Rafting in glacimarine environments

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Abstract: Sources of rafted sediment include glacial, colluvial, fluvial, aeolian and littoral settings. Processes of incorporation and rafting include active incorporation by freezing and passive loading of icebergs and sea ice, and rafting by vegetation, principally marine algae. Passively loaded sediments, being loose on the ice surface, are more easily lost from the raft and are more likely to form deposits of more than one particle at a time. The effectiveness of passive loading depends on sea ice being in contact with the shore in the spring, when colluvial and fluvial processes are most active, and during winter when aeolian processes occur. Actively incorporated sediment becomes concentrated on the surface of the raft by melting. Other factors that control distribution of rafted sediments include the strength and stability of the raft, its rate of deterioration, its ability to leave the source area, and its movement at sea.

Sediment released from the ice is deposited as single particles (referred to as a 'drop'), agglomerations of more than one particle ('dump'), frozen aggregates, and deposits from the melting of sediment-laden ice in contact with the sea floor. Characteristics that have been used to distinguish the mode of rafting include grain size, sorting, shape, surface features (especially striations), fabric, fauna associated with the environment of origin, distance of transport, and water depth. The most reliable in assessing origin from an iceberg is probably the presence of striations, although none is conclusive, and interpretations must be made cautiously.

Rafted sediment is a major component of the total sediment deposited in the glacimarine environment (Ruddiman (1977) estimates 2 × 10^{14} m^3 to the deep Atlantic during 3 Ma). Rafted sediment provides information about the oceanic and terrestrial environment. Many workers have used the presence of coarse particles in sediment as evidence of ice rafting and, therefore, of glacial conditions. In this paper 'glacial' refers to those environments influenced by seasonal or permanent ice cover. This definition allows consideration of both glacier and sea ice, the effects of which, in terms of rafting, are often indistinguishable.

The study of rafting of sediment to the marine environment is difficult because (1) the processes dictate unpredictable events especially in the short term, (2) some assumptions about the processes and forms resulting are not straightforward, and (3) observation and measurement of the process is difficult or impossible. This paper is a review of rafting in glacimarine settings; it draws on the established literature and is illustrated with material from studies especially in the eastern Canadian Arctic. It is necessarily qualitative.

Discussion is organized around a conceptual model (Fig. 1) indicating sources of material, processes and deposits (cf. Campbell & Collin (1958) Table 1). Each of glacial, periglacial and paraglacial (in the sense of Church & Ryder 1972) settings is an important source for rafted sediments. The rafting itself is carried out by icebergs, sea ice and vegetation. The role of sea ice and icebergs in rafting may be divided into active and passive processes. In the active role, sediment is incorporated into the ice by freezing or associated with flow. In passive ice rafting, the ice is simply the agent for distributing sediment placed on it by other processes.

The only significant agent of organic rafting in the glacimarine setting is marine algae of the littoral environment, although small amounts may be related to driftwood and marine mammals (Emery 1963), including polar bears (McKinlay 1976). Material rafted by human activity, especially clinker, scale and ash from steamships, is found with naturally rafted sediments (Huggett & Kidd 1983), and in the North Atlantic is the largest modern source of sediment (Kidd & Huggett 1981). The author's students have found clinker in samples from the sea floor near Pangnirtung, Baffin Island probably associated with the annual 'Arctic Patrol' (Marriott 1940), but it is easily recognized and forms a minor, recent component of the rafted sediment record.

Four types of sedimentary deposits are distinguished in Fig. 1. Following the terminology of Thomas & Connell (1985), a 'drop' is a single particle deposited from rafting, and a 'dump' consists of a number of particles laid down in a

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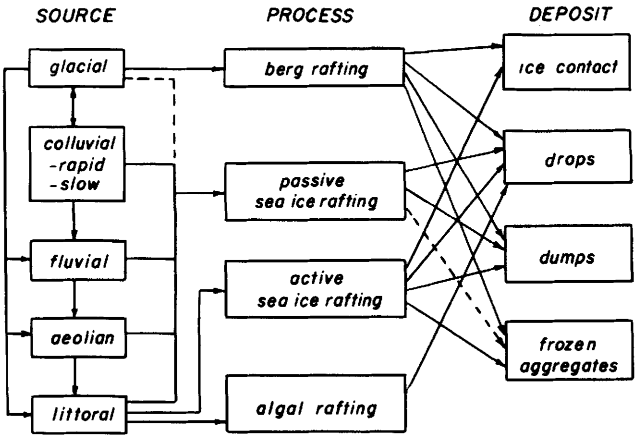

Fig. 1. Conceptual model of sources, processes and forms related to rafting in the glacimarine environment. Extraterrestrial and volcanic sources are not included because these materials are difficult to distinguish as rafted. Floating of particles on surface tension (Hume 1964; Syvitski & van Everdingen 1981) and rafting by trees and mammals (including humans) is not included because these are of minor importance in glacial seas.

single event. 'Frozen aggregates' are held together by interstitial ice and may be released by breaking apart of the ice raft (Huggett & Kidd 1983). 'Ice contact' sediments are deposited by ice in contact with the sea floor. This more general term is used here in preference to 'ice-berg till' (Dreimanis 1979) because the sediment is not associated directly with glacial deposition whether it originates from icebergs or sea ice.

Table 1. *Characteristics of rafted sediments*

Characteristic	Rafting agent		
	Iceberg	Sea Ice	Vegetation
Form	drops, dumps, frozen aggregates, ice contact deposits		mainly drops
Grain size	clay to boulder	clay to boulder maximum size about 4 m	pebbles some sand
Sorting	poor	poor to good	–
Shape	normally angular may be rounded faceted	angular to rounded	normally rounded
Surface features	may be striated	striations less likely	–
Fabric	random except in frozen aggregates and ice contact deposits		random
Fauna	absent	shallow water fauna attached or mixed with rock boring	short-lived holdfasts
Friable and weathered material	absent	may be present	absent
Distance of transport	more distally	more proximally	most proximally
Water depth	deep	shallow to deep	

Sources and modes of rafting

Glaciers and icebergs

Sediment is actively incorporated in glacial ice primarily as basal debris, or passively as a result of colluvial processes on valley sides above the glacier (Anderson *et al.* 1980). The amount of sediment incorporated at the base of the ice varies depending largely on the thermal regime of the glacier (Dowdeswell 1987; Dowdeswell & Dowdeswell 1989). Some cold-based glaciers (Shaw 1977) and those experiencing transition from warm to cold base along their length (Weertman 1961) incorporate a zone of up to several tens of metres of sediment-rich ice. Much of the sediment from warm-based glaciers and from floating ice shelves is lost by melting before icebergs calve (Piper 1976; Elverhøi & Roaldset 1983; Dunbar *et al.* 1985).

Many workers have observed that only a few per cent of icebergs carry visible sediment (Anderson *et al.* 1980), although Kindle (1924) cautioned about the reliability of these observations with the analogy of judging the number of barnacles on a ship by looking at the topsides. The question of the amount of sediment released from bergs is compounded by the near impossibility of direct measurement. However, rates of melt can be calculated (Huppert 1980; Dowdeswell & Murray 1990), and statistics are available on the decline in numbers of bergs away from source areas (Andrews & Matsch 1983), so that rates of deposition may be inferred (Dowdeswell & Murray 1990).

Fluvial and colluvial sources

Timing of delivery of fluvial and colluvial sediments to the shore is of critical importance in determining the potential for passive loading of sea ice. Even at high latitudes, these processes are well established by June. For example, in eastern Baffin Island, rockfalls, landslides and avalanches are most frequent at that time (Church *et al.* 1979; Fig. 2a). Similarly, slow colluvial processes (gelefluction) are most active during about the same period (Washburn 1967; Fig. 2b). In some cases slow colluvial processes may contribute sediment directly to sea ice, but it is more likely that they act as feeders to streams.

Stream flow records (for example, Fig. 2c) indicate that nival melt may begin in May, normally is greatest in June, and in the high Arctic may last into July (Woo 1986). Melt runoff usually lasts only a few weeks (Woo & Steer 1986). Glacial streams peak later in the summer (Church 1974; Gilbert *et al.* 1987).

The breakup of shorefast sea ice in the Arctic normally occurs between mid-June and mid-August after these peaks of sediment production. Except where the inflow of warm water or breakup of dirty ice on intertidal flats has

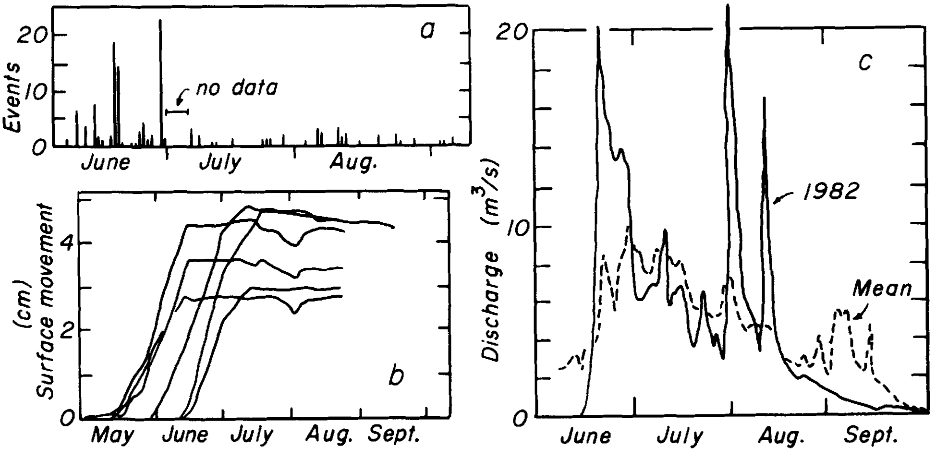

Fig. 2. (a) Number of events of rockfall at Ekalugad Fiord, Baffin Island, June 4 to September 9, 1967 (redrawn from Church *et al.* 1979). (b) Selected measurements of cumulative slope movement near Mesters Vig, northeast Greenland 1957 to 1961 (redrawn from Washburn 1967). (c) Daily mean stream discharge, Duval River at Pangnirtung, Baffin Island 1982 and mean for the years 1974 to 1982 (from the Water Survey of Canada and Gilbert unpublished data).

removed ice from near the shore, there is a reasonable chance in a given year that fluvial and colluvial processes may deliver sediment to the sea ice before breakup. Stefansson (1921) describes an accumulation of sediment ranging from mud to boulders of more than 45 kg on the ice of Peary Channel which he ascribes to landslides. The role of snow in aiding delivery of rock fall to ice is discussed by Ferguson (1970) and Luckman (1975). Howarth & Bones (1972) suggest that sea ice encourages colluvial processes by oversteepening talus slopes at water level as it removes sediment. Kindle (1924) and Keys (1978) describe the delivery of stream-borne sediment to sea ice on Greenland and Ellesmere Island. The largest amounts of sediment delivered to sea ice are from the major rivers of the U.S.S.R. (Fuchs & Whittard 1930) and the Alaskan north slope (Reimnitz & Bruder 1972), although most is lost as the ice melts in situ and little is rafted to sea.

Because the process involves coordination of supply of sediment before breakup of sea ice, spatial and temporal variations are great. It is probably impossible to assign a relative importance to these sources in comparison to loading that occurs by other means.

Aeolian sources

Kindle (1924) describes accounts of dust, sand and gravel carried by wind 'far out on the sea ice' (p. 258) (cf. Campbell & Collin 1958). Bentley (1979) and Barrett *et al.* (1983) conclude that aeolian sand deposited initially on sea ice is the major source of sediment on the floor of western McMurdo Sound.

Many surfaces in glacial regions have a cover of sparsely vegetated sediment and there is, therefore, abundant material for aeolian erosion. The major sources are ice-contact sediments exposed by glacial retreat, and fluvial and lacustrine sediments exposed during periods of low water (Molnia 1983; Migala & Sobik 1984). The presence of water, ice and snow is important in determining the availability of these sediments. During winter aeolian processes are probably most effective because winds are generally strongest and source areas are most exposed due to drying (McKenna Neuman & Gilbert 1986). Cold air temperature results in higher air density and, therefore, in lower critical erosion velocities. Except in temperate alpine environments, snow cover is not sufficient to retard erosion significantly, and blowing snow is itself an agent of erosion (McKenna Neuman 1987).

Except for fine grains (less than about 0.02 to 0.05 mm: Reineck & Singh 1975) which are transported in suspension, aeolian transport of sediment to sea requires a solid substrate of sea ice on which saltation, creep and movement by impact from other grains occur (Fig. 3). Thus, only where sea ice is in contact with the shore can coarse-grained aeolian sediment be carried to sea. At Pangnirtung Fiord, Baffin Island in May 1986 patches of silt, sand and gravel were found up to 5 km from their source on the sandur of Weasel River at the head of the fjord (Fig. 4). Much of the sediment (including cobbles to 128 mm diameter) occurred in pockets where it was trapped around sastrugi and snow dunes (cf. Bentley 1979; Nedell *et al.* 1987).

Under ideal conditions of unlimited supply of dry sand, aeolian rates of transport may reach 1 Mg h^{-1} per metre width at wind speeds of 16 m s^{-1} (Bagnold 1941). This may be compared with values in the range about 0.05 to 5 Mg h^{-1} bedload discharge of sand in water discharge of 0.1 to 1 m^3 s^{-1} per metre width (Blench 1969). Thus, at different times during one year especially on sandurs, aeolian transport of sand may rival fluvial transport of sand.

Melting of sea ice involves the establishment of pools, some of which melt through the ice to form conduits which drain the nearby ice (Barry & Jacobs 1978). It is possible that a significant amount of aeolian sediment deposited on the ice surface is released to the water beneath in a short time during this event, and that it may form a distinct annual layer in the glacimarine sediment (Fig. 5). If so, the layers are important stratigraphic markers in sediments in which varves are rare and in which bioturbation often destroys subtle laminations. If annual aeolian/rafted layers can be recognized, they offer the opportunity for study of year-to-year variation in climatic events over long periods.

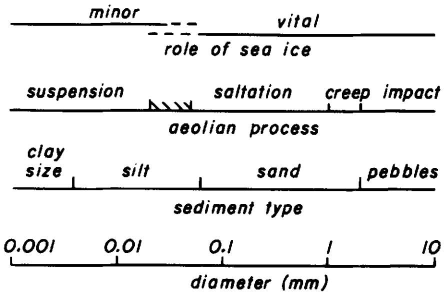

Fig. 3. Relation between grain size and mode of aeolian transport in moderate winds. For transport to sea of particles intermittently or continuously in contact with the surface, the presence of sea ice is vital.

Fig. 4. Aeolian sand and silt on the ice of Pangnirtung Fiord 5 km from the head and about 8 to 15 km from the source on the sandur of Weasel River, June 1986. Photograph by C. McKenna Neuman.

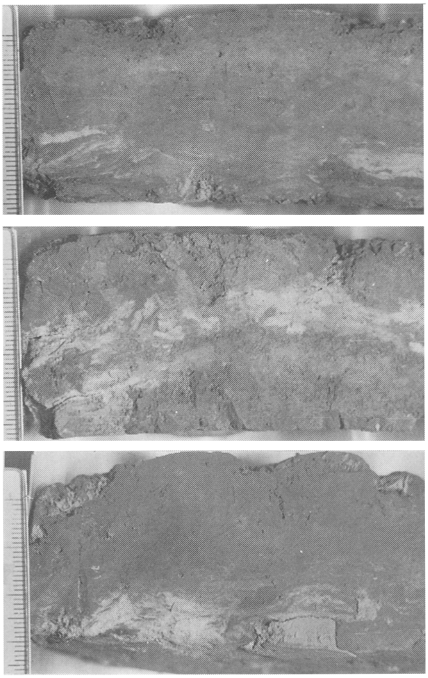

Fig. 5. Ekman grab samples from the floor of Maktak Fiord, eastern Baffin Island at 7 km (upper), 10 km (middle) and 14 km (lower) from a large aeolian deposit (Gilbert & McKenna Neuman 1986) at the head of the fiord. Scale in millimetre divisions. Quartz-rich white sand layers are interpreted as aeolian. Bioturbation following deposition has partially mixed sediments from above and below. Rates of aeolian sedimentation are about 8 to 17 kg m^{-2} a^{-1}.

Littoral environment

The littoral environment is the only one in which both active and passive processes of sediment incorporation into sea ice are important.

Active incorporation occurs in three ways. First, sediment may be frozen onto the bottom of sea ice as it rests on the sea floor. This occurs in shallow water (inshore of the isobath of maximum seasonal ice thickness) and is best developed where there are broad intertidal flats. Commonly, sediment accreted into sea ice by this means is found interlayered with clear ice (Fig. 6). The sediment-rich bands form when the ice is in contact with the sea floor at low tide, and the clear bands when the ice floats off at high tide. In a study of ice on intertidal flats at Pangnirtung, Baffin Island in June 1986 active incorporation accounted for sediment accumulation in the zone within about 200 m of shore of up to 670 g l^{-1} (Fig. 7).

The size distribution of the sediment incorporated by active freezing reflects that on the sea floor because little or no sorting occurs. Commonly, grain size varies from clay size to pebbles. Larger cobbles and boulders are also incorporated into ice as it breaks and refreezes around the stones especially during the highest tides (Fig. 8). The process is described in detail by Rosen (1979) and Lauriol & Gray (1980).

The second active process involves incorporation into the growing sea ice of fine-grained sediment suspended in the sea water. The sediment is scavenged from the water column or sea floor in shallow, turbulent areas by frazil and anchor ice formed in initial stages of freeze-up (Campbell & Collin 1958; Ackley 1982; Barnes *et al.* 1982; Dionne 1984; Reimnitz & Kempema 1987, 1988). The loosely held sediment is later frozen into the ice canopy to form a more-or-less uniformly darkened layer. In the Pangnirtung study (Fig. 7) this ice was observed to occur intermittently in the outer areas of the intertidal flats. Sediment concentrations were variable, but were generally less than several tens of grams per litre. Values of up to 1.6 g l^{-1} from sea ice on the Alaskan north slope have been measured by Barnes *et al.* (1982).

The third process of active incorporation involves the formation of anchor ice on the sea floor (Dayton *et al.* 1969; Reimnitz *et al.* 1987; Reimnitz & Kempena 1988). This is best developed in moderately shallow water where radiative cooling from the sea water and sea floor to a cold sky allows ice (often as frazil) to accumulate especially on rocky bottoms. Areas exposed at low tide may also be cooled sufficiently that, when the tide comes in, ice formation on the

Fig. 6. Ice on intertidal flats at Pangnirtung, Baffin Island during breakup, June 15, 1986, showing actively incorporated sediment by basal freezing and passively loaded sediment on the tops of the pans. Photograph by B.A. Crawford.

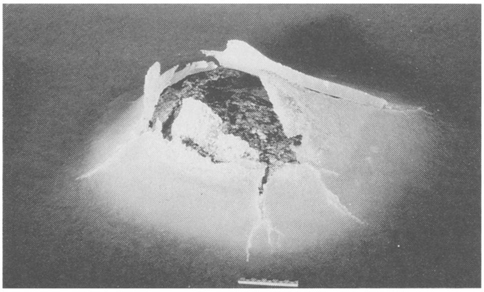

Fig. 8. Photograph of sea ice around a boulder during freeze-up, September 1983 in McBeth Fiord, eastern Baffin Island showing the method of incorporation of cobbles and boulders into sea ice. Two tidal cycles have occurred since the ice began to form. At low tide ice has broken over the top of the boulder, and is freezing around the sides. As the ice thickens, buoyancy increases and the boulder is lifted on each tidal cycle. Scale is 10 cm long.

substrate occurs. Fresh water inflow beneath the surface may also cause anchor ice formation (Sadler & Serson 1981).

As isolated patches of anchor ice grow, their buoyancy may be sufficient to lift sediments and plants from the sea floor. Scour by floating ice may also break a crust of anchor ice allowing it to float upward. These parcels become incorporated in the sea ice cover on the water surface (Reimnitz *et al.* 1987).

Passive loading of sea ice in the littoral environment occurs mainly in spring, as currents generated by rising tides force sediment eroded from the sea floor immediately below through cracks and melt channels onto the ice surface. Wind and waves may be locally important in passive loading at any time during the period of ice cover, but especially when ice is driven ashore, forming gouges and mixing with material from beaches (Andrews & Matsch 1983). However, unless this occurs at low tide, and the tidal range is sufficient to allow refloating, the forces involved usually ground the ice so securely that rafting of its load back to sea is precluded (Kovacs & Sodhi 1980).

In a study at Pangnirtung during breakup in

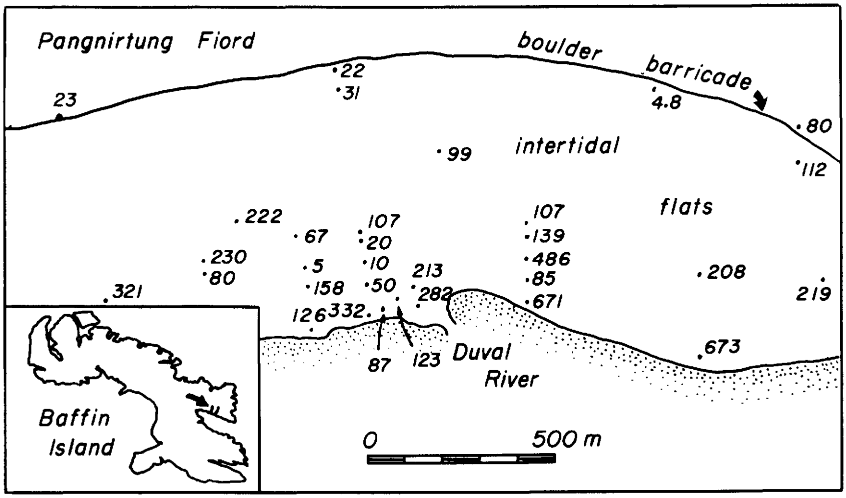

Fig. 7. Intertidal flats at Pangnirtung, June 1986 showing concentrations in g l^{-1} of actively incorporated sediment in sea ice.

June 1986, upwelling through the ice during each rising tide (range 4 to 7 m) allowed sand and small pebbles to be deposited on the ice surface (Fig. 9). During a period of two weeks up to 10 cm of sediment was deposited on the ice, occasionally as much as 1 cm per tidal cycle (cf. Dionne 1984).

Rafting by vegetation

In many glacimarine settings, especially of subpolar regions, benthic macroalgae are an important component of both the littoral and sea floor environments (e.g. Lee 1980). Algae with gas-filled vesicles in the thalli exert lift on the particles to which they are attached. For example in a study of the alga *Fucus vesiculosus* at Pangnirtung Fiord, Gilbert (1984a) showed that when a plant grows to about three times the weight of the stone to which it is attached, it is capable of floating the stone (Fig. 10).

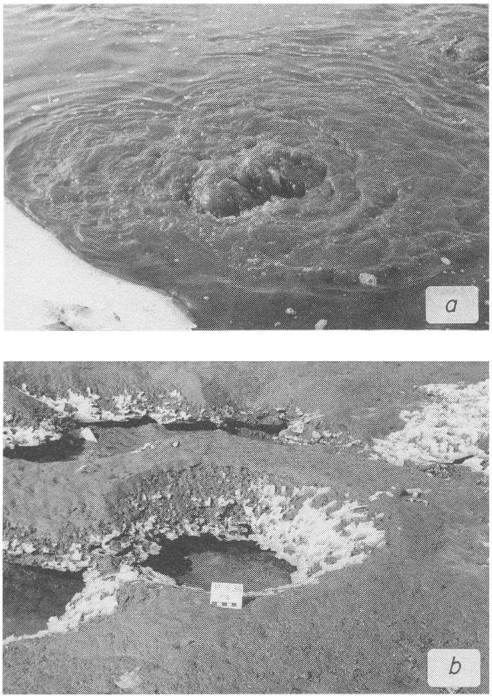

Fig. 9. Melt hole in sea ice on the intertidal flats at Pangnirtung, Baffin Island (a) during rising tide on June 6, 1986 and (b) with sand and gravel deposited as a result of passive loading by June 14. Maximum sediment accumulation is about 10 cm. Scale is 10 cm long. Photographs by B.A. Crawford.

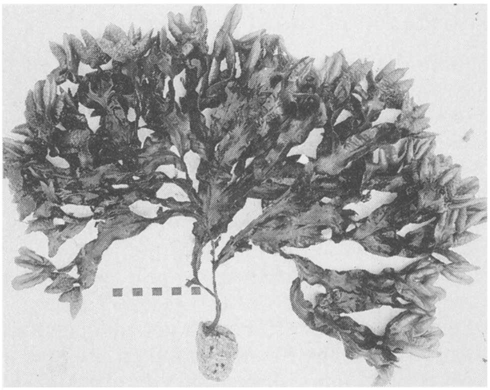

Fig. 10. Stone weighing 72 g with attached alga, *Fucus vesiculosus* sufficient to float it from Pangnirtung Fiord, Baffin Island. Scale is in centimetre divisions.

In addition, the algae are more easily frozen into sea ice around the plants, or as anchor ice accumulates on the plants at some depth below the water surface. In the latter case, the additional buoyancy may be enough to float a stone and attached algae. Beds of algae, some with attached stones and shells, are common in the sea ice of littoral areas.

It is impossible to assess the relative importance of ice rafting and algal rafting in the glacimarine record because both processes are variable in space and time, and are sufficiently unpredictable to make measurement difficult. As well, the evidence of algal transport, namely holdfasts and their marks (Huggett & Kidd 1983), is quickly obliterated by consuming organisms when the plant dies. However, as an indication of algal rafting, rafted particles larger than 16 mm from 49 samples from the floor of Pangnirtung Fiord, Baffin Island were examined. Evidence of *Fucus* holdfasts was found on 5 of 157 particles varying in diameter from 42 to 61 mm.

The presence of coarse particles rafted by algae in glacimarine sediments is important even if the actual amount of sediment moved is small. The occurrence of large particles in a fine-grained sediment far from the source is evidence of rafting and is used to define glacial environments. Emery (1963) shows that the zone of algal rafting overlaps much of the subpolar area of ice rafting in the Northern Hemisphere. Thus, caution must be exercised to distinguish ice rafting and algal rafting when environmental interpretations are being made (Menard 1953).

Distribution of rafted sediment

Rafting depends on a number of factors which must act together. They include:

the availability of material to be rafted;

the amount and distribution of sediment in and on the raft;

the strength and stability of the raft;

the rate of deterioration of the raft;

the freedom of movement of the raft;

patterns and persistence of winds, waves and currents to drive the raft (cf. Piper 1976).

Since most of these factors are independent of each other, their combination to produce a deposit of rafted sediment at any given place and time is normally completely unpredictable, although longer-term average conditions (especially oceanic circulation) produce recognizable patterns over large areas.

One of the critical factors is the relative amount of sediment and ice in the raft, because even a modest concentration of sediment is sufficient to prevent the raft from floating (Fig. 11). Some of the measured concentrations due to active incorporation on the intertidal flats at Pangnirtung (Fig. 7) substantially exceed these values, and ice submerged by its sediment load probably is common in nearshore areas. Single large particles may also be incorporated in sea ice (Fig. 8). Drake & McCann (1982) present calculations that indicate a clean pan of about 35 m diameter and 1 to 2 m thick may float a single boulder of 5 to 6 m diameter. Gilbert & Aitken (1981) show that only boulders smaller than about 4 m diameter are lifted by sea ice. The basal ice of glaciers may also carry concentrations of sediment more than sufficient to sink that part of the ice (Shaw 1977; Dowdeswell 1986; Dowdeswell & Dowdeswell 1989).

Submerged, sediment-laden sea ice or icebergs may be transported by bottom currents in the glacimarine environment. In the absence of direct observation, Josenhans & Woodworth-Lynas (1988) suggest that small furrows in the bottom sediments of the Labrador shelf may originate from movement of submerged ice. Rafting by submerged ice may distribute sediment in very different patterns than occurs in circulation driven by surface currents and winds. Submerged ice that comes to rest probably deposits sediments having the characteristics of frozen aggregates or ice contact deposits.

The distribution of sediment in the raft is important. Sediment actively frozen into bergs or sea ice is usually released by melting out one particle at a time. Thus, particles are much more likely to be deposited as single drop stones over a wider area and a longer time (Dowdeswell & Murray 1990).

Both actively and passively derived sediment may be concentrated on the ice surface by melting during summer. The relation between melt-

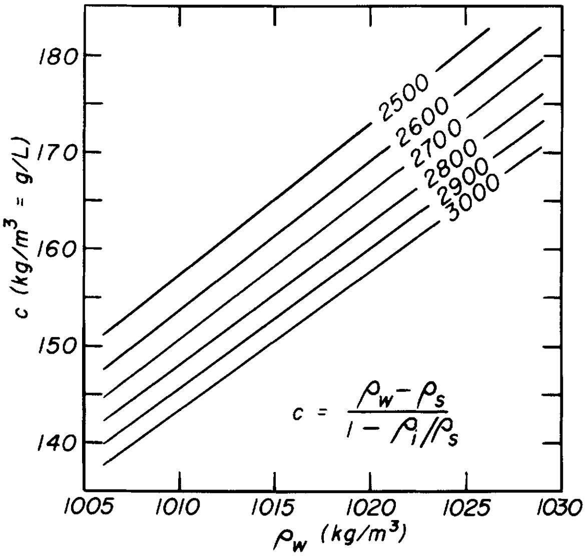

Fig. 11. Concentration, c, of sediment with density, ρ_s, from 2500 to 3000 kg m^{-3} in ice of density, ρ_i = 910 kg m^{-3} necessary to sink the ice in sea water having density, ρ_w.

ing of ice and its sediment content or cover is a complex one involving feedback between the energy balance (especially the net radiation component), characteristics of the sediment (thermal conductivity, shear strength, water content, albedo) and morphology of the melting forms (height, shape, roughness) (Drewry 1972). Even a light dust cover greatly reduces albedo (Woo & Dubreuil 1985). Single particles larger than about 40 mm diameter (depending on surface conditions) accelerate melting (Simmons *et al.* 1986). A complete cover of sediment increases the melting rate until its thickness is about 15 to 25 mm (Driedger 1981), with maximum increase when the cover is about 3 mm thick. Thickness of about 100 mm may decrease the melt rate by 4 to 5 times that of clear ice (Nakawo & Young 1981). The most effective particle size in reducing melt is in the range 0.6 to 5 mm (Drewry 1972). Smaller particles flow down slope more easily to expose ice beneath; larger particles allow freer circulation of air and thus heat through the sediment cover.

Sediment concentrated on the surface of sea ice may retard melting of pans in the nearshore until after breakup has occurred offshore. In some cases this may give these sediment-laden, nearshore pans greater opportunity to drift to sea.

New ice may form on the bottom of berg or sea ice, especially during winter in cold polar seas (Fuchs & Whittard 1930). Thus, although sediment does not move up through the ice, the melting above and freezing below eventually leads to a concentration of sediment on the ice surface. Rafts bearing sediment loads on their surface may be more unstable (the more dense sediment is above the centre of gravity). They are more likely to turn over and dump their loads in a single event (Vorren *et al.* 1983).

Ice rafting is most successful when the raft remains intact until it moves some distance from the source. Icebergs are efficient agents of transportation because of their size. Sea ice presents a different picture. In some cases unconsolidated slush ice that forms in early winter may transport large quantities of sediment before a solid landfast ice sheet forms (Reimnitz & Kempema 1987). However, the dispersal of stable pans produced during spring breakup is affected by several factors (Barry & Jacobs 1978). Most of the sediment load is obtained in the nearshore area where ice normally breaks up and melts first. The relatively stable barrier of ice offshore prevents the escape of sediment-laden pans from near shore. This mechanism has been proposed to explain the presence of boulder barricades at the seaward limit of intertidal flats (Rosen 1979). In the study at Pangnirtung, almost none of the ice in the intertidal area (Fig. 7) was observed to escape to sea for this reason. Domack's (1988) model of regions dominated by sea ice rafting near shore and by berg rafting at a much greater distance (especially at the limit of and beyond the sea ice zone) reflects this difference in behaviour of bergs and sea ice.

Deposition from rafts

Sediment is lost from the raft (1) by melting from the ice, (2) by washing or falling especially from unstable rafts, or (3) by sinking as a result of sediment concentration by melting or as a sediment-charged portion breaks off (Campbell & Collin 1958; Ovenshine 1970; Vorren *et al.* 1983). Loss from vegetation rafts occurs as a portion of the raft is damaged so that floatation is decreased or as the holdfast deteriorates.

Once free of the raft, single sediment particles fall through the water at rates that depend on (in decreasing order of importance) the size of the particle, its shape, its submerged weight, and the viscosity of the water which is a function of temperature. Various empirical and theoretical solutions (e.g. Dietrich 1982) produce a range of settling velocities depending on these characteristics (Fig. 12).

There are very few observations on the distance through which acceleration occurs (that is, the distance to reach terminal fall velocity). From values for larger particles given by Herbich (1981) and shown in Fig. 12, it can be seen that within a short distance even large particles reach terminal velocity, and the energy expended by a particle striking the bottom is independent of depth beyond this distance.

The behaviour of clouds of falling sediment as might be released from the surface of a raft during overturn has received some attention because of concern about the disposal of dredge spoil in deep water. Krishnappan (1975) presents a model for settling of a cloud of sediment which involves an entrainment phase during which the cloud spreads laterally and a settling phase below where velocity and size of the cloud remain constant. Results of calculations of these characteristics from Krishnappan's experiments with sand clouds are shown in Fig. 13. The effect of mixed grain-size dumps can be accounted for by a method of partitioning. Although the calculations are extrapolated from small-scale laboratory experiments, they indicate that the spreading may occur to considerable depth and that the area over which the

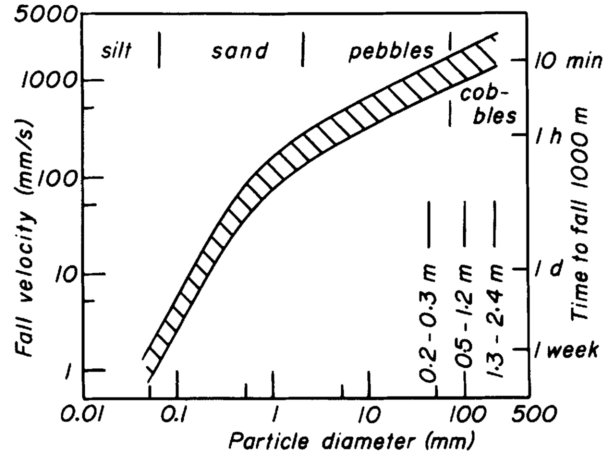

Fig. 12. Range in settling velocities for rounded to angular particles of density 2650 to 2900 kg m^{-3} in sea water at 0°C (after Dietrich 1982). Values in lower right are the distances to reach terminal fall velocity for particles accelerating from rest (Herbich 1981).

sediment spreads in deep water may be rather large.

Rafted deposits

Free-falling particles may significantly turbate soft sediments into which they fall (Thomas & Connell 1985). The amount of disturbance is a function of the kinetic energy of the particle on impact, the angle of impact determined by the slope of the sea floor and the fluttering motion of the particle during descent, the projection of the particle into the sediment based on its shape, and the strength of the host sediment. Impact turbation may rival bioturbation as a mechanism for mixing sediments. For example, measure-

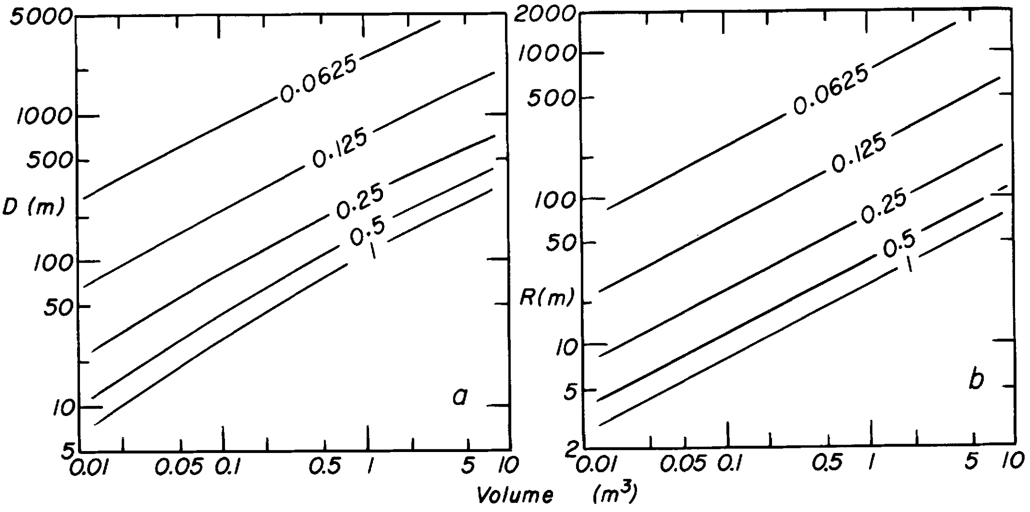

Fig. 13. (a) Depth, D, to which entrainment and spreading of settling clouds occurs and (b) radius, R, of the cloud at that depth for volumes of sediment having grain diameters between 0.0625 and 1 mm. Calculated from analysis by Krishnappan (1975). In water depths greater than D, R, is approximately constant. In water depths less than D, the cloud has radius, $r = Rz/D$ where z is the depth in metres.

ment of the concentration of drop stones larger than 2 mm in Cambridge Fiord, northern Baffin Island, indicates as many as several hundred particles per litre at some sites (Fig. 14). While fine laminations observed in cores from the same area are only slightly disturbed by the smaller stones, particles larger than about 8 mm have sufficiently mixed the fine sediments that laminations may be obliterated.

A dump of sediment of mixed grain size spreads over a wide area (Fig. 13) and separates by grain size to form a graded bed. This deposit might be mistaken for a turbidite, even when seen in an extensive exposure. Lateral separation by grain size also occurs if transport by currents occurs during descent (Krishnappan 1975), thus concentrating coarse sediments near the dump site and fine sediments distally (Hoskin & Valencia 1976; Elverhøi & Roaldset 1983). As a result, a dump may appear similar to the distally fining deposit from a turbid plume issuing from a glacier or stream.

Rafted sediment deposited as a frozen aggregate is difficult to identify because the ice has normally melted long before examination of the feature in a core or section. Indirect evidence includes the occurrence of sedimentary structures preserved when the sediment was incorporated into the raft. For example, active incorporation of beach sands by sea ice may involve little disruption (Medcof & Thomas 1974). Other indicators are the presence of a sharp and irregular boundary between the aggregate and the host sediment, and the presence of an open structure among the particles, previously filled with ice (Thomas & Connell 1985).

The blocks of well sorted medium sand shown in Fig. 15 suggest deposition as frozen aggregates. Dark layers are concentrations of heavy minerals, especially garnet. Similar laminations are common on beaches in the region. The two

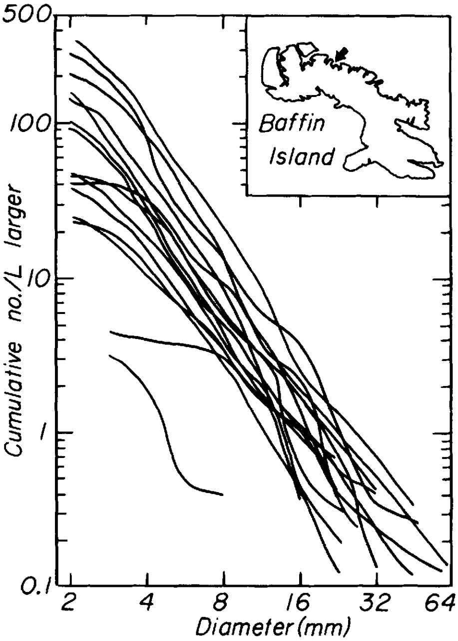

Fig. 14. Number of drop stones in the sediments of Cambridge Fiord, northern Baffin Island. Samples were recovered along the length of the fiord with a vanVeen grab sampler and washed on a 2 mm sieve to recover the drop stones (from Gilbert 1984b).

Fig. 15. Frozen aggregates of beach sand at 5.36 m below the sediment surface (top of photograph) in a piston core recovered 26 km from the head of Cambridge Fiord, northern Baffin Island. (Scale is in 2.5 mm divisions.)

blocks appear to have been deposited at the same time from the same frozen piece.

Distinguishing rafting agents

Because many authors have used the presence of rafted debris as an indicator of glacial history, separation of modes of rafting is important. A list of the major characteristics used to distinguish rafting agents is given in Table 1. Both bergs and sea ice deposit drops, dumps, frozen aggregates and ice-contact sediments, although large dumps may be more characteristic of bergs.

Grain size is a poor indicator of the rafting agent (Fig. 16). It has been suggested that fine-grained sediments are absent in sea-ice rafted debris (Schermerhorn 1974), but both active and passive processes may involve loading with considerable amounts of silt and clay-sized particles (Barnes *et al.* 1982). Very large particles (more than about 4 m diameter) are most likely incorporated and transported only by icebergs (Snelgrove 1984). Vegetation rafting is confined mainly to pebbles, although some sand may be involved. Berg rafted sediments are more likely to be poorly sorted with up to one half or more mud (Hoskin & Valencia 1976; Anderson *et al.* 1980), although sediments of colluvial origin may also include a wide range in a single event (Stefansson 1921).

Shape may be used to suggest origin, but it is not definitive. Well rounded particles are typical of beach and nearshore areas, but in polar environments frost shattering, protection from wave action by ice, and isostatic adjustment to expose new material mean that sedimentary particles may be less rounded (Nichols 1961; McCann & Owens 1969). As well, sediment loaded by colluvial processes is normally angular (Luckman 1975). Angular particles are typical of glaciers, but rounding does occur (Snelgrove 1984; Dowdeswell & Dowdeswell 1989), especially during transport by water within or on the ice.

Fabric is useful for separating dumps and drops from subglacial till (Domack & Lawson 1985). However, it cannot distinguish ice contact sediment from subglacial till, or iceberg rafting from sea ice rafting.

The presence of striations and facets on particles is probably the best evidence of glacial origin (Huggett & Kidd 1983), although sediment transported above the base of the glacier does not become striated (Domack *et al.* 1980). On the other hand, the action of ice push on shores is known to create striations indistinguishable from those made by glaciers (Nichols 1961; McLellan 1971; Dionne 1985).

The presence of shallow water macrofauna within rafted sediments (Reineck 1976) or evidence of shallow water organisms, such as

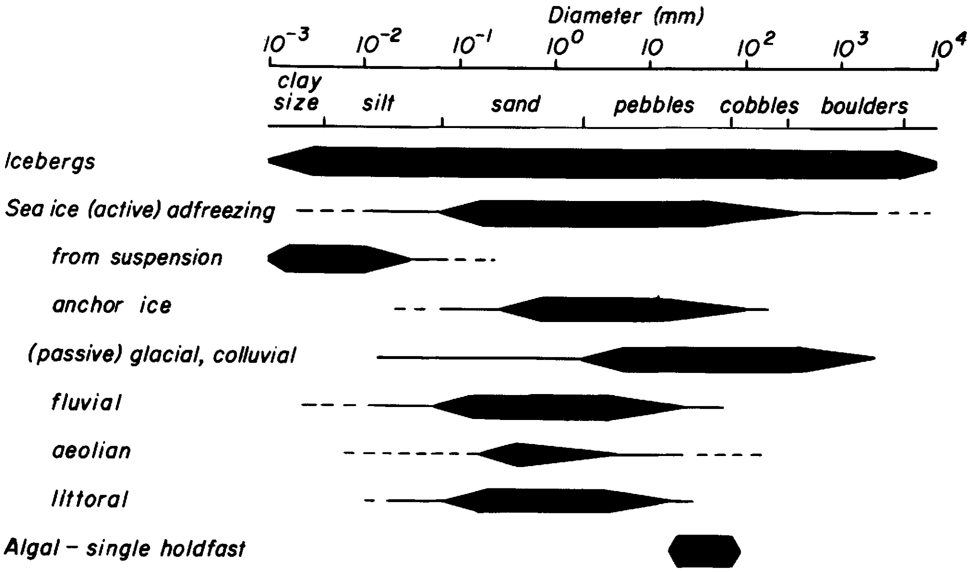

Fig. 16. Approximate range in grain size of rafted sediments.

rock borings (Huggett & Kidd 1983) is an indication of sea ice rafting. However, the absence of these is not a good indication of berg rafting because it is also absent in sea ice loaded by colluvial, aeolian and fluvial processes.

Sea ice is likely to deposit its sediment closer to the point of origin. Because the rafts are smaller, they melt more quickly and are more easily upset (Schermerhorn 1974; Spjeldnaes 1981; Domack 1988). Nevertheless, there are large areas of the sea over which both range together and, except in shallow water where bergs are unable to float, distance of transport may be an unreliable indication of source.

It is clear that sand and gravel particles deposited in deep water at some distance from source areas, and not associated with gravitational processes, are rafted. Even if it is possible to eliminate vegetation rafting by sediment size (Fig. 16) or by the amount of material involved, it is often very difficult to separate sea ice and iceberg rafting conclusively (Spjeldnaes 1981; Reimnitz & Kempema 1988). Conclusions about the extent of glaciation based on interpretation of rafted sediment must be made cautiously.

Conclusions

1. Rafted sediments are a major (in some cases the largest) component of glacimarine sediment. They provide information on the extent of glacial processes and the characteristics of glacial seas.

2. The major agents of rafting are icebergs and sea ice, although algal rafting may be significant locally. Drop stones from algae may not be distinguishable from ice rafted sediment.

3. The major sources of rafted sediments are glacial erosion, colluvial, fluvial and aeolian processes and the littoral environment.

4. Loading occurs by active freezing of sediment to sea ice (as a result of contact with the sea floor, scavenging by frazil ice or formation of anchor ice) or in glacier ice associated with flow, and by passive loading onto the ice surface. Passive loading occurs when sea ice persists unbroken from shore during the period in which loading processes are most active (during nival melt for colluvial and fluvial processes and throughout winter for aeolian processes).

5. Since actively incorporated sediment is distributed through bergs or sea ice, it is more likely to be freed as individual drop stones melting out one at a time, unless it has been concentrated on the ice surface by melting. In this case it behaves as passively loaded sediment, more likely to be dumped *en masse* as a result of slope failure, runoff or upset.

6. Sediment dispersed through berg or sea ice may be concentrated on the ice surface by melting from the top, in some cases while freezing is occurring at the base. Rates of melting depend in part on the ice cover, with accelerated melting caused by sediment load until a thickness of about 15 to 20 mm is reached, after which melting is retarded.

7. Dispersal of sand and mud from a single dumping event may be widespread as a result of the behaviour of the cloud of falling sediment, and if settling velocity is small enough to allow currents to spread the falling material. Resulting deposits may be confused with those created by turbid plumes in underflow or overflow.

8. Deposition of non-buoyant frozen aggregates may occur. As part of this process, the transport of slightly negatively buoyant frozen blocks along the sea floor may raft sediment long distances in patterns different from those at the surface. Preservation of sedimentary structures in frozen aggregates may offer an indication of the origin and mode of transport.

9. Distinguishing the agent of rafting from the nature of the sedimentary deposit is often difficult. The most reliable indication of iceberg rafting is the presence of striated facets on large particles. Other indicators that may be useful in particular circumstances include particle shape, associated shallow water fauna, presence of friable or weathered material, and the depth of water. Just as the presence of rafted debris may not be directly correlated with glacial events, its absence may not be related to non-glacial environments.

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