allow direct access near the grounding-line.

Debris transport and release are discussed in Section 14.3.2.2. Most particles are released by melt-out and descend to the seafloor in rock falls, grain falls or suspension settling of individual particles. There is a significant lack of particle clusters as can occur when icebergs roll (Section 14.5.5). Because layers at higher elevations within the ice move farther from the grounding-line than lower layers, as happens when you slide the top of a deck of cards, englacial debris encased high above the ice sole can be transported farther than basal debris. Consequently, on the sea floor, ISRD may show a change in provenance as well as shape (from more rounded to more angular) with distance from the grounding-line (Powell, 1984). Deposits from these processes have been termed shelfstone muds or shelfstone diamictons depending on their texture (Powell, 1984).

These remote grounding-lines were observed for the first time recently, by using a submersible, remotely operated vehicle (ROV) at a glacier in Antarctica (Powell *et al.*, 1992). At this glacier, the process of sedimentation sea-ward of the grounding-line involved only direct undermelt of the floating glacier-tongue to produce shelfstone diamicton as glaciomarine sediment overlying subglacial (lodgement?) till (Plate 14.8). The till had ridge-and-furrow morphology and streamlined features indicative of subglacial deposition that revealed little modification after the glacier had detached from the bed. The till was immediately draped by the shelfstone diamicton. Basal debris thinned from 20 to 10 m over a down-flow distance of 1.8 km where the glacier remained in contact with the sea floor, giving an approximate rate of till deposition land-ward of the grounding-line of 0.9 cm $year^{-1}$. Sea-ward of the grounding-line, all basal debris had been melted out by seawater within about 2 km, giving a rate of glaciomarine sedimentation of 1.8 cm $year^{-1}$ (Powell *et al.*, 1992).

14.5.5. Iceberg Rafting

Rafting of glacial debris by icebergs can provide a significant volume of glaciomarine sediment and IBRD and can be used to indicate glaciation when mixed with marine sediment of a non-glacial origin. The distribution of icebergs in the circum-Antarctic and North Atlantic (Figs 14.1 and 14.10) clearly demonstrate the far-travelled character of IBRD. Hence, while local climatic conditions may not be suitable for glaciation, IBRD can be deposited with 'warm water' sediments when icebergs are transported quickly by ocean currents. Fourier shapes of silt particles can distinguish IBRD from glaciofluvial quartz grains (Cai *et al.*, 1992) and this technique could be useful for distinguishing iceberg-rafted silt from hemipelagic muds accumulated from other sources on continental shelves.

Melt-out, sediment gravity flows and dumping are the main mechanisms of debris release from icebergs, and they contribute particles of clay to boulder size to the sea floor by rock fall, grain fall and suspension

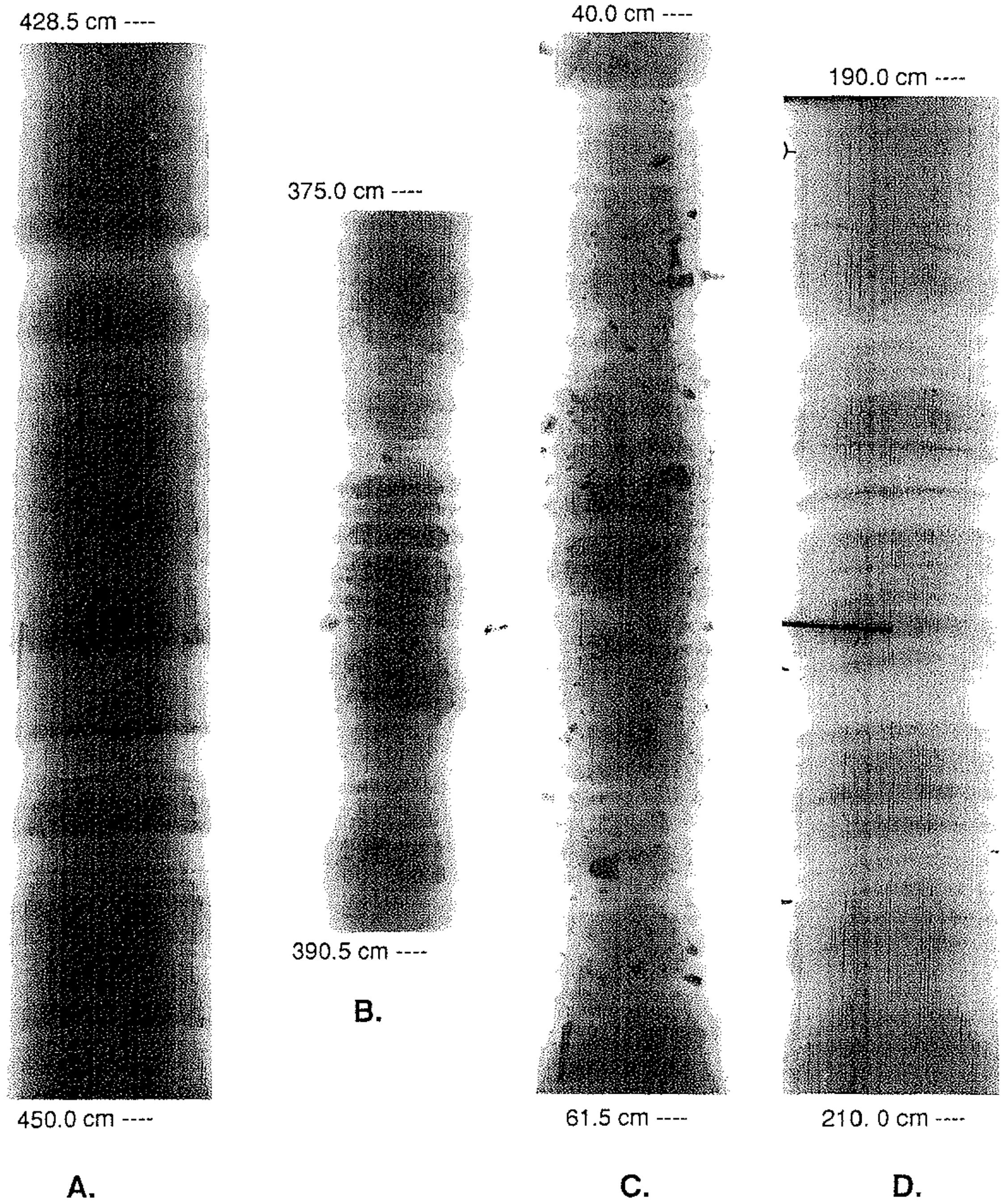


PLATE 14.7. X-radiographs of typical Antarctic glaciomarine sediment. Sediment cores are arranged from left to right in order to demonstrate end-member and transitional compositions of siliciclastic (left) and biogenic(right) sediment. Core A (428.5–450.0 cm) is rhythmically laminated silty mud with minor ice-rafted debris from the inner basin of a fjord some 2 km from the glacier terminus; Gerlache Strait, Antarctic Peninsula. Core B (375.0–390.5 cm) is randomly bedded, sandy to silty mud with variable ice-rafted debris at the same position as core A. Core C (40.0–61.5 cm) is siliceous (diatom) pebbly mud with significant ice-rafted debris and is a good example of two-component mixing that occurs some 19 km from the glaciated coast of the Antarctic Peninsula, Gerlache Strait. Core D (190.0–210.0 cm) is a laminated biosiliceous (diatom) ooze with minor ice-rafted debris, at some 35 km from the glaciated coast of George V Land.

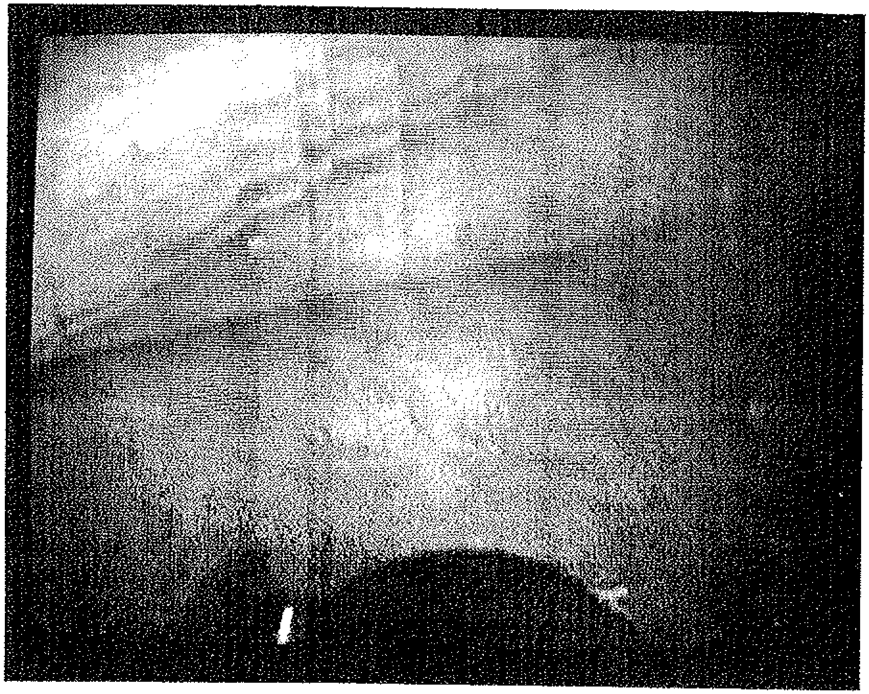


PLATE 14.8. View of a grounding line at a floating glacier-tongue of a polar outlet glacier from the East Antarctic Ice Sheet. Sediment in the foreground is a glaciomarine shelfstone diamicton formed by rainout of debris from melting out of the base of the floating tongue. The glaciomarine diamicton is draped over a stiff subglacial diamicton. In the background, beyond the grounding line which cuts diagonally across the photograph, the glacier is exposed as a vertical cliff because, here, a crevasse extended up from the base of the floating tongue. Basal debris layers dip to the left toward the base of the glacier and extend up to 20 m above the base. Vertical lines in the glacier face are runnels melted by water rising up the face in a similar way to those described for temperate termini (Powell and Molnia, 1989; reprinted by permission of Elsevier Science Publishers BV, Amsterdam). View is 3 to 4 m across.

settling. The ultimate volume of IBRD depends on the size and numbers of icebergs and also their residence times (horizontal velocity), debris concentrations, melting rates and rolling rates. Because melting rates in air are slower than those in water, icebergs eventually get top-heavy and roll. Submarine melting rates vary around and under the iceberg depending on the surface attitudes and currents (Weeks and Campbell, 1973; Gade, 1979; Russell-Head, 1980; Budd *et al.*, 1980; Huppert, 1980). Iceberg rolling is more common for smaller icebergs than for larger tabular icebergs because the latter have a more uniform distribution of mass and they are thus more stable in waves and wind (Weeks and Mellor, 1978). Predicting rates of ice rafting is very difficult, but Drewry and Cooper (1981) and Dowdeswell and Murray (1990) have modelled hypothetical debris-release histories for icebergs which may vary from one situation to the next.

Potentially, IBRD contributions vary with climatic and glacial regime. Large tabular icebergs produced in Greenland and Antarctica can travel to quite low latitudes (30°N, 40°S) (Figs 14.1 and 14.10) but those produced by large ice shelves contain little debris since the debris is preferentially released from the ice shelf near the grounding-line. However, reports from Antarctica of large icebergs with very thick debris layers are not uncommon. Those with higher debris contents are mainly derived from shorter floating termini and floating glacier-tongues, especially those formed by outlet glaciers (Anderson *et al.*, 1980a; Drewry, 1986). Most debris from Antarctic icebergs is thought to be released on the continental shelf due to long residence times and recirculating drift paths of

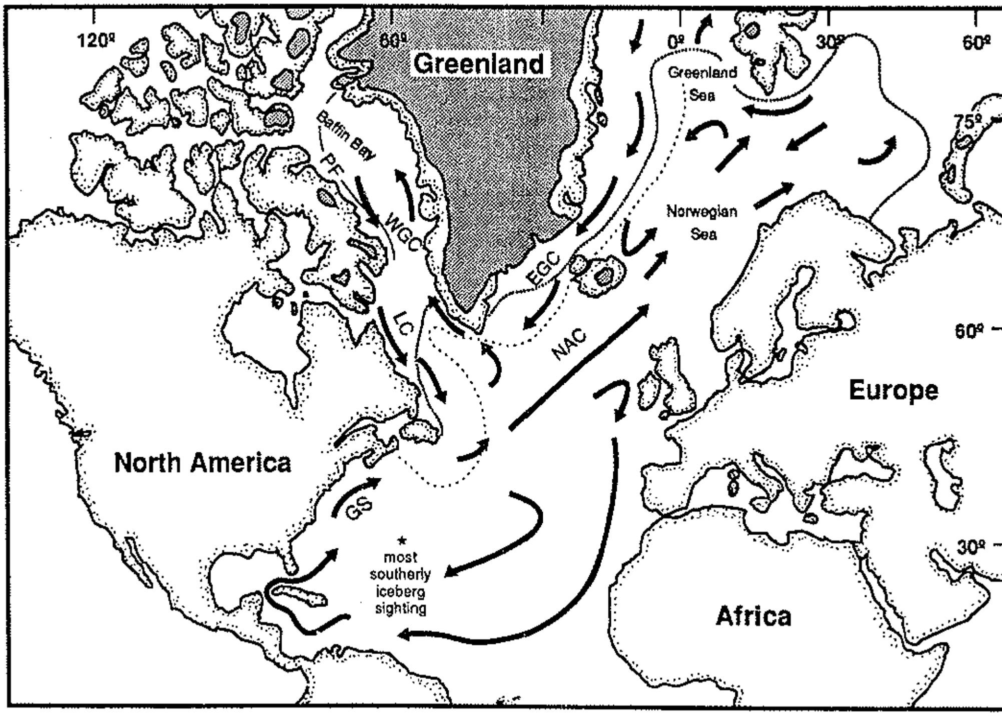


FIG. 14.10. Map of the North Atlantic region and adjacent glaciated (shaded area) and non-glaciated land masses. Dashed line in ocean is maximum limit of seasonal pack ice while dotted line is southern extent of drifting icebergs. Solid line in northern Baffin Bay is the location of the polar front. EGC = East Greenland Current, GS = Gulf Stream, WGC = West Greenland Current, LC = Labrador Current, NAC = North Atlantic Current.

icebergs offshore (Ørheim, 1980; Vinje, 1980; Keys, 1984). Occasionally, icebergs collect in restricted areas because sea ice prevents them drifting after they calve. These iceberg tongues have tabular icebergs separated by lower-lying fast ice. In the North Atlantic large icebergs from Greenland are transported far south by the Labrador Current and can cause problems to shipping lanes (e.g. the 'Titanic' disaster).

Tidewater termini produce smaller icebergs than floating termini and with iceberg transport lengths of only a few tens of kilometres in cool-temperate climates. In boreal and polar-tundra climates these distances can be much longer (Robe, 1980). Because of the rapid melting rate and the coincidence of calving and grounding-lines for tidewater termini, roll-dump structures are common. These structures can be used to infer this environment rather than a floating terminus at a grounding-line since dumping of large amounts of debris en masse is not likely to occur from the base of an ice shelf (Powell, 1984, 1991).

In iceberg zones of temperate termini, icebergs may be concentrated by various processes (Section 14.5.6); but even when iceberg flux through the ice proximal area is high, the area may remain totally covered with icebergs if their production rate is sufficiently high. High glacier flow speeds and calving speeds can be maintained even under cool-temperate climates, especially during rapid grounding-line retreat. In some situations, rates of IBRD production may be predicted to be high near a tidewater cliff and in others, rates could increase slightly farther away. The resulting deposit on the sea floor depends on the relative rates of IBRD and mud or biogenic production; in other words the problem can be modelled as two component mixing.

14.5.6. Two Component Mixing

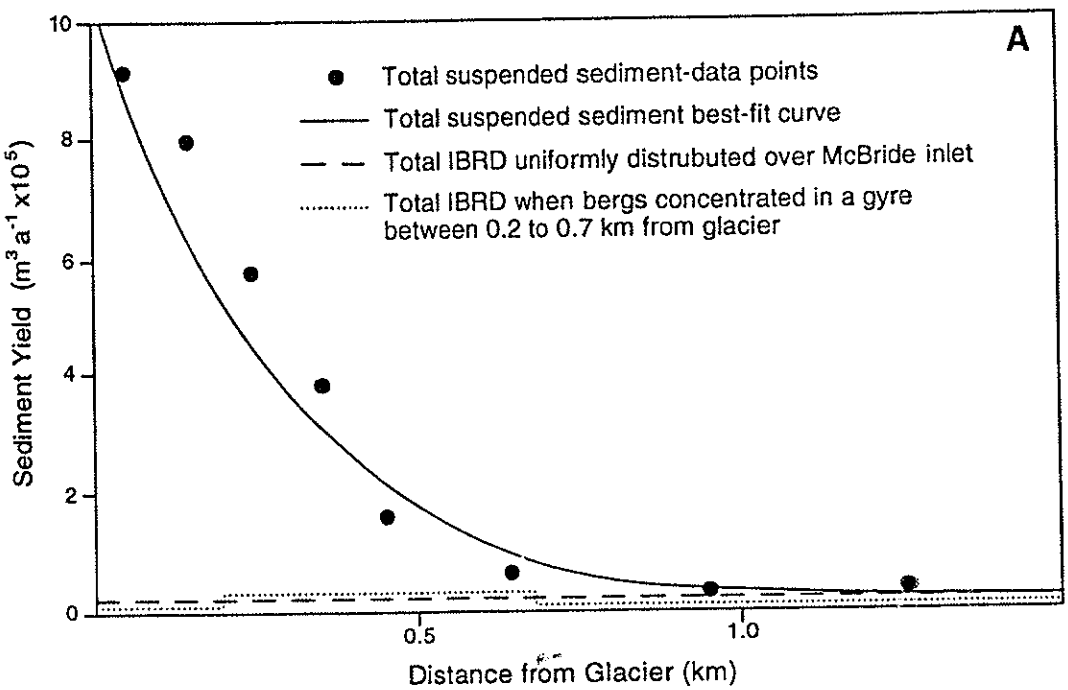
'Two component mixing' refers to the admixture of sediment particles derived from processes which are sufficiently distinct that they can be recognized by the character of the particles they contribute to a deposit. The two most prevalent in the glaciomarine environment are ice rafting and various types of current derived sediment, such as that produced by vertical settling from suspension in low density turbidity flows and bottom currents. The coarser (gravel and sand) ice-rafted component can usually be recognized by poor sorting of the glacially transported debris. However, ice-rafted silt and clay is more difficult to discern from hemipelagic particles in bottom sediment. Current-derived components can be expected to have undergone some sorting during the transport and depositional processes and can often be distinguished by stratification and by modal sizes within the sediment. It is possible to distinguish sediment from different sources by separating particle-size populations using detailed size analysis (e.g. Piper, 1976; Kelly, 1986). Shapes of quartz silt particles can be used to distinguish IBRD from silt from other sources (Cai *et al.*, 1992) (Menzies, 1995b, Chapter 13). Depending upon the relative contribution of the sediment components, various sediments are produced ranging from pebbly muds (i.e. compound glaciomarine sediments (Anderson *et al.*, 1980b) or bergstone muds (Powell, 1984)) to diamictons. In temperate and subpolar settings the current-derived component is often supplied by meltwater discharges along or adjacent to glacier termini or deltaic systems.

The manner in which texture of bottom sediment changes can be qualitatively inferred by superimposing a curve of rates of IBRD production onto those for settling of sediment from overflow plumes (Fig. 14.11). At McBride Glacier, Alaska, icebergs melt within a few kilometres of a grounding-line and thus IBRD production is 'high' (Powell, 1991). IBRD volumes become significant with respect to plume deposits beyond approximately one kilometre of the grounding-line (Fig. 14.11) and a bergstone diamicton may accumulate beyond a more proximal bergstone mud zone. That yield includes all sizes of IBRD but, when only the gravel component of IBRD is considered, the IBRD percentage in bottom sediment is much lower (Fig. 14.11). Therefore, the lonestones which are easily observed in outcrops or cores, are only a small proportion of IBRD that a unit may actually contain.

In most temperate and subpolar regimes fluvial sediment production will dominate iceberg rafting and bergstone muds accumulate. If icebergs are concentrated for any reason, then the proportion of IBRD can increase to produce a bergstone diamicton interfingering with bergstone mud. Circumstances in which a bergstone diamicton may form are, for example, when: (1) meltwater discharge is low due to a cold glacial regime, (2) seasonal discharge decreases but iceberg rafting continues, (3) there is a very high, continuous flux of icebergs in proximal areas, and (4) icebergs occur in a continuous flux in more distal areas where rates of accumulation of marine rock flour have decreased exponentially. Icebergs may be concentrated or have a high continuous flux in situations such as: (1) in constricted basins (e.g. trapped by sills) (Plate 14.9), (2) in oceanic eddies, (3) during catastrophic retreats, (4) during storms blowing icebergs onshore, and (5) during periods when sea ice freezes-in high concentrations of icebergs. Rates of accumulation of the two sediment source-components will probably differ in fjords and on continental shelves because the latter are more open systems.

In polar environments a third, biogenic, component is involved. Because of the limited role of meltwater sedimentation, dropstone diamictons are found in a variety of situations such as near ice shelf grounding-lines and bays with restricted circulation. In general, sediment accumulation rates again decrease from the termini of valley glaciers. This allows for the accumulation of biosiliceous pebbly muds, mixtures of ice rafted and biogenic sediment derived from phytoplankton (Fig. 14.19).

In some areas, coarse IBRD and bioclastic debris are concentrated without a fine-grain component. These deposits have been termed residual glaciomarine sediments (Anderson *et al.*, 1980b) and occur in nearshore areas or shallow banks where marine currents are strong enough to stop fine-grained sediment from accumulating and bioturbation brings finer sediment to the sediment surface where it is removed by bottom currents.



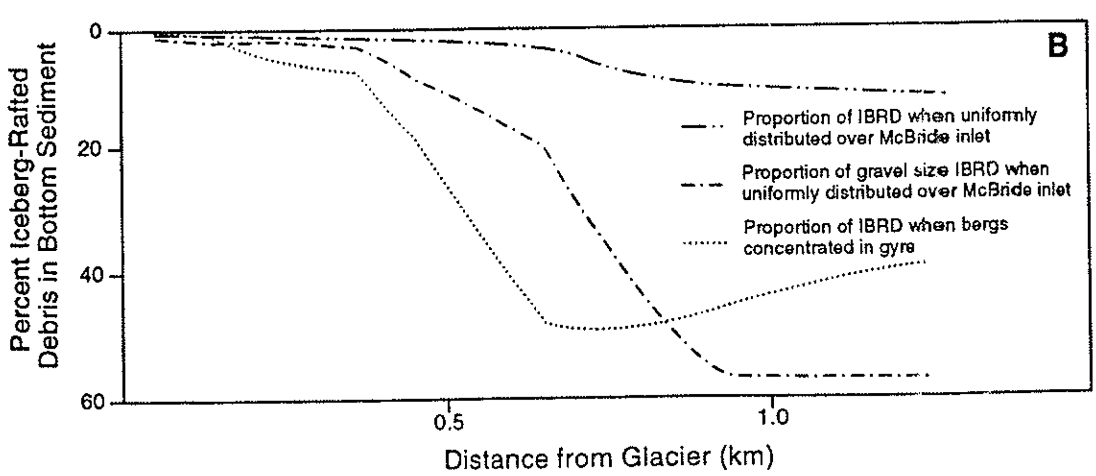


FIG. 14.11. Sediment yields (A) and proportions IBRD in bottom sediment (B) with distance from the tidewater terminus of McBride Glacier, Alaska. In the top plot (A), fields of marine rock flour that settles from suspension are from Cowan and Powell (1991), and IBRD yields are data from Gottler and Powell (1990). In the bottom plot (B), proportions of IBRD in bottom sediment are presented when (1) the total IBRD is distributed uniformly over McBride Inlet, (2) the gravel fraction (taken as the 15% average) of IBRD is distributed uniformly over McBride Inlet, and (3) the total IBRD is concentrated by bergs being in a gyre between 0.2 to 0.7 km from the glacier terminus (after Powell, 1991). (Reprinted individual figures by permission of the author.)

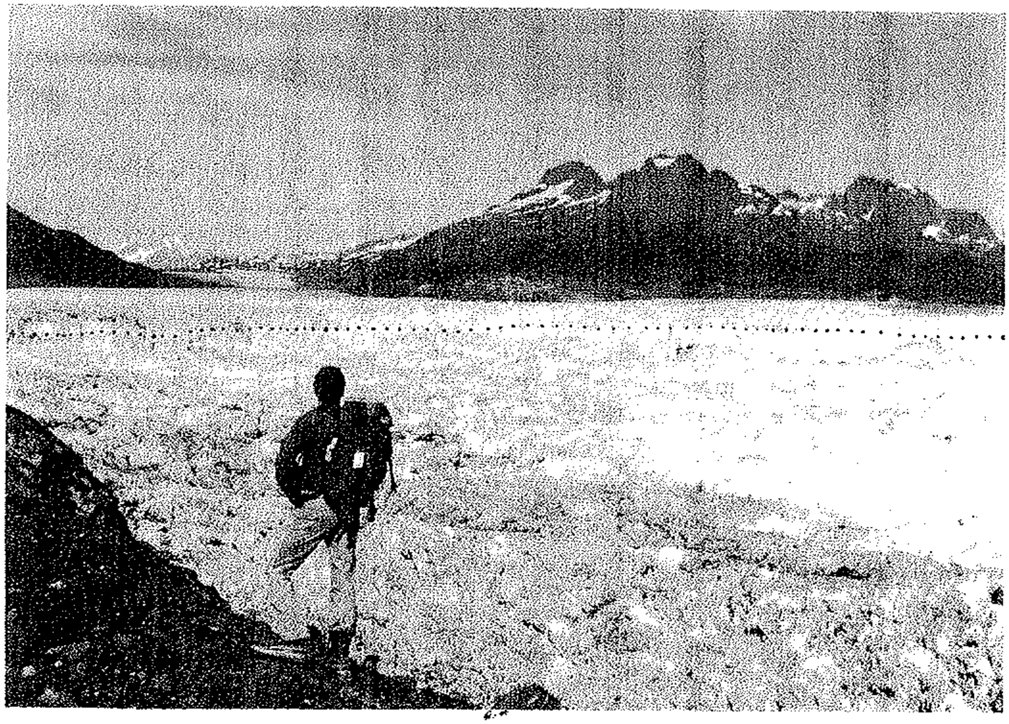


PLATE 14.9. Calving bay of the Columbia Glacier, Alaska (July, 1989). The bay is choked with recently calved icebergs which are prevented from drifting out of the bay by a shallow sill. The concentration of floating icebergs covers 100% of the bay's surface and results in conditions that would be the closest thing to a temperate ice shelf. Water depths within this iceberg-covered bay are up to 380 m deep (Krimmel and Meier, 1989). The dotted line indicates the water line against the grounded tidewater cliff.

In the absence of significant meltwater, most siliciclastic sediment is introduced either as IRD or as sediment gravity flows that are generated by instabilities associated with grounding-line processes (calving, ploughing, etc.). Because of the influence of dilute sediment gravity flows, basin topography plays a major role in the distribution of sediment facies within Antarctic fjords. Silled basins retain siliciclastic sediments within ice-proximal areas and prevent dilution of biogenic detritus within the outer fjord basins (Plate 14.8). Fjords which lack sills contain a more transitional facies gradient between ice-proximal siliciclastic sediment and distal biogenic and ice-rafted deposits (Fig. 14.9).

14.5.7. Sea-Ice Rafting

Formation of sea ice most commonly occurs when air temperatures drop below the freezing point of marine or brackish water. Its formation expels salt and produces brines that form gravity flows that sink down through the water column due to their density and cold temperatures (Lewis and Perkin, 1982). Sea ice can also form under cool-temperate climates where during winter in protected areas heavy snow falls produce a slush of ice crystals in surface water which then freezes as temperatures drop. One problem still remaining is to distinguish SIRD deposited under glacial conditions from that deposited where sea ice forms without glaciers on land (e.g. Reimnitz and Kempema, 1988; Gilbert, 1990).

Sea ice is important to glaciomarine sedimentation because it (1) damps waves (especially during winter storms), (2) changes water column structure, (3) influences biotic productivity, (4) traps icebergs, (5) transports sediment, (6) scours and turbates bottom sediment, (7) deforms shoreline sediment, and (8) reduces calving rates. Sediment can be incorporated into sea ice by (1) freezing of an ice foot, (2) anchor-ice lifting, (3) trapping of sediment during frazil ice/slush ice formation, (4) bottom erosion, (5) stream wash overs, (6) wave wash overs, (7) offshore aeolian transport, and (8) rock falls and slides (see Table 14.1) (e.g. Gilbert, 1983; Reimnitz *et al.*, 1987; Kempema *et al.*, 1989).

SIRD is being increasingly considered as an important component of glaciomarine sediments in distal areas or under colder conditions where other siliciclastic input is low (Clarke *et al.*, 1980; Clarke and Hanson, 1983; Barnes and Reimnitz, 1982; Barnes *et al.*, 1982a; Barrett *et al.*, 1983; Reimnitz and Kempema, 1988; Elverhøi *et al.*, 1989). Sea-ice rafting is important in Arctic seas because of the shallow depths of broad continental shelves and a coastal environment that includes deltaic and wave-dominated systems. Near deltaic systems sea ice has coarse debris incorporated from the sea floor during pressure ridge accretion while waves supply a great deal of fine detritus that is incorporated during freeze-up. Much of the sediment in Arctic sea ice is fine-grained and derived from the littoral zone. Concentrations vary between about 0.05 to 3000 mg l^{-1} (Campbell and Collin, 1958; Barnes *et al.*, 1982a; Elverhøi *et al.*, 1989) but few studies have been conducted. Conditions of debris transport and release are similar to those discussed for icebergs; however, in general, in any one climatic regime icebergs tend to survive longer than sea ice. Antarctic sea ice can be the dominant source of poorly sorted, coarse debris within fjords along the Antarctic Peninsula (Fig. 14.9). Here, rock falls onto fast-ice are common from the rugged cliffs and valley sides during the austral spring. Break-up into pack ice results in the transport of the debris out into fjords during the late spring and early summer. In areas of supercooled bottom water, anchor ice formation can pluck biogenic and coarse detritus from the sea floor. This debris is frozen to the bottom of overlying ice shelves or pack ice and is later rafted further out to sea (Kellogg and Kellogg, 1988). All of the shallow coastal zone surrounding Antarctica (from 0–10 m water depth) has been plucked clean by anchor ice formation (Picken, 1985). The exceptions to this are areas of strong tidal currents. Aeolian transport of fine-sand onto pack and even shelf ice is also significant, especially adjacent to the Dry Valleys along the southwest margin of the Ross Sea (Barrett *et al.*, 1983; Bartek and Anderson, 1991).

14.5.8. Transitional Environments

Transitional environments between terrestrial and fully marine environments mainly include beach, delta, estuarine and tidal flat systems. In some areas these systems are quite important in producing a sedimentary record as they are large depocentres. They can also provide important data for inferring climatic conditions, for example, large deltas are not produced in polar climates but can be significant in cool-temperate and boreal climates. The presence of ice in coastal settings can produce a distinctive record. For example, when grounding-lines of tidewater cliffs are directly at sea level, very gravelly, poorly sorted beaches are formed (Znachko-Yavorskiy, 1964; Sugden and John, 1976); icebergs and sea ice leave deposits such as boulder lines, and deform intertidal sediment into gouges, wallows and push ridges (Barnes, 1982; Barnes *et al.*, 1982a; Reimnitz and Kempema, 1982; Dionne, 1987; Dionne and Brodeur, 1988).

Beaches. In cool-temperate climates, beach sediments range from gravelly to sandy and include all of the features of non-glaciomarine beaches with evidence of ice action superimposed (e.g. Molnia and Wheeler, 1978; Hayes and Michel, 1989). Beach ridge systems with associated aeolian dunes and various forms of spits are common, as are barrier island systems associated with large deltas.

In contrast, beach sediments in more extreme climates tend to have particles that are slightly angular and moderate to poorly sorted, probably due to waves being damped by sea ice (e.g. Powell, 1981b). These beaches have superimposed ice-push ridges and evidence of periglacial action (Nichols, 1961, 1968; Dubrovin, 1979; Gregory *et al.*, 1984). Antarctic beaches are rare and consequently are often sites for penguin rookeries. Because of the penguin activity, old raised beaches have sorted gravels (penguin nests) and penguin bones and mummified carcasses (Whitehouse *et al.*, 1989).

Deltas and estuaries. In order that deltas form, fluvial discharges have to be moderately large. In general, streams in polar ice cap climates do not have high discharges (typically <2 $m^3 s^{-1}$) nor do they carry high sediment loads (Chinn and Oliver, 1983; Howard-Williams and Vincent, 1986; Gallagher and Burton, 1988; Mosley, 1988). Small glaciofluvial systems that do exist along portions of the Dry Valleys are ephemeral and have sediment yields of 5.9 t km^{-2} $year^{-1}$; this is two orders of magnitude less than glacial outwash systems of the Arctic alpine regions (Mosley, 1988). The Antarctic fluvial systems have extensive algal mats and filaments and suspended sediment loads typically <10 mg l^{-1} (Howard-Williams and Vincent, 1986). Consequently, with low fluvial sediment delivery, deltas in these conditions are minor and texturally deltaic beach and fluvial sediments are all similar (Powell, 1981b). Certainly, estuaries, lagoons and marshes are not formed.

Large deltas such as the Yukon Delta (Dupre and Thompson, 1979) can form in slightly warmer climates, and the full range of deltas, estuaries, lagoons and marshes form in temperate coastal climates (Hayes and Michel, 1989; Powell and Molnia, 1989). In these latter conditions, fluvial discharges and sediment loads may be very high; for example, peaks of 7×10^3 $m^3 s^{-1}$ and a total sediment production of 1.07×10^{11} kg $year^{-1}$ are reported for the Copper River, Alaska (Reimnitz, 1966). Sedimentary processes and facies at the river mouth and on the delta front are similar to non-glacial situations except for the influence of fluctuations in discharges and indications of ice action; deltas are generally fan deltas (n.b. Syvitski and Skei, 1983; Hayes and Michel, 1989). These deltas can also develop from grounding-line fans once they build to sea level (Powell, 1990). One characteristic that may distinguish fjords headed by grounding-line fans from those headed by deltas is turbidity current channels (Carlson *et al.*, 1989). Both systems have slides/slumps and sediment gravity flows (cf. Prior and Bornhold, 1989; Powell and Molnia, 1989), but only once the gradient of the foreslope increases to that of a delta can turbidity currents build sufficient energy to be erosive on the fjord floor and cut major channels (Carlson *et al.*, 1989). Sediment transport in the buoyant overflow plume has been modelled by Syvitski *et al.* (1988) (Sections 3.4 and 3.5.1).

Grounding-line fans may aggrade and prograde to form ice-contact deltas and, in the transition, when the delta plain is fully intertidal, channels change location during tidal cycles (Powell and Molnia, 1989). Large volumes of sediment may be removed from the delta by this process to produce a pulse of coarser sediment to the sea floor with ebbing tide. A more mature delta can produce laminites with a spring tidal periodicity by tidal draw-down funnelling coarser sediment over the delta brink at low tide (Smith *et al.*, 1990).

Tidal flats. Tidal flats are not common in polar ice cap climates today, mainly because of narrow, high gradient, coast lines in Antarctica. Where low gradient intertidal areas do exist, they are typically the sites of boulder pavements, depending on the onshore movement of floating ice (Hansom, 1983). In warmer climatic conditions, tidal flats can be significant systems with sedimentary characteristics similar to non-glacial tidal flats except for indications of icebergs and sea ice (Bartsch-Winkler and Ovenshine, 1984; Bartsch-Winkler and Schnell, 1984). Some indications of ice are quite distinctive, for example, isolated boulders and boulder barricades, ice wallows and other keel marks and striations and pavements (Dionne, 1985; Aitken *et al.*, 1988; Aitken and Gilbert, 1989).

14.6. OTHER NON-GLACIAL AND MODIFYING PROCESSES

Some non-glacial processes have already been discussed such as sea ice and biological productivity. However, there are several other aspects that deserve to be mentioned.

14.6.1. Hydrographic Processes

In addition to controlling ice melting rates, iceberg paths (Figs 14.1 and 14.10) and paths of turbid freshwater plumes, oceanic currents are important in modifying and redepositing sediments on the sea floor. Hydrography of continental shelves and shallow seas is very site specific in regard to details of circulation patterns, current strengths and whether the environment is dominated by tides or storm waves. Some generalities can be made.

Along-shelf currents commonly transport sediment at velocities that reach several tens of centimetres per second. Those are often geostrophic flows following salinity gradients established by high freshwater influx at the cool-temperate coast and are enhanced by net wave propagation (Powell and Molnia, 1989). These along-shelf currents are capable of transporting sediment to low areas on the shelf. If the lows are glacially-eroded troughs that cross the shelf, then sediment can be subsequently transported offshore to deeper water (tunnel valleys, Menzies, 1995b, Chapter 2). This enhances hemiplagic sedimentation off glaciated coasts.

Dense water can be produced nearshore during winter cooling and, when combined with wind-driven down-welling, can form offshore, gravitationally convective flows. This effect is enhanced by sea ice formation. Alternatively, during summer, strong katabatic winds, flowing from glaciers offshore, cool surface water as well as cause upwelling of deeper water along the coast.

Under Polar Ice Cap climates, the overriding factor of continental shelf hydrography is heat exchange between the atmosphere–ocean and ice–ocean, especially with sea ice formation, but locally with ice shelves. The net result is off-shelf transport and downward salt flux (Jacobs, 1989; Foldvik and Gammelsrod, 1988). These processes drive vertical convection and 'ventilated' dense water across continental shelves and eventually form the ocean bottom waters (Coachman and Aagard, 1974; Jacobs, 1989) which create most of the turnover in deep areas of modern oceans.

Storms are common in high latitudes and waves so propagated may reach over 40 m high. Waves cause sediment erosion and major reworking episodes to alter the original glaciomarine character. They also promulgate mass failure of sediment on shelves. Wave effects are damped by sea ice or high concentrations of icebergs. In fact, storms are one way of concentrating icebergs in local areas, especially at the heads of fjords and in bays.

Tidal currents produce local high frequency fluctuations in near-bottom sediment concentrations that dominate some shelves. Like waves, tidal currents rework glaciomarine sediment, but their deposits will differ.

Shallow water environments of Polar Ice Cap climates are greatly affected by currents of wind or tidal origin, especially bordering ice piedmont settings. Low siliciclastic sediment supply at the calving line allows such currents to effectively sort and transport sediment into broad aprons of sand rich in bioclastic detritus.

Fjords have complex circulation patterns that are strongly influenced by bottom topography. If fjords do not have entrance sills, then waves and tides on a continental shelf propagate into fjords and can be major influences in shallow fjords. Sills significantly influence fjord circulation because deep water renewal is controlled by sill height. Deep water renewal occurs by gravity currents as the density of water outside the sill becomes greater than bottom water of the fjord basin and so flows down into, the basin. Velocities range from 20 to 70 cm s^{-1} (Syvitski *et al.*, 1987). These flows stop anoxia and transport sediment out of, or into a basin (Winters *et al.*, 1985). Tidal currents flowing over sills make sill surfaces sites of erosion and perhaps high biologic activity. Sills form internal waves at the base of the fjord surface layer that may propagate long distances into the fjord and effect the turbid overflow plume. Tidal currents may also scour and redeposit sediment from fjord walls down to the basin floor (Syvitski, 1989) and prevent anchor ice formation. This stratification breaks down during summer as icebergs and pack ice are blown into and out of fjords by winds and storms.

A common characteristic of fjords is to have a maximum homogeneity of water masses during winters by low freshwater input, thermohaline convection due to surface cooling and/or sea ice formation and deep-water renewal (Syvitski *et al.*, 1987). Major stratification within a fjord water column is generated in summer and autumn by surface freshwater flowing seaward as a barotropic flow from streams either at deltas or tidewater termini. Katabatic winds enhance this flow. To balance the hypopycnal outflow, an intermediate extrabasinal water layer either from the continental shelf or the next basin downfjord flows in as a baroclinic current. In polar fjords along the Antarctic Peninsula, stratification is greatest during the spring as sea ice melts and contributes to a low salinity surface layer.

14.6.2. Sediment Gravity Flows

Sites for generating sediment mass movements are depocentres such as grounding-lines or deltas where high sedimentation rates form unstable slopes that collapse. Sediment gravity flows from these depocentres may have short run-out distances because of either significant bottom relief produced by glacial erosion especially in fjords that trap flows, or the continental shelf sloping toward the continent due to either glacial loading (Anderson *et al.*, 1983) or preferential glacial erosion which also traps flows, or a low slope of the continental shelf on which flows rapidly slow. However, events other than high sedimentation rates, such as storm waves, biogenic gas and earthquakes may result in sediment instability and consequent mass movements in subaquatic proglacial areas. Side-entry inputs are important additional sources in fjords.

Processes of sediment redistribution in proglacial regimes are the same as in non-glacial marine settings and large volumes of sediment can be involved (e.g. Kurtz and Anderson, 1980; Wright and Anderson, 1982; Wright *et al.*, 1983; Vorren *et al.*, 1989). Significant features specific to glaciomarine settings are, first, the possibility of producing mass movement deposits on continental shelves on a regular and perhaps rapid basis due to the glacier being on the shelf. In contrast, non-glacial shelves have their sediment sources usually some distance from sites of deposition except during sea level changes. Second, during glacial maxima, glaciers may often extend to the edge of the continental shelf so that all sediment is released directly down the continental slope to form thick wedges of mass flow deposits with glacial signatures such as striated clasts (Hill, 1984; King and Fader, 1986; Piper and Sparkes, 1987; Bonifay and Piper, 1988; Stoker, 1988; Vorren *et al.*, 1989) (Menzies, 1995b, Chapter 5).

14.6.3. Iceberg and Ice Keel Scouring and Turbation

Plough and furrow marks are common on high latitude shelves in northern (Fader *et al.*, 1982; Reimnitz and Kempema, 1982; Vorren *et al.*, 1983; Barnes *et al.*, 1984) and southern hemispheres (Barnes, 1987; Barnes and Lein, 1988). In fact, turbation of bottom sediments can be so intense that primary stratigraphy and biotic communities may be destroyed (Elverhøi, 1984; Gallardo, 1987) and/or remoulded into diamictons (Vorren *et al.*, 1983). Gouges and wallows can also be produced in intertidal areas (cf. Reimnitz and Kempema, 1982; Hansom, 1983) (Menzies, 1995b, Chapter 4).

The depth of iceberg/sea floor interaction in the Antarctic is dependent upon the thickness of tabular icebergs. Studies suggest that the draft of Antarctic icebergs is limited to about 400 m with some non-tabular icebergs as great as 500 m. A variety of iceberg scours, prods and gouges are revealed by side-scan sonar studies on portions of the Antarctic shelf as deep as 500 m (Barnes and Lein, 1988). The effect of this process is to produce an iceberg turbate sediment from relict deposits which lie above 500 m on the shelf. Such deposits would most likely consist of poorly stratified diamictons with convoluted and deformed interbeds of sand or siliceous mud. Fossils would be represented by mixtures of recent and relict assemblages (Menzies, 1995b, Chapters 4 and 5).

14.6.4. Production and Accumulation of Organic Carbon

The content and character of organic carbon in sediments are perhaps one of the more important differences between polar and temperate glaciomarine facies, particularly within mud-dominated facies. Recent studies within shelf and fjord basins in Antarctica have demonstrated that total organic carbon contents within biosiliceous muds can be as high as 3–4% of the total sediment weight (Dunbar, 1988) and up to 6% of the mud fraction (Domack, 1988). Typically, the organic carbon values range between 1–1.5%. More important, however, is the fact that almost all of the carbon is of autochthonous marine (algal) origin. The preserved flux of relatively unstable organic matter in circum-Antarctic basins ranges from 22.8 to 6.45 gC m^{-2} $year^{-1}$ which compares to 4.2 gC m^{-2} $year^{-1}$ in Baffin Island fjords (Syvitski *et al.*, 1990).

The higher Antarctic rate is related to three factors. First, the polar climate inhibits siliciclastic sedimentation and limits dilution of the organic matter. Second, locally high productivity contributes to organic matter loading to the bottom. Third, the cold bottom water temperatures (<–1.5°C) limit bacterial metabolism within and along the sediment–water interface. Although bottom waters are fully oxygenated within Antarctic shelf basins, subsurface conditions within the sediment are anoxic, further limiting consumption of the organic matter. Associated authigenic minerals include abundant framboidal pyrite and, less commonly, calcium hexahydrite (ikaite) (Suess *et al.*, 1982).

The biosiliceous mud/ooze facies is widespread along the Antarctic continental shelf within basins >500 m water depth. This has led several authors to suggest that the preservation of both organic carbon and opaline silica within these sediments has an effect upon global budgets for carbon and silica (Ledford-Hoffman *et al.*, 1986; DeMaster *et al.*, 1988; Dunbar, 1988).

In contrast, temperate and subpolar glaciomarine muds are typically poor in organic carbon, with typical values of <1% (Andrews, 1987d; Syvitski *et al.*, 1990). When Total Organic Carbon (TOC) content exceeds 1%, the organic matter is invariably detrital or reworked, being derived from stable refractory sources such as bedrock and terrestrial vegetation. The reasons for this are related to climate and high rates of siliciclastic sedimentation associated with meltwater input. Restricted productivity due to limited sea ice may also be a contributory factor to low indigenous carbon contents. Warmer water temperatures would also allow for higher rates of bacterial metabolism.

14.6.5. Bioturbation

Infaunal organisms disturb sediment and progressively destroy sedimentary structures with increasing intensity away from a glacier (Gilbert, 1982; Dale *et al.*, 1989). Bioturbation may also allow winnowing of finer sediment (Singer and Anderson, 1984) and may homogenize sediment into a diamicton. The best potential for preserving fine sedimentary structures is where sedimentation rates are high and infaunal populations are consequently absent or low.

14.6.6. Aeolian Sources

Offshore transport of sediment by wind can be locally important in high latitude polar ice cap and polar-tundra climates (e.g. Gilbert, 1982; Barrett *et al.*, 1983). Fine sand and silt is transported to the sea directly, or via sea ice. In more temperate climates, wind also transports sediment offshore (Post, 1976; Molnia, 1990) but it is heavily diluted by sediment from other sources.

14.7. SEDIMENTATION RATES AND FLUXES

Potential siliciclastic sediment yields from glaciated basins decrease with decreasing size of the glaciated basin assuming the following conditions: (1) uniform subglacial sediment/rock types, (2) constant subglacial debris conditions (erosion-transportation-release) during glacial shrinkage, and (3) minimal sediment storage (probably true for marine-ending glaciers except during glacial minimum conditions where extensive land areas are exposed in a drainage basin). A logarithmic decrease in sediment yield with decreasing drainage basin area has been documented (Fig. 14.12) for the retreat of the temperate Muir Glacier, Alaska (Powell, 1991). The sediment yield curve is considered to be a continuous function reflecting rates of glacial erosion and flow.

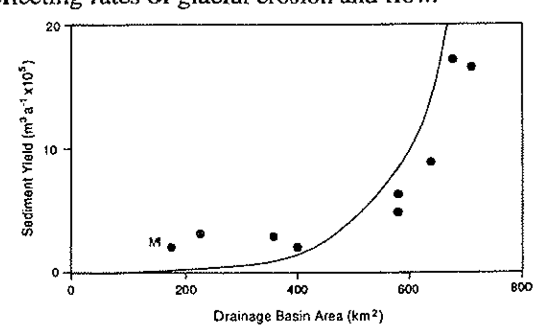


FIG. 14.12. Relation between sediment yields and drainage basin area for a temperate marine-ending glacier. Sediment yields in ice-contact basins (fjord floor basins in which a glacier terminates) are estimated from seismic profiles from Molnia *et al.* (1984) averaged over the period of time Muir Glacier, Alaska, terminated in or at the head of each basin. Drainage basin areas of Muir Glacier, as determined from Brown *et al.* (1982), become smaller during glacial retreat. Curve is a best fit logarithmic function. M is the yield from present McBride Glacier determined from modern sedimentation rates and is consistent with yields estimated for Muir Glacier from seismic profiles (after Powell, 1991; reprinted with permission of the author).

Rates of glacial sediment accumulation on the sea floor depend on the processes of debris release from the glacier (Fig. 14.13). Accumulation rates are a function of: debris content of the glacier and rate of meltout by sea water; subglacial squeeze/push; sediment discharge from basal/subglacial streams; hemipelagic/pelagic suspension settling from meltwater discharges; debris fall from floating ice; marine hydrographic processes for length of transport and washing and winnowing of fine particles; and sea floor processes involved with mass movement. Of these, sedimentation rates from suspension settling are easiest to measure; very few estimates of other processes have been made. An estimate of relative significance of these processes in a temperate situation is that melting of basal debris produces 0.4% of total sediment yield, stream bedload plus mass-flow 30.6%, and suspension settling 69% (Powell and Molnia, 1989). Estimates of annual increases in volume of grounding-line systems in these temperate conditions are of the order of 10^6 to 10^7 m^3 $year^{-1}$ (Powell and Molnia, 1989; Powell, 1990, 1991).

As expected, rates of sedimentation of hemipelagic siliciclastic glacial sediment logarithmically decrease away from grounding lines (Elverhøi *et al.*, 1980, 1989; Andrews *et al.*, 1985; Görlich, 1986; Andrews, 1987d; Syvitski, 1989; Vorren *et al.*, 1989; Cowan and Powell, 1991; Domack *et al.*, 1991b). The low production of siliciclastic sediment in polar ice cap climates means that sedimentation rates become more a function of biosiliceous productivity and dissolution rates. Surface productivity, in turn, is controlled by sea ice fluctuations and subsequent surface layer mixing.

Table 14.3 lists some recent estimates of sedimentation rates for modern Antarctic glaciomarine sediments. Though far from representative of all depositional environments, especially ice proximal settings, the data do give some basis for comparison to other glaciomarine sequences from cool temperate zones. Sediment accumulation rates vary from 3.4 to 0.5 mm $year^{-1}$ over two orders of magnitude. Most rates range between 0.3 to 3.0 mm $year^{-1}$. Because most of these sediments are biosiliceous muds or oozes the rates of siliciclastic sediment accumulation are approximately 70 to 50% of those listed in Table 14.2. The overall low sedimentation rates, however, emphasize the low rate of siliciclastic sediment supply and the absence of extensive fluvial drainage systems.

Another way to measure sedimentation is by determining preserved particle flux which is calculated using sediment bulk density and ^{14}C (long-term) or ^{210}Pb (short-term) chronologies (Table 14.3). Results in the Antarctic demonstrate an order of magnitude variation between 0.471 kg m^{-2} $year^{-1}$ (Amery Trough, Prydz Bay) to 2.285 kg m^{-2} $year^{-1}$ (Andvord Bay, Gerlache Strait, Antarctic Peninsula). In the Ross Sea, the total siliciclastic particle flux ranges from 0.619 kg m^{-2} $year^{-1}$ (Sulzburger Bay) to 0.741 kg m^{-2} $year^{-1}$ (southwest Ross Sea) (Ledford-Hoffman *et al.*, 1986). These results compare to total particle flux values in the Arctic of 0.060 kg m^{-2} $year^{-1}$ (southeast Greenland Shelf), 0.60–1.30 kg m^{-2} $year^{-1}$ (Baffin Island fjords) to 0.40 kg m^{-2} $year^{-1}$ (Baffin Island, shelf trough) (Andrews and Syvitski, 1991).

Total particle flux in the temperate-oceanic fjords of Alaska are several orders of magnitude greater than those of subpolar and polar climates mentioned above (Table 14.4). The results of Powell and others are taken from sediment traps and, therefore, would be expected to represent upper limits to sedimentation when compared to preserved flux as determined by radiogenic isotopic studies of sediment cores. Regardless, the contrasts are extreme and demonstrate the proficiency by which temperate glacial regimes supply sediment into the marine realm.

In order to obtain a reliable comparison of sedimentation rates in different climatic regimes, Cowan and Powell (1991) restricted the data set to those studies that obtained measurements from settling tubes in modern fjord settings and close to grounding lines (Table 14.5). The results show a dramatic 3 to 4 orders of magnitude difference between Temperate and Polar settings, and 1 to 2 orders of magnitude difference between both Temperate and Subpolar settings, as well as Subpolar and Polar settings.

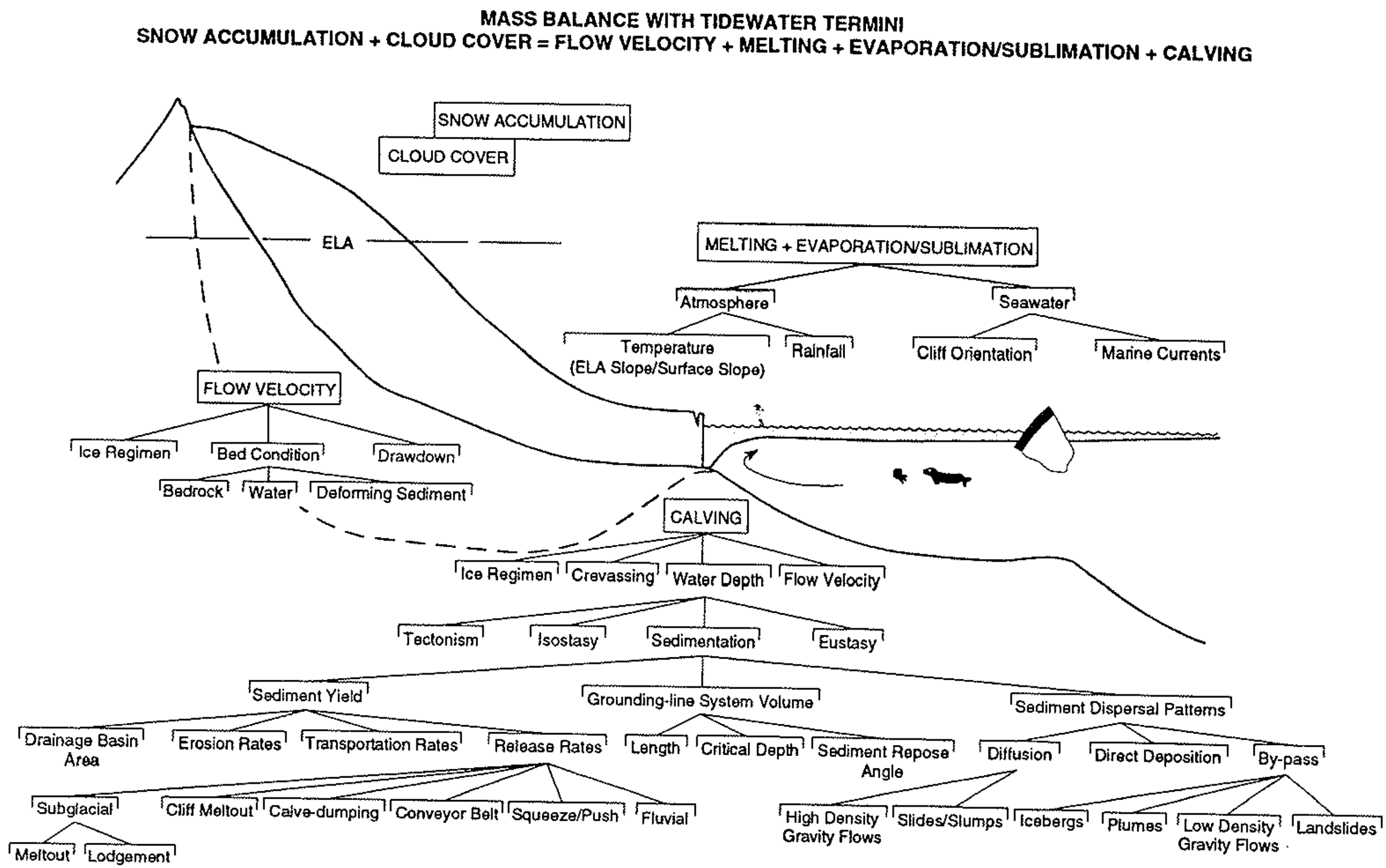


FIG. 14.13. Factors controlling mass balance of glaciers, with tidewater termini emphasizing sedimentological controls at grounding lines (not to scale). Many factors are dependent variables, but feedback loops are not shown for simplicity (after Powell, 1991; reprinted with permission of the author).

TABLE 14.3. Sediment Flux in Glacio-marine Environments

Region Setting	**Distance** from Glacier (km)	**Total Flux** (kg m^{-2} a^{-1})	**Terrigenous Flux** (kg m^{-2} a^{-1})	**Biogenic Flux** (kg m^{-2} a^{-1})
West Antarctica				
Peninsula Fjords*	5 km	2.285	n.a.	n.a.
	16 km	1.219	n.a.	n.a.
Ross Sea**				
SW Basin	n.a.	1.010	0.741	0.268
Sulzburger Bay	n.a.	0.633	0.619	0.013
East Antarctica				
Amery Trough***	70 km	0.471	0.118	0.353
Mertz-Ninnis Trough#	50 km	1.089	0.033	1.056
South East Greenland@				
Shelf Trough	n.a.	0.060-0.080	n.a.	n.a.
Baffin Island, Canada@				
Fjord	n.a.	0.600-1.300	n.a.	n.a.
Shelf Trough	n.a.	0.400	n.a.	n.a.
Shelf	n.a.	<0.050	n.a.	n.a.
Alaska, Glacier Bay†				
Fjord	0	490×10^6		n.a.
	.06	430×10^6		n.a.
	.1	180×10^6		n.a.
	.8	80×10^6		n.a.
	1.0	28×10^6		n.a.
	2.75	14×10^6		n.a.
	4.5	7.4×10^6		n.a.
	8.2	3.7×10^6		n.a.
	11	2.2×10^6		n.a.
	14.5	0.17×10^6		n.a.

*Domack *et al.* (in preparation)
**Ledford-Hoffman *et al.* (1986)
***Domack *et al.* (1991)
#Domack *et al.* (1989)
@Andrews and Syvitski (1991)
†Cai and Powell (unpublished data)

TABLE 14.4. Comparison of Holocene ice-proximal sedimentation rates in fjords or restricted basins (after Cowan and Powell, 1991).

Location	Rate	Distance from Grounding Line	Climatic Classification	Reference
McBride Inlet: Southeast Alaska (tidewater terminus)	200-2000 cm/a	within 1 km	temperate	Cowan and Powell, 1991
Hornsund: West Spitsbergen (tidewater terinus)	10-25 cm/a	within 10 km	subpolar	Görlich, 1986
Kongsfjorden: West Spitsbergen (tidewater terminus)	5-10 cm/a	within 10 km	subpolar	Elverhφi *et al.*, 1983
Mackay Glacier: East Antarctica (floating glacier-tongue)	1.8 cm/a	within 2 km	polar	Powell *et al.*, 1992
Prydz Bay: East Antarctica (inferred sub-ice-shelf or floating glacier-tongue)	0.21 cm/a	within 10 km	polar	Domack *et al.*, 1991

14.8. ADVANCE AND RETREAT OF MARINE-ENDING GLACIERS

Stability of marine-ending glaciers has been undergoing investigation since Mercer (1978) suggested that potential climatic warming (from greenhouse gases) may cause the West Antarctic Ice Sheet to disintegrate rapidly and cause a rise in global sea level.

While stability of marine tidewater termini, ice shelves and terrestrial glacial termini is ultimately a product of mass balance, other variables influence stability and hence the advance and retreat of marine termini (Fig. 14.22). Furthermore, some of these variables involve feedback mechanisms that are more complex in marine systems than in terrestrial systems. As with terrestrial termini, glacial mass balance for marine-ending termini is a function of snow accumulation rate, ablation rate, and glacial flow velocity. Considerations of accumulation rates are the same as for terrestrial systems (Chapter 4).

Non-climatic responses of marine termini have been noted for Pleistocene ice sheets (Mercer, 1968; Hillaire-Marcel *et al.*, 1981). Retreat of some temperate tidewater termini has been documented historically, and some, after reaching a position of maximum retreat, have stabilized or readvanced. Readvances have been attributed to: (1) decrease in size of the ablation area (Mercer, 1961; Post, 1975; Mayo, 1988), (2) increased precipitation at high altitude following a rise in the equilibrium line altitude (ELA) (cf. Field, 1947, 1979; Mercer, 1961; Goldthwait *et al.*, 1963), and (3) accumulation of sediment at the grounding line that decreases water depth and increases terminus stability (Goldthwait *et al.*, 1963; Post, 1975; Powell, 1984, 1991; Mann, 1986; Mayo, 1988; Warren and Hulton, 1990; Warren and Glasser, 1992).

Marine-ending glaciers differ from terrestrial glaciers in that they have longitudinal profiles near their termini that are concave-up rather than convex-up (Mercer, 1961). Once retreat is initiated by a rise in the equilibrium line, the effect can be cumulative because of the near-horizontal glacial profile which continues to lower by the 'draw-down' effect associated with fast flow (e.g. Hughes, 1973, 1983, 1987c).

Water depth at the grounding line is also considered important because of the potential calving instability at tidewater termini (Brown *et al.*, 1982) and rapid migration of the grounding line of floating termini (Thomas and Bentley, 1978). Changes in water depth can be affected by isostasy, eustasy and sediment accumulation/glacial erosion. In theory, a glacier could initiate its own retreat from a maximum advance by crustal loading as the glacier increased in thickness. If the advance is local and independent of global cooling, then a concomitant eustatic lowering of sea level may not occur. Once initiated, retreat may proceed because rebound is not sufficiently rapid. When rebound had decreased water depth glacial advance could then ensue.

Global cooling and glaciation could allow glaciers to advance onto continental shelves simply by eustatic lowering of sea level. If deglaciation of a sufficiently large ice sheet occurs and global sea level rises, then that can force marine-ending glaciers elsewhere to retreat.

The timing of eustatic and isostatic movements is important in the generation of packages of glaciomarine sediments because such movements are forcing functions that initiate or terminate packages (Powell, 1991). They also have the potential of creating their own feedback loops for glacial fluctuations. However, they do not necessarily influence the detailed sedimentary history within packages, which can be more a function of local and regional sedimentary processes (Menzies, 1995b, Chapter 11).

High latitude continental shelf and slope systems are relatively poorly explored but they potentially hold the key to documenting linkages between major ice sheet fluctuations and global climate changes inferred from other records such as deep-sea cores and ice cores. Some models have been suggested to formulate ideas for interpretations of sediment packages with multiple glaciations over high latitude continental shelves (Andrews, 1990; Boulton, 1990; Anderson *et al.*, 1991; Henrich, 1991; Powell, 1991; Syvitski, 1991; Bartek *et al.*, 1991). However, sedimentary systems tracts and their bounding surfaces need to be mapped in detail by high resolution seismic reflection surveys and sediment coring and drilling need to be performed in order to really test the models and ice sheet–climate linkages.

On a large scale, both glacial isostasy and eustatic sea level influence water depth at termini. Eustatic sea-level fluctuations are the largest global influence on local water depth and may be in-phase or out-of-phase with local advances and retreats of marine termini. The crustal depression phase of isostasy is a more significant effect on marine terminus stability than the rebound phase because rebound and its consequent decrease in water depth commonly appear to lag retreat.

It is the effect of rising eustatic sea level that has been suggested to initiate instability in a grounding-line/ice shelf system that results in rapid disintegration of a marine ice sheet, specifically, the West Antarctic Ice Sheet (Hughes, 1973; Thomas and Bentley, 1978). However, models that place more emphasis on longitudinal stresses and basal sliding indicate that marine ice sheets behave less catastrophically and are more stable to climatic and sea level forcings than the earlier models indicate (Alley and Whillans, 1984; Van der Veen, 1985, 1987a, b). The stability of floating portions of ice masses, that is, ice shelves and floating glacier-tongues, is also controlled by calving processes and those secondary factors which influence calving, such as sea ice, tides and storms. Recent studies suggest that climate warming has induced catastrophic break-up of the Wordie Ice Shelf by increased meltwater and hydrostatic pressure within surface crevasses (Doake and Vaughan, 1991).

On a smaller scale, water depth at a terminus is a function of sea-floor topography that can be modified by either glacial erosion (e.g. over-deepening) that creates deep water to force retreat, or glacial deposition (e.g. grounding-line systems of morainal banks, grounding-line fans, ice-contact deltas) at tidewater termini which causes shoaling (Powell, 1991) and decreases calving rate to enhance glacial stability (cf. Meier and Post, 1987). Although as yet unknown, there must be a critical water depth beyond which each tidewater terminus is unstable (e.g. Hughes, 1992b). Sediment accumulation may allow grounding-lines of floating termini to advance due to grounding-line wedge progradation as discussed above, although the relative change in water depth compared with tidewater termini is small due to the geometries at the grounding-lines.

Recent evidence from various sources points to instability of ice sheets on different time scales. The Laurentide Ice Sheet may have collapsed several times over 10 ka intervals to produce ice-rafted debris layers in the North Atlantic (Heinrich, 1988; Bond *et al.*, 1992; Broecker *et al.*, 1992; Andrews and Tedesco, 1992). The surface of the Greenland Ice Sheet appears to have experienced irregular temperature changes (interstadial episodes) over very short time intervals (500 to 2000 years) (Johnsen *et al.*, 1992). A new model of the West Antarctic Ice Sheet indicates that it could be self-regulated by the rates of erosion and transport of subglacial sediment over periods of the order of 100 ka (MacAyeal, 1992). On a longer time-scale, the West Antarctic Ice Sheet may have been absent about 2 Ma ago with open ocean in the area where it is today (Scherer, 1991). Likewise, even the East Antarctic Ice Sheet, which has been considered to have been stable for tens of millions of years, may have been absent as recently as 3 Ma ago (Webb *et al.*, 1984; Barrett *et al.*, 1992), although that is in debate (Marchant *et al.*, 1993).

The questions arise as to the linkages between ice sheets and climate: Which drives what? Our best long-term record of climatic change comes from deep-sea sediment cores and the Milankovitch theory has become the established model, using these data, to explain climatic change during the periods of time when Earth has been in 'icehouse' periods (Fischer, 1981) through its history. At present, this assertion may need re-evaluation, especially in light of the above information in combination with new data from caves (Winograd *et al.*, 1992) that tend to indicate that Milankovitch forcing is not the only process operating (Chapter 2). One way of testing these ideas is to document the glaciomarine record on high latitude continental shelves and slopes (Imbrie *et al.*, 1993).