Annales Societatis Geologorum Poloniae (1986), vol. 56: 433–477
PL ISSN 0208-9068

GLACIMARINE SEDIMENTATION OF MUDS IN HORNSUND FJORD, SPITSBERGEN

Krzysztof Görlich

Institute of Geological Sciences, Polish Academy of Sciences, ul. Senacka 3, 31-002 Kraków

Görlich, K., 1986. Glacimarine sedimentation of muds in Hornsund fjord, Spitsbergen. *Ann. Soc. Geol. Polon.*, 56: 433–477.

Abstract: The Hornsund fjord is fed with sediment-laden meltwaters by eight major tidewater glaciers. Two glacimarine mud facies are described in bottom sediments of Hornsund: (1) laminated mud and (2) homogeneous-bioturbated mud. The laminated mud facies is ice-proximal and is deposited in a ponded mode at high accumulation rates ($\geqslant$ 5 cm/a). It reveals organic-debris-induced lamination and little or no fauna. The clay mineral assemblage in this facies preserves compositional features typical of suspended load discharged from a given source glacier. The homogeneous to bioturbated mud facies is ice-distal and is deposited at relatively low accumulation rates ($\leqslant$ 5 cm/a). It has no organic-induced lamination and is rich in fauna. The composition of clay-mineral assemblage in this facies is not directly related to any particular source of suspended load, due to the conspicuous effects of blending and differential settling of suspension along transportation path. In the homogeneous to bioturbated mud facies, slight seaward coarsening is observed, accompanied by enhanced grain-size bimodality. The latter is due to relatively increasing contribution of ice--rafted debris.

The clay minerals are inherited almost unchanged by bottom sediments from the rocks which crop out in the glaciers' catchments. Lithological indices are introduced to express this property of the glacial environment. Settling of clay suspension is electrochemically coerced by sea water. Flocculation is not preferential toward any specific clay-mineral species or grain size. The suspended load undergoes gravitational sorting which affects also clay fraction. 1Md illite and diagenetic chlorite are preferentially carried in suspension out of the fjord.

Seaward decay of sedimentation rate is exponential but, for individual tidewater-glacier sources, it occurs with different steepness of the rate-distribution curve. This is attributed to compositional and textural differences between the initial suspended loads, and to different values of vertical salinity gradients within near-source parts of the basins. Variable but always large vertical salinity gradients in front of tidewater glaciers enforce mass settling of suspension. This phenomenon is much less pronounced in fjord-type estuaries of glacier-fed rivers due to diffuse halocline and intensive vertical mixing. Hence, efficiency of entrapment of suspension in the near-source zone is the greatest in tidewater-glacier systems.

Key words: Clay minerals, glacimarine mud, suspension settling, tidewater glacier, Spitsbergen.

Manuscript received January 1986, accepted February 1986

INTRODUCTION

Glacimarine sediments received much attention in the last decade because new research vessels, coring devices and analytical methods made them readily accessible for investigations. These sediments constitute a considerable portion of the present-day oceanic sediments and ancient rock sequences. So far, glacimarine sedimentation was studied using methods of facies analysis (e.g., Kravitz, 1976; Anderson *et al.*, 1980; Elverhøi *et al.*, 1980; Elverhøi *et al.*, 1983; Elverhøi, 1984; Gilbert, 1983; Mackiewicz *et al.*, 1984; Powell, 1981) or by applying mineralogical or physico-chemical approach (e.g., Syvitski & Murray, 1981; Syvitski & Macdonald, 1982; Molnia & Hein, 1982; Elverhøi & Roaldset, 1983; Farrow *et al.*, 1983; Osterman & Andrews, 1983).

The above studies yielded a general picture of regional distribution of glacimarine facies in the present-day seas and resulted in comprehensive facies models for glacimarine systems (Anderson *et al.*, 1983; Molnia, 1983; Powell, 1984; Eyles & Miall, 1984). It was found that contribution to marine sediments from meltwater input is negligible in Antarctica, whereas it dominates in Subarctic environments (Anderson *et al.*, 1983). From the latter observation it is implied that glacially controlled sedimentation delivers today much more fine sediments to the oceans in northern high latitudes than it does in Antarctica (Molnia, 1983). It is, however, doubtful whether available data on glacimarine sediments substantiate such general opinions.

In order to explain the observed regularities of clay mineral distribution in northern oceans (Biscaye, 1965; Griffin *et al.*, 1968), more mineralogical studies in the near-source zones are needed. Few such studies have been reported so far (Molnia & Hein, 1982; Osterman & Andrews, 1983).

Too little attention is being paid to those specific features of tidewater-glacier sedimentation which make this process hardly comparable with the marine sedimentation from glacier-fed rivers. In Subarctic environments both these processes are active but in regionally different proportions. For a proper evaluation of the role of meltwater sources in determining of the fine-sediment budget of the high-latitude ocean, it is indispensable that qualitative and quantitative distinction between glacier-fed river and tidewater-glacier sources is made. To the author's knowledge, no such estimates have been attempted so far.

In the present paper, sedimentological data from the Hornsund fjord in Spitsbergen are reported. They illustrate those phenomena within tidewater-glacier sedimentation which strongly bear on the quality and quantity of suspension carried out to the shelf and oceanic basins.

REVIEW OF THE STUDIED PROBLEMS

Intensity and selectivity of suspension settling. Sedimentation of terrigenous fine particles in the nearshore zone and over the shelf is governed mainly by gravitational settling within a given dynamic pattern of marine waters. Suspension sedimentation is also controlled by the (selective?) destabilizing action of salinity barrier at the input point of a freshwater source to marine basin. Due to the presence of the salinity front, most of the clays and particulate organic matter are entrapped within near-source area by mass settling of flocculated colloidal suspension from freshet. An attempt is here made to correlate the intensity of settling with vertical salinity gradients in the near-source basins. This is done in order to show that such dependence constitutes a principal factor distinguishing tidewater-glacier sediment inputs from the glacier-fed riverine ones. It will be discussed also, whether the observed differential settling of suspension is controlled by grain-size distribution or rather it is of an electrical nature.

Sediment provenance and mineral inheritance. The provenance of sedimentary material can be readily recognized in bottom sediments of fjords (e.g., Syvitski & Macdonald, 1982) and northern oceanic shelves (Piper & Slatt, 1977; Molnia & Hein, 1982; Kravitz, 1983). The glaciated drainage basins tend to deliver sedimentary material which is transmitted to bottom sediments in a mineralogically unaltered form. Proximal muds receive fine particulate matter that can be easily matched with the composition of source rocks. This is hardly the case for distal muds (*cf.* Molnia & Hein, 1982), because of blending and maturation of clay suspension.

In this paper, distribution of mineral provinces in the bottom sediment of Hornsund is described and explained in terms of mineral inheritance and differential settling.

Organic-induced lamination. The lamination in glacimarine sediments is predominantly of textural nature and originates from gravity flows, meltwater underflows and meltwater overflow fluctuations. This lamination is thus generally aperiodic in glacimarine muds. Grain-size periodicity found by Domack (1984) in Pleistocene glacimarine sequence, and those discussed by other authors, are essentially limited to proximal settings. Periodicity of diurnal "cyclopel" sequences described by Mackiewicz *et al.*, (1984) from relatively distal muds, seems to hold only for areas south of the Arctic Circle where daily temperature fluctuations are pronounced, and only in instances where ice-proximal conditions extend axially along a narrow sedimentary basin. The settling of fines reveals, however, some periodicity that is at least partly independent of short-term fluctuations in intensity and dispersal of turbid plume. This periodicity is reflected by dark monosulphidic lamination in muds. Such lamination is common in Spitsbergen fjords. Elverhøi *et al.*, (1980) link its origin with a time-lag between the maxima of organic production (spring) and clastic sedimentation (summer). Due to its organic-monosulphidic nature, this lamination is liable to early-diagenetic decomposition and disappearance. The present paper shows that formation and preservation of this lamination is related to sediment accumulation rate. It is suggested here also that organic-monosulphidic lamination may be traced in sediment column, even after oxidative decomposition of organic remains and iron monosulphides. This can be done by detecting peaks of concentration of fine-crystalline siderite that forms abundantly in the originally organic-rich layers.

Glacimarine mud facies. Several mud facies were distinguished in Spitsbergen fjords by Elverhøi *et al.* (1980, 1983). These facies differ in texture and degree of homogeneity or bioturbation. Two of these mud facies roughly correspond to those described in the present paper. The designations for the two facies introduced by the above authors, namely distal homogeneous mud and basin mud, will not be used in the present paper. These names are confusing as regards the term "distal" and inaccurate with respect to homogeneity. Instead, the laminated mud facies and homogeneous to bioturbated mud facies are described in Hornsund. Sets of diagnostic and genetic features for these two facies are presented.

SETTING OF THE STUDY AREA

MORPHOLOGY

The Hornsund fjord (further on referred to as Hornsund) is the southernmost major fjord of Spitsbergen (Fig. 1). Hornsund is about 27 km long. It opens to the Greenland Sea without a sill in the bottom. The shelf in the forefield of Hornsund

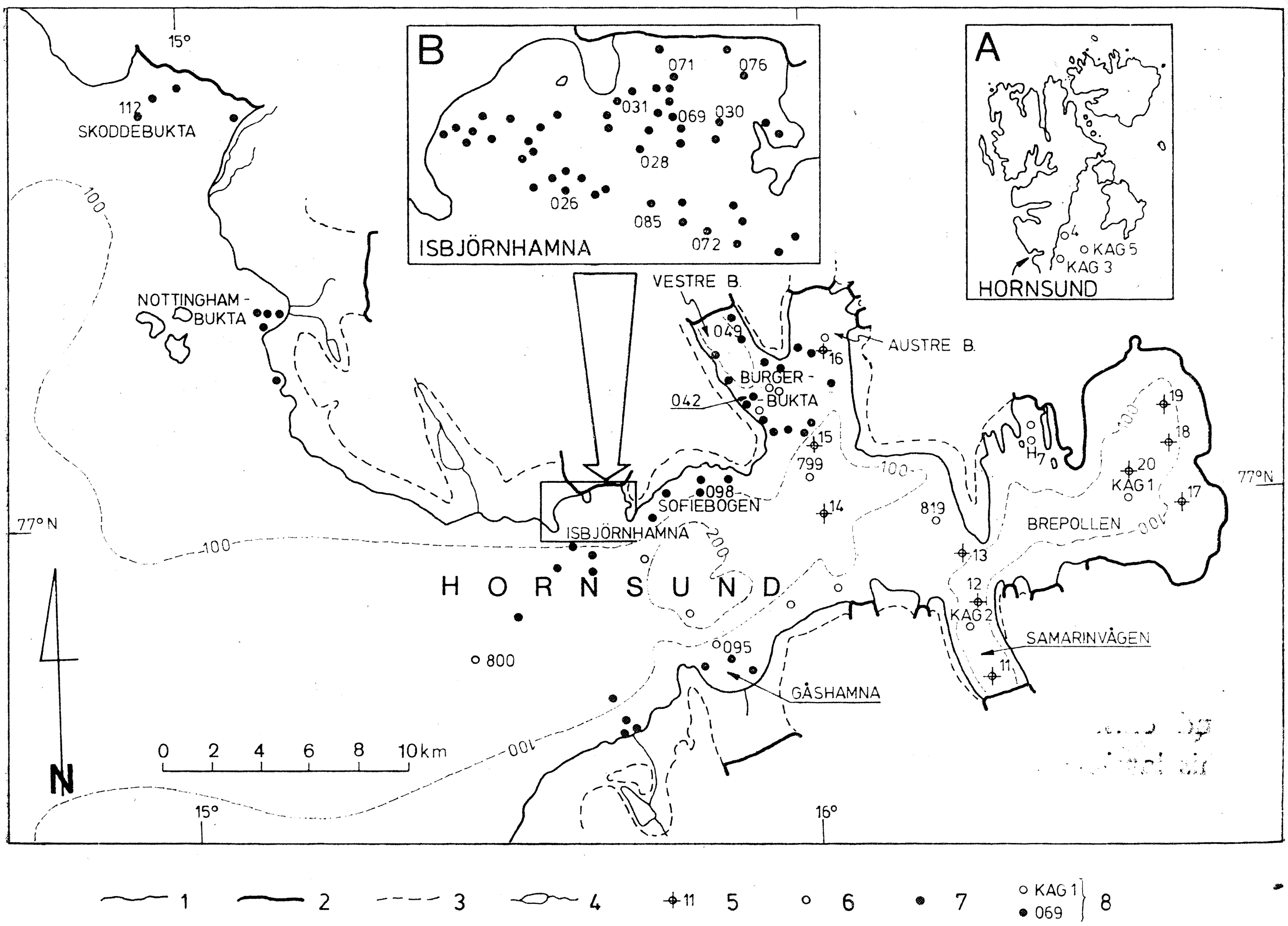


Fig. 1. Sampling sites in Hornsund and on south Spitsbergen shelf. *A* — Spitsbergen with sampling sites in Storfjorden (KAG 3 to KAG 5); *B* — Isbjörnhamna bay; *1* — shoreline; *2* — glacier margins in sea and on land; *3* — 100 m contour line; *4* — rivers and lakes; *5* — coring sites; *6* — samples recovered by J. M. Węsławski, C. Filipowicz, M. Zajączkowski and W. Moskal; *7* — samples recovered by the present author; *8* — sample numbers

is some 70 km wide, with the depths ranging between 100 and 150 m. Within this segment of the shelf, there is a bottom elevation (Hornryggen) in the shape of terminal moraine, attaining 60 m b.s.l. The elevation, in its southern part is cut by longitudinal trough (Hornsunddjupet) down to 250 m in depth.

There are five secondary bays (subbasins) within the fjord. These are: Brepollen, Samarinvågen, Austre and Vestre Burgerbukta, and Isbjörnhamna. All these bays bound on tidewater glaciers. The bays are from 55 m (Isbjörnhamna) to 180 m

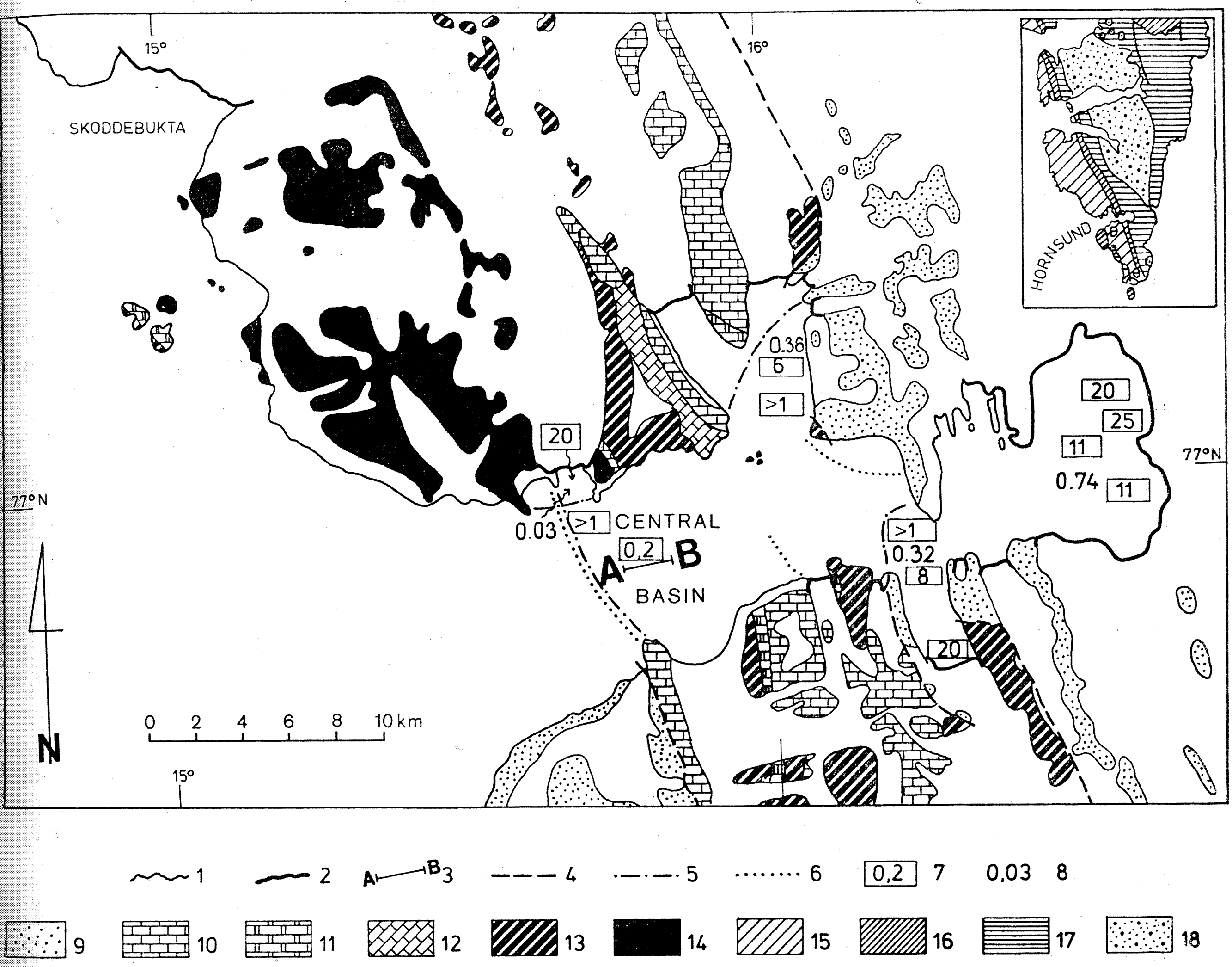


Fig. 2. Generalized lithologies of exposed bedrock in Hornsund catchment and stratigraphic sketch of south Spitsbergen. Indicated are: sills within the fjord, average sediment accumulation rates and values of DR/PM ratios (*cf.* text). Lithologic data are compiled from: Birkenmajer (1960, 1964, 1975, 1977, 1978a, 1978b), Radwański & Birkenmajer (1977), Siedlecka (1968), and Smulikowski (1965). Structural features of the bottom according to seismoacoustic profiling by W. Kowalewski, S. Rudowski & S. M. Zalewski (1983, 1985 — personal communication). *1* — shoreline; *2* — glacier margins in sea and on land; *3* — position of profile in Fig. 8; *4* — boundary delimitating occurrence of metamorphic, metasedimentary and Cambro-Ordovician carbonate rocks in the Hornsund catchment; *5* — morainic sills; *6* — submarine bedrock ridges; *7* — sediment accumulation rate; *8* — DR/PM ratio (see text); *9* — siliciclastic rocks; *10* — limestones; *11* — dolomites; *12* — marbles; *13* — phyllites; *14* — metamorphic rocks of Hecla Hoek Succession; *15* — (inset map) Precambrian through lower Palaeozoic; *16* — upper Palaeozoic; *17* — Mesozoic; *18* — Cainozoic (Tertiary); white areas on land are covered by ice or Quaternary deposits

(Vestre Burgerbukta) deep and are separated by sills from the main trough of the fjord (Fig. 2). The maximum depth of the central basin of Hornsund is about 250 m b.s.l.

GEOLOGY

The outline of geological structure of south Spitsbergen is shown in the inset map of Figure 2. From west to east, Hornsund intersects strata ranging from the Proterozoic Isbjörnhamna Formation through Lower Cretaceous formations. The glacial catchment area encompasses also much younger strata including the Tertiary ones. Quaternary cover of the area is confined to tills of lateral and terminal moraines (mainly of the Holocene glaciation), some outwash deposits, and raised glacimarine sediments.

The sketch in Figure 2 shows dominant lithologies in the area. The high- to low--grade metamorphic rocks of the Hecla Hoek Succession crop out in the western part of the northern coast of Hornsund and in the central part of its southern coast. These are: ortho- and paragneisses, amphibolites, phyllites, muscovite, biotite and chlorite schists, quartzites, and marbles (Smulikowski, 1965). Cambro-Ordovician marbles, limestones and dolostones occur on both coasts of the central fjord (Birkenmajer, 1978a, 1978b; Radwański & Birkenmajer, 1977). In the eastern part of the Hornsund area, upper Palaeozoic to Lower Cretaceous sedimentary rocks are exposed. These comprise conglomerates, sandstones, shales and calcarenites to calcilutites (Birkenmajer, 1964, 1975, 1977, 1984a, 1984b; Siedlecka, 1968). The upper Palaeozoic and Mesozoic sedimentary rocks of the Hornsund area, to the west of the fjord head, were intensely folded in the Carboniferous and Tertiary.

PHYSICAL OCEANOGRAPHY

According to Swerpel (*in print*) and Węsławski *et al.* (1985), three dominant water masses interact within Hornsund and south Spitsbergen shelf: (1) Fjord Surface Water (desalted to 28‰, temperature between 0.5 and 3°C, conventional density 22 to 26) which forms the uppermost, on the average 10 m thick layer, (2) Atlantic Coastal Water (salinity 33.5 to 35‰, temperature 0.5 to 5°C, density 26 to 28) which occurs at intermediate depths between 20 and 100 m, and (3) Atlantic Core Water (salinity above 35‰, temperature over 5°C, density about 28).

Measurements reported by Urbański *et al.* (1980) and Görlich & Stepko (*in print*), indicate that a dense cold bottom water is formed during winter freezing in the troughs of the sheltered bays of Hornsund (Brepollen and Vestre Burgerbukta). This water is exchanged only in the years of abnormally strong inflows of the Atlantic Core Water, as it happened in 1984 (Węsławski *et al.*, 1985).

The structure of water masses bears only indirectly on the course of sedimentation, namely through the dynamic stability of water column and through the abundance and composition of plankton related to the type of water mass (see next paragraph). The enhanced water-column stability during the summer period, hinders mixing and favours lateral dispersal of suspended load. In autumn and winter, this stability is destroyed and deep mixing takes place.

The main input of fresh water and sedimentary material to the fjord follows from eight major tidewater glaciers. It is impossible to evaluate the freshwater discharge in Hornsund due to scarcity of pertinent hydrological data. The seasonal rhythm of glacial discharge is obvious. However, there is some uncertainty concerning winter standstills. In spite of predominately "warm" thermal status of the debouching glaciers, the temperature and salinity measurements (Görlich & Stepko, *in print*) did not unequivocally ascertain the presence of submarine subglacial outflow during winter period in front of the Hansbreen glacier in Isbjörnhamna. However, a limited winter discharge is to be expected from the large bay-head glaciers.

Strong easterly winds that prevail in Hornsund, add drift to the freshet, thus intensifying fjord-type water circulation. Nevertheless, the measured surface-current velocities were seldom higher than 0.5 m/sec, even during persisting eastern winds in excess of 10 m/sec (Swerpel, *in print — b*). Bottom currents are intensified during baroclinic situations. Maximum baroclinic-current speed measured in Isbjörnhamna at 30 m water depth was 0.3 m/sec.

The tide amplitude is up to 1.5 m and tidal cycle apparently does not influence the capacity of meltwater discharge, and only slightly affects the dispersal of turbid plume.

ICE IN THE FJORD

Sea-ice development is influenced by the local vertical stability of water (Görlich & Stepko, *in print*). The fast ice forms over inner Hornsund usually in the end of December and persists till May. In Isbjörnhamna and Gåshamna, the fast-ice sheet is less stable due to episodic destruction by a large oceanic swell. In winter and spring, the fjord is frequently covered by drifting pack-ice fields carried by the Sörkapp Current from the Barents Sea. The sea-ice development in Hornsund is subject to strong fluctuations from year to year.

In Hornsund, ice-rafting is due to the drift of small icebergs and grawlers, and to the presence of thawing pack ice of the Barents Sea origin. Only small icebergs are produced by calving of the Hornsund glaciers. The greatest icebergs observed were ca. 25 m high (a.s.l.) and up to ca. 50 m long. Most icebergs originate from the glaciers in Brepollen and Vestre Burgerbukta. The icebergs plough the morainic sills of Burgerbukta, and often are grounded there. The icebergs frequently remain jammed in northeastern and southeastern parts of Brepollen and in Austre Burgerbukta bay-head. Only few icebergs leave Hornsund. It is thus expected that most of the ice-rafted debris in outer fjord and on the shelf comes from the Barents Sea pack ice.

The fast ice and pack-ice fields control not only the amount and distribution of ice-rafted debris (IRD), and the water-mixing conditions, but they also affect development of organisms. The occurrence of plankton blooms strongly depends on the time and duration of open-water conditions.

PLANKTON AND BENTHOS

The data on zooplankton occurrence in the study area reveal significant correlation between the taxonomic composition of planktonic fauna and hydrological characteristics of main water masses (Węsławski & Kwaśniewski, 1983). On the other hand, the total planktonic biomass is related rather to local characteristics of water (including turbidity). Węsławski *et al.* (1985) report that, in general, Hornsund and south Spitsbergen shelf waters "belong to the poorest in biomass of all the shelf areas" and "plankton samples from Hornsund show the lowest plankton biomass noted in the Arctic". In August 1984, the biomass was the lowest throughout the water column in Hornsund and much higher in the uppermost water layer over the Spitsbergen shelf. This suggests that desalted turbid waters of the bays backed by tidewater glaciers are unfavourable for zooplankton development, is spite of their 80 to 90% O_2 saturation (reported by Węsławski *et al.*, 1985). These facts account for low abundance of pellets and tests in the fjord sediments.

The benthic life is also poorer in the fjord than over the open shelf. The soft-bottom fauna consists predominantly of *Polychaeta* in Hornsund and *Bivalvia* and *Echinodermata* at the Storfjorden stations, marked in Figure 1 as KAG 3 and KAG 4 (Węsławski *et al.*, 1985). The mass occurrence of *Polychaeta* in Hornsund is confined to the distal settings. In the proximal subbasins, no abundant infauna developes.

BOTTOM SEDIMENTS

Four major lithofacies were distinguished in the bottom sediment of Hornsund. These include: stratified sandy mud intercalated with sand and gravel (facies A), laminated mud with organically induced dark lamination and faint coarse-grained streaks (facies B), homogeneous to bioturbated mud, with occasionally preserved organic lamination or spots and coarse-grained intercalations (facies C), and sandy gravel lag with shell debris (facies D).

It is beyond the scope of this paper to discuss all textural and structural varieties of the above facies. Basically, these facies mark transition from ice-proximal (facies A) to ice-distal (facies C) positions, but their distribution strongly depends on bottom morphology and of bottom-current pattern, facies D being totally dependent on these factors.

Facies A forms on the sea bottom narrow belts or lobes along the ice front, which are occasionally overridden by the oscillating glaciers. This zone is migrating with the retreating glaciers and is to be found burried by muds in different parts of the previously glacier-covered basins (what is observed in seismoacoustic profiles). Facies A was not sampled in course of this study. Firstly, because of its proximity to the ice-cliff and secondly, because piston corer did not penetrate coarse sediments.

Facies D occupies some near-shore zones and shallow-water parts of Isbjörnhamna and Gåshamna as well as the open-shelf elevation of Hornryggen. The sediments there are winnowed and reworked by wave action and tidal or baroclinic currents. The facies D developes usually in depressions of exposed bedrock and consists of sand to gravel with boulders and bioclastic components. The clastic material is derived from reworking of till and outwash deposits with admixture of coarse IRD.

Glacimarine mud facies B and C are subject of this study. Facies B: laminated mud, fills ice-proximal troughs in a ponded mode and is thought to originate from vertical settling and low-density gravity flows. Facies C: homogeneous to bioturbated mud, tends to mantle the bottom morphology and is interpreted as the product of suspension settling and ice-rafting.

MATERIALS AND ANALYTICAL PROCEDURES

BOTTOM-SEDIMENT SAMPLES

The marine sediments of Hornsund area were first studied by Antkiewicz and Filipowicz (1978). Numerous sediment samples were collected from the whole Hornsund area by expeditions to Spitsbergen of the Gdańsk University students, led by S. Swerpel and J. M. Węsławski, in the years 1974—1983. The present author collected samples for this study during his wintering in the Polish Polar Station in Hornsund with the Expedition 1982/83, and during the cruise of m/s "Jantar" with the Geophysical Expedition 1985. Both expeditions were organized by the Institute of Geophysics, Polish Academy of Sciences.

Surface-sediment samples were recovered with Van Veen grab, short gravity corer and dredge. This study is based on 80 samples, 20 of which were kindly supplied by J. M. Węsławski and C. Filipowicz. Preliminary results of coring with modified Kullenberg piston corer during m/s "Jantar" 1985 expedition are also included. The cores were maximum 3.20 m long; the deepest sediment penetration was 5.40 m.

The positions of sampling sites are shown in Figure 1. Only samples explicitly referred to in the text are numbered in this figure.

PROCEDURES OF SEMI-QUANTITATIVE XRD ANALYSES

Biscaye's method. Throughout this paper, a method developed by Środoń (1984) is used. However, since previous studies of clays in marine sediments used a method of Biscaye (1965), the latter method will be commented here, and some calculations will be made for comparative purposes.

Studies of clay mineral distribution in marine sediments commonly employ a method of semi-quantitative X-ray diffraction analysis (XRD) developed by Biscaye (1964, 1965). This method was constructed to investigate global distribution patterns of clay minerals in deep-sea sediments (Biscaye, 1965; Griffin *et al.*, 1968). It is also widely used in the studies of high-latitude shallow-marine sediments (e.g., Piper & Slatt, 1977; Molnia & Hein, 1982; Osterman & Andrews, 1983).

Numerous shortcomings of the method were explicitly discussed by Biscaye in his original paper (1965). These confinements arise primarily from the fact that the method: (1) disregards all variable factors influencing XRD peak intensities, included in the diffraction function (instead, it adopts a rigid set of integer weighting factors), and (2) it makes use only of basal reflections of clays in oriented preparations. Hence, Biscaye's method: (1) does not discriminate between mica varieties, (2) it is ambiguous in distinguishing between chlorite and kaolinite in mixtures, and (3) it precludes reliable estimates of non-clay components. The above deficiencies made this method inappropriate to the study of fjord-type sedimentary basin where pronounced compositional variability is unlikely.

Środoń's method. In the present study, the bulk samples and some $< 2\,\mu m$ fractions were semi-quantitatively X-ray analysed for overall mineral composition using a modified technique described by Pawełczyk and Środoń (1978) and Środoń (1984).

The applied method (further on referred to as Środoń's method) relies upon measuring of intensities of the selected (analytical) reflections in disoriented (random) preparations. The measured intensities are related to the respective intensities of

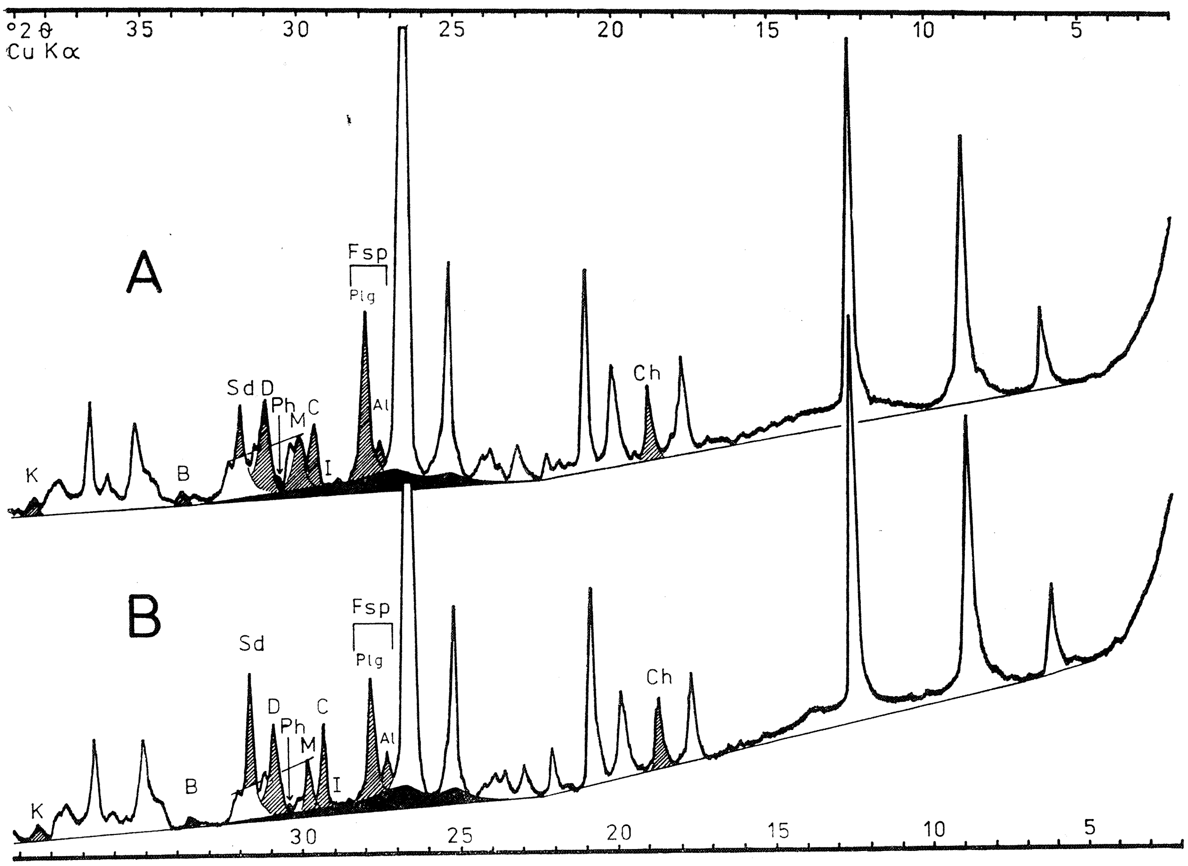


Fig. 3. X-ray patterns of random mounts of bulk sediment samples 069 (*A*) and 072 (*B*). Shaded are analytical reflections used in Środoń's procedure of semi-quantitative XRD analysis (excluding indicated biotite and phlogopite reflections used only for identification). Symbols denote: *K* — kaolinite; *B* — biotite; *Sd* — siderite; *D* — dolomite; *C* — calcite; *Ph* — phlogopite; *M* — muscovite; *I* — illite, *Fsp* — feldspars; *Plg* — plagioclases; *Al* — alkali feldspars; *Ch* — chlorite. Broad black area illustrates 1Md illite reflection, small black reflection is a phlogopite one

pure standards via the formula in: Klug and Alexander (1974, p. 545). In calculations, included are X-ray mass attenuation coefficients of component minerals only, whereas absorption coefficients of the whole samples are omitted as a common denominator. The obtained percent values are thus relative, and are formally normalized to 100% for analysed set of rock-forming crystalline minerals.

The following noncoincident analytical XRD reflections, indicated in Figure 3, were used (first values are *d* spacings in Å, and in parantheses are respective positions of the reflections in 2 theta Cu-Kα): low-temperature quartz — 1.817 (50.2), muscovite $2M_1$ — 3.00 (29.8), illite 1Md — 3.07 (29.1), chlorite — 4.72 (18.9), kaolinite — 2.34 (38.4), siderite — 2.79 (32.1), calcite — 3.04 (29.4), dolomite — 2.89 (30.8), alkali feldspars — 3.23—3.26 (27.6—27.4), plagioclases — 3.15—3.21 (28.3—27.8).

Środoń did not include biotite in his measurements of standard intensities of analytical reflections. Hence, in this study biotite content was determined with separate procedure. To that purpose, the use was made of muscovite/illite (060) reflection 1.498—1.502 Å and the combined biotite (061, 330) plus muscovite/illite ($1.3.\bar{1}\bar{1}$) reflection 1.521—1.528. From the nearly biotite-free $< 2\ \mu m$ fraction of some samples (e.g., sample 800 in Table 1 and Fig. 4), 1.521—1.528 Å reflection was estimated as 1/8 of the 060 muscovite/illite peak. Hence, the following approximate calculation was carried out:

$$\frac{I_{(1.521-1.528)} - \frac{1}{8} I_{(1.498-1.502)}}{I_{(1.498-1.502)}} = \frac{\text{biotite}}{\text{di-micas}},$$

$$\frac{\text{biotite}}{\text{di-micas}} \times \text{di-micas}\ \% = \text{biotite}\ \%.$$

The calculated ratios biotite/di-micas are included in Table 3, and biotite percent contents in Tables 1 and 2. The latter estimate, although very rough (the intensities of analytical reflections for biotite and di-micas are assumed to be equal) is better than not including of biotite, which is definitely present in Hornsund sediments.

XRD PREPARATIONS

The oriented mounts were used to identify and describe clay mineral species. 33 sediment samples were air-dried, and fractions $< 2\ \mu m$ and $< 0.2\ \mu m$ were separated by centrifuging after dialysing them free of salts. Selected samples were treated with sodium acetate buffer (Jackson, 1975) to remove fine carbonates and exchangeable alkaline-earth cations.

The XRD random mounts were prepared by grinding the bulk samples and $< 2\ \mu m$ fractions in laboratory ball-mill and then pressing the powder in the specimen holder against Teflon plate with fine rectangular grooves (Środoń, 1984). The fine-fraction subsamples were sedimented after ultrasonic disaggregation, to produce oriented mounts. Glycolation, K^+ saturation and 1n HCl dissolution were applied to identify clay minerals.

The DRON-3.0 diffractometer was used for XRD analyses, with Ni-filtered Cu-Kα radiation.

Table 1

Mineral composition of sediment samples from Hornsund and south Spitsbergen shelf (in %) from Środoń's method

Sample no.	KAG 1		H7	819		KAG 2	799	072	095[4]	800		112	KAG 3	KAG 4	KAG 5
Area	Brepollen		Brepollen	Adriabukta inner fjord		Outer Samarin-vågen	Outer V. Burger-bukta	Isbjörn-hamna	Gås-hamna	Outer fjord		Skodde-bukta	Hamberg-bukta	Bolt-odden area	Central Stor-fjorden
Depth [m]	95		5	30		105	78	44	55	175		86	85	45	111
Fraction	bulk	2 μm	bulk	bulk	2 μm	bulk	bulk	bulk	bulk	bulk	2 μm	bulk	bulk	bulk	bulk
Quartz	39	24	38	39	15	28	26	23	34	28	20	25	39	42	38
Muscovite[1]	6	13	10	8	19	9	22	18	8	9	24	9	9	6	7
Illite[2]	22	33	26	18	37	25	7	12	6	18	32	11	18	26	25
Biotite[3]	7	6	7	5	7	8	13	16	6	10	3	12	?	6	5
Total micas	35(37)	52	43(36)	31(25)	63	42(30)	42(30)	46(36)	20(14)	37(26)	59(48)	32(34)	27(20)	38(34)	37(28)
Chlorite	2(8)	3	3(8)	2(9)	6	5(13)	7(19)	10(22)	4(10)	5(16)	3(9)	10(17)	4(9)	3(7)	4(13)
Kaolinite	6(3)	11	?(2)	3(2)	6	2(6)	?(0)	2(0)	?(0)	?(0)	?(0)	9(0)	?(2)	?(2)	?(0)
Siderite	4	?	6	3	2	3	2	6	3	5	?	3	6	4	8
Calcite	1	3	1	6	2	14	11	4	23	10	10	3	1	1	1
Dolomite	1	1	1	3	1	2	7	2	7	5	3	14	1	1	1
Alkali fsp	1	1	1	1	1	1	1	3	1	1	1	1	1	1	1
Plagiocl.	11	5	7	12	4	3	4	4	8	9	4	3	21	10	10
Muscovite % / Illite %	0.27	0.39	0.38	0.44	0.51	0.36	3.1	1.5	1.3	0.5	0.75	0.82	0.5	0.23	0.28

[1] Determined as $2M_1$ polytype; [2] determined as 1Md polytype; [3] calculated by separate procedure (see text); [4] winnowed sediment of facies D.
In parantheses are results obtained with Biscaye's method (see text).

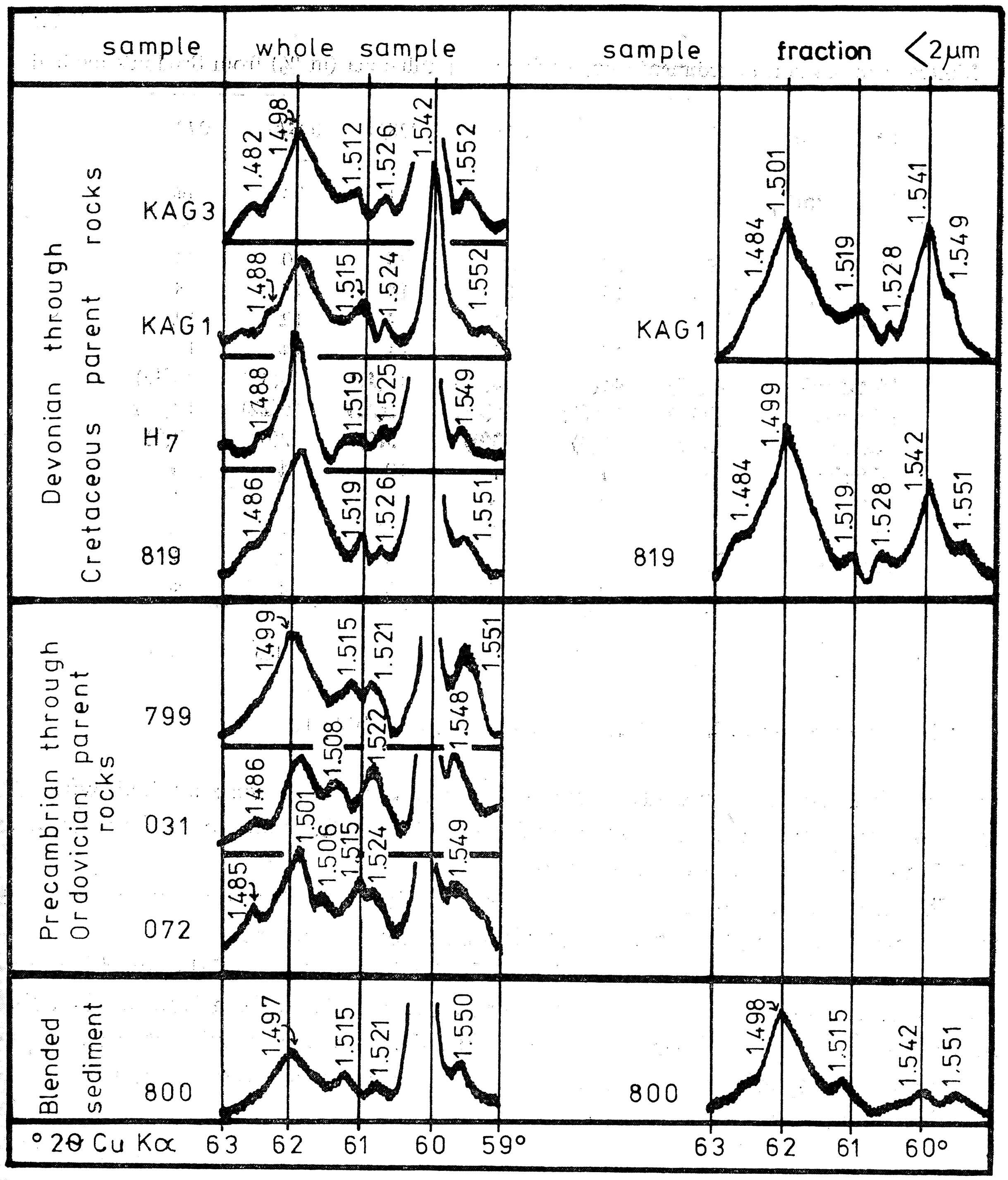


Fig. 4. X-ray 060 (and related) reflections used for distinguishing of tri- from dioctahedral clays and measuring of biotite content. Indicated are sample numbers, *d* spacings in Å and position of XRD reflections in 2 theta Cu-Kα

COMPLEMENTARY ANALYSES

IR spectroscopy was used for identifying mineral composition of 22 centrifuged samples of suspension, since the amounts of the samples were too small for XRD analysis. IR spectroscopy was also applied to sediment samples, primarily to verify the XRD estimates of di- and trioctahedral clays and to identify fine--grained (authigenic) carbonates. Iron content in illites and muscovites was deter-

Table 2

Mineral composition of sediment samples from Isbjörnhamna (in %) from Środoń's method

Sample no.	069	029	031	028/5	028/6	072
Depth [m]	30	12	16	23	23	44
Quartz	28	30	29	26	30	23
Muscovite	18	19	17	18	19	18
Illite	5	5	7	9	12	12
Biotite	18	17	18	14	14	16
Total micas	41(37)	41(35)	42(35)	41(33)	45(37)	46(36)
Chlorite	10(18)	12(21)	11(21)	10(21)	10(22)	10(22)
Kaolinite	3(0)	3(0)	3(0)	3(0)	2(0)	2(0)
Siderite	3	1	1	10	4	6
Calcite	2	4	6	2	2	4
Dolomite	2	2	2	2	1	2
Alkali fsp	3	3	1	3	2	3
Palgioclase	8	4	5	3	4	4
Muscovite % / Illite %	3.6	3.8	2.4	2.0	1.6	1.5

Only bulk samples are included. Other explanations as in Table 1.

mined using infrared spectra. Some 45 sediment samples were studied with IR spectroscopy, either as bulk samples and their < 2 μm fractions or as the fractions: < 0.2 μm, 0.2—2μm, and 2—60 μm.

The infrared spectra for bulk samples and their fine fractions were obtained from KBr pellets in Pye Unicam SP 1200 spectrophotometer.

The scanning electron micrographs of the broken surfaces of the air-dried specimens were obtained with JEOL-JSM-35 microscope. The information on the habit and fabric of clays and coarse-detritial grains, the presence of euhedral microcrystals, and organogenic micro-forms (tests, fecal pellets) were derived from the inspection of micrographs.

QUALITY AND REPRESENTATION OF MINERALOGICAL RESULTS

ERRORS OF XRD ANALYSES

It is attempted here to compare semi-quantitatively different mineral provinces within Hornsund sediments. Both, percent mineral contents and mineral-content ratios used for that purpose, need some comments.

The bulk mineral compositions of the selected samples (estimated with Środoń's method) are given in Tables 1 and 2. Biscaye's and Środoń's methods (see previous section) are compared in these tables, where the Biscaye-method results are shown in brackets. For the purpose of comparison, the Biscaye-method data were recalcu-

Table 3

Ratios of peak intensities and of percent contents of index clay minerals

Ratio		$\frac{K(001)+Ch(002)}{M+I+B(001)}$			$\frac{Ch(001)+Ch(003)}{M+I+B(001)}$			$\frac{\text{biotite}}{\text{di-micas}}$		$\frac{\text{muscovite\%}}{\text{illite \%}}$
Fraction		bulk	< 2 μm	< 0.2 μm	bulk	< 2 μm	< 0.2 μm	bulk	< 2 μm	bulk
Area	Samples									
Brepollen, inner fjord, outer Samarinvågen	KAG 1, H_7, 819, KAG 2	0.82 (0.61–0.94)	0.99 (0.97–1.01)	–	0.31 (0.27–0.35)	0.37 (0.36–0.38)	–	0.22 (0.19–0.25)	0.12	0.36 (0.27–0.44)
Outer V. Burgerbukta	799	1.44	–	–	0.58	–	–	0.44	–	3.1
Isbjörnhamna	071, 069, 030 029, 031, 028 072	1.45 (1.38–1.55)	0.60 (0.50–0.70)	0.24 (0.18–0.28)	0.61 (0.49–0.69)	0.23 (0.19–0.27)	0.14 (0.13–0.14)	0.59 (0.44–0.77)	–	2.5 (1.5–3.8)
Gåshamna	095	1.27	–	–	0.55	–	–	0.45	–	1.3
Outer fjord	800	1.07	0.55	–	0.47	0.22	–	0.38	< 0.05	0.75
Skoddebukta	112	0.98	–	–	0.56	–	–	0.59	–	0.82
Hambergbukta	KAG 3	1.09	–	–	0.43	–	–	–	–	0.5
Boltodden area	KAG 4	0.98	–	–	0.44	–	–	0.20	–	0.23
Central Storfjorden	KAG 5	1.38	–	–	0.59	–	–	0.14	–	0.28

In parantheses ranges of values for a given mineral province. Explanations in text.

lated per total clay content obtained with Środoń's method. Both, for chlorite and micas, the results of the Biscaye and Środoń methods correlate well (correlation coefficients are 0.80 and 0.87 respectively, with $P < 0.001$ in both cases). However, the estimates of Biscaye's method are much higher as regards chlorite and slightly lower as regards micas, compared to those obtained with Środoń's method.

The relative error of the abundances determined in this study may exceed 50% for the low concentrations of the minerals (below approximately 5%). For major components it is assessed to be 10–20%. Such high error is due to:

(1) inaccuracy of adopted mass attenuation coefficients which are not measured but calculated for the assumed chemical formulae of minerals – the coefficients used in this study come from the paper by Pawełczyk and Środoń (1978);

(2) effects of preferential orientation of clay flakes, which is not removed completely by applied intensive grinding and pressing of random mounts with grooved plate,

(3) effect of applying standard intensities taken as a complete set of measurements from Środoń (1984, and unpublished); the minerals encountered in Hornsund sediments certainly differ within distinguished mineral groups from those standards;

(4) diffuse or overlapped analytical reflections of kaolinite,

(5) difficulties in estimating biotite content due to the lack of suitable analytical reflection and measured standard intensity.

Most of the above inaccuracies introduce systematic errors. However, the error due to the effect (3) depends on the intrinsic quality of sedimentary material. It may be neglacted in single-source basins, but may introduce significant errors when comparing remote sedimentary basins. Hence, mineral relations within Isbjörnhamna or Brepollen subbasins are much more reliable than comparisons between those subbasins.

MINERAL CONTENT RATIOS

The ratios between the mineral contents, and for clay minerals also between the selected basal reflections, are used in this paper to discuss mineral relations in the sediment. In Table 3, the ratios between intensities of clay mineral basal reflections (from oriented mounts) and between percent contents of some clay minerals are shown, averaged over groups of samples. The samples are grouped with regard to similarity of these index ratios. The bulk compositions, and also clay mineral assemblages, vary within the distinguished groups, as may be seen from Tables 1 and 2 when compared with Table 3.

The intensity ratios are used here along with percent-content ratios, since much more XRD patterns from oriented specimens than those from random mounts, were obtained in this study. Intensity ratios of XRD reflections do not express relative weight abundances of the respective minerals. Intensity ratios reflect relative changes within clay mineral assemblages (and this only for basal reflections of clays which due to their habit have similar preferred orientation).

Relation between the two measures of relative abundance of clay minerals (i.e., intensity ratios and percent-content ratios) was checked for the samples for which both types of data were available. The linear correlation coefficients between the intensity ratios and percent-content ratios are: 0.75 for the set of data concerning chlorite/micas, and 0.77 for the set of data concerning chlorite + kaolinite/micas, significant at the levels of 0.001 and 0.01 respectively.

Distinct deviation from this linear correlation, encountered in sample 112 from Skoddebukta, suggests intrinsic mineralogical reasons for specific XRD behaviour. Such discrepancies arise due to the effect of heavy atoms (mianly Fe) in the lattice of chlorite on the intensity of basal reflections. The intensity ratio of the even to odd basal reflections of chlorites increases with increasing Fe content. Hence, the 003 line which is used here to calculate percent content of chlorites is affected inversly to the 002 line used to determine intensity ratios in Table 3. The above-mentioned sample 112 reveals much too low intensity ratio kaolinite + chlorite/micas (Table 3) compared with corresponding percent-contet ratio (to be obtained from Table 1). This is explained by the lower iron content in chlorite of sample 112 as compared with the remaining studied samples. The calculated 5.7 atoms per unit cell in chlorite of sample 112 is in fact the lowest of iron contents estimated for chlorites in the samples from Hornsund and Storfjorden. Since such discrepancies concern only the samples from Skoddebukta, they will not negatively influence considerations concerning the sediments of Hornsund.

ANALYTICAL RESULTS

PHYSICAL PROPERTIES OF SEDIMENTS

No systematic grain-size analyses of Hornsund sediments were carried out. In course of preparation of subsamples for mineralogical investigations, the following fractions were separated: < 0.2 μm, 0.2—2 μm, 2—60 μm, 0.06—0.25 mm, 0.25—2 mm, and > 2 mm. Below, representative textural data are given, obtained for surface sediments from homogenized grab samples.

The contents of clay fraction in muds vary between maximum 33% in surface sediments of Brepollen and minimum 14% in Isbjörnhamna. In Burgerbukta, Samarinvågen and inner central fjord, the mud samples contain on the average 25% of clay fraction. Much more variable are contents of finest < 0.2 μm fraction. These range between 14% in Brepollen and 0.5% in Isbjørnhamna. The ratios of fractions < 0.2 μm/0.2—2 μm, are given in Table 4. These ratios are denoted here

Table 4

Average fine-fraction ratios (DR/PM) in sediments of selected areas in Hornsund

Area and samples	DR/PM[1]	Fraction < 60 m	Minimum distance from source glacier	Remarks
Brepollen (17, 18, 19)	0.74	90—97%	1.5—2 km	Different source glaciers on Mesozoic to Cainozoic sedimentary bedrock
Treskellen Narrows (13)	0.32	84%	6 km	Under influence of Brepollen (sedimentary) and Samarinvågen (metasedimentary) sources
Austre Burgerbukta (16a, 16b)	0.36	86 & 95%	3 km	Different source glaciers on lower Palaeozoic through upper Mesozoic low-grade metamorphic to sedimentary bedrock
Isbjörnhamna (071, 030, 031, 026)	0.03	70—85%	0.5—3 km	Single source-glacier on Precambrian through lower Palaeozoic metamorphic and metasedimentary rocks

[1] DR/PM = fraction < 0.2 μm/fraction 0.2—2 μm.

diagenetic-recycled/primary-metamorphic (DR/PM) due to their inferred genetic significance. From Table 4 it is readily seen that these ratios depend on some specific fatures of sedimentary material rather than on the water depth or distance from input point.

The contents of the clay plus silt fraction (< 60 μm) fall within a range 70—97%. This fraction of muds is assumed to be deposited from suspension. The content of fraction > 0.25 mm in the most proximal muds attains 17% in Austre Burgerbukta (sample 16), 15% in Isbjörnhamna (sample 076), and 7% in Brepollen (sample 17). It falls to about zero in central Brepollen, whereas in the axial part of Hornsund it raises again to 11% in the surface sample 13 (Treskelen Narrows) and 15% in sample 800 (outer fjord).

The muds under study are strongly saturated with water. The water contents (determined as weight loss by heating at 110°C), are: Brepollen 40—47%, Vestre Burgerbukta around 30%, Isbjörnhamna 33—49%.

SEDIMENT FABRIC

According to SEM micrographs, the discussed sediments are a mixture of large and fine clay flakes (generally of irregular habit) with angular quartz, silicate and carbonate grains embedded in it. The large flakes are detritial *sensu stricto*, i.e. they originated from mechanical disaggregation of rocks in source areas. On the other hand, some micrographs (Pl. I: 1 and 4) show the origin of the finest fraction of clays. The weathered crystals of feldspar deliver fine irregular illite flakes. The shape of quartz grains (e.g., Pl. I: 3 and 5) shows the unequivocally glacial origin of coarse sedimentary material, although conchoidal fracturing is seldom observed.

Parallel patterns of the large (above 5 μm) clay flakes (Pl. I: 2, 4, 6) are either crystals cleaved by weathering or face-to-face associations. According to numerous observations and theoretical analysis (Yariv & Cross, 1979, p. 357), face-to-face coagulation of clay particles is a slow process and kinetic barrier may inhibit its spontaneous occurrence. The face-to-face mode in SEM micrographs is thus in general artifact, due to applied shearing and drying during sample preparation.

The fecal pellets are practically absent in the laminated mud facies B. Some, may be seen in the SEM micrographs of the samples from homogeneous-to-bioturbated mud facies C (Pl. II: 1—4). Also, only in facies C the microfaunal tests (or their casts) were found (Pl. II: 3—6).

NON-CLAY MINERALS

General remarks

There are distinct differences in mineral composition of bottom sediments in Hornsund. These differences can be seen in Tables 1 and 2, where samples are ordered in a sequence from inner fjord (Brepollen, sample KAG 1) to outer fjord (sample 800), and from western coast of south Spitsbergen (Skoddebukta, sample 112) to its eastern coast (Storfjorden, samples KAG 3 to KAG 5). The mineralogical differences concern contents of carbonates (and within carbonates, of calcite relative dolomite), plagioclases, and composition of clay-mineral assemblage.

Carbonates

Only calcite and dolomite may be used to discuss provenance and sediment mixing since, as will be shown, siderite is commonly authigenic in the studied sediments. Any significant contribution of recent biogenic calcite was ruled out by mineralogically analysing Balanus and Chlamys shells found in the sediments (sample 095 from facies D), and by studying SEM micrographs for frequency of occurrence of calcareous tests. Neither aragonite identified in the shells, nor abundant calcareous tests were identified in the studied sediments in quantities exceeding detection limit of the analytical methods applied.

The XRD-determined calcite plus dolomite contents for the respective subbasins in Hornsund area, are: 2% in Brepollen, 3—8% in Isbjörnhamna, 16% in Samarinvågen, whereas 20—25% in Vestre Burgerbukta (see Tables 1 and 2). The calcite plus dolomite content in the inner central fjord (Adriabukta, sample 819) is 9% and in the outer central fjord (sample 800) 15%.

The selected bulk sediment samples which were subject to sodium acetate treatment, indicate slightly different carbonate contents than those obtained from XRD data. However, by comparing XRD estimates with weight losses by acetate dissolution, one readily sees that in the ice-proximal basins practically all carbonates are dissolved in two runs of acetate treatment. For instance, in Samarinvågen — 12 to 18% is lost after acetate treatment against 19% total carbonates determined with XRD, in Isbjörnhamna — 10% is lost after acetate treatment against average 9% total carbonates determined with XRD. This suggests dominating presence of fine-grained carbonates in these subbasins. On the contrary, in the central fjord only a fraction of carbonates is dissolved in acetate (e.g., in sample 819 from Adriabukta — 2% loss after acetate against 12% total carbonates determined with XRD). This in turn, suggests that most carbonates are concentrated in coarse fraction.

The carbonate content in sediments of Hornsund may be directly related to the lithology of source area only for the bays where single sediment source dominates (see Figs. 2, 10). Brepollen and Isbjörnhamna catchments are dominated by non--carbonate rocks. Samarinvågen and Gåshamna receive much carbonate material from the Ordovician formations building the Hornsundtind and Tsjebysjevfjellet mountain groups. Vestre Burgerbukta catchment is dominated by Cambro-Ordovician limestones, dolostones and marbles. In these ice-proximal single-source subbasins, carbonates are concentrated in fine fraction (acetate soluble), and thus were presumably deposited from suspension. In distal locations of central fjord (e.g., samples 819 and 800), carbonates are concentrated in coarse fraction (resistant to acetate treatment) which suggests deposition in form of IRD (ice-rafted debris).

Carbonate versus siliciclastic components

Semi-quantitative conclusions may be drawn from comparison between quartz content and calcite plus dolomite content (index Q/C+D). Other useful indices are: feldspar content to calcite plus dolomite content (index Fsp/C+D) and quartz to feldspar contents (index Q/Fsp). Average values of these ratios for the selected subbasins of Hornsund, are given in Table 5 (in parentheses are values for < 2 μm fraction). In the last column of Table 5, the abbreviated names of mineral facies are given, since the values of this table were used to distinguish these facies.

Table 5

Average mineral-content ratios for non-clay minerals in bottom sediments of subbasins of Hornsund and adjacent shelf

Area	Q/C+D[1]	Fsp/C+D	Q/Fsp	Mineral facies
Brepollen	19.3 (6.0)[2]	5.0 (1.5)	3.6 (4.0)	Q-Fsp
Samarinvågen	1.8	0.3	7.0	Cb-Q
Adriabukta	4.3 (5.0)	1.4 (1.7)	3.0 (3.0)	Q-Fsp-Cb
V. Burgerbukta	1.3	0.3	5.2	Cb-Q
Inner Isbjörnhamna	5.8	1.6	4.2	Q-Fsp-Cb
Outer Isbjörnhamna	3.8	1.2	3.3	Q-Fsp-Cb
Gåshamna	1.1	0.3	3.8	Cb-Fsp
Outer fjord	1.9 (1.5)	0.7 (0.4)	2.8 (4.0)	Cb-Fsp
Skoddebukta	1.5	0.2	6.3	Cb-Q
Storfjorden	19.8	7.2	3.0	Q-Fsp

[1] For mineral symbols see text and captions to Figs. 3 and 7.
[2] In parantheses are values for fraction < 2 μm.

The values in Table 5 support previous observation that most carbonates are concentrated in fine fraction in the ice-proximal subbasins, and in coarse fraction in the central fjord (compare values for Brepollen with those for inner and outer central fjord for bulk samples and clay fraction). The comparison of values for Brepollen and Adriabukta (inner central fjord) indicates that overall increase in carbonate content between these two locations (see Table 1) occurs within coarse fraction (> 2 μm).

Another feature of indices in Table 5 is that they adequately show similar provenance of sediments in Brepollen and Storfjorden. The correspondence between the pertinent indices is striking when one regards the geographic position of the two areas, separated from each other by Spitsbergen mainland (or rather by two bifurcating glaciers: Hornbreen and Hambergbreen). The only link between Brepollen and Storfjorden is the lithology of bedrock in the source area built of predominantly siliciclastic rocks of Mesozoic and Cainozoic ages.

CLAY MINERALS

Description of clay mineral species

Illite, muscovite, and expandable clays. Illite and muscovite are major components of clay fraction in the bottom mud of Hornsund. Their relative abundance is different in various subbasins of the study area. Two polytypes of dioctahedral micas were determined in Hornsund samples, viz. 1Md and $2M_1$ (Fig. 3). It is not possible in mineral mixture to evaluate precisely what part of illite is the 1M or 1Md variety and how much of it is crystallized in $2M_1$ form. However, it is justified as a first approximation to treat illite as 1Md polytype and muscovite as $2M_1$ polytype. Such approach is adopted throughout this paper.

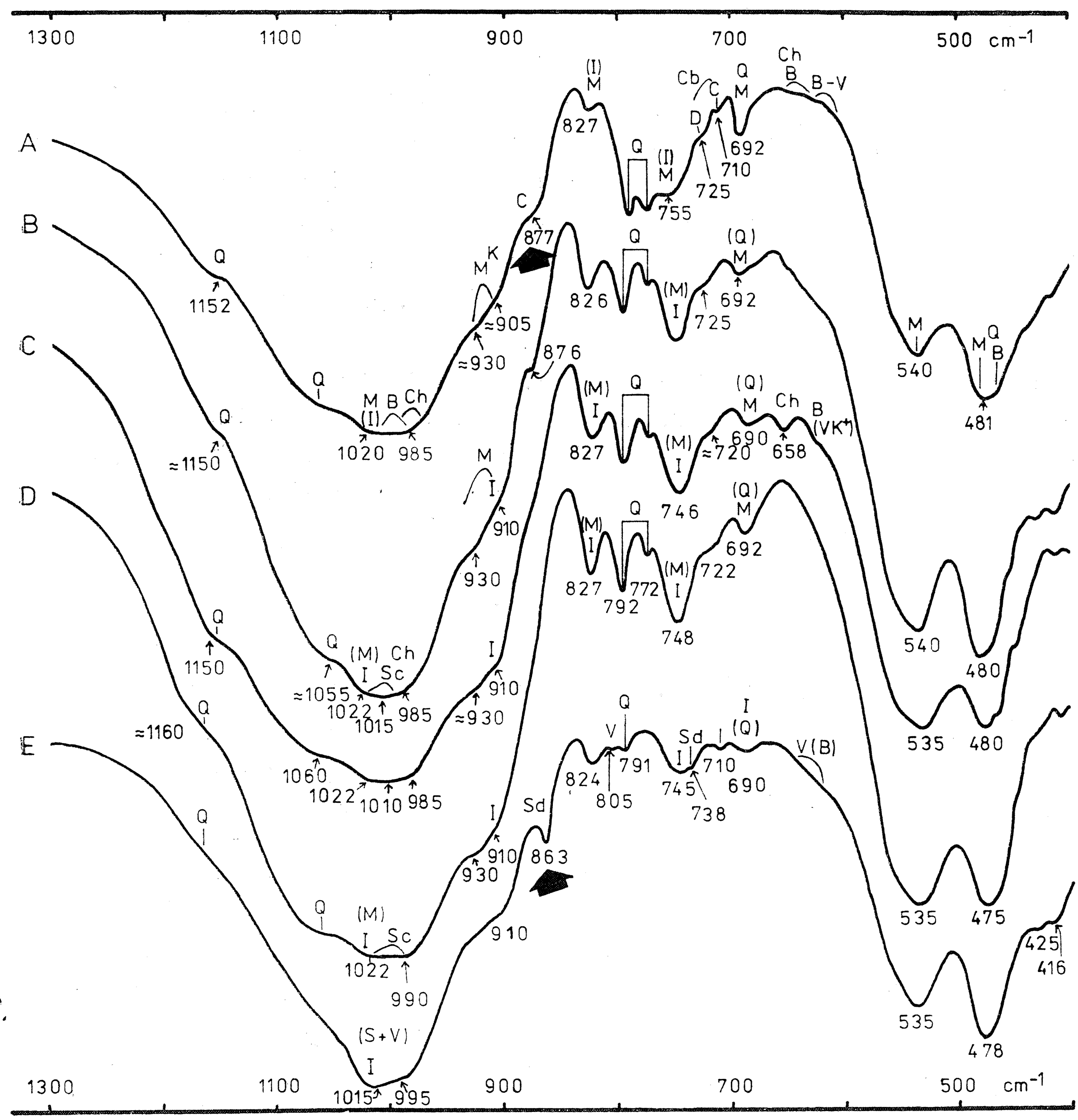


Fig. 5. Infrared spectra of variously treated and untreated fractions of sediment sample 069. *A* — bulk sample; *B* — untreated fraction $< 2\ \mu m$; *C* — K^+ saturated fraction $< 2\ \mu m$; *D* — 1n HCl treated fraction $< 2\ \mu m$; *E* — untreated fraction $< 0.2\ \mu m$. Arrows indicate: calcite band in spectrum *A* and siderite band in spectrum *E*. Symbols for minerals as in Fig. 3

Dioctahedral micas in Hornsund sediments differ in chemical composition. IR spectroscopy was used to determine iron plus magnesium substitution (Fe+Mg) in octahedral layers, using data of Środoń *et al.* (in preparation). In the muds of Hornsund, two chemical varieties of dioctahedral micas are encountered. The low (Fe+Mg) variety, found abundantly only in Isbjörnhamna, is characterized by the following sequence of absorption bands: 930 cm^{-1}, 827 cm^{-1} much less intensive than the 755 cm^{-1} band, 540 cm^{-1}, and 480 cm^{-1}. On the other hand, the high (Fe+Mg) variety exhibits the following bands: 910 cm^{-1}, 824 cm^{-1} equal in intensity to the 745 cm^{-1} band, 535 cm^{-1}, and 478 cm^{-1} (compare spectra in Figs. 5B and 5E).

The above chemical varieties may be readily identified with $2M_1$ muscovite and 1Md illite, respectively. This conclusion is supported by observation that the low (Fe+Mg) variety, referred to muscovite, occurs only in > 0.2 μm fraction, whereas in the < 0.2 μm fraction, exclusively high (Fe+Mg) variety is found (see Fig. 5).

From diagram of Stubican and Roy, modified by Środoń *et al.* (*in*: Środoń *et al., in preparation*), 0.10 $(Fe+Mg)/O_{10}(OH)_2$ for muscovite, and 0.20 $(Fe+Mg)/$ $/O_{10}(OH)_2$ for illite were obtained. In Isbjörnhamna, both varieties occur: low (Fe+Mg) muscovite in coarse-clay fraction and high (Fe+Mg) illite in fine-clay fraction. In the remaining subbasins of Hornsund (viz. Brepollen, Samarinvågen, and Burgerbukta) nearly exclusively the high (Fe+Mg) variety is encountered according to IR spectra.

No expandable clay minerals were detected in sediments of the study area, besides traces of smectite or randomly interstratified illite/smectite in sample 071 (from Isbjörnhamna) and some high-illitic illite/smectite in Brepollen samples (compare Fig. 6B). Expandabilities of illitic minerals were checked using the XRD intensity ratio $Ir = I(001)/I(003)_{\text{air-dry}}: I(001)/I(003)_{\text{glycolated}}$ (Środoń, 1981). In Hornsund, these ratios do not exceed 1.3, which indicates nearly zero expandability.

Biotite. The biotite was unambiguously identified under microscope but it could not be described in this mineral mixture, since its diagnostic reflections are weak and the line 060 is obscured by quartz reflection (Fig. 4). The highest content of biotite is observed in the sediments of Isbjörnhamna (average 16%), and also in Vestre Burgerbukta and Skoddebukta (see Tables 1 and 2).

Chlorite. The chlorites were described using 060 reflection from random mounts and by comparing intensities of basal reflections from oriented mounts (allowing for the effect of kaolinite on the even-order chlorite reflections by subtracting respective peak intensities from 1n HCl treated samples). The data from Brown and Brindley (1980, pp. 344—345, tables 5.15 and 5.16) were used to determine iron distribution between the 2:1 and brucitic layers, and total iron and ferrous iron contents in the unit cell. Chlorites in all the studied samples have iron atoms distributed symmetrically between octahedral sheets. According to XRD data, the mean chlorite composition for Hornsund sediments is $(Mg, Al_{4.2}Fe_{7.8})(Si, Al)_8O_{20}(OH)_{16}$. The Fe content ranges from 7.2 to 9.2 per unit cell. The iron content of chlorites

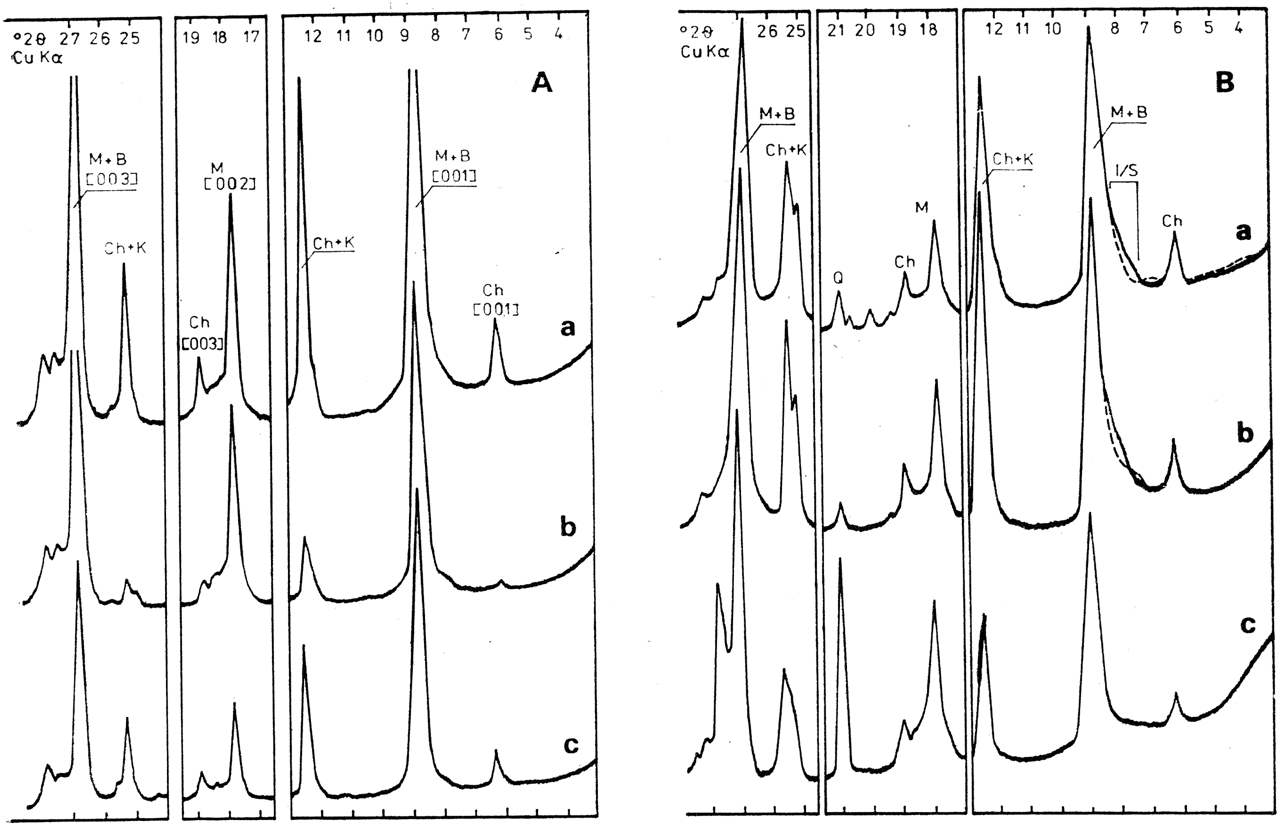


Fig. 6. X-ray patterns of oriented mounts of < 2 μm fractions. *A* — sample 072: *a* — natural air-dry, *b* — 1n HCl treated air-dry, *c* — K^+ saturated glycolated (identical to that of natural glycolated); *B* — natural samples: *a* — KAG 1 from Brepollen, *b* — 819 from inner central fjord (Adriabukta), *c* — 800 from outer central fjord. In *B*, dotted line shows profiles for glycolated preparations (for sample 800 there is no change in profile after glycolation)

in the fraction < 0.2 μm (on an average 5.0 Fe atoms per unit cell) is significantly lower than that for the < 2 μm fraction. The ferrous ion content was calculated using *b* parameter. The 060 spacings for chlorites range from 1.548 to 1.552 Å and the corresponding *b* paıameters are 9.228 and 9.312 Å. The formula of Engelhardt (*fide*, Brown & Brindley, 1980, p. 341) gives for such parameters 4.9 to 6.6 of Fe^{2+} ions per unit cell.

From Table 3 (chlorite to micas intensity ratio) one may read that in Brepollen chlorites are of the size comparable with very fine (*cf.* Tables 1 and 4) flakes of mica, whereas in Isbjörnhamna, chlorites are bound with fractions coarser than 2 μm. This is in accord with the fact that chlorites of metamorphic rocks (Isbjörnhamna area) develope larger flakes than diagenetic chlorites from sedimentary rocks (Brepollen area).

Kaolinite. Kaolinite content in Hornsund sediments is generally very low. In Isbjörnhamna, judging from Fig. 6A-b, it is probably even lower than that indicated in Table 2, based on the intensity of the combined ($20\bar{2}$, $1\bar{3}1$) reflection. The highest kaolinite contents are observed in Brepollen and inner fjord sediments (Table 1 and Fig. 6B-a, b). However, also for these samples absorption bands of kaolinite were not identified in the IR spectra.

Variability of clay mineral assemblages

The uniformity of mineral composition of the sediments from Isbjörnhamna (Table 2) is striking when compared to what is observed for the whole study area (Table 1). Nevertheless, in Isbjörnhamna along a short distance of 1 km, which separates the samples 069 and 072 (Table 2) some compositional trends are visible. The abundance of trioctahedral micas decreases, while that of illite increases. Consequently, the ratio between the dioctahedral and trioctahedral micas grows with increasing distance from the source. The ratio between muscovite and illite contents decreases from ca. 4.3 in Isbjörnhamna (sample 029 in Table 2) to about 1.5 on the outer side of the Isbjörnhamna sill (sample 072 in Tables 1 and 2) and to about 0.6 in the outer part of central fjord (sample 800 in Table 1).

In order to explain the observed increase in illite on expense of biotite and muscovite, the author has analysed 22 samples of suspension collected from surface waters along the profiles from Isbjörnhamna to fjord, by various suspension concentrations and in various positions with respect to the axis of the sediment-laden plume (details in Görlich, *in print*). It was expected that depletion in muscovite and biotite will be correspondingly marked in suspension from surface waterlayer. However, no unequivocal conclusions could be drawn from the compositional variability of suspension.

According to IR spectra (see Fig. 7), as could be anticipated, the suspension at the stations in distal positions (e.g. sample 085) is depleted in quartz and carbonates with respect to the most proximal stations (e.g. sample 076). Nevertheless, no distinct change in composition of clay mineral assemblage is suspension is observed along the transportation path within Isbjörnhamna, in spite of the overall decrease of suspension concentration by about an order of magnitude (*cf.* Fig. 7). In distal part of the bay, there is still domination of muscovite (or rather Al-rich mica variety; see p. 454) over illite (Fe+Mg-rich dioctahedral mica variety). It may be only inferred from comparison of suspension spectra that there is a slight increase in illite content relative muscovite with growing distance from the source. The preferential settling out from suspension of trioctahedral mineral species is a little more pronounced. In the most proximal samples a trioctahedral band 3535 cm^{-1} is observed (not shown in Fig. 7). This band is substituted in the spectra of distal suspension by a dioctahedral band 3620 cm^{-1}. This is quite in accord with compositional changes in the sediment which is seawards gradually depleted in biotite and chlorite.

Tables 1 to 3 inform of the following specific features of clay mineral assemblages in the samples from different parts of Hornsund and south Spitsbergen shelf. Firstly, the clay mineral assemblages maintain their compositional integrity within the single-source subbasins extending from 2 km (Isbjörnhamna) to 10 km (Brepollen) off the glacial meltwater input. For a given single-source subbasin, respective clay mineral assemblages reveal characteristic proportions between components (tri--micas/di-micas, chlorite/micas, kaolinite/micas); both in the bulk samples and in grain fractions. Secondly, it depends on the primary composition of sedimentary material, whether textural segregation of clay minerals takes place during transport.

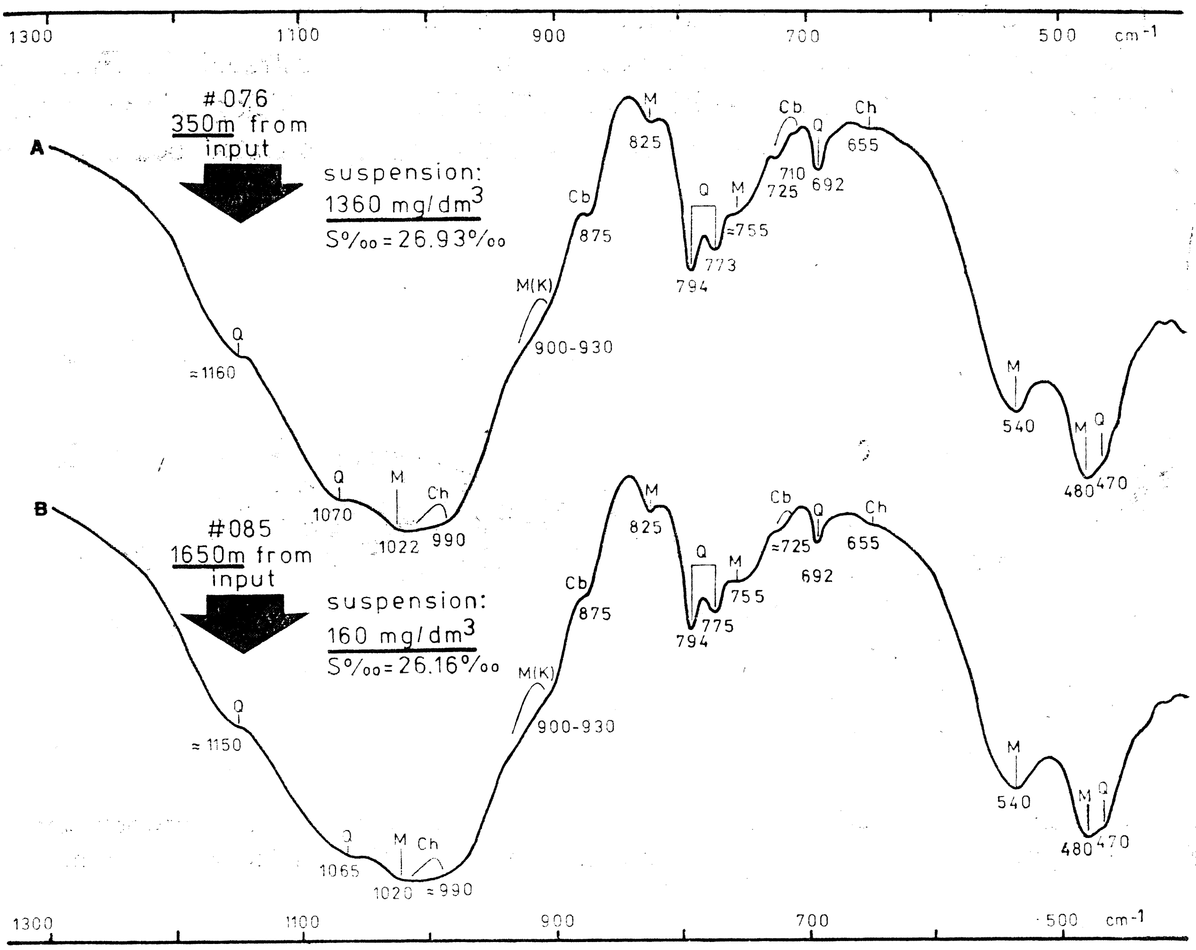


Fig. 7. Infrared spectra of bulk suspension from surface water samples collected at the stations 076 (*A*) and 085 (*B*). Symbols denote: *Q* — quartz, *M* — dioctahedral micas, *Cb* — carbonates; other symbols as in Fig. 3

Clay-grade illite is separated from muscovite, biotite and chlorite along the transportation path in Isbjörnhamna. This does not, however, occur in Brepollen area where mica content is more or less the same, but illite dominates (Tables 1 and 3). Here, the muscovite/illite ratio changes randomly within the range 0.27 to 0.44.

SEDIMENT ACCUMULATION RATES

ESTIMATION OF RATES

The sediment cores from Hornsund area have not yet been dated. An attempt to evaluate average sediment accumulation rates is thus based in this study on:

(1) thickness of glacimarine-mud drape on the bedrock, till or proximal-outwash substratum interpreted from continuous seismoacoustic profiles (CSP),

(2) depth of corer penetration into glacimarine muds,

(3) dating of glacial advances and retreats in the fjord and its side-bays, based on Quaternary-geological results and position of glacier fronts mapped in twentieth century,

(4) separation of black monosulphidic laminae in facies B.

The calculation based on seismoacoustic profiles is illustrated in Figure 8 (the profile is to be located as A—B in Fig. 2). The 25 to 30 m thick glacimarine mud sequence in Figure 8 is interpreted as being deposited on exarated bedrock after

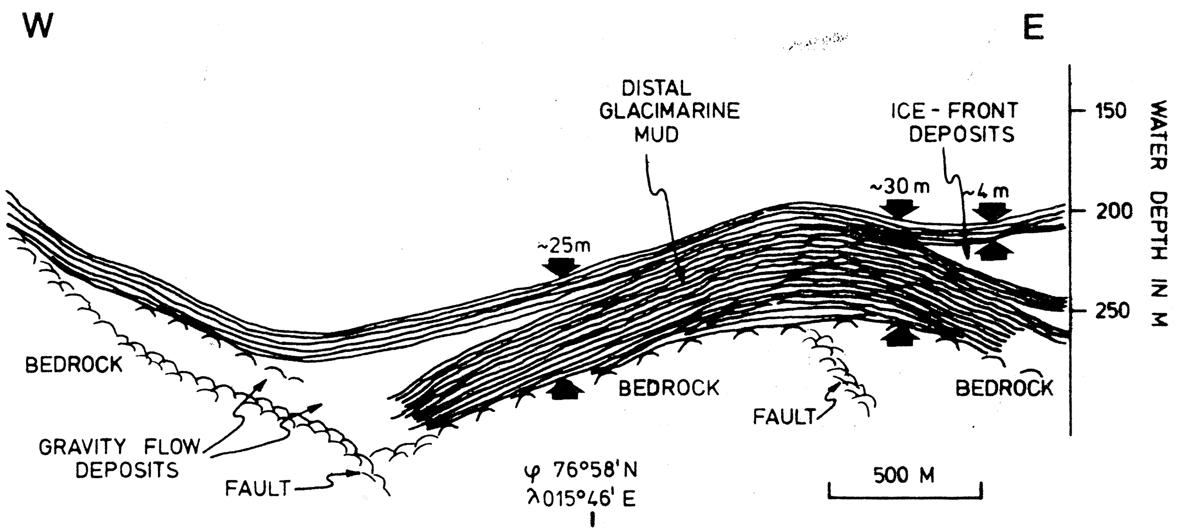


Fig. 8. Thickness of glacimarine muds in the central basin of Hornsund based on interpretation of seismoacoustic profile *A—B* (position in Fig. 2). Adopted seismic velocity is 1.7 m/ms

the retreat of the Hornsund glacier from its maximum Vistulian extent dated at about 40,000 years B.P. (Lindner *et al.*, 1983a, 1983b). This yields an average accumulation rate of about 0.2 cm/a for this sequence. The above sequence is interfingering with (presumably) proximal glacimarine deposits laid down during the Holocene maximum of glacial advance dated ca. at 3,500—2,000 years B.P. (*op. cit.*). These proximal deposits are overlain by 4 m thick distal glacimarine unit. For this unit, we obtain also the average rate of about 0.2 cm/a. Relatively low average rates for central basin of Hornsund point to the fast glacial retreat after both the above maxima. The site in central basin must have remained for only a short time in the ice-proximal position and no thick, rapidly accumulated mud was deposited.

It must be stressed that during the monotonous glacial retreat, the momentary sedimentation rate is expected to decrease quasi-exponentially and, hence, the present-day sedimentation rate in the central basin is probably much lower than the average one, calculated for a thousand-of-years time span. Obviously, the error of averaging is only minor for accumulation rates estimated in proximal subbasins, based on sediment units deposited during the last tens of years.

The rates for proximal sites were calculated using the depths of maximum penetration of the corer into the mud, assuming that the corer was halted by coarse--grained proximal outwash. If it were not the case, the continuous mud unit would be thicker than penetrated by the corer and thus the calculated rates are treated as the minimum ones. The estimated rates were checked against the mean distance between the monosulphicic layers. The results agree roughly, although this verification should not be regarded too strictly as there are numerous indications of intervening gravity flows.

The second factor, i.e. timing of the latest glacial retreat from a given proximal site, could be taken from the existing cartographic data, since in most cases the proximal sites were exposed from beneath ice after the first geodetic mapping in 1936.

From the above considerations it follows that the calculated accumulation rates are rather lower than actual ones in proximal settings, whereas higher than the true ones in distal settings. Referring to the following paragraph, it should thus be noted that sedimentation decay presented in this study will rather be biased towards underestimation.

SEDIMENTATION-DECAY CURVE

All the calculated accumulation rates are shown in Figure 2 (in rectangles). In spite of the complex morphology of the fjord, the above estimated rates fit well to the law of exponential decay of suspension settling, formulated by Syvitski (*fide*, Farrow *et al.*, 1983). According to this law, sedimentation rate (ST) at a given distance (x) from the source glacier (or glacial-river mouth), is:

$$\log(ST) = mx + \log(b) \text{ or}$$

$$\log(ST) = \frac{mx}{10} + \log(10^{-2}\, b) \text{ when } (\mathrm{ST}) < 10^{-2}\, b,$$

where m is a coefficient and b stands for ice-front sedimentation rate (the rate measured at the most proximal site of mud accumulation). In order to obtain the best fit to the Hornsund data, different values of m were accepted in the above formulae. For Brepollen and Samarinvågen areas, the best-fit m is equal to -0.1, for Burgerbukta and fjord stations it is -0.2, and for Isbjörnhamna it is -0.4 (Table 6). The accumulation rates evaluated from acutal sediment thicknesses, and sedimentation rates calculated according to Syvitski's formulae are gathered in the cited table. One may correlate the best-fit coefficient m (which determines the steepness or rapidity of sedimentation rate decay) either with the openness of sedimentary subbasins or with the grain-size composition of sedimentary material, or both. For the most restricted basins (Brepollen, Samarinvågen) the coefficient $m = -0.1$, for the most open bay (Isbjørnhamna) $m = -0.4$. Simultaneously, in Brepollen, the finest fraction of clays dominates (Table 4), whereas in Isbjørnhamna mainly coarse clays (with abundant flakes $> 2\ \mu$m) are present. Thus, the coefficient m may be used to classify the single-source subbasins in the order of efficiency of settling of glacial meltwater load but without specifying whether efficiency is controlled by oceanographic conditions (e.g., salinity gradients) or rather by grain-size distribution in suspension.

DISCUSSION

SETTLING OF SUSPENSION

Agents controlling settling

These considerations will concern only clay-mineral components of suspended load, which have active surfaces and are sensitive towards the change of chemical environment. Before discussing selectivness of settling, the author will consider the agents controlling settling *en masse* of suspension. A perforated-conveyor-belt model of suspension transport is accepted here to hold for the tidewater-glacier sedimentary systems. This model includes horizontal flux of suspension in thin surface layer of water, and vertical flux of suspension below halocline (Fig. 9).

From the results of this study, two principle factors appear to govern settling of clays: (1) primary composition of meltwater load (mainly its texture), and (2) vertical salinity gradient in the forefield basin of tidewater glacier. The former effect is gravitational. The effect of high salinity gradient on sedimentation of clays consists in bringing concentrated meltwater load in contact with concentrated sea--water electrolyte before suspension becomes diluted enough not to undergo significant flocculation (Fig. 9). On the contrary, when salinity gradient is low (i.e., contact between the freshet and sea water is diffuse), a very diluted suspension is brought into sea water of ionic strength sufficiently high to cause coagulation. The latter situation results in a relatively small number of minute floccules.

Slope of sedimentation decay

Both the above agents (texture and salinity) influence the steepness of the exponential curve of turbidity decay and, simultaneously, of sedimentation rate decay. The sedimentation rate will decrease rapidly seawards, when: (1) clay fraction is dominated by coarse clay flakes (pure textural, i.e. gravitational effect) and/or (2) salinity gradient between marine waters and meltwater plume is large (electrochemical effect). On the contrary, the decay of sedimentation rate will be slower, when: (1) clay fraction is dominated by finest grains and/or (2) salinity gradient at an input point is moderated by hindered fjord-type water exchange in restricted bays and by large meltwater discharge.

Salinity gradient depends on the thickness of the so-called freshwater cap and sharpness of its contact with underlying marine waters. The freshwater cap is the thickest and its lower contact the most diffuse in Brepollen; the reverse situation is valid for Isbjörnhamna (Fig. 9, this paper, and Urbański *et al.*, 1980, figs. 3, 4).

It is obvious that if the above factors (original grain-size spectrum and salinity gradient) act in opposite directions, then their effects should cancel; if they are in phase, the effect should be amplified. Such amplified effect is indeed observed in Isbjörnhamna, where both, coarse clays dominate (Table 4) and salinity gradient is the largest (Fig. 9C). To illustrate this amplified net effect we have values of coefficient m of Syvitski's formula (Table 6 and foregoing section). The pertinent coefficient for Isbjörnhamna is equal to -0.4, and thus extraordinarily high (its absolute

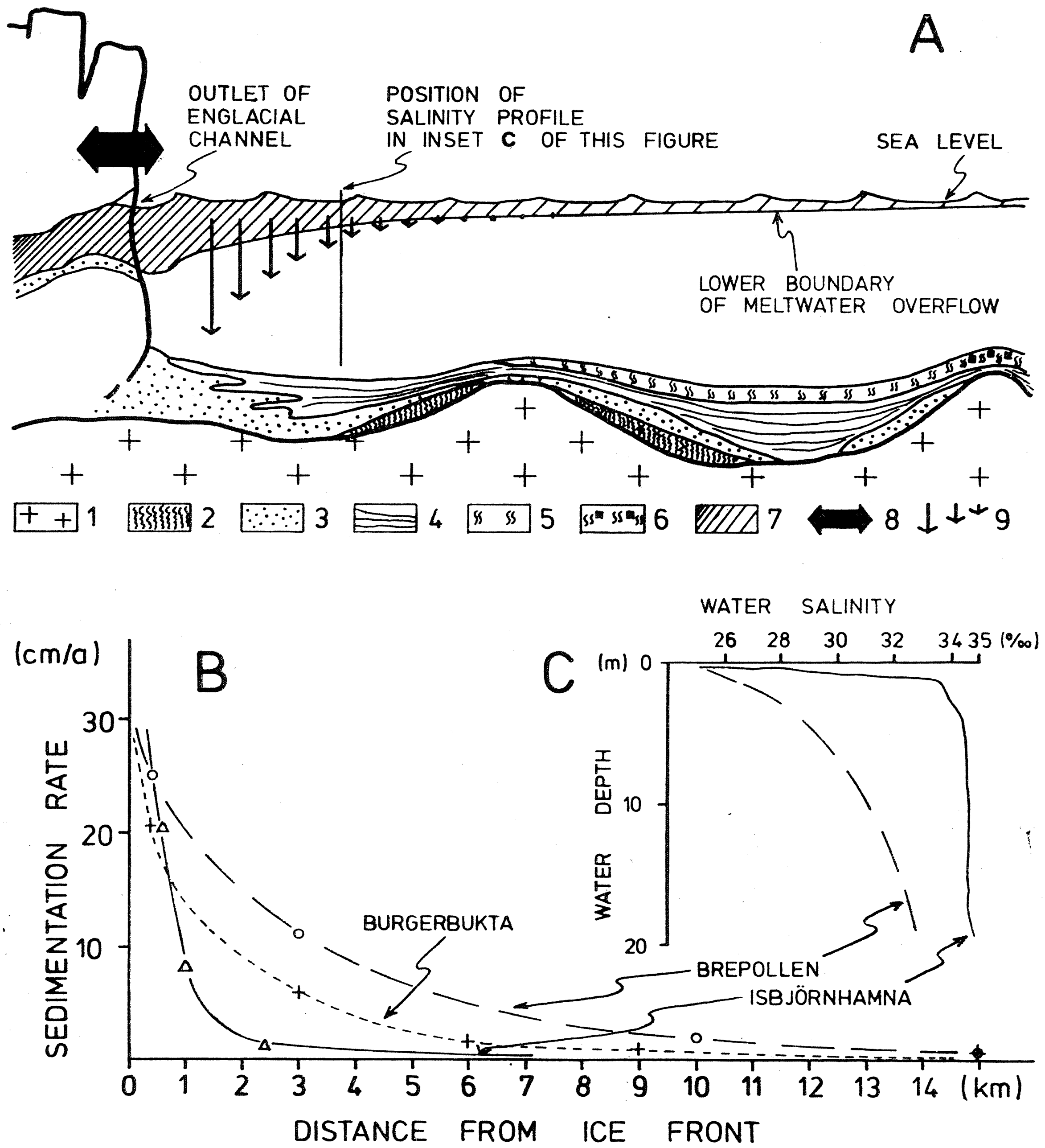


Fig. 9. Relations between sedimentation rate and facies distribution as well as between vertica salinity gradient and sedimentation rate, in front of tidewater glacier. *A* — cross-section of the margin of the grounded retreating tidewater glacier and of the near-source fragment of fjord; indicated are distribution and succession of facies as well as pattern of suspension settling at an outlet of englacial channel: *1* — bedrock, *2* — till (mainly surge deposits), *3* — submarine outwash (facies A); *4* — laminated mud (facies B), *5* — homogeneous to bioturbated mud (facies C), *6* — increased IRD occurrence in facies C, *7* — sediment-laden meltwater plume gradually diluted, *8* — ice-front of oscillatory retreating glacier, *9* — arrows indicating intensity of suspension settling; *B* — empirical curves of accumulation-rate decay based on estimated accumulation rates for selected subbasins of Hornsund (Isbjörnhamna, Burgerbukta and Brepollen); *C* — measured vertical salinity profiles in Isbjörnhamna and Brepollen (according to Urbański *et al.*, 1980)

Table 6

Comparison of accumulation rates estimated from sediment thicknesses with sedimentation rates derived from Syvitski's formula

Area	Main source glacier	Station	Source rate b [cm/a]	Coefficient [m]	Estimated accumulation rate [cm/a]	Syvitski's sedimentation rate [cm/a]
Brepollen	Horn	20	25	−0.1	11	12
Treskellen Narrows	Horn	13	25	−0.1	< 1	3
Outer Samarinvågen	Samarin	12	20	−0.1	8	8
Austre Burgerbukta	Mühlbacher	16	20*)	−0.2	6	5
Austre Burgerbukta	Mühlbacher	15	20*)	−0.2	1	1
Fjord	Mühlbacher	14	20*)	−0.2	0.5	0.4
Fjord	Horn, Samarin Mühlbacher	A-B	25	−0.2	0.2	0.1
Isbjörnhamna	Hans	22	20	−0.4	7	8
Isbjörnhamna	Hans	21	20	−0.4	1	3

*) Assumed source sedimentation rates (other rates are taken from estimates based on ice--proximal coring data).

value). For the glacier-fed river sedimentation in fjords of British Columbia, studied by Syvitski and Murray (1981) and Farrow *et al.* (1983), this coefficient is −0.1.

It is well documented that electrochemical flocculation plays an important role in settling of clays in front of tidewater glaciers (e.g. Gilbert, 1983). Nevertheless, the question whether vertical salinity gradient exerts significant influence on the slope of decay curve for suspended load, is not definitely settled. This question might have been solved if we had an example of the above-mentioned cancelling effect (beside the example of amplification recorded in Isbjörnhamna). Unfortunately, all studied subbasins seem to represent "in-phase" situations.

Selective settling

As regards the selectivness of settling, it is also connected both, with gravitational effects (characteristic modal size of non-aggregated flakes of a given clay-mineral species) and electrochemical effects (specific ability to build large floccules).

The results of this study suggest that selectiveness in settling of clays exists and is primarily governed by gravitational effect. Within the fjord, the clay minerals of typically large flakes (biotite, muscovite, kaolinite, metamorphic chlorite) preferentially settle out of suspension, leading to increased concentration in the more distal sites, of clay minerals of typically smaller sizes (e.g. illite, diagenetic chlorite).

This is illustrated in Table 1 by the clay-mineral composition of the outer-fjord sample 800, where illite content is high (at the level typical of Brepollen samples) and much increased in comparison with the samples from the nearest source area in Isbjörnhamna (sample 072), whereas biotite content in sample 800 is the lowest of those recorded in Hornsund. Additional illustration is given in Table 2, where along the short distance, preferred settling of biotite is conspicuous.

The latter example seems to question electrochemical controls on differential settling. In case of favoured flocculation of the smallest flakes, illite should compete in settling rate with large biotite, muscovite or chlorite flakes, cancelling the effect of gravitational sorting. In Isbjörnhamna, this is not the case. On the other hand, high concentration of the finest clays (higher than in Isbjörnhamna and outer fjord) was found in the inner fjord and Brepollen sediments, suggesting their enhanced settling by flocculation (compare Table 4 and Pl. I). In addition, approximately parallel settling of all clay mineral species, independently of their characteristic grain sizes, which takes place in Brepollen and inner fjord (as indicated by ratios for bulk samples and fraction $< 2\ \mu m$ in Table 3), also suggests that settling of the finest particles is electrochemically aided. Low zooplankton concentration (*cf.* section Setting), and rarely encountered (in SEM micrographs) pellets within the fjord sediments, rule out possibility of biological control over pattern of settling of the finest clays. Thus, flocculation must anyway be invoked to explain the picture.

Essential features of suspension settling

The author concludes that electrochemical flocculation (occurring by compaction of electrical double layer at the charged surface of clay minerals brought into marine electrolyte) takes intensively place in the studied sedimentary basin. However, it does not favour any clay species or size, *ergo* it is not selective. This flocculation is responsible for mass settling of the finest clays in the proximity of glacial source (see Table 4 and Fig. 9).

The concentration of the finest clays is high in the sediment only if such clays are proportionally abundant in the source suspension and if specific grain sizes for the clay minerals present do not differ much (as is the case in Brepollen). If the finest clays are diluted with respect to other components of the suspension, and modal grain-sizes for clay mineral species differ largely, then gravitational sorting of various clay minerals becomes apparent along the transportation path (as is the case in Isbjörnhamna). Due to this inferred non-selective property of coerced settling, the finest clays indeed travel the longest distances and are relatively concentrated in the most distal site. This is illustrated by composition of distal samples (in Table 1 samples: 800, KAG 4, KAG 5), as well as by distribution of clay mineral provinces in Figure 10.

It is also a specific feature of coerced settling of fine suspension in the proximal basins that glacimarine muds are deposited nearly independently of water depth, i.e. also in the sites close to the sea level (see Tables 1 and 2 for samples H_7, 029, and 031).

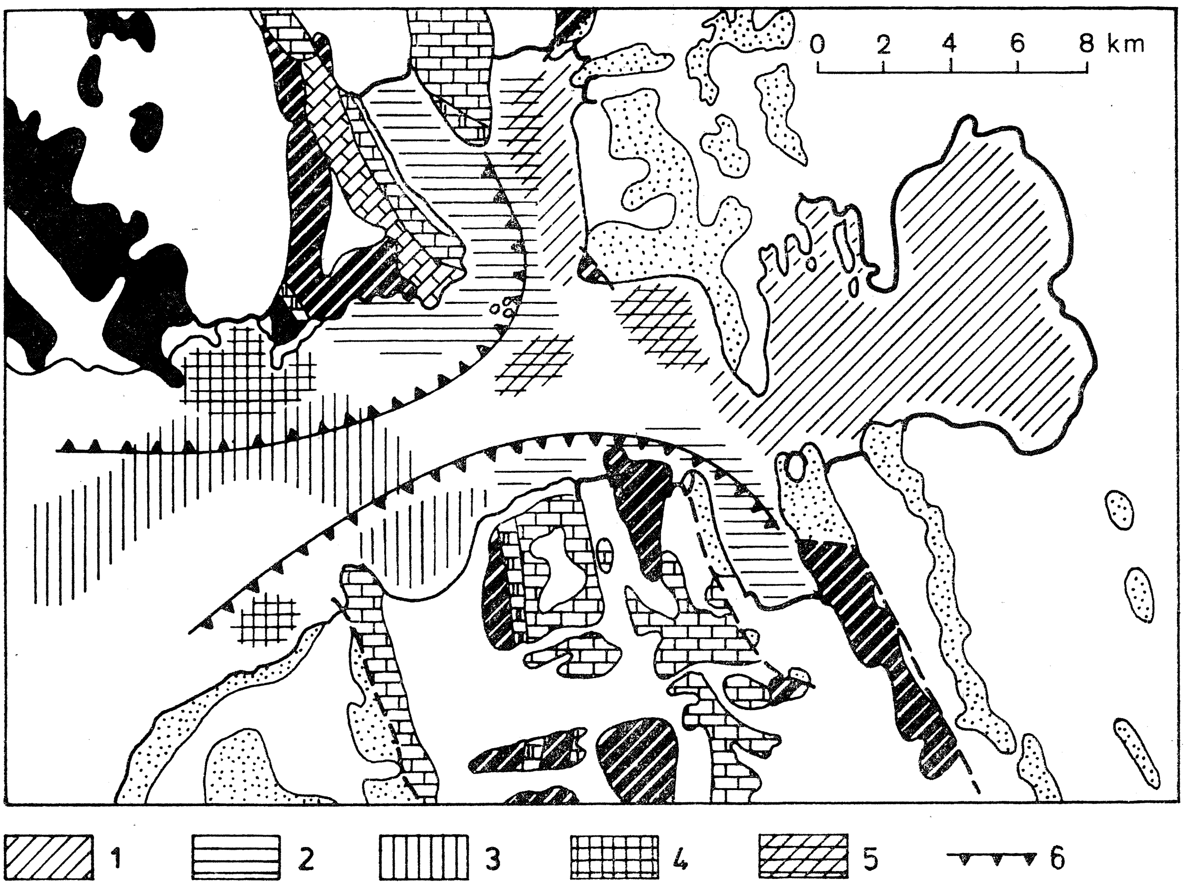


Fig. 10. Sketch of mineral-facies distribution (mineral provinces) in Hornsund sediments. Lithology and other symbols on land same as in Fig. 2. Symbols within the fjord: *1* — quartz-feldspar facies (Q-Fsp); *2* — carbonate-quartz facies (Cb-Q); *3* — carbonate-feldspar facies (Cb-Fsp); *4* — inherited (mixed on land) quartz-feldspar-carbonate facies (Q-Fsp-Cb inh); *5* — fjord-mixed quartz-feldspar-carbonate facies (Q-Fsp-Cb mix); *6* — boundary between the muscovite-biotite facies (M-B) on the cogged side of the line, and illite facies (I) on the smooth side of the line

SEDIMENT PROVENANCE AND MINERAL INHERITANCE

Mineral provinces

Basing on the present results, one can distinguish mineral facies and, respectively, mineral provinces in glacimarine muds of Hornsund. These provinces are tightly related to the bedrock lithologies in the individual source areas or they represent blended sediment supplied from multiple sources.

The following provinces are distinguished according to the distribution of the non-clay mineral facies (Table 5 and Fig. 10):

(1) quartz-feldspar or siliciclastic province in Brepollen, Austre Burgerbukta, and Storfjorden,

(2) quartz-feldspar-carbonate province in Isbjörnhamna and inner central fjord,

(3) carbonate-quartz province in Vestre Burgerbukta and Samarinvågen,

(4) carbonate feldspar province in Gåshamna and outer fjord.

The second of the distinguished provinces is genetically inhomogeneous (Fig. 10). In Isbjörnhamna it reflects mixed composition of original fluvioglacial sedimentary material, whereas in central fjord and also in central Burgerbukta, it results from blending of sedimentary materials coming from different sources as meltwater load (Burgerbukta) or ice-rafted debris (central fjord).

Independently of the above non-clay provinces, one can distinguish in Hornsund two basic clay mineral facies and corresponding provinces (Tables 1 to 4, Fig. 10):

(1) muscovite-biotite facies, with domination of $2M_1$ muscovite and biotite over 1Md illite, low di-micas/tri-micas ratio, and low DR/PM ratio (see p. 449 and Table 4),

(2) illite facies, with domination of clay-grade 1Md illite, high di-micas/tri-micas ratio, and high DR/PM ratio.

The corresponding mineral provinces are located as follows:

(1) muscovite-biotite province in Isbjörnhamna, Gåshamna, Vestre Burgerbukta, Sofiebogen, inner Samarinvågen, and Skoddebukta.

(2) illite province in Brepollen, outer Samarinvågen, Austre Burgerbukta, central fjord, and Storfjorden.

As regards kaolinite plus chlorite to micas, and chlorite to micas ratios (Table 3), outer fjord and Storfjorden areas reveal intermediate compositions of clay mineral assemblages.

Distribution of clay mineral provinces is determined by extent of sedimentary domination of individual glacial sources. These single-source provinces occupy the forefields of glaciers, and their lateral extent depends on intensity of meltwater input and the slope of suspended load decay (see previous section). Hence, the illite facies, related to Brepollen and Austre Burgerbukta sources (with high meltwater input and coefficient m between -0.1 and -0.2) dominates with purely single--source features up to a distance of about 10 km. The muscovite-biotite facies, related in the first place to Isbjörnhamna and Vestre Burgerbukta sources (both characterized by low meltwater input) extends with the single-source features up to about 2 km in Isbjörnhamna (by the decay coefficient $m = -0.4$) and up to about 5 km in Vestre Burgerbukta (by estimated decay coefficient $m = -0.2$).

Inherited features of clay minerals

Distinct boundaries between the provinces suggest that clay minerals are inherited from the different source rocks unchanged (and thus nonhomogenized) by the currently operating processes. Neither weathering, nor soil processes, nor marine environment, are effective in obliterating the inherited features. These diagnostic features, transmitted in Hornsund from the bedrock to sediment, are for the Precambrian to lower Palaeozoic first-cycle clays, as follows: (1) distinctly polymodal grain-size spectrum of clay mineral assemblage, (2) dominance of $2M_1$ muscovite over clay-grade illite, (3) abundance of ferrous-ferric chlorite and biotite, (4) the ratio: dioctahedral/trioctahedral micas between 1.0 and 2.5, (5) DR/PM ratio $\ll 0.2$. The respective finger-print features of the upper Palaeozoic to Mesozoic (to some extent also Tertiary) multiple-cycle and diagenetic clays are: (1) conform grain-size spectra of the clay mineral species, (2) dominance of 1Md illite rich in (Fe+Mg) over muscovite, biotite and large-flake metamorphic chlorite, (3) relative abundance of kaolinite, (4) more than 2.5-fold prevalence of dioctahedral over trioctahedral micas, (5) small overall size of clays, DR/PM ratio $\geqslant 0.2$.

As regards the DR/PM ratio, its denotion here: diagenetic-recycled/primary-metamorphic must not be taken literatim. By no means the fraction $< 0.2\ \mu m$ is composed solely of diagenetic and multiply recycled clays, as well as fraction $0.2-2\ \mu m$ is not composed exclusively of primary and metamorphic clays. The values of the ratio imply only that for the studied Subarctic fjord, one can distinguish sedimentary material derived of metamorphic rocks from the material eroded from siliciclastic bedrock, basing on the texture of clay fraction. For the studied sediments, the boundary value is arbitrarily chosen at DR/PM = 0.2. This, however, cannot be universal value, and presumably the only universal property of this ratio is that within subpolar and polar zones the sediments with clastic source rocks have this ratio much higher than those with metamorphic or igneous source rocks.

Genetic interpretation of mineral provinces

The discrepancy between the distribution of non-clay and clay mineral provinces in the sediments of Hornsund invokes obvious conclusion of different dominant transportation modes for the two mineral groups in distal zones. The non-clay minerals are predominantly ice-rafted, whereas clay minerals are mainly carried within suspension.

The driving forces for suspension dispersal are: (1) meltwater jet in proximal zone, and (2) fjord-type circulation plus drift currents in distal zone.

Intensity of ice rafting is generally low in Hornsund. Two main ice carriers of sedimentary material, i.e. glacier ice and pack ice from the Barents Sea, tend to distribute IRD randomly over the fjord bottom. The most concentrated deposition from icebergs and grawlers occurs in some proximal sites where icebergs are frequently trapped by sills and shoreline morphology. Such situation is observed in inner Austre Burgerbukta as well as northeastern and southeastern parts of Brepollen. There, facies B (laminated mud) contains numerous drop-stones. Nevertheless, in majority of proximal subbasins, in facies B iceberg drop-stones are rare. This situation is the net effect of high sedimentation rate of meltwater suspension and, on the other hand, low intensity of glacier calving and moderate concentration of debris in glacial ice. The intensity of supply of sedimentary material from icebergs is more or less uniform beyond the above-mentioned proximal areas. This intensity falls to about zero over the shelf.

In turn, ice rafting by the Barents Sea pack ice is most intensive over the Hornsund shelf and in the central part of the fjord. This may be induced from the frequency maps of pack ice cover in Hornsund (Görlich & Stepko, *in print*).

The complementary pattern of iceberg and pack ice sedimentations in Hornsund results in overall random supply of ice-rafted debris to the sediments of the fjord and shelf.

One may expect that by exponentially decreasing rate of suspension sedimentation and approximately random distribution of IRD sedimentation rate, there should be noted relative enrichment of muds in non-clay components with growing distance from the source. This may indeed be observed when one compares the proximal samples from Hornsund (at distances up to 3 km from the nearest sediment source, water depths in a range 5 to 105 m, listed in Tables 1 and 2) with the more distal samples (at distances from 8 to 10 km from the nearest sediment source, water depths in a range 78 to 175 m, listed in Table 1). For the first set of samples, average clay mineral content is 54%, while for the other (more distal) it equals to 43%. A similar tendency is observed in Storfjorden (Spitsbergen east coast), where sample KAG 4 (at a distance of 10 km from the nearest sediment source, water depth 45 m) contains 43% of clays, whereas sample KAG 5 (at a distance of 50 km from the nearest sediment source, water depth 111 m) contains 41% of clays.

The above effect blurs grain-size distribution which in areas with high intensity of ice-rafting cannot be used to establish sediment mixing ratios for multiple sources. This is in contrast to what can be done by the absence of significant ice-rafting (e.g., in Howe Sound, British Columbia in paper by Syvitski & Macdonald, 1982). The bimodal grain-size distribution and inversed lateral gradation of texture are, by the presence of ice-rafting, inherent properties of the single-source glacimarine sediments.

ORGANIC-INDUCED LAMINATION

Lamination in facies B

Milimeter-thick organic-rich surface lamina was observed in sediment samples recovered from the deepest parts of Isbjörnhamna in May-June. The dark "monosulphidic" lamination is common in the high-accumulation-rate sediments (facies B). This lamination is only occasionally encountered in the low-accumulation-rate mud (facies C). Seasonal condensation of organic material in sediment, giving rise to monosulphidic lamination, is well explained by Elverhøi *et al.* (1980); see p. 435. The concentrated organic matter will produce reduced laminae only in sites where: (1) bottom water exchange is restricted (although water itself is oxygenated), (2) no sediment winnowing occurs, (3) sediment mixing by bioturbation is negligible, and (4) sediment accumulation rate is high enough to inhibit oxidation at the sediment surface.

Low accumulation rate and, hence, long exposition of sediment to the oxidative environment of sediment/water interface, as well as intensive bioturbation, inhibit formation of dark laminae. However, in case of strongly restricted fjords where oxygen-depleted bottom water is present, anoxic conditions may freely develope also in the sediment of low accumulation rate. There, reducing environment involves bulk sediment rather than separate laminae.

The conditions favouring formation of organic lamination are fulfilled within the troughs in forefields of the Hornsund tidewater glaciers (in facies B).

Authigenic siderite

With the above described laminae, related are increased concentrations of siderite. For example, a 1.5 cm thick layer of "organic mud" (sample 028/5 in Table 2), contains 10% siderite while the underlying bright-bege clayey mud (sample 028/6 in Table 2) contains 4% siderite. The black lamina of sample 799 contains 5% siderite, compared with 2% siderite in the bright layer of the same sample.

The authigenic siderite may be seen in IR spectra, concentrated in the finest fraction of the 2 cm thick top (spring) dark layer extracted from sample 069 (Fig. 5E). SEM investigation of the siderite-bearing samples from Isbjörnhamna did not, however, unambiguously confirm the presence of euhedral siderite microcrystals. One may infer such presence in the sample 069 (Pl. I: 1 and 3) where besides gypsum crystals (crystallized during air-drying of the sample) there are rhombohedral (?) microcrystals of siderite (marked with S).

In the samples from the areas of low accumulation rate (e.g., samples 819 from Adriabukta in inner Hornsund, and 042 from Vestre Burgerbukta) the sediment is relatively enriched in organic remains and fecal pellets. In spite of their feasible oxidation (due to slow burial) which inhibits formation of reducing conditions in the bulk sediment or laminae, some authigenic siderite is observed in the immediate vicinity of fecal pellets or organic remains (Pl. II: 1 and 4).

There is a general tendency in the studied muds to dissolve carbonates. The content of carbonates decreases with sediment depth in all the cores studied. The fact is best seen in rapidly accumulating sediments of Brepollen (facies B) where average drop of carbonate content (for 4 studied cores) is 0.6 wt. % per meter of sediment depth. Since the discussed cores represent roughly 30 years of deposition, and only decreasing-downcore carbonate distribution is observed, it is unlikely that this drop results from the compositional fluctuations within sedimentary material. Thus there are enough carbonate ions in interstitial waters to react in the locally reducing environments with ferrous ions to precipitate as siderite; a process commonly observed in the estuarine-marine muds, e.g. in the Baltic Sea (Görlich *et al.*, 1978).

By the excess carbonate ions in pore waters, the siderite will not dissolve when by deeper burial, oxidative Eh is again established. Hence, the stratiform presence of concentrations of fine siderite will mark primarily organic lamination.

It must be noted that the studied glacimarine muds are never reducing in their bulk. They attain at most post-oxic (Berner, 1981) stage. Only locally, in dark laminae or spots, sulphate reduction occurs, indicated by presence of black, X-ray amorphous iron monosulphides. By scarce production of sulphide ions, pyrite is rarely formed. It is observed in SEM micrographs only in the immediate vicinity, of organic remains and never as complete framboids (Pl. II: 6). For the same reason, i.e. defficiency of reduced sulphur, it is not anticipated that iron monosulphides will eventually be transformed to pyrite. The very unstable monosulphides will rather be oxidized and hydrolize during proceeding burial, to form ferric oxyhydroxides. Thus, siderite is the only relatively stable form testifying to primary organic-induced lamination.

GLACIMARINE MUD FACIES

Different glacimarine mud facies are distinguished in fjord sediments and different names and diagostic features are ascribed to them (see also p. 435). These facies may be distal in the scale of the fjord or in the meaning of Elverhøi *et al.* (1983) but they are proximal according to classification of Boulton and Deynoux (1981) or when one considers glacimarine sediments as being delimitated laterally by drop-stone occurrence. Hence, the present author rather refers to them as laminated mud (facies B) and homogeneous to bioturbated mud (facies C).

Only the above two facies are discussed in this paper, although the author is aware of much more complex than bipartite, lithology of muds within the fjord. The reason for such simplifying approach is that this study aimed at demonstrating primary processes of deposition of meltwater and ice-rafted loads. Both, small-scale redistribution of sediments by gravity flows and local pecularities of sedimentation, had to be included only when they obscured general pattern of sedimentation.

The following descriptive and genetic features are ascribed in this study to the laminated mud facies (B):

(1) presence of dark organic-induced lamination,

(2) common textural lamination produced by pulses of sediment-laden meltwater plume,

(3) scarce infauna,

(4) very rare plankton tests and fecal pellets,

(5) clay mineral assemblage preserving features typical of the muds in the most proximal positions for the given source glacier,

(6) accumulation rate greater than 5 cm/a,

(7) ponded filling of bottom depressions.

Respective features of the homogeneous to bioturbated mud facies (C) are as follows:

(1) occasionally dark monosulphidic spots, locally dark lamination,

(2) very faint textural inhomogeneities, occasionally with coarse intercalations,

(3) strong bioturbation by the abundant fauna,

(4) plankton tests and fecal pellets present in limited quantities,

(5) clay mineral assemblage of blended nature; no direct correspondence to the muds in the proximity of feeding glaciers,

(6) accumulation rates lower than 5 cm/a,

(7) deposition mode mantling bottom topography.

The above features are better applicable to distinguishing of facies sequences in cored sediment column than to precise delimitating of lateral facies distribution in the present-day sediment.

CONSLUSIONS

The process of mass entrapment of terrigenous fines in the proximity of tidewater glaciers in Hornsund includes:

(1) strong destabilizing action of sea water on suspension which is introduced

to the fjord with the meltwater freshet; the destabilization (of electrochemical nature) is not selective towards any of the component clay mineral species,

(2) gravitational grain-size sorting along the transportation pathway which occurs also within the fine fraction of suspension (< 2 μm) and is ultimately responsible for the observed selective settling of different clay species.

The steepness of suspension decay is various for individual tidewater-glacier sources and is related here both to the different intensity of destabilizing action of sea water (depending on the vertical salinity gradient in the discharge area, Fig. 9) and to the different quality of the initial suspended load (reflected by the grain--size distributions within the clay fraction).

Due to the non-selective property of the coerced (electrochemical) settling of clays, the clay minerals of typically finest grain sizes (e.g., 1Md illite, diagenetic chlorite) travel the longest way and are preferentially carried out of the fjord onto the shelf.

Sediments of the distinguished provinces in Hornsund bottom very well reflect lithologies of the bedrock in the catchments of feeding glaciers. Hence, the provenance of sediments is easy to be traced back. Four non-clay mineral facies are distinguished in Hornsund: quartz-feldspar (siliciclastic), quartz-feldspar-carbonate, carbonate-quartz, and carbonate-feldspar ones.

Clay-mineral facies could also be distinguished and the respective provinces delimitated in the Hornsund bottom (Fig. 10). Boundaries of the non-clay mineral provinces do not fit to those of clay mineral ones. This is primarily due to: (1) different behaviours of clays (with physicochemically active surfaces) and non-clays (with inert surfaces) during settling of suspension in the proximal sites, and (2) different modes of transportation of clay and non-clay minerals in distal settings. The clay minerals are mainly carried in suspension, whereas the non-clay ones by ice-rafting.

Analysis of clay minerals in sediments of individual single-source subbasins, proved that clays in glacimarine sediments preserve distinct features of the source rocks on land. These features are, in the Subarctic environment of Hornsund catchment, nearly directly transmitted to the marine muds, apparently not altered by weathering, soil process or impact of marine environment.

Siderite authigenesis (mainly in the rapidly accumulating muds of facies B) is found to be related with the initial presence of concentrated organic matter in the sediment. Authigenic siderite occurs in dark (monosulphidic) laminae and also locally in the vicinity of organic remains or fecal pellets in the anoxic spots. Contrary to iron monosulphides which are very unstable in oxidative environments, siderite is supposed to be preserved in the sediment after the rise of Eh during proceeding burial.

Presence or absence of organic-induced lamination, of benthic life and organic remains, different sediment accumulation rates, as well as integrity of source finger--printing clay mineral assemblages, constitute criteria for distinguishing two glacimarine mud facies in Hornsund.

Strong positive dependence of sedimentation rate on the vertical salinity gradient in the forefields of tidewater glaciers, is a main factor controlling efficiency of settling of suspension. These vertical salinity gradients are, as a rule, greater in the forefields of tidewater glaciers compared to estuaries of glacier-fed rivers, resulting is mass entrapment of suspension in the near-source zones of tidewater-glacier systems. Due to this effect, the tidewater-glacier systems contribute only minor quantities of sediments to the shelf and ocean, contrary to the glacier-fed river sources. This, and not extremely low meltwater discharge, may account for the scarcity of the meltwater sediments in most of the Arctic basins and on the Antarctic shelf.

Acknowledgements

Cooperation in the field of Dr. S. Rudowski and all the members of the Spitsbergen Expedition 1982/83 (organized by the Institute of Geophysics, Polish Academy of Sciences) is gratefully appreciated.

Thanks are due to Cezary Filipowicz, Wojciech Moskal, Jan Marcin Węsławski and Marek Zajączkowski for providing the author with sediment samples from Brepollen, Samarinvågen and Storfjorden.

The author expresses his thanks to Dr. S. M. Zalewski and ing. W. Kowalewski from the Institute of Geophysics, Polish Academy of Sciences, for their cooperation during m/s „Jantar" cruise in 1985, and specially for offering access to sediment cores and CSP profiles from Hornsund.

The present research was carried out within the project B.8/MR.I.29 supervised by Prof. K. Birkenmajer.

The paper largely benefited from reviewing by Dr. S. J. Porębski, Dr. J. Środoń and Prof. R. Gradziński, whose help is gratefully acknowledged.

REFERENCES

Anderson, J. B., Brake, C., Domack, E. W., Meyers, N. & Wright, R., 1983. Development of a polar glacial-marine sedimentation model from Antarctic Quaternary deposits and glaciological information. In: Molnia, B. F (ed.), *Glacial-marine sedimentation*. Plenum Press, New York, pp. 233—264.

Anderson, J. B., Kurtz, D. D., Domack, E. W. & Balshaw, K. M., 1980. Glacial and glacial marine sediments of the Antarctic continental shelf. *J. Geol.*, 88: 399—414.

Antkiewicz, A. & Filipowicz, C., 1978. Bottom sediments of Isbjörnhamna. (In Polish.) *Sympozjum Polarne Polskiego Towarzystwa Geograficznego*, Gdańsk—Gdynia. 2: 7—11.

Berner. R. A., 1981. A new geochemical classification of sedimentary environments. *J. Sedim. Petrol.*, 51: 359—365.

Birkenmajer, K., 1960. Geological sketch of the Hornsund area. *Intern. Geol. Congr. XXI Sess. Norden, Suppl. to Guide, Exc. A. 16, Aspects of the Geology of Svalbard*, pp. 1—12.

Birkenmajer, K., 1964. Devonian, Carboniferous and Permian formations of Hornsund, Vestspitsbergen. *Stud. Geol. Polon.*, 11: 47—123.

Birkenmajer, K., 1975. Jurassic and Lower Cretaceous sedimentary formations of SW Torell Land, Spitsbergen. *Stud. Geol. Polon.*, 44: 7—43.

Birkenmajer, K., 1977. Triassic sedimentary formations of the Hornsund area, Spitsbergen. *Stud. Geol. Polon.*, 51: 7—74.

Birkenmajer, K., 1978a. Cambrian succession in south Spitsbergen. *Stud. Geol. Polon.*, 59: 7—46.

Birkenmajer, K., 1978b. Ordovician succession in south Spitsbergen. *Stud. Geol. Polon.*, 59: 47—82.

Birkenmajer, K., 1984a. Mid-Carboniferous red beds at Hornsund, south Spitsbergen: their sedimentary environment and source area. *Stud. Geol. Polon.*, 80: 7—23.

Birkenmajer, K., 1984b. Cyclic sedimentation in mixed alluvial to marginal-marine conditions: the Treskelodden Formation (?Upper Carboniferous and Lower Parmian) at Hornsund, south Spitsbergen. *Stud. Geol. Polon.*, 80: 25–45.

Biscaye, P. E., 1964. Distinction between kaolinite and chlorite in recent sediments by X-ray diffraction. *Am. Miner.*, 49: 1281–1289.

Biscaye, P. E., 1965. Mineralogy and sedimentation of recent deep sea clays in the Atlantic Ocean and adjacent seas and oceans. *Geol. Soc. Am. Bull.*, 76: 803–832.

Boulton, G. S. & Deynoux, M., 1981. Sedimentation in glacial environments and the identification of tills and tillites in ancient sedimentary sequences. *Precam. Res.*, 15: 397–422.

Brown, G. & Brindley, G. W., 1980. X-ray diffraction procedures for clay mineral identification. In: Brown, G. & Brindley, G. W. (eds.), Crystal structures of clay minerals and their X-ray identification. *Mineralogical Society, London, Monograph*, 5: 305–359.

Domack, E. W., 1984. Rhythmically bedded glaciomarine sediments on Whidbey Island, Washington. *J. Sedim. Petrol.*, 54: 589–602.

Elverhøi, A., 1984. Glacigenic and associated marine sediments in the Weddell Sea, fjords of Spitsbergen and the Barents Sea. *Mar. Geol.*, 57: 53–88.

Elverhøi, A., Liestøl, O. & Nagy, J., 1980. Glacial erosion, sedimentation and microfauna in the inner part of Kongsfjorden, Spitsbergen. *Norsk Polarinstitutt Skrifter*, 172: 33–60.

Elverhøi, A., Lønne, Q. & Seland, R., 1983. Glaciomarine sedimentation in a modern fjord environment, Spitsbergen. *Polar. Res.*, 1: 127–149.

Elverhøi, A. & Roaldset, E., 1983. Glaciomarine sediments and suspended particulate matter, Weddell Sea shelf, Antarctica. *Polar Res.*, 1:1–21.

Eyles, N. & Miall, A. D., 1984. Glacial facies. In: Walker, R. G. (ed.), *Facies models.* Geoscience Canada, Reprint Series, 1: 15–39.

Farrow, G. E., Syvitski, J. P. M. & Tunncliffe, V., 1983. Suspended particulate loading on the macrobenthos in a highly turbid fjord: Knight Inlet, British Columbia. *Can. F. Fisheries Aquatic Sci.*, 40: 273–288.

Gilbert, R., 1983. Sedimentary processes of Canadian Arctic fjords. *Sedim. Geol.*, 36: 147–175.

Gorlich, K., *in print*. Sedimentation of glacimarine muds by the tidewater glacier, Hornsund fjord, Spitsbergen. In: *Spitsbergen '84*. Ossolineum, Wrocław.

Görlich, K., Görlich E. A. & Stoch, L., 1978. Iron in the Baltic Sea clay sediments. *Miner. Polon.*, 9: 63–79.

Görlich, K. & Stepko, W., *in print*. Hydrological phenomena related to sea-ice formation and presence, observed in Hornsund, Spitsbergen. In: *Spitsbergen '84,* Ossolineum, Wrocław.

Griffin, J. J., Windom, H. & Goldberg, E. D., 1968. The distribution of clay minerals in the World Ocean. *Deep-Sea Res.*, 15: 433–459.

Jackson, M. L., 1975. *Soil chemical analysis – advanced course.* Publ. by the author, Madison, Wisconsin, 895 pp.

Klug, H. P. & Alexander, L. E., 1974. *X-ray diffraction procedures for polycrystalline and amorphous materials.* Wiley, New York, 966 pp.

Kravitz, J. H., 1976. Textural and mineralogical characteristics of the surficial sediments of Kane Basin. *J. Sedim. Petrol.*, 46: 710–725.

Kravitz, J. H., 1983. Glacial and glacial-marine sediment lithofacies of the Kane Basin. In: Molnia, B. F. (ed.), *Glacial-marine sedimentation.* Plenum Press, New York, pp. 401–450.

Lindner, L., Marks, L. & Pękala, K., 1983a. Quaternary glaciations of South Spitsbergen and their corralation with Scandinavian glaciations in Poland. *Acta Geol. Polon.*, 33: 169–182.

Lindner, L., Marks, L. & Pękala, K., 1983b. Late Quaternary glacial episodes in the Hornsund region of Spitsbergen. *Boreas*, 13: 35–47.

Mackiewicz, N. E., Powell, R. D., Carlson, P. R. & Molnia, B. F., 1984. Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. *Mar. Geol.*, 57: 113–147.

Molnia, B. F., 1983. Subarctic glacial-marine sedimentation: a model. In: Molnia, B. F. (ed.), *Glacial-marine sedimentation.* Plenum Press, New York, pp. 95–144.

Molnia, B. F. & Hein, J., 1982. Clay mineralogy of a glacially dominated shelf: northwestern Gulf of Alaska. *J. Sedim. Petrol.*, 52: 515–527.

Osterman, L. E. & Andrews, J. T., 1983. Changes in glacial-marine sedimentation in core HV77-159, Frobisher Bay, Baffin Island N.W.T.: a record of proximal, distal, and ice-rafting glacial marine environments. In: Molnia, B. F. (ed.), *Glacial-marine sedimentation.* Plenum Press, New York, pp. 451—494.

Pawełczyk, J. & Środoń, J., 1978. Quantitative X-ray analysis of mineral composition in shales. (In Polish.) *I Konf. Minerały i Surowce Ilaste, Bolesławiec,* pp. 211—220.

Piper, D. J. W. & Slatt, R. M., 1977. Late Quaternary clay-mineral distribution on the eastern margin of Canada. *Geol. Soc. Amer. Bull.* 88: 267—272.

Powell, R. D., 1981. A model for sedimentation by tidewater glaciers. *Ann. Glaciology,* 2: 129—134.

Powell, R. D., 1984. Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater sediments based on Quaternary examples. *Mar. Geol.,* 57: 1—52.

Radwański, A. & Birkenmajer, K., 1977. Oolitic/pisolitic dolostones from the late Precambrian of south Spitsbergen: their sedimentary environment and diagenesis. *Acta Geol. Polon.,* 27: 1—39.

Siedlecka, A., 1968. Lithology and sedimentary environment of the Hyrnefjellet Beds and the Treskelodden Beds (Late Palaeozoic) at Treskelen, Hornsund, Vestspitsbergen. *Stud. Geol. Polon.,* 21: 53—95.

Smulikowski, W., 1965. Petrology and some structural data of lower metamorphic formations of the Hecla Hoek Succession in Hornsund, Vestspitsbergen. *Stud. Geol. Polon.,* 18: 7—107.

Środoń, J., 1981. X-ray identification of randomly interstratified illite-smectite in mixtures with discrete illite. *Clay Minerals,* 16: 297—304.

Środoń, J., 1984. Mixed-layer illite-smectite in low-temperature diagenesis: data from Miocene of the Carpathian Foredeep. *Clay Minerals,* 19: 205—215.

Środoń, J., Eberl, D. D., Northrop, H. R., Mingchou Lee & Nadeau, P. H. *Sericites from the Silverton Caldera, San Juan Mountains, Colorado.* Manuscript in preparation.

Swerpel, S., *in print (a).* The Hornsund fjord: water masses. *Pol. Polar Res.*

Swerpel, S., *in print (b).* On water transport in the Hornsund fjord coastal zone. *Pol. Polar Res.*

Syvitski, J. P. M. & Macdonald, R. D., 1982. Sediment character and provenance in a complex fjord; Howe Sound, British Columbia. *Can. J. Earth Sci.,* 19: 1025—1044.

Syvitski, J. P. M. & Murray, J. W., 1981. Particle interaction in fjord suspended sediment. *Mar. Geol.,* 39: 215—242.

Urbański, J. Neugenbauer, E., Spacjer, R. & Falkowska, L., 1980. Physico-chemical characteristic of the waters of Hornsund Fjord on south-west Spitsbergen (Svalbard Archipelago) in the summer season 1979. *Pol. Polar Res.,* 1: 43—52.

Węsławski, J. M. & Kwaśniewski, S., 1983. Application of biological indicators for determination of the reach and origin of sea currents within the region of Spitsbergen. *Pol. Arch. Hydrobiol.,* 30: 189—197.

Węsławski, J. M., Zajączkowski, M., Moroz, R. & Kwaśniewski, S., 1985. *Preliminary report from the cruise of m/s "Jantar" to South Spitsbergen area — August 1984.* Manuscript in the Institute of Oceanology, Polish Academy of Sciences, Sopot.

Yariv, S. & Cross, H., 1979. *Geochemistry of colloid systems.* Springer Verlag, Berlin, 450 pp.

Streszczenie

GLACJALNO-MORSKA SEDYMENTACJA MUŁÓW WE FIORDZIE HORNSUND NA SPITSBERGENIE

Krzysztof Görlich

Abstrakt: Zbadano mineralogicznie i sedymentologicznie muły morsko-lodowcowe zalegające dno fiordu Hornsund na Spitsbergenie. W osadzie wyróżniono prowincje mineralne odzwierciedlające pochodzenie materiału osadowego. Określono przestrzenny rozkład stopy akumulacji mułów.

Wykładniczy spadek stopy akumulacji w kierunku od czoła lodowca zachodzi w poszczególnych basenach przyczołowych z różną prędkością. Prędkości te zależą od pionowego gradientu zasolenia wody morskiej i wyjściowego składu zawiesiny. Materiał osadowy jest segregowany grawitacyjnie także w obrębie frakcji ilastej. Szczególnie silna flokulacja zawiesiny w przyczołowych basenach lodowców uchodzących do morza powoduje, że zasadnicza część materiału osadowego zostaje zatrzymana w strefie przyźródłowej i nie dostaje się do oceanu.

Przeprowadzono badania mineralogiczne i sedymentologiczne mułów zalegających dno fiordu Hornsund na Spitsbergenie (Fig. 1). Muły te są deponowane z zawiesiny wprowadzanej bezpośrednio do fiordu przez osiem głównych lodowców rejonu Hornsundu oraz z materiału wytapianego z lodu lodowcowego i lodu morskiego.

Badania mineralogiczne (Fig. 3—7; Tab. 1—5) pozwoliły na opisanie głównych minerałów i wydzielenie facji mineralnych w osadzie, oddzielnie dla minerałów nieilastych i dla minerałów ilastych.

Na podstawie zawartości minerałów nieilastych wyróżniono następujące cztery facje mineralne (Fig. 2, 10; Tab. 1, 2, 5): (1) siliciklastyczną, (2) bogatą w kwarc, skalenie i węglany, (3) węglanowo-kwarcową i (4) węglanowo-skaleniową. Ustalenie rozkładu tych facji w osadach dennych fiordu doprowadziło do wyznaczenia wyraźnie wyodrębniających się prowincji mineralnych. Prowincja siliciklastyczna obejmuje dna zatok Brepollen i Austre Burgerbukta (a poza Hornsundem — Storfjorden), prowincja kwarcowo-skaleniowo-węglanowa rozciąga się w zatoce Isbjörnhamna i wewnętrznej części fiordu (poza Hornsundem w Skoddebukta), a prowincja bogata w węglany (facje 3 i 4) w zatokach Samarinvågen, Vestre Burgerbukta, Gåshamna i zewnętrznym fiordzie.

Opierając się na minerałach ilastych (Tab. 1—3, 5) wyróżniono dwie facje mineralne: (1) muskowitowo-biotytową i (2) illitową (Fig. 10).

Facja muskowitowo-biotytowa charakteryzuje się dominacją pierwotnych i metamorficznych mik: muskowitu i biotytu (Tab. 1, 2), dużymi charakterystycznymi rozmiarami blaszek minerałów ilastych (wyrażonymi niskim stosunkiem zawartości frakcji $< 0{,}2$ μm do zawartości frakcji 0,2—2 μm (Tab. 4) oraz niskim stosunkiem zawartości mik dioktaedrycznych do zawartości mik trioktaedrycznych (Tab. 3).

Facja illitowa charakteryzuje się dominacją bardzo drobnoziarnistego illitu koncentrującego się we frakcji $< 0{,}2$ μm (Tab. 1, 2 i 4), niewielką ilością grubodetrytycznych mik (wyrażoną wysokim stosunkiem zawartości frakcji $< 0{,}2$ μm do zawartości frakcji 0,2—2 μm (Tab. 4) oraz wysokim stosunkiem zawartości mik dioktaedrycznych do zawartości mik trioktaedrycznych (Tab. 3).

Facja muskowitowo-biotytowa występuje w Isbjörnhamna, Gåshamna, Vestre Bugerbukta, wewnętrznej części Samarinvågen oraz poza Hornsundem w Skoddebukta. Facja illitowa występuje w Brepollen, zewnętrznej części Samarinvågen, Austre Burgerbukta i w centralnych nieckach fiordu, a poza Hornsundem — w Storfjorden (na wschód od południowego Spitsbergenu — Fig. 10).

Prowincje mineralogiczne ściśle związane są z litologią skał podłoża w obszarach alimentacyjnych lodowców (Fig. 2). Ten fakt oraz cechy zespołów minerałów ilastych sugerują, że minerały w osadach fiordu Hornsund są dziedziczone ze skał podłoża na lądzie w formie niemal nie zmienionej. Wietrzenie, procesy glebowe ani oddziaływanie środowiska morskiego nie zacierają podstawowych cech mineralogicznych skał macierzystych. Obecne warunki klimatyczne Spitsbergenu i transport lodowcowy działają zachowawczo w stosunku do minerałów ilastych w materiale osadowym.

Rozmieszczenie prowincji mineralnych określa zasięg dominującej sedymentacji z poszczególnych źródeł lodowcowych. Wyraźnie wyodrębnione prowincje mineralne zajmują bezpośrednie przedpola lodowców, a ich poziomy zasięg zależy od wydajności wypływów wód lodowcowych oraz od prędkości wypadania materiału osadowego z zawiesiny (Fig. 9). Obie te cechy są różne dla poszczególnych przyczołowych basenów sedymentacyjnych w Hornsundzie.

W badanych akwenach Hornsundu stężenia zawiesiny, a zatem i stopy sedymentacji, maleją wykładniczo z odległością od czół lodowców (Tab. 6), zgodnie z równaniem zaproponowanym przez Syvitskiego (*in*: Farrow *et al.*, 1983). Spadek stopy sedymentacji odbywa się z różną prędkością, określoną w równaniu Syvitskiego współczynnikiem *m*. Wartości tego współczynnika dla różnych akwenów Hornsundu zostały obliczone w tej pracy przez dopasowanie równania Syvitskiego do rzeczywistych prędkości sedymentacji uzyskanych z interpretacji zapisów ciągłego profilowania sejsmoakustycznego, głębokości penetracji próbnika rdzeniowego oraz odstępów (rocznych) między ciemnymi laminami. Zestawienie współczynników *m* dla akwenów Hornsundu podaje Tabela 6.

Różnice prędkości spadku stopy sedymentacji tłumaczy się w tej pracy działaniem dwóch czynników:

(1) różnicami w pionowym gradiencie zasolenia w proksymalnych częściach akwenów,

(2) różnicami w rozkładzie uziarnienia wyjściowej zawiesiny dostarczanej przez wody lodowcowe.

W Isbjörnhamna, gdzie współczynnik $m = -0{,}4$ (szybki spadek stopy sedymentacji), gradienty zasolenia są duże i dominują minerały ilaste o charakterystycznych dużych rozmiarach blaszek. W Brepollen, gdzie współczynnik $m = -0{,}1$ (powolny spadek stopy sedymentacji), gradienty zasolenia są małe i dominują najdrobniejsze minerały ilaste.

Z powyższych obserwacji autor wnioskuje, że sedymentacja zawiesiny z wód lodowcowych we fiordzie Hornsund jest wymuszona elektrochemicznie przez flokulację blaszek minerałów ilastych w wodzie morskiej. Intensywność flokulacji zależy prawdopodobnie od pionowych gradientów zasolenia w strefie przylodowcowej akwenów (Fig. 9). Flokulacja nie jest selektywna, to jest nie preferuje ona żadnego minerału ani charakterystycznej wielkości ziarn. W związku z tym zawiesina podlega jedynie segregacji grawitacyjnej i najdalej transportowane są drobne ziarna mineralne. Minerały ilaste obecne w zawiesinie mają różne charakterystyczne rozmiary ziarn: biotyt, muskowit i chloryt metamorficzny — duże, illit i chloryt diagenetyczny — małe. Grawitacyjna segregacja prowadzi zatem do wzbogacenia osadów dystalnych w illit.

Na podstawie opisu cech sedymentacyjnych i mineralnych wydzielono w drobnoziarnistych osadach fiordu Hornsund dwie litofacje: (1) muły laminowane i (2) homogeniczne muły zbioturbowane. Badania mineralogiczne oraz obliczenia tempa sedymentacji w różnych częściach badanego basenu sedymentacyjnego pozwoliły na określenie genezy i cech diagnostycznych wyróżnionych w Hornsundzie facji.

Dla facji mułów laminowanych są to: (1) obecność cienkich lamin organiczno-monosiarczkowych, (2) powszechna laminacja ziarnowa powstająca na skutek fluktuacji strumienia wód lodowcowych doprowadzających zawiesinę, (3) uboga infauna, (4) niemal zupełny brak szkieletów planktonu i grudek kałowych, (5) zespół minerałów ilastych o cechach identycznych z zespołem tych minerałów w najbardziej proksymalnych mułach danego akwenu, (6) stopa akumulacji ponad 5 cm/rok, (7) płaskie wypełnianie zagłębień dna basenu sedymentacyjnego.

Dla facji zbioturbowanych mułów homogenicznych określono następujące cechy diagnostyczne: (1) rzadko rozmieszczone ciemne plamy monosiarczkowe i sporadycznie laminy, (2) niewyraźna niejednorodność teksturalna, miejscami gruboziarniste wkładki, (3) silna bioturbacja związana z obecnością licznej infauny i epifauny, (4) sporadycznie spotykane szkielety planktonu i grudki kałowe, (5) mieszany charakter zespołu minerałów ilastych nie wykazujący silnego związku ze składem zespołu tych minerałów w bardziej proksymalnych mułach, (6) stopa akumulacji poniżej 5 cm/rok, (7) oblekające pokrywanie topografii dna.

Ciemna laminacja mułów związana jest z występowaniem monosiarczków żelaza. W niniejszej pracy wykazano, że ciemnym laminom towarzyszy obecność autigenicznego syderytu (Fig. 3, 5; Pl. I, II). W przeciwieństwie do nietrwałych monosiarczków, syderyt pozostanie w osadzie nawet po zmianie środowiska na utleniające (co następuje w miarę pogrzebania) i może być zatem użyty jako świadectwo istnienia pierwotnej laminacji pochodzącej z nagromadzenia substancji organicznej.

Silna prosta zależność stopy sedymentacji od pionowego gradientu zasolenia na przedpolu lodowców uchodzących do morza jest głównym czynnikiem decydującym o efektywności osiadania zawiesiny. Ten efekt powoduje, że lodowce uchodzące do morza, których akweny przyczołowe charakteryzują się z zasady bardzo dużymi pionowymi gradientami zasolenia, dostarczają znikomych ilości osadu na szelf. Ta cecha różni zasadniczo lodowcowo-morskie źródła osadu od źródeł rzecznych, zasilanych przez lodowce kończące się na lądzie. Wydaje się, że intensywne wyłapywanie zawiesiny w pobliżu źródła, co jest charakterystyczną cechą lodowców uchodzących do morza, jest główną przyczyną znikomego rozprzestrzenienia osadów pochodzących z wód lodowcowych w basenach mórz arktycznych i na szelfie antarktycznym.

EXPLANATIONS OF PLATES

Plate I

SEM micrographs of bulk air-dried sediment samples from Hornsund

1 — Sample 069, Isbjörnhamna, from topmost layer (2 cm thick) of black organic mud. Rhombohedral grain of (alkali?) feldspar slightly sericitized (A). In lower left and upper left tabular grains of plagioclases (?). Euhedral microcrystals of siderite in the upper right corner (S)

2 — Sample 069, same specimen. Large clay flakes cleaved by weathering (or disaggregated) embedded in fine, aggregated clays (C). In the upper right corner euhedral crystal of siderite (or pyrite) — S, and unidentified spherical particles — D

3 — Sample 069, same specimen. Angular quartz grains (Q) within weakly but predominantly face-to-face aggregated clay matrix. Authigenic siderite (S) and secondary gypsum crystals (X) crystallized due to air-drying of the sample

4 — Sample 072, outer slope of Isbjörnhamna sill. Irregular to pseudohexagonal clay flakes (C) often edge-to-face aggregated. Tabular and rhombohedral (A) weathered feldspar grains. Unidentified spherical particle (D) at the bottom of the picture.

5 — Sample H_7, Brepollen. Compact aggregation of fine clay flakes tightly adhering to coarser grains. Compare this micrograph with the next one in the same scale

6 — Sample 800, outer central Hornsund fjord. Loose texture of larger (compared with the above sample H_7) clay flakes which do not adhere to the faces of angular coarse grains

Plate II

SEM micrographs of organogenic components and associated authigenic minerals in the sediments of Hornsund (1—4) and eastern coast of south Spitsbergen (5 and 6).

1 — Sample 819, Adriabukta — inner Hornsund fjord, sediment of low (< 1 cm/a) accumulation rate. Pelletized clay mineral aggregation (P) in the central part of the micrograph. Authigenic siderite crystal (S)

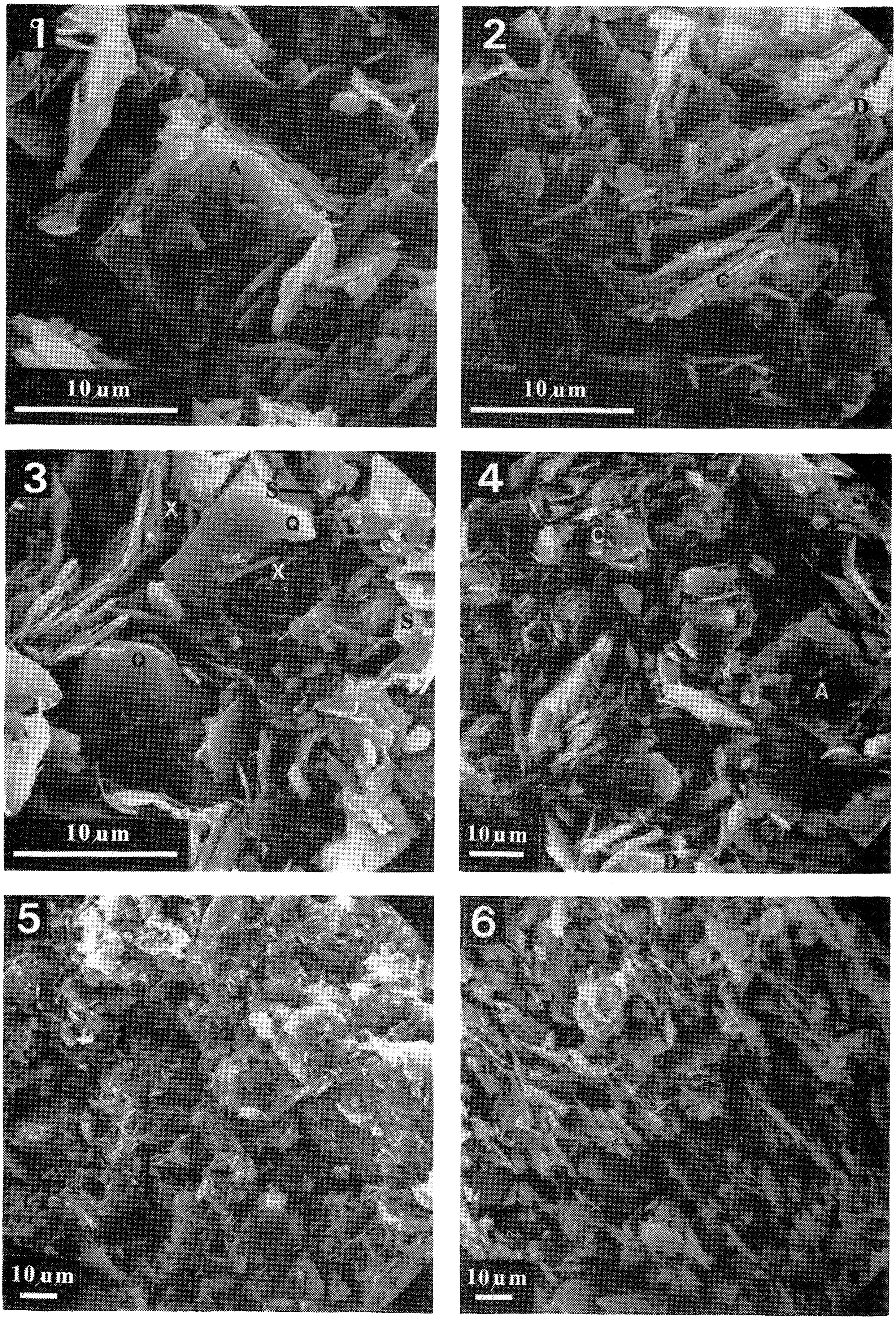
2 — Sample 042, outer Burgerbukta, sediment of low (< 1 cm/a) accumulation rate. Fecal pellet (P) in the central part of the micrograph and authigenic siderite or pyrite crystal (S; FeS_2?)

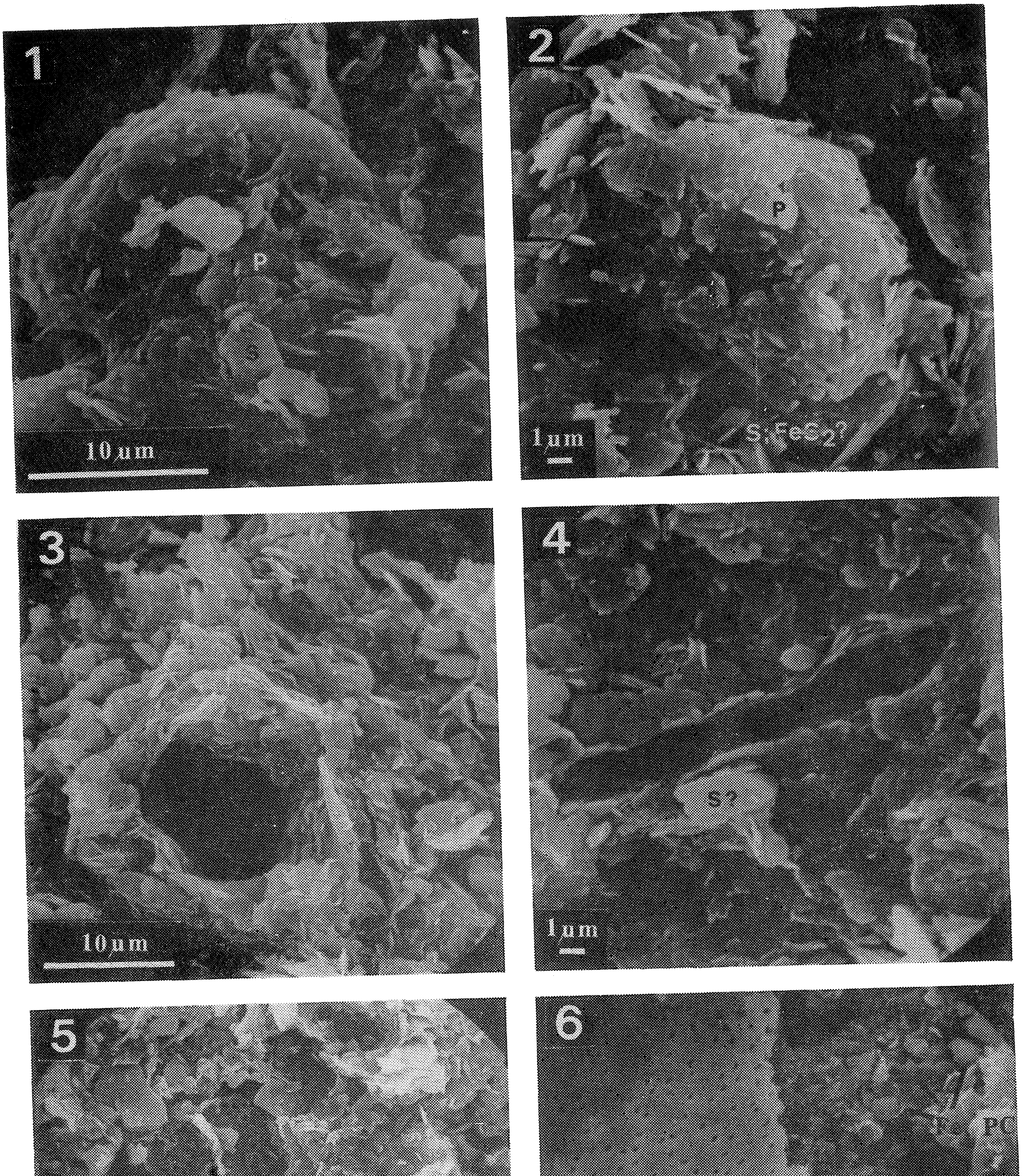
3 — Sample 819, same specimen as in 1. Agglutinated test (?) embedded in clay matrix

4 — Sample 819, same specimen. Strongly aggregated clays with microfaunal cast and associated siderite crystal (S?)

5 — Sample KAG 3, Hambergbukta, eastern coast of south Spitsbergen. Strongly aggregated (pelletized?) fine clays and a fragment of diatom test (no diatoms were observed in Hornsund sediments)

6 — Sample KAG 5, Storfjorden, eastern coast of south Spitsbergen. A fragment of diatom test with concentration of authigenic pyrite crystals (Fe) within pelletized clays (PC)


1
A
S
10 µm
2
D
S
C
10 µm
3
X
S
Q
X
S
Q
10 µm
4
C
A
D
10 µm
5
10 µm
6
10 µm


1
P
S
10 µm
2
P
1 µm
S; FeS$_2$?
3
10 µm
4
S?
1 µm
5
10 µm
6
Fe
PC
PC
10 µm