CARBONATE DEPOSITION DURING THE LATE PROTEROZOIC ERA: AN EXAMPLE FROM SPITSBERGEN

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ABSTRACT. Carbonate sediments reflect the physico-chemical and biological circumstances of their formation; thus, features of limestones and dolomites may provide insights into both environmental and evolutionary change through geological time. The Upper Proterozoic (approx 800–700 Ma) Akademikerbreen Group, Spitsbergen, comprises 2000 m of carbonates, with only minor intercalations of quartz arenite and shale. Although Proterozoic carbonates are often seen as predominantly dolomitic, the Akademikerbreen Group is about 45 percent limestone. Stromatolites are conspicuous in outcrop but constitute only 25 percent of the total section. Micrites and coarser intraclastic carbonates derived mainly from micritic precursors comprise 60 percent of the group, while oolites make up the remaining 15 percent. Distinctive sedimentary features of the group include giant (up to 16 mm) ooids, very early diagenetic calcite nodules and cements, micrites containing subaqueous shrinkage cracks filled with equant microspar cement, and strong $^{13}$C enrichment in both carbonates and co-occurring organic matter. The principal features of Akademikerbreen carbonates are widely distributed in coeval successions. However, these rocks appear to differ from older limestones and dolomites in their relative abundance of grainstones and, perhaps, micrites, as well as their paucity of tufa-like laminates and columnar or coniform stromatolites that preserve petrographic evidence of in situ precipitation as a dominant means of carbonate accretion. Upper Proterozoic carbonates also differ from Paleozoic accumulations, but the transition is not abrupt. Most changes accompanying the Proterozoic/Phanerozoic transition can be interpreted in terms of the consequences rather than the causes of metazoan and metaphyte evolution, including the evolution of biomineralization. Carbonate sedimentology reinforces data from other sources which indicate the last 200 to 300 Ma of the Proterozoic Eon was a distinctive interval of Earth history.

INTRODUCTION

In the present day oceans, CaCO$_3$ precipitation is predominantly a biological process mediated by skeleton-forming invertebrates, algae, and protozoa (Bathurst, 1975). Although the earliest known skeletonized organisms appeared near the end of the Proterozoic Eon, approx 600 Ma ago (Germs, 1972; Grant, 1990; Grant, Knoll, and Germs, in press), thick and laterally extensive carbonates are characteristic components of earlier Proterozoic sedimentary successions. How do these Proterozoic carbonates compare or contrast with the predominantly biogenic limestones and dolomites of the Phanerozoic Eon?

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Grotzinger (1989) recently reviewed Proterozoic carbonate deposits, focusing in particular on well-studied Lower Proterozoic ramps and rimmed platforms from Canada and South Africa. In these successions, dolomite is the predominant mineralogy and stromatolites the principal lithological feature. Although admittedly overgeneralized, this characterization appears to provide a good first order description of Early Proterozoic carbonates in general. But what about the later Proterozoic? Is it, like the earlier part of the eon, characterized by extensive stromatolites, tufa-like laminates, and dolomitic lithologies? Does it presage Phanerozoic-type deposits? Or, is the Late Proterozoic a distinctive era whose patterns of carbonate deposition must be considered as distinct from, if related to, both earlier and later patterns? In this paper, we discuss the mineralogy, lithologies, and characteristic sedimentary features of thick and laterally extensive carbonates from the Upper Proterozoic Akademikerbreen Group, northeastern Spitsbergen, and use this succession as the basis for asking more general questions about carbonate deposition during the Late Proterozoic Era.

**THE AKADEMIKBEREEN GROUP**

The Upper Proterozoic Lomfjorden Supergroup of northeastern Spitsbergen comprises nearly 6000 m of siliciclastic and carbonate rocks deposited in a rapidly subsiding intracratonic basin (Harland and Wilson, 1956; Wilson, 1958, 1961). Its two principal units, the predominantly siliciclastic Veteranen and largely calcareous Akademikerbreen groups, and their constituent formations can be traced along strike for some 120 km with only minor changes in facies (fig. 1): along with originally contiguous successions in neighboring Nordaustlandet and central East Greenland, it forms part of a significant locus of Late Proterozoic sedimentation whose original lateral extent was at least 650, and probably more than 1000, km. Although there are no radiometric dates that closely constrain depositional ages, acritarchs (Knoll, 1982a: Knoll and Swett, 1985; Butterfield, Knoll, and Swett, 1988; and unpublished data) and stromatolites (Raaben, 1969) both suggest a late Late Riphean age, perhaps 800 to 700 Ma ago, for Akademikerbreen deposition, with Veteranen sedimentation beginning on the order of 50 Ma earlier. This age estimate is consistent with the succession's stratigraphic position directly beneath Varangian tillites of the Polarisbreen Group (Wilson and Harland, 1964; Hambrey, 1982), as well as with chemostratigraphic data linking Akademikerbreen carbonates to other Late Riphean successions (Knoll and others, 1986; Derry and others, 1989).

**Pre-Akademikerbreen Carbonates**

Although massive limestones and dolomites begin at the base of the Akademikerbreen Group, three stratigraphically significant carbonate horizons are intercalated among the predominantly cross-bedded and rippled quartz arenites and carbonaceous shales of the 3800 m thick Veteranen Group (Wilson, 1958). The lowermost Veteranen unit ex-
Fig. 1. Map showing the outcrop distribution of Akademikerbreen Group rocks in northeastern Spitsbergen, as well as their stratigraphic equivalents in Nordaustlandet (after Hjelle and Lauritzen, 1982). Starred arrows indicate localities visited during the course of this investigation.
posed in northeastern Spitsbergen is a 250 to 300 m section of the Kortbreen Formation in which carbonates are interbedded with subordinate, thin-bedded carbonaceous shales, and sandstones (Wilson, 1958). In the Skinfjøsebreen area, where we examined these beds, the carbonates are predominantly laminated micrites and calcsiliites. Some beds display low amplitude cross-lamination, and there are intercalations of oolitic and microphytolitic limestone. Dolomitization is pervasive in the lowermost 10 to 20 m exposed (Wilson, 1958) but of minor importance higher in the section. The Kortbreen limestones become increasingly shaley in their uppermost 20 m and are succeeded abruptly by quartz arenites of the upper Kortbreen Formation.

A second carbonate-dominated sequence occurs about 2000 m above the first. Like the Kortbreen succession, the Bogen Limestone Member of the Kingbreen Formation (called the Cavendishryggen Limestones by Wilson, 1958) comprises approx 250 m of thin- to tabular-bedded micrites, calcsiliites, and calcarenitic, microphytolitic, and oolitic grainstones interleaved with carbonaceous shales and thin beds and lenses of quartz arenite (Wilson, 1958). Low amplitude cross-bedding, including herringbone cross-bedding, is common in the grainstones. Once again, dolomitization is pervasive in the basal few meters of the member; higher in the section dolomite and limestone are about equal in abundance.

The uppermost carbonates of the Veteranen Group occur in its youngest unit, the Oxfordbreen Formation (Wilson, 1958). In a sense, Oxfordbreen carbonates presage the onset of massive carbonate deposition in the overlying Akademikerbreen Group. Oxfordbreen carbonates consist mainly of decimeter to meter thick limestone beds within the predominantly shaley and sandy section. Dominant carbonate lithologies include cross-bedded calcarenites and oolites, as well as flake conglomerates. In some horizons laterally linked domal stromatolites up 0.5 m thick are intercalated among the grainstones. Dolomitization is uncommon in these beds, although thin, platy dolomitic shales do occur among the intervening shales.

**Akademikerbreen Carbonates**

The Akademikerbreen Group (fig. 2) constitutes the principal accumulation of Upper Riphean carbonate in northeastern Spitsbergen. Akademikerbreen carbonates were first described in stratigraphic detail by Wilson (1961), and they are currently the subject of continuing research (Swett and Knoll, 1989; Fairchild, Knoll, and Swett, in preparation). In this paper, we confine ourselves to a brief description of the group, stressing several major features relevant to a more general discussion of Upper Proterozoic carbonates.

**Grusdevbreen Formation.**—In northeastern Spitsbergen, more or less vertically continuous carbonates commence with the 850 m Grusdevbreen Formation, the basal unit of the Akademikerbreen Group. The lower 275 to 300 m of the formation comprises 6 to 15 cm beds of
fine grained intraclastic calcarenites and calcisiltites separated by thin intercalations of micrite and micritic shale. The calcarenites frequently exhibit unidirectional cross-lamination and, on bedding surfaces, low amplitude symmetrical ripples. This succession is interrupted at infrequent intervals by coarse intraformational conglomerates a few centimeters thick; stromatolites are rare.

Lower Grusdievebreen carbonates are characterized by a distinctive pattern of ovoid or irregular limestone nodules within otherwise dolomitic beds (Wilson, 1961; fig. 4A). Limestone distribution sometimes reflects sedimentary structures such as ripple marks or bedding, but, just as often, calcitic "islands" bear no obvious relationship to other features; limestone-dolomite contacts are generally diffuse. We interpret these beds as originally limy deposits in which CaCO₃ cementation to form nodules or irregular beds occurred very early in diagenesis. (Indeed, the coarse flat-pebble conglomerates—a lithology otherwise unusual in this succession—appear to reflect the intraformational reworking of already cemented carbonates.) Dolomitization subsequently affected much of the sequence, but pore waters were unable to alter sediments whose porosity had been occluded by the earlier cement. The Grusdievebreen "mottled" carbonates are succeeded by 100 m of tabular bedded, fine-grained black limestones with subordinate intercalated flake conglomerates.

The upper half of the formation consists predominantly of 5 to 25 cm beds of cross-bedded calcarenite, oolitic and microphytolitic grainstones, and flake conglomerates, with minor intercalations of micrite. Stratiform to columnar stromatolites appear in the upper 100 m of the formation, becoming more abundant (locally up to 30 percent of the sequence) toward its top. Pervasive dolomitization is largely restricted to these stromatolitic horizons.

Grusdievebreen sediments evidently accumulated in shallow subtidal environments (shoaling toward the top of the formation) subject to persistent wave or current activity and occasional storms.

Svanbergfjellet Formation.—The Grusdievebreen Formation is overlain without obvious unconformity by the Svanbergfjellet Formation, a lithologically variable sequence 450 to 550 m thick. The basal Lower Dolomite Member (approx 150 m thick) is a tidal flat/lagoonal complex consisting of dolomitic microbial laminites interbedded with thin dolarenites, dolomicrites, and intraclastic flake conglomerate, along with subordinate thickness of their limestone counterparts. Fenestrae and erosional surfaces up to several centimeters deep document frequent exposure, while silicified microfossil populations independently attest to
peritidal conditions. Domal stromatolites and small domed-shaped bioherms of radiating digitate stromatolites occur sporadically throughout the member. Centimeter-scale phosphorite nodules formed during early diagenesis (prior to carbonate cementation) in voids such as shrinkage cracks and the intergranular areas of flake conglomerates deposited under shallow subtidal conditions.

The member is capped by an 8 m thick biostrome of columnar stromatolites (*Minjaria* sp.) which can be traced without evident change throughout the outcrop area of the formation. This is the lowermost significant stromatolite build-up in the Lomfjorden Supergroup, and its broad lateral extent brings to mind the craton-wide distribution of columnar stromatolitic biostromes in the Atar Group of West Africa (Bertrand-Sarfati and Moussine-Pouchkine, 1988). Bertrand-Sarfati and Moussine-Pouchkine concluded that such horizons result from the rapid transgression of an epicratonic sea over a surface of minimal relief, an interpretation that fits the Svanbergfjellet biostromes well.

The overlying Lower Limestone Member consists primarily of laminated to 20 to 30 cm thick, tabular bedded micrites and calcisiliites, with occasional fine-grained calcarenites and only rare microbially laminated beds (fig. 3B). Wilson (1961, p. 100) called attention to the “extreme regularity of bedding” in this member, a probable result of deposition below at least fair weather wave base. This 150 m member is predominantly limestone, although variably dolomitized beds occur, especially in its lower part. Fine-grained limestones commonly contain shrinkage cracks filled by microspar cement.

The third member of the Svanbergfjellet Formation is the Algal Dolomite Member, a 120 to 150 m unit characterized by black to maroon shales interbedded with 1 to 3 m high, cabbage-shaped, irregularly columnar stromatolitic bioherms sometimes linked to form laterally extensive biostromes with undulating bedding surfaces. The stromatolites consist almost entirely of dolomite. The virtually complete separation of siliciclastic sediments and carbonates into distinct beds is problematic. Perhaps the shales record discrete pulses of continental run-off that smothered microbial mat communities. Alternatively, Grotzinger (personal commun.) has suggested similarities between this member and shallowing upward shale/carbonate parasequences in the Lower Proterozoic Rocknest Formation.

The Algal Dolomite Member is overlain gradationally by the Upper Limestone Member, a 70 to 150 m sequence consisting of dark, tabular bedded micrites and shale, with interbedded megarippled calc- and dolarenites and a few calcirudites. Like the Lower Limestone Member,

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Fig. 3(A)  Backlundtoppen nunatak, near the southern limit of Akademikerbreen outcrop, showing exposures of the Draken and Backlundtoppen formations. Dark strata on the left belong to the Polarisbreen Group. (B) Tabular-bedded micrite in the Lower Limestone Member of the Svanbergfjellet Formation; person for scale. (C) Outcrop of the Draken Conglomerate Formation, showing broad, lenticular beds of intraclastic dolomite and thin, interbedded dolomicrite; person for scale.
Fig. 4(A)  Mottled limestone/dolomite of the lower Grusdievbreen Formation; dark patches are calcitic, light are dolomitic.
(B and C) Subaqueous shrinkage cracks in fine-grained carbonate from the Backlundtoppen (B) and Draken Conglomerate formations. Note in (C) that the microspar-filled cracks are limited to the lutitic bed. Scale bar in (C) = 20 cm for (A), = 7 cm in (B), and = 10 cm in (C).
this unit appears to have accumulated in a relatively quiet subtidal environment subject to occasional storms of varying magnitude.

In sum, the Svanbergfjellet Formation begins with a regressive facies but higher in the section documents one and perhaps two transgressions whose sedimentary products range from dolomites formed in wave-swept peritidal environments to limestones deposited under subtidal conditions below wave base for all but the strongest storms.

Draken Conglomerate Formation.—Abruptly overlying the Svanbergfjellet Formation is the 150 to 250 m thick Draken Conglomerate Formation. Almost entirely dolomitic, this formation is characterized by broadly lenticular beds of intraclastic dolomite, with sparse to abundant rudite clasts of dolomicrite set in a dolarenitic matrix (figs. 3C, 6A). These units are interbedded with subequal thicknesses of dolomicrite, along with rare microbially laminated beds up to a few meters thick. Oolites and microphytolites are significant in the middle part of the succession, and stratiform stromatolites, some disrupted by soft-sediment deformation and tepees, are common near the top. Desiccation cracks occur at a few horizons in the sequence, but petrological evidence for emergence (leaching, artesian cements) is common (Fairchild, Knoll, and Swett, in preparation).

Columnar stromatolites are rare, except for a conspicuous horizon of 5 m thick bioherms (Inzeria sp.) near the base of the formation in the southern region of outcrop (Swett and Knoll, 1985; fig. 5). Although basal Draken Formation by definition (Wilson, 1961), the sequence that includes the bioherms consists mainly of tabular bedded microphytolitic grainstones and micrites containing subaqueous shrinkage cracks, much like the upper Svanbergfjellet Formation. Thus, the columnar stromatolites accreted subtidally prior to the major regression that ushered in typical Draken sedimentation. The stromatolites display an unusual petrographic fabric—their laminae are defined by populations of carbonate-lined filamentous sheaths, much like Girvanella (Raaben, 1969; Swett and Knoll, 1985). These structures provide evidence for in situ carbonate precipitation in stromatolite accretion, although interlaminar dolomictes contain scattered siliciclastic grains and vase-shaped planktonic microfossils indicative of trapping and binding (Swett and Knoll, 1985; Fairchild, Knoll, and Swett, in preparation).

Although in situ stromatolites constitute a limited percentage of the remaining section, a fair proportion of the abundant carbonate clasts are ripped up and redeposited mat fragments (as determined from silicified clasts containing preserved mat populations, Knoll, 1982b).

The Draken Formation represents an extensive tidal flat/lagoonal complex deposited in a series of protected subtidal to inter- and supratidal environments; muds and microbial mats deposited during relatively quiet periods were ripped up and redeposited during storms. Oolites record periods of minor transgression.

Backlundtoppen Formation.—The uppermost formation of the Akademikerbreen Group is the 500 to 600 m thick Backlundtoppen Forma-
Fig. 5. Outcrop view of the uppermost Svanbergfjellet and lower Draken Conglomerate formations on Svanbergfjellet nunatak. By Wilson’s (1961) definition, the formational boundary lies at the base of the buff quartz arenite bed visible at the bottom of the talus apron in the middle of the section. Above that are conspicuous columnar stromatolitic bioherms and, about 5 m higher, a sharp transition from dark thin-bedded limestones to the intraclastic, tidal flat dolomite characteristic of Draken carbonates.

tion. Basal Backlundtoppen carbonates include 30 m of black, tabular bedded micrites (commonly with microspar-filled syneresis cracks) overlain by some 150 m of cross-bedded and megarippled oolites and pisolites (actually consisting of unusually large ooids; Swett and Knoll, 1989; fig. 6B and C). The coated grains are predominantly calcareous, although partial to complete dolomitization and silicification are not uncommon. Micrites and, less abundantly, stromatolites and associated flake conglomerates occur as interbedded components of the oolitic/pisolitic unit (Swett and Knoll, 1989). A 50 m sequence of columnar stromatolitic and intraclastic dolomite overlies the oolites, and this in turn is overlain by 60 m of mixed black micrites, oolitic limestones, calcarenites, and small oncèles. This heterogeneous sequence gives way to 180 to 250 m of massive stromatolitic dolomites and interbedded intraclastic dolomites derived primarily from eroded stromatolites. This unit is overlain by 25 m of gray shaley limestone, 60 m of shaley to sandy
Fig. 6(A)  Intraclastic flake conglomerate sitting atop erosionally truncated dolomicroite in the Draken Conglomerate Formation. (B) Cross-bedded oolite (pisolite) in Backlundtoppen equivalents from central East Greenland. (C) Pisolite with microspar cement from the same unit as (B). Bar in (B) = 6 cm in (A), = 10 cm in (B), and = 4 cm in (C).
siliciclastic beds, and a laterally extensive 10 to 30 m thick unit of laterally linked coniform stromatolites (*Ephyaltes* sp.; Knoll, Swett, and Burkhardt, 1989). The uppermost carbonate beds are dolomitic and are marked by syn-sedimentary breccias throughout northeastern Spitsbergen (Fairchild and Hambrey, 1984).

The predominantly oolitic and pisolithic lower half of the formation records shallow subtidal environments subject to persistent wave or current activity; endolithic microfossils preserved in silicified grains are closely similar to populations found in Recent ooids from the Bahama Banks (see Knoll, Swett, and Burkhardt, 1989; and Green, Knoll, and Swett, 1988, on correlative beds from East Greenland). The massive dolomitic stromatolite and intraclastic flake conglomerate unit appears to represent shallow subtidal conditions of intermittent rather than persistent wave action. Quiet subtidal conditions at the close of Akademikerbreen time are documented by the laterally persistent coniform biostratomes (Hoffman, 1976; Bertrand-Sarfati and Moussine-Pouchkine, 1985, 1988; Kerans and Donaldson, 1989). It may be noted that the largest along-strike differences observed within the Akademikerbreen Group and its equivalents are found in the uppermost 300 m. The thick stromatolitic dolomites that dominate upper Backlundstoppa strata in Spitsbergen do not occur in central East Greenland. Instead one finds carbonate turbidites interpreted by Herrington and Fairchild (1989) as indications of renewed basinal extension and subsidence.

**DISCUSSION**

Akademikerbreen outcrop is essentially two dimensional, providing little direct evidence of facies variation perpendicular to strike. Nonetheless, vertical facies relationships strongly suggest comparison to Read's (1985) models of carbonate ramps. Lower Grusdievbreen sediments document shallow subtidal sand flats with little evidence of lagoons, consistent with their interpretation as a fringing ramp. By upper Grusdievbreen times, the system had evolved into a ramp-barrier complex. Broad tidal flats merged seaward into lagoons or coastal embayments. The seaward barriers for these embayments were oolitic shoals and, at times, build-ups of columnar stromatolites. Deeper off-shore environments received carbonate mud and, during storms, coarser material exported from coastal environments. Coniform stromatolites accreted in quiet subtidal environments where coarse, traction load sediments were absent. Three transgression-regression cycles can be documented within the Akademikerbreen Group, dividing the section into packages whose thickness and carbonate facies succession is broadly reminiscent of Lower Paleozoic Grand Cycles (Aitken, 1978).

The Akademikerbreen Group differs from the Lower Proterozoic carbonates discussed by Grotzinger (1989). Limestones are common; stromatolites are not lithologically dominant; grainstones abound; and tufa-like laminates are rare or absent. The carbon isotopic composition of Akademikerbreen carbonates also stands in sharp contrast to those of
most Lower Proterozoic examples; Spitsbergen carbonates and organic matter show marked enrichment in $^{13}\text{C}$ relative to both older and younger rocks (Knoll and others, 1986; compare to Schidlowksi, Hayes, and Kaplan, 1983).

An obvious question arises. In a critique of autobiography, Clive James (1985, p. 12) framed the issue squarely: "He who abandons his claim to be unique is even less bearable when he claims to be representative". The Akademikerbreen Group, after all, is only a single succession. Is it anomalous, or do Upper Proterozoic, especially 600 to 800 Ma old, carbonates in general differ from older as well as younger accumulations? In the following paragraphs, we discuss this question with regard to mineralogy, lithology, and characteristic sedimentary features.

**Mineralogy**

On the basis of field and petrographic determinations, we estimate that Akademikerbreen carbonates are about 45 percent limestone (fig. 2). (This refers to dominant mineralogy; many limestones contain a small percentage of dolomite, and dolomites frequently contain later diagenetic and/or relict calcite.) Although the relationship is statistical rather than absolute, mineralogy appears to correlate with facies. Carbonates containing evidence of inter- to supratidal deposition are commonly dolomitized, as are stromatolitic beds. Dolomitization of bioherms is commonly pervasive, but where it is not, the selective dolomitization of originally organic-rich laminae can sometimes be observed (fig. 7; see Gebelein and Hoffman, 1973). Thinly bedded, fine-grained carbonates representing off-shore sedimentation tend to be limestone, while environmentally intermediate facies exhibit variable degrees of dolomitization.

A detailed analysis of limestone/dolomite distributions in Upper Proterozoic successions is beyond the scope of this paper, but examination of two recent compendia on Proterozoic sedimentation (Campbell, 1981; Preiss, 1987) suggests that the Spitsbergen succession is fairly typical of its age. Consider two Upper Proterozoic successions from Canada, the Little Dal Group of the Mackenzie Mountains and the possibly correlative Shaler Group exposed on Victoria Island. The Little Dal Group is informative because both platform and basinal facies are represented (Aitken, 1981). Basinal nodular and rhythmitic carbonates are mostly limestone; shallow, high energy platform deposits in the lower half of the group are also predominantly calcitic, although dolomites occur. On the other hand, stromatolitic reefs, tidal flat carbonates, and carbonates associated with evaporites tend to be dolomitic (Aitken, 1981).

Deep basinal deposits are not found in the Shaler Group, but a comparable pattern of limestone/dolomite distribution is nonetheless observed (Young, 1981). Thick, microbially laminated, tidal flat carbon-
Fig. 7. Columnar stromatolites showing preferential dolomitization of microbial laminae, especially clear in microbial mat bridges between adjacent columns. In both photographs, dolomite is light colored and calcite, dark. From the Lower Dolomite Member of the Svanbergfjellet Formation (B) and Svanbergfjellet equivalents in central East Greenland (A). Units in scale in (B) = 2 cm for (A) and 1 cm for (B).
ates of the Glenelg Formation are pervasively dolomitized, as are columnar stromatolitic bioherms in the upper part of the formation and carbonates associated with evaporites higher in the section (Young, 1981). In contrast, oolitic and intraclastic grainstones, as well as fine-grained subtidal carbonates, of the Reynolds Point and Wynniatt formations contain much limestone.

The pattern extends to the Adelaide Geosyncline of South Australia, where the predominantly peritidal to (?)non-marine River Wakefield, Skilligalee, and upper Brighton carbonates are essentially dolomitic, while shelf to basinal carbonates of the Trezona, lower Brighton, Etina, and Wonoka formations are predominantly limestone (Preiss, 1987). To the north, carbonates associated with evaporites in the Gillen member of the Bitter Springs Formation are dolomitic (Lindsay, 1987). In the overlying Loves Creek Member, Southgate (1989) documented a series of shallowing upward carbonate cycles and noted (p. 336) that field relationships “suggest that dolomite occurs in areas of increased emergence and salinity. This is consistent with dolomite always comprising the thin-bedded dolostone facies at the tops of cycles.”

Limestones characterize a number of other Upper Proterozoic shelf to basinal successions, including the Biri Formation, Norway (Tucker, 1983a); the Campanuladal Formation, Greenland (Adams and Cowie, 1953); and the Kuibis and Schwarrrand subgroups of the Nama Group, Namibia (Germs, 1983).

In general, Upper Proterozoic carbonates are quite variable in composition, with limestone being a major component of many sequences. Although we stress the statistical and not absolute nature of the pattern, dolomite does appear to correlate highly with evidence for evaporite, inter- to supratidal, and playa lake sedimentation, as well as originally organic-rich sediments such as stromatolites; carbonates deposited in basinal environments below storm wave base are often limestone (but not always; for example, Herrington and Fairchild, 1989). Thus, the Akademikerbreen Group does not appear to present an unusual or distorted picture of Upper Proterozoic limestone and dolomite distributions.

In earlier Proterozoic formations, dolomite far exceeds limestone in abundance; however, when carbonate depositional environments are considered, a similar distributional pattern appears to emerge. Most Lower Proterozoic carbonates are dolomitic, but then, many are abundantly stromatolitic or tufa-like (Grotzinger, 1989). Basinal carbonates from North America (Ricketts and Donaldson, 1981; Hoffman, 1989); southern Africa (Beukes, 1983), and Australia (Plumb and others, 1981) include substantial amounts of limestone. The same is true for Middle Proterozoic carbonates. The exceptionally thick carbonate accumulations of the McArthur Basin, northern Australia, are overwhelmingly dolomitic, and they are also largely peritidal evaporitic and playa lake deposits (Muir, Lock, and von der Borsch, 1980). Shelf and basinal carbonates of the same age are predominantly calcareous. Indeed,
already in 1981, Delaney noted that a statistical pattern of basinal limestone and peritidal dolomite is common in Middle Proterozoic successions.

It is admittedly simplistic to lump all dolomites together as a group. Numerous authors (for example, papers in Zenger, Dunham, and Ethington, 1980; Morrow, 1982; Land, 1985; Hardie, 1987) have stressed that there are “dolomites and dolomites,” reflecting a variety of diagenetic processes and environments. Nonetheless, many Proterozoic dolomites are similar to one another in that they preserve depositional fabrics, implying dolomitization early in diagenesis, before deep burial (for example, Tucker, 1982, 1983b; Grotzinger and Read, 1983; Zempolich, Wilkinson, and Lohman, 1988; Kerans and Donaldson, 1989; Fairchild, Knoll, and Swett, in preparation). Later diagenetic dolomitization may not show such a strong environmental bias.

Students of Paleozoic carbonates will not find the environmental distributions of Proterozoic limestone and dolomite unusual. Indeed, they are comparable to those documented repeatedly in Cambro-Ordovician successions (for example, Cowie and Adams, 1957; Aitken, 1978; Swett, 1981; Cecile, 1982; Demico, 1985; Ineson and Peel, 1987; Ross and others, 1989; James and others, 1989) and interpreted in terms of sediment permeability and hydrological regimes (Land, 1985; Hardie, 1987). In the context of this discussion, it is noteworthy that dolomites appear to be as abundant in Cambrian carbonates as they are in the Upper Riphean and Vendian—indeed in Spitsbergen and East Greenland, Cambrian carbonates have a higher proportion of dolomite than their Proterozoic counterparts (Cowie and Adams, 1957; Swett, 1981). To the extent that early diagenetic, fabric-retentive dolomitization decreased during the Early Paleozoic, it would appear to have done so largely in the Ordovician, coincident with a major radiation of heavily skeletonized benthic invertebrates and algae.

Overall, it does appear, as earlier analyses indicated (Veizer, 1978, and references cited therein), that dolomites are more abundant in Proterozoic carbonates than they are in Paleozoic ones; however, at least from the Middle Proterozoic onward, the approximate constancy of the probability of dolomitization within a given depositional environment points to the possibility that, to a large extent, later Proterozoic/Phanerozoic trends in the abundance of dolomite reflect changes in the relative representation of different depositional environments rather than some major change in ocean chemistry. The changing ratio of limestone to dolomite observed through this interval very likely reflects the shift from a world dominated by non-skeletal carbonate deposition, in which carbonates were precipitated abundantly at the edges of the oceans, to one dominated by skeletonization, where shelves and subtidal epeiric environments became the primary loci of carbonate accumulation (Veizer and Hoefs, 1976; Veizer, 1978).
Lithology

In the Akademikerbreen Group, stromatolites, oncolites, and microbial laminates make up only about 25 percent of all carbonates (fig. 2). Oolites and pisoliths contribute another 15 percent, and the remainder—60 percent of the total—consists of micrites and arenites or rudites derived for the most part by the erosion and redeposition of micrite (fig. 2). Stromatolites are visually conspicuous in outcrop, but they are volumetrically subordinate. Grotzinger (1989) noted that apart from flake conglomerates and other intraclastic carbonates associated with stromatolites, carbonate grainstones constitute a limited proportion of many Lower Proterozoic successions. In contrast, although many Upper Proterozoic carbonates are more stromatolite-rich than those in Spitsbergen, micrite and grainstone-dominated carbonates are abundant in packages of this age (Aitken, 1981; Young, 1981; Teitz and Mountjoy, 1989). In part, this apparent dichotomy may relate to the preponderance of rimmed platforms in Lower Proterozoic examples and ramps in Upper Proterozoic packages; as Grotzinger (1989) points out, the Lower Proterozoic Monteville ramp, South Africa, contains relatively abundant grainstones. Still, the relative abundance of grainstones and bedded micrites in youngest Proterozoic carbonates appears unlikely to reflect only rim versus ramp sampling.

What is the source of the micritic sediments and clasts that dominate the Spitsbergen (and other) sections? We can identify three possibilities. One is that the muds are finely clastic carbonates originally precipitated in microbial mats; that is, the predominant Late Proterozoic carbonate factory was benthos much as it is presumed to have been during the Paleozoic Era. One way of testing this hypothesis is to examine carbonate arenites and rudites petrographically. Silicified arenites and rudites occur throughout the sequence, and these commonly preserve original clast textures. If stromatolite erosion is the major source of carbonate mud in the Akademikerbreen Group, one might expect that coarser clasts would preserve stromatolitic microstructure. Mat debris does occur in many rudites, particularly those intimately associated with stromatolites; however, many arenite and rudite clasts appear to be ripped up un laminated mud, not microbial mat fragments. If precipitation in microbial mats was indeed the predominant means of removing carbonate from solution prior to the radiation of skeletonized animals and protists, then it is necessary to hypothesize that most mats disintegrated completely to mud, which was then redeposited as sand or rudite grains that retain no textural memory of their original deposition.

An alternative hypothesis is that prior to the evolution of biomineralization, chemical precipitation directly out of the water column was common in warm, shallow platforms of moderate to high productivity. The phenomenon of whitings in the modern Bahama Banks and elsewhere has generated much interest during the last 25 yrs. Recent isotopic and SEM data indicate that while the reworking of aragonite
needles precipitated by benthic green algae may contribute to suspended load carbonates, a substantial portion of whiting carbonate is generated by the spontaneous precipitation of aragonite needles when local water masses reach a critical level of supersaturation (Shinn and others, 1989, and references cited therein). While debate over marine whittings may continue (see, for example, Morse and others, 1984; Shinn, 1985; Shinn and others, 1989), there is no question that whiting carbonates precipitate from the water column in lakes (Wiedemann and others, 1985, and references cited therein).

Perkins, Lloyd, and Kerr (1986) recently proposed that non-skeletal carbonate precipitation has been important periodically in the Phanerozoic, at times and in places where calcareous skeleton builders were not abundant. These authors termed such precipitation “default sedimentation,” suggesting that when skeleton precipitation is insufficient to keep ambient waters below critical supersaturation levels, non-skeletal sedimentation occurs “by default.” Although Perkins and his colleagues restricted their discussion to the Phanerozoic, the logic of their thesis is most compelling when applied to the Proterozoic. Certainly, the abundance of micrite and micrite-derived sediments in the Akademikerbreen Group is consistent with “default” hypothesis, as are the appreciable thicknesses of oolites. The whiting hypothesis is also consistent with Grotzinger’s (1989) suggestion that carbonate saturation in Proterozoic oceans was actually higher than it is today.

A third possible explanation for Late Proterozoic micrites is that they formed by the decomposition of calcareous skeletons of macroscopic green or red algae, much the way that some micrites form today. It is difficult to reject this hypothesis unequivocally, but there is little petrographic evidence for skeleton-forming algae until the very end of the Proterozoic (Riding and Voronova, 1984; Grant, Knoll, and Germs, in press). Thus, while we cannot dismiss it entirely, we can muster no petrographic or paleontological evidence in support of this third hypothesis and consider it the least tenable of the three possibilities.

It is impossible at this point to choose unequivocally between microbial mats and carbonate precipitates formed in the water column as the primary source of Akademikerbreen micrites, but the abundance of mud relative to stromatolites (which, clearly, were often lithified within millimeters of their mat surfaces), the abundance of oolites and pisoliths, and the undoubted occurrence today of spontaneous carbonate precipitation from critically supersaturated lake and marine waters suggest to us that whittings demand serious consideration as a principal source of carbonate sediments during the Late Proterozoic Era (see also Grotzinger, 1989; Fairchild, 1989). Much as microbial CO₂ utilization may have promoted in situ carbonate precipitation in microbial mats, so may phytoplankton blooms have driven precipitation within the water column. We do not argue that microbial mats were insignificant sites of carbonate precipitation, but rather that they may not have been the only or even the most important sites in Late Proterozoic oceans. Indeed, it is
possible that much of the fine-grained carbonate found in Upper Proterozoic stromatolitic laminae is itself suspended load carbonate trapped and bound rather than precipitated in situ by mat building communities (see, for example, Fairchild, 1989).

A question arises concerning the comparison of Upper versus earlier Proterozoic carbonates. Grotzinger (1989) emphasized the importance of tufa-like laminates and columnar to coniform stromatolites preserving precipitate textures in Early Proterozoic carbonates. Tufa-like (possibly abiogenic) carbonates are uncommon in post-Early Proterozoic marine carbonates, although they occur in rocks as young as about 1000 Ma (for example, Strother, Knoll, and Barghoorn, 1983; Grey and Thorne, 1985; Swett and Butterfield, unpublished observations). Columnar and coniform stromatolites with well preserved precipitate textures are also documented in Middle and early Late Proterozoic successions (for example, Walter, Krylov, and Muir, 1988, p. 99; Kerans and Donaldson, 1989), but their abundance appears to decrease after about 850 Ma. One might speculate that Late Riphean decreases in pCO₂ altered CaCO₃ saturation levels in the oceans, tipping the balance toward carbonate removal by whittings rather than via precipitation within benthic microbial communities. The hypothesis of decreasing pCO₂ is consistent with evidence for episodes of continental glaciation beginning approx 850 Ma ago (Kasting, 1987). Anomalously high rates of organic carbon burial provide a possible mechanism for carbon dioxide drawdown (Knoll and others, 1986).

Thus, the possibility exists that Upper Proterozoic carbonates differ not only from their Phanerozoic counterparts but also from older examples. Principal “carbonate factories” may have shifted from benthic microbial communities to fertile coastal water columns during the course of the Proterozoic Eon, although in environments such as those represented by the Little Dal and basal Draken bioherms, CaCO₃ precipitation on the surfaces of microorganisms contributed to stromatolite accretion in settings where suspended load micrite was absent or produced at infrequent intervals.

It is not unreasonable to think that ocean and atmospheric chemistry should have changed during the nearly 2000 Ma duration of the Proterozoic Eon, and the detailed comparative study of Lower, Middle, and Upper Proterozoic carbonates provides a promising means of addressing this question (Grotzinger, 1989). To the extent that changes in pCO₂ or other parameters can be documented, these may provide physico-chemical explanations for observed changes in stromatolite morphology and microstructure through time (for example, the decline of Conophyton after about 1000 Ma ago, interpreted by Walter and Heys, 1985, as a consequence of early metazoan activity; see Grotzinger, 1990).

An additional (and by no means mutually exclusive) possibility is that the Middle to Late Proterozoic radiation of seaweeds caused a significant decrease in the aerial distribution of microbial mats, espe-
cially in subtidal environments of low to moderate wave or current energy. What must have been lawns of cladophoralean green algae are preserved in siliciclastic mudstones in the upper Svanbergfjellet Formation (Butterfield, Knoll, and Swett, 1988); comparable algal growth on carbonate platforms could have contributed to the accumulation and stabilization of shallow subtidal micrites and intraclastic grainstones.

Characteristic Sedimentary Features

It is well known that stromatolites are more characteristic of Proterozoic carbonates than they are of Phanerozoic successions (Awramik, 1971; Walter and Heys, 1985), but are there other hallmarks of pre-Phanerozoic carbonate sedimentation? Several features prominent in the Akademikerbreen Group are uncommon in younger limestones and dolomites; these include originally aragonitic primary marine pisoids (overgrown ooids) up to 16 mm in diameter; pervasive micritic or microspar cements in oolites, pisolites, and other lithologies; abundant subaqueous shrinkage cracks in fine-grained carbonates generally filled by microspar cement; mottled limestone/dolomite of the type seen in the lower Grusdievbreen Formation; the preservation of fine (mm scale) lamination in shallow subtidal micrites (and siliciclastic mudstones); and strong enrichment in $^{13}$C ($\delta^{13}$C = +5 to +8 permil, with concomitantly light organic carbon).

Giant ooids (primary marine pisoids).—The large size and microspar cements of Akademikerbreen ooids have been described in detail elsewhere (Swett and Knoll, 1989). We note here only that unusually large (by Phanerozoic standards) ooids are abundant in this succession (fig. 6B and C) and have been reported from a number of other Upper Proterozoic carbonates (citations in Swett and Knoll, 1989). In the Backlundtoppen Formation, ooids smaller than 1 mm are rare, whereas grains larger than this have been reported from only a handful of post-Tommotian marine carbonates. Akademikerbreen ooids are cemented mainly by equant microspar comparable to that found in coeval shrinkage cracks: microspar cemented oolites also occur in the Cambrian (Swett, 1981), but younger marine oolites are predominantly oosparite. These relationships reinforce the suggestion that the oceans overlying Late Proterozoic platforms were highly supersaturated with respect to CaCO$_3$, resulting in growth of large ooids and rapid subsequent cementation (Swett and Knoll, 1989).

Subaqueous shrinkage cracks.—The prevalence of shrinkage cracks that close both at their tops and bottoms (thought to be the result of subaqueous dewatering; for example, Horodyski, 1982; fig. 4B and C) and the preservation of millimeter-scale laminae in fine-grained lithologies can both be related to the absence of significant bioturbation prior to the latest Proterozoic. The microspar in-filling of shrinkage cracks and other void spaces further attests to a high degree of CaCO$_3$ supersaturation in near-surface pore waters; the bending of micrite laminae
around crack cements indicates that cementation preceded sediment compaction (Fairchild and Hambrey, 1984). Subaqueous shrinkage cracks are widely distributed in fine-grained Proterozoic carbonates deposited in shallow subtidal environments (for example, Aitken, 1981; Young, 1981; Horodyski, 1983; Jackson, 1986).

*Mottled limestone/dolomite.*—We have observed mottled carbonates of the Grusdievbreen type (fig. 4A) in the Shaler Group and the Bitter Springs Formation (Southgate, personal demonstration, 1987). Once again, early cementation, perhaps related to Late Proterozoic CaCO₃ saturation levels, must have contributed significantly to the formation of this feature. Mottled limestone/dolomites occur only when nodular cementation was curtailed prior to the complete cementation of a bed. In fact, the mottling phenomenon may not be more widespread simply because pre-dolomitization cementation often proceeded to completion in shallow subtidal environments. As noted above, such early cementation may contribute significantly to the retention of calcitic mineralogy in many Upper Proterozoic micrites and grainstones. A concomitant absence of bioturbation is also required to produce this feature.

In general, the distinctive features of Upper Proterozoic carbonates appear to relate to the insignificance of animals as agents of carbonate precipitation and bioturbation, as well as decreasing pCO₂ and, possibly, the sedimentary consequences of seaweed radiation.

¹³C enrichment.—As reported by Knoll and others (1986) and corroborated in analyses of petrographic microsamples by Fairchild and Spiro (1987), many Akademikerbreen and correlative carbonates are strongly enriched in ¹³C relative to older or younger carbonates. This isotopic signature is largely independent of both facies and mineralogy and undoubtedly reflects mainly depositional rather than diagenetic conditions of origin. Within the Akademikerbreen succession, there is an excursion, seen in both carbonate and organic carbon, to lighter values (−2 to 0 permil) centered on the Lower Dolomite Member of the Svanbergfjellet Formation. This complicates interpretation in that lithologically and micropaleontologically similar carbonates in the Draken Conglomerate Formation have the ¹³C-enriched signature characteristic of the rest of the group. Other negative excursions occur in stratigraphic association with glaciogenic sediments in the overlying Polarisbreen Group and in organic carbon within two Veteranen horizons: Knoll and others (1986) have suggested that, by analogy with the Polarisbreen pattern, negative isotopic excursions in the Veteranen and Akademikerbreen groups might reflect pre-Vendian ice ages in other parts of the world. Recent stratigraphically controlled analyses of Upper Riphean and Vendian carbonates from Namibia (Kaufman and others, in press) as well as scattered data from other localities lend support to this hypothesis (Kaufman and Knoll, 1989).

Increasing data from other locations (Zempolich, Wilkinson, and Lohman 1988; Strauss and others, in press; Kaufman and others, in
press: Butterfield and others, unpublished data) support both the global nature of Late Riphean $^{13}$C enrichment in carbonates and organic carbon and the occurrence of negative excursions within this interval. This isotopic signature differentiates 850 Ma and younger Proterozoic carbonates from both earlier Proterozoic and Lower Paleozoic limestones and dolomites; its origin is not completely understood, but it must reflect, in no small part, high (but fluctuating) rates of organic carbon burial (Knoll and others, 1986).

CONCLUSIONS

Viewed from one perspective, Proterozoic and Paleozoic carbonate accumulations are remarkably similar, reflecting the overriding influence of tectonics and eustasy on deposition (Grotzinger, 1989; James and others, 1989). At another level, however, we can observe secular changes in carbonate lithologies that appear to reflect both environmental and biological evolution (Veizer, 1978; Grotzinger, 1989).

Characteristic features of carbonates changed during the Proterozoic-Phanerozoic transition, but we find little evidence to support the idea of a major Late Proterozoic “Dolomite Event” that altered the chemical composition of seawater and made skeletonization possible (Riding, 1982; Kazmierczak, Ittekkot, and Degens, 1985). Rather, just the opposite seems likely—that evident changes in carbonate lithologies reflect the increasing importance of skeletal precipitation and bioturbation (Veizer, 1978; Holland, 1984). As calcareous skeleton-forming animals, algae, and protozoans proliferated, CaCO$_3$ supersaturation levels decreased, reducing the frequency and importance of features such as large ooids, rapidly precipitated microspar cements, and, of course, whitings. Changes in the principal loci of carbonate deposition contributed to secular changes in dolomite content. We cannot exclude the possibility that changing fluxes of sea water through hydrothermal ridge systems, reduced rates of organic carbon burial, and/or changing pCO$_2$ altered sea water chemistry during this interval, but the demonstration of any such effect must take into account changing environmental distributions of carbonate deposition.

At the same time, subtle but distinctive differences between Upper and earlier Proterozoic carbonates indicate that neither organisms nor environments were static throughout the long Proterozoic Eon. Both decreases in atmospheric carbon dioxide levels and increases in the abundance, ecological distribution, and competitive abilities of macroscopic algae probably contributed to the distinctive features of Akademikerbreen and other 800 to 550 Ma old limestones and dolomites.

Much sedimentological, paleontological, and geochemical research remains to be done before the nature of Late Proterozoic carbonate sedimentation can be understood in detail; however, the limited data currently available reinforce evidence from C and Sr isotopes, glacial deposits, fossils, and widespread extensional basins that the final 200 to
300 Ma of the Proterozoic Eon was a distinctive and unusual period in Earth history (Knoll, 1990).

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