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STRATIGRAPHY AND SEDIMENTATION OF THE OTTO FIORD FORMATION — a major Mississippian-Pennsylvanian evaporite of subaqueous origin in the Canadian Arctic Archipelago

W. W. Nassichuk and Graham R. Davies

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Preface

This report contributes to an important function of the Geological Survey, namely the assessment of abundance and distribution of our hydrocarbon resources. As geological models become increasingly refined through interpretation and synthesis of new data, so do statistical assessments become increasingly realistic. Through several decades of field exploration, a rather sophisticated model has evolved of the stratigraphic framework of the Sverdrup Basin, a major geological province within the Canadian Arctic Archipelago. The basin is an elongated depression parallel to the northern margin of the stable part of the North American continent and contains up to 13 000 m of strata ranging from Mississippian to Eocene. Significant hydrocarbon reserves are known to be contained within it.

This report presents data from the Otto Fiord Formation, a unit which includes Mississippian and Pennsylvanian evaporites, limestones, shales and sandstones, and thus provides new information on the early paleogeography and tectonosedimentary history of the Sverdrup Basin. The Otto Fiord rocks were probably deposited in a subaqueous hypersaline marine environment. The formation is richly fossiliferous and, on the basis of data from several fossil groups, regional and intercontinental biostratigraphic correlations have been more clearly defined.

D.J. McLaren
Director General
Geological Survey of Canada

Ottawa, April 1978

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Abstract

The Otto Fiord Formation, composed of anhydrite, limestone, shale and sandstone of Late Mississippian to Middle Pennsylvanian age, occurs widely in the interior of the Sverdrup Basin, where its distribution appears to have been influenced by previous or contemporaneous tectonic events. It is overlain by Middle Pennsylvanian to Lower Permian limestone, siltstone and shale of the Hare Fiord Formation, and the two formations combined define the Basinal Clastic and Evaporitic Belt. The Otto Fiord Formation attains a thickness of 420 m in northern Ellesmere Island but is considerably thinner in eastern Axel Heiberg Island. In both of these areas it is exposed in normal stratigraphic successions. Elsewhere in Axel Heiberg Island, and in Ellef Ringnes, Amund Ringnes and Melville islands, it outcrops only where it is exposed in more than 100 evaporitic diapirs. The diapirs have been known for more than two decades but the presence of an active halite core beneath a bedded sulphate cap has been known only since 1972 when an exploratory well, Hoodoo L-41, was drilled southeast of Hoodoo Dome proper on Ellef Ringnes Island.

In this report stratigraphic relationships and depositional environments of the Otto Fiord are described, and interpretations of the early paleogeography and tectono-sedimentary history of the Sverdrup Basin are presented. The Otto Fiord probably was deposited in a subaqueous, hypersaline marine environment. This interpretation comes at a time when the field of evaporite sedimentology is dominated by examples of alleged paleo-sabkhas.

In northern Ellesmere Island the Otto Fiord Formation is abundantly fossiliferous and on the basis of data from ammonoids and foraminifers, including fusulinaceans, rather refined biostratigraphic correlations have been made between circum-Arctic regions and standard Carboniferous successions in Europe and midcontinental North America. Rather marked diachroneities at the top of the Otto Fiord Formation, over relatively short distances in the Hare Fiord area of northern Ellesmere Island, coincide with the development of extensive carbonate mounds at the base of the Hare Fiord Formation. Evaporite deposition may have continued behind structural or carbonate barriers in some areas after normal marine conditions became established in others.

Résumé

La formation d'Otto Fiord composée d'anhydrite, de calcaire, de schiste argileux, et de grès, et se située entre le Mississippien supérieur et le Pennsylvanien moyen, occupe une grande étendue du bassin de Sverdrup; il semble que des phases tectoniques contemporaines ou anciennes aient influencé sa répartition. Les calcaires, siltstones et schistes argileux de la formation de Hare Fiord dont l'âge s'échelonne entre le Pennsylvanien moyen et le Permien inférieur la recouvrent, et les deux formations combinées constituent la zone clastique d'évaporites du bassin. La formation d'Otto Fiord atteint une épaisseur de 420 m dans le nord de l'île Ellesmere mais elle s'amincit considérablement dans l'est de l'île Axel Heiberg. Dans ces deux zones, ses affleurements indiquent une succession stratigraphique normale. Mais ailleurs dans l'île Axel Heiberg, et dans les îles Ellef Ringnes, Amund Ringnes et Melville, elle n'apparaît que sous forme de diapirs, plus d'une centaine au total, formés par les évaporites. On a découvert les diapirs il y a plus de vingt ans, mais on ne connaît la présence d'un noyau actif de halite surmonté d'une calotte de roches sulfatées et stratifiées que depuis 1972, c'est-à-dire l'époque à laquelle on a foré un puits d'exploration, Hoodoo L-41, au sud-est du dôme de Hoodoo qui est situé sur l'île Ellef Ringnes.

Dans le présent rapport, on décrit les relations stratigraphiques et les milieux de sédimentation du fiord Otto, et on donne certaines interprétations relatives à la paléogéographie et à l'histoire tectonique et sédimentaire ancienne du bassin de Sverdrup. On pense que la formation d'Otto Fiord s'est constituée dans un environnement sous-marin hypersalin. Nous émettons cette hypothèse, à un moment où la sédimentologie des roches évaporitiques comporte un très grand nombre d'exemples de paléo-sabkhas.

Dans le nord de l'île Ellesmere, la formation d'Otto Fiord est très fossilifère, et les ammonoides et les foraminifères, y compris les fusulinacées, ont permis d'établir des corrélations biostratigraphiques très précises entre la région circumpolaire, et les successions classique du Carbonifère de l'Europe et de la partie centrale du continent nord-américain. Au sommet de la formation d'Otto Fiord, il existe des séries sédimentaires transgressives, qui se prolongent sur des distances relativement faibles dans la zone du fiord Hare dans le nord de l'île Ellesmere, et où l'on rencontre d'importants dômes de carbonates à la base de la formation de Hare Fiord. Il est possible que les évaporites aient continué à se déposer dans certaines zones derrière des barrières structurales ou de carbonates, après qu'un milieu marin normal caractérisait d'autres zones.

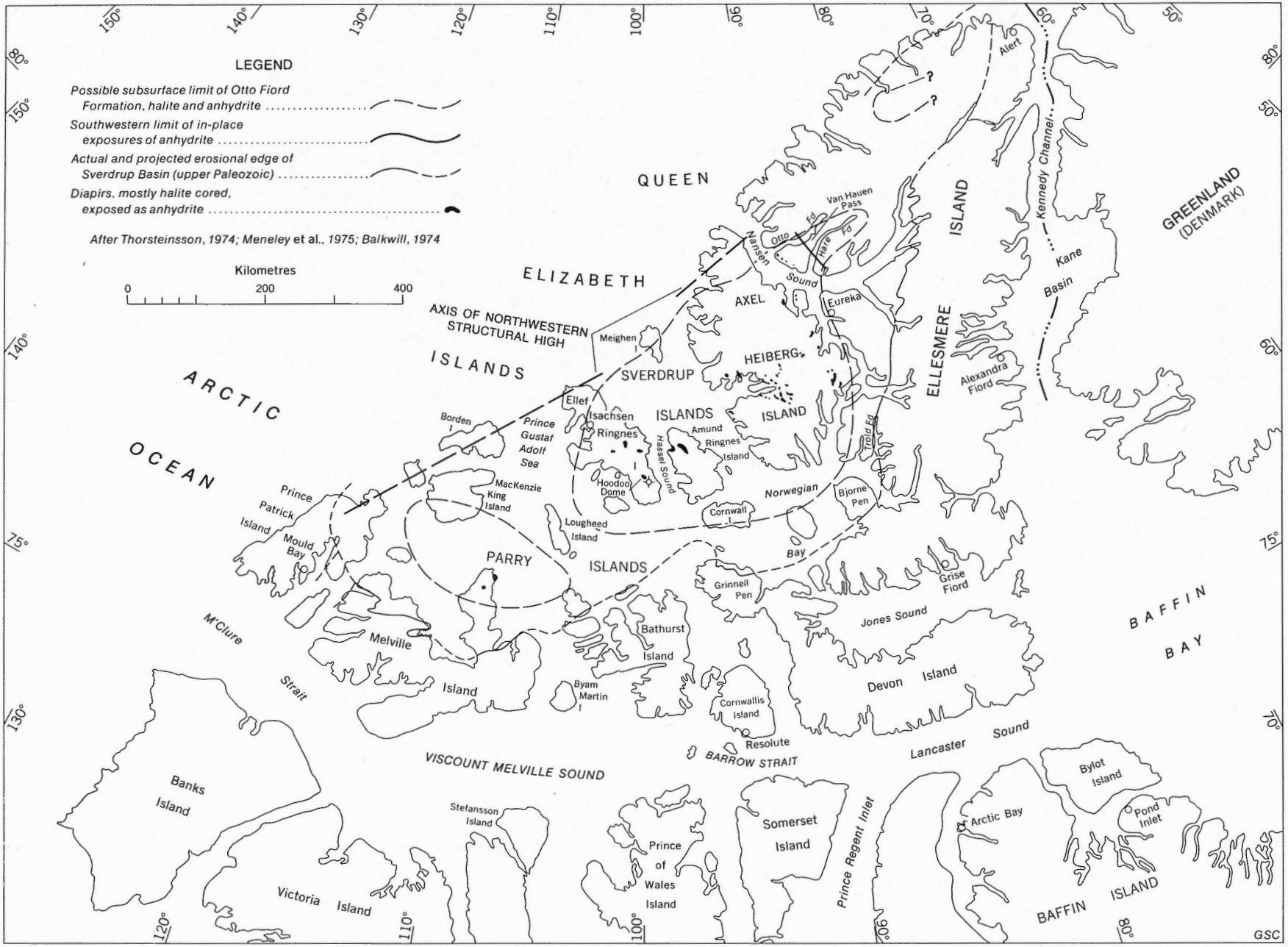


Figure 1. Index map of the northern Canadian Arctic Archipelago showing projected distribution of Carboniferous evaporite facies (Otto Fiord Formation and equivalents) in the Sverdrup Basin (modified from Meneley et al., 1975), and the distribution of Otto Fiord diapirs.

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INTRODUCTION

The Otto Fiord Formation was defined by Thorsteinsson (1974) for a sequence of interbedded anhydrite, limestone and shale in interior regions of the Sverdrup Basin. In an earlier summary of upper Paleozoic stratigraphy in the Sverdrup Basin, Thorsteinsson (1970) used the formational name informally. The type section of the Otto Fiord Formation is near van Hauen Pass on northern Ellesmere Island and is shown on Geological Survey of Canada Map 1309A (Otto Fiord) by Thorsteinsson and Trettin (1972).

This report records the results of field work begun by Nassichuk in 1969 and continued by both of us in the 1971, 1972 and 1974 field seasons. Summary papers resulting from this field work and dealing with faunas, stratigraphy and sedimentology of the Otto Fiord Formation have been published (Nassichuk, 1975; Davies and Nassichuk, 1975). In addition, specialized papers dealing with evaporite chemistry and genesis have been completed (Davies and Krouse, 1975; Davies, 1977c). In the present study Nassichuk is responsible principally for regional stratigraphy and biostratigraphy and Davies for sedimentology and environmental interpretations.

Professor N.C. Wardlaw of the University of Calgary and his student D.L. Christie spent several weeks in 1973 studying the Otto Fiord Formation at Hare Fiord, Ellesmere Island. Their documentations, which include an unpublished M.Sc. thesis by Christie (1975) and a sedimentological report by Wardlaw and Christie (1975), complement our research in the type area for the Otto Fiord Formation.

Deposition of the Otto Fiord Formation represents an important sequence of events in the sedimentological and structural history of the Sverdrup Basin. In addition to providing insight into the early paleogeography and tectonic-sedimentary history of the basin, the formation also affected later structural events involving evaporite diapirism and detachment. The purpose of this paper is to document regional stratigraphic and biochronological relationships of the Otto Fiord Formation and to interpret depositional environments through detailed sedimentological studies. Of even broader significance, the report also deals with evaporites per se. In this respect, at a time when the field of evaporite sedimentology is dominated by examples of paleo-sabkhas, the Otto Fiord evaporites are some of a few considered to be of marine subaqueous origin.

We have examined the Otto Fiord Formation in considerable detail in areas where it is preserved best in bedded sequence; that is, in the vicinity of Hare Fiord on northern Ellesmere Island, and also nearby in the Blue Mountains and the Krieger Mountains of northern Ellesmere Island. In addition, some data are available from Axel Heiberg, Ellef Ringnes, Amund Ringnes and Melville islands. Most of our data from the last four islands have been derived from surface exposures of evaporitic piercement structures and from drill core.

We are indebted to many scientific associates for important data used in this report. B.L. Mamet, University of Montreal, identified numerous collections of foraminifers and algae and J.L. Carter of the Carnegie Institute, Pittsburgh, identified brachiopods. K.P. Bender of Phillips-Universitat, Marburg, identified conodonts from the type area of the Otto Fiord Formation at van Hauen Pass, Ellesmere Island, and C.A. Ross of Western Washington State College identified fusulinaceans from the Blue Mountains, Ellesmere Island. N.C. Wardlaw of the University of Calgary provided sedimentological field information which complemented our data from the type area of the Otto Fiord Formation. Also, J.L. Hardie of Johns Hopkins University and B.C. Schreiber of Lamont Doherty Geological Observatory, provided two illustrations (Figs. 45, 46) used in this report.

REGIONAL STRATIGRAPHIC FRAMEWORK: THE SVERDRUP BASIN

The Sverdrup Basin was defined by Fortier (1957) as a regional depression occupying a large part of the northwestern Queen Elizabeth Islands (Fig. 1) and containing an essentially conformable sequence of upper Paleozoic, Mesozoic and Tertiary strata. Earlier, the term 'Sverdrup Basin' was employed without discussion in the legend and map of a paper by Fortier and Morley (1956). Fortier's (1957) general discussion was followed by descriptive accounts by Thorsteinsson and Tozer (1960, 1970), Fortier et al. (1963), Douglas et al. (1963), Tozer and Thorsteinsson (1964), Plauchut (1971), Trettin et al. (1972), Nassichuk (1972, 1975) and Thorsteinsson (1974). In summary, the Sverdrup Basin is an elongate, northeasterly trending pericratonic depression containing a maximum thickness of 13 000 m of Lower Carboniferous (Viséan) to lower Tertiary (Eocene) rocks, which rest unconformably on Devonian to Precambrian rocks of the Franklinian Geosyncline. The latter rocks were deformed by the Ellesmerian Orogeny between Late Devonian (Frasnian) and Early Carboniferous (Tournaisian) time. Along with the Franklinian Geosyncline the Sverdrup Basin constitutes the Innuitian orogenic system of Fortier et al. (1954) embracing the Canadian Arctic Islands and northern Greenland.

Recently completed regional mapping and structural studies in the Sverdrup Basin by Thorsteinsson (1974, p. 3), Trettin et al. (1972) and Balkwill (1974), as well as seismic and drilling programs completed by Panarctic Oils Ltd. and other commercial groups, have provided sufficient data to interpret the early depositional history of the basin. According to Balkwill (ibid.), the geometry and the depositional fabric of the basin suggest that it may have evolved through episodes of crustal fracturing, which resulted in periodic subsidence and filling accompanied by volcanic extrusion and intrusion. Trettin et al. (1972) indicated that the basic tectonic framework of the generally unstable Sverdrup Basin was inherited from early Paleozoic tectonic patterns and that the configuration of the basin varied periodically throughout its history. According to Trettin et al. (ibid.), the axial trough of the Sverdrup Basin coincides approximately with the position of the lower Paleozoic Hazen Trough in the underlying Franklinian Geosyncline and a structural high bordering the axial trough on the northwest side of the basin coincides with a lower Paleozoic structural high. This northwestern structural high is illustrated clearly in a cross-section of Ellef Ringnes Island (Meneley et al., 1975, Fig. 7; Fig. 2, this report). Similarly, other tectonic elements, which may represent rejuvenation of earlier Paleozoic structures and which affect sedimentary patterns within the basin, have been identified.

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The Cornwall Arch, a positive feature in Late Cretaceous time, trends north-northwesterly across the axis of the basin and underlies Cornwall and Amund Ringnes islands. This arch parallels a late Paleozoic carbonate facies belt stated by Meneley et al. (ibid.) to extend across the basin axis in the vicinity of Lougheed and Mackenzie King islands, dividing basin-centre evaporites of the Carboniferous Otto Fiord Formation into two or more depocentres. This carbonate belt, or 'central basin platform', probably was localized by a 'basement' ridge; the belt also parallels a major fracture system (Plauchut, 1971, Fig. 1). The trend of these crosscutting arches or ridges is generally parallel with pre-Sverdrup Basin tectonic lineaments such as the Cornwallis Fold Belt exposed on Cornwallis, Devon and Bathurst islands, the Princess Margaret Arch (Trettin et al., 1972) on Axel Heiberg Island, and possibly the Bache Peninsula Arch (Kerr, 1968) on east-central Ellesmere Island.

The Sverdrup Basin, which now is a structural synclinorium, was subjected to two episodes of deformation. Between Late Pennsylvanian and Early Permian time, the Melvillian Disturbance initiated localized folding and faulting in marginal regions of the basin; folding has been recognized in the southwestern part of the basin in the vicinity of northwestern Melville Island (Tozer and Thorsteinsson, 1964) and faulting on northern Ellef Ringnes Island (Meneley et al., 1975). Structural deformation also is apparent between Pennsylvanian and Permian strata near the eastern margin of the Sverdrup Basin at Trold Fiord (Thorsteinsson, 1970) and at Bjorne Peninsula (N.E. Haimila, pers. com., 1976). Most of the major structural characteristics presently observed within the basin, such as broad asymmetrical folds, abundant normal faults and some reverse faults, were imprinted during the Eurekan Orogeny, an event dominated by widespread vertical movements between Late Cretaceous (Senonian) and middle Tertiary (Miocene) time. On Axel Heiberg and Ellesmere islands, where major upper Paleozoic and Mesozoic depocentres were situated, deformation was particularly intense. Elsewhere it was relatively mild, especially in western parts of the basin. In some central parts of the basin, detachment occurred above Carboniferous evaporites; locally intrusion of evaporites into the cores of folds and along faults affected structural style. However, intrusion of large halite-cored diapirs in the central and southern axes of the basin (Fig. 2) probably was not triggered by the Eurekan Orogeny but may have been activated by normal sediment loading as early as Jurassic time. Whereas the Eurekan Orogeny generally is considered to mark the climax of widespread marine deposition within the basin, it is particularly clear that the basin ceased to exist as a separate depositional entity even earlier, in the Early Cretaceous, when transgressions progressed far beyond the basin boundaries that existed in pre-Cretaceous time.

Finally in this discussion of the Sverdrup Basin as an assumed discrete depositional entity, even though structural and stratigraphic information is abundant little is known of its relationships in the context of circumpolar tectonics and deposition. It is evident that upper Paleozoic and lower Mesozoic strata in the Sverdrup Basin, in northeastern Greenland (Wandell Sea Basin) and in some of the islands of Svalbard and Novaya Zemlya are remarkably similar. The last two groups of islands are on the Barents Shelf; Svalbard is some 800 km northeast of northernmost Ellesmere Island and Novaya Zemlya is an equal distance beyond Svalbard. Clearly, all regions were affected by comparable tectonism, and deposition occurred under similar environmental conditions during much of the late Paleozoic. It is less clear, however, whether deposition took place in a series of isolated successor basins as implied by Churkin (1969), or whether the present isolation reflects transection of a large common basin through differential movements of tectonic plates during Tertiary time. Some authors have postulated a post-Permian northward movement of Greenland relative to Ellesmere Island (in Harland, 1966); it is equally plausible that Svalbard migrated east along the Nansen Fracture Zone away from the Arctic Mid-Oceanic Ridge.

Outcrop distribution shows that the Sverdrup Basin extended at least 1000 km from northeastern Prince Patrick Island to northeastern Ellesmere Island. The maximum width of the basin, from southwestern Ellesmere Island to northwestern Axel Heiberg Island, is about 290 km. These must be considered minimum dimensions since recently recovered subsurface data from the westernmost Queen Elizabeth Islands indicate that deposition extended far beyond present outcrop limits. The present eastern and southern erosional margins of the Sverdrup Basin are bounded by deformed Devonian and older Paleozoic rocks that are, in turn, bounded to the east and southeast by northern extensions of the Canadian Shield. A prominent re-entrant along the southern margin of the basin between eastern Melville Island and Grinnell Peninsula, Devon Island, may reflect lower Paleozoic structural highs in the Parry Islands and Cornwallis fold belts. The latter fold belt, considered by Thorsteinsson and Tozer (1960) to be part of the Boothia Uplift, was periodically rejuvenated and was subjected to at least five episodes of deformation between Early Devonian and Tertiary time. The Parry Islands Fold Belt extends in a westerly direction from the Cornwallis Fold Belt and terminates at the southwest corner of the Sverdrup Basin near Eglinton Island. The Eglinton Island area was depicted by Thorsteinsson and Tozer (ibid.) as a graben, related to the Prince Patrick Uplift. According to them (1970), the Prince Patrick Uplift was relatively positive in late Paleozoic and Mesozoic time and may have marked the southwestern limit of the basin.

The northwestern margin of the basin is less clearly defined since much of it is concealed by Tertiary and younger sediments of the Arctic Coastal Plain. Erosional edges, which may parallel depositional margins, are delineated in northwesternmost Ellesmere Island and northwestern Axel Heiberg Island by exposures of Devonian and older Paleozoic sedimentary, igneous and metamorphic rocks (Trettin et al., 1972), against which abut younger Paleozoic rocks of the Sverdrup Basin succession. Thorsteinsson and Tozer (1960, 1970) and Tozer and Thorsteinsson (1964) recognized nearshore characteristics in Mesozoic rocks in the western Queen Elizabeth Islands and suggested that a periodically positive ridge separated the Sverdrup Basin from the Arctic Ocean during late Paleozoic and Mesozoic time. Gregory et al. (1960) suggested on the basis of magnetic and radiometric profiles that an arch, possibly originating in the Franklinian Geosyncline, parallels the edge of the basin in this area. Meneley et al. (1975) showed from data assembled from seismic profiles and drillholes that the basin margin from Prince Patrick Island to Axel Heiberg Island was influenced by a tectonically active fault-bounded ridge (Fig. 2). The trend of this ridge parallels the trend of the mid-Paleozoic Grantland Uplift on Ellesmere Island, and of the outcrop belt of lower Paleozoic metamorphic and other rocks along the northwestern coast of Ellesmere Island (Trettin et al., 1972, Fig. 11). It is important to note that although the intermittently active northwestern marginal ridge may have formed a bathymetric or even topographic (subaerial) barrier at various times during Sverdrup Basin deposition, sedimentation was probably contemporaneous on both sides of the ridge. The ridge undoubtedly was a positive element in Pennsylvanian to Early Permian time (Fig. 2), influencing and restricting marine circulation within the basin and the consequent deposition of Carboniferous evaporites.

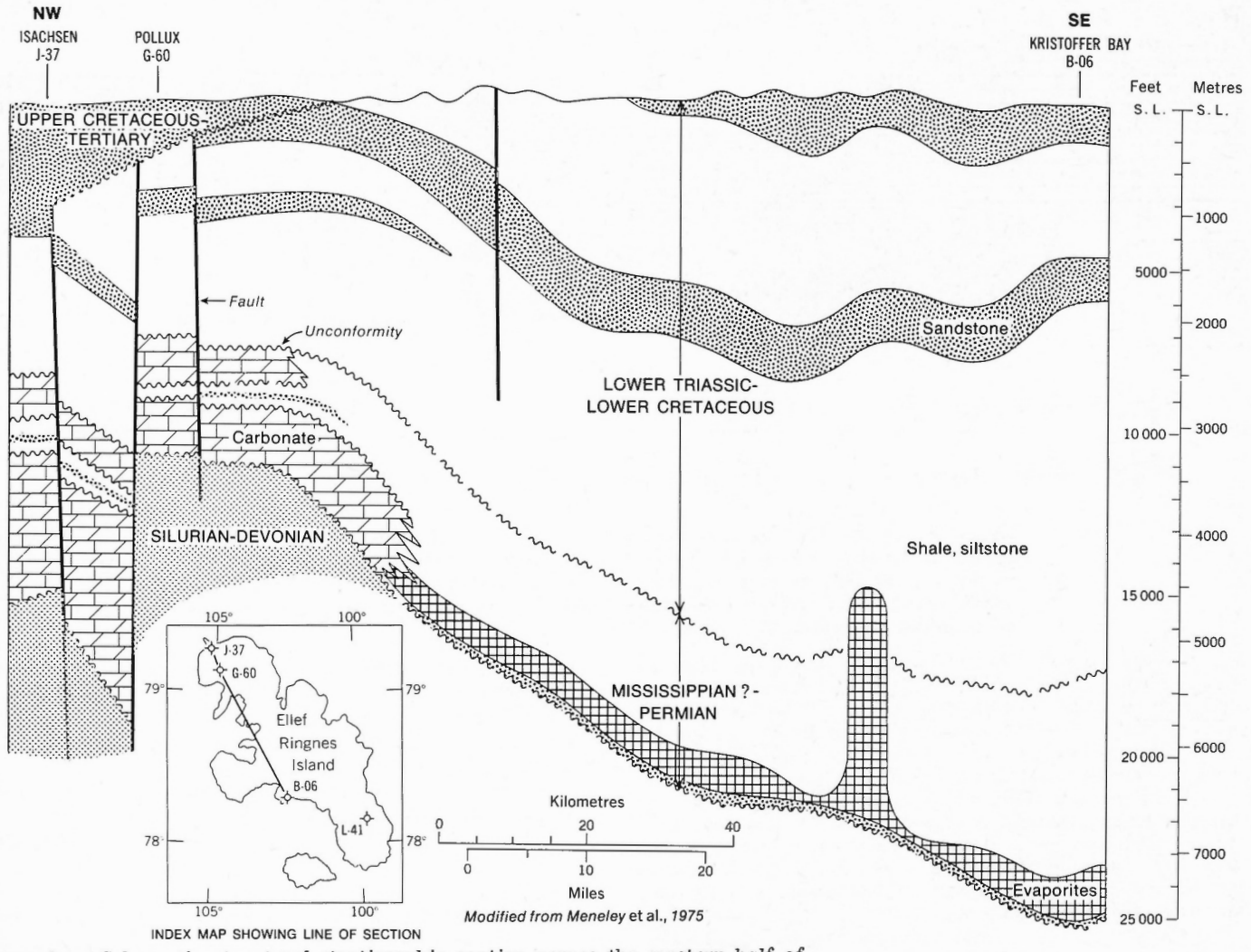


Figure 2. Schematic structural-stratigraphic section across the western half of the Sverdrup Basin at Ellef Ringnes Island (modified from Meneley et al., 1975). The section illustrates the distribution of the Otto Fiord halite facies in the subsurface with local diapirism, and the presence of coeval carbonates over a structural high forming the northwestern rim of the basin.

Reconstruction of both northwestern and northeastern basin margins on northern Ellesmere Island is made difficult by discontinuous outcrops (outliers) of late Paleozoic to Cenozoic age. An extensive linear sequence of upper Paleozoic to Cenozoic strata on the northeastern side of the basin parallels the axis of the basin at Lake Hazen. These strata are separated from the basin proper by the Lake Hazen Fault Zone (Christie, 1964), in which Paleozoic rocks occur at the surface. Similarly, it is not known whether Carboniferous and Permian outliers on the northwestern side of the basin along the north coast of Ellesmere Island are allochthonous or autochthonous. They may have been isolated from the Sverdrup Basin by Tertiary tectonism or, alternatively, they may represent deposition in satellite depressions separated from the Sverdrup Basin (Trettin et al., 1972). Another alternative is that the outliers were once continuous with the Sverdrup Basin and, prior to Tertiary plate movements, with sequences now preserved in Svalbard and parts of Greenland.

The Carboniferous and Permian succession

During Carboniferous and Permian time the Sverdrup Basin depression received less than 20 per cent of its total sedimentary complement; the other 80 per cent was laid down in the Mesozoic. Sediments were derived mainly from sources to the south and east, and major depocentres were situated near the eastern and western margins of the basin. On western Ellesmere Island, near the eastern margin of the basin, some 3000 m of strata occur at the surface and, on Brock Island, near the western margin, more than 3600 m of strata have been identified in the subsurface. In interior regions of the basin, where structural deformation by the Tertiary Eurekan Orogeny is more intense than at the basin edge, between 1200 and 2400 m of strata are preserved. In the interior of the basin, formations are conformable on a regional scale and only local interformational angularity is apparent. Along the margin of the basin, strata usually retain a gentle dip toward the basin interior and many formations are separated by disconformities or mild angular unconformities that reflect overstepping. At several places near the basin margin, angular unconformities reflect important episodes of disturbance. On northwestern Melville Island and west-central Ellesmere Island, for instance, the Carboniferous Canyon Fiord Formation, uplifted and folded by the Melvillian Disturbance in Late Carboniferous or Early Permian time, is overlain unconformably by the Upper Permian Trold Fiord Formation.

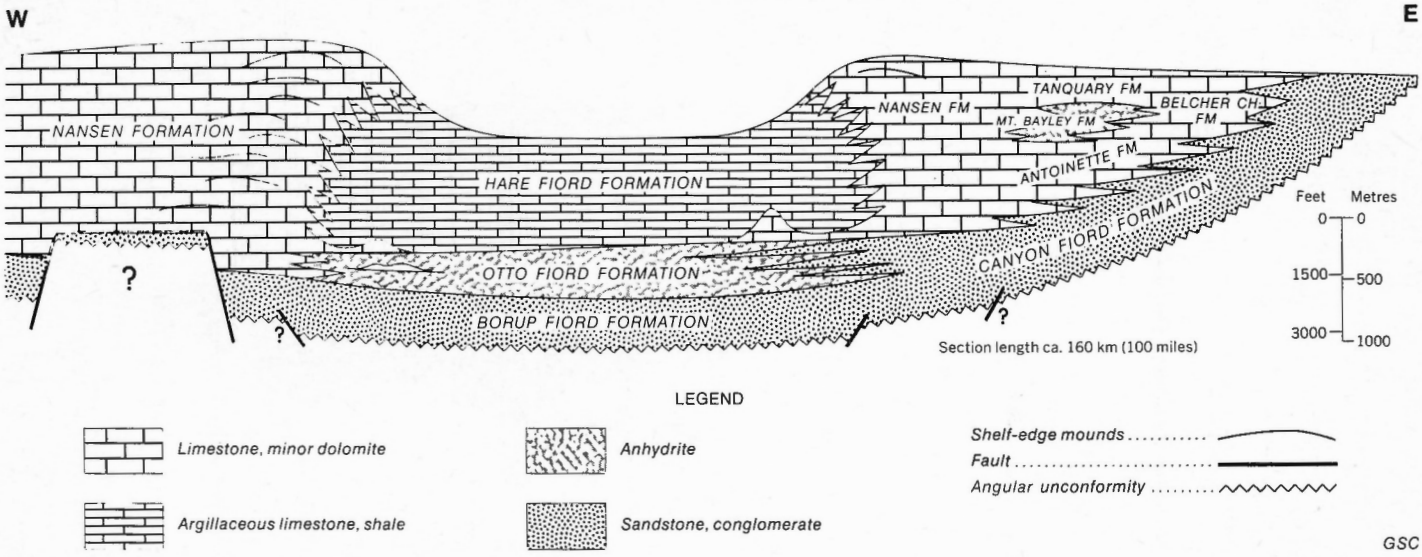


Figure 3. Schematic restored section for Upper Mississippian to Lower Permian lithofacies across the northeasterly trending axis of the Sverdrup Basin in the vicinity of Hare Fiord on northern Ellesmere Island (modified after Thorsteinsson, 1974, Fig. 2). Structural relationships at western end of section are hypothetical, and represent a projection from subsurface data (Fig. 2).

In this report a brief résumé of the Carboniferous-Permian stratigraphic framework in the Sverdrup Basin is followed by more detailed descriptions of the Emma Fiord and Borup Fiord formations, which underlie the evaporitic Otto Fiord Formation, and of the Hare Fiord Formation, which overlies the Otto Fiord Formation.

Thorsteinsson (1970, 1974), who is responsible for establishing a generally accepted model for the deposition of Carboniferous and Permian rocks over much of the basin, recently has completed maps and a comprehensive monograph on the Carboniferous and Permian stratigraphy of Axel Heiberg Island and much of Ellesmere Island (Thorsteinsson, 1974). Thorsteinsson's general stratigraphic scheme is accepted in this report, with minor modifications because of additional data from seismic and drilling activities recently completed within the basin, and from field research carried out by the writers. A schematic illustration of Carboniferous to Lower Permian facies relationships in the northern part of the basin is shown in Figure 3. Detailed biostratigraphic correlations of Carboniferous formations are shown in Figures 4 and 5.

Following initial deposition of the nonmarine Emma Fiord Formation in marginal depressions in the Sverdrup Basin during Viséan time, a major marine transgression began either in late Viséan or early Namurian time and spread throughout the basin. Deposition during this transgression is characterized by reddish weathering conglomerate and sandstone and minor amounts of limestone, which in the interior of the basin are designated the Borup Fiord Formation and at the edge of the basin the Canyon Fiord Formation. Beginning with the deposition of these red-bed units, marine deposition prevailed through the remainder of Carboniferous and Late Permian time. In fact, nonmarine deposition is clearly demonstrable only in part of the Lower Permian Sabine Bay Formation, a partly deltaic, partly marine sandstone, conglomerate, coaly unit that is discontinuous along the southern and eastern margins of the basin.

Thorsteinsson (ibid.) realized that uppermost Mississippian to Lower Permian (Namurian-Artinskian) rocks in the Sverdrup Basin show marked lithological differences between the basin edge and the basin interior, and identified a series of facies belts parallel with the long axis of the basin. Westward from the east edge of the basin, the facies belts and formations contained within them are as follows (see Fig. 3):

(a) Marginal Clastic Belt. Between Cañon Fiord and Bay Fiord on western Ellesmere Island, the Canyon Fiord Formation of Middle Pennsylvanian to Early Permian age is composed entirely of sandstone and conglomerate and forms a narrow, conspicuous belt adjacent to the basin margin.

(b) Marginal Clastic and Carbonate Belt. Basinward from the Marginal Clastic Belt and elsewhere along the eastern and southern margins of the basin, the Canyon Fiord Formation contains limestone beds as well as sandstone and conglomerate beds and forms a couplet with the overlying Lower Permian Belcher Channel Formation (limestone). Within the belt the Canyon Fiord Formation is of Middle Pennsylvanian age. In some parts of the belt, between Cañon Fiord northward to Tanquary Fiord, the Canyon Fiord Formation is overlain successively by the Upper Pennsylvanian Antoinette Formation (limestone), the Lower Permian (Asselian) Mount Bayley Formation (evaporites) and the Lower Permian (Sakmarian) Tanquary Formation (limestone). The last three formations are all lateral equivalents of the Belcher Channel Formation.

(c) Southeastern Carbonate Belt. According to Thorsteinsson (ibid.), the Southeastern Carbonate Belt is represented by the Nansen Formation and is presumed to extend from northern Bjorne Peninsula northward at least to Borup Fiord on the north side of Greely Fiord, where it merges with the Northwestern Carbonate Belt. Upper Paleozoic rocks do not outcrop over much of this belt, and more conclusive evidence concerning the distribution of the belt depends on obtaining additional data from drillholes. Nassichuk (1975) suggested that rocks referred by Thorsteinsson (1974) to the Nansen Formation at Blind Fiord on southwestern Ellesmere Island might better be placed in the Belcher Channel Formation. This implies that the Belcher Channel Formation of the Marginal Clastic and Carbonate Belt grades basinward into the Hare Fiord Formation of the Basinal Clastic and Evaporitic Belt, and that the Southeastern Carbonate Belt is nonexistent in the Blind Fiord area.

(d) Basinal Clastic and Evaporitic Belt. Evaporitic clastic and carbonate rocks of the Upper Mississippian - Middle Pennsylvanian Otto Fiord Formation and overlying shale and limestone of the Middle Pennsylvanian - Lower Permian Hare Fiord Formation define the limits of the Basinal Clastic and Evaporitic Belt. The belt occupies an interior part of the basin and extends southwestward from near the head of Hare Fiord on Ellesmere Island for about 800 km to Sabine Peninsula on Melville Island. In the northern part of the belt, north of Whitsunday Bay on Axel Heiberg Island, both the Otto Fiord and Hare Fiord formations outcrop; south of Whitsunday Bay, the presence of these formations can be demonstrated only from subsurface data and from evaporitic diapirs derived from the Otto Fiord Formation. As discussed earlier, an upper Paleozoic carbonate facies belt that probably extends across the long axis of the basin near Lougheed and Mackenzie King islands indicates that the Basinal Evaporitic and Clastic Belt may not have been continuous, and that Otto Fiord sediments were deposited in two or more depocentres.

Reddish weathering clastic rocks of the Borup Fiord Formation underlie the Otto Fiord Formation on Ellesmere Island, and also occur locally beneath the Nansen Formation of the Northwestern Carbonate Belt along the northern margin of the Basinal Clastic and Evaporitic Belt.

(e) Northwestern Carbonate Belt. This belt is represented by the Upper Mississippian - Lower Permian Nansen Formation, which is composed of up to 2100 m of limestone, the lateral equivalent of the Hare Fiord and Otto Fiord formations in the Basinal Clastic and Evaporitic Belt. The southernmost exposure of the Nansen Formation is at Bunde Fiord on northwestern Axel Heiberg Island and from there it extends northeast for 300 km, at least to Yelverton Pass on northern Ellesmere Island. The Nansen Formation is distributed widely in the subsurface south of the Bunde Fiord area but the full extent of the formation has not yet been outlined by drilling.

All of the above facies belts contain rocks that range in age from Mississippian or Early Pennsylvanian to Early Permian (early Artinskian). Younger Permian rocks, from late Early Permian (late Artinskian) to early Late Permian (Wordian), are superposed on formations contained within these belts but the younger rocks, mainly clastics and fewer carbonates, reflect widely transgressive seas and are not distributed in comparable linear facies belts.

The oldest of the clastic formations that postdate formations in Thorsteinsson's (ibid.) lithofacies belts is the Sabine Bay Formation. The Sabine Bay of Artinskian age contains a few bands of marine limestone and sandstone but is composed mainly of nonmarine quartzose sandstone deposited in a deltaic environment. The Sabine Bay Formation also contains subordinate conglomerate and coal beds, each of which varies in thickness from 1 to 10 cm. It is discontinuous along the southern and eastern margins of the Sverdrup Basin, where it conformably overlies the Belcher Channel Formation in the Marginal Clastic and Carbonate Belt. The Sabine Bay Formation is overlain conformably by the upper Lower Permian (Roadian) Assistance Formation, which contains carbonaceous sandstone and concretionary siliceous limestone bands. In regions where the Sabine Bay Formation is absent the Assistance Formation or its deeper water equivalent, the van Hauen Formation, rests conformably on the Belcher Channel Formation. The van Hauen Formation comprises noncalcareous shales, siltstone and spicular chert and some limestone beds. In other areas of the basin the van Hauen Formation variously overlies other formations in the Basinal Clastic and Evaporitic Belt, the Northwestern Carbonate Belt and the Marginal Clastic and Carbonate Belt.

The youngest Permian formations in the Sverdrup Basin are the lower Upper Permian (Wordian) Trold Fiord and Degerböls formations. The Trold Fiord contains glauconitic sandstone as well as lesser amounts of shale and conglomerate, and occurs in eastern and southern regions of the basin where it is distinctly transgressive and where it directly overlies various older Permian and Carboniferous formations. In some areas it also rests unconformably on lower Paleozoic rocks of the Franklinian Geosyncline. The Degerböls Formation, the basinal equivalent of the Trold Fiord, contains skeletal limestone replete with brachiopods, bryozoans and crinoids and rests on the van Hauen Formation or, where the van Hauen has been truncated, directly on the Nansen Formation.

Thorsteinsson (ibid.) described two formations composed of volcanic rocks on northern Axel Heiberg Island and northern Ellesmere Island. The Audhild Formation occurs between the Borup Fiord Formation and the Nansen Formation on Kleybolte Peninsula, Ellesmere Island and its stratigraphic position suggests a probable Namurian or Bashkirian age. According to Thorsteinsson, the Audhild, which is 312 m thick at the type section, includes basaltic and spilitic flows as well as pyroclastic sediments. The Esayoo Formation was named for basaltic flows and agglomerates above the Nansen Formation 3.2 km northeast of the head of Esayoo Bay, northern Ellesmere Island; it is variously overlain by the van Hauen, Trold Fiord and Degerböls formations. The same volcanics also occur above the Hare Fiord Formation near Wood Glacier in the van Royen Ridges on northern Ellesmere Island. The Esayoo Formation contains basaltic flows and agglomerate (ibid.). We have observed these volcanics adjacent to the Nansen Formation northeast of Esayoo Bay and believe they may in fact be intrusive, rather than extrusive. Thorsteinsson (ibid.) also identified volcanic flows and pyroclastic rocks (unnamed) overlying a thick sequence of Nansen rocks near Emma Fiord on northern Ellesmere Island. Middle Pennsylvanian fossils occur 305 m below the volcanics but the top of the volcanic rocks is a recent erosion surface.

Deposits in the Sverdrup Basin that predate Otto Fiord evaporites

To assess fully the geological and climatic conditions that prevailed during deposition of the evaporitic Otto Fiord Formation, which began in Late Mississippian (early Namurian) time, it is appropriate to review the depositional history of the Sverdrup Basin immediately prior to evaporite deposition. Two formations predate the Otto Fiord Formation in the basin: the nonmarine Emma Fiord Formation and the succeeding partly marine Borup Fiord Formation were deposited between Viséan and early Namurian time.

Emma Fiord Formation

The oldest rocks in the Sverdrup Basin belong to the Viséan Emma Fiord Formation (Fig. 4), which was deposited following a prolonged period of deformation and erosion of the Franklinian Geosyncline during Late Devonian and earliest Carboniferous (Tournaisian) time. The Emma Fiord varies in thickness from less than 100 m to slightly more than 300 m and contains carbonaceous shale and siltstone, lensoid sandstone, fine conglomerate and coal seams.

The Emma Fiord Formation has yielded an abundant spore microflora (Playford and Barss, 1963) as well as abundant and, as yet, undescribed leaf and stem impressions. In addition, undescribed ostracodes and vertebrate bones now are known in the formation (H.W. Pfefferkorn, pers. com., 1974). A few thin limestone beds in the formation contain algae of nonmarine aspect.

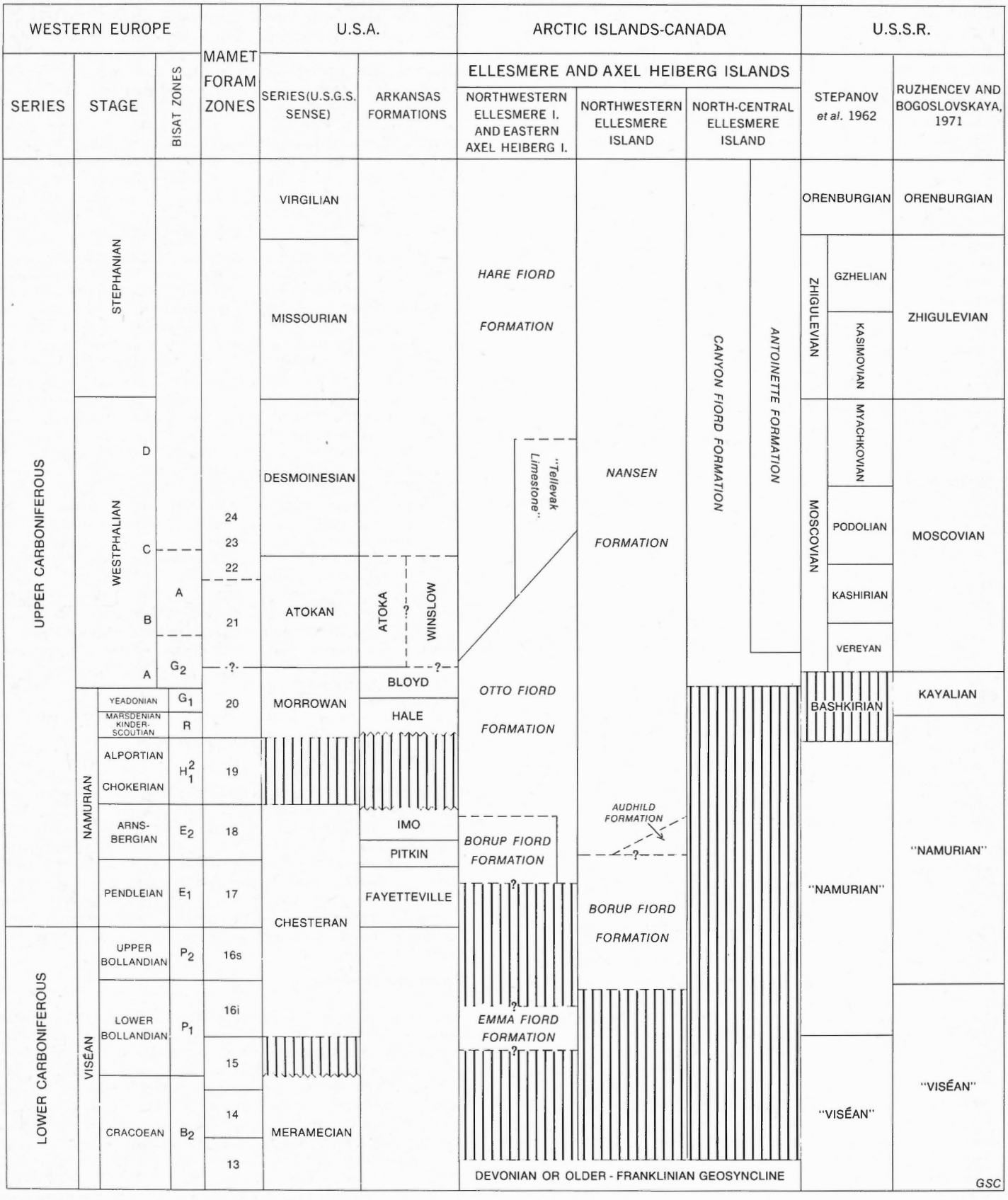


Figure 4. Relationships between Carboniferous formations in the Arctic Islands and standard successions in the North American midcontinent and Europe (modified from Nassichuk, 1975).

The Emma Fiord occurs in two widely isolated regions of the Sverdrup Basin, one on the northwestern side and another along the southern margin. In the former region, isolated occurrences are known to overlie older Paleozoic rocks of the Franklinian Eugeosyncline at the Svartevaeg Cliffs on northeastern Axel Heiberg Island (Kerr and Trettin, 1962), at Kleybolte Peninsula on northwestern Ellesmere Island (Thorsteinsson, 1974) and near Clements Markham Inlet, northern Ellesmere Island (Christie, 1964). The type section of the Emma Fiord is at Kleybolte Peninsula, where it unconformably overlies the Ordovician-Silurian Imina Formation and is overlain conformably by the Borup Fiord Formation. Near the southern margin of the basin the Emma Fiord occurs on Grinnell Peninsula, Devon Island (Kerr, 1975). There it fills synclinal depressions in the Franklinian Miogeosyncline topography· it unconformably overlies the Silurian Allen Bay and Cape Storm formations and is overlain conformably by Middle Pennsylvanian (Moscovian) strata in the Canyon Fiord Formation.

The lack of identifiable marine fossils, the abundance of plant remains, and general rock types suggest that the Emma Fiord sediments were deposited mainly in nonmarine lacustrine and fluvial environments; thin, discontinuous pisolitic and pseudo-oolitic limestone beds in the exposures on Grinnell Peninsula probably are of lacustrine or vadose diagenetic (caliche) origin, and do not indicate a marine transgression; they also contain nonmarine algae. The Emma Fiord appears to have been confined to a near-margin environment in the incipient Sverdrup Basin. Indeed, development of the active depositional phase in the Sverdrup Basin, perhaps initiated by subsidence along major faults, may be recorded by the dramatic increase in clast size in the Canyon Fiord Formation conglomerates, which overlie the Emma Fiord Formation on Grinnell Peninsula, and in the Borup Fiord Formation, which overlies the Emma Fiord elsewhere.

Borup Fiord Formation

The Borup Fiord Formation (Figs. 3, 4, 5) marks the first major marine transgression in the Sverdrup Basin. It contains Late Mississippian (Namurian) fossils in its upper part but no fossils have been recovered from the lower part, which may be nonmarine. The formation is characterized by red- and green-weathering conglomerate, sandstone, siltstone and shale. Locally interbeds and lenses of carbonate rock, including oolitic and bioclastic limestone and dolomite, are abundant in the uppermost part of the formation. The formation varies in thickness from a feather edge to 460 m, and either rests conformably on Emma Fiord rocks or unconformably on deformed lower Paleozoic rocks of the Franklinian Geosyncline. The upper part of the Borup Fiord probably is continuous with basal beds of the lithologically similar Canyon Fiord Formation (Fig. 3), which occurs along the southern and eastern margins of the basin; this inferred relationship has not yet been substantiated by mapping. The Borup Fiord Formation varies considerably from place to place in the basin both in thickness and in lithology, reflecting different provenances. On Kleybolte Peninsula, northern Ellesmere Island, where the Borup Fiord rests on siltstone and shale of the Emma Fiord Formation, the conglomerate clasts are small and red coloration is subdued. On the north side of Hare Fiord, however, where the Borup Fiord rests on Cambrian rocks with angular unconformity, the red is pronounced and conglomerate clasts are large (some greater than 24 cm). This reflects a closer source than at Kleybolte Peninsula: perhaps a lower Paleozoic basement high such as the Grantland Uplift (Trettin, et al., 1972). In the Hare Fiord region the red and coarse, lensoid conglomerates in the Borup Fiord Formation suggest deposition as coalescing alluvial fans, probably by intermittent streams cutting uplifted Cambrian rocks under a climatic regime that was characterized by marked seasonal contrasts (arid, hot summers; wet winters). Marine influence became significant only late in the deposition of Borup Fiord red beds or fanglomerates. The gradational change from a nonmarine or marginal marine environment is marked by the general increase of carbonate units at the top of the Borup Fiord red beds and the interbedding of red sandstone units of Borup Fiord aspect with carbonates of Otto Fiord aspect at the upper contact. Some of the carbonate lenses are identifiable as discrete oolitic and skeletal sand bars or banks enclosed in red beds.

It is interesting to note that carbonate rocks are relatively well developed near the top of the Borup Fiord in regions where the formation is overlain by the evaporitic Otto Fiord Formation. Conversely, where the Otto Fiord has graded into carbonate, such as near the head of Hare Fiord, Borup Fiord rocks beneath the Nansen Formation are almost completely devoid of a carbonate component. At Yelverton Pass on northern Ellesmere Island, the Borup Fiord is thickest and rests on Ordovician carbonate rocks. In this area the formation underlies carbonate rocks of the Nansen Formation and is composed mainly of thin-bedded shale and siltstone, with only small amounts of conglomerate; carbonate rocks are absent from the formation in this area. Unfortunately, the Borup Fiord Formation in the Yelverton Pass area has yet to yield fossils and additional investigations are required. Mayr (pers. com., 1975) recently identified a new formation of Devonian clastic red beds farther to the northwest of Yelverton Pass and it is conceivable that locally Borup Fiord red beds could lie directly upon Devonian red beds.

Age relationships of the Emma Fiord and Borup Fiord formations

Playford and Barss (1963) identified 19 Viséan spore species belonging to the "*Aurita* assemblage" from a single sample in the Emma Fiord Formation from the Svartevaeg Cliffs on northern Axel Heiberg Island. The *Aurita* assemblage was defined in general terms by Playford (1963). More recently, Barss (pers. com., 1975) identified 14 species belonging to the same assemblage from the type section of the Emma Fiord 37 km northeast of the Svartevaeg Cliffs locality at Kleybolte Peninsula, Ellesmere Island. Similarly, Barss (ibid.) identified 51 species from the Emma Fiord Formation at Grinnell Peninsula and indicated a close resemblance to the *Aurita* assemblage except that an important index for the latter assemblage, *Murospora aurita*, is absent. He also indicated that three species (*Knoxisporites stephanephorus, Knoxisporites triradiatus and Secarisporites remotus)* in the Grinnell Peninsula flora have not been reported elsewhere in the *Aurita* assemblage but occur in the Viséan - lower Namurian interval in eastern Canada, the United States and Great Britain.

Fossils are exceedingly rare and have not been reported from the lower and middle parts of the Borup Fiord Formation, but a rather diverse fauna of early Namurian (Late Mississippian) foraminifers has been reported from near the top of the formation in the vicinity of the type section near Stepanow Creek on Hare Fiord in northern Ellesmere Island (Mamet, in Thorsteinsson, 1974). Faunas of comparable age have been identified by Mamet (in Nassichuk, 1975) from near the base of the type section of the Nansen Formation (which is 29 km east of the type section of Borup Fiord and 1 km east of the head of Hare Fiord) and also from the base of the Otto Fiord Formation, which grades into the Nansen Formation near the head of Hare Fiord.

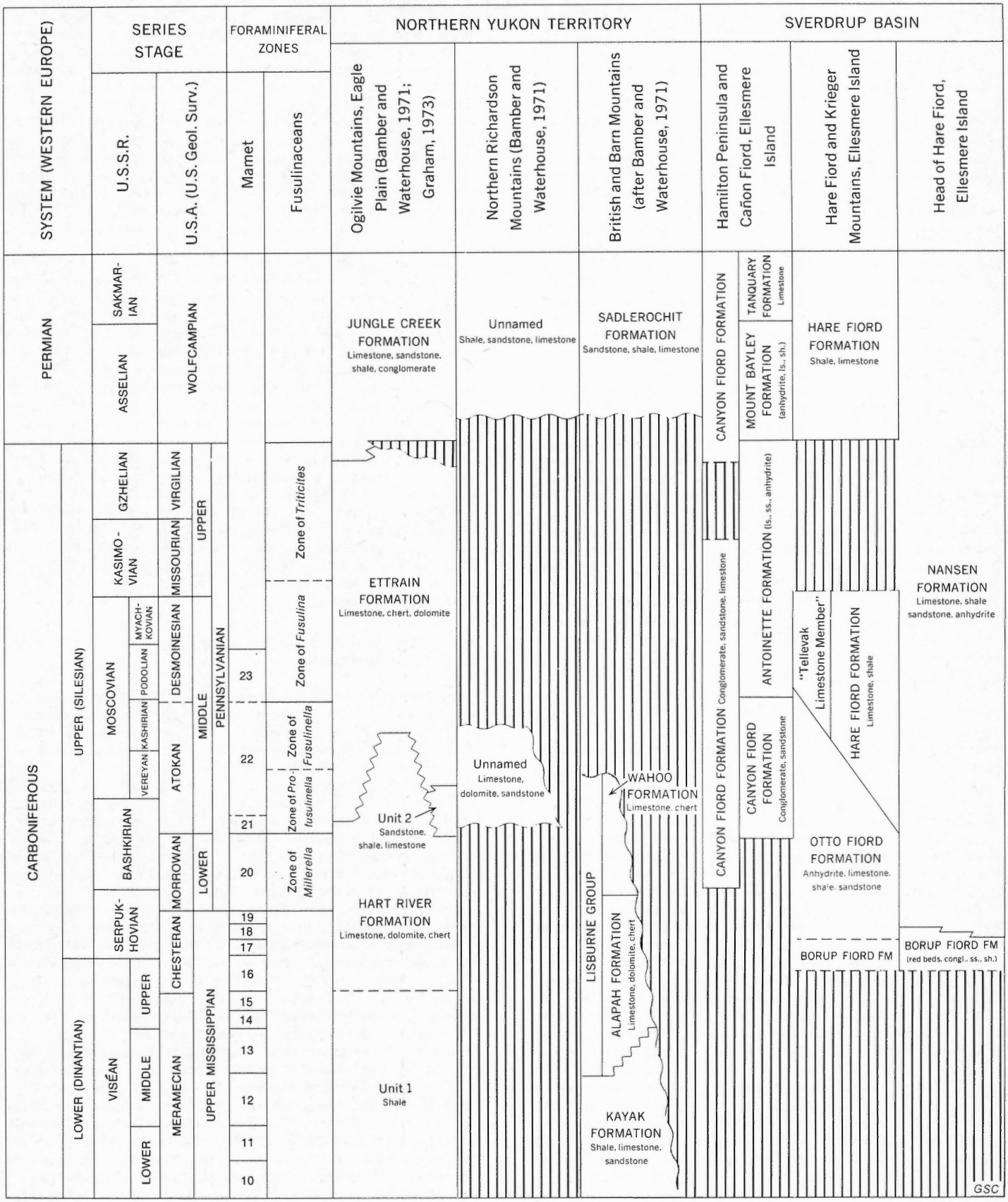


Figure 5. Relationships between Carboniferous formations in the Arctic Islands and northern Yukon Territory (from Nassichuk and Bamber, 1979).

Sediments not unlike those within the Emma Fiord Formation and containing comparable microfloral elements extended over much of the U.S.S.R. (Luber and Waltz, 1941), Spitzbergen (Playford, 1962, 1963) and the Yukon Territory (Hacquebard and Barss, 1957; Barss, 1967) during Viséan time. A particularly striking similarity exists between Mississippian successions in interior regions of the Sverdrup Basin and Vestspitzbergen. In the latter area a sequence of shale and sandstone, designated the Culm beds, corresponds closely in age, lithology and thickness to the Emma Fiord Formation in Arctic Canada (Lyutkevitch, 1937; Orvin, 1940; Gee et al., 1953). Similarly, greenish weathering conglomerate and sandstone near the top of the Culm in Spitzbergen closely resemble the Borup Fiord Formation, which overlies the Emma Fiord Formation in the Sverdrup Basin. Mississippian sandstone, conglomerate, dark grey shale and coaly beds that are comparable with the Emma Fiord Formation also occur in the Terrestrial Group in northeastern Greenland. Further, the lower part of the Lower Marine Group in northeastern Greenland also bears a close temporal and lithological resemblance to the Borup Fiord Formation.

The Emma Fiord can be correlated also with the Kayak Formation in Alaska (Bowsher and Dutro, 1957) and the northern Yukon (Bamber and Waterhouse, 1971), and also with the Mattson Formation and Stoddard Group in northeastern British Columbia, southeastern Yukon and southwestern District of Mackenzie (Bamber et al., 1968; Bamber and Waterhouse, 1971).

At least the upper part of the Borup Fiord Formation was deposited contemporaneously with the Kogruk Formation and Alapah Limestone in the Lisburne Group in northern Alaska (Mamet and Armstrong, 1972), and with the Alapah Limestone and Ettrain Formation in the northern Yukon Territory (Mamet, in Bamber and Waterhouse, 1971).

THE OTTO FIORD FORMATION

Distribution

The Otto Fiord Formation was named for a sequence of interbedded anhydrite, shale and limestone east of van Hauen Pass on northern Ellesmere Island (Figs. 6, 7). Although other primary evaporitic rock types are not known from surface exposures, a halite facies has been identified recently from the subsurface at Ellef Ringnes Island (Davies and Nassichuk, 1975; Davies, 1975a). In the type area the Otto Fiord has a thickness of about 420 m. The base of the section is marked by a major thrust fault; at various places dark grey shale of the Lower Triassic Blind Fiord Formation and limestone of the Upper Permian Degerbols Formation are visible in the footwall of the fault. Some 22 km east of the type section, just west of the mouth of Stepanow Creek on the north side of Hare Fiord, the base of the Otto Fiord is unfaulted. At this locality the Otto Fiord is 240 m thick and rests conformably on red beds of the Borup Fiord Formation. Wardlaw and Christie (1975) indicated that the Otto Fiord Formation is 600 m thick locally northeast from the mouth of Stepanow Creek and they also reported a dramatic thickening of carbonate units within the Otto Fiord. In this area the Otto Fiord Formation grades to carbonates of the Nansen Formation and we believe that the local dramatic thickening reported by Wardlaw and Christie may be due partly to the appearance of thickening foreslope tongues of carbonate from the Nansen Formation.

Typical Otto Fiord rocks near van Hauen Pass are overlain conformably by the Middle Pennsylvanian-Lower Permian Hare Fiord Formation, a dark grey weathering unit of thin-bedded cherty limestone, siltstone and shale with thick developments of fossiliferous fenestellid bryozoan mounds near the base (Fig. 8). Whereas the shale and siltstone within the Hare Fiord appear to reflect deposition in a starved basin, the biogenic mounds may have developed near the shoreline or on topographic highs within the starved basin.

As indicated previously, the Otto Fiord and the Hare Fiord formations define the Basinal Clastic and Evaporitic Belt (Thorsteinsson, 1974). This belt delimits a distinctive elongate lithofacies in the interior of the basin, which extends from near Chapman Glacier about 40 km east of the head of Hare Fiord to Melville Island several hundred kilometres to the southwest. Near the northern and northwestern edges of the Basinal Clastic and Evaporitic Belt, both the Otto Fiord and Hare Fiord formations grade to a thick sequence of shelf carbonates, the Nansen Formation, which is the principal formation in Thorsteinsson's (ibid.) Northwestern Carbonate Belt (Fig. 23).

Whereas both the Otto Fiord and Hare Fiord formations outcrop in relatively little disturbed, bedded sequences between Hare Fiord and Whitsunday Bay on eastern Axel Heiberg Island, it is clear that at least the Otto Fiord is distributed far more extensively than surface exposures indicate, for it is the source of evaporites in all of the more than 100 evaporitic piercement structures on Axel Heiberg, Ellef Ringnes, Amund Ringnes and Melville islands (Fig. 1). According to Tozer and Thorsteinsson (1964), large evaporitic piercement structures are particularly common where the basal beds in the overlying Hare Fiord Formation are shale and siltstone, but they are encountered only rarely in areas where relatively competent limestone beds overlie the Otto Fiord Formation. This relationship, however, simply may reflect the mutual depositional control and distribution of the Otto Fiord evaporites and overlying Hare Fiord argillaceous sediments, rather than control by differences in competence.

Where the Otto Fiord Formation is not exposed in continuous successions at the surface, its lithology has been assessed from rock types observed in piercement structures. We have not, however, attempted to review the events which ultimately led to the formation of these structures. For a general review of piercement structures within the basin, reference is made to Thorsteinsson (1970), who has categorized occurrences in several different structural settings, and to Balkwill (in press), who has studied various domes in the Ringnes Islands in some detail. In addition, important contributions to understanding the structural relationships and mechanics of evaporite emplacement of piercement structures have been made by Norris (1963), Schwerdtner and Clark (1967), Hoen (1964), and Gould and De Mille (1964).

Although the Otto Fiord Formation is composed principally of anhydrite and limestone with lesser amounts of shale in the vicinity of van Hauen Pass, Ellesmere Island, quartzose sandstone units are up to 30 m thick adjacent to some major carbonate mound developments within the formation. These sandstone units thicken to 100 m about 32 km south and southeast, between the Blue Mountains and the Krieger Mountains. The sandstone from the latter area weathers variously from medium grey to pink and contains a variety of sedimentary structures, including ripple marks and crossbeds. On the basis of current directions determined in the field, we believe that sandstone units at van Hauen Pass were derived from a northerly or northeasterly source whereas those in the Krieger Mountains were derived from the southeast. We have recognized similar sandstones in the Otto Fiord Formation north of Whitsunday Bay on eastern Axel Heiberg Island and in large bedded inclusions within the evaporite core of Barrow Dome on northeastern Melville Island. Additionally, Greiner (1963) reported 30 m of light red to pink weathering, well bedded sandstone on the western edge of the core of Malloch Dome on Ellef Ringnes Island. Roots

(1963) also described interbeds of red-weathering sandstone and banded 'gypsum' in a diapir west of Flat Sound on Axel Heiberg Island.

The regional significance of the sandstone is not entirely clear but its widespread distribution probably reflects tectonic instability in the Sverdrup Basin depression, which probably is rift controlled. Comparable red sandstone units in the older Borup Fiord Formation in the basin interior, which may in fact be continuous with the Canyon Fiord Formation, reflect a comparable tectonic instability. It is possible that sandstones in the Otto Fiord Formation are lobate extensions of the Canyon Fiord Formation (see Fig. 3).

A thick succession of halite is known to occur within the Otto Fiord Formation on Ellef Ringnes Island (see Figs. 1, 2); more than 3600 m of halite were encountered in a drillhole near Hoodoo Dome (Davies and Nassichuk, 1975; Davies, 1975a). Considering the proximity of the well to Hoodoo Dome, the thickness encountered clearly does not reflect depositional thickness; halite apparently was mobilized in this area as early as Jurassic time. Further discussion of halite in the Otto Fiord is presented in a later section.

Evaporitic rocks of Carboniferous age are known also from two other formations in the Sverdrup Basin. Thorsteinsson (1974) documented some 60 m of Middle Pennsylvanian anhydrite in the Canyon Fiord Formation near the eastern edge of the Basin at Trold Fiord, western Ellesmere Island. Trettin (1969) reported bedded anhydrite associated with shale, limestone and sandstone in an unnamed formation (Unit 10 of Trettin, ibid.) on the northern side of the basin at M'Clintock Inlet on the north coast of Ellesmere Island. Mayr (pers. com., 1976) recently discovered a thick succession of evaporites northwest of Yelverton Pass, northern Ellesmere Island, which are probably of Carboniferous age. Relationships between these evaporites and the Otto Fiord Formation have not been assessed fully, but we are of the opinion that they represent deposition in isolated marginal or near-marginal depressions contemporaneous with deposition of the Otto Fiord Formation.

In addition to the Carboniferous evaporites, Permian evaporites have been recorded from northern areas of the basin. The lowermost Permian (Asselian) Mount Bayley Formation near the eastern edge of the Sverdrup Basin on northern Ellesmere Island extends from near the mouth of Tanquary Fiord southward to Sawtooth Range east of Fosheim Peninsula. Also, an unnamed unit of anhydrite that probably is of earliest Permian age extends discontinuously along the north coast of Ellesmere Island from the northeast side of M'Clintock Inlet easterly to Clements Markham Inlet (Christie, 1964).

Intercontinental correlation

Even though the Otto Fiord Formation contains thick anhydrite units reflecting intense hypersaline conditions, practically all limestone interbeds contain an abundant, non-depauperate biota. A wide variety of ammonoids, brachiopods, conodonts, solitary corals, crinoids, trilobites, gastropods, ostracodes, bryozoans, fusulinids and other calcareous foraminifers and algae have been collected. Some of these groups have been studied in considerable detail and have yielded important data on age relationships of the Otto Fiord: ammonoids (Nassichuk, 1975), fusulinaceans (Ross, pers. com., 1976), foraminifers and algae (Mamet, in Nassichuk, 1975, and in this report), conodonts (Bender, in Nassichuk, ibid., and in this report), brachiopods (Carter, 1975, and in this report), and crinoids (Strimple and Nassichuk, 1974).

Following sections of this report discuss in some detail the biostratigraphic significance of fossil collections from specific localities throughout the basin. Prior to that discussion, the Otto Fiord Formation is compared with other circumpolar evaporite units of similar age and with standard Carboniferous successions elsewhere in North America and Europe.

The Otto Fiord Formation ranges in age from Late Mississippian (late Chesteran) to Middle Pennsylvanian (early Desmoinesian) or, in terms of western European chronological standards, from early Namurian (lower *Eumorphoceras* Zone E) to middle Westphalian (Figs. 4, 5). In terms of current popular usage in the Soviet Union, the range is from early Serpukhovian to middle Moscovian (Kashirian and possibly Podolian). Carter (1975) described 15 brachiopod species from Melville Island and western Axel Heiberg Island in southern regions of the Basinal Clastic and Evaporitic Belt and suggested tentatively the possibility of an even older Mississippian (Viséan) age. A Viséan age, however, cannot yet be substantiated by any other fossil groups.

The lithological similarity between Carboniferous successions in interior regions of the Sverdrup Basin and contemporaneous strata in northeastern Greenland and Svalbard (Vestspitzbergen) invites formation-by-formation comparison. It was indicated earlier that the oldest deposits in the Sverdrup Basin (the Emma Fiord Formation) are correlated with the Terrestrial Group in Greenland and with the lower part of the Culm beds in Spitzbergen. Similarly, the slightly younger Borup Fiord Formation in the Sverdrup Basin is correlated with the lower part of the Lower Marine Group in Greenland and with the upper part of the Culm beds in Spitzbergen. Following this parallel comparison, the Otto Fiord Formation and the lower part of the overlying Hare Fiord Formation closely resemble in lithology and age the middle and upper parts, respectively, of the Lower Marine Group in Greenland. Moreover, the Lower Gypsiferous Series and the overlying Passage Beds of Vestspitzbergen are practically identical with the Otto Fiord and Hare Fiord formations, respectively, in Arctic Canada. Of special significance for the present study is that a particularly striking comparison can be drawn between the evaporitic Ebbadalen Formation of Vestspitzbergen and the Otto Fiord Formation. The Ebbadalen, which was separated from the Lower Gypsiferous Series by Cutbill and Challinor (1965) and which was described comprehensively by Holliday and Cutbill (1972), compares with the Otto Fiord Formation almost on a bed-by-bed basis in terms of lithology, thickness and age.

Late Mississippian as well as Early and Middle Pennsylvanian correlations between North America and Europe, particularly the Soviet Union, continue to be refined (Nassichuk, 1975). Carboniferous strata are being investigated actively in three important areas in the Soviet Union where stratotypes suitable for intercontinental correlation occur; these are the Donetz Basin (Ukraine), the Ural Mountains and the Russian Platform. Of particular relevance to the present study are relationships between Bashkirian stratotypes in the Ural Mountains and Moscovian stratotypes in the Russian Platform; it has not yet been demonstrated satisfactorily that lower Moscovian strata in the Russian Platform immediately postdate upper Bashkirian strata in the Urals. The Bashkirian was abandoned as a standard reference by Ruzhencev and Bogoslovskaya (1969, 1971) because of apparent discontinuities in the succession in the Ural Mountains. Those authors preferred instead the Kayalian Stage defined by Rotai (1957). It is based on a stratotype containing numerous marine and nonmarine cycles between the Namurian and the Moscovian in the Donetz Basin. Clearly the Kayalian is not a viable alternative to the Bashkirian and is rejected by the majority of Soviet stratigraphers because the stratotype is poorly exposed and rather poorly represented by marine faunas.

A widely known foraminiferal zonal scheme advanced by Mamet (1968), Mamet and Skipp (1971), Mamet and Armstrong (1972) and Mamet (pers. com., 1976) is of particular significance in relating the Otto Fiord Formation to standard successions in North America. The Otto Fiord contains Zones 18 (Chesteran), 20 (Morrowan), 21 and 22 (Atokan), and 23 (Desmoinesian). Zone 19, generally thought to be equivalent to the *Homoceras* Zone of Bisat (1924) in western Europe (Mamet et al., 1971), and which coincides with Late Mississippian in North America, has not been recognized in the Otto Fiord Formation. Mamet (pers. com., 1973), however, has identified Zone 19 faunas from carbonate rocks near the base of the Nansen Formation, which grades to the Otto Fiord Formation near the head of Hare Fiord. Although relationships between Chesteran and Morrowan strata in North America and equivalent strata in Europe are moderately well known in the literature, the same cannot be said for relationships between Atokan strata. Since Atokan and younger, early Desmoinesian, faunas are common in the upper part of the Otto Fiord Formation on Ellesmere Island, a brief summary of Atokan relationships, designed to complement that of Nassichuk (1975), follows.

Rotai (1975), on behalf of the Permanent Carboniferous Commission of the Interdepartmental Stratigraphic Committee of the U.S.S.R., presented a stratigraphic scale of the U.S.S.R. Carboniferous System to the VIIIth International Congress on Carboniferous Stratigraphy in Moscow. He correlated the Lower Pennsylvanian Morrowan Series of North America with the Bashkirian of the Soviet Union and the combined Atokan and Desmoinesian Series (Middle Pennsylvanian) of North America with the Moscovian in the Soviet Union. Nassichuk (1975) showed that a comparable procedure generally was subscribed to by Ruzhencev and Bogoslovskaya (1971), Thompson (1964) and others. Brazhnikova et al. (1967), however, concluded on the basis of studies completed in the Donetz Basin that Mamet's Zones 21 and 22 (Atokan) contain faunas largely confined to the Bashkirian. Eisenverg (pers. com., 1975), in general agreement with Brazhnikova et al., considers the Atokan to be represented in the upper Bashkirian and the lowermost Moscovian (Vereyan) and the Desmoinesian to range from upper Vereyan to the top of the Moscovian (Myachkovian).

We are of the opinion that the Winslow Formation in Arkansas, which is depicted as Morrowan in Rotai's (1957) scheme, is equivalent to the Atoka Formation in Oklahoma; in the type area the Atoka Formation is largely devoid of fossils but Mamet (pers. com., 1974) has recognized Zone 22 foraminifers near the base. The Winslow contains the ammonoid *Winslowoceras* in direct association with *Fusulinella* (Nassichuk, 1975), which Thompson (1960) considered to characterize the upper Atokan and Desmoinesian; *Fusulinella* is confined to the Moscovian in the Soviet Union.

Although fossils of Atokan and Desmoinesian age are common in the upper part of the Otto Fiord Formation near its northern extremity in the Hare Fiord region of Ellesmere Island, they are unknown from the formation elsewhere. In all areas where the Otto Fiord occurs south of the mouth of Hare Fiord, fossils indicate either a Late Mississippian or an Early Pennsylvanian age. Rather dramatic changes in the time of deposition of the upper part of the Otto Fiord are apparent between the type area near van Hauen Pass and the Krieger Mountains some 32 km east-southeast. Whereas Otto Fiord occurrences near van Hauen Pass are adjacent to the northwestern edge of Thorsteinsson's (1974) Basinal Clastic and Evaporitic Belt, those in the Krieger Mountains are adjacent to the opposite (southeastern) edge of the belt. The uppermost beds in the Otto Fiord Formation in the type area are considered to be early Atokan or slightly older (Morrowan) but in the Krieger Mountains the upper part of the formation is late Atokan or early Desmoinesian. The biostratigraphy and sedimentology of specific sections in both areas are discussed in following paragraphs.

As stated earlier, Carter (1975) reported Viséan brachiopods from the Otto Fiord in southerly regions of the Sverdrup Basin. Marine strata of Viséan age had not been reported previously from the Sverdrup Basin but nonmarine clastic rocks containing spores and other plant remains are known from the Emma Fiord Formation in northern regions of Devon, Axel Heiberg and Ellesmere islands. Because the presence of Viséan marine rocks bears on the early history of the Sverdrup Basin, it is hoped that studies of other fossil groups will provide additional data to corroborate such a date. Carter's (ibid.) brachiopods were derived from 'blocks' associated with diapiric anhydrite at Barrow Dome on Melville Island and at 'South Fiord Dome' on western Axel Heiberg Island. Nassichuk (1975) described a variety of ammonoids from several localities within Barrow Dome and from a single locality at 'South Fiord Dome' and all ammonoids provided data suggestive of a post-Viséan, that is, a late Chesteran or even younger, Morrowan age. At present, we have no information concerning the age of the oldest rocks in the Otto Fiord Formation nor of rocks directly below the Otto Fiord Formation in southerly regions of the Basinal Clastic and Evaporitic Belt at Melville Island, and on the Ringnes Islands and western Axel Heiberg Island. In northern regions of the Basinal Clastic and Evaporitic Belt at Hare Fiord, however, strata within the Borup Fiord Formation directly beneath the Otto Fiord Formation are known to contain Late Mississippian (early Namurian) foraminifers belonging to Zone 18 (see Mamet, in Thorsteinsson, 1974). Mamet (in Nassichuk, 1975) also reported Zone 18 faunas from the lower part of the Otto Fiord Formation and the lower part of its lateral equivalent, the Nansen Formation, east of the head of Hare Fiord. Although it is conceivable that deposition of Otto Fiord evaporites began earlier in southern parts of the basin than it did in northern regions, evidence is still required.

Throughout much of the region in which the Otto Fiord Formation is exposed in relatively little disturbed stratigraphic successions from Hare Fiord on northern Ellesmere Island southward to Whitsunday Bay on eastern Axel Heiberg Island, the base of the section is disrupted by thrust faults. The lower contact with the Borup Fiord Formation is preserved intact only at Stepanow Creek on the north side of Hare Fiord and east of the head of Hare Fiord; locally in the latter area, a rather distinctive erosional topography is displayed on the top of the Borup Fiord. Some evidence for interfingering between the lower part of the Otto Fiord and the upper part of the Borup Fiord also is apparent locally east of the head of Hare Fiord. The upper contact with the Hare Fiord Formation generally is well exposed and conformable throughout this outcrop belt but locally the Hare Fiord may either downcut into or interfinger with the Otto Fiord.

Ellesmere Island

Since the Otto Fiord Formation is exposed in normal stratigraphic successions mainly on northwestern Ellesmere Island, most of the stratigraphic, paleontological and sedimentological information has been gathered from this area. Although several fault-associated diapirs of Otto Fiord anhydrite occur on northwestern Ellesmere Island, they have not been examined in detail or have not yielded fossils.

In this report, data from extensive fossil collections made from stratigraphic sections in the type area at van Hauen Pass, from the Blue Mountains, and from an outcrop belt extending east-northeast from the head of Hare Fiord are documented. In all three of these regions, the Otto

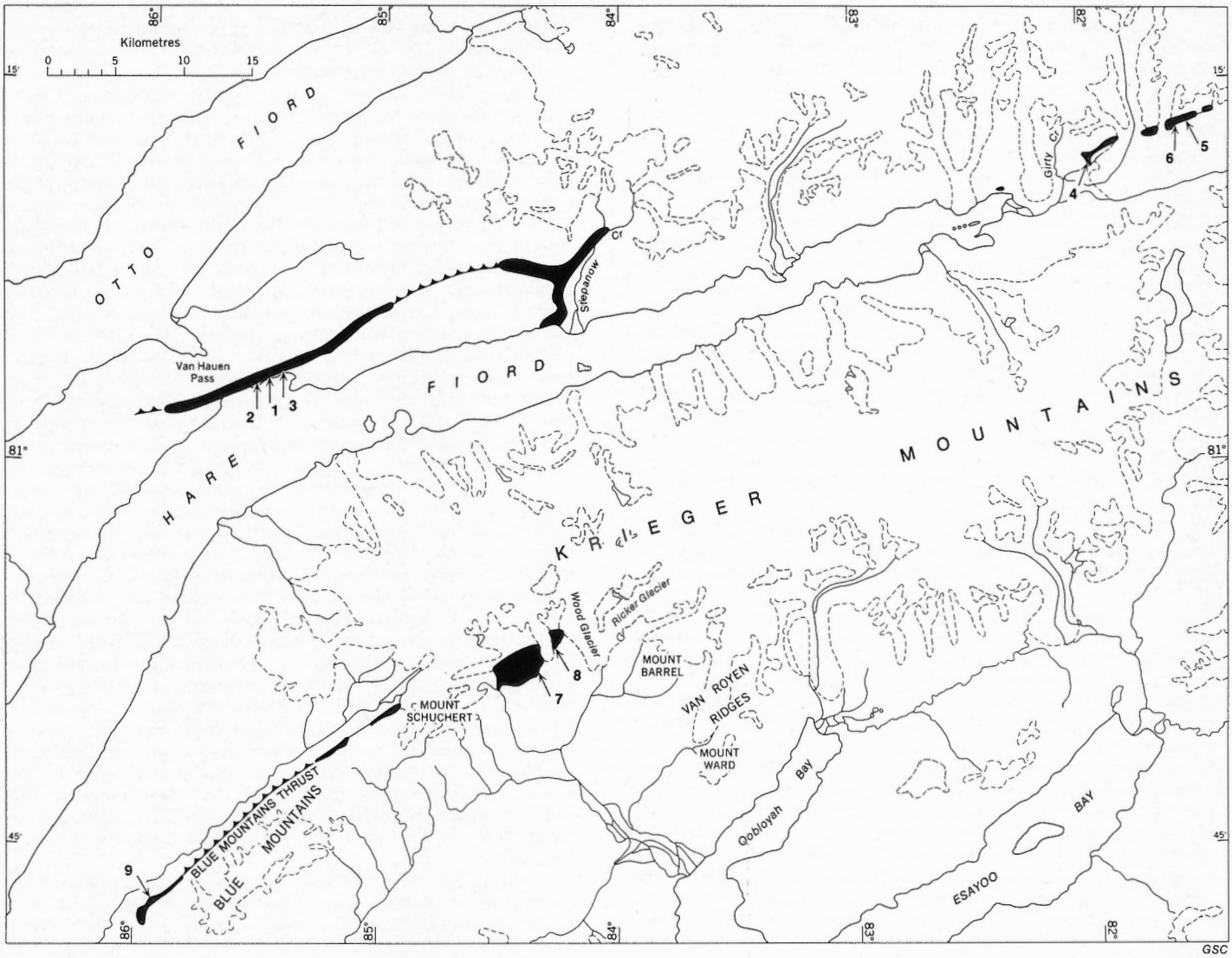


Figure 6. Index map showing distribution of the Otto Fiord Formation (solid black) and locations for measured sections 1 - 9 in the Hare Fiord and Blue Mountains areas of northern Ellesmere Island.

Fiord Formation is slightly more than 400 m thick and is overlain conformably either by carbonate rocks or by shale and siltstone of the Hare Fiord Formation. Near van Hauen Pass and in the Blue Mountains the base of the Otto Fiord has been disrupted by faults and the lower contact cannot be seen. East of the head of Hare Fiord, however, the formation rests conformably on red-weathering sandstone and limestone of the Borup Fiord Formation. Thorsteinsson (1974) considered the top of the Otto Fiord Formation to be at the top of the highest anhydrite unit in the type section. It is clear from our studies that transition from the Otto Fiord to the Hare Fiord progressed uniformly and continuously and we have no evidence for a regional disconformity between the two formations. Locally, however, the Hare Fiord appears to downcut into the Otto Fiord. It is clear also that rock types at the base of the Hare Fiord were produced in a different chemical environment than is indicated for evaporites of the Otto Fiord. This difference probably reflects an increased influx of open-marine waters in Hare Fiord time. Based on the assumption that a considerable environmental change occurred between deposition of the two formations, and to facilitate objective discussion, we follow Thorsteinsson (ibid.) in considering that the top of the Otto Fiord coincides with the top of the youngest anhydrite at least in areas where the Otto Fiord Formation has a profile resembling that in the type area. We are able to show that the youngest anhydrite beds in the type area of van Hauen Pass are somewhat older than the youngest anhydrite beds in the formation in the Blue Mountains and east of the head of Hare Fiord. In these two areas the youngest anhydrite is of late Atokan or early Desmoinesian age but at van Hauen Pass all available data suggest that the highest anhydrite in the Otto Fiord is early Atokan or older. It is not known definitely, however, whether this apparent age difference reflects diachroneity in the deposition of laterally continuous anhydrite beds or whether deposition occurred at different rates and persisted to different times in isolated depressions, which may have formed differentially depending on local rates of uplift and subsidence.

Type area at van Hauen Pass

Otto Fiord rocks form spectacular white-weathering bluffs facing Hare Fiord on the south side of van Hauen Pass (Figs. 7, 8). The formation is best exposed in the type area directly northeast of van Hauen Pass, where it is up to 400 m thick and extends continuously for nearly 10 km

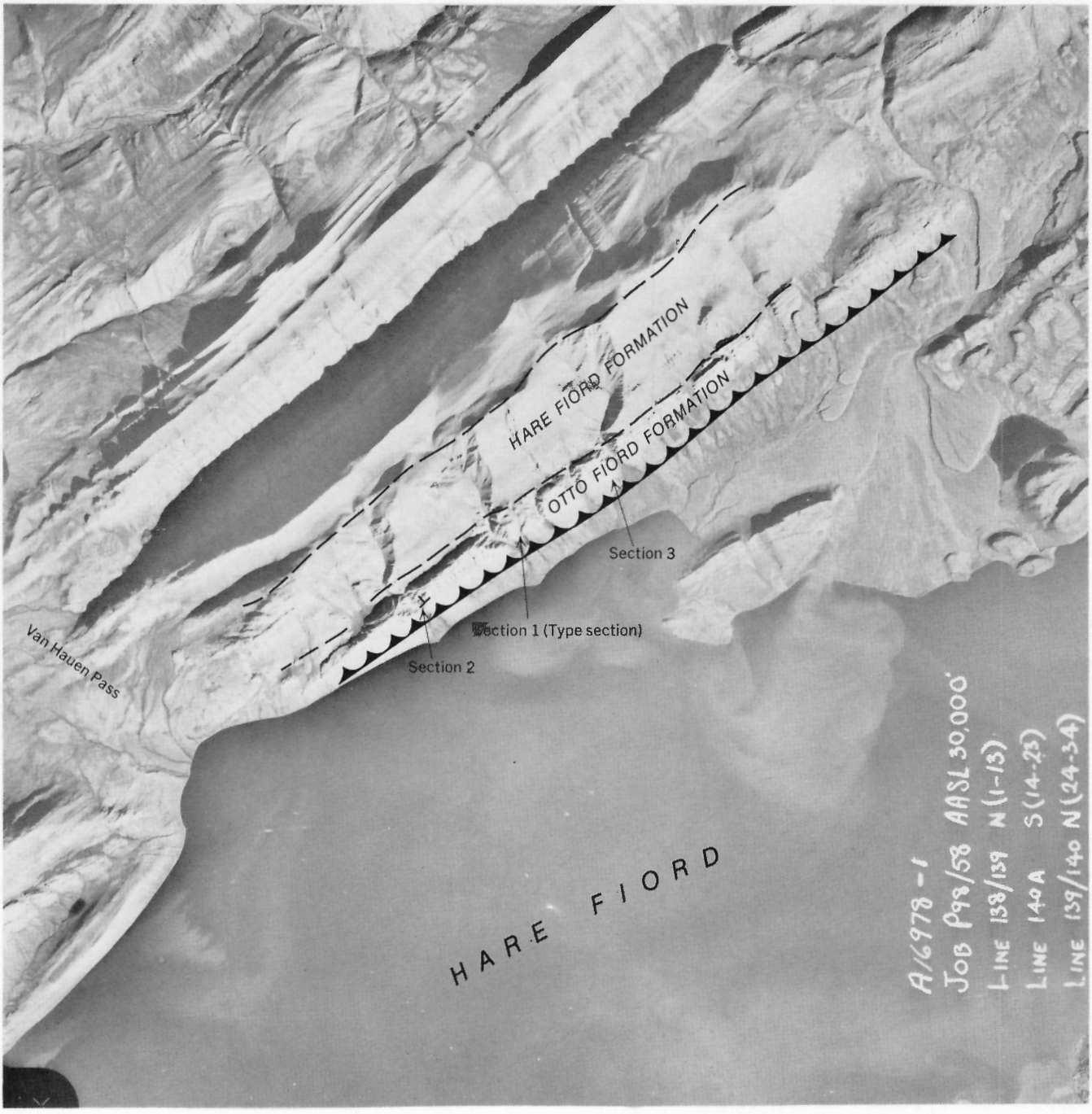


Figure 7. Locations of measured sections 1, 2 and 3 in the type area for the Otto Fiord Formation near van Hauen Pass, northern Ellesmere Island. Scale is approximately 1 cm = 0.7 km. Vertical aerial photograph A16978-1.

approximately parallel with Hare Fiord. Although the formation contains prominent bands of dark grey limestone, its general pale weathering is in sharp contrast to the dark grey weathering shale and limestone of the overlying Hare Fiord Formation. The lower units of the Otto Fiord Formation in the type area have been displaced by a major thrust fault, and truncated Otto Fiord rocks directly overlie Upper Permian limestone of the Degerböls Formation or Lower Triassic shale of the Blind Fiord Formation.

The type section of the Otto Fiord Formation (Sec. 1, Figs. 6, 7) was described by Thorsteinsson (ibid.). Its location is shown on geological Map 1309A (Otto Fiord) by Thorsteinsson and Trettin (1972). The section is 3.5 km northeast of van Hauen Pass, measured from a bay marking the maximum incursion of Hare Fiord into the pass. It is in the gully of a stream which forms the second of two small deltas in Hare Fiord immediately northeast of van Hauen Pass. The type section of the Otto Fiord Formation is 300 m thick. Although considerable biostratigraphic and sedimentological data were obtained from the type section, most of the writers' interpretations are based on data from a thicker section (410 m) measured in an adjacent stream-cut gully about 1 km southwest of the type section (Sec. 2, Fig. 7). To complete the account of the Otto Fiord Formation in the type area, data were accumulated also from a section 4.8 km northeast of van Hauen Pass (Sec. 3, Fig. 7).



Figure 8. Exposure of the Otto Fiord Formation at the type area east of van Hauen Pass, northwestern Ellesmere Island, looking southwest. Hare Fiord in foreground, iceberg-strewn waters of Otto Fiord in background. Base of the Otto Fiord Formation is truncated by a thrust fault; top is overlain by dark carbonates and shales of the Hare Fiord Formation. Note the cyclicity of dark limestone and light anhydrite in the Otto Fiord Formation, general thickening of successive carbonate and anhydrite units, and the three crinoid-algal mounds (large arrows) exposed along one horizon. Discontinuous hematite beds (H) and sandstone units (S) occur between the mounds. Carbonate breccia (B) occupies a truncational structure (small arrows) that cuts down into the upper Otto Fiord Formation at the east end of the exposure belt. Note also the load displacement (L) of beds below the large algal mound at far left. GSC 199247.

The thicker measured section of the Otto Fiord Formation (Sec. 2), described in detail here (Fig. 9), is composed mainly of rhythmically alternating units of dark limestone and light anhydrite, which may be traced almost continuously throughout the van Hauen Pass outcrop belt. About 180 m above the base of the section, four limestone mounds or reefs up to 30 m thick are exposed along one limestone marker horizon (Fig. 8). Units of recessive sandstone and red hematitic anhydrite are discontinuous in the intermound section along the same horizon.

Successive limestone and anhydrite units generally thicken upward through the section, but more subtly in the upper 100 m. Limestone units from less than 1 to 5 m thick and anhydrite units 7 to 13 m thick are more common in the lower 100 m of the section. Higher in the section some bedded limestone units are 20 to 25 m thick and some anhydrite units are 25 to 30 m thick; anhydrite flanking the limestone mound near the middle of the section reaches a composite thickness of 50 m.

Several dykes of intrusive diabase(?) cut vertically through the Otto Fiord section at van Hauen Pass. The dykes are only a few metres wide, and are found at the head of erosional gullies along the outcrop belt. This physiographic relationship is of interest since it demonstrates that the diabase retreated with a higher rate of erosion compared with the enclosing anhydrite. The apparently anomalous relationship is the result of the relatively arid arctic climate in this part of Ellesmere Island; hydration and dissolution of the anhydrite is retarded, whereas the jointed diabase is shattered and the fragments dislodged by ice-wedging.

Lithological cycles

The rhythmic alternation of limestone and anhydrite units in the Otto Fiord Formation at van Hauen Pass and the repetition within these units of similar lithologies and fabrics indicate a cyclic or rhythmic pattern of sedimentation. To give some coherence to the description of the Otto Fiord rocks in the type area and in other areas of Ellesmere Island, a composite lithological cycle (Fig. 10) is described, followed by an account of other variables.

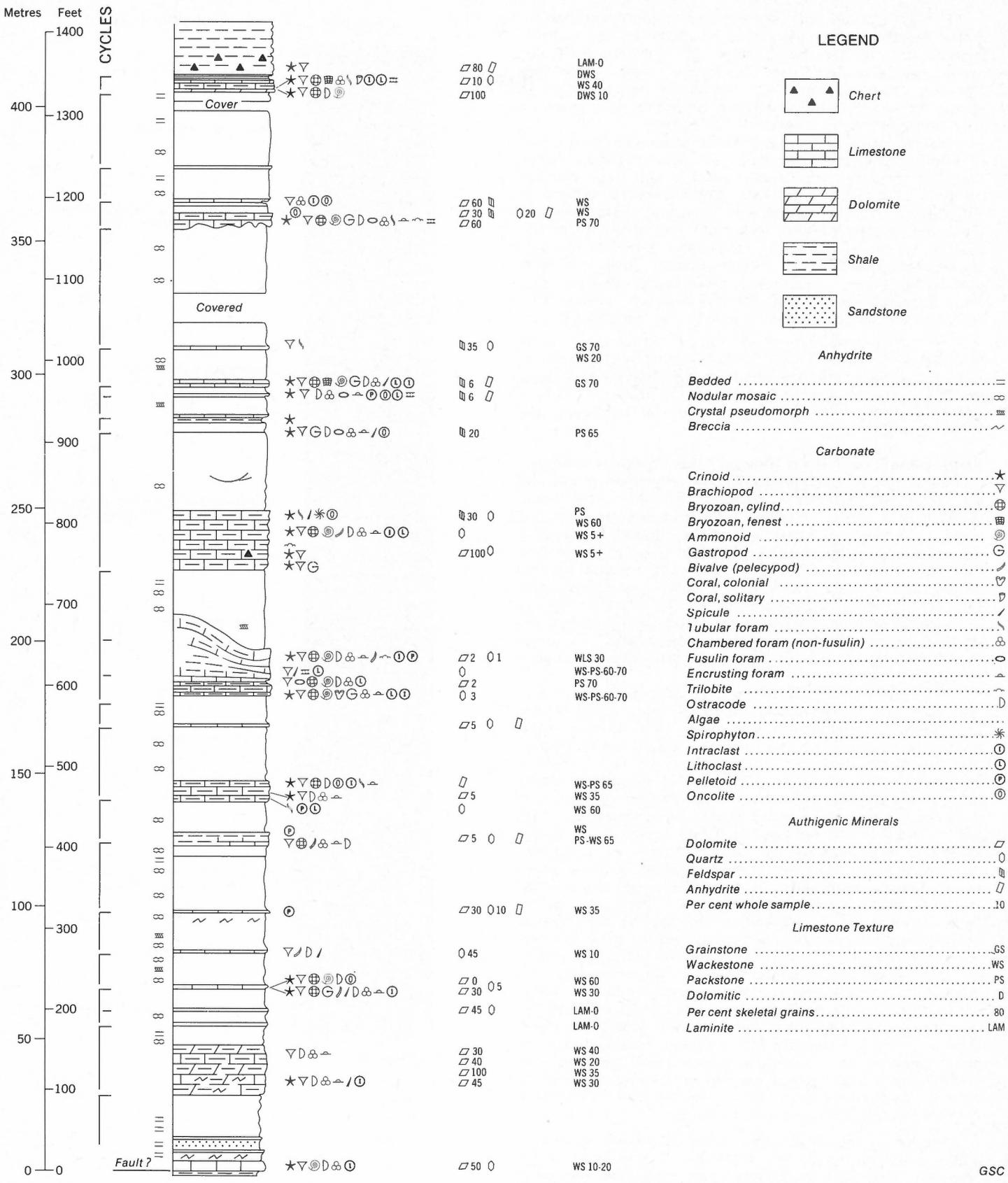


Figure 9. Lithological summary of the Otto Fiord Formation in Section 2, at the type area east of van Hauen Pass. Biotic variety of individual limestone units is generalized for the whole unit, without differentiation into a vertical sequence (see Fig. 10). Lenticular unit in the middle of the section represents an algal mound.

In the type area all limestone units containing marine shelly fossils are considered to be separate cycle components and, on this basis, about 20 limestone-anhydrite cycles may be distinguished. The cycles range in thickness from 6 m to more than 50 m. The anhydrite component is always thicker than the underlying limestone component of the cycles, usually by a factor of two to ten.

Identification of the 'type' lithological cycle as a lower limestone member and an upper anhydrite member, rather than the reverse, was based initially on field characteristics. The contact between anhydrite and limestone at the base of each limestone unit is always sharp, whereas the upper contact between limestone and anhydrite at the top of limestone units is commonly (but not always) gradational through an interlaminated zone of both lithological types. Subsequent petrographic analysis and the interpretation of depositional sequence support this choice as the logical sedimentary rhythm in these rocks.

The characteristics of a typical lithological cycle (Fig. 10) in the Otto Fiord Formation in the type area at van Hauen Pass are described in the following section, beginning at the base of the cycle. Figures 11 to 16 illustrate some of these characteristics for carbonate and anhydrite units.

Basal contact and lower beds of the limestone member. The basal contact between the top of an anhydrite unit and the overlying limestone member of the next cycle invariably is sharp, with no intergradation of the two rocks. The sulphate within 1 m of the contact in places is hydrated to granular, recessive gypsum (Fig. 10), suggesting that the limestone unit was an active, permeable aquifer during the late stage of erosional unloading of these rocks. This contact may be erosional or solutional but commonly it is planar with only a few major irregularities suggesting infilling of an irregular surface by carbonates. A major exception to this, however, is the anhydrite substrate below limestone mounds.

The lowermost part of the typical limestone unit is very dark grey to black and commonly laminated (uniform and continuous laminae), with open vugs several millimetres in diameter at the surface. Small, later diagenetic fractures filled with calcite are common. The lower carbonate beds usually are composed of calcite or partly dolomitized calcite microspar or pseudospar forming well developed, regular laminae several millimetres to 1 cm thick. Boundaries between laminae often are microstylolitic, and marked by concentrations of flocculant organic detritus, opaque material (sulphide, organic argillaceous), fecal pellets (Fig. 11) or pelloids, and a few small euhedral quartz crystals authigenic on quartz silt nuclei. Some laminae are composed of equant to size-graded, brown (transmitted light) calcite crystals; in some coarsely recrystallized fabrics, calcite crystals reach 80 μm in size.

All calcite and dolomite fabrics in these laminites are diagenetic; other diagenetic events include calcite filling of fractures, authigenic growth of anhydrite crystal aggregates by replacement of laminae, and gypsum fracture-filling. Calcite fracture-fill predates replacement by anhydrite crystals. Stylolites, some vertical, predate gypsum fracture fills. Present surface weathering of anhydrite crystal aggregates within 1 cm of the exposed rock surface produces the solution vugs noted earlier.

Higher in the basal beds, although calcite spar laminates may still occur, the carbonate of some units is less regularly laminated and contains more bioclasts and more argillaceous (and organic?) material. Fecal pellets and pelloids, rounded micritic intraclasts, tubular foraminifers and various algal bioclasts may be common components in a wackestone matrix (Fig. 11). A few bioclasts

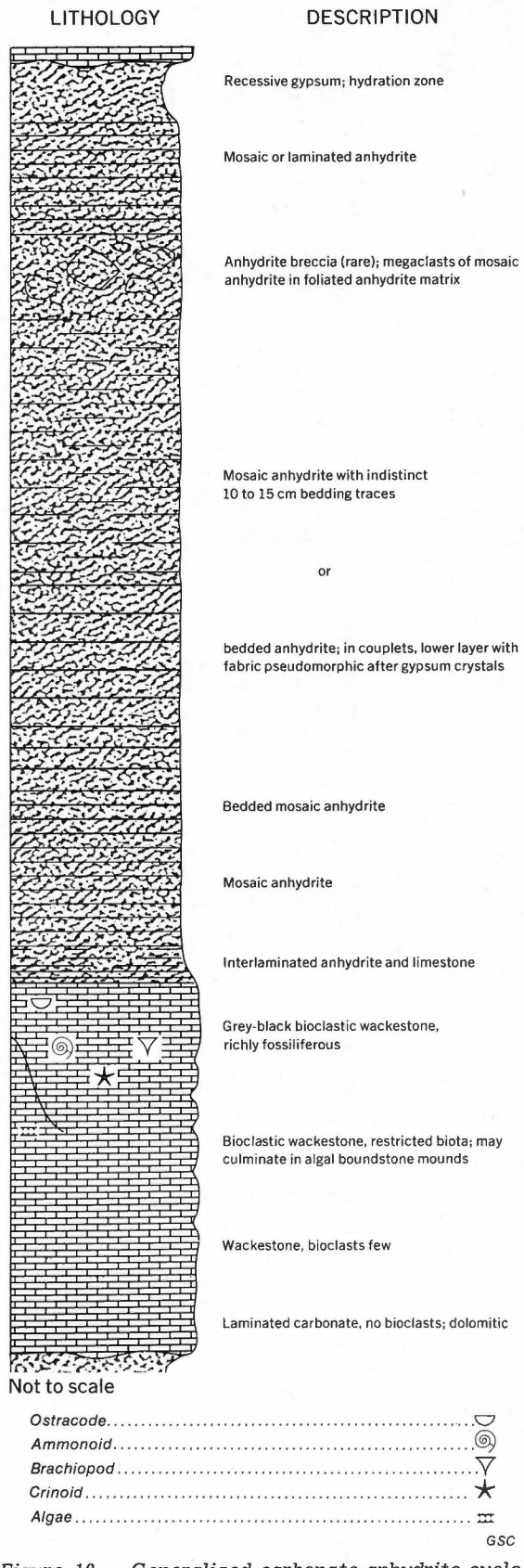
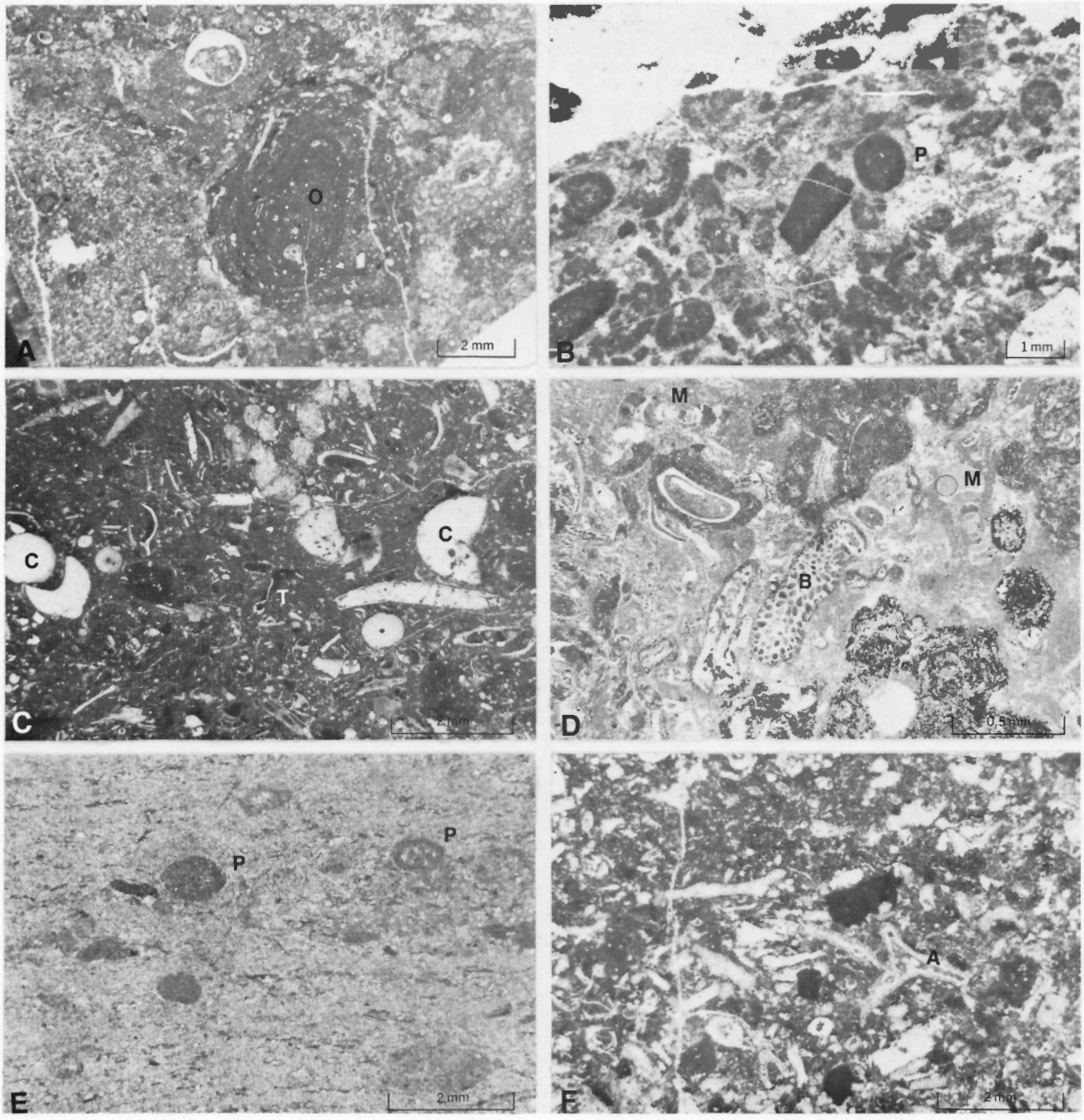


Figure 10. Generalized carbonate anhydrite cycle for the Otto Fiord Formation in the type area (Sec. 2) near van Hauen Pass. Cycle thickness varies from about 8 m to more than 50 m.



Pellet ..... **P** *Crinoid* ..... **C** *Algae* ..... **A** *Oncoid* ..... **O**

Tribolite ..... **T** *Bryozoan* ..... **B** *Ammonoids* ..... **M**

A. Indistinctly laminated unfossiliferous limestone from the basal beds of a carbonate unit; limestone is clastic-textured, and contains only a few micritic pellets and scattered organic detritus.

B. Algal wackestone and packstone containing abundant beresellid or kamaenid algae (A) from the lower beds of a limestone unit transitional from the basal unfossiliferous beds into the marine microfacies.

C. Skeletal wackestone with a varied marine biota of crinoids, bryozoans, brachiopods, trilobites, foraminifers and other organisms in a bioturbated, spicular matrix; characteristic of beds in the middle and upper parts of limestone units.

D. Skeletal wackestone with some algal oncoids, bryozoans (B), ammonoids (A), brachiopods and other bioclasts in a spicular matrix, from the marine component of a limestone unit.

E. Algal oncoid (O) in a bioturbated wackestone matrix in the upper beds of a limestone unit.

F. Fecal pellets from the uppermost beds of a limestone unit, immediately below and transitional into the interlaminated zone of limestone and anhydrite. Pellets of this type are characteristic of thin limestone beds within anhydrite units, and apparently record the existence of a nonbenthonic biota of hypersaline aspect.

Figure 11. Characteristic lithologies of Otto Fiord carbonate units in the type area near van Hauen Pass (photomicrographs from Sec. 2 samples).

of echinoderms, bryozoans, brachiopods and other foraminifers also may be present. However, the percentage of bioclasts in the lowermost beds of each unit is invariably less (10 - 30%) than in overlying beds.

Many of the bioclastic wackestones are partly dolomitized; in some, 30 per cent or more of the micritic matrix may be dolomitized. Other authigenic minerals include euhedral feldspar (Na or K) and quartz.

Middle beds of the limestone member. Many carbonate units contain thin, recessive shale beds and shaly partings between carbonate beds. These argillaceous beds tend to be concentrated in the middle of individual carbonate units. The shale is calcareous, greenish grey to grey-black, rarely reddish grey and, in places, encloses lensoid pods of bioclastic limestone; the shale also contains scattered ammonoids and other megafossils. Carbonate beds above, below and within the shaly interval generally are composed of well bedded, dark grey to black bioclastic limestone. They rarely show surface solution vug development.

Limestone beds in the lower midunit are bioclastic wackestones containing 10 to 40 per cent by volume of skeletal grains, whereas carbonates in the upper part of the midunit are bioclastic wackestones and packstones with 60 per cent or more of skeletal grains.

In the lower beds of the unit, common bioclasts include echinoderm (crinoid) columnals and spines, a variety of brachiopod valves and spines, ostracodes, tubular foraminifers and other chambered benthonic foraminifers (Fig. 11); algae are also common in some places. Less common are ammonoids and bryozoans. Algal-bryozoan oncoids and bored molluscan(?) bioclasts are common in the upper beds of units but rare in the lower beds. Larger bioclasts are enclosed in a matrix of unsorted, fine-grained bioclastic detritus and dark micritic carbonate. The preservation of micritic pelletoids and pelloids within articulated brachiopods and other microenvironments where they have been protected from compaction suggests that much of the fine matrix sediment originally was in a pelloidal form. Rounded intraclasts of bioclastic mudstone are common in some beds.

The homogenized sediments, lack of primary depositional structures and rare preserved burrows attest to a high degree of bioturbation in these carbonate beds.

With some exceptions, the middle and upper beds of typical carbonate units are not dolomitized extensively. Where dolomitization has occurred, it affects only a small amount of the matrix sediment, leaving most bioclasts unaffected. Authigenic quartz and feldspar in the form of small euhedral crystals occur in variable amounts, with feldspar generally more abundant in units higher in the section. Authigenic feldspar crystals are developed preferentially within the matrix of micritic intraclasts.

In the upper beds of the middle part of the carbonate unit, the bioclast component increases to 60 per cent or more of the rock, in wackestone to packstone limestone textures. Biotic variety is great (Fig. 11); components include ammonoids, trilobites, echinoderms (crinoids), brachiopods, bryozoans, including fenestellid forms, molluscs, ostracodes, tubular and chambered foraminifers, including many encrusting forms such as *Tuberitina* and *Tetrataxis*, and a variety of algal types. Spicules, usually as monaxon forms but with triad and tetrad forms apparently increasing higher in the section, are a new and significant component of the fine matrix sediment in these upper beds. Extensively bored clasts of molluscs, echinoderms, brachiopods and bryozoans also are more abundant in the upper beds. Rounded micritic intraclasts, described earlier for lower beds, still are present. Intense bioturbation has homogenized the sediment and left traces in the form of burrows.

The upper midunit limestones are argillaceous and, in some places, silty; however, the proportion of primary detrital quartz silt is difficult to estimate as extensive quartz and feldspar authigenesis has obscured the detrital component. Some samples contain 35 per cent by volume of authigenic silicates. As in lower beds, authigenic feldspars are developed preferentially but not exclusively in micrite intraclasts.

In some thicker carbonate units, some evidence for primary sedimentary structures formed by mechanical transport or current reworking of grains is preserved, although modified by solution compaction and other diagenetic processes.

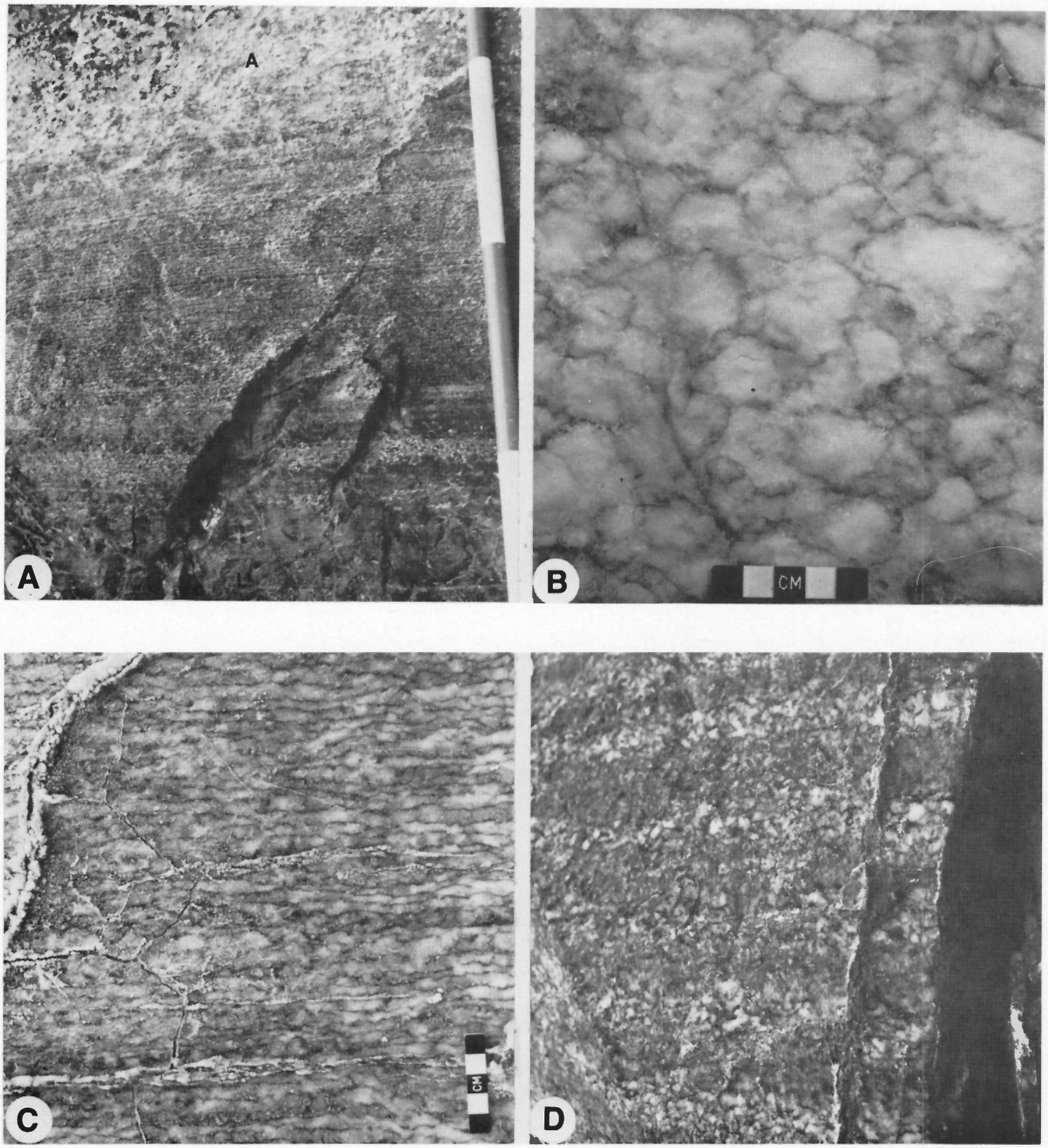
Upper beds of the limestone member. The uppermost beds of limestone members are very dark grey to black and generally are thin bedded to irregularly laminated. Black, rounded clasts up to several centimetres in diameter are common at the upper surface of the topmost beds, and are the most persistent and characteristic of all components in these carbonate beds. Surface indications of sulphate nodule or crystal aggregate solution and fracturing are more frequent in these uppermost beds than in underlying midunit beds.

In thin section, the topmost beds are found to be characterized by rounded to irregularly shaped oncoids (Fig. 11) of intergrown filamentous and segmented algae, encrusting bryozoans and encrusting foraminifers; the foraminifers appear to play only a minor and passive role in construction of the oncoids. Nuclei of oncoids include gastropods and other bioclasts. Included in the latter are all the biotic components listed earlier for the middle beds of the limestone members. Calcispheres, however, which probably are of algal origin, are a new common component. In some samples small ammonoids and beresellid algae are very prominent components of the oncoidal microfacies. Textures range from clast-rich wackestone to packstone; the matrix is characterized by spicules, some of sponge origin. Bored clasts of molluscs and other organisms are associated with the oncoids, as are coated bioclasts; the micritic coatings appear to be bound filamentous algae. Higher in the Otto Fiord section a few fusulinacean foraminifers appear in the upper beds of the limestone members.

Authigenic quartz and feldspar form a variable but persistent diagenetic element in these uppermost rocks. Dolomitization is not common, but where it does affect the oncoidal microfacies, it replaces preferentially the fine-grained micritic matrix. Anhydrite in the form of solution vug fillings, replacement fabrics and fracture fillings reappears as a significant diagenetic component in these upper rocks.

Limestone and anhydrite transition. The contact between limestone and anhydrite at the top of limestone members usually is gradational but is abrupt in places. Even where abrupt, however, the volume of carbonate impurities in the anhydrite immediately above the contact suggests a closer depositional relationship between the anhydrite and limestone at the top of the carbonate unit than at the base. Thus, for the typical cycle, a gradational limestone-to-anhydrite contact is considered the norm.

In several cycles, the limestone to anhydrite transition consists of a zone, as much as 1 m thick, of interlaminated limestone and anhydrite (Figs. 10, 12A). The proportion of carbonate decreases upward as anhydrite becomes dominant. The carbonate laminae are predominantly calcite (X-ray diffraction) and are dark grey to black, laterally extensive and persistent, but irregular in thickness on a small scale. This irregularity is a reflection of the nodular habit of the interlaminated anhydrite. Individual anhydrite



A. Interlaminated calcite and anhydrite at a transitional zone from the top of a limestone unit (L) into the base of an overlying anhydrite unit (A). Note the lateral continuity and parallelism of the laminae, the progressive upward increase in the volume of anhydrite relative to calcite, and trend toward nodular fabric in the anhydrite layers. Scale divisions are in feet (ca. 30 cm).

B. Nodular mosaic fabric typical of most anhydrite units in the Otto Fiord Formation. Impurities defining nodule boundaries are mainly subhedral calcite crystals with some organic detritus. In many units of nodular mosaic anhydrite, indistinct traces of bedding at 10 to 15 cm intervals are present.

C. Nodular laminated anhydrite formed by the lateral coalescence of diagenetic anhydrite nodules within primary depositional laminae of sulphate; primary lamination is recorded by layers of calcite and other impurities.

D. Bedding nodular mosaic fabric in anhydrite in the lower part of an anhydrite unit about 80 m from the base of Section 2. This fabric is transitional between unbedded nodular mosaic at the base of the unit and well developed bedded anhydrite above (Fig. 13). Bed units in this view are about 15 cm thick.

Figure 12. Lithologies of Otto Fiord anhydrite units in the type area near van Hauen Pass (Sec. 2).

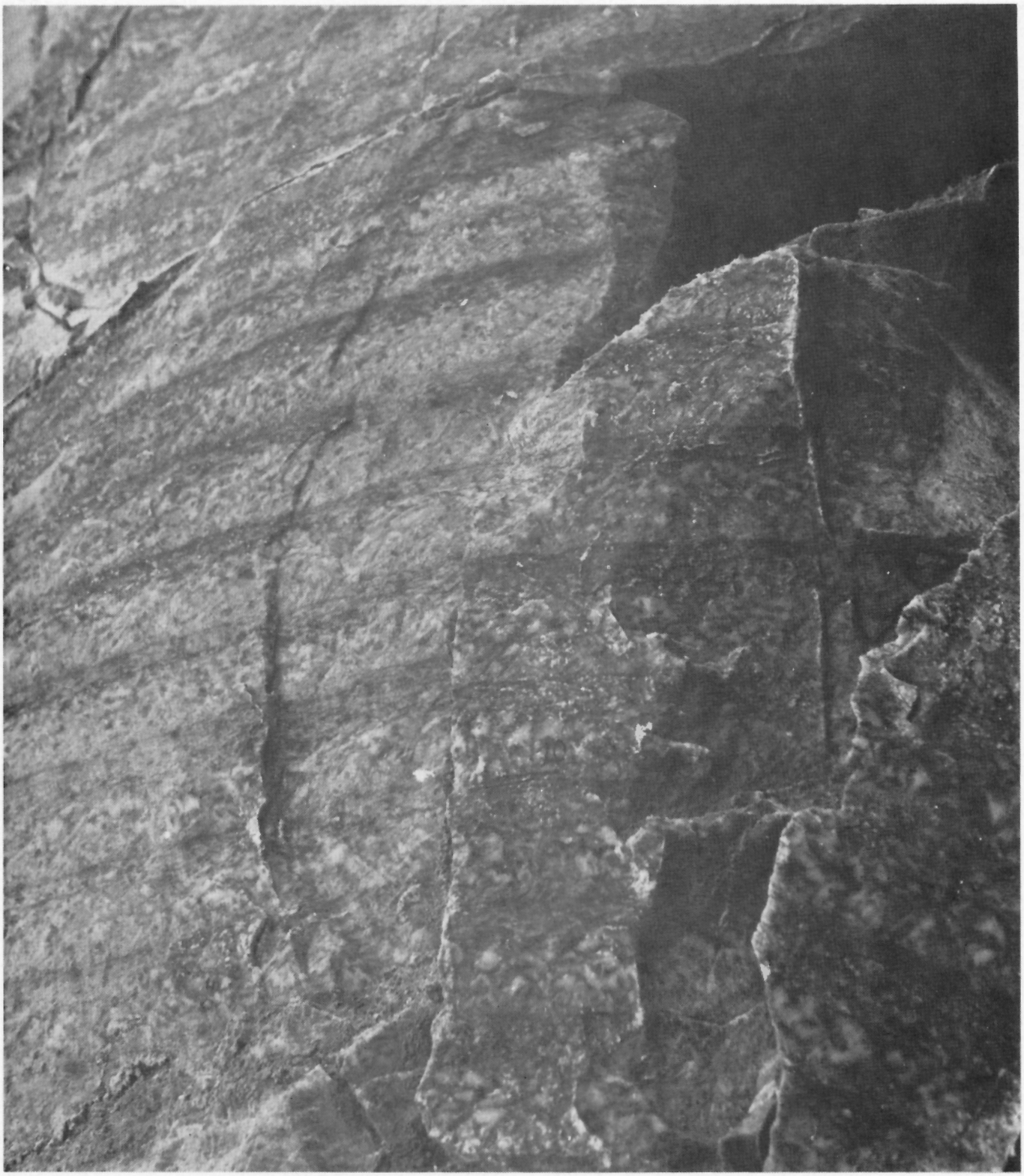


Figure 13. Rhythmically bedded anhydrite from a unit about 80 m from the base of Section 2 near van Hauen Pass. Each bed is about 15 cm thick and is composed of a thicker lower layer containing pseudomorphs after gypsum crystals, which grades upward into a thinner upper layer containing more calcitic impurities. Thick units of nodular mosaic anhydrite often contain indistinct traces of bedding at similar intervals; most of the anhydrite in the Otto Fiord Formation formed originally as bedded gypsum with this depositional character. GSC 199261.

laminae usually consist of small laterally coalesced nodules, which apparently both replaced and displaced the adjoining or enclosing carbonate layers. The interlaminated zone grades upward into carbonate-rich nodular mosaic anhydrite in which the nodule boundaries are outlined by abundant calcitic inclusions. This zone is in turn followed by relatively pure nodular mosaic of bedded anhydrite.

A variation of the interlaminated contact occurs where the top of the carbonate unit is overlain directly by nodular mosaic anhydrite with a large volume of carbonate (and possibly other) impurities. These impurities are predominantly calcite (X-ray analysis). Indistinct bedding outlined by the confluence of nodule boundaries and a concentration of impurities at 10 to 15 cm intervals is apparent in some of these basal units; in many places they are overlain by well bedded anhydrite in which beds of the same thickness are clearly defined.

In a few cycles the lower 60 cm of calcium sulphate above the carbonate-anhydrite contact has been hydrated to gypsum and is recessive in weathered profile. This effect is identical to that found in the sulphate immediately below most carbonate units; its cause is discussed later.

Anhydrite member of typical cycles. Above the basal interlaminated zone, anhydrite units are characterized by nodular mosaic fabrics (Fig. 12B). Nodule size varies considerably, and without any recognizable systematic pattern. In some units nodule size increases upward from small (several centimetres in length) at the base of the section to large (crude mean 6 cm long by 3 cm thick) higher in the section; this trend, however, may be reversed or repeated higher in the same unit. Nodule boundaries are defined by inclusions or impurities, commonly calcitic, which tend to weather out in slight relief.

The occurrence of bed forms ranging from laminae to beds up to 30 cm thick is of greater significance than the overall nodular mosaic habit of most of the anhydrite. Individual anhydrite laminae (Fig. 12C) usually consist of laterally coalesced nodules, with the bedding trace outlined by a concentration of impurities. The laminae are persistent in lateral extent and thickness. Sets of laminae and thin beds occur in most units, but without any apparent consistent pattern, although there is a trend toward decrease in bed thickness and thus to an increase in the proportion of thin beds and laminae in the upper part of some anhydrite units.

In many units bedding traces are obliterated almost totally by the later development of nodular mosaic fabrics, and can be recognized only by the concentration of impurities (usually carbonate) at regularly spaced and repeated intervals. Once recognized, bedding is discernible in the field even in the units that are composed entirely of nodular mosaic anhydrite (see Fig. 16A). Most often, bedding traces in nodular mosaic anhydrite are spaced 10 to 15 cm apart, but range up to a spacing of 25 to 30 cm. Within the constraints of outcrop width, individual beds in well exposed units appear to be laterally persistent and uniform in thickness, at least for distances of about 50 m.

The identification of these traces as true bedding planes is supported in some units by well preserved and distinct rhythmic bedding with average thicknesses in the 10 to 15 cm range (Fig. 13). In these units, a basal zone of inclusion-rich, nodular mosaic anhydrite about 1 m thick with small nodules 2 to 3 cm long is overlain by increasingly distinct beds of anhydrite averaging about 15 cm in thickness. The lowest of these beds retain an over-all nodular mosaic fabric, although a twofold subdivision of the beds generally is apparent (Fig. 12D). Initially the subdivisions appear to reflect varying concentrations and distribution of impurities (mainly carbonate) in the anhydrite, resulting in a distinct rhythmic banding on the exposed surface. In overlying beds, however, the internal organization becomes more distinct, and two rhythmically repeated fabrics are clearly defined (Figs. 13 - 15). The lower of these occupies 65 to 75 per cent of the bed, and is characterized by linear or bladelike sets of dark inclusions within the anhydrite. Many of the inclusions are small euhedral calcite crystals; others are claylike or organic detritus. They are arranged in several patterns:

(1) As broad bladelike sets up to 4 cm long and 7 mm wide, widening upward, and frequently inclined at from 40 to 50 degrees to bedding. Some terminate with a local concentration of dark impurities resembling a solution surface (Fig. 14A).
(2) As smaller bladelike sets of inclusions in V-shaped pairs, diverging upward at 25 to 45 degrees (Fig. 14B).
(3) As parallel sets of inclusions 2 to 3 m apart and 1 cm or so long, with inclusion-free anhydrite between (Fig. 14A). These sets are variously oriented, from vertical to horizontal relative to bedding.

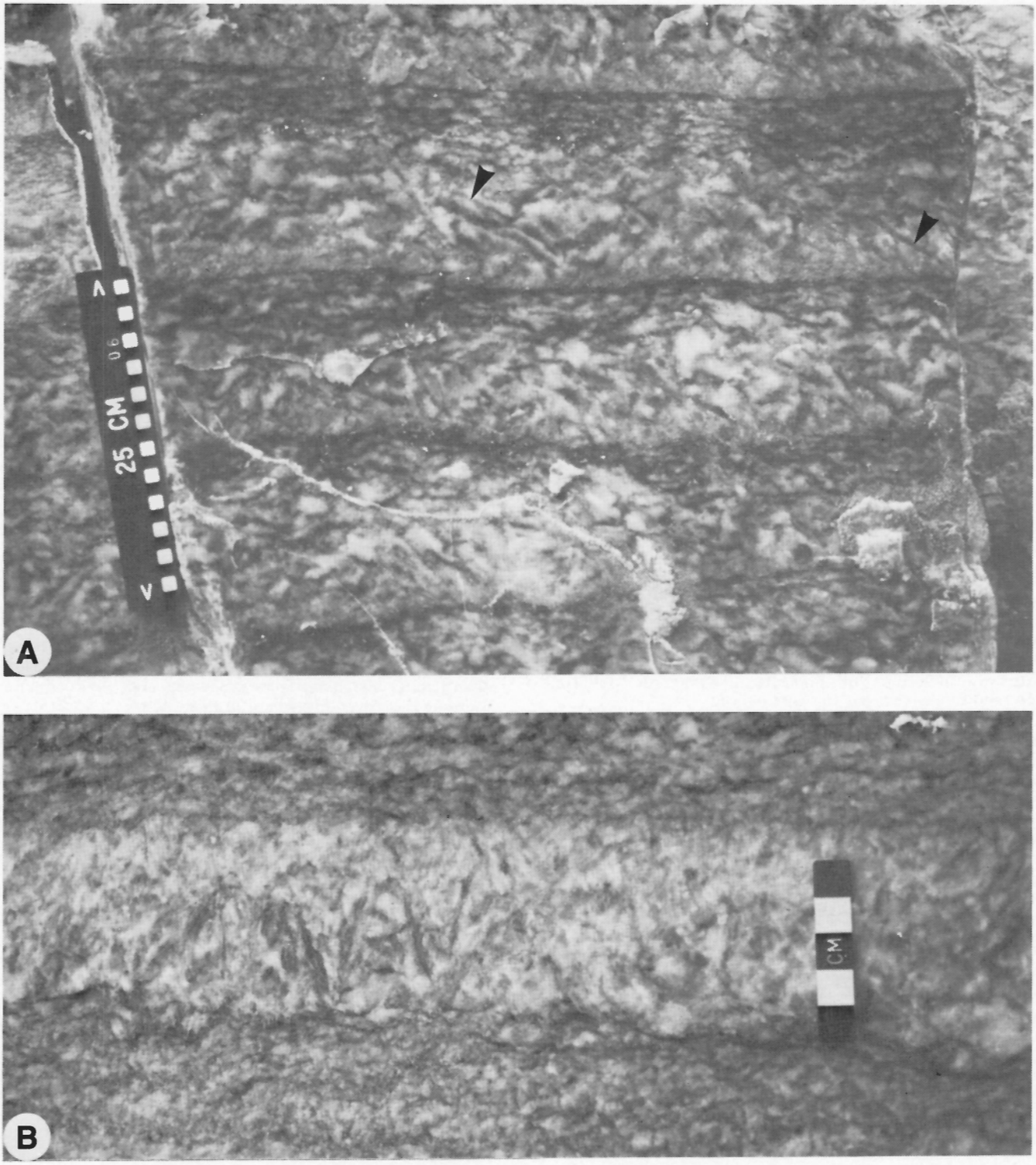
The linear fabrics defined by the inclusions are smaller and more randomly orientated in the lower part of the beds, and become larger and more clearly defined as inclined bladed sets in the upper part.

The inclusion fabrics are interpreted to be pseudomorphic after primary or early diagenetic gypsum crystals. A fabric analog has been illustrated by Hardie and Eugster (1971, Fig. 4, p. 193) from the Miocene Solfifera gypsum beds of Sicily. Further discussion of depositional and diagenetic processes and environments of these beds is given elsewhere in this report.

The upper quarter to third of each bed is composed of darker, mottled, inclusion-rich anhydrite with a crude horizontal fabric (Fig. 15). This fabric is formed by discontinuous lenticular and laterally coalesced nodules of anhydrite, interspaced with irregular stringers of dark (organic?) detritus and inclusion-rich anhydrite. This upper layer is interpreted as a layer of mixed composition (calcium carbonate, calcium sulphate, minor clays and organic detritus) in which sulphate nodules grew during early diagenesis.

Fifty beds, averaging about 12 cm in thickness, were counted in the lower 6 m in one well exposed, bedded anhydrite unit in the type area. Above this interval, beds gradually lose character and regularity, and are replaced higher in the unit by less distinctly bedded, thin-bedded to laminated nodular anhydrite. The uppermost 50 cm of this unit below the next limestone unit is irregularly laminated, with anhydrite layers of laterally coalesced nodules separated by irregular, discontinuous concentrations of dark impurities. In other units bed thickness increases upward to 20 cm about 4 m above the base of the unit. Some thicker beds may contain subrhythms suggesting minor reversals or changes in the depositional trend.

Anhydrite breccia. About 90 m above the base of the Otto Fiord section at van Hauen Pass, an anhydrite lithoclast breccia about 3 m thick is interbedded near the middle of a 15 m thick anhydrite unit (Fig. 16B). This breccia zone is underlain by undisturbed, thick-bedded anhydrite and overlain by undisturbed, thin-bedded and nodular-mosaic anhydrite and thus appears to be synsedimentary. The breccia clasts are angular to subangular, up to 40 cm long, and composed of laminated and nodular mosaic anhydrite. The larger blocks are set in a matrix of jumbled, smaller, angular anhydrite clasts and finer textured anhydrite



A. Closer field view of bedded anhydrite from same locality as Fig. 13. Indistinct linear inclusions of microcrystalline calcite define pseudomorphs after gypsum crystals (arrows) in the lower layers of each rhythmic bed. Note the gradational transition from the crystal pseudomorph layer into the overlying inclusion-rich subnodular layer, and the contrastingly sharp contact at the top of the subnodular layer.

B. Pseudomorphs of upwardly diverging gypsum crystals in anhydrite from an isolated bed in the lower part of Section 2; compare with Miocene bedded gypsum fabrics of Fig. 45.

Figure 14. Crystal pseudomorph fabrics in bedded anhydrite, Section 2.

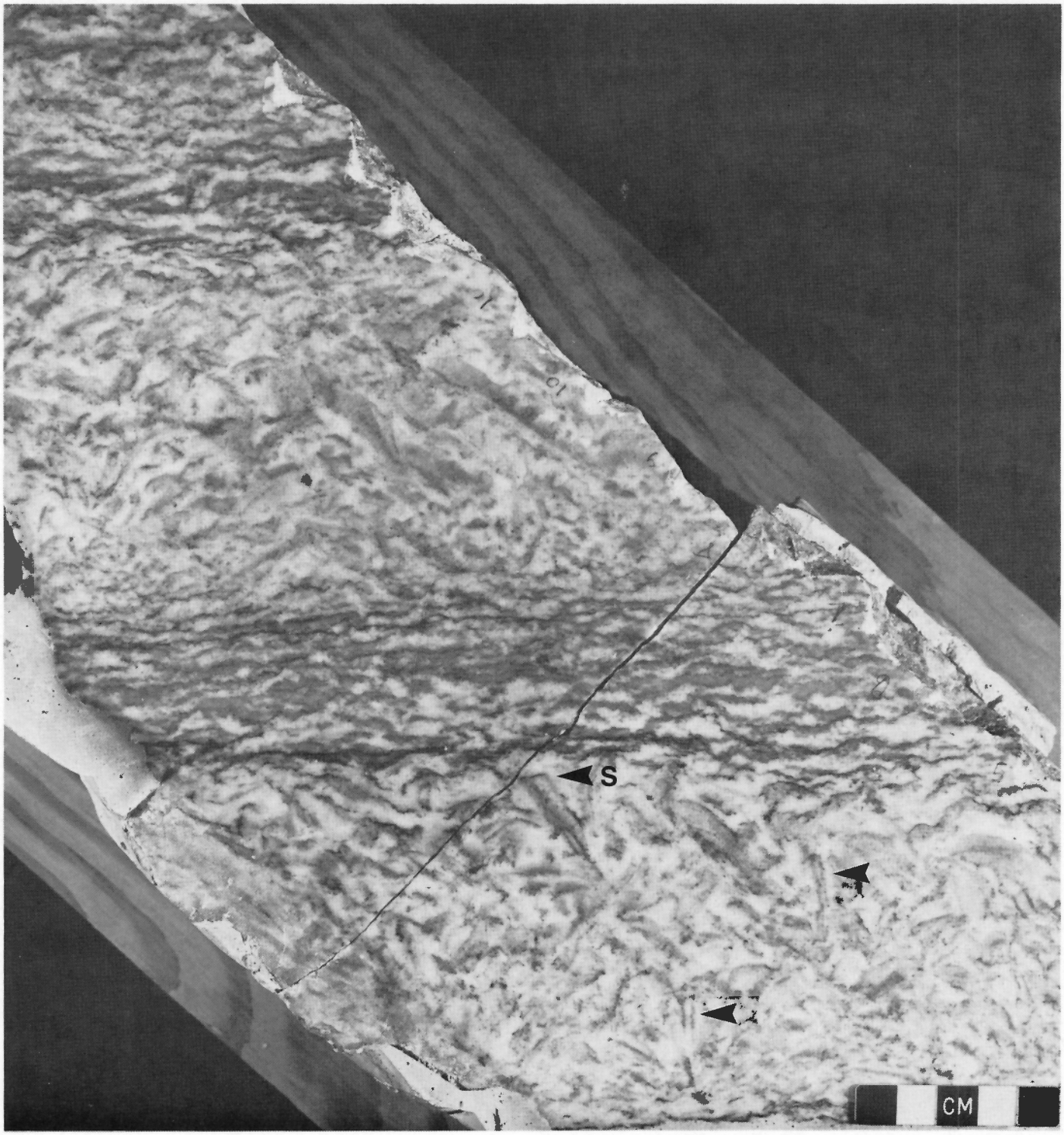
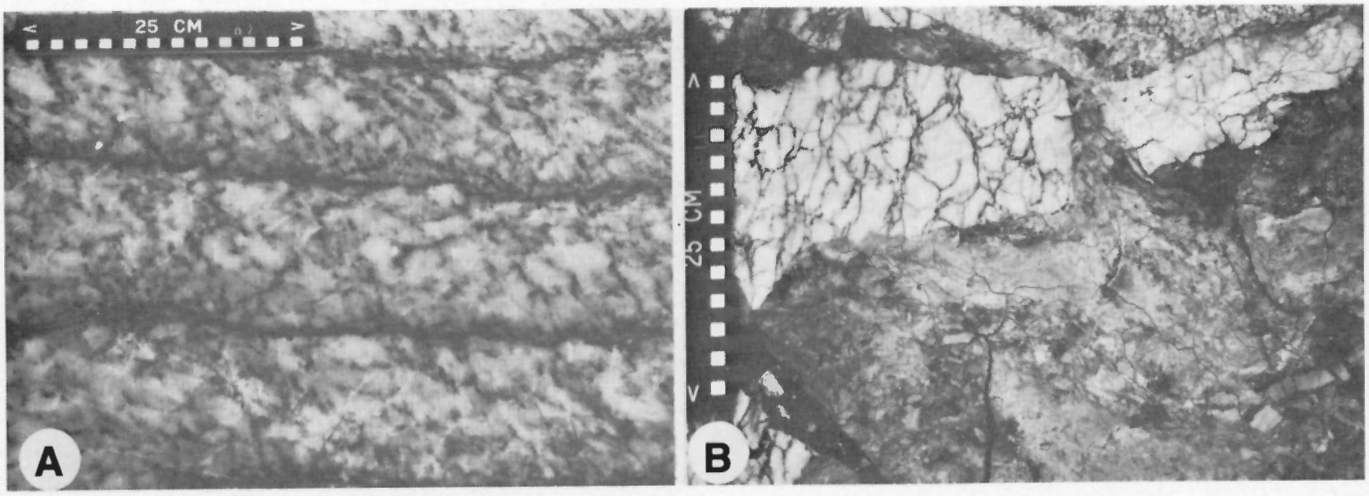


Figure 15. Detail of bedded anhydrite fabrics from lower part of Section 2; polished slab. Pseudomorphs after gypsum crystals (arrows) are defined mainly by microcrystalline calcite inclusions in linear arrangements. Local concentrations of impurities (S) at the top of pseudomorphs in the upper part of the lower layer may record a solution event at the onset of reduction in hypersalinity levels. The upper, inclusion-rich layer probably records a seasonal or other periodic environmental rhythm of reduced salinity and increased carbonate accumulation; the present fabric suggests extensive replacement and displacement of this layer by early diagenetic sulphate.



A. Distinct traces of primary bedding at about 12 cm spacing in anhydrite that otherwise is characterized by nodular mosaic fabric. Inclusions defining bedding are mainly microcrystalline calcite. Nodular mosaic fabric is inferred to be a diagenetic (postburial?) replacement of bedded crystalline gypsum.

B. Breccia of nodular mosaic anhydrite in a sheared and flow-laminated matrix of anhydrite from the mid-upper part of an anhydrite unit in the lower part of Section 2. This breccia is interpreted as a slump or mass flow phenomenon penecontemporaneous with deposition, as anhydrite below and above it are normally bedded and not disrupted.

Figure 16. Fabrics in anhydrite, Otto Fiord Formation.

groundmass. Flowage has sheared the matrix and wrapped it around the large blocks. This breccia probably is a slump or mass-flow deposit, suggesting deposition of the source material on a slope (Schlager and Bolz, 1977).

Calcite laminites. The uppermost anhydrite unit in Section 2 is overlain by 5 m of dark limestone followed by 2 to 5 m of light calcite laminite. The laminite unit is overlain directly by well bedded, cherty and argillaceous carbonate typical of the Hare Fiord Formation. Essentially identical laminites also are found in several other sections where the Otto Fiord Formation is overlain gradationally and apparently conformably by the Hare Fiord Formation.

A description of calcitic laminites is given in the discussion of stratigraphic section 7b, and their sedimentological significance is documented. The sedimentological characteristics of the calcitic laminites in the type area suggest that the top of the Otto Fiord Formation as a depositional unit logically should be placed at the top of the laminite section in which it is present.

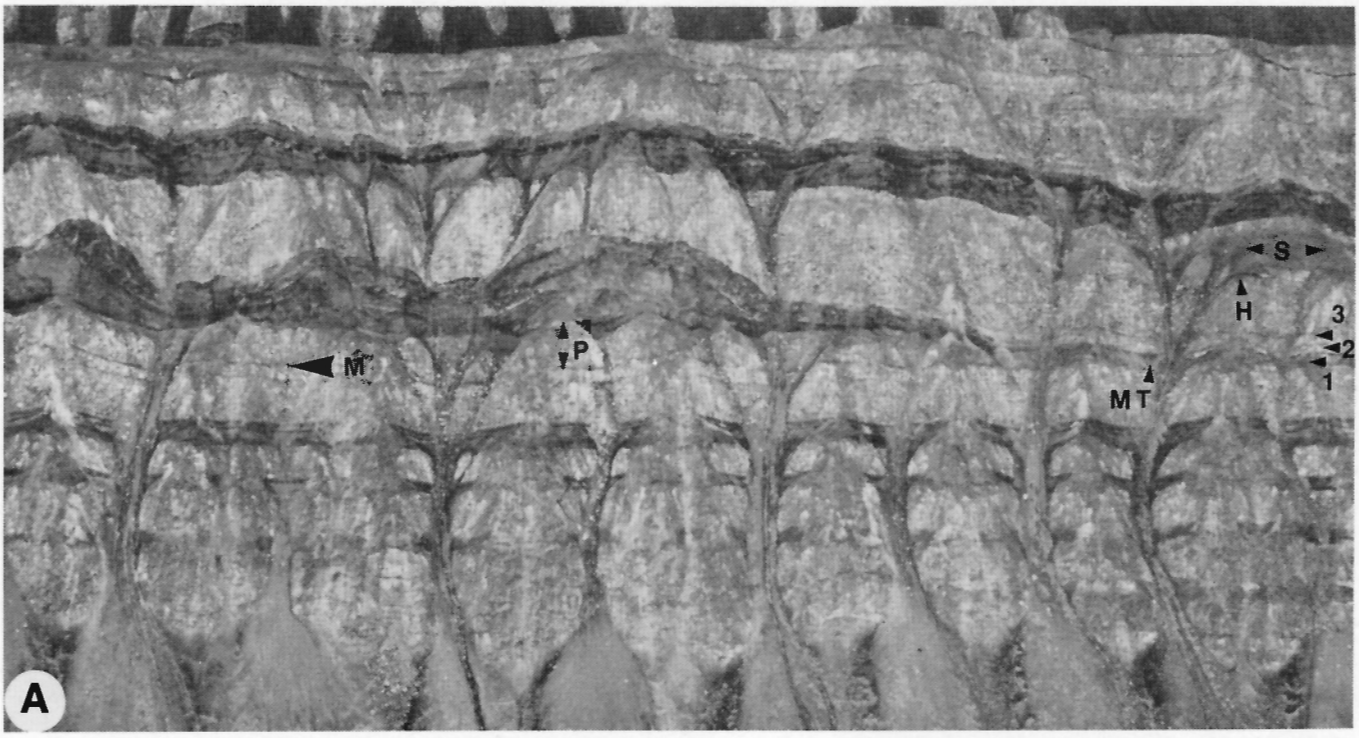
Hydration zones in anhydrite adjacent to limestone units. Many limestone units interbedded with anhydrite have a hydrated zone of gypsum immediately below and, in some places, above the limestone contact. The hydrated zones commonly are 1 m thick and at the exposed surface are composed of recessive granular gypsum. The limestone units apparently acted as aquifers for subsurface water that hydrated the adjacent anhydrite to gypsum. However, as the contemporary Arctic climate is not conducive to hydration of anhydrite (there is little or no hydration of anhydrite at the outcrop surface), and as the permafrost may extend to a depth of 1000 m, this hydration process cannot be related to the present erosional surface and processes. Rather, it appears to be a relict of pre-exposure subsurface water movement, perhaps under milder Tertiary or older climatic regimes. Mossop (1973) has observed similar hydration phenomena in anhydrite interbedded with carbonates in the Baumann Fiord Formation on southern Ellesmere Island.

Limestone mounds

Limestone mounds or reefs occur at several levels within the Otto Fiord Formation in the type area near van Hauen Pass (Fig. 1). About 200 m above the base of the Otto Fiord Formation (Sec. 2), four limestone mounds are exposed within the same limestone marker bed (Fig. 8). The largest is about 30 m thick and 350 m long in the plane of exposure. The mounds have flat bases and irregular upper profiles (Fig. 17). Variations in weathering of internal components reveal that the mounds are composite buildups of three or more vertically stacked and thickened limestone units forming each mound (Fig. 17). Apparent depositional dips at the flanks of the mounds and of internal components range from 25 to 30 degrees. In the intermound area each of the main lenticular limestone units of the mound is represented by thinner, planar beds of limestone separated by units of anhydrite (Fig. 17A).

The mounds are built on elevated blocks or platforms of anhydrite, which have relief of about 14 m above the surface of anhydrite in the intermound areas (Fig. 17A). These platforms are interpreted to be erosional residuals formed during a period of subaerial exposure. The geometry of the irregular surface is defined best by using the thick limestone unit that lies about 30 m below the mounds as a horizontal datum. Two thin limestone marker beds overlain by about 15 m of anhydrite below the centre of the mounds appear to be truncated at the flanks of the anhydrite platform. This apparent truncation is the main criterion for an erosional origin for the anhydrite platforms. The preservation of regular, parallel bedding in at least part of the anhydrite of the platform unit supports this interpretation rather than the alternative concept of mechanical construction of the platforms as banks of detrital sulphate on a more or less horizontal substrate.

Carbonate sediment accumulation after erosion of the underlying anhydrite was controlled to a large degree by the geometry of the eroded surface. Mound building was localized on the platformlike surface of the erosional residuals of anhydrite. Thickest accumulations appear to



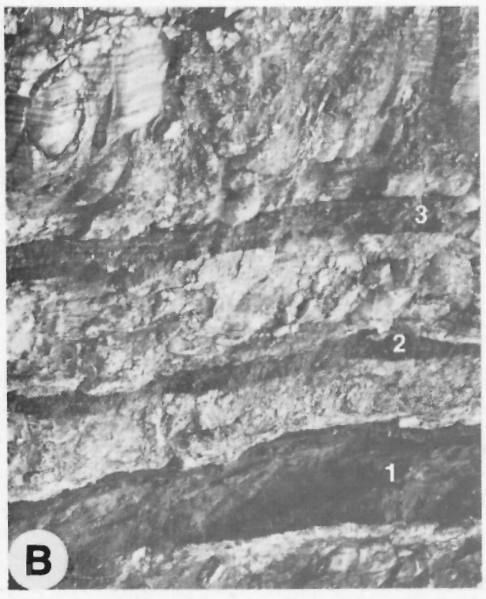


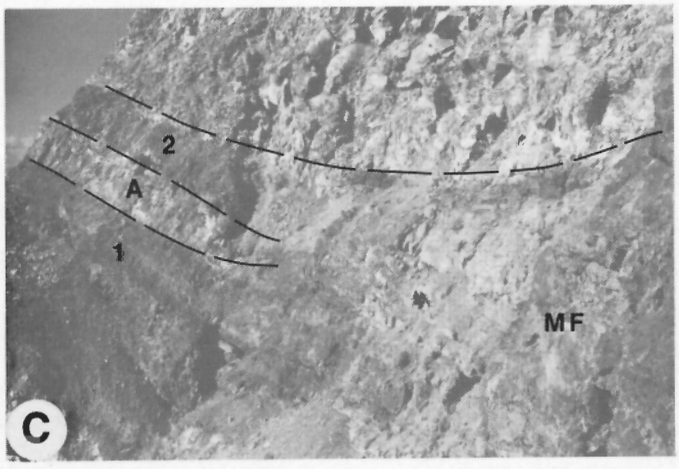
A. General view of mound at far left (west) of Fig. 8. Mound is about 30 m thick. Note strong cyclicity of carbonate and anhydrite units in the formation, and the increase in thickness of successive units. The upper of two thin limestone marker beds (M) below the mound is truncated below the extreme flank of the mound base (MT). The mound is built on a platform of anhydrite (P) about 14 m thick. Internal bedding, some steeply dipping, is visible within the mound. A thick basal limestone unit (1) and two thinner limestone beds (2, 3), with interbedded anhydrite separating them, form the lateral equivalents of the mound in the intermound setting. A discontinuous hematite (H) and a sandstone unit (S) are enclosed in the thick anhydrite unit in the intermound area. GSC 199245.

B. General view of algal mound at far right of Fig. 8. GSC 199253.

Figure 17. Lower Pennsylvanian algal mounds in the Otto Fiord Formation near van Hauen Pass.







A. Oblique aerial view of the eastern flank of the algal mound illustrated in Fig. 17A. Limestone unit 1 is continuous with the lower part of the mound (M); limestone beds 2 and 3 are separated from unit 1 and each other by anhydrite units in the intermound setting but overlap westward (left) onto the flanks of the mound. GSC 199259.

B. Closer field view of intermound limestone units 1, 2 and 3 of 18A, with interbedded anhydrite units that pinch out at the mound flanks. Note the bedding in the overlying anhydrite unit in upper left. GSC 199250.

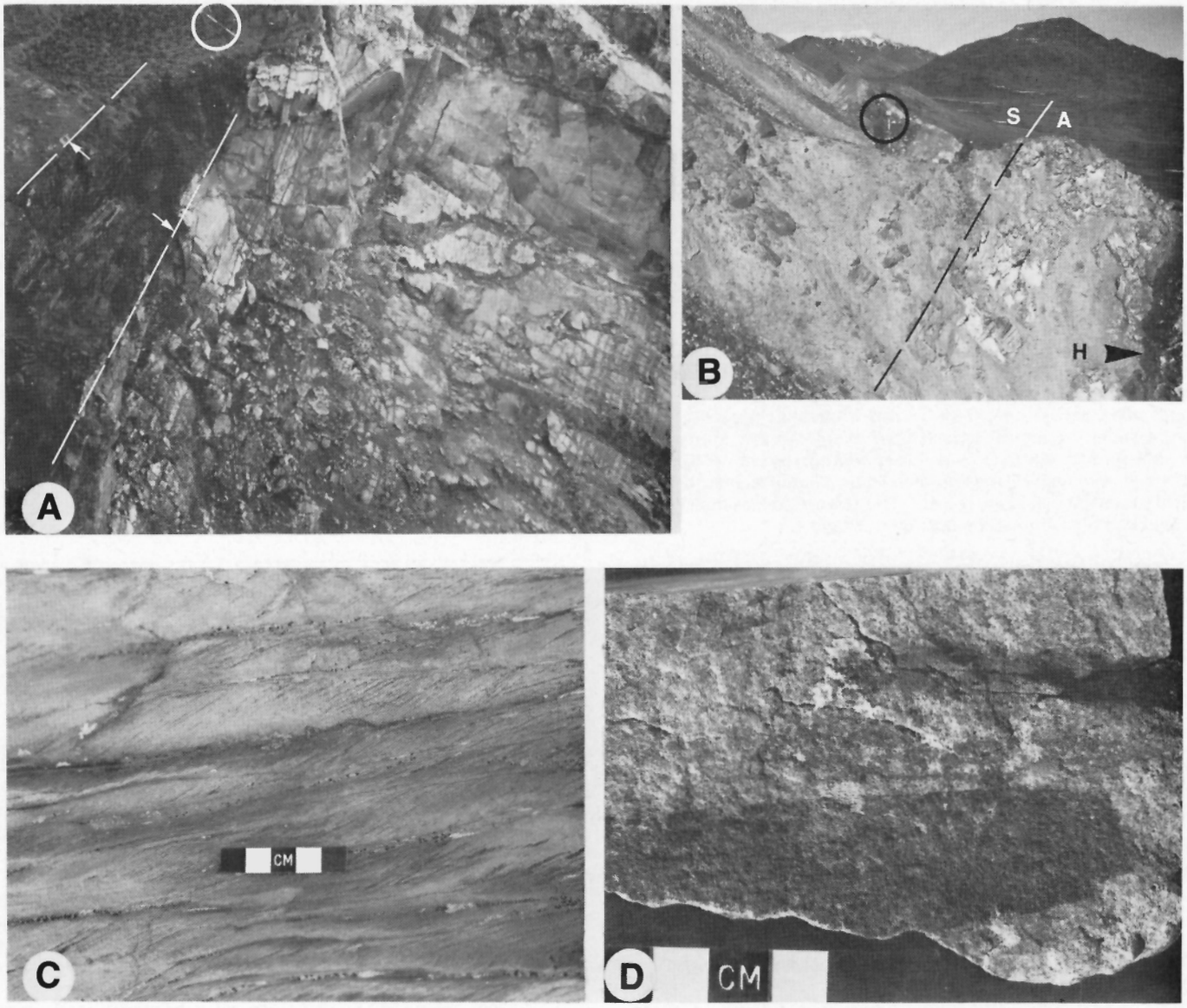
C. Field view of the western flank of a mound (M) showing the pinch-out of an anhydrite unit between the basal intermound limestone unit (1) and the second intermound limestone bed (2). GSC 199254.

Figure 18. Detail of mound flank and intermound relationships.

overlie slight topographic highs on the platform. Rapid thinning of limestone flank beds and the first appearance of off-mound anhydrite interbeds occur above the flanks of the anhydrite platform (Figs. 17, 18).

In the off-mound or intermound setting, the equivalent section is composed of three limestone beds that are lateral extensions of thicker accretionary units in the mound (Fig. 18). The limestone beds are separated by anhydrite units that pinch out between the limestone beds at the mound flanks, with the upper anhydrite units onlapping farther and higher onto the flanks.

The mounds are buried by a thick unit of anhydrite that has a maximum thickness in off-mound settings of about 50 m. This anhydrite in turn is overlain by a prominent shale and limestone unit that is 20 m or more thick. Wardlaw and Christie (1975, p. 160) found that this limestone unit is up to 5 m thicker where it overlies the limestone mounds.



A. Hematitic anhydrite unit (arrows) about 3 m thick overlying bedded anhydrite in the intermound area between algal mounds. Five-foot (1.5 m) rod (circle) provides scale. GSC 199251.

B. Sandstone unit interbedded in anhydrite in the intermound area. Sandstone (S) overlies 5 to 6 m of anhydrite (A) that in turn overlies hematite bed (H). Eastern end of van Hauen Pass outcrop belt. GSC 199255.

C. Ripple-laminated sets in sandstone; some sets of climbing ripples. Same locality as 19B. GSC 199260.

D. Plant stem fragments on bedding plane in sandstone from same locality as 19B. GSC 199256.

Figure 19. Discontinuous hematite beds and sandstone units in intermound areas, van Hauen Pass.

About 30 m above the uppermost limestone bed in the off-mound setting, a discontinuous tabular sandstone unit appears in the section (Figs. 17, 19B). This unit is about 18 m thick and is composed mainly of crossbedded and ripple-laminated calcite-cemented quartz sand (Fig. 19C). The sandstone also contains carbonized plant fragments (Fig. 19D) and marine skeletal fragments. Green shale and siltstone beds are common in the lower part of the unit.

Close to the flanks of the mounds, the sandstone unit thins out abruptly and is represented over the flanks of the mounds by only a thin, green argillaceous marker bed, which may be traced from the top of the sandstone unit some distance through the anhydrite above the flanks and tops of the mounds. Thinning of the sandstone unit near the mound flanks appears to proceed from the base of the unit upward, consistent with the geometry expected for a sedimentary infill of a depression between adjacent mounds. The irregular surface defined by the sandstone bodies and the thin green marker beds over the mounds may correspond to the depositional topography on the underlying

anhydrite. Significantly, this would require deposition of anhydrite in water depths of at least several tens of metres and thus is consistent with a submarine depositional model.

The discontinuous sandstone bodies are underlain by a very prominent but discontinuous bed of red hematite and hematitic anhydrite (Figs. 17, 19A, B). This bed is 2 to 3 m thick, and is separated from the base of the overlying sandstone by about 7 m of crumbly-weathering anhydrite. The hematite bed thins and finally pinches out completely off the flanks of the limestone mounds, thus conforming to the same distributional pattern as the sandstone unit. This bed is considered to record a period of accumulation and preservation of iron oxides, probably in hydrous mineral form originally, on the deep floor of the intermound depressions. Alternatively it may represent a residue after solution of salt.

The discontinuous hematite and sandstone bodies are overlain by anhydrite several metres thick, which is continuous across the tops of the limestone mounds. Thus the complex series of depositional events in the intermound or off-mound setting was terminated by a return to sulphate evaporite accumulation in a continuous blanket with little depositional relief. The entire succession finally is overlain by a limestone unit 20 m thick.

Internal lithological succession. The limestone mounds in the Otto Fiord Formation at van Hauen Pass are characterized internally by a vertical lithological sequence from crinoid-rich skeletal wackestones at the base, to a main algal mound phase, and are capped by units of skeletal wackestone with a varied marine biota (Fig. 20). In more detail, the sequence from the base upward is as follows:

1. Contact with underlying anhydrite (or granular gypsum in hydrated zone within 1 m of contact) is sharp, with some irregularities apparently caused by local infilling of depressions on the eroded surface by carbonate sediment.

2. The basal unit of the mound is formed by 2 to 5 m of well bedded, dark limestone, in part argillaceous and recessive; the unit thins toward the flanks of the mound and apparently is continuous with the lowest and thickest of the three correlative limestone beds in the intermound setting (Fig. 18). This basal limestone unit contains a rich marine biota dominated by crinoids (pelmatozoans); other common components are brachiopods, ammonoids, bryozoans, gastropods, ostracodes and foraminifers. Darker micritic sediment lumps, perhaps of fecal origin, are common. Matrix is bioturbated spicular, pelletoidal wackestone. Many skeletal clasts are bored. In beds higher in the unit, sediments were reworked by currents into poorly graded beds of crinoid grainstone and crinoid-bryozoan packstones with traces of quartz silt and sand. Uppermost beds include finer grained, spicule- and calcisphere-rich bioturbated wackestone with large, thick-walled ostracodes and the first observed algae.

3. The main topographic expression of the mounds is the result of buildup by branching tubular algae. The dominant varieties are beresellid and kamaenid types that have some affinities with dasycladacean algae but may be unrelated to that family (Mamet and Rudloff, 1972). Preserved fabrics range from open framestones of algal mesh with patches of pelletoidal wackestone matrix, to aggregates of broken algal clasts in grainstone and packstone textures. Other skeletal components in the algal mound facies are brachiopods, fenestellid bryozoans, crinoids, encrusting foraminifers, ostracodes, and several other types of algae including *Komia* (or *Ungdarella*). Internal hard grounds also may be present in these rocks. Where the primary skeletal frame is preserved, the algal branches commonly are encrusted by a fringe of isopachous, inclusion-rich fibrous calcite. This calcite is similar to fibrous calcite cements in Pennsylvanian and Permian reef rocks higher in the Sverdrup Basin succession, which are interpreted to be of submarine origin and probably neomorphic after magnesian calcite (Davies, 1977b).

Another characteristic of the algal mound facies is the predominance in some units of dark, fan-shaped masses of radial-fibrous calcite (Fig. 20; also Wardlaw and Christie, 1975, Figs. 10 - 13). On exposed rock faces, the dark radial-fibrous calcite contrasts with lighter coloured, pore-filling diagenetic calcite and dolomite, producing a distinctive mottled fabric (Fig. 20). In thin section, the radial-fibrous calcite is seen to be pseudomorphic after radiating cones of a crystalline precursor that grew in primary cavities within the algal frame, in other cavities perhaps of solutional origin and, in part, by replacement. The radial-fibrous calcite fabric is postdated by zoned iron-rich calcite spar and zoned dolomite crystals, and pore-filling and replacive quartz (chalcedony), anhydrite and fluorite.

Essentially identical pore-filling and replacive growths of botryoidal and spherulitic calcite have been found in Middle Pennsylvanian and Lower Permian reef rocks in the Sverdrup Basin; these fabrics are interpreted to be neomorphic after aragonite of submarine origin (Davies, 1977b).

Figure 20. Characteristic lithologies of Lower Pennsylvanian crinoid-algal mounds in the Otto Fiord Formation near van Hauen Pass.

A. Crinoid-rich skeletal wackestone and packstone from the lower beds of an algal mound. Brachiopods, ammonoids (A) and bryozoans are other obvious components of this marinelike limestone.

B. Similar to 20A, but mechanically sorted. Crinoid-bryozoan packstone from the lower beds of an algal mound.

C. Field view of algal mound core showing distinctive mottled fabric composed of dark masses of recrystallized radial-fibrous calcite and lighter patches of later diagenetic calcite and dolomite. Note subhorizontal trend in fabric that parallels primary bedding.

D. Algal frame in mound core with recrystallized fringe of fibrous cement (F), probably of early submarine origin, followed by later diagenetic calcite spar. Insert, at same scale, shows detail of some well preserved beresellid and kamaenid algae from the mound core.

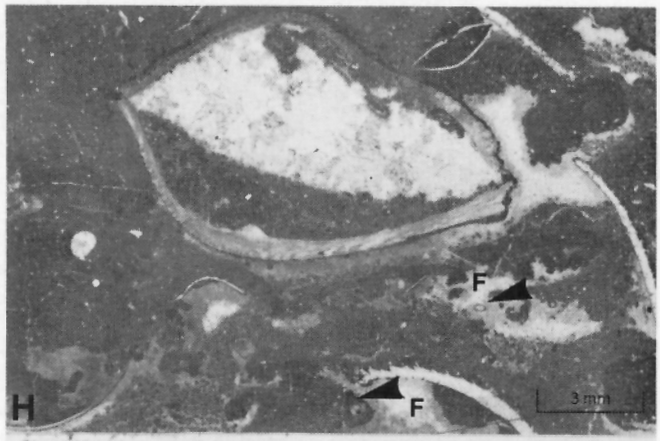
E. Recrystallized fans and cones of radial-fibrous cement fabrics (R) in the main algal mound-building core. Algae (A) form an indistinct organic frame. Calcite-replaced radial-fibrous cements probably were emplaced originally as aragonite.

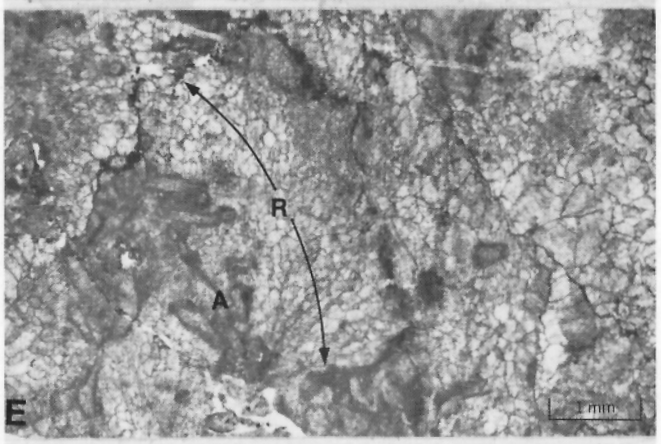
F. Algal framework in mound core. Patches of primary pelletoidal sediment are partly replaced by fans of radial-fibrous calcite (R).

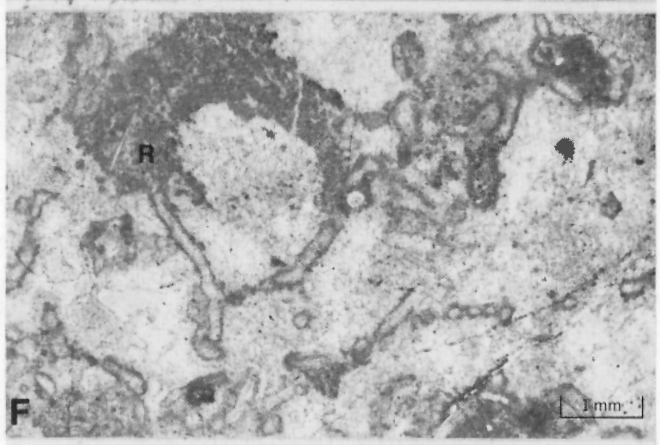
G. Articulated brachiopods in a micritic matrix of limestone from the upper beds of an algal mound. Note the preservation of delicate spines of the brachiopod at right centre that are embedded in the matrix; this type of preservation attests to the autochthonous nature of this fauna.

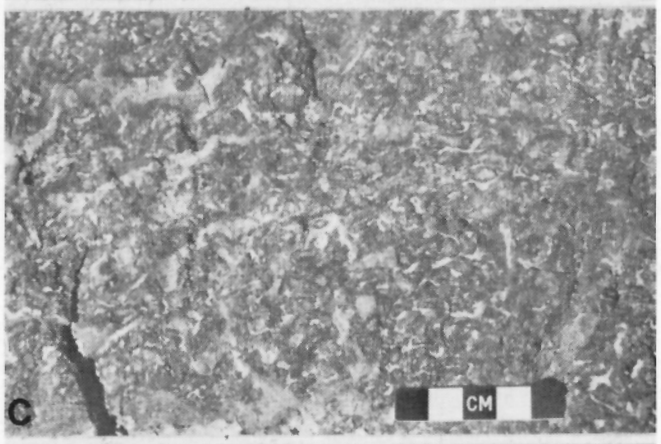
H. Brachiopod in limestone near the top of an algal mound with geopetal infilling of valves by pelletoidal sediment, and early (submarine) cementation by fibrous calcite (F) in primary shelter pores and dewatering compactional cavities.

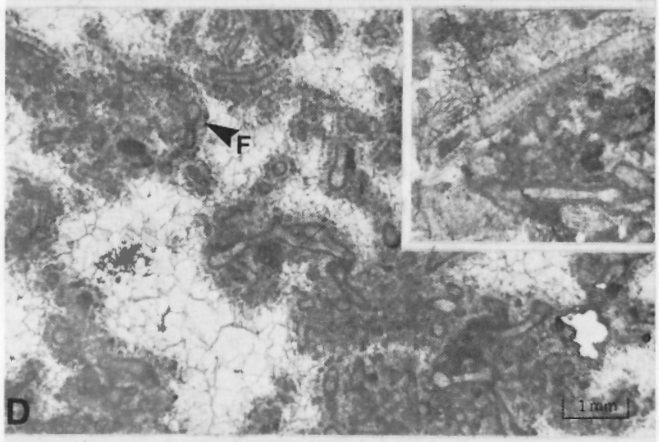

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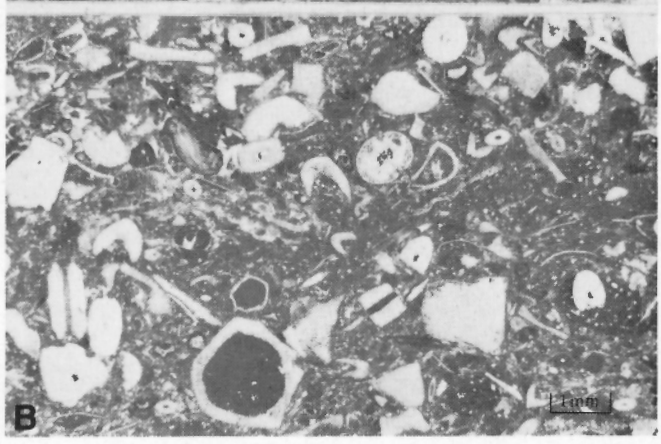

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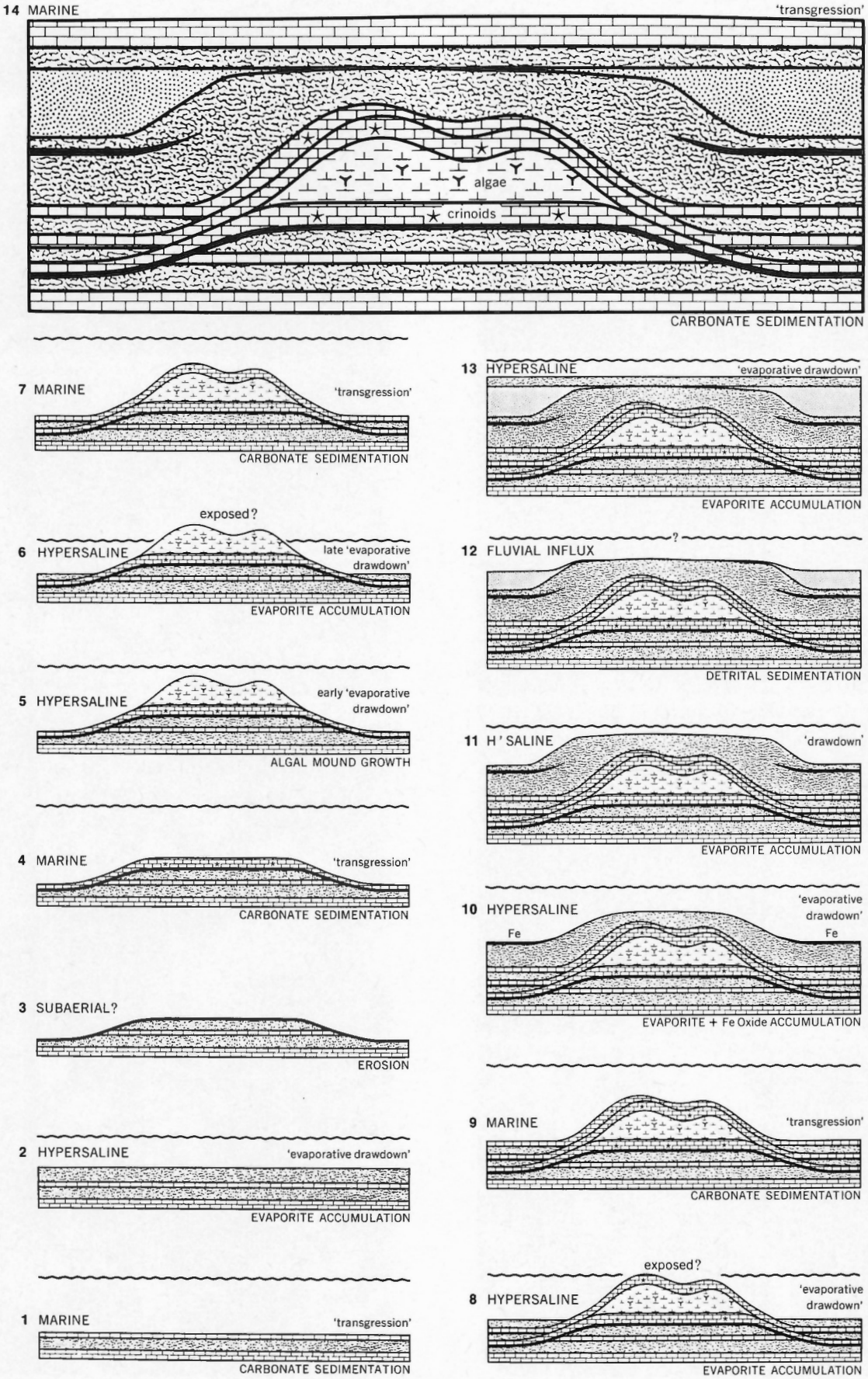


Figure 21. Interpretation of the sequence of events culminating in the formation of the Lower Pennsylvanian crinoid-algal mounds and adjacent lithological units near van Hauen Pass. Stages 1 to 13 include at least four 'transgressive-regressive' cycles that correspond to marine ingression followed by evaporative drawdown of brine sea level. Stage 14 records the beginning of a fifth cycle.

Large selenite (gypsum) crystals occur at several horizons within the algal mound units. These are considered to be of very late diagenetic origin, postdating unloading of the Otto Fiord Formation by erosion in Late Cretaceous and Tertiary time (see Davies, 1976, for summary of burial history and postburial diagenesis of overlying rocks). However, the distribution of the selenite crystals may coincide with original sulphate-rich depositional discontinuities within the mound mass.

4. The algal mound facies is overlain by several units, 3 to 4 m thick, of dark, argillaceous skeletal wackestone containing a varied marine biota. These units drape across the tops of the mounds and thin out at the flanks. They apparently are continuous with the two thin limestone beds in the intermound setting (Fig. 18). Common skeletal clasts include crinoids, large brachiopods, sponges and sponge spicules, ammonoids, bryozoans and foraminiferids. The matrix contains abundant pelletoids and spicules. Whole sponges and brachiopod valves with delicate spines still intact and embedded in the matrix attest to the autochthonous nature of this marine biota and of the sediment (Fig. 20). Irregular fenestrae and skeletal shelter pores in these rocks commonly are rimmed by fibrous calcite cement, apparently of submarine origin. Some fenestrae, however, lack this cement lining; they may have been formed by subaerial (vadose) processes. Remaining primary pores in this facies are filled by later diagenetic zoned ferroan calcite and zoned dolomite. Significantly, radial-fibrous calcite fabrics of the type that characterize the algal mound facies are absent from these capping beds.

The uppermost bedding plane of the capping skeletal limestone is very dark and pyritic, and on its surface are many large brachiopods, articulated segments of crinoid columnals, ammonoids, and other clasts of marine organisms. This surface probably represents a marine hardground. It is overlain directly by anhydrite or by local hydrated zone of gypsum. This contact is sharp, and lacks the interlaminated limestone and anhydrite characteristic of normally bedded limestone-anhydrite cycles in this formation.

Evolution of the algal mounds. The crinoid-algal mounds in the Otto Fiord Formation at van Hauen Pass are interpreted as having formed by the following sequence of events keyed to the numbered sequence in Figure 21:

1 - 3. Rhythmic deposition of marine limestones and gypsum was interrupted by a period of subaerial solution and erosion, leaving an irregular surface of elevated 'platforms' of residual gypsum and depressions 14 to 15 m deep. The alternative of submarine erosion cannot be dismissed, but is unlikely because of the lack of irregular erosion in other cycles that apparently were exposed periodically to normal marine water. However, an influx of hyposaline water could have produced this erosional-solutional event in a subaqueous environment.

4. The unconformity surface was transgressed by seawater and a varied marine biota was established on all substrates; clay and silt introduced by the sea or reworked from the erosional surface were incorporated into the basal carbonate layers. Organic productivity and hence carbonate sediment accumulation rates were higher on the elevated substrates, probably through a combination of advantageous photic levels, favourable nutrient supply and lower turbidity. Crinoid communities colonized the submarine highs and increased the skeletal carbonate supply. In the depressions, crinoids were less common, rates of production and accumulation were lower, and thinner beds of sediment accumulated.

5. The transgressive maximum with its normal marine salinities and biota was followed by a progressive increase in salinity and depression of regional sea levels under conditions of net evaporation ('evaporative drawdown': Maiklem, 1971). The controls for this rhythmic pattern probably were mainly tectonic, reflecting continuing evolution of the riftlike Sverdrup depression. As salinities increased, the varied marine biota on the seafloor was succeeded initially by an increasingly restricted, euryhaline biota of sponges, ostracodes and calcisphere-producing algae, and finally by beresellid and kamaenid algae. With the high nutrient supply of the hypersaline water mass, the algae flourished and constructed the main mound buildup. The calcified algal mesh in part was cemented into a rigid framework by submarine cements.

Pervasive cementation and replacement in primary pores and sediments by spherulitic and botryoidal aragonite was confined to the algal mound phase, suggesting strong contemporaneous environmental and substrate controls. This early diagenetic process may have resulted from the internal circulation of hypersaline water derived from the enclosing water body and sediments. However, it is possible that this process was contemporaneous with Event 6 (concurrent with sulphate precipitation), as hypersaline water from which gypsum was precipitated should be enriched in Mg relative to Ca, a factor favourable to aragonite precipitation at the expense of calcite (Berner, 1975).

6. With continued net evaporation, skeletal carbonate production was inhibited, and eventually gypsum supersaturation levels were reached; gypsum precipitated and grew as bedded crystal aggregates in the intermound areas and at least part way up the flanks of the mounds. It is possible that at the evaporitic maximum the tops of the algal mounds were exposed, although there is no diagnostic evidence in the mound rocks for such an event.

7. This evaporitic cycle was terminated by renewed marine ingression and deepening. Some gypsum deposits higher on the flanks of the mound may have been eroded or dissolved. The mound surface and the intermound substrates again were invaded by marine organisms. Organic productivity was greater on the elevated surface of the mounds, and thus a thicker sediment lens accumulated there. A thin bed of marine limestone was deposited in the intermound areas.

8. With progressive increase in salinity, gypsum saturation was reached and gypsum once again accumulated in the off-mound setting. The absence of an intermediate algal mound component in this cycle, in contrast with Event 5, may have been caused by differences in water depth, salinity or rates of sea level change, or probably a combination of these factors.

9. Event 7 was repeated, producing the second of the thin intermound limestone beds and another lenticular mound unit of marine limestone.

10. The carbonate subcycle of Event 9 was followed by deposition of a thick sequence of bedded gypsum, which apparently draped over and buried the mounds. The thickness of the sulphate and the topography of the mounds indicate that the original water depth for this depositional cycle exceeded 30 m.

After deposition of about 30 m of gypsum in the intermound areas and burial of the mound by many metres of gypsum, an event occurred that resulted in the accumulation of 2 to 3 m of iron oxides and gypsum, later converted to hematite and hematite-stained anhydrite. This accumulation was confined to the intermound depressions, but a thin trace of hematite may be followed and for a short distance toward and above flanks of the mound. The source of the iron oxides is open to conjecture. Hematite staining of evaporite-associated basinal

carbonates of the Silurian Niagara Formation in the Michigan Basin is attributed to the early loss of detrital organic matter at the sediment-water interface under conditions of very slow 'basinal' sedimentation, resulting in inhibition of bacterial reduction of sedimented hydrous iron oxides during subsequent burial; later dehydration to red iron oxides completed the staining cycle (Mesolella et al., 1974). In the Otto Fiord rocks, red iron oxide staining also is found in limestone and argillaceous limestone units interbedded with anhydrite, but these are minor occurrences compared with the discontinuous hematite bed in the intermound setting under discussion. Wardlaw and Christie (1975, p. 166) suggested that erosion of ferruginous sandstones at the basin edge may have provided a source for iron and later for the overlying sandstone. The lack of detrital grains in the hematite unit precludes the movement of strong traction currents into this area from a fluvial source, yet an apparent increase in opaque argillaceous and organic impurities (visual estimate) in the hematitic anhydrite would support a surface density outflow of low-salinity, clay-rich water into and across the hypersaline water mass at this time. Iron oxides and other iron complexes introduced in this surface stratum may have precipitated from within the mixing zone between the two water masses, and have been preserved preferentially on the floors of the deep depressions.

An alternative that must be considered is that the iron enrichment is a residue of solution of salt and perhaps potash-grade evaporite minerals. However, there is no other structural or textural evidence to support this conjecture.

11. Iron oxide accumulation was accompanied and followed by continued gypsum precipitation and crystal growth, while maintaining the irregular depositional topography of the buried mounds and depressions.

12. A major influx of detrital sediment swept into the hypersaline basin and infilled the partly gypsum-filled topographic depressions between the gypsum-covered mounds. Skeletal grains in the sandstone indicate reworking of contemporaneous or earlier marine sediments at the margin of the hypersaline basin, and large plant fragments in the same sands indicate a source from a vegetated landmass. The sedimentary structures in the sands record transport from the north or northeast. Basinward progradation of the detrital wedge is indicated by the grain size, which coarsens upward from shale and silt to medium-grained sand. The detrital sediments eventually filled the depressions and reduced the submarine relief to a more or less planar surface, with a thin trace of green shale overlying the slight elevation above the former mounds.

This pulse of detrital sediment represents a phase of renewed tectonic uplift and basin subsidence perhaps to the east and northeast, where similar but thicker sandstone units are interbedded with Otto Fiord anhydrite close to the eastern margin of the basin. Progressive but episodic deepening of the Sverdrup depression through this time, consistent with this interpretation, is recorded by the thickening of successive anhydrite and limestone units, and the increase in marine biotic components in limestones higher in the section.

13. Deposition of the detrital sediments and levelling of the submarine topography was followed by accumulation of a blanket of gypsum of more or less equal thickness across the entire area.

14. The termination of the evaporitic subcycle and the initiation of another major cycle was marked by deposition of the next thick and continuous skeletal limestone unit, with slight thickening over the site of the buried mounds, indicating that a slight topographic expression of the mounds still existed on the sea floor.

Carbonate breccias

At the eastern end of the outcrop belt east of van Hauen Pass, the upper units of the Otto Fiord Formation are truncated over a distance of about 2 km by a broad, apparently concave-upward structure, which is partly filled with carbonate breccia. The structure removed at least 120 m, and perhaps as much as 140 m, of the upper Otto Fiord section, which is composed mainly of thick units of anhydrite with less than 20 per cent of thinner carbonate and shale interbeds (Fig. 8).

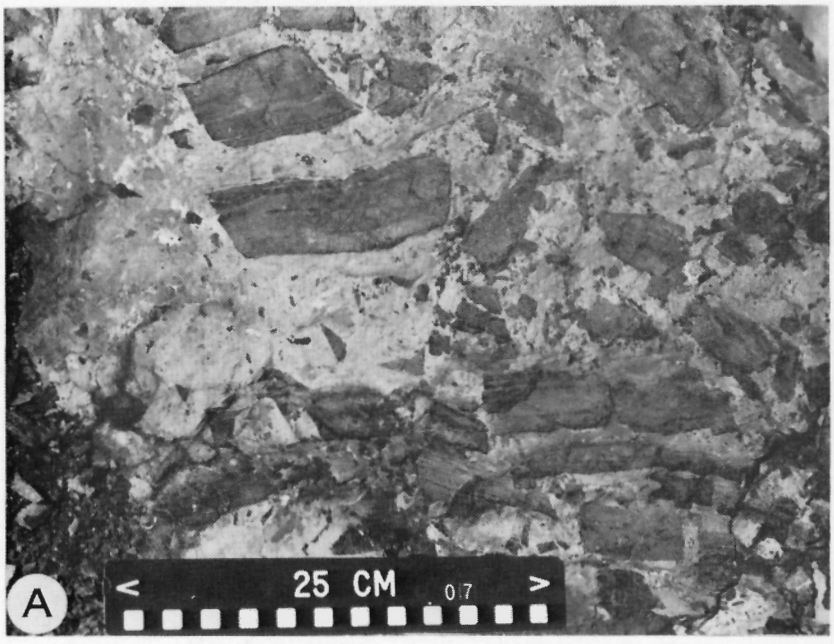
This breccia is of more than local significance, since similar carbonate breccias with a crystalline dolomite matrix are found in many other areas either in close association with evaporites or where evaporites may have been present. Commonly, lead and zinc and other types of mineralization occur within these dolomitic breccias; examples in northern Canada include Pine Point, Robb Lake and the Polaris orebody. Mineralization of this type has not been found in the Otto Fiord breccias. Generally, the spatial relationships between this type of breccia and evaporites cannot be seen or documented; in the example described here for the Otto Fiord Formation, a direct relationship between solution of evaporites and collapse of overlying strata may be demonstrated and, further, it is apparent that these processes occurred after burial of the evaporites.

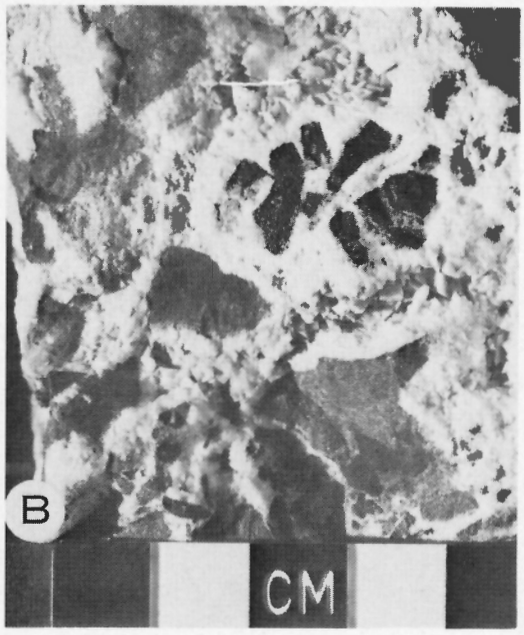
The breccia is composed of angular clasts of dark limestone and dolomite set in a matrix that varies from dolomitized carbonate detritus to coarsely crystalline white dolomite (Fig. 22). The detrital matrix contains a polymictic assemblage of small lithoclasts, some of which are totally replaced by very finely crystalline dolomite, and others that are unaltered limestone. Isolated bioclasts including crinoid columnals and fusulinacean foraminifers also occur in this type of matrix; these bioclasts are interpreted to be residuals of fracturing and solution of more poorly indurated, fossiliferous calcareous shale or argillaceous limestone, rather than having any environmental significance in the brecciation process.

The coarsely crystalline white dolomite is zoned, with an inner iron-poor core and a zoned iron-rich outer phase. This sequence is the reverse of that of iron-zoned dolomite in reefal carbonate rocks of Pennsylvanian and Early Permian age elsewhere on Ellesmere Island (Davies, 1976; Davies et al., in prep.). The white dolomite is postdated by a minor amount of final pore-filling, low-iron calcite. Small fractures that cut these breccias are filled either with ferroan dolomite or calcite; where they intersect, the calcite-filled fractures displace the dolomite-filled fractures.

Carbonate blocks and clasts in the lower part of the breccia unit are similar in composition to limestone found interbedded in the Otto Fiord Formation. However, higher in the breccia pile, it is obvious that a major part of the breccia is derived from lower units of the overlying Hare Fiord Formation. Narrow chimneys or solution collapse conduits extend upward from the breccia pile into otherwise undisturbed Hare Fiord limestone and bedded chert. Coarse white dolomite and dolomitized breccia clasts are found in these vertical structures, some of which extend for at least 50 m and probably much farther upward into the Hare Fiord Formation.

If it is assumed that the entire breccia mass, including the vertical extensions into the Hare Fiord Formation, was formed by the same solution-collapse process, and more or less synchronously, then pre-Hare Fiord solution of the Otto Fiord evaporites is excluded as a viable process. Thus, the solution of the evaporites in this area appears to have been a subsurface phenomenon that occurred, at the earliest, after deposition and induration of the overlying







A. Partly dolomitized carbonate breccia at a stratigraphic level corresponding with the uppermost Otto Fiord-lowermost Hare Fiord formations. Matrix is dolomitized carbonate detritus and solution residuum.

B. Coarsely crystalline white dolomite cement between dolomitized carbonate clasts from a solution-collapse breccia pipe extending upward through the Hare Fiord Formation.

C. Polymictic breccia grading to right into undisturbed bedded carbonate.

Figure 22. Carbonate solution-collapse breccias above the truncated upper Otto Fiord Formation, eastern end of van Hauen Pass exposure belt.

Hare Fiord Formation. The source of water for the solution event, or the control of its localized effect, is open to conjecture.

A possible early source in a postburial setting is the water released during dehydration of gypsum to anhydrite. This should happen at a depth of 1000 m, depending on geothermal gradients and other factors. Calculations by Mossop (1973, Fig. 78) show that a unit volume of gypsum should release about half of its volume as water during conversion to the denser anhydrite polymorph. During the subsurface conversion of a thick deposit of gypsum to anhydrite, assuming that most of the Otto Fiord sulphate evaporites were deposited as gypsum, there must have been the potential for the release of a very large volume of water. If movement of this water was localized by early fracture systems and if solution of some of the sulphate occurred under the subsurface flow regime, the present geometry of the post - Otto Fiord breccia could have been achieved.

An alternative to this relatively early solutional process is solution and collapse much later related to migration of subsurface water along fracture systems that intersected the Otto Fiord Formation. The regional fracture pattern in the general area of the breccia near van Hauen Pass includes a set, now partly paralleled by diabase dykes, that is aligned normal to the strike of the present Otto Fiord cliff exposure near van Hauen Pass. Lateral migration of groundwater localized at the intersection of a regional fracture and the relatively impermeable Otto Fiord sulphate evaporites, could provide the necessary conditions for localized channellike solution of the sulphate, and consequent collapse of insoluble interbeds and of overlying Hare Fiord rocks. It is possible that subsurface water conditions of this type could be most active during shallow burial after extensive erosional unloading.

Biostratigraphy

In the type area, directly east of van Hauen Pass, fossils have been collected systematically from Sections 1, 2 and 3 (Fig. 7). Everywhere in the type area, faults are found at the base of the Otto Fiord Formation; thus we are unable to interpret precisely the time at which evaporite deposition began. However, about 70 km along strike from the type section, east of the head of Hare Fiord, faults are absent from the base of the Otto Fiord and foraminifers and algae of Late Mississippian age are abundant. Our data show that in the type area the greatest part of the formation was deposited during Early Pennsylvanian (Morrowan) time but that the uppermost part of the formation



Figure 23. Vertical aerial photograph mosaic showing distribution and stratigraphic relationships of the Otto Fiord Formation east of the head of Hare Fiord, Ellesmere Island. The type section of the Nansen Formation, adjacent to Girty Creek, contains about 1800 m of carbonate rocks ranging from Upper Mississippian (Chesteran) to Lower Permian (Wolfcampian). Note that in the vicinity of Section 4, east of Girty Creek, the evaporitic Otto Fiord Formation grades to the lower part of the Nansen Formation. An oblique view of the Otto Fiord Formation between A and A' showing the north wall of Hare Fiord east of Girty Creek and emphasizing evaporite to carbonate lithofacies change is shown in Fig. 24. A16785-115, A16788-139.

was laid down during the earliest Middle Pennsylvanian (Atokan). In terms of Mamet's zonal scheme (Mamet and Skipp, 1971), faunas representing Zone 20 (Morrowan) and possibly Zone 21 (lower Atokan) are present in the type area. An extreme upper limit to the age of the Otto Fiord in the type area is derived indirectly from fossils in lower beds of the overlying Hare Fiord Formation; Nassichuk (1975) reported Atokan ammonoids immediately above the type section of the Otto Fiord.

Section 1 (type section; Fig. 7); 3.5 km northeast of van Hauen Pass (81°03'30"N; 85°29'W). The type section has a thickness of 300 m. Although there are no fossil data from the lowest exposures of the type section, conodonts have been recovered from near the top of the formation and near the base of the overlying Hare Fiord Formation. These indicate a Morrowan to early Atokan age (Bender, pers. com., 1973). Conodonts from the type section are as follows:

GSC locality C-23895; 30 m below the top of the Otto Fiord Formation:

Gnathodus ouachitensis (Harlton)
Idiognathodus sp.
Ozarkodina sp.
Hindeodella sp.
Ligonodina sp.
Lonchodina sp.
Age: Morrowan (Bloydian) to ?early Atokan

GSC locality C-24402; 1.5 m above the base of the Hare Fiord Formation, which overlies the Otto Fiord Formation:

Gnathodus ouachitensis (Harlton)
Idiognathodus cf. *I. claviformis* Gunnell
Hindeodella sp.
Metalonchodina bidentata (Gunnell)
Ligonodina sp.
Age: Morrowan (Bloydian) to ?early Atokan

Although ammonoids have not been described from the type section of the Otto Fiord, several species are known from near the base of the type section of the overlying Hare Fiord Formation at the same locality. Nassichuk (1975) identified the following Atokan species from GSC locality 47867, 27 m above the base of the Hare Fiord Formation:

Metapronorites ellesmerensis Nassichuk
Phaneroceras lenticulare Plummer and Scott
Syngastrioceras smithwickense (Plummer and Scott)
Diaboloceras involutum Nassichuk

Section 2 (Fig. 7); 2.5 km northeast of van Hauen Pass. The Otto Fiord Formation has a measured thickness of 402 m and contains a diverse fauna in most limestone beds; ammonoids, foraminifers, algae, conodonts and brachiopods are listed in the appendix of this report. The ammonoids *Brannerоceras branneri* and *Gastrioceras* sp., both found 173 m and 327 m above the base of the formation, indicate an Early Pennsylvanian, Morrowan (Bloydian = early Bashkirian) age.

Mamet (see Appendix) identified a variety of calcareous foraminifers, algae and incertae sedis from 15 horizons between 173 and 393 m above the base and concluded that all were of late Early Pennsylvanian (Morrowan, Zone 20) or early Middle Pennsylvanian (early Atokan, Zone 21) age. Included are representatives of *Asteroarchaediscus, Biseriella, Bradyina, Calcisphaera, Climacammina, Diplosphaerina, Endothyra, Eotuberitina, Neoarchaediscus, Palaeotextularia, Planospirodiscus, Pseudoendothyra, Pseudoglomospira, Pseudoissinella, Rectangulina, Tetrataxis, Vicinesphaera, Parathurammina* and numerous others, including primitive representatives of *Pseudostaffella*. Mamet (pers. com., 1974) commented: "I have never previously encountered *Beresella-Pseudobradyina*? below Zone 21 (Atokan) in North America. *Pseudostaffella* is also not encountered below Zone 21, but in Eurasia this fusulinid is common in the lower Bashkirian = Morrowan (Zone 20) interval. *Beresella* has not been previously recorded from as low as Zone 20, but in several of these samples it is associated with definite Zone 20 faunas."

Conodonts were recovered from five horizons between 71 and 241 m above the base and these were identified by Bender (Appendix), who indicated a Morrowan age; dominant species in the interval include *Streptognathodus lateralis, Gnathodus nodulifera* and *Idiognathodus delicatus.* Other collections from near the top of the formation contained *Ozarkodina delicatula* and *Gnathodus ouachitensis* and Bender (Appendix) suggested a Morrowan - early Atokan age.

Finally, brachiopods were identified from seven levels between 71 and 393 m above the base of the Otto Fiord by Carter (Appendix), who indicated an age of "Morrowan and possibly younger". Thus, although there are as yet no faunal data from the lower 70 m of the Otto Fiord Formation in Section 2, all available data from higher in the formation suggest a Morrowan to early Atokan age.

Section 3 (Fig. 7); 4.8 km northeast of van Hauen Pass. A single limestone mound nearly 10 m thick near the middle of the Otto Fiord Formation contains abundant foraminifers and algae as shown in the following list by Mamet (pers. com., 1975):

GSC locality C-20530; 210 m above base of the Otto Fiord section:

Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Asteroarchaediscus aff. *A. gnomellus* Brenckle
Beresella sp.
Biseriella sp.
Bradyina sp.
Bradyina of the group *B. cribrostomata* Rauser-Chernoussova and Reitlinger
Calcisphaera sp.
Diplosphaerina sp.
Endothyra sp.
Eotuberitina sp.
cf. *Komia*? sp.
Palaeotextularia of the group *P. longisepta* Lipina
Planoendothyra sp.
Pseudobradyina? sp.
Pseudoendothyra sp.
primitive *Pseudostaffella* sp.
Tetrataxis sp.
Ungdarella sp.
Age: Early Pennsylvanian (Morrowan) or earliest Middle Pennsylvanian (early Atokan); Zone 20 or lower part of Zone 21

East of the head of Hare Fiord

Sedimentology

Otto Fiord rocks east of the head of Hare Fiord have been examined over a narrow outcrop belt that extends more or less continuously for about 16 km to the east of Girty Creek, which is 3.2 km east of the head of Hare Fiord (Figs. 6, 23). Otto Fiord rocks have been mapped an additional 11.2 km farther east to within 9.6 km of Chapman Glacier, but the formation has not been examined in that area for the present study (see geological Map 1309A, Otto Fiord, by Thorsteinsson and Trettin, 1972).

The Otto Fiord Formation shows lithological characteristics and stratigraphic relationships that are similar to those in the type area throughout much of the outcrop belt east of the head of Hare Fiord. In this area, sections of rhythmically alternating beds of anhydrite and limestone attain thicknesses of from 350 to 400 m. The Otto Fiord overlies conglomeratic beds of the red-weathering Borup Fiord Formation, which locally exceed 300 m in thickness and which rest unconformably on Cambrian rocks of the Franklinian Geosyncline. The Otto Fiord Formation is overlain by shale and argillaceous limestone of the Hare Fiord Formation. Some 3.2 km east of the head of Hare Fiord, in the vicinity of Girty Creek (Fig. 23), which is the site of the type section of the Nansen Formation, a complex facies change from Otto Fiord evaporites to Nansen carbonates is exposed (Figs. 23 - 25). This evaporite to carbonate facies change, which is modified by local reef facies, is indicative of deposition proximal to the northwestern edge of the depositional 'basin'. East of the head of Hare Fiord the data from the Otto Fiord Formation were obtained from several measured sections; one of these (Sec. 4, Fig. 23) is close to Girty Creek and the evaporite-carbonate facies change, and the other (Sec. 5, Fig. 23) is farther east in an area where the Otto Fiord assumes a more 'typical' lithological profile. Specifically, Section 4 is 2.2 km and Section 5 10.9 km east of Girty Creek.

At Section 4 the Borup Fiord Formation is overlain by 180 m of bedded carbonate rocks that are continuous with carbonate strata in the lowest part of the type Nansen Formation 2.2 km along strike to the west; however, at this locality they are considered to be a facies of the Otto Fiord Formation. This bedded platform unit is overlain by more typical Otto Fiord beds of anhydrite, limestone and shale. The bedded platform unit is characterized by well bedded, dark, skeletal wackestone and dolomitic wackestone containing a shallow marine biota. The most common skeletal components are tubular and encrusting foraminifers and beresellid and other types of algae, with fewer crinoids, brachiopods, gastropods, ostracodes, bryozoans, fusulinacean foraminifers and calcispheres. A few beds contain ooids; most contain abundant micritic pelletoids. Several discontinuous beds of anhydrite up to several metres thick are interbedded within the platform unit.

About 1 km along strike to the east, this carbonate platform unit disappears abruptly and, in its place, anhydrite and lensoid masses of rubbly weathering limestone appear (Fig. 25). The nature of the abrupt change is difficult to decipher. It is either a fault or a very abrupt facies change; if a fault, it may be synsedimentary, predating evaporite deposition, because the beds in overlying anhydrite and limestone cycles carry across the break without disturbance. Another alternative is that the lensoid limestone masses enclosed in anhydrite are slump masses separated from the bedded limestone unit by small listric faults. However, the geometry of these lenticular limestone bodies (Fig. 25) suggests that they are more likely to be in-place mounds or reefs similar to those found in the Otto Fiord Formation at van Hauen Pass and also in the region some 11 km east of the head of Hare Fiord. The geometry of the lenticular limestones illustrated in Figure 25 is traced from colour slides; they appear to be stacked and 'stepping' back toward the former platform edge. About 6 km along strike to the east, where the Otto Fiord Formation is developed fully and overlain by argillaceous limestone, shale and siltstone of the basinal Hare Fiord Formation, four or five prominent limestone units are interbedded with the anhydrite; these bedded limestone units can be traced back along strike to the west where it is seen that they correspond approximately with the four or five limestone mounds. Similar lenticular limestone units are interbedded with Otto Fiord anhydrite higher in the section above the stacked mounds and these also are interpreted to be in-place mounds (Fig. 25). Both groups of mounds, if interpreted correctly, probably are constructed largely of beresellid and kamaenid algae, similar to the mounds at van Hauen Pass, as correlative limestones along strike contain many of these algae, some in submarine-cemented framestone fabrics (Fig. 27).

Where it is continuous, the top of the carbonate platform at the base of the Otto Fiord is overlain by about 180 m of interbedded anhydrite, limestone, and black, red and green shale or argillaceous limestone, which closely resemble typical Otto Fiord rocks (Fig. 24). The lower 50 to 80 m of this bedded succession grades westward into bedded limestone, which appears to be of the same type as the underlying platform unit. Along strike to the east, the lower beds of anhydrite in the Otto Fiord section interfinger with and surround another small limestone mass interpreted as a reef (Fig. 24) with a composite thickness of about 12 m.

The 180 m thick Otto Fiord section above the platform unit is interrupted about 100 m above the base by a prominent, discontinuous limestone unit that has a planar base and an irregular upper surface (unit 1, Figs. 24, 25). It varies from 40 to 50 m thick where it overlies the bedded anhydrite, but thins rapidly to the east, so that 500 m along strike it is indistinguishable from other thin limestone units interbedded in anhydrite. To the west, this irregular limestone unit thickens and, within a few hundred metres, may be traced into the base of a steeply dipping tongue of limestone that continues up the dip (depositional) for at least several hundred metres before becoming unrecognizable. Apparent depositional relief on the upper surface of this steeply dipping limestone tongue and its subhorizontal extension into the anhydrite section is at least 100 m.

Limited sampling through the irregular limestone unit where it is interbedded with anhydrite indicates that beresellid algal framestone fabrics with early fibrous cements are common in at least the central part of the unit. These fabrics are identical to those found in the van Hauen Pass mounds. The upper beds include pelletoidal skeletal packstones with algae as a common component.

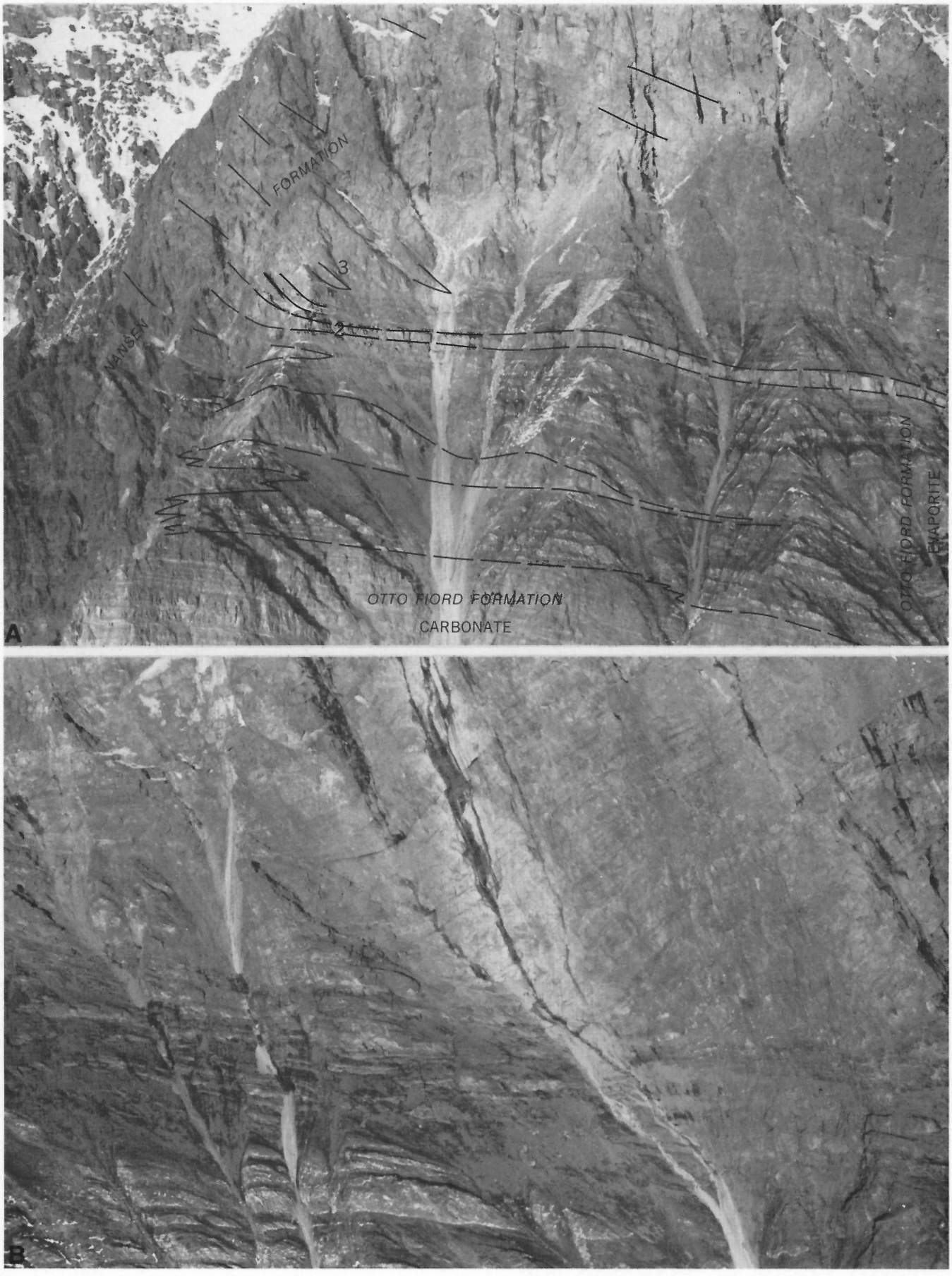
The irregular limestone unit is overlain by well bedded anhydrite, limestone, and red and green shaly units of typical Otto Fiord evaporite facies. Because of the irregular topography of the underlying carbonate, thickness of this evaporite section varies from 50 m to more than 100 m. To the east the evaporite section is continuous with the upper part of the fully developed Otto Fiord Formation, which is more than 300 m thick. To the west, however, a

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Figure 24. North wall of Hare Fiord east of Girty Creek illustrating anhydrite and carbonate facies relationships at or close to the northwestern margin of the Otto Fiord evaporite depositional 'basin'.

A. Major anhydrite to carbonate facies change. Thickness of visible section is about 500 m. Section 4 was measured up the prominent talus shute at left centre. Interbedded anhydrite, limestone and shales of the Otto Fiord Formation at lower right (east) grade to the left (west) into carbonates. Discontinuous limestone tongues (1 - 3) projecting outward from this facies transition are keyed to Fig. 25. Note the steep depositional dips within these carbonate tongues and in the massive limestones of the overlying basinward-prograding Nansen Formation.

B. Local mounding of promontory in prograding shelf-edge carbonates of Nansen Formation above Otto Fiord anhydrite, shale and limestone; corresponds with upper right of Fig. 25.


FORMATION
3
2
NANSEN
OTTO FIORD FORMATION
CARBONATE
OTTO FIORD FORMATION
EVAPORITE
A
B

more dramatic facies change is revealed where anhydrite beds pinch out between tongues of limestone that are extensions of steeply dipping shelf foreslope limestone units (Figs. 24, 25). Present dips on bedding planes in these foreslope rocks exceed 30 degrees; extensive submarine cementation by fibrous cements appears to be a controlling factor in stabilization of these steep depositional slopes. The rapidly thinning tongues of limestone that extend outward into the anhydrite section are characterized once again by beresellid algal framestone fabrics with early fibrous calcite cements. Depositional relief on the carbonate foreslope units, determined by tracing bedding planes on photographs of the unscaleable cliffs above and west of the section, is estimated to increase within each successive unit and to be at least 400 m for the uppermost limestone-anhydrite cycles.

The evaporite section is terminated by several more prominent and resistant limestone units that are the extensions of foreslope tongues thinning out to the west (Figs. 24 - 26). Thin-bedded, argillaceous limestone immediately below these more massive units contains a crinoid-rich, varied marine biota and is identical to many other rocks of the 'basinal' Hare Fiord Formation that normally overlie the Otto Fiord Formation. The massive discontinuous limestone strata above the evaporites include limestone breccias and graded beds interpreted as debris sheets and turbidites shed downslope from the adjacent reef and shelf edge (Davies, 1977a). The most prominent of these debris beds is overlain by about 50 m of dark, recessive, argillaceous limestone typical of the Hare Fiord Formation (Fig. 26B). Again, the Hare Fiord rocks grade laterally to the west into the same type of steeply dipping foreslope beds of the shelf edge, with depositional relief on the uppermost bedding planes exceeding 450 m.

This thin unit of Hare Fiord rocks of deeper water aspect is overlain abruptly by more than 1000 m of light-coloured limestone of the Nansen Formation (Figs. 24 - 26). Throughout this mass of limestone, steeply dipping bedding planes may be followed from colour photographs to delineate a complex of prograding foreslope sediment tongues and local reeflike masses. These reeflike masses, however, may be local spurs or promontories along the shelf-to-basin front, truncated randomly by the fiord wall (Fig. 24B). A few kilometres east of this area the massive shelf limestone beds of the Hare Fiord Formation in turn grade by facies change into basinal rocks of the Hare Fiord Formation of normal thickness (500 m).

In summary, the exposure at Section 4 (Fig. 23) reveals the complexities that may be present at a shelf-to-trough transition from marine carbonates to 'basinal' evaporites. Additionally, it emphasizes for these evaporites the depth of water during deposition of interbedded limestone. Significantly, this entire section records progressive deepening of successive transgressive events into the Sverdrup basinal depression. The relief on lowermost evaporite units is from a few metres to 300 m or more at termination of evaporite deposition, to 450 m or more for deep-water sediments of Hare Fiord type, prior to the progradation of massive shelf-edge sediments into this marginal part of the basin floor.

Section 5 (Figs. 6, 23) was measured 10.9 km east of Girty Creek, near the eastern end of the east-northeast trending 'belt' of Otto Fiord rocks (Fig. 27) exposed along the northern wall of the glacier-filled valley that is the landward extension of Hare Fiord. The Otto Fiord Formation in Section 5 (Fig. 28) is 365 m thick to the top of the uppermost but discontinuous anhydrite unit; the highest thick and continuous anhydrite unit is 290 m above the base of the formation. The Otto Fiord is underlain by red conglomerates, sandstone and other rocks of the Borup Fiord Formation, and is overlain by at least 400 m (estimated) of dark, well bedded cherty limestone of basin slope facies in the Hare Fiord Formation. The top of the Hare Fiord Formation is interbedded with and overlain by light-coloured massive limestone of the prograding, shelf-forming Nansen Formation (Fig. 27). Several kilometres east of Section 5, a major facies transition from Hare Fiord to Nansen rocks is exposed (Fig. 23).

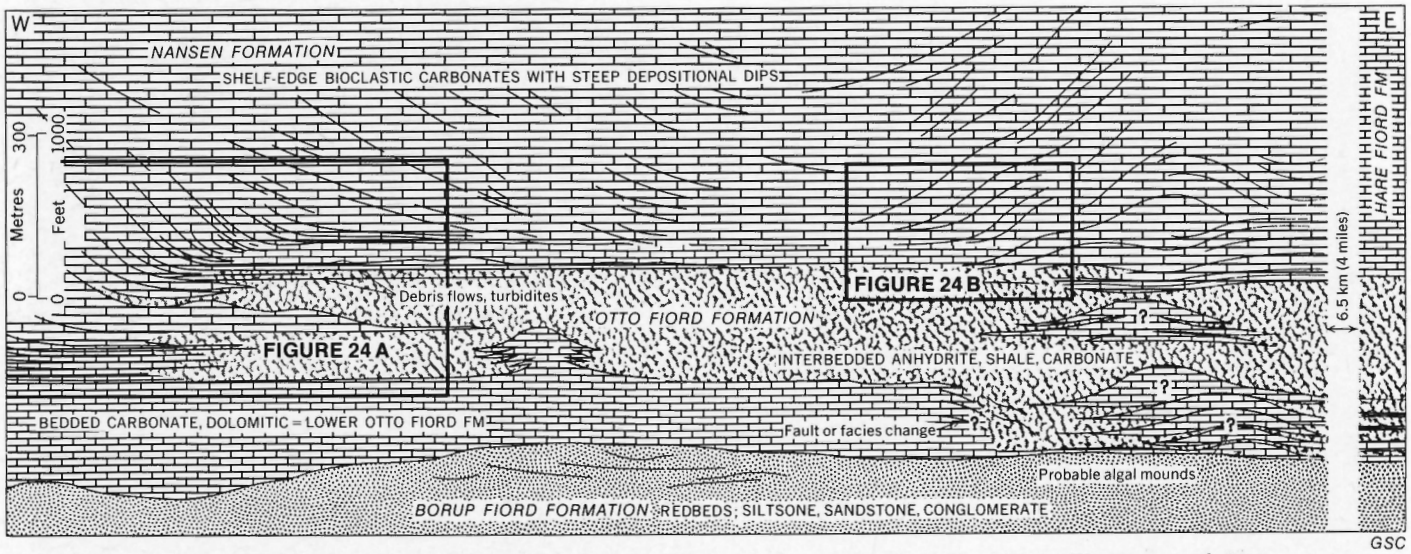
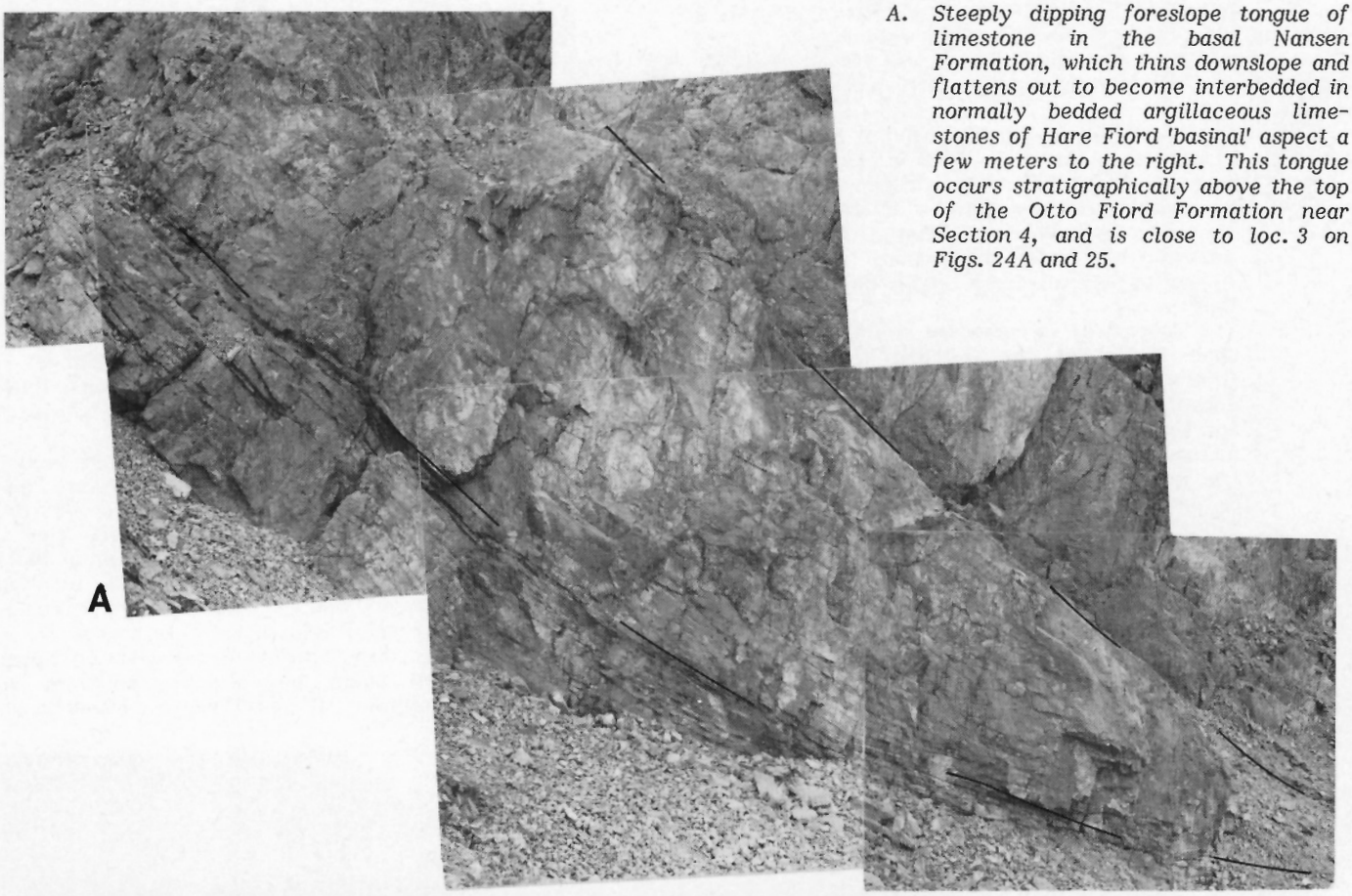
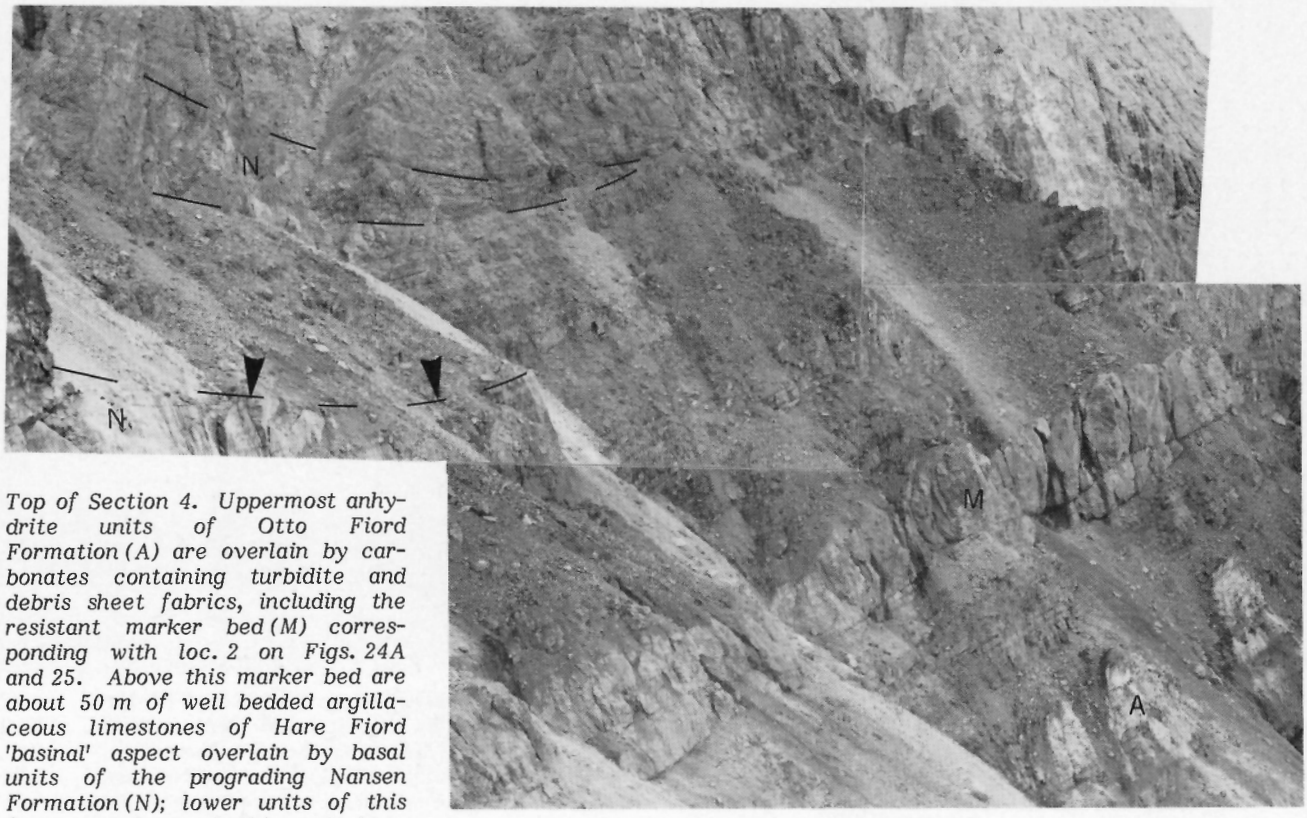


Figure 25. Carbonate-anhydrite facies relationships along the north wall of Hare Fiord east of Girty Creek (Fig. 23). The section is 2.5 km long; Fig. 24A covers the western third of this section. The appended lithological column at right represents the more normal succession, away from this complex facies change, which is exposed about 6.5 km along the fiord to the east (see Fig. 27). To place this section in regional facies perspective, it may be compared with the westward facies boundary of the Otto Fiord Formation shown in Fig. 3. The exposure spans three formations ranging in age from late Mississippian to Early Permian.



A. *Steeply dipping foreslope tongue of limestone in the basal Nansen Formation, which thins downslope and flattens out to become interbedded in normally bedded argillaceous limestones of Hare Fiord 'basinal' aspect a few meters to the right. This tongue occurs stratigraphically above the top of the Otto Fiord Formation near Section 4, and is close to loc. 3 on Figs. 24A and 25.*



B. *Top of Section 4. Uppermost anhydrite units of Otto Fiord Formation (A) are overlain by carbonates containing turbidite and debris sheet fabrics, including the resistant marker bed (M) corresponding with loc. 2 on Figs. 24A and 25. Above this marker bed are about 50 m of well bedded argillaceous limestones of Hare Fiord 'basinal' aspect overlain by basal units of the prograding Nansen Formation (N); lower units of this formation (arrows) thin out down depositional slope and flatten out into more argillaceous Hare Fiord rocks.*

Figure 26. Bedding relationships and characteristic lithologies of Otto Fiord rocks and overlying units in Section 4, east of Girty Creek.

The Otto Fiord Formation in the vicinity of Section 5 east of Girty Creek bears a general resemblance to the same formation in the type area at van Hauen Pass but several important differences can be distinguished:

(1) The lower half of the formation at Section 5 and elsewhere in the same 'belt' is characterized by four prominent and thick limestone units (Fig. 27) with relatively thin units of interbedded anhydrite (poorly differentiated in Section 5 because of local topographic irregularities). Anhydrite beds are more thickly developed in the type area.

(2) The upper half is composed mainly of recessive shale and other fine detrital sediments, with fewer interbeds of the sandstone, limestone and anhydrite that characterize the upper half of the formation in the type area.

(3) In the vicinity of Section 5 the contact between the Borup Fiord Formation and the Otto Fiord Formation is exposed. In the type area pre-Otto Fiord rocks are not exposed.

(4) At least one small carbonate mound is present near Section 5 and the pre-Otto Fiord topography on Borup Fiord detrital rocks appears to have controlled its localization (Fig. 28). Mounds in the Otto Fiord Formation in the type area do not appear to have been influenced by Borup Fiord topography.

The contact between red beds in the Borup Fiord Formation and the overlying Otto Fiord Formation is irregular and complex in the general area of Section 5. The contact varies in style along strike from apparently conformable and lithologically gradational, to unconformable with as much as 60 m of relief developed on the Borup Fiord red-bed surface. Individual limestone and anhydrite units of the Otto Fiord Formation thicken and thin depositionally over this irregular surface (Fig. 28).

At least one of the more abrupt changes in relief on the Borup Fiord rock surface appears to have been formed by pre-evaporite fault displacement. Beds of red conglomerate and other detrital rocks in the Borup Fiord Formation on the upthrown side are truncated at a near-vertical contact, while beds of limestone and anhydrite in the overlying Otto Fiord Formation thicken abruptly on the downthrown side. Total relief over this structure is about 3 m.

The largest structure (Fig. 28) east of Section 5, at the easternmost end of the outcrop belt, is less clearly identified as a fault-generated feature. Maximum relief across the structure at the contact between the two formations is 60 m. Unfortunately, the sloping contact surface is recessive and mostly covered by talus. Discontinuous lenses and beds of sheared anhydrite occur at or close to the contact; this shearing is consistent with postevaporite fault displacement or rejuvenation along a pre-evaporite fault plane, but also is consistent with differential compaction of interbedded anhydrite and



Figure 27. Typical exposure of the Otto Fiord Formation near Section 6 at eastern end of the glacier-filled headward extension of Hare Fiord (Fig. 23). Field of view is about 1 km wide. Note rhythmic interbedding of thick carbonate units and anhydrite in lower half of Otto Fiord Formation, and more shaly (light-coloured, recessive interval) section in the upper half. Borup Fiord red beds are exposed at the extreme base of the cliff, and well bedded cherty carbonates of the Hare Fiord Formation lie above the Otto Fiord rocks. Massive limestone tongues of the prograding Nansen Formation occur at the top of the cliff. Section 6 was measured up the gully at right centre. GSC 199252.

limestone over the irregular Borup Fiord topography under deep burial conditions. Consequently, although the large structure probably formed initially by pre-evaporite fault displacement, its origin solely by depositional relief and erosion or its modification by erosion after faulting cannot be discounted.

Where the contact between the two formations is not complicated by high-relief topography, the lithological character of the two formations immediately above and below the contact suggests that they are conformable and gradational. Although this is inconsistent with the evidence for unconformity, it suggests that development of relief on the Borup Fiord red-bed surface, whether by faulting, erosion (and perhaps partly depositional?), or both, was nearly contemporaneous with the onset of Otto Fiord evaporite and carbonate deposition. As these events also are more or less contemporaneous with the first marine incursion into the Sverdrup Basin in this locality, perhaps there was a close genetic relationship between faulting, basin development and marine transgression in this tectonosedimentary setting.

In Section 5 beds and discontinuous lenses of red-stained limestone containing marine microfossils (foraminifers, ostracodes) are found interbedded with the red detrital rocks of the Borup Fiord Formation. A limestone sample from near the top of the Borup Fiord, 15 m below the base of the Otto Fiord, yielded a Late Mississippian (Zone 18) microfauna. Fossils of comparable age occur at sample localities at the base of the Otto Fiord section. Limestone near the top of the Borup Fiord Formation is generally oolitic and sandy (Fig. 29A), with many ooids nucleated on quartz grains. These ooids are largely recrystallized and replaced by calcite spar, suggesting early exposure to fresh water in vadose diagenetic environments. The limestone is interbedded with, or enclosed by, thicker units of crossbedded, mottled red and green conglomerate, sandstone, siltstone and shale. The sandstone is composed mainly of subangular clasts of quartz and chert, with a few of shale and other rock types. The sand grains have a surface coating of iron oxides. Cements in sandstone lower in the section are syntaxial quartz and calcite but calcite becomes the dominant cement in sandstones closer to the Otto Fiord contact.

Sandstones in the Borup Fiord Formation within 1 m of the Otto Fiord contact appear to be bleached and lack the definite red staining of lower units. The first Otto Fiord anhydrite unit above the contact is discontinuous and contains some discontinuous and deformed limestone beds. Some structural displacement along this contact after deposition is suggested by these irregularities.

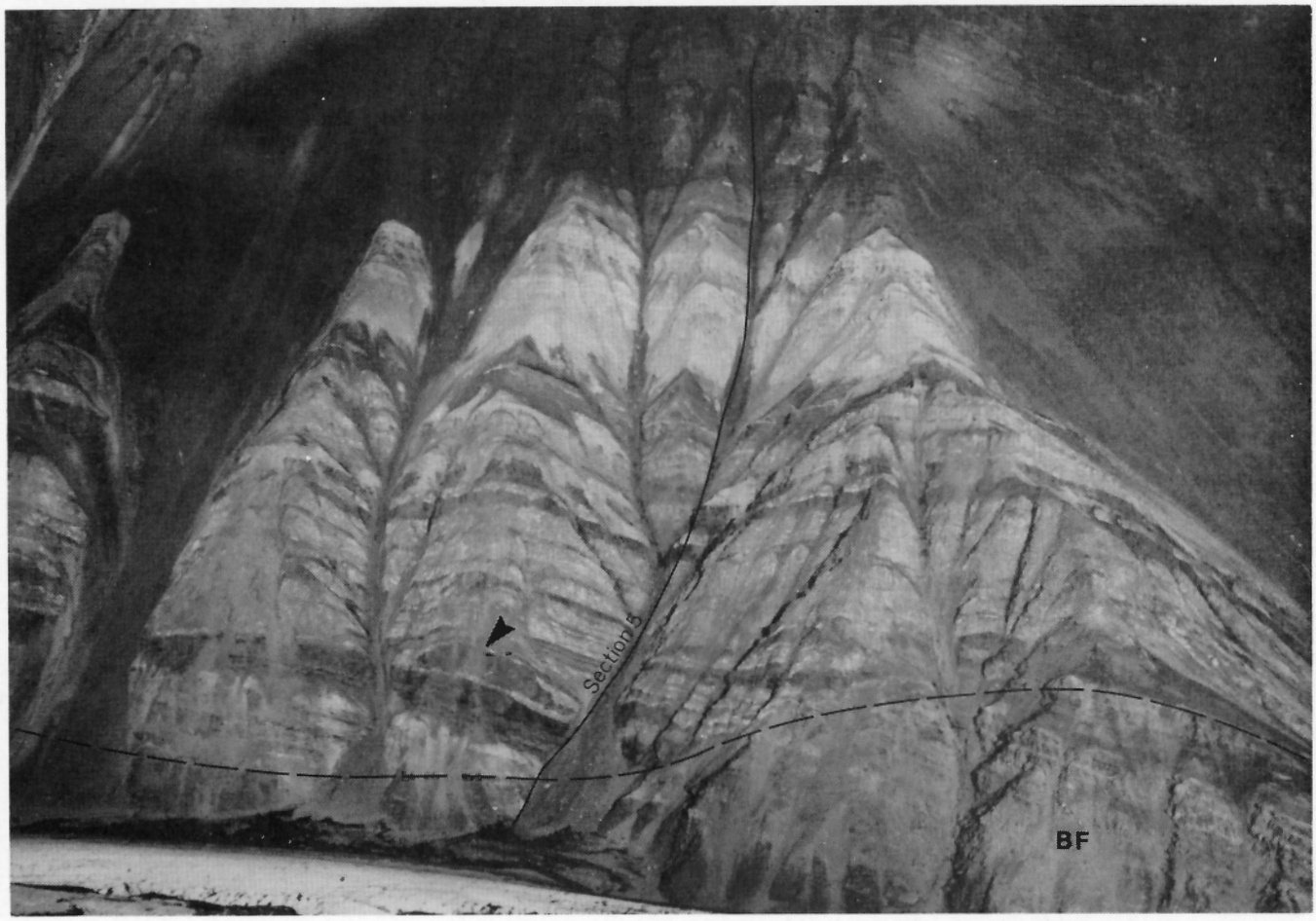
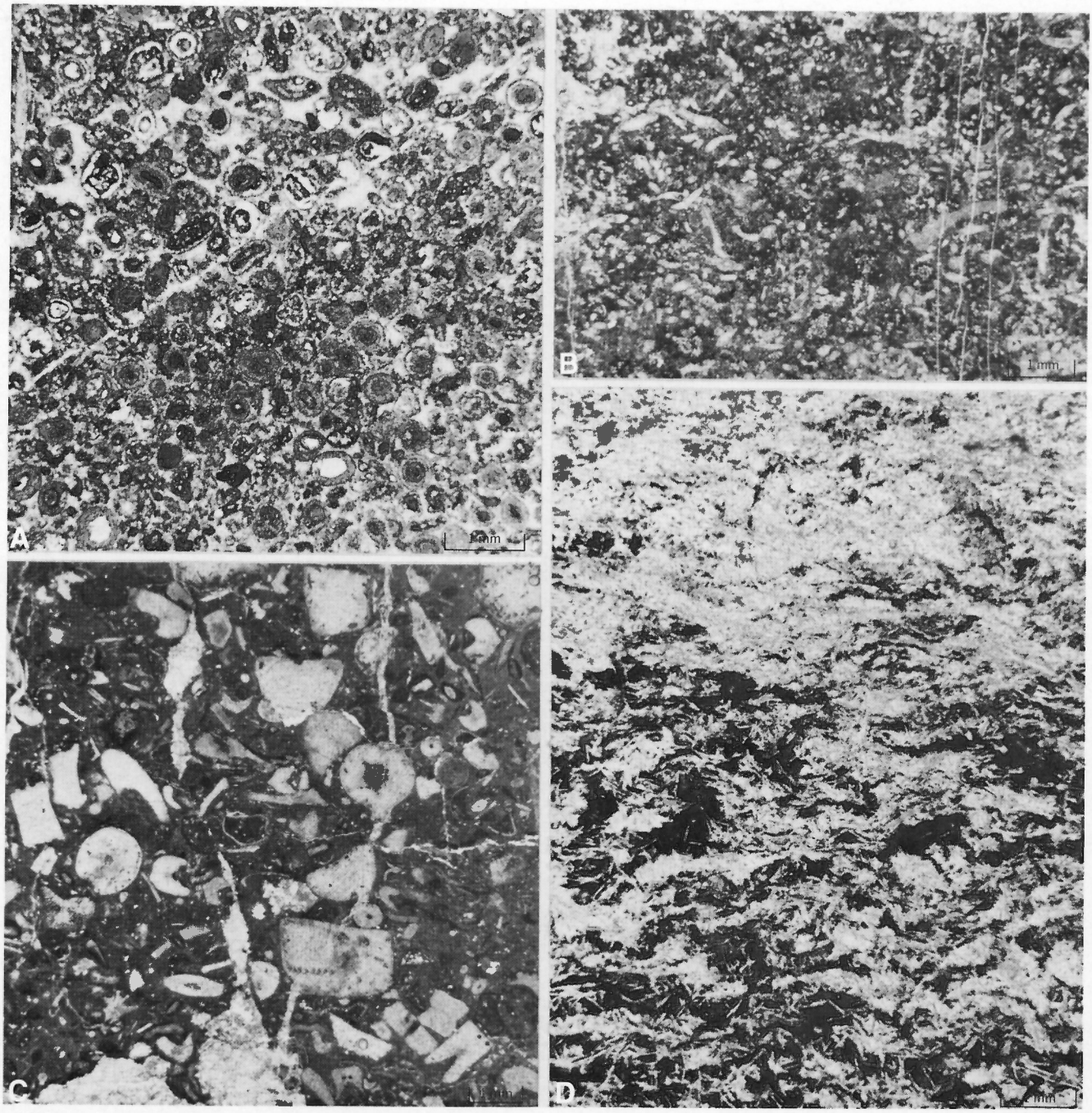


Figure 28. Exposure of the Otto Fiord Formation at Section 5, about 1 km east of the area illustrated in Fig. 27. Section 5 was measured up the gully at right centre. About 60 m of relief are present at the contact between Borup Fiord red beds (BF) and the interbedded carbonates and anhydrite of the Otto Fiord Formation. Beds in the Otto Fiord pinch out against the drape over this structure. Note the limestone mound (arrow) enclosed within the Otto Fiord rocks off the flank of the former topographic high. GSC 199249.



A. Oolitic grainstone from upper beds of the Borup Fiord Formation. Some ooids are nucleated on quartz sand grains, others on algae, foraminifers and other clasts.

B. Algal packstone of hypersaline aspect from a carbonate unit in the Otto Fiord Formation in Section 5. Algae are beresellid and kamaenid types.

C. Crinoid-rich wackestone from the Otto Fiord Formation with accessory brachiopod and ostracode clasts in a micritic matrix.

D. Vertical transition from micritic carbonate to anhydrite as a product of diagenetic replacement and displacement of the carbonate by anhydrite. Petrographic relationships indicate that much of the anhydrite at transitional contacts from carbonate to anhydrite units is of replacement origin, and often results in a crude nodular mosaic fabric.

Figure 29. Characteristic lithologies of Borup Fiord and Otto Fiord rocks in Section 5.

The Otto Fiord Formation at Section 5 differs lithologically from sections in the type area (Secs. 1, 2, 3; Fig. 7) and the sections near Wood Glacier (Secs. 7, 8; Fig. 6) mainly in the volume of shale in the upper part. From the base of the section to about 190 m above the base, anhydrite units compose slightly more than half the thickness, and limestone units form most of the remainder. Shale and sandstone are very minor components. From 190 m to the uppermost anhydrite bed defining the top of the section at 365 m, shale forms more than half the formation and anhydrite accounts for about 15 per cent by thickness. The remainder is composed of limestone and sandstone.

Because of the depositional complexities imposed by the irregular topography on the upper surface of the Borup Fiord red beds, and the general loss of identity of major limestone units toward the eastern end of the outcrop belt at this locality, Section 5 does not illustrate clearly the cyclic nature of limestone-anhydrite or limestone-anhydrite-detrital sediment rhythms found in other sections. However, there are sufficient lithological data from this section to characterize the composition of the limestones in the section and the fabric of the anhydrite units.

The first limestone unit with a varied marine biota in the Otto Fiord Formation at Section 5 occurs about 4 m above the Borup Fiord contact. A limestone unit 4 m thick that begins at that stratigraphic level is composed at its base of bioturbated silty wackestone and packstone containing bioclasts of crinoids, bryozoans, brachiopods, gastropods, foraminifers and a few algae. Micritic intraclasts and lumps are of several types. The top of the unit is composed of very sandy bioclastic grainstone with a variety of foraminifers and other bioclasts.

Other limestone units in the lower 190 m of the formation in Section 5 vary from a few metres to 16 m thick. Where sampling density is sufficient to define it, a general succession may be distinguished from algal-rich limestone of hypersaline aspect (Fig. 29B) at the base of units, to limestone with a more varied marine biota within the units. Crinoids, brachiopods, solitary corals, trilobites, ostracodes and foraminifers are present in these more marinelike carbonates (Fig. 29C). Alternatively much of the entire unit may be dominated by beresellid, kamaenid or other types of red algae, or by fine-grained lime mudstone with only a few scattered clasts of crinoids and brachiopods. Uppermost beds may be dominated by algae or a mixed algal-crinoid wackestone.

Anhydrite units in the lower 190 m of the formation range from about 4 to 15 m thick. The most common fabrics are distorted or compacted nodular mosaics and laminated anhydrite. Crystal pseudomorph fabrics after gypsum have not been observed in this section. Most fabrics appear to have been affected by structural or compactional shear deformation, again perhaps caused by differential compaction over the irregular Borup Fiord substrate. Traces of elemental sulphur are present in some beds. Replacement and displacement of micritic carbonate by anhydrite is common in the transition zone between limestone and anhydrite units (Fig. 29D).

In the upper 180 m, black, green and red shale and argillaceous limestone make up more than 50 per cent of the section by thickness. Sandstone, limestone and anhydrite make up the remainder, with anhydrite contributing only about 15 per cent of the total. The shale forms units up to 17 m thick and is recessive, fissile and laminated. Interbedded sandstone units up to 5 m thick are buff to green and calcite cemented; a few bioclasts, authigenic dolomite and some glauconite are present in some of this sandstone.

Limestones in the upper half of the Otto Fiord section are dark, argillaceous and, in places, are fine-grained mudstones with a few large bioclasts. Some, however, contain abundant algae and foraminifers, including fusulinaceans in increasing abundance in stratigraphically higher units. Crinoids, brachiopods, bryozoans and trilobites are present in limestone with more varied marine biotas.

In the upper 60 m of the formation anhydrite is rare; only four thin and discontinuous beds occur. The anhydrite unit 3 m thick that is used to define the top of the section 365 m above the Borup Fiord - Otto Fiord contact also is discontinuous. It pinches out either depositionally or by postdepositional deformation to the west of the measured section. Within the upper 25 m of section, the rocks comprise well bedded, dark, argillaceous and cherty limestone with shaly interbeds, more typical of the overlying Hare Fiord Formation than normal Otto Fiord rocks. Thus the use of the uppermost and discontinuous anhydrite unit to define the top of the Otto Fiord Formation may be erroneous. Apart from this terminological problem, however, the significant factor is that the contact between the Otto Fiord and Hare Fiord formations in this section appears to be conformable and lithologically gradational.

Only the basal 60 m of the overlying Hare Fiord Formation above the uppermost anhydrite bed were measured in this section. The basal Hare Fiord rocks comprise bedded, argillaceous and cherty, bioclastic and micritic limestone with shaly partings, typical of Hare Fiord rocks elsewhere in the shelf-slope setting. The rocks become thicker bedded, less argillaceous and more resistant higher in the basal section.

A summary of the principal features of the Otto Fiord Formation in Section 5 shows that there is a strong dilution by shale of the more typical limestone-anhydrite component in the upper half of the section. The recognition of discrete cycles or depositional rhythms in this section is not as obvious as in the type section of the formation or as in Section 8 (Fig. 6) at Wood Glacier. A basic limestone-anhydrite rhythm is apparent, however, in the lower half of the section, and six or seven limestone-anhydrite-shale cycles, perhaps comparable with the limestone-anhydrite-mudstone/sandstone cycles of the Wood Glacier section, are recognizable in the upper half; several incomplete cycles of limestone-shale also are present.

Limestone mounds in Section 5. Several small limestone mounds are in the lower part of the Otto Fiord Formation in Section 5, but only one is exposed clearly. This mound, about 40 m above the base of the formation, is about 12 m high and 55 m long in the plane of exposure. It is separated from an underlying thick limestone unit by about 5 m of anhydrite and is overlain by anhydrite and limestone, which drape over the top and flanks of the mound (Fig. 28).

The mound is composed of rubbly weathering, dark limestone intensely recrystallized to a calcite spar mosaic. However, relict fabrics in the rock reveal that originally it was characterized by the same type of radial fibrous cement fabrics found in the main algal mound facies of the van Hauen Pass buildups. No preserved algal frame has been recognized in these rocks, but relict textures suggest that algae were the primary mound builders. The mound limestone thins into flank beds, and is continuous with a normally bedded argillaceous limestone about 3 m thick in the off-mound setting. This coeval limestone contains a varied marine biota including crinoids, corals and brachiopods.

The substrate control for the localization of this mound is a slight thickening of the thick limestone unit that directly underlies the mound. A more significant substrate control appears to be the local topographic relief on the Borup Fiord red beds (Fig. 28). Thus apparent structural displacement by faulting prior to deposition of the evaporites and marine carbonates effectively has controlled the localization of carbonate buildups in the overlying succession.

Biostratigraphy

East of the head of Hare Fiord, the Otto Fiord Formation ranges in age from Late Mississippian (Chesteran) to Middle Pennsylvanian (Atokan and possibly early Desmoinesian). In terms of Mamet's foraminiferal zonal scheme (in Fig. 4), Zones 18 (Chesteran), 20 (Morrowan), 21 and 22 (Atokan) and possibly 23 (Desmoinesian) are present. Although Zone 19 (uppermost Chesteran) is in the lower part of the type section of the Nansen Formation, which grades to the Otto Fiord Formation in the immediate vicinity of Section 4 (see Mamet, in Nassichuk, 1975), it has not been identified from the Otto Fiord Formation proper. According to Mamet et al. (1971), Zone 19, which is characterized by *Eosigmoilina*?, represents the time equivalent of Bisat's (1924) *Homoceras* Zone H. Zone H has not been clearly identified throughout much of North America and may coincide with a major hiatus between the Mississippian and Pennsylvanian. Mamet (in Thorsteinsson, 1974; and in Nassichuk, 1975) has identified Chesteran Zone 18 foraminifers and algae from limestone near the top of the type section of the Borup Fiord Formation at Stepanow Creek, 30 km west of Girty Creek (Fig. 6). Faunas of comparable age also occur in the Borup Fiord Formation east of Girty Creek and also near the base of the type section of the Nansen Formation at Girty Creek. These data support field observations of interfingering between the Otto Fiord and Nansen formations but also tend to indicate an uninterrupted transition from Borup Fiord deposition to Nansen and contemporaneous Otto Fiord deposition.

East of the head of Hare Fiord all of our biostratigraphic information for the Otto Fiord Formation has come from fossils collected from the previously described Sections 4 and 5 and also from lithologically undescribed Section 6 (Fig. 23).

Section 4 (Fig. 23); 2.2 km along strike to the east from Girty Creek. Although the Borup Fiord Formation is exposed beneath the Otto Fiord Formation at this location, the basal beds of the Otto Fiord appear contorted and may be faulted. In spite of that, the Otto Fiord has a thickness of 351 m, which is comparable to the thickness farther east along strike where the section definitely is unfaulted. Immediately west of Section 4, in the vicinity of Girty Creek, Otto Fiord rocks grade to carbonates and shale of the Nansen Formation. Calcareous foraminifers and algae collected from near the base and near the top of the Otto Fiord Formation have been identified by Mamet (pers. com., 1975) as follows:

GSC locality C-31017; 13.5 m above the base of the Otto Fiord Formation:

apterrinellids
Asphaltina sp.
Asphaltina cordillerensis Mamet, in Petryk and Mamet (1972)
Asteroarchaediscus sp.
beresellid algae
Bradyina sp.
Calcisphaera sp.
Endothyra sp.
Eostaffella sp.
Eotuberitina sp.
primitive form of *Globivalvulina*
Millerella sp. (sensu stricto)

Age: Zone 20 or possibly 21; Early Pennsylvanian (Morrowan) or possibly early Middle Pennsylvanian (Atokan)

GSC locality C-31018; 21.6 m above the base of the Otto Fiord Formation:

apterrinellids
Asphaltina sp.
Asphaltina cordillerensis Mamet, in Petryk and Mamet (1972)
Asteroarchaediscus sp.
beresellid algae
Bradyina sp.
Calcisphaera sp.
Climacammina sp.
Eostaffella sp.
Globivalvulina sp. (primitive form)
Lipinella sp.
Millerella sp. (sensu stricto)
Neoarchaediscus sp.
"Palaeotextularia" sp.
Planospirodiscus sp.
Stacheoides sp.
Zellerina sp.

Age: Zone 20 or possibly Zone 21; Early Pennsylvanian or possibly early Middle Pennsylvanian (Atokan)

GSC locality C-31024; 71 m above the base of the Otto Fiord Formation:

apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Bradyina sp.
Calcisphaera sp.
Climacammina sp.
Endothyra sp.
Eostaffella sp.
Globivalvulina sp.
Pseudostaffella sp.
Tetrataxis sp.
Zellerina sp.

Age: Zone 21 or 22; Middle Pennsylvanian (Atokan)

GSC locality C-31051; 28.5 m above the base of the Hare Fiord Formation (measured above the highest anhydrite bed in the Otto Fiord):

abundant fusulines and beresellid algae

Age: Zone 23?; Middle Pennsylvanian (Desmoinesian)

Section 5 (Fig. 23); 10.9 km along strike to the east from Girty Creek. This section, contained between red-weathering sandstone of the underlying Borup Fiord Formation and shaly, dark grey weathering limestone of the Hare Fiord Formation, is 360 m thick. The Hare Fiord Formation was mapped locally as part of the Nansen Formation by Thorsteinsson and Trettin (1972). The top of the Otto Fiord is considered to be at the top of the highest anhydrite bed in the section. Calcareous foraminifers and algae collected from near the top of the Borup Fiord Formation directly below the Otto Fiord and from near the base and top of the Otto Fiord Formation at Section 5 have been identified by Mamet (pers. com., 1976) as follows:

GSC locality C-31068; 15 m below the top of the Borup Fiord Formation:

apterrinellids
Asteroarchaediscus sp. (very abundant)
Biseriella sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Calcisphaera sp.
Endothyra sp.
Globoendothyra sp.
Neoarchaediscus sp.
Pseudoglomospira sp.
Tetrataxis sp.
Zellerina sp.

Age: Zone 18; Late Mississippian (Chesteran)

GSC locality C-31077; 4 m above the base of the Otto Fiord Formation:

apterrinellids
Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Asteroarchaediscus ovoides (Rauser-Chernoussova)
Calcisphaera sp.
Endothyra sp.
Eotuberitina sp.
Fourstonella sp.
Neoarchaediscus parvus (Rauser-Chernoussova)
Palaeotextularia sp.
new stacheiin algae (similar to C-31017)
Tetrataxis sp.
Tuberitina sp.
Zellerina sp.
Age: Zone 18 or Zone 20; Late Mississippian (Chesteran) or Early Pennsylvanian (Morrowan)

GSC locality C-31080; 8 m above the base of the Otto Fiord Formation:

Asteroarchaediscus sp. (extremely abundant)
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Biseriella sp.
Bradyina sp.
Calcisphaera sp.
Climacammina sp.
Endothyra sp.
Eostaffella sp.
Eostaffellina sp.
Neoarchaediscus sp.
Millerella sp.
Palaeotextularia sp.
Tetrataxis sp.
Zellerina sp.
Age: Zone 20; Early Pennsylvanian (Morrowan)

GSC locality C-31088; 33 m above the base of the Otto Fiord Formation:

apterrinellids
Asteroarchaediscus sp. (abundant)
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Beresella sp.
Biseriella sp.
Calcisphaera sp.
Climacammina sp.
Eoschubertella sp.
Eotuberitina sp.
Globivalvulina of the group *G. bulloides* (Brady)
Neoarchaediscus sp.
Neoarchaediscus grandis (Reitlinger)
Planospirodiscus sp.
Pseudoendothyra sp.
Pseudostaffella sp.
Tetrataxis sp.
Zellerina sp.
Age: Zone 21 or 22; Middle Pennsylvanian (Atokan)

GSC locality C-31131; 331 m above the base of the Otto Fiord Formation:

Asteroarchaediscus sp.
Calcisphaera sp.
Climacammina sp.
Deckerella sp.
Endothyra sp.
Globivalvulina sp.
Komia sp.
Haplophragmina sp.
Neoarchaediscus sp.
Neostaffella sp.
Pseudostaffella sp.
Shartymophycus? sp.
Tetrataxis sp.
Tuberitina sp.
Tuberitina bulbacea sp.
Age: Zone 23?; Middle Pennsylvanian (Desmoinesian)

Section 6 (Fig. 23); 9.6 km along strike to the east from Girty Creek. The Otto Fiord Formation at this location is about as thick (360 m) as it is at Section 5. Although the Borup Formation beneath the Otto Fiord does not outcrop at Section 6, it is estimated from lithological comparisons made with Section 5 that the Borup Fiord - Otto Fiord contact is about 6 m below the base of Section 6. Fossil specimens have been recovered only from near the base of this section because the upper part of the formation is exposed in an inaccessible cliff. Calcareous foraminifers and algae from near the base of the Otto Fiord were identified by Mamet (in Nassichuk, 1975) as follows:

GSC locality C-20460; 0.6 m above the base of the exposed section or an estimated 6.6 m above the true base of the Otto Fiord Formation:

Asphaltina sp.
Asteroarchaediscus sp.
Asteroarchaediscus gnomellus Brenckle
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Biseriella sp.
Biseriella parva (Chernycheva)
Calcisphaera laevis Williamson
Calcisphaera pachysphaerica (Pronina)
Climacammina sp.
Earlandia sp.
Endothyra sp.
Eostaffella sp.
cf. *Eostaffellina* sp.
Neoarchaediscus sp.
Neoarchaediscus parvus (Rauser-Tschernoussova)
Neoarchaediscus parvus regularis (Rauser-Tschernoussova)
Palaeotextularia sp.
Planospirodiscus taimyricus Sossipatrova
Planospirodiscus sp.
Pseudoendothyra sp.
Stacheoides sp.
Tetrataxis sp.
Zellerina sp.
Zellerina discoidea (Girty)
Age: Zone 18; Late Mississippian (Chesteran)

GSC locality C-20462; 3.3 m above base of the section:

apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Biseriella parva (Chernycheva)
Calcisphaera sp.
Earlandia sp.
Endothyra sp.
Eotuberitina sp.
Pseudoglomospira sp.
Age: Zone 18; Late Mississippian (Chesteran)

GSC locality C-20507; 74 m above base of the section:

apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Calcisphaera sp.
Earlandia sp.
Endothyra sp.
Neoarchaediscus sp.
Tetrataxis sp.
Tetrataxis of the group *T. conica* Ehrenberg emend von Möller
Age: $\geq$ Zone 18; Late Mississippian (Chesteran) or Early Pennsylvanian (Morrowan)

Blue Mountains Belt

Sedimentology

A narrow belt of Otto Fiord rocks parallels the Blue Mountains Thrust and outcrops discontinuously for some 38.4 km from Hare Fiord Diapir in the southwestern Blue Mountains to Wood Glacier, midway between Mount Schuchert and Mount Barrel (Fig. 6) in the northeast. This outcrop belt, which is outlined on geological Map 1311A (Greely Fiord west; Thorsteinsson, 1971a), is approximately parallel to a previously discussed belt along the north side of Hare Fiord extending from van Hauen Pass easterly beyond the head of Hare Fiord. Whereas Otto Fiord rocks in the latter area are on the northwest side of the Basinal Clastic and Evaporitic Belt, those in the Blue Mountains are on the southeast side.

Throughout the long but discontinuous 'Blue Mountains Belt', the lower contact of the Otto Fiord Formation is covered and rocks beneath the Otto Fiord have not been examined. Relationships between the Otto Fiord Formation and overlying strata, however, can be assessed more readily. The Otto Fiord invariably is overlain by the Hare Fiord Formation in the 'Blue Mountains Belt'. In some places, the lower part of the Hare Fiord contains argillaceous limestone, shale and siltstone, as is the case at van Hauen Pass, but in many places the Otto Fiord is overlain by the 'Tellevak Limestone Member' of the Hare Fiord Formation. This member is composed of Middle Pennsylvanian bryozoan mounds or reefs.

Near the eastern extremity of the 'Blue Mountains Belt', the Otto Fiord is exposed in a structurally complex terrane which clearly was affected by the northeasterly extension of the Blue Mountains Thrust fault, which in turn was controlled apparently by detachment within evaporite units in the Otto Fiord Formation (Secs. 7, 8; Fig. 6) a few kilometres west of Wood Glacier in an eastward-trending valley about 9 km east-northeast of Mount Schuchert. Another incomplete section (Sec. 9) was measured near the southwestern extremity of the Blue Mountains.

Structure in the Wood Glacier area is an east-northeast trending, faulted, asymmetric anticline with the northern limb either dipping steeply or (locally) slightly overturned. Beds of the Otto Fiord Formation, particularly the anhydrite units and enclosed thin carbonate beds, have been folded, sheared and faulted. Even though the upper and lower contacts of the formation in this area are obscured by faults, it was determined that the formation is overlain variously by thin-bedded argillaceous limestone and siltstone of typical Hare Fiord character or by thick-bedded bryozoan mounds ('Tellevak Limestone') that are local within the Hare Fiord Formation. Curiously, where the Otto Fiord is overlain by carbonate mounds, a succession of pink-weathering sandstones more than 30 m thick and similar to those reported by Greiner (1963) from Malloch Dome, Ellef Ringnes Island, occurs in the upper part of the Otto Fiord Formation. Elsewhere, where mounds are absent, sandstone units are either thinly developed or absent from the Otto Fiord Formation.

Section 7 (Fig. 6); 3.8 km west of terminus of Wood Glacier. Section 7 was measured up the south side of a major east-west valley about 9.5 km east-northeast of Mount Schuchert. The section is described here in two parts, Section 7a and Section 7b, each slightly more than 200 m thick. The two subsections are separated from each other by a major fault and, although most contacts are obscured, the upper subsection is overlain conformably by an argillaceous limestone and siltstone facies of the Hare Fiord Formation. Paleontological data from throughout the two subsections provide a correlation across the fault that indicates a repetition in the upper part of the Otto Fiord section. In spite of this apparent repetition, no reliable data were obtained concerning the possibility of local lateral displacement of strata. Since both sections complement each other and show different lithological features because of differences in preservation or possibly local differences in depositional environment, the sedimentology of each is described separately.

Section 7a includes about 230 m of interbedded anhydrite, limestone and thin sandstone units. About 140 m of this subsection is anhydrite in units of various thicknesses up to about 30 m, whereas the remainder is composed of carbonate units averaging about 4.5 m thick and a few sandstone or finer terrigenous clastic units averaging about 8 m thick. Between 6 and 8 anhydrite-carbonate or anhydrite-sandstone-carbonate cycles probably are present in this incomplete section.

As in the type section of the Otto Fiord Formation, the proportion of bioclastic or skeletal components in carbonate units in Section 7a increases upward. The lowermost carbonate beds are dark, thin bedded to lensoid in form, and contain a small percentage by volume of ostracodes, a few echinoderm clasts, small brachiopods, tubular and encrusting foraminifers, a few chambered foraminifers and a variety of algae. Pelletoidal grains are common and a few units are composed almost entirely of fecal pellet grainstone and packstones enclosed in anhydrite. Most carbonate units in Sections 7a and 7b contain stylolites and, in places, are severely fractured; these fractures now occur as calcite-filled veinlets.

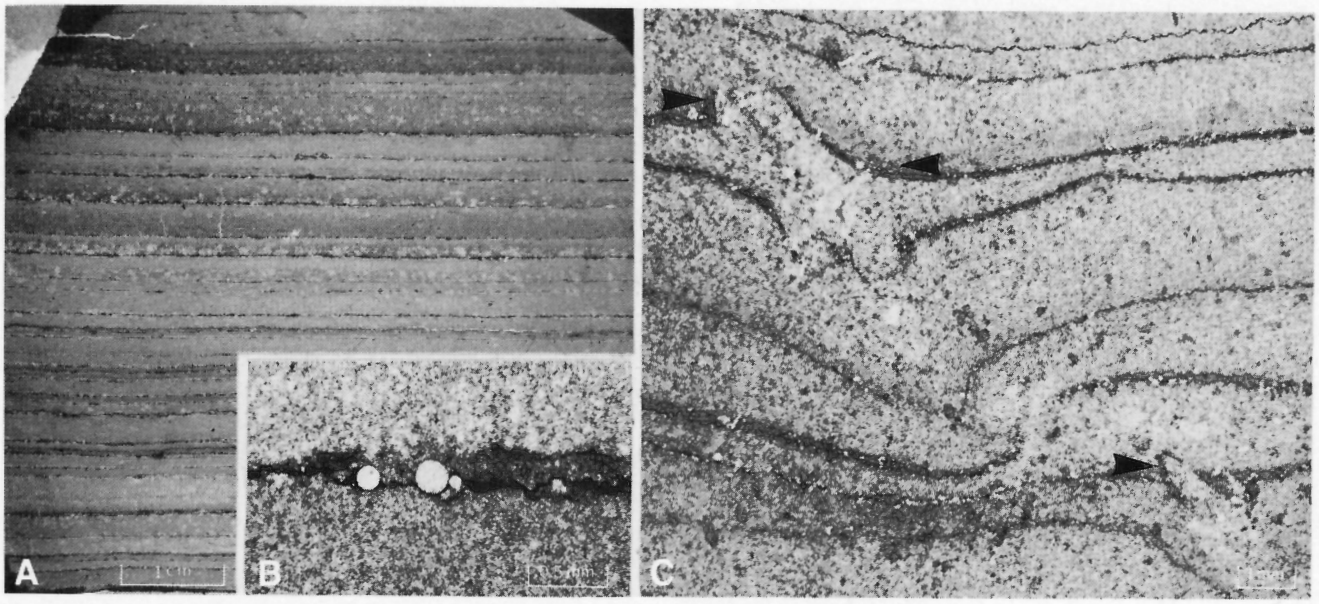
The uppermost carbonate beds in Section 7a show a lithological sequence similar to the idealized cycle or rhythm described earlier for the carbonate units in the type area for the Otto Fiord Formation at van Hauen Pass. Some of these stratigraphically higher carbonate units are underlain by medium- to fine-grained, calcite- and silica-cemented crossbedded sandstones. These sandstones are, in turn, underlain or interbedded with green calcareous mudstone and siltstone, which are either flat laminated or ripple laminated. The generalized lithological sequence within the upper carbonate bed is:

Base: Black, laminated micritic lime mudstone to wackestone with very few bioclasts; disseminated pyrite.

Middle: Pelletoidal, bioclastic algal wackestone to boundstone, with a lower percentage by volume of clasts; usually not dolomitized; dominated in places by beresellid or other dasycladacean and red algae. Brachiopods, a few crinoids, encrusting and other foraminifers, ostracodes, and a few gastropods are associated with the algae in some places, and these may form an interconnected growth network suggesting a boundstone fabric.

Top: Pelletoidal, bioclastic, oncoidal wackestone to packstone, 50 to 65 per cent grains by volume, with a silty micritic matrix in part extensively replaced by very finely crystalline dolomite. Bioclasts, in decreasing volume, include echinoderms (crinoids), bryozoans, brachiopods, various algae (beresellids, kamaenids, red algae), encrusting, tubular and chambered foraminifers, ostracodes, gastropods and a few spicules. Oncoids are irregular to ovoid in shape and, rarely, are nucleated on gastropods, and with an algal component in the envelope. Authigenic feldspar is a common diagenetic component within the oncoids.

Section 7b was measured well above a thick zone of deformation that separates it from 7a. This subsection includes about 205 m of interbedded anhydrite, carbonate sandstone and other detrital sediments in the upper Otto Fiord Formation, and passes upward into deep-water argillaceous limestone, siltstone and shale of the overlying Hare Fiord Formation. About 800 m west of this subsection a large Pennsylvanian bryozoan carbonate mound ('Tellevak Limestone Member' of the Hare Fiord Formation) directly overlies the top of the Otto Fiord Formation and is overlain



A. Calcitic laminite from the top of the Otto Fiord Formation at Section 2. Note the concentration of inclusions at interlaminar boundaries; most boundaries now are microstylolites. Darker laminae near the top of the photomicrograph contain bitumen inclusions in intercrystalline micropores.

B. Organic microspheres of unknown biological affinity and other flocculent organic detritus at the base of a calcitic lamina from Section 7b.

C. Calcite-replaced pseudomorphs (arrows) of gypsum crystals that grew by displacement in the original laminated sediments. Compare with Fig. 19 of Hardie and Eugster (1971) for possible Miocene gypsum analog.

Figure 30. Lithological characteristics of calcitic laminites.

by shales, argillaceous limestone and other Hare Fiord rocks that include carbonate turbidites above the flanks of the mound.

The upper part of the Otto Fiord Formation in Section 7b contains about six recognizable carbonate-anhydrite-mudstone and sandstone cycles, and terminates just below the Hare Fiord Formation contact with a thick unit of calcitic laminite. The two thick anhydrite-sandstone-carbonate cycles in the upper part of this section appear to correspond to the thick cyclic units of similar lithology to be described in the Wood Glacier section (Sec. 8). However, the thick, laminated mudstone, calcareous mudstone and shale, which characterize the lower part of each detrital sediment cycle in the Wood Glacier section, appear to be missing or severely attenuated in Section 7b, which is only slightly more than 1 km distant. Apart from this difference, other major characteristics of these thick sandstone-carbonate units are similar to those to be described for the better exposed Wood Glacier section.

A limestone unit 3 m thick in the lower part of the upper sandstone-carbonate cycle deserves separate attention. This bed contains tabulate and rugose corals associated with beresellid algal boundstone, and is correlated with a lensoid carbonate bed of beresellid algal boundstone in the Wood Glacier section. The lithological succession in this unit is:

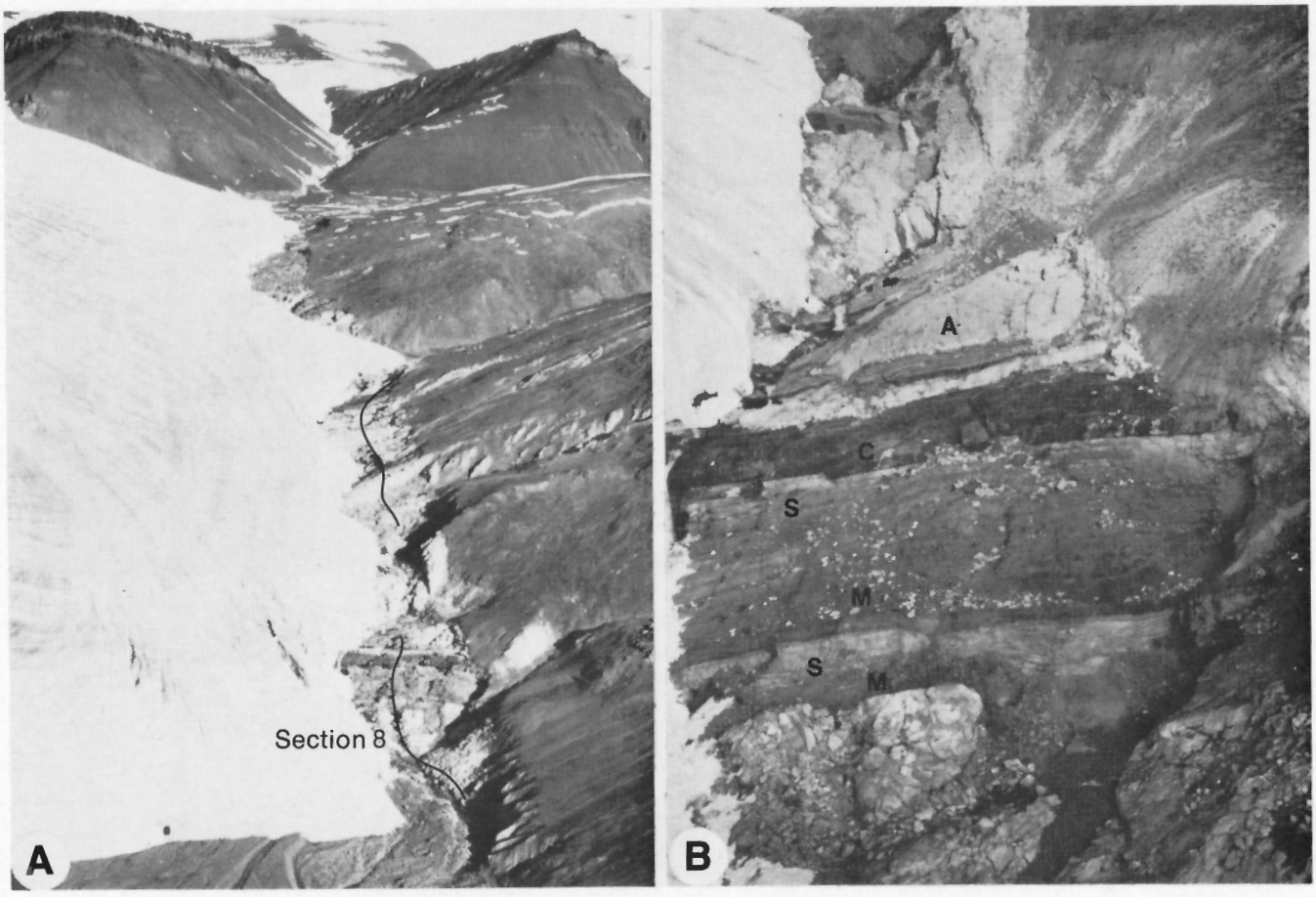
Base: Rounded-bioclast, tabulate-coral, algal grainstone, with early cementation by fibrous cement. Contains up to 3 per cent of authigenic quartz and possibly some feldspar. *Cladochonus*-like tabulate corals are common, overlying or associated with tabulate coral-algal boundstone; same type of tabulate corals enclosed in a growth network of beresellid and red algae; associated tabular and chambered foraminifers; 2 per cent of authigenic quartz.

Middle and upper: Pelletoidal, bioclastic dasycladacean-algal packstone to boundstone, with beresellid algae dominant; few brachiopods, tubular and chambered foraminifers, overlying rugose coral and algal boundstone, with some macrocrystalline, late diagenetic dolomite; beresellid and other dasycladacean algae are abundant, forming a growth network in which solitary(?) rugose corals are enclosed. Fusulinacean and other foraminifers are present.

Top: Thin layer of selectively silicified, partly dolomitized (macrocrystalline, late diagenetic dolomite) algal boundstone, with a few encrusting foraminifers within the algal layers; this structure appears 'stromatolitic' yet is formed largely of skeletal or precipitated carbonate rather than of bound particulate sediment.

The uppermost 30 m of Section 7b, above the two thick anhydrite-sandstone-carbonate cycles, are composed of calcareous silty sandstone overlain by about 4.5 m of poorly exposed carbonate terminating in an oncoidal and bioclastic wackestone containing an exceedingly varied biota, and overlain, in turn, by a thick unit of calcite laminite. Discontinuous outcrops of laminite occur through about 24 m of section. These laminites are similar to the laminites at the top of the Otto Fiord Formation in the type area; laminites from several localities are illustrated in Figure 30.

The laminites are characterized by very even parallel laminae that appear to be laterally persistent (Fig. 30A). They are composed of subhedral calcite crystals, which tend to vary slightly but persistently in size within each lamina; mean size for larger crystals lies in the range 40 to 60 μm. Normal and reverse grading in crystal size is found in the laminae. Boundaries between laminae are defined by dispersed opaque sulphide grains, organic staining, and by euhedral authigenic quartz crystals, many or all of which



A. General view of structurally complex section. GSC 199248.

B. Closer view of a thick mudstone-sandstone-carbonate (M-S-C) unit that corresponds with the upper part of the reconstructed measured section; upper carbonate is overlain by folded anhydrite (A) with carbonate interbed. GSC 199246.

Figure 31. Exposure of the Otto Fiord Formation at Section 8, marginal to a tongue of Wood Glacier in the van Royen Ridges (Fig. 6).

probably are nucleated on detrital quartz grains; detrital matter occurs at the base of the laminae. Lamina thickness varies but, in the lower part of the laminite sequence, it ranges between less than 1 mm to about 2 mm. Many interlamina boundaries are stylolitic.

The uppermost laminite contains disrupted laminae, some of which retain the pseudomorphic ghosts of gypsum crystals cutting across them (Fig. 30C). The crystals were inclined at a steep angle to the laminae; this inclination is similar to the inclination of apparent gypsum pseudomorphs in bedded anhydrite in the type area for the Otto Fiord Formation. The gypsum pseudomorphs at the top of the laminite sequence provide one argument for placing the top of the Otto Fiord Formation at the top of the laminite unit (which also is at the top of section in the type area) rather than at the top of the uppermost anhydrite unit.

Section 8 ('Wood Glacier Section'); 1.6 km west of Wood Glacier. A composite section of the Otto Fiord Formation was measured 1.6 km west of the western margin of Wood Glacier (Fig. 6). The section is located in a thin strip of Otto Fiord and younger rocks exposed by lateral ice retreat along the edge of a narrow, southerly directed, unnamed glacier fed from the same ice field as Wood Glacier (Fig. 31). Ice cliffs approximately 15 m high form the western limit of exposure. Rocks in this area are folded and faulted, reflecting the influence of the Blue Mountains Thrust and minor related faults. Deformation is particularly intense within anhydrite units of the Otto Fiord Formation; distorted limestone beds within the deformed anhydrites reflect the severity of deformation.

The section includes about 405 m of Otto Fiord rocks in continuous section but both the lower and upper contacts of the formation are covered. An additional 53 m of Otto Fiord strata were measured directly below the base of the 'Tellevak Limestone', a succession of moundlike bryozoan carbonates that locally form the lower part of the Hare Fiord Formation. However, this 53 m segment of Otto Fiord strata cannot be fitted into sequence with the larger section in the same general area, nor can it be related to the upper part of the nearby Section 7 in which the Otto Fiord Formation grades transitionally upward to 'typical' Hare Fiord strata where mounds of the 'Tellevak Limestone' are absent. The contact between the top of this short section and the base of the 'Tellevak Limestone' mound appears to be a fault or series of faults. However, although this area is complicated structurally, the Otto Fiord section exposed is important because of its contribution to an understanding of the biostratigraphy, facies, sedimentology and depositional origins of the formation.

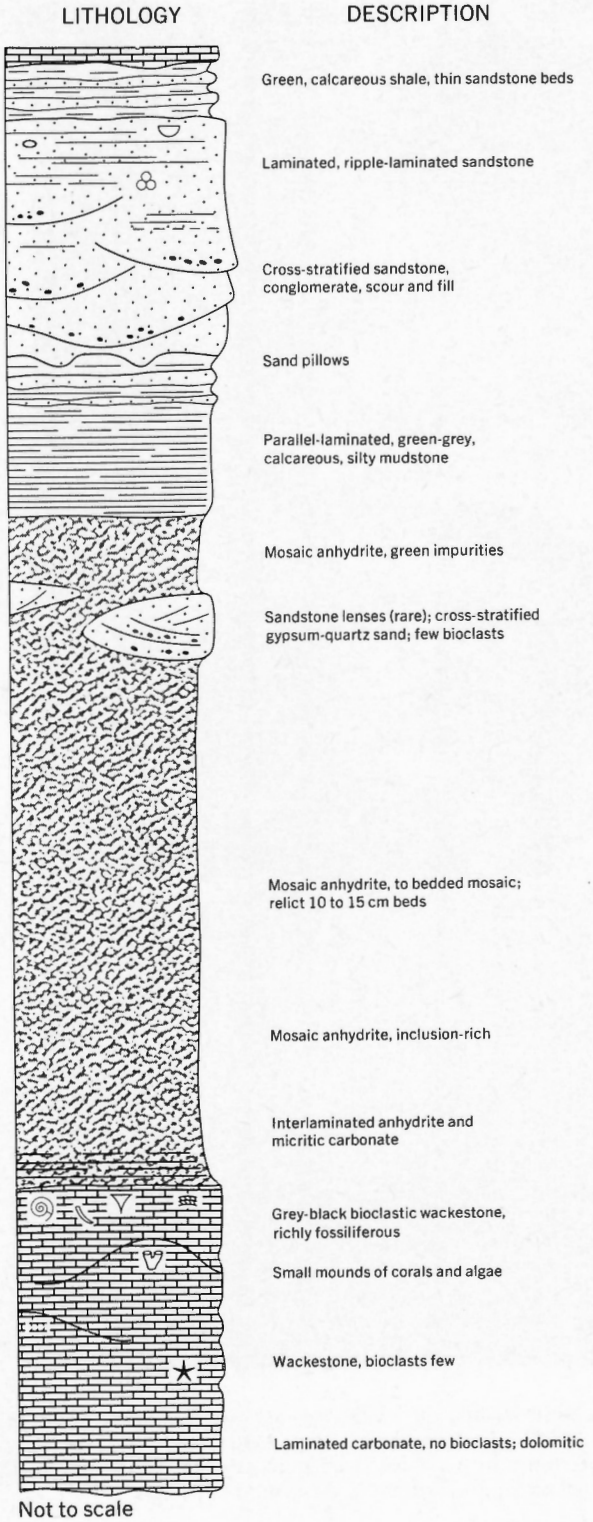


Figure 32. Generalized carbonate-anhydrite-mudstone-sandstone cycle for the Otto Fiord Formation in Section 8.

The lowest exposures in Section 8 are contorted, south-dipping carbonates about 1070 m north of the terminus of the glacier flanked by the section. The base of this carbonate unit is fault bounded and is underlain by a structurally complicated zone of recessive anhydrite and other rock types in a strike valley. The section extends from the basal carbonate updip toward the toe of the glacier to the axis of a synclinal fold. An auxiliary section comprises stratigraphically higher units from north-dipping beds on the south flank of the syncline, measured northward from the toe of the glacier.

At least five cycles of anhydrite and carbonate, or anhydrite, mudstone, sandstone and carbonate are present in this section (Fig. 32). Within the upper two thick and complex cycles, several incomplete subcycles also can be defined. In the lower 140 m of the section carbonate and anhydrite units are dominant and shale and sandstone form only subordinate lithological components. The carbonate units are from less than 1 m to 15.5 m thick; anhydrite units are 4 to 46 m thick. The thickest carbonate-anhydrite cycle in this part of the section is 64 m, with a thickness ratio for carbonate:anhydrite of 1:3.

The lithological succession within these carbonate units is similar to that found in the limestone units of the Otto Fiord Formation in the type area. The following sequence applies to most of the carbonate units in the lower part of Wood Glacier section.

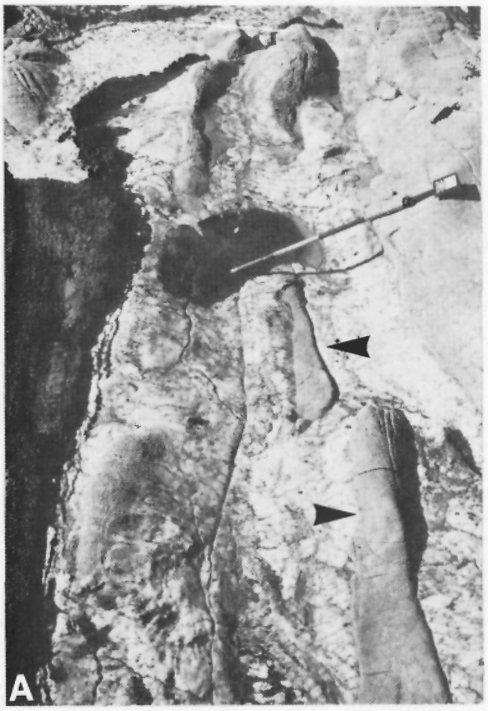
Base: The contact with underlying anhydrite is abrupt. The lowermost beds are very dark grey to black, argillaceous to silty limestone with disseminated pyrite. The beds, in places, are irregularly laminated, and are either unfossiliferous or contain a few foraminifers and ostracodes of hypersaline aspect. Replacive anhydrite crystal aggregates weather out to leave surface vugs.

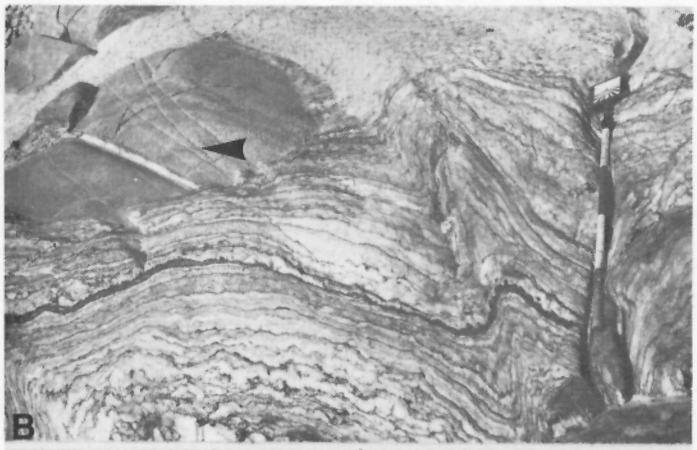
Middle: These beds consist of argillaceous limestone, containing a more diverse biota than underlying beds; bioclasts include crinoids, ammonoids, brachiopods, gastropods, ostracodes, spicules and a variety of foraminifers including fusulinaceans. Algae, particularly beresellids, are abundant. Limestone textural types are pelletoidal, intraclastic and bioclastic wackestone and packstone. Some beds are irregular, but generally the beds are wavy and lenticular with shaly partings. Stylolites are common.

Upper: These are argillaceous limestone, generally of bioclast-rich wackestone and packstone with an abundant and diverse biota of crinoids (most abundant; includes articulated calyxes), brachiopods, trilobites, gastropods, bryozoans, ostracodes and foraminifers (but not fusulinaceans). Algal oncoids are common in the uppermost beds. Textures include wackestone and packstone. In some beds approximately 60 per cent of the rock matrix is dolomitized. Authigenic quartz and feldspar replace some of the algal oncoids.

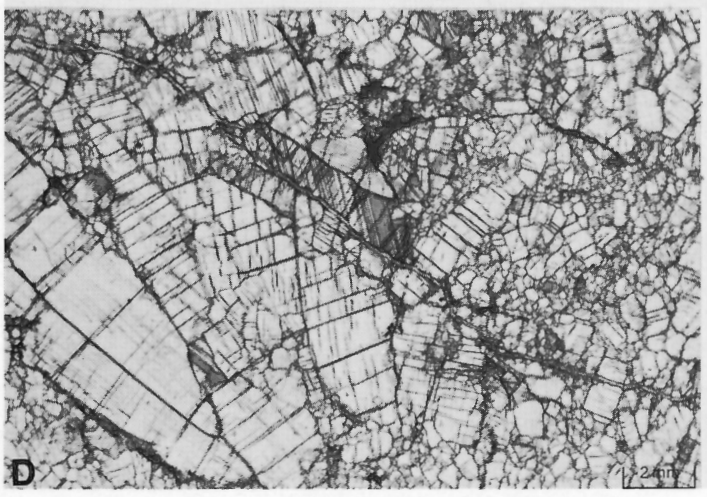
Carbonate-anhydrite transition: The contact between carbonate and overlying anhydrite in the Wood Glacier section usually is transitional. At some places the transition zone includes interlaminated anhydrite and limestone, or a zone of calcite laminite or lime mudstone. Replacive aggregates of anhydrite and dolomite crystals are present in this zone; the anhydrite crystals are hydrated and dissolved at exposed surfaces to leave vugs.

Anhydrite: Nodular mosaic anhydrite is the most common fabric type in the anhydrite units of Section 8. Traces of bedding at 10 to 15 cm intervals, similar to anhydrite units in the type area, appear several metres from the base of each major anhydrite unit. The crystal-pseudomorph fabrics found in some of the anhydrite beds in the type area have not been seen in the Wood Glacier section, but the similarity in bedding thickness and style suggests a similar depositional control. Bedding planes are indicated by concentrations of limestone granules and other impurities. The internal fabric of the beds is nodular mosaic anhydrite (see Fig. 16A).











A. Deformed anhydrite and limestone bed within a thick unit of anhydrite. Anhydrite is weathering differentially in two forms: in resistant lenticular pods (arrows); and in more normal, slightly recessive profile. Traces of inclusion-defined nodular mosaic fabrics are found throughout both of these weathering forms. Their differences in crystal fabric are illustrated in 33D and E. Field scale in feet (1 ft. = 0.3 m).

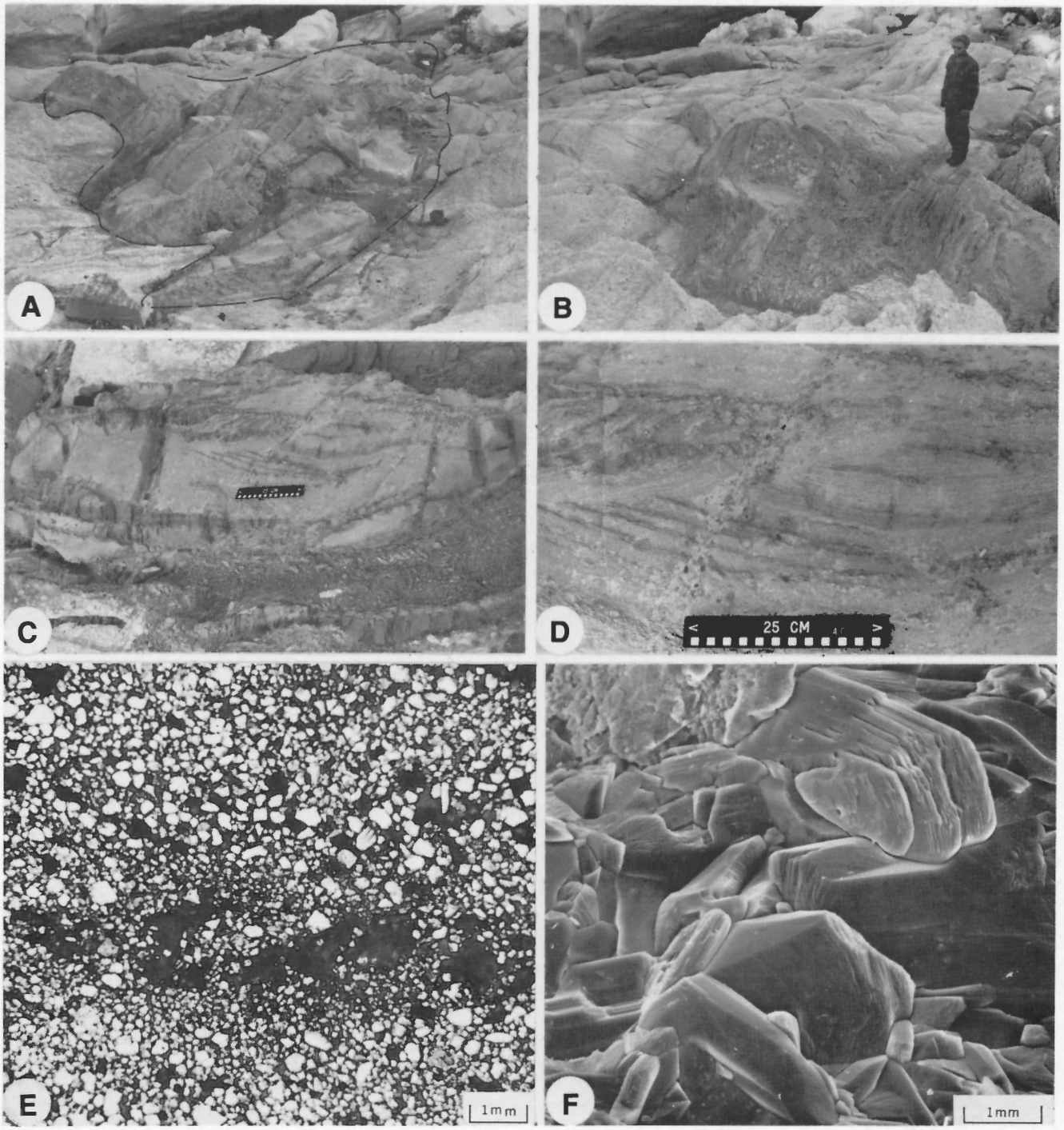
B. Deformed and pseudolaminated anhydrite with an isolated resistant pod (arrow) cut by a dilational fracture.

C. Intensely deformed and sheared anhydrite and structurally brecciated limestone bed. Blocks of limestone have been rotated and displaced in near-random orientation, with sheared and flow-laminated anhydrite enclosing each block.

D. Photomicrograph of resistant anhyarite pod similar to those in 33A. Anhydrite has recrystallized under tectonic stress into very large porphyroblastic crystals with localized zones of smaller, mechanically deformed crystals.

E. Photomicrograph of normal-weathered anhydrite that surrounds resistant, coarsely crystalline pods. Same scale as 33D. Lineation is poorly defined. Differences in resistance to weathering obviously are the result of the slower weathering rate of the larger crystals in the resistant pods.

Figure 33. Structurally deformed carbonate and anhydrite in the Otto Fiord Formation in Section 8.



A. General view of a detrital sulphate-quartz sand body enclosed in anhydrite in the upper part of Section 8. Scale is provided by comparison with 34B.

B. Closer view of the upper part of the sand body shown in 34A, illustrating the enclosing nature of the adjacent anhydrite; the anhydrite retains a trace of bedding in otherwise nodular mosaic fabric.

C. Trough crossbedded detrital sulphate-quartz sand in an isolated pod in anhydrite. Sandstone is underlain by thin lenticular bed of recessive shale.

D. Closer view of 34C showing scour and trough crossbedding. More resistant laminae appear to contain more quartz sand. Larger clasts and mudlumps occur at base of scours.

E. Photomicrograph of detrital sulphate-quartz sandstone. Darker clasts are grains and lumps of detrital sulphate, some partly replaced.

F. Scanning electron photomicrograph of detrital sulphate-quartz sandstone from pod similar to 34D. Fabric is composed of interpenetrating sulphate crystals (gypsum, anhydrite?) that grew as cement, possibly in parts as syntaxial overgrowths solution. Clast at upper left probably is a quartz grain with cement overgrowths.

Figure 34. Lithological characteristics of isolated detrital sulphate-quartz sand pods in anhydrite unit of Section 8.

Some anhydrite units contain extremely contorted carbonate interbeds a few tens of centimetres thick (Fig. 33A, B, C). Some of these beds are overturned and brecciated. Rock types include calcite laminites and partly dolomitized oncoidal, bioclastic and oolitic wackestone. Fecal pellets are common in some beds. The enclosing anhydrite also is intensely deformed, weathering with a 'boudinage'-like profile of resistant lenticular pods of coarsely crystalline, porphyroblastic anhydrite set in a recessive matrix of flow-laminated, finely crystalline anhydrite (Fig. 33D, E). Relict, undistorted mosaic fabric is preserved in the lenticular pods of coarse anhydrite. In contrast, all degrees of distortion of mosaic fabrics are found in the sheared and flowed gypsum. End product of this distortional process appears to be a pseudolaminated, inclusion-rich gypsum wrapped around the pods of resistant anhydrite. A complex of metamorphic textures, fabrics and mineralogies thus are present in these structurally deformed sulphates.

The upper 265 m of the Wood Glacier section include two major cycles composed of limestone, anhydrite, mudstone and sandstone. Each major cycle contains several incomplete subcycles. The following simplified lithological succession (Fig. 32) applies in a general way to both major cycles.

Base of the carbonate unit: Lithological cycles are interpreted as beginning with carbonate units, comparable with the carbonate-anhydrite cycles of the lower part of the Wood Glacier section. The principal difference is that the carbonate units in the upper part of the section lie directly and abruptly on sandstone or shale of the underlying detrital units. Limestone beds in the lower part of the carbonate units typically are composed of silty wackestone with a few tubular foraminifers, ostracodes and other bioclasts, rounded carbonate grains and, rarely, ooids. The last are significant since they are one of the very few examples of oolitic sediment in the Otto Fiord Formation, and may mark the inflection point between regression and transgression in these cycles. The oolitic sediments occur in cross-stratified beds.

Middle beds of carbonate units: The basal beds grade upward to a variety of wackestones, packstones and boundstones with more bioclasts and a more varied biota. Algae, particularly beresellids and red algae, are common. Fusulinaceans normally are present at this level and even lower, but are absent from the uppermost beds of carbonate units throughout the Otto Fiord Formation. Rugose or tabulate coral-algal boundstone in small bioherms about 2 m high also are at this horizon in some units.

Upper beds of carbonate units: Limestones in the upper part of carbonate units are black to dark grey, abundantly fossiliferous, bioclastic and oncoidal wackestones and packstones. Biota includes ammonoids, trilobites, brachiopods, crinoids, bryozoans, gastropods, a variety of foraminifers (but not fusulinaceans), ostracodes and some algae; some bioclasts are bored. Some matrix is slightly silty, and partly (selectively) dolomitized. Bedding is wavy to lenticular and contains shaly partings.

Carbonate-anhydrite transition: The transition from the top of the marine bioclastic limestone into overlying anhydrite is marked by a thin bed (less than 30 cm thick) of calcitic laminite with fecal pellets in packstone and grainstone textures but without bioclasts.

Anhydrite unit: The lowermost part of anhydrite units has an unbedded nodular mosaic fabric, but each grades upward through a few metres into bedded anhydrite. Bedding traces recorded by dark impurities are spaced 10 to 15 cm apart. Internal structure of beds again is a nodular mosaic fabric, without distinct crystal pseudomorph fabrics. Anhydrite units in the upper part of the Wood Glacier section are as much as 56 m thick.

About 8 m above the base of the lower of the two major anhydrite units in the upper Wood Glacier section, a series of resistant, crossbedded sandstone pods (Figs. 32, 34) are enclosed in the anhydrite. The lenticular pods are up to 3.5 m thick (Fig. 34A, B); several pods occur along strike at slightly different stratigraphic levels, but all are enclosed within anhydrite. Large, low-angle crossbeds and scour structures characterize the internal structure of the pods (Fig. 34C, D). Compositionally, the pods are calcite- and gypsum-cemented sandstone, with subangular quartz grains (Fig. 34E) in size-graded beds or laminae. The largest grains are 400 µm (mean) in diameter and the mean size for all grains lies in the 100 to 160 µm range. Opaque grains and 'heavy mineral' grains are concentrated in some thin laminae; most of the grains are rounded, and yellowish in reflected light. Larger lithoclasts (intraclasts?) of gypsum (brown in transmitted light) are common. One sample contains a coral or bryozoan skeletal fragment.

The calcite cement is poikilotopic in habit and appears to have corrosion contacts with the quartz grains. The gypsum cement occurs as interpenetrating fibrous crystals (Fig. 34F), which have corrosion-contact relationships with quartz grains.

These lenticular, crossbedded, gypsum-lithoclastic quartz sandstones with gypsum and calcite cements are unique to this part of the section. They are the first sign of the change from evaporitic to detrital sedimentation within the depositional cycle.

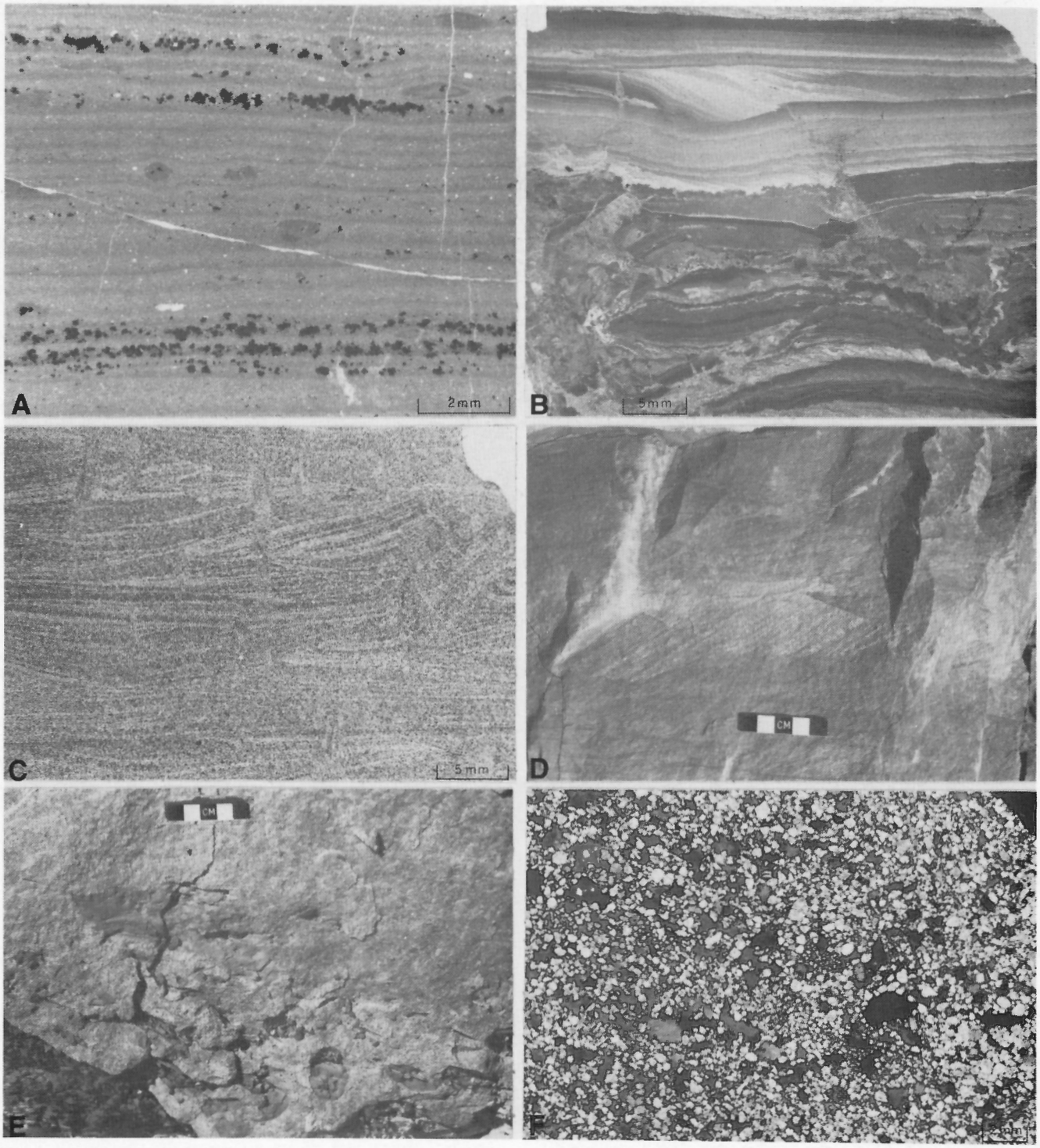
Basal part of detrital sediment unit: Anhydrite in the upper part of the Wood Glacier section is overlain gradationally or abruptly by a unit of fine- to coarse-grained detrital rocks coarsening upward. Commonly the basal part of this unit is a thick sequence of green, parallel-laminated to flat-bedded mudstone, calcareous mudstone or calcareous shale (Fig. 35A); calcareous mudstone is dominant. There are no bioclasts. Disseminated crystalline pyrite is common, normally concentrated at regular intervals within laminae. Some ripple crosslamination occurs at the top of some units (Fig. 35B); soft-sediment load structures are found in laminated mudstone below flow-rolled sandstone in overlying beds. Fine-grained, silty sandstone or sandy siltstone in thin beds may be interbedded with mudstone, particularly higher in sequence. Sandstone-siltstone beds often are ripple laminated. Thickness of these units varies from 6 to 15 m.

Middle part of detrital unit: The mudstone grades upward into greenish, crossbedded calcareous sandstone, which is silty, argillaceous and micaceous (Fig. 35C, D). Sand rolls and load structures are found at the base of this sandstone section, which is finer grained than the overlying sandstones of the upper unit. Some of the sandstone beds contain bioclasts of ostracodes, brachiopod fragments, foraminifers (including fusulinaceans) and algae.

Upper part of detrital unit: The sandstones, showing load casts, in the middle of the detrital unit grade upward to coarse- to medium-grained, trough- and crossbedded lithoclastic sandstone (Fig. 35E, F). The sandstone beds contain large crinoid-rich mudstone clasts, and silicified and dolomitized gypsum lithoclasts; the coarsest sands and largest lithoclasts occur as lag deposits at the base of scours and channel structures. Grain size in individual depositional units decreases upward into laminated finer grained sandstone, which may be truncated by a channel or scour of the next overlying fining-upward depositional unit.

Thick crossbed sets occur in some lenticular masses of coarse sandstone. Porous sands of channel and scour origin are cemented by ferroan dolomite.

Top of detrital unit: The detrital unit is capped, in places, by a thin zone of interbedded green laminated shale, shaly mudstone, siltstone and sandstone; generally these are finer sediments and more recessive than underlying units. Some beds contain a few brachiopods, fusulinaceans, tubular foraminifers and ostracodes. Some of the sandstone and siltstone beds are crossbedded or ripple-laminated. Sands are cemented by calcite.



A. Laminated silty mudstone from the basal part of a mudstone-sandstone cycle. Indistinct pellets or clasts within laminae may be fecal pellets or other detrital lumps.

B. Laminated and ripple-cross-laminated siltstone from fine-grained detrital unit below sandstone. Fabric suggests extensive load deformation and early soft-sediment displacement by microfaults.

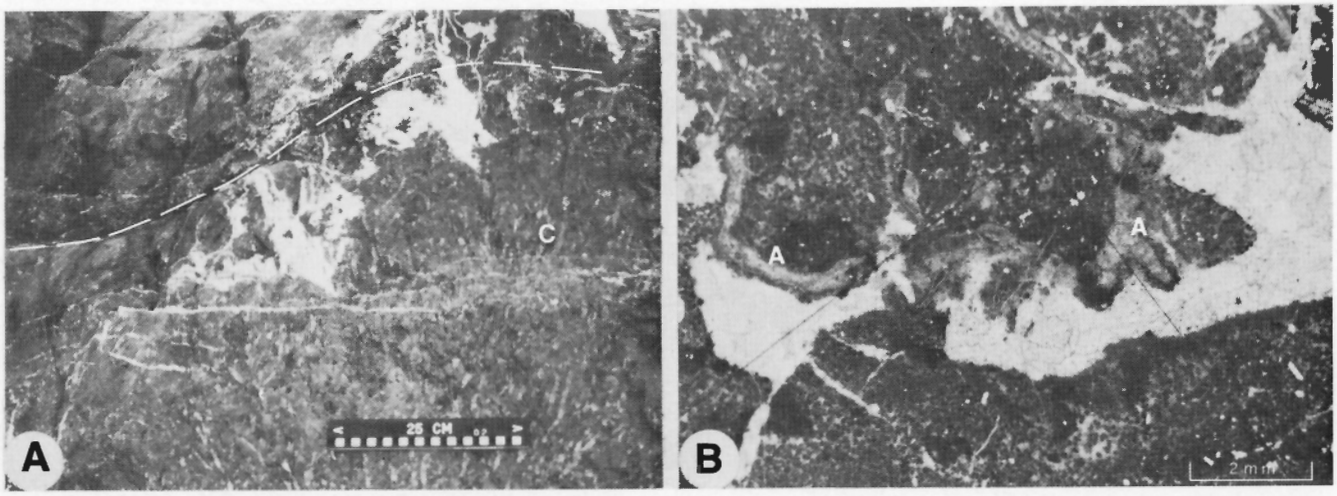
C. Small scour structures in fine-grained sandstone in detrital unit. Vertical structures may be burrows or gas vents.

D. Larger scale crosslamination and scour structures in coarse-grained pink-weathering sandstone in the upper part of a mudstone-sandstone cycle.

E. Large mudstone clasts at the base of a sandstone-filled channel cut in mudstone in the upper part of a mudstone-sandstone cycle.

F. Photomicrograph of coarse sanastone from the upper part of a mudstone-sandstone cycle. Opaque and darker clasts are mudstone lumps and detrital sulphate clasts.

Figure 35. Lithological characteristics of detrital sediment units (mudstone-sandstone) in the Otto Fiord Formation at Section 8.



A. Field photograph of large in-place colonial coral (C) with local relief of about 30 cm, overlain by algal-rich skeletal limestone. Coarse white calcite patches and fracture-fillings are later diagenetic phenomena.

B. Photomicrograph of algal framestone fabric immediately above coral mass. Primary shelter cavity formed below algae (A) has floor of pelletoidal sediment, and is filled by later diagenetic calcite spar.

Figure 36. Small colonial coral and algal mounds in carbonate units of the Otto Fiord Formation in Section 8.

Total thickness of sandstone beds from the middle part of the detrital unit to the top ranges from 9 to 18 m.

Coral-algal mounds in carbonate units: Small mounds constructed of coral and algal framestone fabrics occur in several of the carbonate units in the upper part of the Wood Glacier section (Fig. 36). Relief in the mound during growth was less than 1 m.

In one unit small mounds built by beresellid algae overlie basal beds of wackestone containing agglutinate tubular foraminifers, small brachiopods, spicules and a variety of algae (Fig. 36B). The mounds are composed of in-place framestone networks of beresellid algae and of accumulations of algal clasts with packstone textures; fusulinaceans, chambered and encrusting foraminifers, a few echinoderm spines and abundant pelletoids are other components. Irregular patches of white calcite are common; some fill irregular, sediment-floored 'stromatactis'-like fenestrae. Fibrous calcite cement, some radial in form, is nucleated on some of the beresellid algal framestone and lines some of the fenestrae; this is similar to fibrous cements identified as early submarine in origin in carbonate buildups of overlying formations (Davies, 1977c).

Small coral mounds or discrete colonial coral heads up to 1 m high are found in the upper part of one of the carbonate units (Fig. 36A). These mounds are composed of in-place colonial rugose corals associated with brachiopods, a few crinoids and ostracodes, and beresellid and red algae. The matrix of the mound rocks is selectively and almost totally replaced by dolomite. Between the coral heads, but within the same bed, are tubular foraminiferal-algal boundstones with a variety of algae, small brachiopods, large ostracodes and many pelletoids. The coral bioherms are overlain by beds of partly dolomitized algal packstone and boundstone containing fusulinaceans, other chambered and encrusting foraminifers, and small brachiopods. Red algae, as well as dasycladacean algae are common to abundant in some of these beds, which total about 2 m in thickness.

Section 9 (southern Blue Mountains section); near the southwest extremity of the Blue Mountains 13 km southeast of Hare Fiord, 3.2 km northeast of Hare Fiord Diapir. Section 9 is incomplete and only about 200 m of strata are exposed (Fig. 37). The upper 55 m are poorly exposed, and the base of the section is limited by surficial cover and the Blue Mountains Thrust, which juxtaposed the Otto Fiord Formation over Triassic rocks. A few kilometres south of the section the Otto Fiord Formation is exposed as a large diapiric mass of anhydrite and gypsum 3 km long and 1.5 km wide (Hare Fiord Diapir).

Even though it is poorly exposed the southern Blue Mountains section of the Otto Fiord Formation is significant because it also contains thick units of sandstone interbedded with anhydrite and limestone. Several limestone-anhydrite and limestone-anhydrite-sandstone cycles may be differentiated. The thickest complete cycle is 55 m thick.

Sandstone units (Fig. 38A, D) are thicker and more common higher in the section, increasing sequentially upward from 5 to 8 to 21 m in thickness. Unlike the Wood Glacier section (Sec. 8), thick units of fine-grained detrital sediments and calcitic mudstone below the sandstone units are absent from the southern Blue Mountains. The sandstones are medium to fine grained, green to buff, and variably parallel laminated to crossbedded (Fig. 38C, D, E). Thin beds of green shale occur within and at the base and top of some of the sandstone units. Load casts, sand rolls and other soft-sediment deformation structures are found at the base of some units. Halite hopper crystal moulds and casts are at a few horizons within the sandstones (Fig. 38E). Thin sections reveal that some of these halite crystals grew poikilotopically within fine-grained sands and silts, enclosing the detrital grains with only minor disruption of primary structures.

The sandstone bodies are composed primarily of rounded to subrounded quartz grains with very little obvious feldspar. Pleochroic green mineral grains, other mafic grains and plates of micaceous minerals are other components; some of these may be partly authigenic. The sandstones are cemented by calcite with some addition from authigenic dolomite and other ferroan carbonates. Traces of glauconitelike authigenic minerals also are present.

Limestone units in this section are similar to those in other sections of the Otto Fiord Formation. Some beds contain a varied biota including crinoids, brachiopods, trilobites, gastropods, ammonoids, ostracodes, *Cladochonus*-type corals, foraminifers, spicules, calcispheres and beresellid and other types of algae. The uppermost beds of some units are characterized by oncoidal wackestones similar to carbonate units in the type area and in the Wood Glacier section.

In several thinner beds of limestone enclosed in anhydrite, large cylindrical fecal pellets are the sole clast component (Fig. 38F). These pellets, up to 2.5 mm long by 0.5 mm wide, retain traces of internal 'canal' structures typical of some modern crustacean and other invertebrate fecal pellets. They may be accumulations from a planktonic fauna thriving in surface waters over a hypersaline substratum.

Authigenic feldspar is abundant in some thin beds and in laminites in the lower limestone units of this section. Dolomite replaces as much as 15 per cent of the matrix of some beds.

Anhydrite units in this section are poorly exposed. Nodular mosaic is the dominant fabric in the anhydrite; bedding of the type found in sections in the type area and in the Wood Glacier area was not observed. The thickest continuous anhydrite unit in the section is 30 m thick.

The top of the section is covered, but sandstone float indicates that another thick sandstone unit occurs at the top. About 15 m of covered section separate the uppermost float blocks of sandstone from the (faulted) base of the overlying but steeply dipping bryozoan reef limestones in the 'Tellevak Limestone Member' of the Hare Fiord Formation.

Biostratigraphy

Within the 'Blue Mountains Belt' most fossils from throughout the Otto Fiord Formation, including ammonoids, foraminifers and crinoids, are indicative of an Early Pennsylvanian (Morrowan, Zone 20) to early Middle Pennsylvanian (Atokan, Zones 21, 22) age. Some foraminiferal data, however, suggest that both ends of this range might be extended to include both the Late Mississippian (Chesteran, Zone 18) and the late Middle Pennsylvanian (Desmoinesian, Zone 23).

Mamet (pers. com., 1974) identified foraminiferal and algal species from a single locality in Section 9 (Fig. 6) in the southern Blue Mountains and suggested a "possible late Chesteran" age; included in the collection are representatives of *Asteroarchaediscus*, *Archaediscus*, *Biseriella*, *Neoarchaediscus* and *Planospirodiscus*. The Otto Fiord Formation at Section 9 is overlain by the 'Tellevak Limestone Member' of the Hare Fiord Formation, which has yielded ammonoids as well as foraminifers, including fusulinaceans, of Atokan age (Mamet, in Nassichuk, 1975).

In northeastern regions of the 'Blue Mountains Belt' (Secs. 7, 8; Fig. 6) some fossil data suggest that at least locally the upper part of the Otto Fiord Formation may have been deposited contemporaneously with part of the 'Tellevak Limestone Member' of the Hare Fiord Formation. Mamet (pers. com., 1976) has identified foraminifers, including fusulinaceans of Desmoinesian, Zone 23 age from "*Glomospira* reefs" near the top of the Otto Fiord Formation in Sections 7 and 8. Strata of comparable age are known also from the top of the Otto Fiord Formation east of Hare Fiord. Of particular significance, however, in the Wood Glacier area is that in Section 7 older pre-Desmoinesian (i.e. Atokan) fossils occur in the lower part of the 'Tellevak Limestone Member' of the Hare Fiord Formation, which generally overlies the Otto Fiord elsewhere. This apparent discrepancy is difficult to reconcile stratigraphically because faulting appears to be particularly intense near the top of the Otto Fiord Formation west of Wood Glacier. It is clear, however, that sedimentological interpretations based on these paleontological data may be premature and that additional paleontological and stratigraphic investigations are required to clarify relationships between the Otto Fiord and Hare Fiord formations near Wood Glacier.

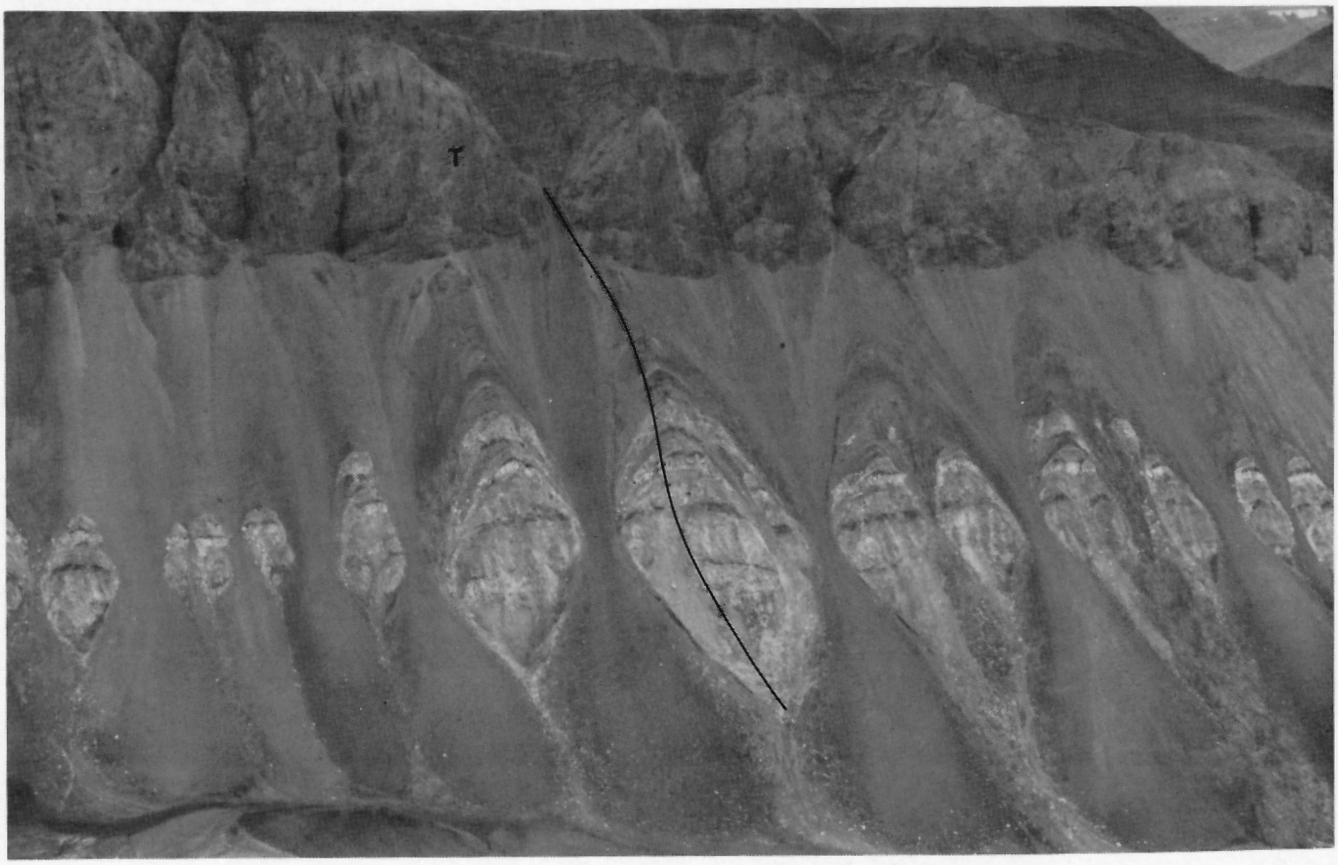
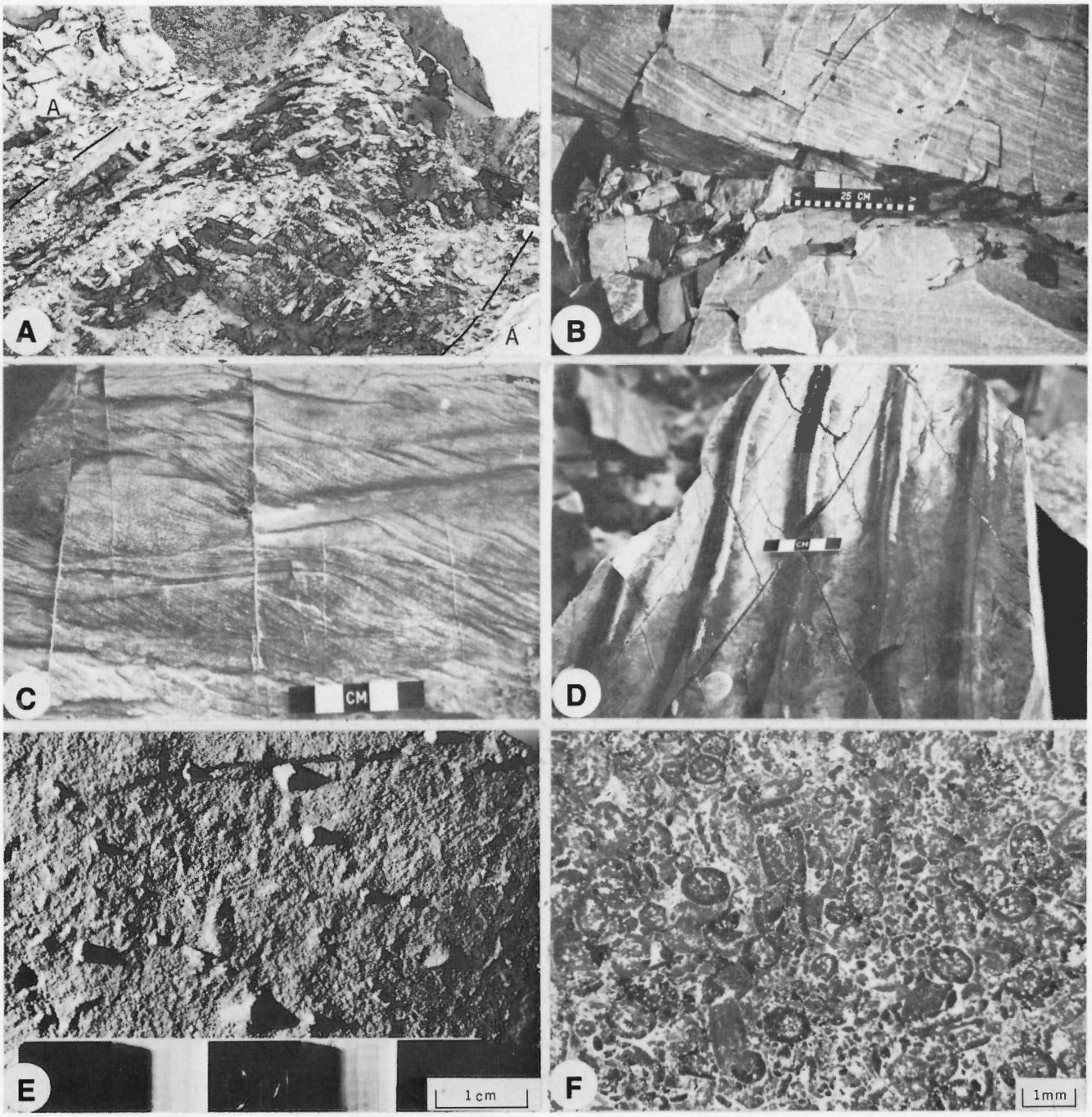


Figure 37. Exposure of the Otto Fiord Formation near Section 9 at the southern end of the Blue Mountains (Fig. 6). The poorly exposed Otto Fiord rocks in this area are truncated at the base by the Blue Mountains fault, which places Pennsylvanian rocks over Triassic (foreground). The resistant limestone mass (T) above the covered interval at midground is a thin remnant of the 'Tellevek Limestone' at the base of the Hare Fiord Formation. About 1 km to the right (south), the Otto Fiord evaporites are exposed in Hare Fiord Diapir. GSC 199264.



A. Field view of sandstone unit interbedded between anhydrite (A) in the upper part of Section 9. Lower part of unit is more shaly and less resistant than upper part.

B. Large-scale crossbedding and scour structures in medium to coarse grained sandstone.

C. Ripple-crosslaminae in sandstone, some in climbing sets.

D. Flat-topped symmetrical ripples on sandstone bedding plane.

E. Casts of halite hopper crystals in sandstone.

F. Fecal pellets or algae in limestone interbedded in anhydrite in the lower part of Section 9. Grains of this type appear to be characteristic of thin limestone beds of hypersaline aspect that are interbedded within thick anhydrite units of the Otto Fiord Formation in all sections.

Figure 38. Characteristic lithologies of Otto Fiord rocks in Section 9.

From near the top of the Otto Fiord Formation at Section 8 (Fig. 6), Mamet (pers. com., 1975) identified *Beresella* sp., *Bradyina* sp., *Calcisphaera* sp., *Climacammina* sp., *Endothyra* sp., *Globivalvulina* sp., *Komia* sp., *Neostaffella* sp., *Ozawainella* sp., *Pseudostaffella* sp. and *Tetrataxis* sp., and commented: "This is certainly the *Neostaffella* level (Moscovian, Desmoinesian Zone 23). Nowhere in the world has an *Asteroarchaediscus-Neoarchaediscus-Planospirodiscus* fauna been described above this level." The 'Tellevak Limestone Member' of the Hare Fiord Formation is about 220 m thick in the immediate vicinity of Section 8 west of Wood Glacier and *Asteroarchaediscus*, *Neoarchaediscus* and *Planospirodiscus* are abundant in faunas that are indicative of an Atokan, Zone 22 age in the lower and middle parts of the unit. In fact, according to Mamet (pers. com., 1974), faunas that are comparable in age to faunas at the top of the Otto Fiord in Section 8, that is Desmoinesian, are encountered only in the upper third of the 'Tellevak Limestone'.

Fusulinaceans from near the top of the Otto Fiord Formation in Section 8 and from the lower part of the 'Tellevak Limestone' in the same general area have been identified by Ross (pers. com., 1975). According to Ross, *Pseudostaffella gorskyi* (Dutkevitch) and *Eostaffella kashira* var. *rhomboides* Rauser from the upper part of the Otto Fiord indicate a Moscovian (Kashirian) age and the upper part of Kashirian may be equivalent to Desmoinesian. Ross also indicated that the following fusulinaceans from the lower 'Tellevak Limestone' are either the same age or slightly older than those in the uppermost part of the Otto Fiord Formation: *Eoschubertella* cf. *E. obscura* var. *mosquensis* Rauser, *Profusulinella (Aljutovella)* cf. *P. (A.) cybae* Lentovich, *Staffella (Parastaffella)* sp., *Pseudostaffella* ex gr. *P. gorskyi* (Dutkevich).

Although probable late Middle Pennsylvanian (Desmoinesian) fossils occur in the upper part of the Otto Fiord Formation near Wood Glacier and also east of the head of Hare Fiord, the formation is no younger than Early Pennsylvanian (Morrowan) or early Middle Pennsylvanian (Atokan) elsewhere in the Sverdrup Basin. It has been indicated previously that in the type area near van Hauen Pass conodonts and ammonoids of Atokan age occur near the bottom of the Hare Fiord Formation just above the Otto Fiord. Similarly, at Stepanow Creek on the north side of Hare Fiord (Fig. 6) Mamet (in Nassichuk, 1975, p. 188) has identified an abundant fauna of *Asteroarchaediscus* sp., *Planospirodiscus* sp. and numerous other foraminifer species indicative of an Atokan (Zones 21, 22) age in Hare Fiord strata just above the top of the Otto Fiord. Considerably farther south on Axel Heiberg Island, near the head of Whitsunday Bay, foraminifers from the Otto Fiord Formation indicate a Morrowan (Zone 20) age.

Section 7 (Fig. 6); 3.8 km west of terminus of Wood Glacier. Section 7 has been described previously as two contemporaneously deposited parts or subsections of a repeated section, each subsection with complementary sedimentological features. Both subsections occur at the generalized locality given for Section 7 above. Foraminifers and algae have been identified from both subsections by Mamet (pers. com., 1975) as follows:

Section 7a

GSC locality C-21142; 9.6 m above the base of Section 7a:
Asteroarchaediscus sp.
Biseriella sp.
Neoarchaediscus sp.
Radiosphaerina sp.
Tetrataxis sp.
Tuberitina sp.

GSC locality C-21144; 42.6 m above the base of Section 7a:
apterrinellids
Bradyina sp.
Eotuberitina sp.
Pseudoglomospira sp.
Tetrataxis sp.

GSC locality C-21151; 145.2 m above the base of Section 7a:
Bradyina sp.
Climacammina sp.
Endothyra sp.
Globivalvulina sp.
Neostaffella sp.
Pseudoglomospira sp.

Mamet commented on these collections: "The lowermost collection (C-21142) in Subsection 'a' contains abundant Archaediscidae indicative of a Morrowan-Atokan age. The uppermost collection (C-21151) is rich in *Pseudoglomospira* and contains *Neostaffella*. It is probably near the Atokan-Desmoinesian boundary, i.e. Zone 22 - ? Zone 23."

Section 7b

GSC locality C-21165; 22.5 m above the base of Section 7b:
Asteroarchaediscus sp.
Eotuberitina sp.
Neoarchaediscus sp.
Tetrataxis sp.

GSC locality C-21168; 63.6 m above the base of Section 7b:
Asteroarchaediscus sp.
Earlandia sp.

GSC locality C-21172; 80 m above the base of Section 7b:
Asteroarchaediscus sp.
Eotuberitina sp.
Neoarchaediscus sp.
Tetrataxis sp.

GSC locality C-21175; 116 m above the base of Section 7b:
Beresella sp.
Endothyra sp.
Komia sp.
Tetrataxis sp.

GSC locality C-21177; 116.5 m above the base of Section 7b:
Beresella sp.
Endothyra sp.
Eotuberitina sp.
Komia sp.
Ozawainella sp.
Pseudostaffella sp.
Tetrataxis sp.
Tuberitina sp.
Shartymophycus sp.

GSC locality C-21178; 116.7 m above the base of Section 7b:
Beresella sp.
Bradyina sp.
Endothyra sp.
Eotuberitina sp.
Komia sp.
Globivalvulina sp.
Tetrataxis sp.
Tuberitina sp.
Turrispiroides sp.

GSC locality C-21183; 144 m above the base of Section 7b:
"*Glomospira reef*" unit
apterrinellids
Pseudoglomospira sp.

According to Mamet, "The stratigraphic position of Section 7b is about the same as Section 7a; that is, the lower part of the subsection contains abundant Archaediscidae of Morrowan-Atokan age and the upper part, in a *'Glomospira* reef' facies, is near the Atokan-Desmoinesian boundary."

Section 8 (Fig. 6; Wood Glacier section); 1.6 km west of Wood Glacier. The sedimentological description of Section 8 explained that it was composed of a series of folded and faulted subsections directly west of Wood Glacier. The total thickness of the Otto Fiord Formation in Section 8 is about 460 m but the lower and upper contacts are obscured by faulting and the formation may be considerably thicker.

The Otto Fiord Formation at Section 8 ranges in age from Early Pennsylvanian (Morrowan) to Middle Pennsylvanian (Desmoinesian?). The introductory discussion of biostratigraphy in the 'Blue Mountains Belt' suggested that, according to some fossil data, the uppermost part of the Otto Fiord Formation is either the same age or slightly younger than the lower part of the 'Tellevak Limestone Member' of the Hare Fiord Formation, which clearly overlies the Otto Fiord elsewhere. Stratigraphic relationships between the Otto Fiord Formation and the 'Tellevak Limestone' require additional attention in the vicinity of Section 8. Certainly though, the upper part of the Otto Fiord Formation in Section 8, probably Desmoinesian in age, is younger than the uppermost part of the Otto Fiord Formation in the type area at van Hauen Pass, where it is either late Morrowan or Atokan. This diachroneity over a relatively short distance could result from continued deposition of evaporites in the vicinity of the northern 'Blue Mountains Belt' behind a structural or reefal carbonate barrier after the onset of more normal marine conditions at van Hauen Pass.

Nassichuk (1975) identified the Early Pennsylvanian (Morrowan) ammonoid *Brannerоceras branneri* (Smith) from GSC locality C-21241, 90 m above the lowest exposure of the composite Section 8. Strimple and Nassichuk (1974) described the crinoid *Anchicrinus planulatus* Moore and Strimple from the same locality but indicated erroneously that the locality was 360 m above the base. Calcareous foraminifers and algae, moderately abundant in most limestone beds within the Otto Fiord, have been identified by Mamet. Fusulinaceans are abundant in limestone strata near the top of the Otto Fiord and a single collection has been identified by Ross. The following are foraminifer and algal identifications provided by Mamet (pers. com., 1975):

GSC locality C-32262; 18 m above the base of composite Section 8:
Asteroarchaediscus sp. (abundant)
Endothyra sp.
cf. *Millerella* sp.
Neoarchaediscus sp.
Planospirodiscus sp.
Zellerina sp.
Age: Zone 20; Early Pennsylvanian (Morrowan)

GSC locality C-32267; 65 m above the base of composite Section 8:
Asteroarchaediscus sp. (abundant)
Biseriella sp.
Earlandia sp.
Eotuberitina sp.
Globivalvulina sp.
Pseudoendothyra sp.
Pseudostaffella sp.
Radiosphaerina sp.
Age: Zone 21; early Middle Pennsylvanian (Atokan)

GSC locality C-32272; 82.5 m above the base of composite Section 8:
apterrinellids
Beresella sp.
Climacammina sp.
Deckerella sp.
Endothyra sp.
Eotuberitina sp.
Komia sp.
Tetrataxis sp.
Tuberitina sp.

GSC locality C-32273; 84 m above the base of composite Section 8:
Calcisphaera sp.
Earlandia sp.
Eotuberitina sp.
Eostaffella sp.
Tetrataxis sp.

GSC locality C-32278; 91.5 m above the base of composite Section 8:
Asteroarchaediscus sp.
apterrinellids
Eostaffella sp.
Eotuberitina sp.
Tetrataxis sp.
Turrispiroides sp.

GSC locality C-32312; 252 m above the base of composite Section 8:
Asteroarchaediscus sp.
beresellid algae
Biseriella sp.
Endothyra sp.
Eotuberitina sp.
cf. *Pseudobradyina* sp.
Tuberitina sp.

GSC locality C-32313; 256.5 m above the base of composite Section 8:
Asteroarchaediscus sp. (very scarce)
apterrinellids
Calcisphaera sp.
Neoarchaediscus sp.
Orthovertella sp.
Volvotextularia sp.

Mamet (ibid.) commented on GSC locality C-32313: "While the base of the section (Morrow-Atoka) is full of Archaediscidae, they gradually dwindle and disappear at this level. In the Russian Platform, the level of disappearance is about at the lower-middle Moscovian junction; we must be approximately at that level at GSC locality C-32313."

GSC locality C-32228; 336 m above the base of composite Section 8:
undetermined new algal genus
apterrinellids
Beresella sp.
Bradyina sp.
Climacammina sp.
Komia sp.
Neostaffella sp.
Ozawainella sp.
Pseudobradyina sp.
Pseudostaffella sp.
Shartymophycus? sp.
Tetrataxis sp.

GSC locality C-32230; 337.5 m above the base of composite Section 8:

"undetermined new algal genus which according to Dr. Korde is related to the kamaenid algae; I do not agree, as there is an outer cortex. It is probably a Beresellidae?"(ibid.)

Beresella sp.
Bradyina sp.
Climacammina sp.
Eolasiodiscus sp.
Ozawainella sp.
Neostaffella sp.
Pseudobradyina sp.
Pseudoendothyra sp.
Tetrataxis sp.
Turrispiroides sp.

GSC locality C-32242; 385.5 m above the base of composite section 8:

"This rock is a good example of what Russian workers call '*Glomospira* reefs'." (ibid.)

apterrinellids
Biseriella sp.
Globivalvulina sp.
Pseudoglomospira sp.

GSC locality C-21230; 387 m above the base of composite Section 8:

apterrinellids
Beresella sp.
Calcisphaera sp.
Climacammina sp.
Endothyra sp.
Globivalvulina sp.
Komia sp.
Neostaffella sp.
Ozawainella sp.
Pseudostaffella sp.
Tetrataxis sp.
a new algal genus?

Age: Zone 23 *Neostaffella* level; Middle Pennsylvanian (Desmoinesian)

GSC locality C-21232; 390 m above the base of composite Section 8:

apterrinellids
Beresella sp.
Calcisphaera sp.
Endothyra sp.
Globivalvulina sp.
Komia sp.
Eotuberitina sp.
Pseudoglomospira sp.
Tetrataxis sp.
a new algal genus?

Age: Zone 23 *Neostaffella* level; Middle Pennsylvanian (Desmoinesian)

Finally, Ross (in Nassichuk, 1975, p. 196) identified the fusulinaceans *Pseudostaffella gorskyi* (Dutkevich) and *Eostaffella kashira* var. *rhomboides* Rauser from GSC locality C-21230, 387 m above the base of composite Section 8 and indicated an early Moscovian (Kashirian) age. According to Ross (pers. com., 1976), the upper part of the Soviet Kashirian probably is equivalent to Desmoinesian.

Section 9 (southern Blue Mountains section); 13 km southeast of Hare Fiord, 3.2 km northeast of Hare Fiord Diapir. Only about 200 m of Otto Fiord strata occur in Section 9. The base of the section is limited by the Blue Mountains Thrust. Faults may also be in a thick covered interval at the top of the section, directly beneath the 'Tellevak Limestone Member' of the Hare Fiord Formation. A single collection of fossil foraminifers from Section 9 was identified by Mamet (pers. com., 1975) as follows:

GSC locality C-10836; 102 m above the base of Section 9:

Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Archaediscus of the group *A. Krestovnikovi*
Biseriella sp.
Biseriella parva (Chernycheva)
Neoarchaediscus sp.
Neoarchaediscus parvus (Rauser-Chernoussova)
Neoarchaediscus parvus of the group *N. incertus* (Grozdilova and Lebedeva)
Planospirodiscus sp.

Age: Zone 18 (or younger); Chesteran-?Morrowan

Ammonoids are absent from the Otto Fiord Formation at Section 9 in the southern Blue Mountains but they are abundant and diversified in the overlying Hare Fiord Formation. Nassichuk (1975, p. 44) reported the following early Middle Pennsylvanian (Atokan) species from red limestone directly above the 'Tellevak Limestone Member' of the Hare Fiord Formation, just above Section 9:

Pseudopronorites arkansiensis (Smith)
Metapronorites ellesmerensis Nassichuk
Syngastrioceras sp. indet.
Phaneroceras lenticulare Plummer and Scott
Diaboloceras neumeieri Quinn and Carr
Diaboloceras involutum Nassichuk

Melville Island biostratigraphy: Barrow Dome

Barrow Dome (Figs. 1, 39) is a circular evaporitic piercement structure on northern Sabine Peninsula, northeastern Melville Island. It is about 6.4 km wide, has an elevation of some 300 m and a topographic relief of about 60 m. At the surface the dome contains a contorted mixture of gypsiferous anhydrite, shale, sandstone and limestone. Some 'blocks' of strata are 150 m thick and are continuous laterally for more than 0.5 km. One section measured by Nassichuk in 1967 includes 82.5 m of interbedded gypsum and anhydrite, black shale and argillaceous limestone, which closely resemble strata in the Otto Fiord Formation on Ellesmere Island. Bedded units of sandstone, some with plant fossils and possibly coal, have also been recorded from within Barrow Dome.

Only one carbonate sample from Barrow Dome has been examined in thin section. This sample, from the north-central side of the dome, is a totally dolomitized ostracode-rich bioclastic wackestone. Skeletal components other than ostracodes are uncommon, but include brachiopods, gastropods and crinoid plate fragments. The sample is interesting for two reasons: first, totally dolomitized skeletal carbonates are uncommon in the Otto Fiord Formation where it is exposed in bedded succession at the surface on Ellesmere Island and, second, it is similar to ostracode-rich but 'dedolomitized' carbonate rocks cored from interbeds in diapiric salt encountered in the Hoodoo L-41 well on Ellef Ringnes Island.

Ammonoids have been collected extensively from Barrow Dome and the various collections show three distinctive ages between Late Mississippian and Early Pennsylvanian (Nassichuk, 1975). The oldest (Mississippian) ammonoids belong to *Cravenoceras tozeri* Nassichuk, which indicates a Chesteran (early Namurian, E_2) age. The following earliest Pennsylvanian, Halian (late Namurian, R_2 or G_1) ammonoids are known from several localities:

Bilinguites canadensis Nassichuk
Bisatoceras hoeni Nassichuk

Slightly younger Early Pennsylvanian ammonoids of Bloydian (Westphalian A) age are by far the most common ammonoids at Barrow Dome. Several thousand specimens of the following have been recovered from five localities:

Brannerоceras branneri (Smith)
Brannerоceras nicholasi Nassichuk
Brannerоceras hillsi Nassichuk
Brannerоceras sp.
Bisatoceras renni Nassichuk
Gastrioceras melvillensis Nassichuk
Gastrioceras liratum Nassichuk
Syngastrioceras oblatum (Miller and Moore)
Melvilloceras sabinensis Nassichuk

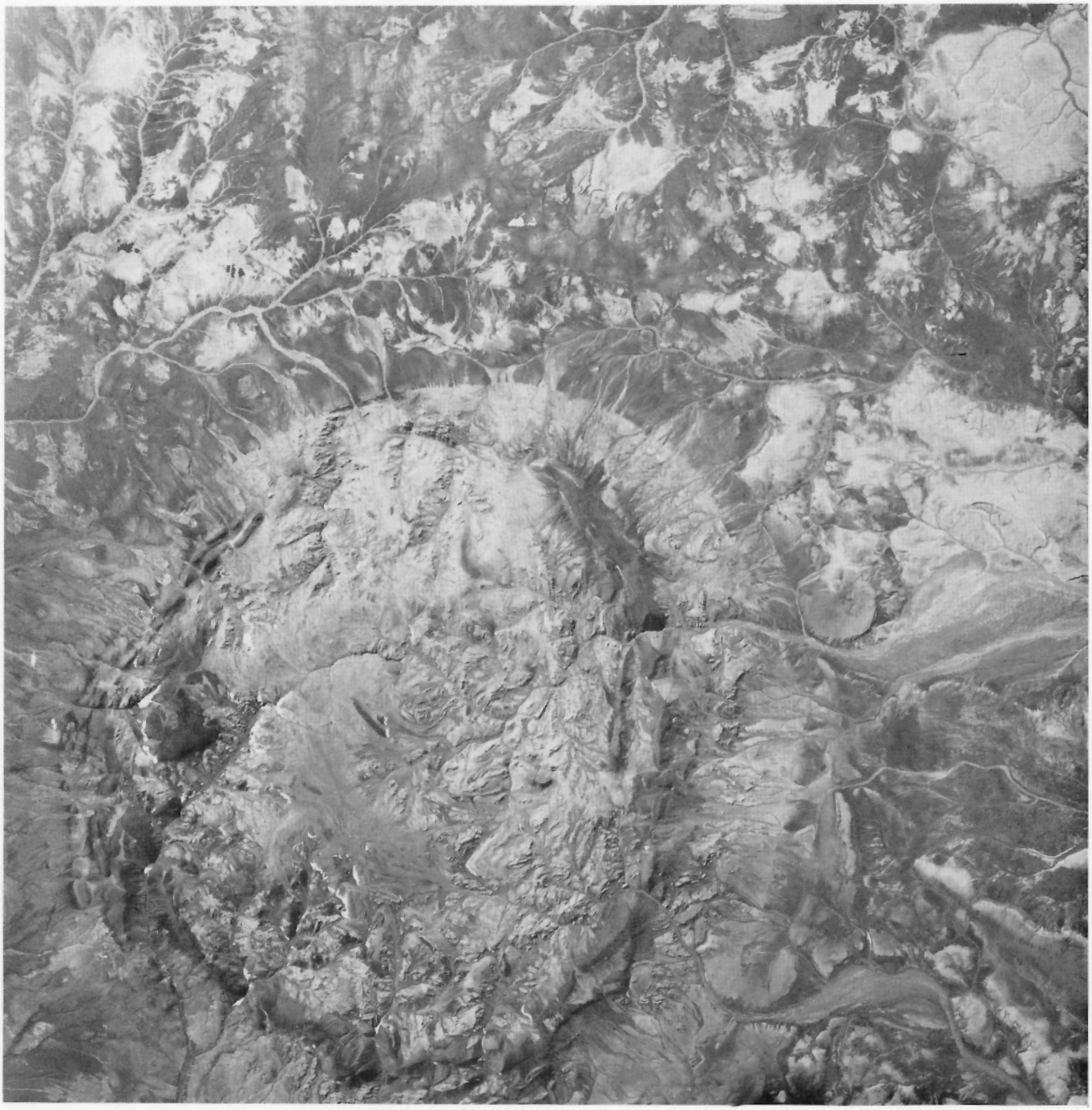


Figure 39. Vertical aerial photograph of Barrow Dome on north-central Sabine Peninsula, Melville Island. Dome is about 6 km in diameter, and is composed mainly of anhydrite and gypsum at the surface. Dark concentric band within the diapir is an igneous intrusion. Note the radial drainage system around the diapir that joins into a more typical dendritic drainage net on the low-relief Mesozoic host rocks. A16764-19.

Carter (in Nassichuk, 1975) identified brachiopods, in a preliminary manner, from two localities at Barrow Dome and suggested that species from one (GSC loc. C-4126) were indicative of a Late Mississippian (Chesteran) age, whereas those from the other (GSC loc. C-4125) indicated Early Pennsylvanian (Halian). Subsequently Carter (1975) described the Chesteran materials from GSC locality C-4126 and very tentatively suggested, on the basis of generic occurrences elsewhere in North America, the possibility of their being older than Namurian or of late Viséan age. The Mississippian brachiopods described by Carter (ibid.) are:

Hustedia n. sp.
?Eomarginifera sp.
Ovatia cf. *O. minor* (Snider)
Composita arctica Carter
Cleiothyridina cf. *Composita offirmata* Beil
Crurithyris cf. *C. acadica* (Bell)
Anthracospirifer borealis Carter
Beecheria formosa (Hall)

Lower Pennsylvanian (Halian) brachiopods identified by Carter (in Nassichuk, 1975) from GSC locality C-4125 include:

Camerisma sp.
Phricodothyris n. sp.?
Rugosochonetes sp.

Other fossils found associated with Otto Fiord rocks at Barrow Dome include corals, ostracodes and plant remains. Bamber (pers. com., 1966) identified the solitary rugose corals *Lophophyllidium* sp. and *Bartychisma* sp. from GSC locality 73161 and suggested a Pennsylvanian age. A single plant stem preserved in pale grey sandstone from GSC locality C-33959 tentatively was identified as *Mesocalamites* sp. by Pfefferkorn (pers. com., 1975), who considered it to be of Namurian age. Finally Copeland (pers. com., 1967) identified the ostracode *Gutschickia*? sp.; the genus is known elsewhere in Pennsylvanian and Permian strata.

Ellef Ringnes Island

Malloch Dome biostratigraphy

Malloch dome is on a short peninsula projecting into Kristoffer Bay on the southwest side of Ellef Ringnes Island. The dome has a maximum diameter of 6.3 km and a maximum elevation of about 120 m. Greiner (1963) reported siltstone, dolomite and limestone as inclusions within the dome. An inclusion of more than 30 m of light red to pale pink sandstone is of special interest. Similar sandstone occurs on the eastern side of the Basinal Clastic and Evaporitic Belt, between the Blue Mountains and the Krieger Mountains on northern Ellesmere Island where the formation outcrops as a normal bedded succession.

A single ammonoid species, *Brannerocerasbranneri* (Smith), is known from Malloch Dome (Nassichuk, 1975). *Branneroceras* is an index for the Lower Pennsylvanian (Morrowan) Bloydian Stage (Westphalian A).

Harker (in Greiner, 1963, p. 566) recognized nine brachiopod species recovered from limestone within Malloch Dome and indicated that they are "probably of Pennsylvanian or Permian age".

Haakon Dome biostratigraphy

Haakon Dome is on the east side of Ellef Ringnes Island, on the eastern edge of Christopher Peninsula. It extends southeasterly from Trio Island; the dome is about 8 km long and 1.6 km wide. Stott (1969) reported abundant limestone inclusions from within the dome. One large inclusion contains more than 300 m of recrystallized, in some places pseudobrecciated or sheared, calcite-spar limestone. Bitumen infilling of secondary pores is common in some samples. Several stages of calcite spar cementation separated by at least one phase of solution are apparent in some of these rocks; bitumen emplacement accompanied by or preceded by dolomite crystal growth postdated this solution event. Many dolomite crystals are partly or almost totally replaced by calcite ('dedolomitization'). In a few samples, intraclasts of limestone preserve original fabrics; some were pelletoidal wackestone containing large ostracode valves, others were packstone or wackestone containing spherulitic grains and ooids.

Brachiopods are abundant in various limestone inclusions in Haakon Dome and one collection has been identified by Waterhouse (in Stott, 1969, p. 16):

Krotovia sp.
Avonia sp.
Plicatifera sp. (or echinoconchid)
Linoproductus sp. (or *Ovatia* sp.)
Coledium cf. *C. erugatum* Grant
poorly preserved buxtoniid, dictyoclostid, diclasmatid, and leioproductid brachiopods
Parallelodon sp.

Waterhouse (ibid.) was unable to provide a definite age for this fauna but indicated that some species resemble forms known from the Mississippian of western Canada (Nelson, 1961). He suggested that early Middle Pennsylvanian (Atokan) was a possibility since a similar fauna is found at the base of the Hare Fiord Formation immediately above the Otto Fiord Formation at GSC locality 56430, near Stepanow Creek on Hare Fiord, northern Ellesmere Island. Nassichuk (1975) described 15 ammonoid species from locality 56430 and documented an Atokan age. Atokan or slightly younger fossils are common near the top of the Otto Fiord Formation east of the head of Hare Fiord and in the Wood Glacier area of northern Ellesmere Island but they have not been dated confidently from within the Otto Fiord outside Ellesmere Island.

Hoodoo Dome sedimentology

Hoodoo Dome (Figs. 1, 40), which McLaren (1963) designated informally "Meteorologist Peninsula Dome", is situated on Meteorologist Peninsula, south-central Ellef Ringnes Island. The structure is about 16 km long and 9 km wide but evaporitic rocks outcrop over a limited 4 km^2 area near the western edge of the dome.

Lower Cretaceous (Neocomian) sandstone of the Isachsen Formation surrounds evaporitic rocks on Hoodoo Dome except along the southern edge of the evaporitic mass where shale from the slightly younger Cretaceous (Albian) Christopher Formation is adjacent to evaporites. Nassichuk and Roy (1975) identified serpulid worm tubes from relatively shallow-water skeletal limestone mounds within the Christopher Formation and concluded that Hoodoo Dome probably was a positive feature at least as far back as Albian time.

Only a few rock samples, mainly limestone, taken from the exposed surface of Hoodoo Dome have been studied in thin section. One is a brachiopod coquinoid limestone cemented by fibrous calcite of probable submarine origin (Davies, 1977c). Other carbonate rocks include bioclastic, intraclastic and pelletoidal grainstones and packstones. Thin-walled brachiopods, ostracodes and rare ammonoids are skeletal components of these rocks. Some clasts are coated with algaelike encrustations. Spherulitic grains

with a radial crystal fabric, which may be recrystallized ooids, are common. Another sample is a calcareous sandstone composed of poorly sorted, medium-grained quartz sand in a coarsely crystalline calcite cement.

Imperial, Panarctic Dome et al Hoodoo L-41 exploratory well---salt discovery

Limestone, dolomite, sandstone and siltstone were encountered within a thick succession of Otto Fiord anhydrite and salt in the Hoodoo L-41 exploratory well (Figs. 40, 41) drilled off the southeastern flank of Hoodoo Dome in 1972 (Davies, 1975a). This well started in sandstone of the Lower Cretaceous Isachsen Formation and entered gypsum or anhydrite at the top of the subsurface extension of the diapir at 337 m. Log characteristics suggest that only the upper 12 m of this sulphate 'cap' are hydrated to gypsum (Fig. 41), and that anhydrite continues to a depth of 370 m.

Between 370 and 415 m about 45 m of limestone were penetrated; thin sections of selected cuttings from this limestone unit show it to have a recrystallized microspar texture, stylolitic in places but with no visible or preserved bioclasts or other primary fabrics. This unit is underlain by about 108 m of anhydrite with interbedded limestone and dolomite and lesser amounts of calcite-cemented sandstone and siltstone; some of the sand grains include pleochroic (dark green to olive-brown) minerals. Some cuttings are composed of coarsely crystalline rhombic calcite suggestive of calcitized dolomite ('dedolomite' fabrics).

At 620 m the well entered halite (logs), and remained in halite-dominant lithologies to a total depth of 4280 m. Interbedded within the steeply inclined halite mass are numerous thin anhydrite, limestone, dolomite and rare sandstone-shale beds. Thin sections of selected cuttings from some interbeds show that the limestone usually is recrystallized microspar, commonly with some detrital and authigenic quartz and scattered authigenic pyrite.

In a few less altered limestone samples, an original wackestone to lime mudstone texture is preserved. Some have a pelletoidal (clotted) fabric, with a few calcite-filled moulds of thin-walled bivalved molluscs, possibly ostracodes, and also calcispheres (spar-filled, no skeletal wall) about 120 µm in diameter. Rosettes of radiating authigenic crystals about 400 µm in diameter, preserved as calcite, are common in some cutting samples (ibid.). Other limestone samples show extensive replacement by euhedral authigenic anhydrite along former fractures.

Salt cored at about 1220 m (core 1) is a coarsely recrystallized halite containing many angular and sheared mudstonelike or clay lithoclasts. Some anhydrite stringers and beds also are present. Dips in the halite core range between 70 and 80 degrees.

At approximate depths of 2835 m and 3260 m, thick (50 and 33 m) gabbroic igneous intrusions, either sills or dykes, were intersected in the salt section (Fig. 41); a short core (core 2) was obtained from the higher of these two intrusives.

Core 3, at a depth of 3675 m, recovered steeply dipping and deformed carbonate and anhydrite with some halite stringers and beds. Some of the anhydrite preserves a primary laminated fabric. Thin sections of limestone from the cored interval reveal that some are diagenetically altered bioclastic, rounded-grain grainstone that contains gastropods (several spire types), pelecypods, brachiopods and small ostracodes. Many of the larger bioclasts have been crushed, and replaced by solution-moulded diagenetic processes. Many rounded grains preserve a radial and concentric pseudospherulitic to pseudo-oolitic texture, but are of indeterminate origin. The present fabric comprises large interlocking anhedral calcite crystals that transect primary grain boundaries. Indistinct remnants of 'cloudy' rhombic cores in many of these calcite crystals suggest

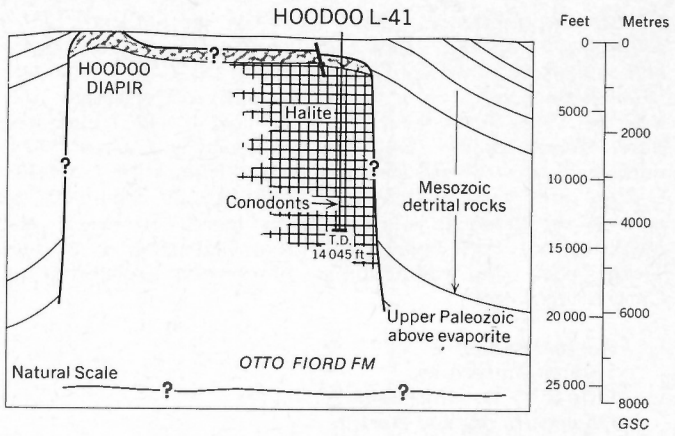


Figure 40. Schematic cross-section across Hoodoo diapir on south-central Ellef Ringnes Island. This section is based on the present exposed dimensions of Hoodoo Dome and its apparent subsurface extension below the Hoodoo L-41 exploratory well, but is not based on seismic data.

strongly that the limestone represents calcitization of coarsely crystalline dolomite, which in turn replaced the original primary sediment or rock. The limestone also yielded the only microfossils (conodonts) found in the Paleozoic rocks in this well.

From GSC locality C-65147, between depths of 3778 and 3780 m, three conodonts were recovered. Two of the specimens are *Cavusgnathus lautus* Gunnell, and the third is a broken piece of a bar-type conodont (Bender, pers. com., 1973). *Cavusgnathus lautus* ranges in age from Early Pennsylvanian to Early Permian and, thus, is not of great stratigraphic value. Bender found this conodont in one other sample from the Otto Fiord Formation, where it is exposed in normal stratigraphic succession near easternmost Hare Fiord on northwestern Ellesmere Island. This outcrop sample is dated on foraminiferal evidence to be of latest Mississippian or Early Pennsylvanian age. *Cavusgnathus lautus* also was found by Bender at different horizons in the Nansen Formation, the Upper Mississippian to Lower Permian carbonate shelf lithofacies exposed in the northern Sverdrup Basin.

The conodont evidence is inconclusive for a precise correlation with the Otto Fiord Formation exposed on Ellesmere Island, but with the evidence from the paleontological data (ammonoids, brachiopods) obtained from surface samples from other diapirs on Melville, Ellef Ringnes, Amund Ringnes and Axel Heiberg islands (Nassichuk, 1975), there is no doubt that the diapirs represent an intrusive halite lithofacies of the Otto Fiord Formation.

Amund Ringnes Island biostratigraphy

Carboniferous ammonoids have been recovered from limestone blocks enclosed in gypsum-anhydrite in a large evaporite piercement in the central northwestern part of Amund Ringnes Island. This structure was referred to informally as "Amund Ringnes piercement dome" by Norris (1963) and "North Cornwall Dome" by Gould and De Mille (1964). It has an exposed length of about 27 km and a maximum width of 12.8 km, a maximum elevation of about 270 m and relief of about 60 m. Rather extensive limestone and dolomite units, some measuring 150 m thick and extending for more than 1 km along strike, are common particularly near the western edge of the dome. Shale and sandstone have not been reported from the dome. Two limestone blocks near the western edge have yielded the ammonoid *Brannerоceras branneri* (Smith), which is indicative of an Early Pennsylvanian (Morrowan) age.

Axel Heiberg Island biostratigraphy

'South Fiord Dome'

The 'South Fiord Dome', a large ovoid diapiric structure with a maximum diameter of about 8 km, is near the head of South Fiord on west-central Axel Heiberg Island. It contains many large limestone blocks that are enveloped in contorted anhydrite and gypsum. Several of these blocks have yielded brachiopods and ammonoids.

A large limestone block near the centre of the dome (GSC loc. 47996) yielded the following ammonoids of Early Pennsylvanian (Halian) age (Namurian R_2 or G_1):

Bisatoceras hoeni Nassichuk
Bilinguites heibergensis Nassichuk
Bilinguites sp.

These ammonoids were examined first by Glenister and Furnish (in Hoen, 1964), who tentatively referred the two

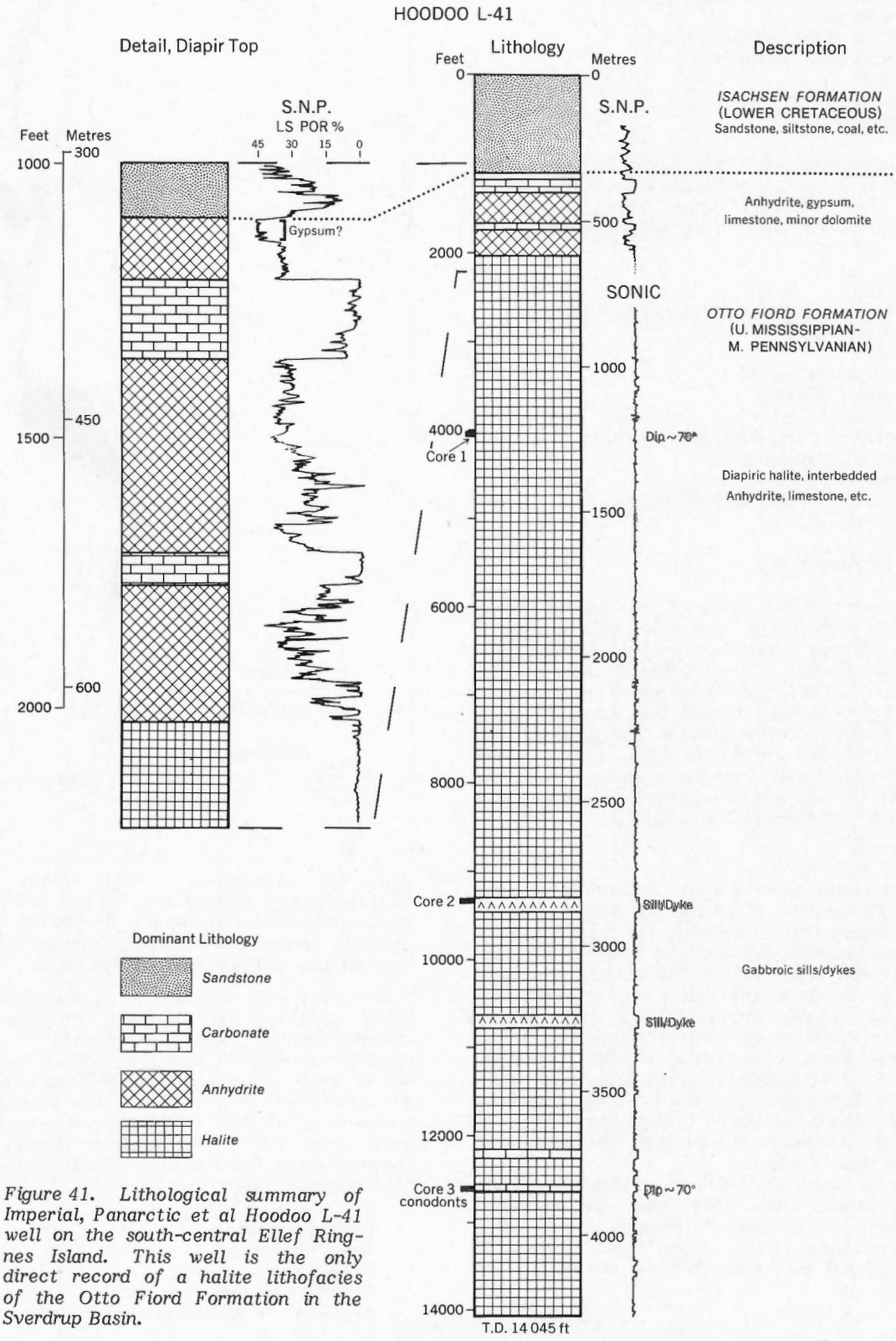


Figure 41. Lithological summary of Imperial, Panarctic et al Hoodoo L-41 well on the south-central Ellef Ringnes Island. This well is the only direct record of a halite lithofacies of the Otto Fiord Formation in the Sverdrup Basin.

species (now referred to *Bilinguites* by Nassichuk, 1975) to *Eumorphoceras carinatum* Schmidt and *Reticuloceras reticulatum* Phillips, respectively, and considered them to be of an Early Pennsylvanian age (middle Namurian). Brachiopods associated with the ammonoids at GSC locality 47996 were identified by Harker (in Hoen, ibid.), who considered them to be Pennsylvanian. The species listed in Hoen (ibid., p. 8) are:

Linoproductus sp. (small form)
Linoproductus aff. *L. cora* (d'Orbigny)
Buxtonia aff. *B. peruviana* (d'Orbigny)
Echinoconchus sp.
Phricodothyris aff. *P. septata* Chronic
Neospirifer ex gr. *N. cameratus*
Dielasma sp.
"*Enteletes*"(?) sp.
"*Reticularia lineata*"

Brachiopods are also abundant and diverse at GSC locality 26403 in another large limestone inclusion within 'South Fiord Dome'. The following species were described by Carter (1975), who estimated a Late Mississippian (Viséan) age:

Overtonia sp.
Setigerites cf. *S. altonensis* (Norwood and Pratten)
?*Eomarginifera* sp.
Productus sp.
Ovatia cf. *O. minor* (Snider)
indeterminate wellerellid
Composita arctica Carter
Cleiothyridina cf. *Composita offirmata* Bell
Crurithyris cf. *C. acadia* (Bell)
Anthracospirifer borealis Carter
"*Brachythyris*" cf. *B. dichotomus* Litvinovich
Phricodothyris n. sp.?
Martinia sp.
Beecheria formosa (Hall)

According to Carter (ibid.), seven of the species at this locality in 'South Fiord Dome' also occur at a locality in Barrow Dome, Melville Island (see previous discussion of faunas at Barrow Dome). Carter indicated that because the fauna of the 'South Fiord Dome' was more diverse, a Viséan assignment was more certain than for the Barrow Dome fauna. Further, Carter stated: "The genera *Overtonia* Thomas, 1914 and *Setigerites* Girty, 1939, are not known from strata younger than the Viséan, whereas the genera *Productus* Sowerby, 1814 and *Martinia* McCoy, 1844 are not known from strata older than Viséan...."

Whitsunday Bay

A narrow discontinuous 'belt' of Otto Fiord rocks has been mapped from the vicinity of Whitsunday Bay on southeastern Axel Heiberg Island northward for about 80 km to Gibs Fiord on east-central Axel Heiberg Island (Fig. 42 and Map 1302A, Eureka Sound North; Thorsteinsson, 1971b). Through much of this belt, the Otto Fiord is exposed in diapirs that are related either to the Stolz Thrust (Whitsunday Bay Diapir) or to major anticlinal structures (Gibs Diapir and Mokka Fiord Diapir). Locally, however, particularly north of Whitsunday Bay and east of Buchanan Lake, nondiapiric Otto Fiord rocks are in bedded sequences beneath the Hare Fiord Formation; in such places only the upper part of the formation is preserved, the lower part being a locus for thrust faults.

About 11.2 km north of the head of Whitsunday Bay, about 120 m of bedded Otto Fiord rocks were measured between the base of the section (marked by the Stolz Thrust) and the base of the Hare Fiord Formation. Information regarding the oldest rocks in the Otto Fiord

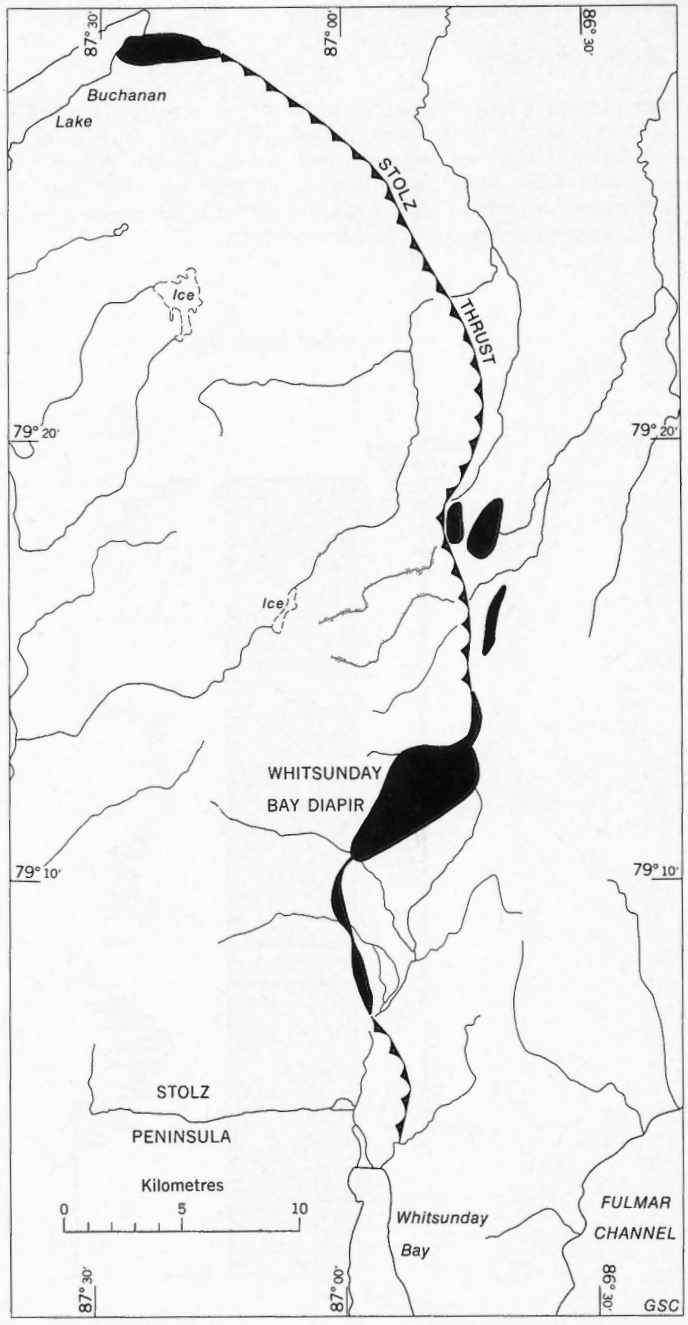


Figure 42. Distribution of Otto Fiord rocks between Whitsunday Bay and Buchanan Lake in eastern Axel Heiberg Island. Otto Fiord strata are thinned by truncation along the Stolz Thrust. All fossil data from the Otto Fiord in this area indicate an Early Pennsylvanian age.

Formation is not available from this area. About 105 m below the base of the Hare Fiord Formation, a 1.5 m bed of pink-weathering, slightly conglomeratic sandstone containing brachiopods (GSC loc. C-4009) is overlain by a 0.6 m bed of sandy limestone containing calcareous foraminifers and algae (GSC loc. C-4009a). The brachiopods from the sandstone (GSC loc. C-4009) were examined by Carter (pers. com., 1970), who reported poorly preserved representatives of the following species indicative only of a "Pennsylvanian" age:

Phricodothyris sp.
Wellerella? sp.
Krotovia? sp.
linoproductid gen. et sp. indet.
athyridid or martiniid gen. et sp. indet.

Calcareous foraminifers and algae from GSC locality C-4009a, less than 1 m stratigraphically above the listed brachiopods from GSC locality C-4009, were identified by Mamet (pers. com., 1970), who suggested an Early Pennsylvanian, Morrowan (Zone 20) age:

Asteroarchaediscus sp.
Neoarchaediscus sp.
Neoarchaediscus postrugosus (Reitlinger)
Planospirodiscus sp.

Basal beds of the overlying Hare Fiord Formation yielded nine ammonoid species dated by Nassichuk (1975) as Atokan; thus, an upper limit for the age of the Otto Fiord Formation at Whitsunday Bay is provided.

ELEMENTAL AND ISOTOPIC ANALYSES OF OTTO FIORD ROCKS

Strontium content of Otto Fiord anhydrite

Fourteen analyses were run for strontium content through two successive beds of anhydrite containing gypsum-crystal pseudomorph fabrics (Fig. 43) from the type area. The results averaged 1060 ppm Sr within a narrow range of 1040 to 1120 ppm. There is no significant variation in vertical distribution of the strontium values through either of the two beds of anhydrite.

Any discussion of the significance of this level of Sr in the anhydrite is dependent on whether or not it is close to the strontium content of the primary sulphate mineral, in this case gypsum. Butler (1973, p. 450) argued that, in the absence of dolomitization during early diagenesis, the replacement of gypsum by anhydrite during burial and in the presence of undiluted pore brines should result in an anhydrite containing close to the same level of Sr as the primary gypsum. It has been noted that carbonates interbedded with the Otto Fiord anhydrite, particularly in the type area, contain only a small percentage of dolomite. Thus, it is assumed tentatively that the primary gypsum crystals that grew at the sediment-brine interface had an Sr content of 1000 to 1100 ppm.

From analyses of gypsum from the Persian Gulf and other areas, Butler (ibid.) calculated the partition coefficients for strontium in gypsum precipitated from marine-derived brines. This temperature-dependent coefficient is about 0.2 at 20°C and 0.22 at 10°C; these temperatures cover the probable range for a shallow subaqueous brine. Applying these data in a very general way to the assumed Sr values of the primary gypsum, we obtain an approximate Sr^{+2}/Ca^{+2} ratio for the precipitating brine of 2.0 to 2.4×10^{-2}.

Butler (ibid., Fig. 5) has shown that for Persian Gulf (Abu Dhabi) lagoonal to low upper intertidal brines, concentrated from 1.2 to 3.4 times normal marine water, the ratio Sr^{+2}/Ca^{+2} of the brine before the onset of gypsum precipitation is relatively constant at about 1.85×10^{-2}. Higher in the intertidal zone, brine concentrations exceed 3.4 times and gypsum begins to precipitate; brines in this condition have an Sr^{+2}/Ca^{+2} ratio of about 2.2×10^{-2}. In high intertidal and supratidal settings, where brine concentrations may be much higher, the Sr^{+2}/Ca^{+2} ratio of the brine often exceeds 3.0×10^{-2}.

Obviously the temperature regime on an intertidal and emerged substrate such as the Persian Gulf sabkha and environs is considerably higher than that expected for a subaqueous, albeit shallow, brine and partition coefficients for strontium in gypsum precipitated in the two environments will be different. However, since these are only average figures, and as the partition coefficient does not vary greatly with temperature, the following generalizations may be made (from Butler's data):

(1) The average value of 1060 ppm Sr in the two beds of Otto Fiord anhydrite, if similar to the original Sr content of the primary precipitate, is consistent with gypsum grown from a brine concentrated by evaporation from 3.4 to about 7 times.
(2) The consistency of the Sr distribution through the two beds suggests that there have not been extreme variations in brine composition, and this is consistent with a subaqueous brine environment rather than a high intertidal to supratidal (sabkha) setting with its extreme variations in brine concentration and Sr^{+2}/Ca^{+2} ratios (Butler, ibid., Fig. 5).
(3) The Sr^{+2}/Ca^{+2} ratio (2.0 to 2.4×10^{-2}) of the precipitating brine calculated from the Sr content of the Otto Fiord anhydrite is close to the Sr^{+2}/Ca^{+2} ratio of 2.2×10^{-2} shown by Butler (ibid.) to be the expected level in brines from which gypsum is precipitating in the absence of dolomitization; again, this is consistent with the Otto Fiord rocks.

Sulphur isotopic composition of Otto Fiord anhydrite

In an earlier paper (Davies and Krouse, 1975), the results of 24 analyses of Otto Fiord anhydrite for sulphur isotope ratios ($^{34}S/^{32}S$, expressed as $\delta^{34}S‰$) were listed and discussed. No additional analyses have been made. The general statement on sulphur isotope terminology and methodology used in the earlier paper (ibid., p. 221) will not be repeated here.

The sulphur isotope values of the Otto Fiord sulphates average $+14.6 \pm 1.9‰$ $\delta^{34}S$. As a generalization, the older samples, of earliest Pennsylvanian age, are slightly heavier isotopically (more ^{34}S) than the youngest samples of Middle Pennsylvanian age. This trend and the mean values are consistent with the published curves for the secular variation in sulphur isotope ratios of sulphate rocks through geological time (Fig. 44; see Holser and Kaplan, 1966; Davies and Krouse, 1975, Fig. 1). The average $\delta^{34}S$ values of other Paleozoic sulphate evaporites in the Arctic Archipelago that are either older or younger than the Otto Fiord Formation also conform closely to this secular curve (Fig. 44).

Of greater relevance to the interpretation of depositional environments and processes for the Otto Fiord anhydrite is the pattern of $\delta^{34}S$ distribution found in 14 analyses through two beds of anhydrite with gypsum-crystal pseudomorph fabrics from the type area (Fig. 3). These beds are the same units sampled and analyzed for strontium; they are estimated to be middle Early Pennsylvanian in age.

As described in an earlier section, each bed has a lower and thicker layer of anhydrite in which pseudomorphs after inclined crystals of gypsum are defined by inclusions, and an upper and thinner layer with an indistinct, inclusion-rich submosaic anhydrite fabric. At the base of each pseudomorph layer, the $\delta^{34}S$ value is +16.2 to +16.3. The anhydrite then becomes progressively isotopically lighter upward through the pseudomorph layer to a mean of between +12.0 and +12.5 $\delta^{34}S$ at or just below the base of the upper, inclusion-rich layer. Analyses within the upper layer show a slight trend of increasing $\delta^{34}S$.

If it is assumed that the consistency of this trend in the two beds is not fortuitous, and that the values obtained do not differ significantly from the isotopic composition of the original gypsum, several conclusions may be drawn.

(1) The vertical trend of decreasing $\delta^{34}S$ suggests that fractionation occurred during precipitation of the primary gypsum crystals, resulting in a preferential uptake of the heavy sulphur isotope (^{34}S) into the precipitate and a consequent progressive decrease of the $^{34}S/^{32}S$ ratio in the residual brine; later forming crystals therefore contained less ^{34}S.

(2) The differential of about $4^{o}/_{oo}$ $\delta^{34}S$ within one crystal layer is greater than the expected fractionation occurring during crystallization of gypsum as documented by Thode and Monster (1965) and Nielsen (1972). However, Butler et al. (1973, Fig. 2) have shown that sulphur isotopic fractionation up to $4^{o}/_{oo}$ $\delta^{34}S$ occurs in gypsum precipitated at an advanced stage of brine evolution and crystallization. These data (ibid.) are derived from a study of sulphate evaporites in the Persian Gulf sabkha and may not be directly applicable to fractionation processes in a subaqueous brine environment. However, they

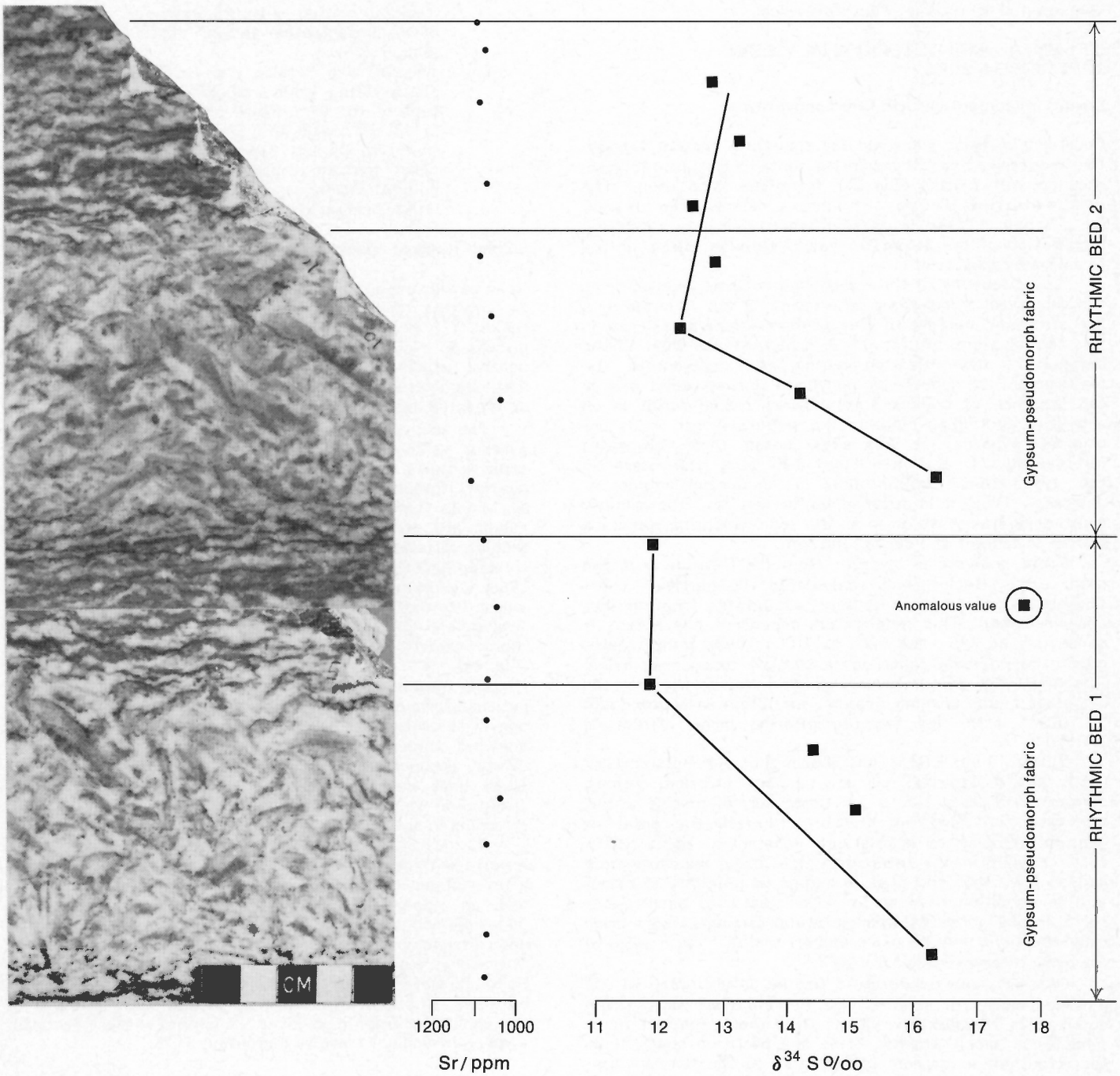


Figure 43. Plots of Sr in ppm and $\delta^{34}S$ for 14 samples through two beds of anhydrite containing gypsum-crystal pseudomorphs from Section 2, near van Hauen Pass. The beds probably are Early Pennsylvanian. Note the consistency of the Sr data, and the decrease in $\delta^{34}S$ value from the base to the top of the pseudomorph layers (modified from Davies and Krouse, 1975, Fig. 3).

appear to be the only available recent data on gypsum crystallization and isotopic fractionation, and do substantiate that this degree of fractionation occurs in natural systems.

(3) Significantly, Butler et al. (1973) found that modern sabkha sulphates show a vertical trend of increasing $\delta^{34}S$, and they suggested that this trend might be used to identify supratidal cycles in older evaporite deposits. In contrast, the reverse trend is found at least at an intrabed level for the Otto Fiord anhydrites that are interpreted from other evidence to be of subaqueous brine origin. Whether this trend holds true for complete units of bedded and other altered types of anhydrite in the Otto Fiord Formation is unknown.

(4) The trend of decreasing $\delta^{34}S$ for each of the Otto Fiord anhydrite beds is interpreted to result from fractionation during crystal growth of gypsum in a more or less 'closed' brine system, isolated temporarily from the 'infinite' source of dissolved oceanic sulphate.

Other chemical analyses

In addition to the analyses for strontium and sulphur isotopes, selected anhydrite and carbonate samples were analyzed for Ca, Mg, Na, Zn, V, Co, Cu, Ni, Fe and Pb. Christie (1975) analyzed 31 samples of Otto Fiord limestone for Ca, Mg, Sr, PO_4, Cu, Pb and Zn; he concluded that none showed significant enrichment in any of the metallic elements. Only one of the samples we analyzed appears to show significant enrichment; a dense, black, possibly slightly argillaceous, thin limestone interbed in anhydrite from the type area contained 1300 ppm Zn (and less than 500 ppm Pb). Without more extensive sampling and analysis, further comment on this result is meaningless.

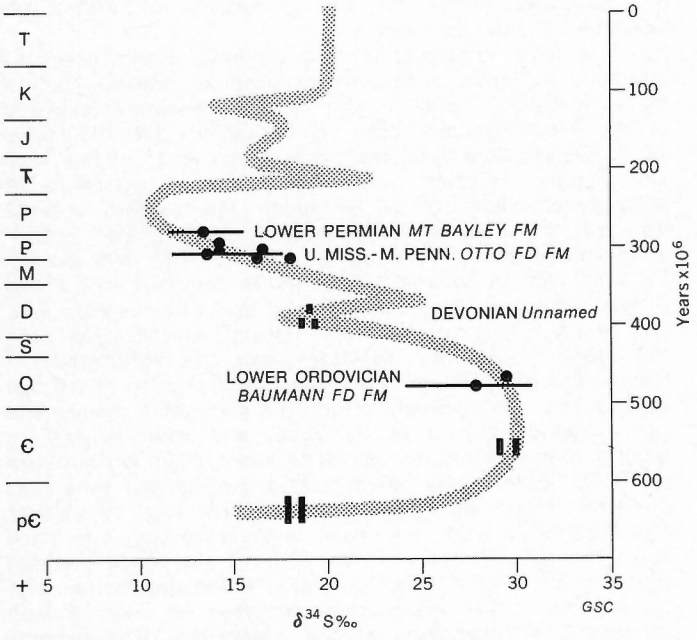


Figure 44. Sulphur isotope ratios for the Otto Fiord Formation and other Arctic sulphate evaporites overlain on the secular curve for sulphur isotope variations published by Holser and Kaplan (1966). Note the general conformity between the Arctic data and the curve (from Davies and Krouse, 1975, Fig. 1).

The close spatial relationships between evaporites, carbonates and red beds shown by the Otto Fiord and Borup Fiord formations are conducive to the formation of economically significant copper ores and, possibly, other metallic mineral zones. Although no analytical data suggest the presence of any mineralization, these lithofacies on northwestern Ellesmere Island deserve more careful and detailed geochemical surveying.

DEPOSITIONAL ENVIRONMENTS OF EVAPORITES

Evaporites in the geological sense are sedimentary deposits composed mainly of minerals that precipitated from saline solutions (brines) concentrated by evaporation. The usual source for areally extensive and stratigraphically thick evaporite deposits is seawater. Saline groundwater is the source of many smaller evaporite deposits in present continental regions. Evaporite minerals may precipitate within a saline body of water, at a sediment-water interface, or interstitially within carbonate or other types of sediment under standing saline water or subaerially. Because many evaporite minerals are highly soluble and generally unstable, one mineral phase may dissolve or replace another on a large scale and as a very early diagenetic phenomenon. However, some replacements may be later diagenetic products not related spatially or temporally to penecontemporaneous evaporitic conditions.

An obvious primary condition for the generation of evaporitic minerals and deposits is climatic rigour and net evaporation. The evaporitic sodium sulphate deposits in seasonally frozen lakes in Antarctica, western Canada and elsewhere show that net evaporation is not exclusively a condition of low latitudes. However, major calcium sulphate and sodium chloride deposits in the geological evaporite deposits and most significant modern ones are within two broad latitudinal belts extending between about 10 and 40 degrees north and south of the Equator. These belts coincide approximately with the subtropical high-pressure atmospheric circulation zones (Gordon, 1975). Distribution of evaporite deposits within these belts is controlled by local climatic and geomorphic factors. The equatorial region, corresponding to the zone of low-pressure atmospheric circulation, specifically is an area of few or no significant evaporite deposits.

A second primary condition for the formation of geologically significant evaporite deposits is large bodies of saline water within the climatic belts of net evaporation. These bodies of water must be sufficiently separated from the relatively infinite salt reservoir of open seas or oceanic water that evaporation will lead to hypersalinity and eventual saturation in solutes. These conditions often are linked directly to the tectonic and geomorphic setting and the evolution of continental plates. Obviously the shallower the water the faster it will evaporate. A corollary of this concept is that under ideal conditions where the saline water is partly or completely isolated and net evaporation exceeds influx of marine and fresh waters, the saline brine level will be lowered progressively relative to the contemporaneous sea level. Maiklem (1971) has termed this process 'evaporative drawdown'. At the same time sediments deposited around the margins of the evaporating water mass usually will be exposed at the evaporitic maximum. Further, if changes in sea level in an initially deep saline basin are large, for example, Hsü's (in Hsü et al., 1973) interpretation of Messinian sea levels in the Mediterranean, isostatic adjustment of the crust may be an expected consequence of drawdown.

The major problem in the field of evaporite geology today is to correctly interpret depositional environments and processes leading to the accumulation of major anhydrite and halite evaporite deposits. Unlike many carbonate

and terrigenous detrital sedimentary deposits, modern analogues for depositional environments of thick and extensive evaporites are lacking. This lack is partly an accident of time, since the present Holocene transgressive flooding of the continental shelves is not conducive to the development of large-scale evaporitic environments. However, it also is due to the present stage of tectonic evolution of the continents in relation to sea level; for example, present major rift troughs in northeastern Africa are accumulating nonevaporitic sediments, yet are underlain by thick sequences of halite-dominant evaporites of Tertiary to Quaternary age.

Another factor that has played a major role in the formation of evaporite deposits through geological time is the variation of global climate from glacial to thermal maxima. Many authors use the secular distribution of evaporite deposits to document the episodic thermal maxima, in contrast with the record of glacial deposits at thermal minima (e.g. Meyerhoff, 1970, Fig. 4). However, to isolate climatic fluctuations as the principal control of temporal variability in evaporite distribution may be misleading, since climatic variations may be linked intimately to major tectonic events that, in turn, may be controlled by extraterrestrial events. Sloss (in Sloss and Speed, 1974, and earlier papers) demonstrated that major tectonic events were synchronous with generation of similar stratigraphic sequences for different cratons; phases of craton emergence and tectonic oscillation (Sloss and Speed, 1974) generally coincide with incidence of major evaporite deposits, although inevitably there are exceptions.

Another factor that may have affected the temporal distribution of evaporite deposits is the composition of seawater. Holland (1974) and others reasoned that the similarity in bulk composition of evaporite deposits through time argues against significant variations in the composition of seawater. However, temporal variations in the sulphur isotope compositions of seawater (Holser and Kaplan, 1966; Davies and Krouse, 1975) and possibly of other elements in their isotopes are but one suggestion of possible compositional variations in seawater chemistry. Further, the vast masses of sodium chloride and sodium locked up in evaporite deposits in Cambro-Ordovician, Devonian, Permian, and Triassic-Jurassic times alone must have had some impact on global oceanic salt budgets (see also Kinsman, 1975a, p. 377). Again, variations in oceanic salinity might be expected between glacial maxima, when vast masses of fresh water were locked up as ice, leaving a net gain in total oceanic solutes. Although this type of speculation adds little to the real understanding of evaporite formation, the general problem of secular variations in oceanic salinity is worthy of further evaluation, as it is pertinent not only to evaporite genesis but also to biotic evolution and extinctions.

The influence of tectonics on the original distribution of major evaporite deposits is exemplified best by the close association between rift troughs (proto-oceans) and thick halite-dominant evaporite deposits. This association has been recognized clearly only in the last few years, concomitant with the resolution of seafloor spreading and plate movements as fundamental global tectonic processes. Kinsman (1975a, b) has synthesized some of these new data and concepts into a model for the evolution of rift valley basins into proto-oceans in which evaporite deposits, within latitudinal constraints, are a predicted basal component of the sedimentary fill in rift basins and 'geosynclines'. As an example, the Upper Triassic to Cretaceous salt deposits flanking the northern and southern Atlantic continental masses probably were deposited in grabens below sea level, formed in the early stages of continental rupture and divergence (Burk, 1975; and others).

'Barred basin' models for evaporite genesis

The hypothesis that evaporite deposits accumulated subaqueously in an evaporating basin environment separated from a marine water source by a shallow bar has been proposed with several variations since the mid-19th century (see Kirkland and Evans, 1973, p. 1 - 4, for a short historical review). Many of the depositional models that use this hypothesis have been relatively simple, yet the general concept is applicable to many continental-interior evaporite basins and deposits (Michigan Basin, Elk Point Basin, Paradox Basin, many others) and to the major evaporite deposits in rift basins and former rifted continental margins (Mediterranean, Red Sea, Dead Sea, Atlantic north and south).

The general validity of this type of depositional model is supported by tectonic and stratigraphic evidence that many thick evaporite deposits accumulated in relatively deep or subsiding basins. A marine source is fundamental to this model and is generally supported by the nature of interbedded or associated lithofacies. Moreover, the model requires a long-term, albeit episodic, influx of marine water as the source of solutes.

Perhaps the most significant new conceptual variation of the 'barred basin' hypothesis is that the brine level in the evaporating basin may have been depressed well below the level of the contemporaneous marine sea level source---the concept of 'evaporative drawdown' (Maiklem, 1971). Lowered brine levels during the evaporitic maxima require that the inflow volume of marine water (or other water sources) be greatly exceeded by the rate of evaporation, and that the 'bar' be a near-complete seal. For many interior basins such as the Elk Point and Michigan, this seal may have been formed largely by carbonate buildup across the marine seaway; also, tectonic events and periodic lowering of the oceanic sea level relative to the sill depth of the basin may have been significant factors. For rift basins, tectonic and local structural factors must have controlled access of oceanic water into the troughs, yet their initial geometry and climatic setting (long linear troughs in continental interiors) were ideal for the generation of evaporites (within latitudinal constraints).

The most controversial and recently documented example of a 'barred basin with evaporative drawdown' or a 'desiccated deep basin' in which evaporites were deposited is the Mediterranean area in Messinian (late Miocene) time. Geophysical data, and cores from Leg 13 of the Deep Sea Drilling Program have established the existence of widespread evaporites of Messinian age beneath a large area of the present Mediterranean Sea. These halite-dominant evaporites are up to 1000 m thick and exceed 1 000 000 km^3 in volume. Thick halite-dominant evaporites of the same age also underlie large areas of the Red Sea. Hsü et al. (1973) proposed that during Late Miocene time the entry of Atlantic seawater into the Mediterranean trough was blocked and the basin evaporated to near-total desiccation. Sedimentary structures and other phenomena were used by Hs. u et al. (ibid., and other papers) to suggest that at least the anhydrite and gypsum encountered by DSDP cores were emplaced at sabkha-like emergent surfaces. However, many of the criteria used to support this hypothesis were equivocal, and controversy continues as to the actual depth of the floor of the Mediterranean trough relative to Atlantic sea level in Messinian time, and to whether total desiccation occurred or was indeed necessary for deposition of the evaporites. The present consensus is that the Messinian evaporites were deposited in a spectrum of environments (Schreiber, 1974) ranging from emergent sabkhas in marginal areas to relatively deeper subaqueous brine environments. Most if not all of the thick Messinian salt in the Mediterranean (and the Red Sea area) probably accumulated below an extensive body of brine, the surface of which was depressed below the level of the Messinian Atlantic Ocean by evaporative drawdown.

Sabkha model of evaporite genesis

The current upsurge in interest in the sedimentology of evaporites originated largely in the discovery less than 15 years ago of anhydrite in the sabkha sediments along the Trucial Coast of the Persian Gulf. The principal stimulus to what was to become one of the more controversial issues in sedimentology was provided by Shearman (1966), who proposed that many evaporite deposits, particularly those interbedded with carbonates, may be the product of diagenetic emplacement of anhydrite within a sediment wedge below a sabkha (supratidal) surface. The characteristic succession for a sabkha-generated evaporite is considered to be a basal lagoonal carbonate (subtidal environment), grading up into an algal or stromatolitic carbonate (intertidal environment) and overlain by nodular and enterolithic anhydrite (supratidal environment). The sabkha model has been applied since 1966 to a wide variety of evaporite deposits in the geological record, unfortunately often without very critical assessment of the regional and local tectonic and facies settings of the evaporite, and without questioning some of the sedimentological criteria. As an example, the presence of nodular anhydrite fabrics often has been used uncritically as a criterion for a sabkha origin, although this is increasingly recognized as untenable (Dean et al., 1975).

Depositional spectrum

Currently evaporite deposition is seen less from one extreme view or the other, that is, deep basin versus sabkha, and more as the product of a spectrum or depositional continuum ranging from subaerial to subaqueous environments. After studying exposed Messinian evaporites in Sicily and comparing these rocks with the cored Messinian evaporites below the Mediterranean, Schreiber (1974) was one of the first to document this more balanced and realistic view.

In summary, the emerging consensus is that evaporites are formed in a variety of environments ranging from subaerial sabkhas to subaqueous basins. Further, evaporation of the enclosed seawater mass results in lowering of sea level, or evaporative drawdown. The consequences of this process are that sabkha-like emergent environments must prograde around the periphery of the brine-filled basin, and that the more soluble halite must be concentrated in and precipitate out in the initially deeper axial zones of the basin.

Otto Fiord evaporites

In an earlier summary paper and abstract, we proposed a subaqueous depositional model for the anhydrite of the Otto Fiord Formation (Davies and Nassichuk, 1972, 1975). Wardlaw and Christie (1975) came to essentially the same conclusion on the basis of data accumulated by Christie (1975) for a Master of Science thesis study of Otto Fiord rocks along Hare Fiord. Our reasons (see table) for this conclusion are based on the following observations:

1. Axial position. The Otto Fiord evaporites occupy the axial, basin-centre position in the former Sverdrup trough. They overlie a thick succession of continental-derived red beds that incorporate the first evidence for marine transgression into the basin only tens of metres below the first anhydrite unit. The conglomeratic red beds probably were shed from fault scarps formed at initiation of the crustal downwarp leading to formation of the trough. Structures near Section 5, east of the head of Hare Fiord, suggest that local fault displacement probably was contemporaneous with the onset of evaporate deposition.

The Otto Fiord evaporites are, in part, coeval with carbonate rocks of shallow-water shelf aspect, at least toward the northwestern margin of the basin. Depositional relief on the coeval carbonates relative to the upper surfaces of interbedded anhydrite units increases upsection from a few metres to more than 200 m. In total, these facies relationships and axial distribution of the evaporites are consistent with a saline basin undergoing periodic evaporative drawdown, rather than emplacement on an emergent sabkha-like surface.

2. Thickness of anhydrite units. Anhydrite units in the Otto Fiord Formation typically are more than 10 m thick; many units exceed 30 to 40 m and a few are close to 60 m thick. Although extrapolation of expected thicknesses of anhydrite from the modern Trucial Coast sabkhas is dependent on predictions of tectonic and eustatic changes, sabkha cycles conceptually should be composed of anhydrite units only a few metres thick. Most anhydrite or gypsum successions interpreted as paleo-sabkhas have sulphate units ranging in thickness from less than 1 m (Shearman, 1966, Fig. 3, uppermost Jurassic; Llewellyn and Stabbins, 1970, Lower Carboniferous) to about 4 m (Wood and Wolfe, 1969, Jurassic-Cretaceous; Bosellini and Hardie, 1973, Upper Permian). Exceptions are the anhydrite units of the Lower Ordovician Baumann Fiord Formation on Ellesmere Island, which were interpreted by Mossop (1973) as paleo-sabkha deposits; these anhydrite units average 2 to 3 m in thickness but include some that are 30 m thick. However, the interpretation of some of these thick anhydrite units as products of sabkha emplacement is open to debate. In summary, the thickness of major Otto Fiord anhydrite units appears to be greater than that expected or predicted for an emergent sabkha environment, while being conceptually consistent with the volume of sulphate that could be produced by a body of brine (with renewal) undergoing increasing evaporation and drawdown.

3. Marine biotic composition of interbedded limestones. Limestone units interbedded with Otto Fiord anhydrite units typically contain a marine to hypersaline (euryhaline) biota up to the contact with overlying anhydrite. Further, the typical vertical succession of biotas appears to record evolution of depositional environments from initial hypersaline (?transgressive mixing) to more normal marine conditions with an abundant and varied biota, followed finally by increasing hypersalinity (algae, oncoids). The gradational transition from carbonate to anhydrite through an interlaminated zone records the onset of sulphate supersaturation in the hypersaline brine; calcitic laminae may be partly the product of postdepositional bacterial reduction of precipitated sulphate (Friedman, 1972). Although the lagoonal carbonate facies of a typical sabkha cycle should also contain marine fossils, and the intertidal facies should contain euryhaline biotas, the biotic composition and succession in the Otto Fiord carbonate rocks are consistent with the proposed subaqueous depositional model.

4. Bathymetric relief on interbedded marine carbonates. The interpretation of depositional environments and processes for the crinoid-algal mounds in the Otto Fiord Formation at van Hauen Pass (Fig. 21) requires onset of deposition of bedded anhydrite in at least 20 to 30 m of hypersaline water. Continuing evaporative drawdown coupled with shallowing by accretion of gypsum beds on the floor of the saline sea probably reduced these depths progressively toward the end of the evaporite depositional cycle.

At Section 4 (near Girty Creek), relief on carbonate units interbedded with anhydrite increases upsection from a few metres to more than 200 m at the end of Otto Fiord deposition. This degree of relief may be translated into approximations to minimum water depths during deposition of the carbonate rocks. If evaporative drawdown is a

Table 1. Comparison of characteristics of Otto Fiord marine carbonate-anhydrite cycles and an idealized sabkha cycle

	Otto Fiord cycle Submarine environment		Sabkha cycle Subaerial environment
1.	Basin *axis* distribution	1.	Basin *margin* (shelf) distribution
2.	Shallow-marine carbonates 'landward' of evaporites	2.	Terrigenous/continental rocks 'landward' of evaporites
3.	Major salt facies in basin centre?	3.	Marine (carbonate) facies in basin centre
4.	Thick cycles common (10 to 50+m)	4.	Thin cycles common (1 to 3 m)
5.	High depositional topographic relief on some interbedded carbonates (30 to 400+m)	5.	Low depositional/topographic relief for sabkha complex
6.	Marine biota in carbonate units, continuing up to anhydrite contact; biotic variety increases upward through unit	6.	Marine biota in basal (subtidal) carbonate units, decreasing in variety and volume into intertidal subfacies
7.	Dolomite not common; late diagenetic when present	7.	Dolomite common; early diagenetic
8.	Carbonate-anhydrite transition marked by thin, parallel laminae of both phases, desiccation structures; stromatolites absent	8.	Intertidal transition marked by algal laminae (irregular), stromatolites, desiccation structures
9.	Discrete nodules and enterolithic layers of anhydrite absent at carbonate to anhydrite transition	9.	Nodules and enterolithic layers of anhydrite common at transition and in supratidal facies
10.	Regular parallel bedding 10 to 15 cm thick common in otherwise massive anhydrite units	10.	Regular thick bedding not found in sabkha deposits
11.	Fabrics in anhydrite beds pseudomorphic after large vertical to inclined gypsum crystals	11.	Beds of subvertical gypsum crystals not found in sabkha deposits
12.	Some thin-bedded and parallel-laminated anhydrite	12.	Thin-bedded and parallel-laminated anhydrite not common, or absent
13.	Nodular mosaic fabric common in thick anhydrite cycles; considered secondary (diagenetic) fabric	13.	Coalesced nodular fabric present; primary to very early diagenetic

viable mechanism, and was operative in the Otto Fiord basin, accumulation of anhydrite units above interbedded carbonates occurred under brine levels lower than the carbonate-forming sea levels. However, the lack of criteria supporting total desiccation at the evaporitic maxima and the increasingly extreme fluctuations of sea level required for alternating deposition of marine carbonates and sabkha evaporites argue persuasively against the extreme condition of total desiccation for evaporite accumulation. Rather, precipitation from an evaporating brine and crystal growth at the sedimentbrine interface under evaporative drawdown is a more reasonable mechanism.

5. Bedded anhydrite. The preservation of bedding, often at 10 to 15 cm intervals, in many units of Otto Fiord anhydrite records a depositional environment consistent with a subaqueous brine origin and is inconsistent with sabkha-like emplacement mechanisms. Although only a few bedded anhydrite units preserve internal detail of former gypsum crystal fabrics in these beds, traces of bedding intervals in thick units of nodular mosaic anhydrite suggest that many of these units also were formed by a similar process. Individual beds are traceable as continuous planar entities as far as local exposures allow and thus appear comparable to the basic bedding characteristics of other sedimentary rock types of deeper water subaqueous origin. Schreiber (1974, Fig. 23) has illustrated continuity of bedded Miocene gypsum with similar crystal fabrics over 15 km in Sicily.

Hardie and Eugster (1971) and Schreiber (1974) have described and illustrated Miocene examples of bedded gypsum deposits which they considered to have been formed by subaqueous crystal growth at the sediment interface below a hypersaline water body. Similar bedded gypsum fabrics form today in natural and artificial saline lakes and evaporating pans (Schreiber, ibid.). The selenitic gypsum crystals grow in divergent clusters upward from the sediment substrate, so that inclined crystal orientations are common; a similar pattern is seen in the crystal pseudomorph fabrics in Otto Fiord bedded anhydrite. Bedded crystal aggregates of this type are not found in emergent sabkha environments.

6. Absence of subaerial indicators. Diagnostic evidence for subaerial exposure below, within or above anhydrite units in the Otto Fiord Formation is absent. No desiccation cracks, flat-pebble conglomerates, discrete nodules or enterolithic layers of anhydrite, algal stromatolites or laminae, or other sedimentary structures indicative of subaerial exposure have been found in any part of the succession, including specifically the transitional zones from carbonate to anhydrite that would be expected to record periodic intertidal exposure in a sabkha cycle. The absence of these indicators is consistent with a subaqueous depositional model.

7. Rarity of dolomite. Dolomite is a rare component of the Otto Fiord carbonates that are interbedded with anhydrite. Where present, it is encountered more

commonly in the lower unfossiliferous beds of limestone units, and usually in fabrics suggesting a later diagenetic rather than penecontemporaneous origin. In contrast, dolomitization of aragonite-dominant sabkha sediments is common on the Trucial Coast of the Persian Gulf and in similar supratidal sediment wedges elsewhere (Bahamas, Shark Bay). Wardlaw and Christie (1975, p. 169) argued that the relative absence of dolomite in the Otto Fiord Formation may be expected because, for each sedimentary cycle, the marine carbonate sediments are sealed and buried by the relatively impermeable beds of gypsum and thus effectively isolated from the dolomitizing potential of the overlying calcium- and sulphate-deficient brine. Sabkha sediments, on the other hand, are open to repeated flood recharge and evaporative pumping mechanisms that create efficient lateral and vertical migration of saline groundwater resulting in sulphate precipitation and extensive dolomitization.

Collectively, the characteristics of the Otto Fiord Formation discussed above are persuasive arguments for a subaqueous hypersaline depositional origin for the Otto Fiord anhydrite units. Few of them stand alone as unequivocal criteria for such an origin; perhaps only the bedded anhydrite with crystal pseudomorph fabrics may be close to a diagnostic criterion for this type of environment.

REGIONAL TECTONIC AND ENVIRONMENTAL SETTING OF THE SVERDRUP BASIN IN OTTO FIORD TIME

Paleomagnetic reconstruction by Smith et al. (1973) and Roy (1973) suggest that in Early Carboniferous time the developing Sverdrup trough lay between 10° and 25° N, with the axial trend of the trough roughly parallel with the contemporaneous equator. Assuming the same type of global atmospheric circulation and climatic patterns as at present, the incipient Sverdrup trough lay within the northern subtropical, high-pressure belt characterized by warm seasonal climates and an excess of evaporation over precipitation.

The paleomagnetic reconstructions by Smith et al. (1973) also show that in Early Devonian time the area encompassed by the present Arctic Archipelago lay slightly south of the Equator. The equatorial climatic zone, therefore, must have migrated southward across this region during Late Devonian and Mississippian time. Significantly a trace of this humid, tropical climate and the accompanying prolific land plant growth is preserved in isolated remnants of the Viséan Emma Fiord Formation on northwestern Ellesmere Island and northern Grinnell Peninsula of Devon Island. On Grinnell Peninsula, black, nonmarine carbonaceous and coaly shale with abundant plant fossils lies in erosional depressions on older Paleozoic carbonates, and is overlain abruptly by coarse red conglomerate of the Sverdrup Basin-margin Canyon Fiord Formation.

This succession, therefore, in part records the latitudinal change from an equatorial humid climate to a subtropical, seasonally arid climatic region, and coincides with the onset of deepening of the Sverdrup Basin.

Formation of the Sverdrup Basin began at least by Late Mississippian time. It developed as an elongate, pericratonic depression superimposed on deformed and eroded rocks of the Franklinian Geosyncline. The actual mechanism of basin formation is not known, but probably involved crustal thinning and downwarping in a linear zone approximately coincident with the former axis of the Hazen Trough of the Franklinian Geosyncline. The eastern and southern boundaries of the developing basin were formed by emergent and eroded Devonian and older Paleozoic rocks. In many areas, the Sverdrup Basin boundary roughly parallels the former Ordovician to Devonian Franklinian Geosyncline shelf edge that, in turn, probably was controlled and localized by deep-seated structural elements. The western and northwestern boundaries of the basin are less clearly defined; they probably were active tectonically and variable in position and topographic expression throughout the history of the basin. The northwestern boundary of the basin generally coincides with the former position of the tectonically active early Paleozoic Pearya Geanticline (Trettin et al., 1972); eroded residuals of this structural element define the present erosional margin of the basin on Ellesmere and Axel Heiberg islands, and underlie the younger Paleozoic and Mesozoic rocks of the Sverdrup succession in the subsurface below Brock, Ellef Ringnes and other westernmost islands. An unknown factor is the presence or absence of a proto-Arctic Ocean west of this boundary in late Paleozoic time.

Active subsidence of the Sverdrup trough probably took place along faults or fault zones that developed surface relief and provided the source for the thick wedges of coarse conglomerates, sand and finer clastics that were shed across parts of the northern trough (Borup Fiord Formation). Climatic aridity, probably seasonal in type, resulted in the formation of ferric oxide coatings on clasts, and preservation of these continental-derived rocks as classical red beds. Initial faulting along the northwestern boundary of the Sverdrup Basin apparently was accompanied by volcanism and extrusion of vesicular flows (Audhild volcanics) preserved on northwesternmost Ellesmere Island.

Flooding of the deepening Sverdrup trough by marine water in Late Mississippian time heralded the onset of the first marine sediment accumulation. The source of the marine water is open to speculation. The biota of upper Paleozoic limestone in the Sverdrup Basin has close affinities with the biota of the Russian Platform and other Arctic regions and less with that of the midcontinent of North America. This pattern suggests that marine water entered the Sverdrup trough by way of one or more seaways entering the northern or northwestern margin of the trough or, perhaps, by flooding over a shallow western and northwestern sill from a proto-Arctic Ocean. This interpretation is supported in a crude way by the distribution of halite and anhydrite within the basin. Salt deposits are restricted to one or more depocentres in the central and southern part of the basin, whereas anhydrite is found in a belt that extends across northwestern Ellesmere Island and can be correlated with anhydrite on Spitzbergen to the north. Unfortunately, the amount of anhydrite that is associated with the major depocentres cannot be estimated. However, if this pattern of the more soluble salts occurring in the southern part of the basin holds true and compares favourably with the distribution of evaporite mineral species in the evaporite basins such as the Devonian Elk Point Basin, the conclusion is that it supports a marine water source from the north, rather than a general spillover from a proto-Arctic Ocean to the west.

Initiation of marine sedimentation and first evaporites, northern basin

Because Otto Fiord rocks in normal stratigraphic succession are exposed principally on northwestern Ellesmere Island, the interpretation of depositional events must, of necessity, concentrate on the carbonate-anhydrite and carbonate-anyhdrite-shale/sandstone rhythms of that region. Inferences are drawn, where relevant, to salt deposition in the central and southern basins.

The first marine transgressions into the northern basin in the vicinity of eastern Hare Fiord were coincident with continuing accumulation of iron oxide-coated conglomerates, sandstones and finer detrital sediments (upper

units of Borup Fiord Formation). Very shallow water conditions are indicated by the discontinuous lenses of oolitic grainstones mixed with marine skeletal debris of shallow-water, perhaps euryhaline, aspect (algae, abundant foraminifers of a few species).

The first thick laterally extensive and relatively pure marine limestone appears above the Borup Fiord red beds at the base of the Otto Fiord Formation. Structural evidence for local fault displacement of the red beds before deposition of these carbonates and earliest evaporites has been documented, yet elsewhere the apparent transitional boundary between the two formations suggests near-contemporaneity of the end of red-bed deposition and the onset of the major marine phase.

The basal limestone units in the Otto Fiord Formation, and the basal 'platform' carbonate succession east of Girty Creek (Sec. 4; Fig. 6) are characterized by sediments of shallow-water aspect (abundant algae, some ooids). Bedding is planar and thus inferred depositional relief is low. Field relationships and regional facies patterns for the Girty Creek exposures indicate that the shallow-marine carbonates at the base of the section are the basin-margin equivalent of the Otto Fiord evaporites. Higher in that section the first thin anhydrite beds appear, pinching out between carbonate beds but finally burying the platform.

Deposition of sulphate evaporites in the lower, low-relief part of the Otto Fiord Formation is assumed to have been initiated by a reduction in replenishment of the marine transgressive sea. Whether this was purely a function of tectonics, of eustatic oceanic sea level changes, or simply a product of intense evaporation over the shallow, elongate trough far from the seaway entrance, is debatable. Whatever the cause, concentration of the residual seawater into a gypsum-saturated brine, accompanied by a reduction in brine depth by evaporative drawdown, resulted in accumulation of the first deposits of calcium sulphate. Some of this sulphate may have precipitated from the air-water interface or from a mixing zone between compositionally stratified water masses, whereas some may have nucleated and grown as gypsum crystals at the sediment-brine interface. Gypsum is inferred to have been the principal initial sulphate product under the temperature and pressure conditions of the subaqueous hypersaline setting.

Deposition of anhydrite

The intertonguing between carbonates of the Nansen Formation and anhydrite of the Otto Fiord Formation near the Girty Creek facies change demonstrates depositional relief in excess of 200 m on steeply dipping carbonate units that thin out downslope between essentially horizontally bedded anhydrite units. Relief on the carbonate units appears to increase progressively for successive interbeds up through the section, apparently as a response to progressively deeper marine transgression into the basin.

Evaporite units that pinch out against the sloping toes of these steeply dipping carbonate units again are inferred to result from subaqueous precipitation or crystal growth in brines undergoing evaporative drawdown. A corollary of this process is that the upper parts of the adjacent steeply inclined carbonate must have been exposed subaerially at least at the evaporitic maxima.

Similar reasoning applies to the rhythmic deposition of carbonate and anhydrite units in the type area and throughout the interval in which the crinoid-algal mounds formed. The regional trend toward progressively deeper marine transgressions correlative with the later stages of Otto Fiord deposition is supplied by the progressive thickening of successive limestone and anhydrite units in the type area and other sections.

Accumulation of halite in the central and southern sectors of the Sverdrup Basin probably coincided with periods of extreme evaporative drawdown in the northern basin. The central and southern basins or subbasins may have been bathymetrically deeper than the northern basin, and may have acted as a gravitational sink for halite-saturated brines. Active subsidence may have been greater in the central and southern basins than in the northern region in late Paleozoic time; certainly this is true for Mesozoic time.

Deposition of lithological cycles

Within the general tectonic and depositional framework outlined above for the Otto Fiord Formation, deposition of the several types of lithological cycles that characterize the formation are considered in more detail as follows.

Carbonate-anhydrite cycles

In the type area and in the lower part of most other sections throughout the Otto Fiord Formation, alternation of thick limestone and anhydrite units is characteristic. Because of the abrupt contact between the two rock types at the base of limestone units, and the common interlaminated and gradational contact at the top of the same units, the depositional cycle begins with carbonate sediment accumulation.

The sharp contact between anhydrite and limestone at the top of anhydrite units is interpreted as recording renewed transgression of marine source water into the Sverdrup trough. The contact commonly is regular and subplanar, with rare evidence for local solution or scour depressions at the anhydrite contact. The sharpness of the contact may reflect some degree of resolution of sulphates by the lower salinity transgressive sea.

The lowermost beds of carbonate in typical cycles are unfossiliferous, black, in places pyritic limestone with vague depositional laminae. Clotted micrite fabrics and ovoid to cylindrical pellets of apparent fecal origin, added to the lack of skeletal debris and preservation of laminae in these beds, indicate that the depositional environment was inhospitable for benthonic life. The pellets, if fecal, may have been produced by planktonic or other nonbenthonic organisms. Two general interpretations are derived from these observations: (a) initial transgression was rapid and deep, discouraging early colonization by benthonic organisms; and (b) initial transgression resulted in gradual mixing of marine water with the residual evaporitic brine mass, probably with some compositional and density stratification, so that evolution of the water mass to more marine composition that would support benthonic life was a gradational process. In this circumstance, some of the laminated calcitic sediments may have been formed diagenetically by bacterial reduction of gypsum precipitated from the residual brines during this transitional phase (compare with calcitic sediments of Red Sea: Friedman, 1972). We favour this general interpretation of initial transgressive mixing; it does not require total desiccation of the evaporite basin.

With continued influx of marine water, benthonic organisms were able to colonize the carbonate substrates. Where the transgressive sea was relatively shallow and perhaps more saline than normal seawater, algae of various types were prominent components of the colonizing biota. In thicker limestone units that accumulated in deeper transgressive pulses of normal marine salinity, a varied and prolific biota characterized by crinoids, brachiopods, bryozoans, foraminifers, sponges and rare ammonoids and trilobites was established on the seafloor. The greater mass of many limestone units accumulated under these conditions of a more or less normal marine environment.

The first evidence of increasing salinity with consequent biotic changes and reduction in prolific skeletal carbonate production is recorded in the carbonate pile by an increase in several types of algae, of intensely bored skeletal grains indicative of slower burial rates, and the growth of algal-foram-bryozoan oncoids. Limestone beds with these characteristics in the upper part of carbonate units are, in turn, overlain by thinner beds of dark, commonly pyritic limestone containing algae, ostracodes and tubular foraminifers of euryhaline aspect. Fecal pellets of the type found in the lowermost limestone beds again appear in some of the uppermost beds.

By the time the first thin laminae of gypsum were preserved at the transitional zone from limestone to anhydrite, strongly evaporitic conditions with sea level drawdown must have been established across the basin over sufficient time for the brine to become supersaturated with calcium sulphate. Some of the calcitic laminae preserved in the transition zone from carbonate-dominant to sulphate-dominant deposits may have been formed diagenetically by bacterial reduction of precipitated sulphate (ibid.). As the proportion of gypsum in the interlaminated zone increased, correlative with increasing hypersalinity, laminae became less distinct and the first unlaminated beds of gypsum were deposited. The nodular mosaic fabric preserved in the lower part of the anhydrite units obviously was not a primary depositional fabric, but a diagenetic reorganization of the sulphate and the minor amount of calcite and other impurities in the sediment. Whether this diagenetic reorganization into nodular mosaic fabric occurred while the sediment was still dominantly gypsum or whether it was a byproduct of the conversion of gypsum to anhydrite is not known. One clue to this type of diagenetic process and the formation of nodules is provided by samples from Section 5 (Fig. 6) in which progressive diagenetic replacement of micritic limestone (carbonate muds) by a crystalline aggregate of anhydrite culminates in an almost pure anhydrite rock with a crude nodular mosaic fabric. Centres of replacement in the carbonate began as small rosettes of anhydrite crystals, which lost their identity as replacement proceeded. Nodular mosaic fabrics may have formed in a similar way by anhydrite replacement of a mixed gypsum-calcite mush.

Primary depositional bedding, commonly spaced at 10 to 15 cm intervals, is preserved in many units of Otto Fiord anhydrite. The inclusion-defined pseudomorphs after gypsum crystals preserved in some of these beds show that some, if not all, of the bedded deposits formed largely as dense aggregates of gypsum crystals that grew upward from the sediment-brine interface. It is inferred that, where such crystal pseudomorph fabrics are not preserved but traces of the same thickness and type of bedding are found in massive units of nodular mosaic anhydrite, later diagenetic reorganization and conversion of gypsum to anhydrite have been more extreme and have obliterated the primary gypsum fabrics.

Perhaps more than any other characteristic of the Otto Fiord evaporites, the bedded gypsum pseudomorph fabrics are considered a criterion for accumulation in a subaqueous hypersaline environment.

Bedded gypsum deposits that appear to be similar to the original depositional character of the Otto Fiord bedded crystal-pseudomorph anhydrite have been described from the Miocene of Sicily by Hardie and Eugster (1971, Figs. 4 - 6) and Schreiber (1974). The Miocene gypsum crystals (Fig. 45) were interpreted as having grown subaqueously at the sediment-brine interface. Many of the crystals grew as upward-diverging 'swallowtail' twins, so that crystals inclined to the plane of bedding are dominant over vertical crystals; the same pattern is shown by the crystal pseudomorph fabrics in Otto Fiord anhydrite (Fig. 15). Hardie and Eugster (1971, Fig. 21B, p. 215) compared the Miocene gypsum growth fabrics with selenitic gypsum crystals that grew on the floor of Marion Lake in South Australia. Schreiber (1974) described the same Miocene gypsum beds but in more detail and in stratigraphic context. She showed (ibid., Fig. 8) that these beds are part of a succession (lower 'Gessi inferiori') that are the lateral and, apparently, 'basinal' equivalent of limestones; a similar relationship was established for the Otto Fiord anhydrites. Schreiber also demonstrated precise correlation of beds of gypsum crystals up to 5 cm thick between sections that are 15 km apart (Fig. 46), thus adding further support to her interpretation that these bedded deposits accumulated subaqueously by crystal growth at a sediment-brine interface. Similarities between the Miocene crystal fabrics and gypsum crystals formed in artificial salt pans in California (ibid., Fig. 89) confirm this conclusion.

The thinner, inclusion-rich upper layer of each bedding couplet is interpreted as representing a reduction in salinity toward more normal marine levels. Partial solution of the uppermost faces of the bedded gypsum crystals may have occurred at this time; some crystal pseudomorphs in the Otto Fiord anhydrite are capped with irregular concentrations of impurities (Fig. 15) that may record such a process. Solutional events were recorded similarly in subaqueous gypsum crystals in Marion Lake and the Miocene rocks of Sicily (Hardie and Eugster, 1971; Schreiber, 1974). Precipitated and mechanically introduced micritic carbonate may have accumulated above the gypsum crystals at this time, later to be invaded and largely replaced by early diagenetic (syngenetic) gypsum at the onset of the next evaporitic bedding cycle.

Anhydrite breccias in at least one unit in the type area of the Otto Fiord Formation are interpreted as subaqueous mass flow of slump breccias. They are underlain and overlain by undisturbed bedded and laminated anhydrite and thus were formed penecontemporaneously with deposition. Schlager and Bolz (1977) have described similar breccias in the Permian Zechstein anhydrite in Germany and attributed their formation to mass flow and slumping on the foreslope of a constructional sulphate platform; the same environmental setting cannot be demonstrated for the Otto Fiord anhydrite and breccias.

The anhydrite depositional episode for each cycle was terminated by renewed marine transgression and the subsequent onset of carbonate sedimentation of the next cycle.

Carbonate-anhydrite-shale/sandstone cycles

In Section 8 near Wood Glacier, thick mudstone and sandstone units are interbedded with anhydrite and limestone in the upper part of the Otto Fiord Formation. The detrital rock units commonly occur above thick anhydrite units and are capped by limestone units. Because many of the limestone units in this section have gradational contacts with overlying anhydrite, similar to the carbonate-anhydrite cycles of the type area and elsewhere, the depositional cycle is interpreted as beginning with carbonates and terminating at the top of the detrital sediment pile.

Environments and depositional processes are inferred to be the same for the carbonate and anhydrite units in this type of cycle as they are in the basic carbonate-anhydrite cycles discussed previously. One significant difference is that, in at least one of the major cycles, the base of the limestone unit overlying a thick sandstone sequence is composed of oolitic grainstone indicative of an environment of much shallower water and more agitated than normally found in Otto Fiord limestones. This difference,



Figure 45. Bedded selenite gypsum crystals with interbeds of fine gypsum sand from the Upper Miocene Solfifera Series of Sicily, polished slab. Note that there is an apparent preference for crystal growth or preservation to be inclined to the left; a similar trend is shown by the crystal pseudomorph fabrics in the Otto Fiord anhydrite beds (see Fig. 15). Hardie and Eugster (1971) interpret a subaqueous accretionary mechanism for these gypsum crystals. Reproduced courtesy of L.A. Hardie (from Hardie and Eugster, 1971, Fig. 4).

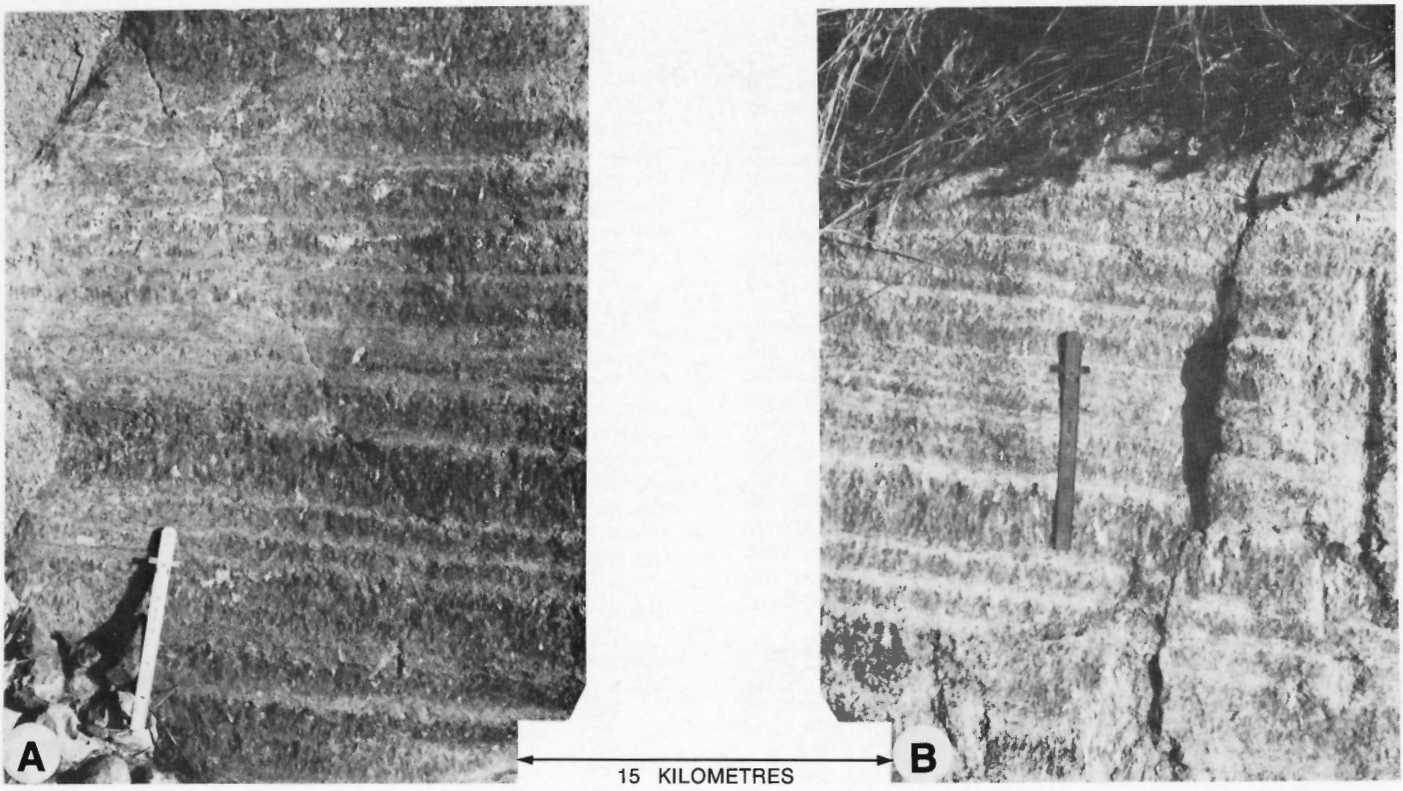


Figure 46. Matched beds of gypsum crystals 15 km apart from the Upper Miocene 'Gessi inferiori' in the Cattolics Basin of Sicily (from Schreiber, 1974, Fig. 23). This correlation demonstrates the lateral continuity and extent of the gypsum beds, and thus supports the subaqueous depositional mechanism proposed for these beds by Schreiber. Although the beds are thinner and less regular than those found in the Otto Fiord sulphates, they are considered to be analogs for the bedded crystal-pseudomorph sulphates of the Otto Fiord. Reproduced courtesy of B.C. Schreiber.

however, is not unexpected since the section is estimated to be close to the shallower eastern or northeastern margin of the evaporite basin and, further, the detrital sediment pile on which the carbonates accumulated may have built up into a subaerial or at least very shallow submarine setting.

Other differences include growths of colonial coral heads and more abundant algae throughout some limestone units, all suggestive of generally shallower water conditions compared with the limestone in the type area.

With the development of hypersaline conditions in the basin and the transition to sulphate accumulation, bedded and massive gypsum accumulated above the carbonates. In one thick anhydrite unit, the first sign of approaching detrital sediment input into the evaporite basin was deposition of apparently channelized, crossbedded sandstone composed of detrital gypsum and quartz sand. These lenticular gypsarenite units are interpreted as subaqueous channel fill or terminal fan deposits emplaced in front of a prograding delta wedge.

The detrital sediment unit is a major coarsening-upward succession with some fine sediments at the top. The fine, parallel laminated mudstone and shale at the base of the succession overlie anhydrite, and grade upward into coarse, scour-and-fill and crossbedded sandstone. These deposits are considered to be relatively localized deltaic wedges shed from eroding older Paleozoic rocks to the east and northeast. Fluvial streams may have been ephemeral, strongly influenced by abnormal seasonal precipitation; modern analogs include the ephemeral rivers and deltaic wedges entering Shark Bay in Western Australia (Davies, 1970a). Although neither of these modern deltas is associated with thick evaporite deposits, one of them, the Wooramel River delta, enters a hypersaline embayment and contains gypsum crystals and ephemeral halite layers (Davies, 1970b). Streams and deltaic deposits of this type, built out actively only during abnormal flooding episodes (hurricanes), appear to satisfy the condition of fluvial deltaic progradation into a hypersaline environment, which otherwise seems conceptually ambiguous.

Generation of the deltaic wedges probably was initiated by renewed uplift along faults close to the eastern and northern margin of the Sverdrup trough. However, since there is no way of demonstrating such a close genetic relationship between tectonics and sedimentation for these rocks, the simpler interpretation, that they represent the terminal phase of a regressive sequence closer to an eastern detrital sediment source, is retained.

Although a precise time-correlation has not been established, it is probable that the thick shale units in the upper part of Section 5 may be the distal and deeper subaqueous lithofacies equivalent of the coarsening-upward prograding detrital wedges of Section 8. However, as Section 5 is located about 65 km northeast of Section 8, and as the geometry of the evaporite basin in the vicinity of Section 5 is inferred to be narrowing northeastward, the sources of the detrital sediment in the two sections probably was local, although triggered and controlled by the same regional tectonic-climatic pulses.

The sandstone units in Section 9 in the southern Blue Mountains lack the thick packages of fine sediment in the coarsening-upward detrital cycles of Section 8. They may represent nearshore reworking and deposition of sand as bars or barriers. The discontinuous sandstone units between the crinoid-algal mounds in the type area near van Hauen Pass, now about 30 km north of the sandstone exposure of Section 8 (not allowing for palinspastic adjustment), may be the product of a major pulse of detrital sediment and fluvial outflow across the northern part of the evaporite basin. The few directional data for sediment transport from the intermound sandstone units are consistent with a source from the northeast.

Significance of calcite laminites

In at least three sections where the Otto Fiord Formation is overlain apparently conformably and transitionally by cherty argillaceous limestone of the Hare Fiord Formation, the uppermost units of the Otto Fiord Formation are composed of light-coloured calcite laminites (Fig. 30). Examples are found in Section 2 in the type area, in Section 7b in the Krieger Mountains, and in an undescribed section near Stepanow Creek on Hare Fiord (Fig. 1). In Section 2 the laminite is about 2.5 m thick, and as much as 24 m thick in Section 7b.

The laminites are composed of parallel-laminated layers of subhedral to euhedral calcite crystals in the general size range of 20 to 50 µm. Within each lamina, crystal size may vary either by decreasing ('normal' grading) or increasing ('reverse' grading) upward, or by increase followed by decrease. The base of many laminae may include detrital calcisphere-like grains (Fig. 30B), flocculent organic detritus, or a few quartz silt grains; the latter usually serve as nuclei for later authigenic quartz crystal growth. The boundaries between adjacent laminae commonly are stylolitic on a microscale. Many laminae are porous, with intercrystalline porosity generally occluded by bitumen.

In Section 7b the uppermost laminite unit contains textures interpreted to be pseudomorphic after early diagenetic calcite crystals (Fig. 30C), but now preserved as calcite. The gypsum crystals grew by replacement and displacement within the laminated sediment. This occurrence lends further weight to the evidence that the calcitic laminites are more closely related genetically to the Otto Fiord depositional environment and sediments than to the overlying Hare Fiord Formation. From a sedimentological point of view, therefore, the top of the Otto Fiord Formation should be placed at the top of the uppermost calcite laminite, if laminite is present, or the top of the uppermost anhydrite unit if it is absent.

The textural evidence indicates that the calcite fabric of the laminite is a diagenetic phenomenon superimposed on a primary depositional lamination. The lateral persistence and even, parallel-laminated morphology of these rocks, and their apparent restriction to the specific environmental conditions marking the transition from evaporite-dominant to marine-dominant sedimentation, suggest that the primary component was a chemical sediment. There appears to be a choice: the primary sediment was either carbonate or sulphate. If carbonate, any precipitate or accumulate would probably be very fine grained crystalline material ('micrite'), which would retain a dense, darker signature after diagenesis. However, this is equivocal, and an aragonitic or calcitic precipitational precursor cannot totally be discounted.

The alternative mechanism for calcite laminite formation is that the laminae accumulated as sedimented gypsum and were replaced diagenetically by calcite. The apparent gypsum pseudomorphs preserved now as calcite texturally similar to the laminite calcite is direct proof of at least an early interstitial hypersaline brine condition, and of calcite replacement of gypsum as a viable mechanism.

The calcitic laminites are at least superficially similar in fabric and primary layering, if not present composition, to laminated gypsum (balatine) and carbonate in the Miocene evaporites of Sicily (Hardie and Eugster, 1971, Figs. 8, 19). Discrete selenitic gypsum crystals grow at inclined angles by displacement and probably also by replacement in these Miocene gypsum laminites (ibid., Fig. 19); they are comparable to the pseudomorphic fabrics found in the Otto Fiord laminites. Further, normal and reverse size grading of gypsum crystals are found in the Miocene laminae.

Hardie and Eugster (ibid., p. 217) interpreted the Miocene laminated gypsum deposits as mechanical storm layers rather than chemical varves, with the gypsum of detrital origin and the carbonate that forms the basal part of the laminae of skeletal origin. Schreiber (1974), on the other hand, working with the same rocks, argued more convincingly for a precipitational mechanism for the microcrystalline gypsum, modified by syngenetic processes. The authors of both papers, however, interpret the depositional setting as shallow subaqueous and hypersaline.

A mechanism for early diagenetic (syngenetic) replacement of precipitated subaqueous gypsum by calcite is bacterial sulphate reduction (Friedman, 1972; and others). By this mechanism, sulphate and reducing bacteria use the sulphate of gypsum as an oxidant for the detrital organic matter required for metabolism. Carbon dioxide produced in the bacterial oxidation of the organic matter combines with calcium from the decomposing gypsum to produce calcite. Hydrogen sulphide gas is produced as the byproduct of this degradation process. Frequently, the primary precipitated gypsum is preserved in the oxygen-rich waters of the shallow margins of evaporite basins, but is bacterially reduced to calcite in the oxygen-depleted deeper basin.

Carbon and oxygen isotopic analyses may help in establishing whether reduction of sulphate to form calcite was a probable mechanism for the Otto Fiord laminites.

We speculate that the calcitic laminites at the top of the Otto Fiord Formation represent the last pulse of hypersaline conditions imposed on the Sverdrup trough by evaporative drawdown. Unlike preceding evaporitic pulses, however, the depth of marine water was such that saturation with calcium sulphate did not proceed much beyond the threshold before marine conditions once again became dominant. The laminites, therefore, record an abortive pulse of evaporite accumulation and the onset of regional marine-basinal conditions.

STRUCTURAL AND ECONOMIC SIGNIFICANCE OF THE OTTO FIORD FORMATION

A large proportion of the estimated 15 trillion cubic feet of natural gas discovered in the Canadian Arctic Archipelago is in Triassic sandstone reservoirs in anticlines that probably have cores of Otto Fiord salt. The anticlines probably were formed by halokinesis rather than by regional tectonic compressional stresses (H.R. Balkwill, pers. com., 1976). Movements of salt may have been initiated by increased loading by deltaic-marine detrital sediments in Triassic time. Evaporites thus have played a prominent role in the entrapment of gas in the Arctic Islands. To date, no significant oil or gas has been discovered on the flanks of diapirs formed by Otto Fiord evaporites.

No evidence has been found for potash mineralization within evaporite units or significant enrichment in Pb, Zn or other metals in associated carbonates.

During the Eurekan Orogeny in Tertiary time, evaporites of the Otto Fiord Formation on Ellesmere Island and elsewhere acted as décollement surfaces for the development of asymmetric folds and associated reverse faults. Some of these reverse faults were mapped by Thorsteinsson (1974) as thrust faults (for example, the Blue Mountains Thrust fault). Locally diapiric masses of evaporites, perhaps with salt cores but exposed at the surface only as sulphate, were intruded at sites along these faults; the Hare Fiord Diapir, at the southern end of the Blue Mountains fault, is one example. The Otto Fiord evaporites undoubtedly acted as detachment surfaces for structures in other parts of the Sverdrup Basin during the Eurekan Orogeny.

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Appendix

FOSSIL COLLECTIONS FROM THE OTTO FIORD FORMATION TYPE AREA, NORTHEAST OF VANHAUEN PASS

Extensive fossil collections have been assembled from the Otto Fiord Formation in the type area northeast of van Hauen Pass, northern Ellesmere Island. The following were collected from Section 2, which is 2.5 km northeast of van Hauen Pass. At Section 2 the Otto Fiord is 402 m thick; it is overlain conformably by the Hare Fiord Formation but the base of the Otto Fiord is truncated by a major fault. Thus all measurements are taken from the faulted bottom of the section, which may not necessarily be close to the depositional base of the formation.

Ammonoid identifications by W.W. Nassichuk

GSC locality C-10677; 173 m above base of section:
Brannerоceras branneri (Smith)
Age: Early Pennsylvanian, Morrowan (Bloydian= Bashkirian)

GSC locality C-10690; 327 m above base of section:
Brannerоceras branneri (Smith)
Gastrioceras sp.
Age: Early Pennsylvanian, Morrowan (Bloydian= Bashkirian)

Calcareous foraminifers, algae and incertae sedis identifications by B.L. Mamet

GSC locality C-10677; 173 m above base of section:
apterrinellids
Asphaltina sp.
Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Biseriella sp.
Bradyina sp.
Bradyina cribrostomata Rauser-Chernoussova and Reitlinger
Calcisphaera sp.
Climacammina sp.
Diplosphaerina sp.
Endothyra sp.
Eotuberitina sp.
Neoarchaediscus sp.
Palaeotextularia sp.
Planospirodiscus sp.
Pseudoendothyra sp.
Pseudoglomospira sp.
Pseudoissinella sp.
primitive *Pseudostaffella* sp.
red algae
Rectangulina sp.
Tetrataxis sp.
Vicinesphaera sp.

GSC locality C-29163; 174 m above base of section:
Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Diplosphaerina sp.
Endothyra sp.
Neoarchaediscus sp.
Planoendothyra sp.
Planospirodiscus sp.
Tetrataxis sp.

GSC locality C-29164; 175 m above base of section:
Asteroarchaediscus sp.
Biseriella sp.
Calcisphaera sp.
Earlandia sp.
Endothyra sp.
new kamaenid genus
Neoarchaediscus sp.
Parathurammina sp.
Planospirodiscus sp.
Tetrataxis sp.

GSC locality C-10678; 182 m above base of section:
apterrinellids
beresellid filaments
Beresella sp.
Eotuberitina sp.
Neoarchaediscus sp.
Tetrataxis sp.
Uraloporella sp.

GSC locality C-29165; 185 m above base of section:
apterrinellids
Asteroarchaediscus sp.
Calcisphaera sp.
Diplosphaerina sp.
Eolasiodiscus sp.
Neoarchaediscus sp.
Planospirodiscus sp.
Pseudoglomospira sp.
Tetrataxis sp.

GSC locality C-10680; 188 m above base of section:
apterrinellids
archaediscids
calcisphaerids
endothyrids
Pseudoglomospira sp.
Tetrataxis sp.

GSC locality C-10681; 188.5 m above base of section:
apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Diplosphaerina sp.
Endothyra sp.
Endothyra of the group *E. similis* Rauser-Chernoussova and Reitlinger
Neoarchaediscus sp.
Planoendothyra sp.
Pseudoglomospira sp.
Tetrataxis sp.

GSC locality C-10684; 230 m above base of section:
apterrinellids
Asteroarchaediscus sp.
Calcisphaera sp.
Endothyra sp.
Neoarchaediscus sp.
Neoarchaediscus parvus (Rauser-Chernoussova)
Planoendothyra sp.
Planospirodiscus sp.
Tetrataxis sp.
Tetrataxis of the group *T. angusta* Vissarionova

GSC locality C-10686; 240 m above base of section:
apterrinellids
Eotuberitina sp.
Neoarchaediscus sp.
Orthovertella sp.
Planospirodiscus sp.

GSC locality C-29166; 255 m above base of section:
cf. *Anthracoporella* sp.
apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Bradyina sp.
Calcisphaera laevis Williamson
Endothyra sp.
Eostaffella sp.
Millerella sp.
Neoarchaediscus sp.
ozawainellids
Pseudoendothyra sp.
Pseudoglomospira sp.
Stacheiinae

GSC locality C-10689; 289 m above base of section:
cf. *Anthracoporella* sp.
apterrinellids
Asteroarchaediscus sp.
Asteroarchaediscus baschkiricus (Krestovnikov and Teodorovitch)
Asteroarchaediscus cf. *A. gnomellus* Brenckle
Calcisphaera sp.
Diplosphaerina sp.
Earlandia sp.
Endothyra sp.
Eotuberitina sp.
Neoarchaediscus sp.
Ortonella sp.
Rectangulina sp.
Stacheiinae
Tetrataxis sp.
Ungdarellaceae

GSC locality C-29167; 309 m above base of section:
algae, undetermined genus
cf. *Anthracoporella* sp.
Asphaltina sp.
Asteroarchaediscus sp.
Beresella sp.
beresellid algae
Bradyina sp.
Calcisphaera sp.
Diplosphaerina sp.
Eostaffella sp.
Eotuberitina sp.
kamaenid algae
Millerella sp.
Neoarchaediscus sp.
ozawainellids
Pseudostaffella sp.
Zellerina sp.

GSC locality C-29168; 315 m above base of section:
cf. *Anthracoporella* sp.
Asteroarchaediscus sp.
Diplosphaerina sp.
Eotuberitina sp.
Neoarchaediscus sp.
ozawainellids
Zellerina sp.

GSC locality C-10690; 349 m above base of section:
apterrinellids
Asteroarchaediscus sp.
Biseriella sp.
Calcisphaera sp.
Earlandia sp.
Endothyra sp.
Neoarchaediscus sp.
Tetrataxis sp.

GSC locality C-10692; 393 m above base of section:
cf. *Anthracoporella* sp.
apterrinellids
Asphaltina sp.
Asteroarchaediscus sp.
Calcisphaera sp.
Eotuberitina sp.
Endothyra sp.
cf. *Komia* sp.
Neoarchaediscus sp.
Neoarchaediscus grandis (Reitlinger)
ozawainellids
Rectangulina sp.
Tetrataxis sp.
Stacheiinae
Ungdarellaceae
Age: (GSC locs. C-10677 - C-10692) late Early Pennsylvanian or early Middle Pennsylvanian; late Morrowan or early Atokan (Zone 20 or lower part of Zone 21)

Conodont identification by K.P. Bender

GSC locality C-10676; 71 m above base of section:
Streptognathodus lateralis Higgins and Bouckaert
Gnathodus noduliferus (Ellison and Graves)
Idiognathodus delicatus Gunnell
Hindeodella sp.
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian)

GSC locality C-10677; 173 m above base of section:
Streptognathodus lateralis Higgins and Bouckaert
Gnathodus noduliferus (Ellison and Graves)
Idiognathodus delicatus Gunnell
Ligonodina sp.
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian)

GSC locality C-10681; 188 m above base of section:
Idiognathodus delicatus Gunnell
Ozarkodina delicatula (Stauffer and Plummer)
Hindeodella sp.
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian)

GSC locality C-10684; 230 m above base of section:
Streptognathodus lateralis Higgins and Bouckaert
Gnathodus noduliferus (Ellison and Graves)
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian)

GSC locality C-10686; 240 m above base of section:
Streptognathodus lateralis Higgins and Bouckaert
Idiognathodus delicatus Gunnell
Hindeodella sp.
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian)

GSC locality C-10690; 349 m above base of section:
Gnathodus ouachitensis (Harlton)
Age: Early Pennsylvanian; Morrowan (Bloydian=Bashkirian) to Middle Pennsylvanian (early Atokan=early Moscovian)

GSC locality C-10691; 360 m above base of section:
Ozarkodina delicatula (Stauffer and Plummer)
Age: Early Pennsylvanian; Morrowan (Bloydian) to Middle Pennsylvanian (early Atokan)

GSC locality C-10692; 393 m above base of section:
Gnathodus ouachitensis (Harlton)
Ozarkodina delicatula (Stauffer and Plummer)
Age: Early Pennsylvanian; Morrowan (Bloydian) to Middle Pennsylvanian (early Atokan)

Brachiopod identifications by J.L. Carter

GSC locality C-10676; 71 m above base of section:
Crurithyris sp.
terebratulid gen. et sp. indet.

GSC locality C-10677; 173 m above base of section:
Rhipidomella sp.
Stenoscisma sp.
?*Martinia* sp.
indeterminate productid

GSC locality C-10681; 188 m above base of section:
Rhipidomella sp.
indeterminate davidsoniacean
?*Krotovia* cf. *K. pustulata* (Keyserling)
"*Marginifera*" *timanica* (Chernychev)
Karawankina sp.
?*Kutorginella* sp.
Wellerella sp.
Stenoscisma sp.
Coledium cf. *C. globulina* (Phillips)
?*Nucleospira* sp.
Neospirifer poststriatus (Nikitin)
Meristorygma arctica Carter
Tangshanella sp.
Martinia sp.
Phricodothyris sp.

GSC locality C-10682; 220 m above base of section:
Dielasma sp.

GSC locality C-10689; 289 m above base of section:
"*Marginifera*" cf. *Productus pseudoartiensis* Stuckenberg

GSC locality C-10690; 349 m above base of section:
Rhipidomella sp.
indeterminate small davidsoniacean similar to that of GSC loc. C-10681
Rugosochonetes sp.
"*Marginifera*" cf. *Productus pseudoartiensis* Stuckenberg
"*Marginifera*" cf. *M. pulchra* Rotai
Stenoscisma sp.
Coledium cf. *C. globulina* (Phillips)
Meristorygma arctica Carter
?*Eomartiniopsis* sp.
Crurithyris sp.
indeterminate small terebratulid

GSC locality C-10692; 393 m above base of section:
Rhipidomella sp.
"*Marginifera*" cf. *M. pulchra* Rotai
indeterminate small stenoscismatacean
?*Tangshanella* sp.
Crurithyris sp.
indeterminate small terebratulid
Age: According to Carter, "Brachiopods from the above seven localities indicate an Early Pennsylvanian, Morrowan (Bashkirian) to Upper Pennsylvanian, Missourian (Gzhelian) age; further age refinements will depend on larger collection of fossils that are suitably preserved for serial sectioning."

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