ABSTRACT OF THESIS

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Title of Thesis PETROLOGY OF THE CAMBRO-ORDOVICIAN SUCCESSION OF THE NORTH-WEST HIGHLANDS OF SCOTLAND

The Cambro-Ordovician sedimentary succession of North-west Scotland is interpreted as the deposits of a gradually transgressive sea. The basal unconformity of the Cambro-Ordovician oversteps Lewisian metamorphic and Torridonian sedimentary rocks and is perceived as the erosional front of this transgression. Revised thicknesses are given for the Eribol sandstones, the An t-Sron siltstones, and the Durness carbonates which successively overlie the unconformity. This sedimentary succession probably reflects an increasing isolation from detrital sources rather than increasing depth of water.

Various tube-like trace fossils occurring normal to the bedding in the "Pipe Rock" member of the Eribol sandstones are believed to be the result of the burrowing action of suspension-feeding organisms with the variations in "pipe" structures due to differences in sediment or water characteristics rather than differing species of animals.

An abnormally high potash content discovered by the Geological Survey in the An t-Sron siltstones and confirmed in the present study is tentatively explained as an authigenic enrichment, possibly having derived the necessary potassium through dolomitization of overlying illitic limestones.

The thick succession of carbonate rocks which comprises the bulk of the total Cambro-Ordovician sequence is interpreted as biochemical and perhaps chemical carbonate deposits of a shelf or platform environment. Oolitic and algal stromatolitic horizons interspersed throughout much of the sequence suggest that the rate of deposition approximately paralleled the rate of subsidence with the result that most of the carbonates are probably shallow water deposits.

Diagenesis has profoundly affected the characters of the succession. Pore space is virtually absent from the sandstones, having been reduced principally by pressure solution and redeposition processes. Authigenesis in the sandstones mainly involves quartz and feldspar but minor authigenic replacements by dolomite, pyrite, leucoxene, and glauconite were also noted. Diagenesis in the carbonate rocks is considerably more complex but may be summarized in outline form as follows:

1. RECRYSTALLIZATION
2. DOLOMITIZATION
3. SILICIFICATION
4. CALCITIZATION
5. DOLOMITIZATION

? 4-6. PYRITIZATION

Any one or more of these stages of diagenesis may be missed in a particular lithology but the relative order of the events is remarkably consistent. Recognition of this paragenetic order permits the solution of several perplexing problems regarding the field relationships of the limestones, dolostones, and cherts.

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PETROLOGY OF THE CAMBRO-ORDOVICIAN SUCCESSION OF THE NORTHWEST HIGHLANDS OF SCOTLAND

by

KEENE SWETT, B.S., M.S.

Thesis submitted for the DEGREE OF DOCTOR OF PHILOSOPHY
of the UNIVERSITY of EDINBURGH in the FACULTY OF SCIENCE.
1965.

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Eden Grove

A portion of this thesis concerning the petrologic interpretations of the oolitic horizons has been accepted for publication by the <u>Journal of Sedimentary Petrology</u>. With the exception of very minor alterations to that paper suggested by referees and incorporated herein, this thesis is entirely my own work.

April, 1965. Keene Swett

TABLE OF CONTENTS

Page

I INTRODUCTION

II ERIBOL SANDSTONES

III AN T-SRON FORMATION

IV DURNESS CARBONATES

V SUMMARY

I. INTRODUCTION

A. GENERAL STATEMENT

The topic of this thesis, the Petrology of the Cambro-Ordovician Succession of the Northwest Highlands of Scotland, was suggested to the author by Dr. A. Hallam during the spring of 1962. Preliminary field reconnaissance during late summer of the same year confirmed the suitability of the project and brought into focus a generalized scheme of study. The plan was basically to perform a detailed examination, description, and petrologic study of as complete a vertical sequence as could be established in the Cambro-Ordovician units and aimed at a reconstruction of the depositional and post-depositional history of the rocks. Original plans for comparative studies of several vertical sections were abandoned because of the abundance of significant material in the Durness area and to avoid the pitfall of a study involving so many samples that detailed petrographic examinations become impossible.

Cursory examinations were made of stratigraphic sections other than at Durness and some attention has been given to these but the bulk of the petrographic work has been restricted to the composite stratigraphic sequence of the Durness - Eribol area. Remapping of the areal geology with the benefit of air photographs and further paleontological investigations would undoubtedly offer much valuable information concerning the Cambro-Ordovician succession. However further mapping or paleontological studies remains outside the scope of the present thesis except where these subjects have influenced the location of the measured vertical succession or have affected the petrology of the rocks.

B. HISTORICAL BACKGROUND

In 1774, T. Pennant first recorded the presence of limestone and marble in Ross-shire and Sutherland. In the nearly two centuries since that time many geologists have given consideration to various aspects of the sedimentary carbonate and quartzitic rocks of the northwest Highlands with most of their efforts being directed toward structural and stratigraphic problems. These rocks, now generally accepted as Cambro-Ordovician in age, have commanded at least brief attention from several pioneers in the field of geology. Among the notables who have contributed to the literature are Bonney, Geikie, Lapworth, Murchison, Sedgwick, Nicol and MacCulloch.

In the Geological Survey Memoir (B.N. Peach et. al., 1907), J. Horne gives an extremely interesting and full review of geological investigations in this area prior to 1900. No additions or improvements to this account can be offered here and accordingly, the interested reader is directed to that Memoir.

In addition to its review of early investigations, the

Survey Memoir (1907) also affords the most comprehensive work yet published on the Cambro-Ordovician rocks of the Highlands. A general account of the stratigraphy, structures, paleontology, and petrography is given and despite a number of inaccuracies, it remains a highly significant contribution to the knowledge of the succession. Little further work has been done on the Cambro-Ordovician sequence with the exception of various proposals concerning the stratigraphic correlations between the Scottish succession and rocks of probable similar age in Greenland (Poulsen, 1951, and Cowie, 1960), Spitzbergen (Gobbet & Wilson, 1960, and Wilson, 1961), and North America (Grabau, 1916). These contributions will be considered further in conjunction with discussions of the stratigraphy.

C.B. Crampton (1958) has discussed the fabrics and mineral orientation of dolomite and calcite marbles relative to regional tectonics in the area between Loch Broom and Loch Glencoul. Phemister, in the British Regional Geology Handbook of the Northern Highlands (1960), has reviewed the Cambrian and Ordovician systems but has offered few, if any, new thoughts on the subject.

Geological mapping of the area commenced with MacCulloch in 1840 and apparently culminated with the efforts of B.N. Peach et. al. in 1889 which led ultimately to the production of Sheet 114 of the Geological Survey of Scotland.

C. STRATIGRAPHY

Quartzitic sandstones, dolomitic siltstones, and dolomitic and calcitic carbonate rocks of Cambrian and Ordovician age crop out in normal sequence along a narrow belt approximately 85 miles in length reaching from Loch Kishorn in the south (N57°20', W5°47') to Loch Eribol in the north (N58°23', W4°40' - 4°50') (Fig. I). These units are also found in faulted outliers at Durness, Achiltibuie, and on the Isle of Skye near Broadford and at Ord.

The relative rarity of significant fossils in the sandstones and carbonates and their absence from the overlying and underlying units caused considerable early controversy regarding their age. Present knowledge permits the following generalized account of the age relationships. To the west and presumably widespread at depth is the oldest unit, the Lewisian metamorphic complex. The Lewisian is mainly composed of orthogneiss probably derived from a granite - gabbro - peridotite complex (Phemister, 1960) but the system also includes some schists recognized as altered sediments.

Also to the west and unconformably overlying the Lewisian are several thousand feet of Torridonian arkoses. The Eribol sandstones (see p. 8 & 9) unconformably overstep both the Lewisian and Torridonian. The age of the Eribol sandstones is accepted as Lower Cambrian due to the discovery of Olenellus by C. Lapworth in 1885 (publ. 1888) in the dolomitic siltstones of the An t-Sron formation which immediately and

FIGURE I

Outline map of northern Scotland showing outcrop area of Cambro-Ordovician rocks (stippled). The Cambro-Ordovician units overlie Lewisian gneiss and Torridonian arkoses to the west and dip generally E.S.E. beneath overthrust Moine schists.

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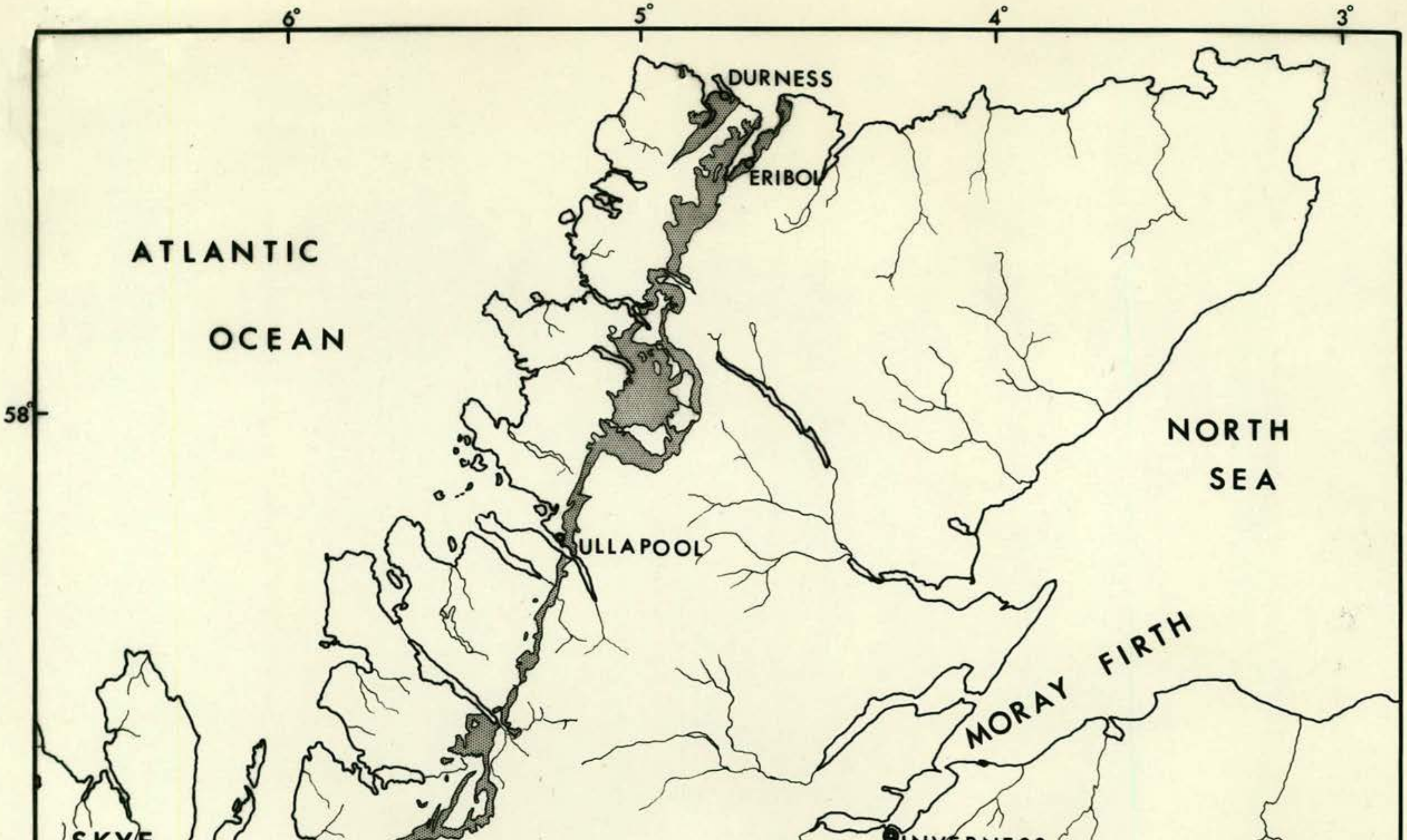
TUB SIZED AIR DRIED


6°
5°
4°
3°
58°
57°
DURNESS
ERIBOL
ATLANTIC
OCEAN
NORTH
SEA
ULLAPOOL
MORAY FIRTH
SKYE
INVERNESS
RHUM
0
10
20
MILES

conformably overlies the Eribol sandstones. The dolomitic siltstones of the An t-Sron formation grade conformably upward into somewhat in excess of 4000 feet of carbonate rocks (maximum exposed thickness) which are Ordovician in age, at least in the upper portion.

The age of the carbonate sequence is largely based on a molluscan assemblage and a doubtful asaphoid trilobite (Phemister, 1960). Poulsen (1951) prefers to assign to these units a Lower Ordovician age and cites the remarkable similarity between the faunal assemblages of the East Greenland Beekmantown, the "Durness Series" and the Beekmantown of the St. Lawrence geosyncline. He lists the following species common to East Greenland and Northwest Scotland (those marked with an asterisk are also known from the Beekmantown of the St. Lawrence geosyncline).

*_Archaioscyphia minganensis_ (Bill.)
Gyromena sp.
Helocotoma n. sp.
*_Hormotoma artemesia_ (Bill.)
*_Maclurites oceanus_ (Bill.)
Maclurites peachi (Salter)
Maclurites sp. (operculum)
Ophileta calcifera (Bill.)
*_Solenospira prisca_ (Bill.)
*_Trochonema calphurnium_ (Bill.)
Proterocameroceral n. sp.

**Protocycloceras lamarcki* (Bill.)

**Protocycloceras mendax* (Salter)

**Euchasma blumenbachi* (Bill.) var. II

On the basis of both the faunal and lithologic characters, Poulsen draws the correlations as shown in Page 6a. and supports Grabau's view (1932) that the Scottish "Durness Series" and the Greenland and North American Beekmantown units were deposited in the same geosyncline. The fauna is also closely related to the Pennsylvanian Beekmantown (Phemister, 1960) which is accepted by most American geologists as Lower Ordovician (Canadian).

The Durness carbonates are everywhere truncated on their upper surface by the Moine Thrust plane or other faulted Caledonian structures. The obscurity of the Moine thrust plane fostered considerable early controversy regarding the age relationships of the units of the northwest Highlands. The age relationships between the Moine Schists and the Torridonian arkoses is still a questionable issue for both can be demonstrated to lie unconformably on the Lewisian gneiss but the direct relationships of the Moine and Torridonian units are never seen. Current thought on this subject considers the Moine and Torridonian rocks possibly equivalent in age with the Moine schists representing a metamorphosed basinal facies of the Torridonian arkosic sediments (M.R.W. Johnson, 1964, personal communication).

A misinterpretation which caused prolonged controversy was

FIGURE II

CORRELATION OF THE CAMBRO-ORDOVICIAN OF NORTHWEST SCOTLAND, GREENLAND, AND ENGLAND

(after Poulsen, 1951)

NORTHWEST SCOTLAND	EAST GREENLAND	ENGLISH ?
Durine Gp.	Narwhale Sound fm.	Llanvirn ?
Croisphuill Gp. Balnakiel Gp. Sangamore Gp. Sailmhor Gp.	Cape Weber fm.	Arenig
HIATUS	HIATUS	M. and U. Tremadoc
	Cass Fjord fm.	Lower Tremadoc
Eilean Dubh Gp.	Dolomite Point Dolomite	Age not determined
Grudie Gp.	Hyolithes Creek Dolomite	
Serpulite Grit Fucoid Beds	Ella Island fm.	Lower Cambrian
Pipe Rock	Pipe-rock Bastion fm.	

the early correlation of the Torridonian arkoses with the Devonian Old Red Sandstone arkoses. The discovery of the Olenellus horizon (Lapworth, 1888) in the strata overlying the Torridonian sandstones ended this dispute.

Within the Cambro-Ordovician sequence, exposures of the arenaceous units are, quite naturally, much better than those of the carbonate rocks owing not only to their greater resistance to weathering and erosion but also to their greater competence in tectonic environments. A total vertical section of the sandstones is continuously exposed in several localities. The argillaceous and carbonate rocks which, unfortunately, include the greater part of the Cambro-Ordovician succession are, at best, poorly exposed with but one exception of a partial section of roughly 2000 feet of continuous exposure along the coast at Balnakiel Bay.

The commonly accepted stratigraphic terminology for the Cambro-Ordovician sequence as presented by Peach et. al. (1907) is summarized in the following outline.

Calcareous Series (youngest)
  VII Durine Group
  VI Croisphuill Group
  V Balnakiel Group
  IV Sangamore Group
  III Sailmhor Group
  II Eilean Dubh Group
  I Grudaidh Group

- Middle Series (partly calcareous and partly arenaceous)
  - Upper Zone (Serpulite Grit)
  - Lower Zone (Fucoid Beds)
- Arenaceous (quartzite) Series
  - Upper Zone (Pipe Rock)
    - Subzone V Ordinary pipes, somewhat large.
    - Subzone IV Exceedingly numerous ordinary pipes.
    - Subzone III Trumpet pipes.
    - Subzone II Ordinary pipes.
    - Subzone I Small pipes.
  - Lower Zone (false-bedded quartzites)

Torridonian Arkoses	(Precambrian)
Lewisian Gneiss	(Precambrian)

The present writer would favour a broader grouping of the stratigraphic units but revision of established nomenclature in a petrologic study which lacks detailed consideration of paleontological factors or areal distribution is clearly inappropriate. Certain minor revisions and redefinitions of the stratigraphic units are tentatively offered and a more rigorous adherence to recent views regarding the separation of rock and time units and regarding "group" terms seems advisable. The following stratigraphic divisions, retaining former terms (where practical) are suggested.

Durness Carbonates (formation)
    Durine member
    Croisphuill member
    Balnakiel member
    Sangamore member
    Sailmhor member
    Eilean Dubh member
    Grudaidh member
An t-Sron formation (= Middle Series or Passage Series)
    "Serpulite Grit" member
    "Fucoid Beds" (Planolites) member
Eribol sandstone (formation)
    "Pipe Rock" member
    Lower member

Torridonian Group (Precambrian)

A comparison with the nomenclature of Peach et. al. (1907) will show that the major alteration proposed is the substitution of "formations" for "Series" and "members" for "Groups" or "Zones". Type locality names are given to the formations but with the exception of the An t-Sron formation, all of the names have been used earlier. In as much as the bulk of the Durness formation is dolostone or dolomitic, and since the Eribol sandstones include no "quartzites", the substitution of the terms "carbonates" and "sandstones" seems easily justifiable. The only remaining change proposed is the elimination of the "subzones" within the "Pipe Rock" member. The subzones of Peach et. al. (1907) are neither laterally continuous nor everywhere in the same order and therefore are misleading as formal stratigraphic divisions.

A second difference of opinion between earlier stratigraphers and the present writer concerns the stratigraphic thickness of various units. Table I presents a comparison of the stratigraphic intervals recorded in the Durness - Eribol area by Peach et. al. (1907), Phemister (1960), and the present writer. Other workers have remarked on the low stratigraphic thickness figures assigned to the various "Groups" by Peach et. al. (1907) (Gobbet and Wilson, 1960) but no details of this discrepency have yet been published.

D. METHODS AND PROCEDURES EMPLOYED

In any petrological study which is to involve large areal extents of rocks, it is necessary to devise a plan for taking a representative sample of the rocks. For a sedimentary sequence, essentially two possibilities exist. One choice is a study based on a vertical section. The second method, quite naturally, is a lateral facies oriented study. Confronted with an outcrop pattern characterized by a relatively minor amount of stratigraphy extending over a considerable lateral distance as with the Cambro-Ordovician succession of Scotland (4000 feet X 100 miles) it is natural to assume that a facies study would be

TABLE I

VARIATIONS IN RECORDED STRATIGRAPHIC INTERVALS

Stratigraphic Unit	Peach et. al. (1907)	Phemister (1960)	This thesis
DURINE GP.		180'	610'
CROISPHUILL GP.		325'	475'
BALNAKIEL GP.		215'	1005'
SANGAMORE GP.	1500'	200'	607'
SAILMHOR GP.		315'	553'
EILEAN DUBH GP.		280'	656'
GRUDAIDH GP.		75'	196'
SERPULITE GRIT	30'	30'	86'
FUCOID BEDS	50'	50'	
PIPE ROCK	350'	280'	319'
LOWER QUARTZITES	200'	230'	255'
TOTAL	2100'	2180'	4762'

the most profitable. The ultimate decision, however, must rely on the availability of vertical and lateral variations. In the Cambro-Ordovician succession, the lack of lateral variation and the difficulties of correlation caused by relatively poor exposures and complex structural complications make a lateral facies study highly impractical. Thus the project becomes self-determined by its own characters as a petrological study of mainly a single vertical stratigraphic section (see appendix).

The detailed measurements and descriptions of the stratigraphy as presented in the appendix to this thesis were accomplished with the aid of a Jacob's staff. Since this technique is seldom employed by geologists in Britain, it is felt worthwhile to describe and evaluate the procedure both for its intrinsic value and to alleviate misgivings that the method of measurement might be responsible for the discrepency between the stratigraphic thicknesses recorded here and those in earlier reports.

The Jacob's staff is constructed of wood ¾" X 1½" X 4' 10-5/8" so that with a Brunton compass placed on top of the squared end of the upright staff, the line of sight through the compass is exactly 5 feet above the ground. By setting the Brunton's clinometer reading equal to the dip of the strata and inclining the staff in the direction of the dip (normal to strike) until the levelling bubble is centred, the line of sight through the compass intersects the ground surface exactly five feet stratigraphically above the base of the staff. The staff is painted alternate colours at 1 foot intervals for measuring details of a magnitude less than 5 feet (see Plate 1-C).

The accuracy of this method, as well as most other methods, fails when the dip of the strata is roughly parallel to the topography. Fortunately, in the sequence considered here, this parallelism was very seldom encountered. The reliability of the results obtained with a Jacob's staff compare very favourably with those obtained with either a steel tape or a plane table and the Jacob's staff has the advantage of giving direct stratigraphic thicknesses (without calculations) and the advantage that it may be conveniently employed by a lone worker.

Samples for the petrographic studies were taken contemporaneously with the measurement and description of the stratigraphic succession at intervals determined by the degree of lithologic variation. In addition to the samples taken from each unit as representative of the general lithology, an effort was also made to sample local diagenetic phenomena, local lithologic variations and the contact features between adjacent lithologies.

Each sample was labelled with a letter representing the portion of the composite sequence ("Q" = Lower quartzose sandstones, "P" = Pipe Rock, "I" = An t-Sron formation, and "D" = Durness carbonates). In addition to the letter designation, a number was assigned to each sample corresponding to the stratigraphic distance between the sample and the base of the unit designated by the letter. These identification labels are retained on all samples throughout the study and afford an immediate stratigraphic reference from any specimen label whether it be a thin section, hand sample, etched surface, insoluble residue, or chemical analysis. For example, a sample label "D-1465" indicates that the specimen was taken 1465' stratigraphically above the base of the Durness carbonate sequence.

The samples collected were subjected to various laboratory study procedures depending upon their lithologies. Thin sections were prepared of nearly all specimens and the carbonate thin sections were all stained with Alizarin red "S" and potassium ferricyanide to make easier the distinction between calcite, dolomite, and ferroandolomite (Evamy, 1963).

Other laboratory procedures involved the study of polished and varnished surfaces, etched surfaces, X-ray identifications of carbonate and clay minerals, insoluble residues from the carbonates and chemical analyses of selected samples. Some of these procedures proved most unprofitable in terms of the ultimate aim of the study - to interpret the depositional and post-depositional history of the rocks. Unless the discovery of the limitations of these various procedures can be considered of value in such a study, considerable effort and time was wastefully expended on them.

II. ERIBOL SANDSTONES

A. GENERAL STATEMENT

The sandstones that comprise the lower few hundred feet of the Cambro-Ordovician sequence have been variously termed "Quartz Rocks", "Eribol Quartzites", "Ord Quartzites", or, more generally, the "Cambrian Quartzites". The label "quartzites" is misleading in two respects: first, the term quartzite is now generally reserved for metamorphic rocks, and second, all of the rocks involved are not, mineralogically, even orthoquartzites. With respect to the first objection, the matter of precedence must be weighed against more recent refinements in sandstone nomenclature. However, the resolution of this argument to either the term quartzite or orthoquartzite fails in that both of these terms present a slightly inaccurate concept of the lithology. Many of the sandstones are compositionally arkosic or sub-arkosic (after Pettijohn, 1957).

The term "quartzose sandstones" would offer a more accurate picture of the lithology and avoids the metamorphic and compositional connotations of quartzites. However, a "type" locality name seems preferable to a purely descriptive stratigraphic name. Loch Eribol is a logical choice for the type area since the largest areal exposure and the complete vertical sequence are well demonstrated there.

The Eribol sandstones are one of the more spectacular rock units of the northwest Highlands due to the whiteness of their appearance, described by Dr. J. MacCulloch in 1814,

"....... as if for ever retaining the snows of winter".

These quartzose sandstones are commonly and easily divided into two divisions: a lower, current-bedded or false-bedded member and an upper member known as the "Pipe Rock" because of its characteristic cylindrical rod-like structures normal to the bedding, believed to have resulted from the burrowing action of some organism. The term "pipe" is derived from Nicol's (1857) comparison of the structures to pipe stems.

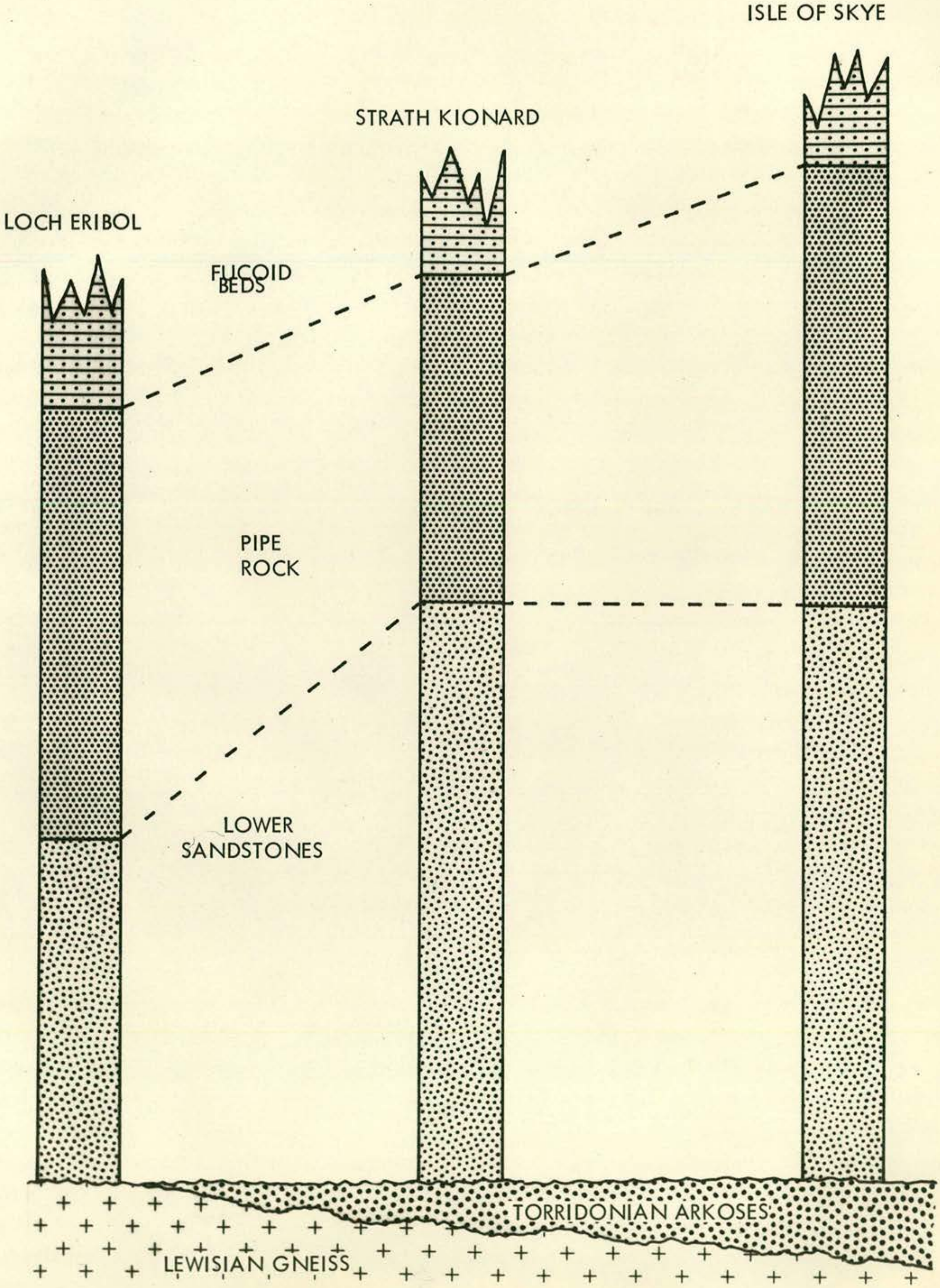
The base of the Eribol sandstones is marked by a pronounced planar unconformity with the Eribol sandstones overstepping the Lewisian gneiss and the gently folded Torridonian arkoses. The planar aspect of this unconformity was interpreted by J. Horne (Peach et. al., 1907, p. 369) to indicate a very long interval of erosion during which the ancient land surface was ...."planed down to the sea-level or perhaps just beneath it". Recent investigations have established that marine planation of platforms greater than one-third of a mile in width must result from a rather continuous submergence (i.e. a transgressive sea) (Bradley, 1958). The total sequence of the Cambro-Ordovician conforms to a concept of very gradual transgressive marine processes by the stratigraphic superposition above the planar unconformity of the sandstones, then siltstones and eventually by a thick carbonate sequence. This does not necessarily imply a continuous deepening of the water. Oolitic horizons and algal structures found throughout the carbonates indicate water conditions probably no deeper than for the sandstone deposition and it seems probable indeed that the stratigraphic variations in lithology express an ever increasing distance from detrital sources rather than an increasing depth of water. The concept of a gradual marine transgression also offers the most satisfactory explanation for the remarkable uniformity in thickness of the sandstones between Loch Eribol and Ord on the Isle of Skye. Three rather equally spaced sections were measured during the study (Figure II) and these show only a slight thickening to the south. The variation of less than 200 feet over the 100 mile length of outcrop suggests that either the sands are a very tabular deposit (as might be expected with gradual transgression) or that the present outcrop exposures trend nearly perfectly parallel to the strand line of a wedge-shaped deposit. A sheet-like geometry of the sandstones seems more probable.

B. PETROGRAPHY

The compositional range of the sandstones encompasses three of the categories within the classification scheme of Pettijohn (1957, p. 291.) The categories within which the Eribol sandstones fall are, in order of decreasing abundance,

FIGURE II

Vertical Scale 1" = 100'



orthoquartzite, subarkose, and arkose. It should perhaps be pointed out that the modal analyses of the rocks given in Table II are not necessarily a true assessment of the quantitative distribution of rock types for the author must admit to a sampling bias favouring rocks which appeared petrographically interesting. This bias has probably magnified the proportion of the feldspathic rocks relative to the orthoquartzites.

The vertical variation shown in the modal analyses does reflect a consistent and actual increase in the compositional maturity (based on percent feldspar) of the sediment as one passes upward through the succession. Induration and pressure solution processes have so profoundly altered the textural aspects of the sandstones that an evaluation of textural maturity (Folk, 1951) is impossible.

Grain size within the sandstones varies from the rare instance of pebbly conglomerates in the Lower member to the equally rare instance of siltstones. Most of the rocks fall in the medium range of the Wentworth scale (0.25 mm - 0.5 mm) but the degree of sorting is not as high as the mineralogy might suggest and isolated pebble-size grains are commonplace. Disaggregation of the sandstones is impossible due to their high degree of induration. This renders size and sorting studies possible only from thin section and thus impractical relative to the knowledge offered. It is, however, possible to make the general observations from the thin sections,

TABLE II

MODAL ANALYSES - SANDSTONES

(Minimum of 1000 points each - calculated to 100%)

Thin Section*	Qtz.	Micr.	Orth.	Plag.	Chl. Ser. Mus.	Ilm.	Mag.	Hem.	Leu.	Zir.
Q-2	73.1	15.1	7.0	3.5	1.3	-	-	-	-	-
Q-6	90.4	0.8	8.2	0.3	0.3	-	-	-	-	-
Q-25	88.5	3.1	7.1	1.0	0.3	-	-	-	-	-
Q-50	88.9	6.6	3.7	0.6	0.2	-	-	-	-	-
Q-75	83.5	16.5				-	-	-	-	-
Q-86	36.1	1.5	8.6	3.4	46.7	-	0.8	-	2.8	-
Q-90	88.1	10.5			-	-	1.4	-	-	-
Q-150	77.1	9.2	9.4	2.1	1.3	-	-	-	0.8	0.1
Q-205	90.8	0.8	3.1	0.4	3.0	-	0.5	0.3	0.6	0.5
Q-245	95.6	0.8	2.2	0.4	1.0	-	-	-	-	-
P-30	92.0	-	-	-	7.7	0.3	-	-	-	-
P-95	95.0	-	-	-	3.6	-	1.4		Tr	-
P-105	93.3	-	-	-	5.1	0.4	-	1.2	-	-
P-135	85.6	1.1	5.2	-	7.8	-	-	-	-	0.3
P-160	92.9	4.3		-	2.7	Tr	Tr	Tr	-	Tr
P-206	92.3	7.7		-	-	-	-	-	-	-
P-290	95.7	4.3		-	-	-	-	-	-	-
P-311	95.9	4.1		-	-	-	-	-	-	-

* Q = Lower member; P = "Pipe Rock" member.

Numbers indicate stratigraphic interval between sample and base of the member designated by the letter.

without detailed measurements, that grain size decreases slightly and the degree of sorting and mineralogical maturity generally increases as the succession is ascended.

C. SEDIMENTARY STRUCTURES

A conspicuous feature of the Lower member of the Eribol sandstones is their cross-bedded or current-bedded character.

> "Cross-bedding is a structure confined to a single sedimentation unit (Otto, 1938, p. 575.) consisting of internal bedding, called foreset bedding, inclined to the principal surface of accumulation" (Potter and Pettijohn, 1963).

The foreset beds of the Lower member of the Eribol sandstones are the tabular variety (after Potter and Pettijohn, 1963) or planar variety (after J.R.L. Allen, 1963) (Plate 1-A-B). The cross-bedded units range in thickness from 3 inches to 4 feet with the average unit approximately 14 inches thick.

The cross-bedding conforms most closely to the "omicron" cross-stratification of Allen (1963) which he interprets as the probable result of migrating trains of large-scale asymmetrical ripple marks (= megaripples of Van Straaten, 1953). Further applying the descriptive terminology of Allen, the "cosets" may be designated as "grouped", "large scale", "non-erosional", "planar", "concordant" and "homogeneous".

The migration of sand waves does not, in itself, predestine the deposition of cross-bedded units unless accompanied by a sand supply exceeding the transportation capacity of the currents. The erosional surfaces between the cross-bedded units impose the additional restriction that the ripple should gain from the bed-load of the currents a volume of sediment less than the volume of the ripple body. In this way, the ripple will undergo erosion on the stoss side to produce the erosional surface between two units (Allen, 1963).

Observation of cross-bedding in both modern and ancient sediments ...."shows that the maximum dip of a foreset

> bed is parallel or subparallel to the direction of the average local current vector - recognizing, of course, that minor secondary or eddy currents may display marked deviation from this trend" (Potter &

Pettijohn, 1963). If the genetic interpretation of migrating, large-scale ripple trains is accepted for the Eribol sandstones, the validity of the assumed coincidence between maximum dip direction and local current direction is enhanced. The possibility of backset beds in migrating ripples would seem possible only by the up-current regression of antidunes. Upcurrent migration of antidunes seems unlikely within the velocity spectra of most marine shelf environments. Deviations in dip directions due to secondary or eddy currents would be far less common in migrating marine ripples than in fluvial environments.

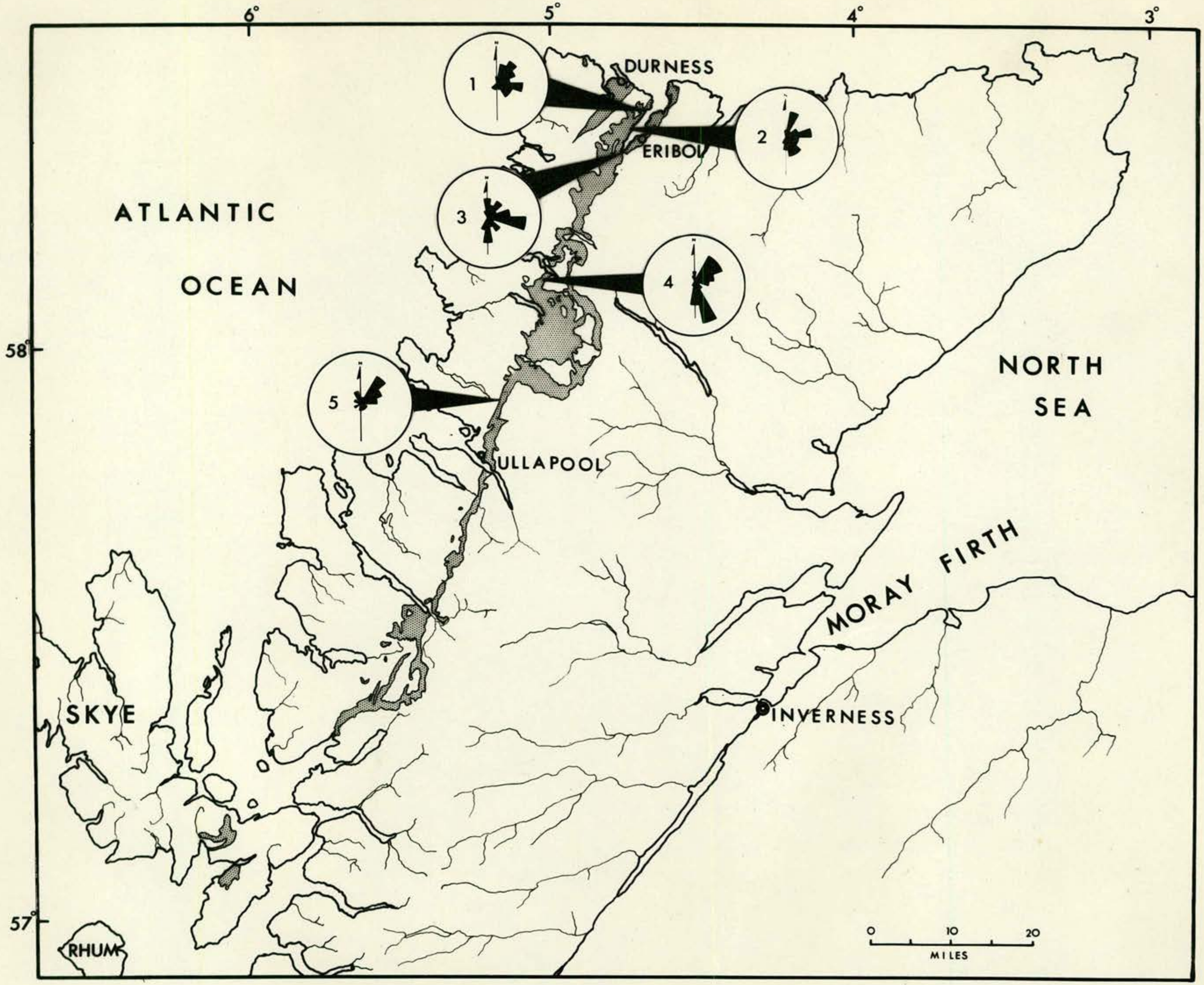
The results of the measurements of dip directions of the cross-bedding are shown in Fig. 3. Measurements of the dip of the cross-bedding were mostly determined on exposed dip surfaces. In each of the rose-diagram plots (Fig. 3) the regional dip, as determined from the erosional surfaces and the non-cross-bedded units, has been eliminated by rotation to the horizontal by manipulation on the Schmidt net. Each plot represents fifty or more dip direction measurements. At the outcrop, an attempt was made to sample both the stratigraphic and facies variations of the cross-bedding. No consistent trends were detected in either the stratigraphic or facies variations suggesting that, while exhibiting "unit" directional variation, the agents and conditions of deposition changed very little.

Perhaps the most striking feature of the current patterns indicated by the rose diagrams of the maximum dip directions is the almost total restriction of the depositional current directions to the north, east, and south. The near total lack of westerly dip directions combined with the consistent western direction of overstep of the basal unconformity strongly supports a general provenance direction for the detritus to the west. The depositional currents probably included both wave-induced and tidal currents. If the interpretation of the general north-south strand line bordered by land to the west is accepted, the on-shore tidal currents must have been either erosional or too weak to cause the migration of the megaripples.

FIGURE III

Rose diagram plots of current-bedding measurements. Each rose-plot represents fifty or more measurements of maximum dip directions of current bedding. Graphical extension of 20° segments is proportional in direction and length to the frequency of maximum dip direction measurements. In each plot, the regional dip has been eliminated by manipulation on a Schmidt stereographic net.

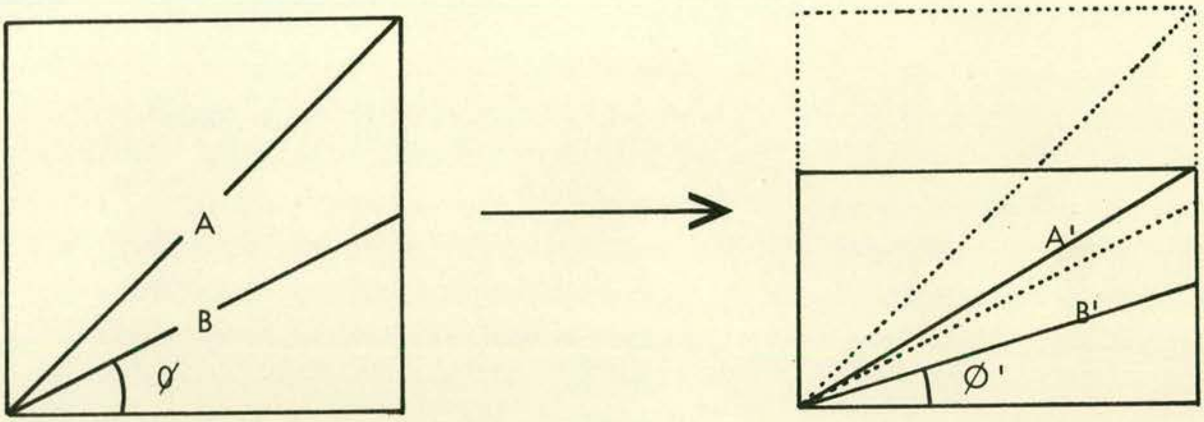
LOCATION 1	N.W. Loch Eribol. (N.C. 456643) (Regional dip of 4° to N90°E)
LOCATION 2	W. Loch Eribol. (N.C. 413615) (Regional dip of 9° to S38°E)
LOCATION 3	S.W. Loch Eribol. (N.C. 393554) (Regional dip of 14° to S45°E)
LOCATION 4	Old Drumrunie Lodge (N.C. 163057) (Regional dip of 12° to S75°E)
LOCATION 5	Cnoc Coir a' Bhaic (N.C. 240297) (Regional dip of 10° to S65°E)


6°
5°
4°
3°
58°
57°
DURNESS
ERIBOL
1
2
3
4
5
N
ATLANTIC
OCEAN
NORTH
SEA
ULLAPOOL
MORAY FIRTH
INVERNESS
SKYE
RHUM
0
10
20
MILES

Compaction and cementation have reduced the porosity of the sandstones to negligible proportions. The reduction in pore space has been accomplished largely by the mechanism of vertical compression and compaction. If the rocks originally possessed roughly 40 percent porosity as is common in recent sands (Lee, 1919, p. 121.), the deposit would have been approximately 40 percent thicker than the present rock units. To the knowledge of this writer, the effects of volume reduction by vertical compaction upon sedimentary structures (e.g. current-bedding) have received little if any attention. This subject would seem worthy of at least brief consideration although it becomes clear that this does not affect the results of current direction studies. The angle of dip of the cross-stratified units of the Lower member of the Eribol sandstones must have been reduced during compaction of the sediment. The magnitude of the angular reduction of the dip value depends, of course, upon their original deviation from the horizontal. Strictly vertical compaction will have no effect on either horizontal or vertical structures. Figure IV demonstrates the hypothetical effect of a 40 percent vertical compaction on intermediate dip angles. For this compaction factor, there would be a maximum angular reduction of roughly 14° for compaction acting upon an original dip of 45° on the foreset beds. The reduced dip angle can only approach zero, never reach it, and thus, the dip direction can

FIGURE IV

Hypothetical effect of a 40% compaction on dip angles of current-bedding.



$$\tan \text{Ø}' = 0.6 \tan \text{Ø}$$

or

$\tan \text{Ø}'$ (depositional angle) = $1.66 \times \tan \text{Ø}''$ (fossil angle)

(assuming a 40% compaction of the rock)

never be altered by vertical compaction of horizontal units. The effect of compaction could be of considerable significance in studies where the angle of repose is important to the interpretations.

The "Pipe Rock" member is easily distinguished from the Lower member in the field by its lack of current-bedding and by the common occurrence of cylindrical rod-like structures oriented normal to the bedding (Plate 1-C). J. Nicol was responsible for first applying the term "pipes" to these structures but Dr. J. MacCulloch gives an earlier, if not the first, account of these features from the "Quartz Rocks" of the Assynt district. (MacCulloch, 1814, pp. 461-462.).

> "An occurrence of equal importance and greater singularity is that of imbedded cynlindrical bodies which they (the sandstones) exhibit". .... "If I might venture on a comparison as vulgar as it is explanatory of this appearance, I would compare it to the two sections of a piece of larded meat" ... ... "it probably arises from the remains of some animal, a Sabella or other marine worm."

The "Pipe Rock" sandstones have been divided into five subzones by Peach et. al. (1907, p. 372-373) based on different forms of pipes which they ascribe to the burrowing actions of differing annelid species. Their five zones are:

I. "Small pipes" (lowest)
II. "Ordinary pipes"
III. "Trumpet pipes"
IV. "Exceedingly numerous ordinary pipes"
V. "Somewhat larger than the ordinary pipes"

The descriptive material of Peach et. al. is very sparse and no mention is made of the facies variations in the distribution of these structures. It is the opinion of this writer that the various forms of pipe structures could all be the result of the burrowing actions of a single species of organism (possibly annelids) rather than the result of several species. The variability of form in the pipes seems equally well explained as the response of a single species to differences in environmental conditions such as sediment packing or perhaps the availability of food in the water above the sediment surface.

The individual pipes are essentially normal to the bedding, parallel to each other and all possess a cylindrical or rod-like character ranging from 3 mm to 15 mm in diameter and attaining lengths in excess of a meter. Length measurements are very difficult to obtain because fractures, and hence exposures, seldom, if ever, exactly parallel the vertical pipe structures for their full length. In those horizons where the pipes are closely packed, it is also often difficult to determine if only one pipe is being traced.

The pipe structures may be central to a larger deformation phenomenon associated with the cylindrical core and show a concentric downward deflection of the adjacent bedding planes toward the "core" thus producing the "trumpet pipes" (Plate 1-D) or the pipe structure may simply trucnate the bedding laminae sharply and without perceptible deformation. The pipes may be so widely spaced that they are difficult to find or may be so closely packed as to actually overlap each other (Plate 1-E).

As indicated in Table II, the upper sandstones or the "Pipe Rock" member lithologies are more quartzose than the Lower member. Generally, the pipes appearance is a weathered feature (Plate 1-F) reflecting a very slight mineralogical difference within the rod-like cores of a slightly lower sericite content. This mineralogical difference is so slight that it is often impossible to detect petrographically yet differential weathering has revealed their presence. In certain instances where the host sandstones are generally less sericitic, the pipes may be extremely difficult to detect even on weathered surfaces and it is often difficult to decide if pipes are present or if one is looking at a surface which has been merely water streaked. On freshly broken surfaces the structures are very poorly visible and polished or varnished surfaces are only sometimes helpful.

The "trumpet pipes" are the result of a downward deflection of the sediment laminae adjacent to the central pipe core (Plate 1-D). In plan view, this appears as concentric

PLATE 1

A. Current-bedding in the Lower sandstone member of the Eribol sandstones at coastal exposures on the west shore of Loch Eribol.

B. Current-bedding in the Lower sandstone member of the Eribol sandstones showing some of the smaller cross-laminations within the larger current-bedded units.

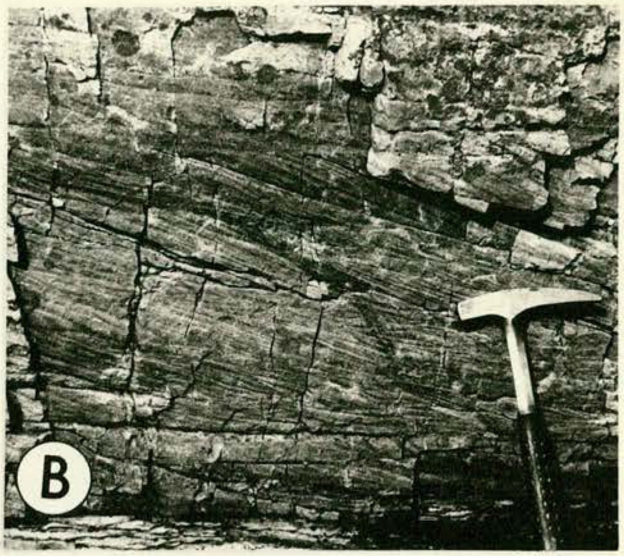
C. Weathered exposure of the "ordinary pipes" within the Pipe Rock member of the Eribol sandstones. Painted intervals on the Jacob's staff are 1 foot each.

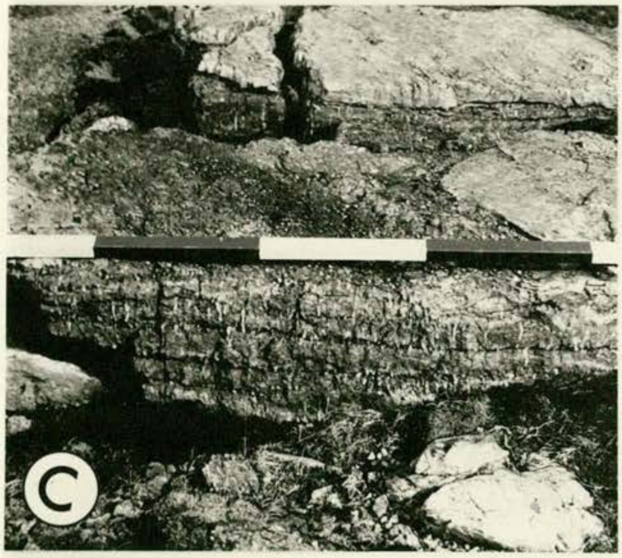
D. Cross-sectional exposure of the "trumpet pipes" within the Pipe Rock member of the Eribol sandstones. Note the downward deflection of the sediment laminae adjacent to the "pipe" cores.

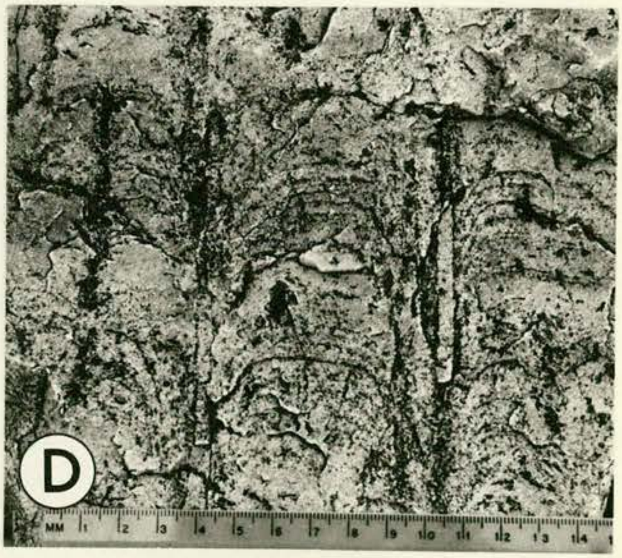
E. Bedding-plane exposure of close-packed "ordinary pipes" showing their occasional overlap.

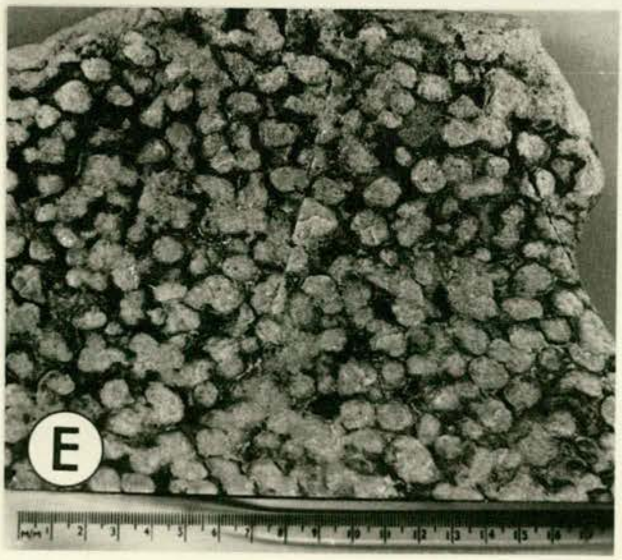
F. Bedding-plane exposure of "ordinary pipes" showing their projection above the general surface caused by differential weathering of their slightly greater quartz/clay content than the inter-pipe lithology.

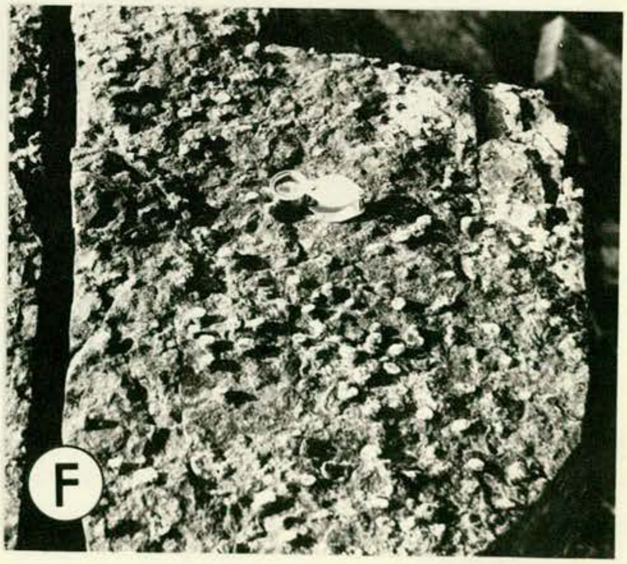
PLATE 1











rings which encircle the core (Plate 2-A) and may occasionally produce an overall structure up to 4 inches in diameter (Plates 2-B & 2-C). More normally, the "trumpet pipes" effect a marginal deformation only ½ to ¾ inches wide (1 to 1½ inch diameter) (Plate 1-D). There is never an indication of a bending of the pipes or of a "U"-shaped tube as is common with some burrowing annelids (e.g. *Arenicola* sp.). It seems possible that the trumpet pipes represent a greater original porosity of the sand through which the organisms were burrowing or they could reflect a longer period of residence within the tube by the organism. The conformity in size of the central "core" of the trumpet pipes to the cores of the "ordinary" pipes renders doubtful the suggestion of Peach et. al. (1907) that these were produced by a different species.

Peach et. al. (1907, p. 372) offered the following views concerning the pipe structures:

> "These (the pipes) are attributable to burrowing annelids, and they not only indicate the presence of such animals during the deposition of the quartzite, but also of sufficient organic matter having been mixed with the sand where they occur to furnish nourishment to the worms."

This statement is not necessarily valid, for the pipes may represent merely a shelter or resting place for the organism and need not be a feeding burrow. On ethological and ecological grounds, W. Schäfer (1962) is able to divide recent trails and burrows into four types based on their characters: 1) food gain; that is scavenging, grazing, eating of sediment, filtering, catching with tentacles, 2) erection of a dwelling (housing paved with a mucous lining), 3) flight (escape burrow), or 4) search for a resting place.

The ubiquitous vertical and nearly straight character of the pipes (Plate 2-D) discourages the interpretation that these burrows are the ramifying trails of grazing, scavenging, eating or filtering of sediment. There is no evidence of a mucous-lined borehole. The mineralogy and colour of the sandstones clearly does not support the view of an animal feeding from the sediment unless the animal was extremely effective in cleaning the sediment both laterally and vertically. The spacing and geometry of the pipes does not fulfill this requirement. It seems far more likely that the pipes are the trace fossils of animals which were suspension feeders and not feeding from the sediment of their vertical burrows.

Although annelids are perhaps a likely animal to have been responsible for the burrows (Lessertisseur, 1955; Westergard, 1931) there are several other organisms which might have been responsible. Pelecypods, gastropods, scaphopods, holothurians, coelenterates, plants, and even ascending gas bubbles have been reported to form similar if not identical structures. Fenton and Fenton (1934) show fairly convincing evidence that recent phoronids construct tubes which are similar in shape, structure, size, and spacing to the Skolithos trace fossils as originally defined by Haldeman in 1840 (Fucoides ?linearis or Scolithus linearis):

> "Stem simple (never branched), rectilinear surface nearly even, dimater 1/8 to 1/4 inch, length several feet, cylindrical or compressed. Locality - Pennsylvania."

Similar trace fossils have been reported from the Cambrian and/or Ordovician rocks of Europe, N. America, Greenland, Sweden, and Tasmania (Moore, 1962, Treatise Vol. "W").

There are no known criteria by which these burrows may be assigned to a particular animal. Richter (1921) has diminished the credibility of the ascending gas bubble hypothesis by arguing that it fails to explain the close-packed, never-connected tubes. He has proposed that the tubes are constructed vertically from the base upwards, of sand grains cemented during sedimentation by a mucous secretion of a suspension-feeding annelid (c.f. Sabellaria) which occupied the tube head up and, in colonies, built up "reefs" of sand tubes. This hypothesis also loses favour as an explanation for the pipes because of the following factors: 1) the tubes of the pipe rocks are very straight and regular in their diameters, 2) the sediment containing the pipes is conspicuously flat bedded with no sign of "reef" structures, 3) the burrowing character of the pipes is indicated by the sharp truncation of bedding laminae (e.g. the "ordinary" pipes) and by the deflection of bedding (e.g. the "trumpet" pipes.

Karcz (1963) has suggested that the presence of organic matter upon which the organisms were feeding may have caused the difference in stratification between the flat-bedded "Pipe Rock" sandstones and the current-bedded Lower sandstones. He invokes a complex explanation of organic binding agents but to his proposals it would seem necessary to add .... and the organisms which were feeding on the organic material also consumed the feldspar minerals. Karcz's hypothesis fails to consider the close correlation between the presence of the pipe structures and the mineralogical maturity of the sediment as well as with the variation in bedding structures. An environmental control causing the bedding characteristics, the mineralogical variation, and the presence or absence of the pipes seems a much more likely explanation than that of organic binding agents controlled the patterns of deposition. Karcz was perhaps partly misled in his hypothesis of organic binding by the views of Peach et. al. (1907) that the sediment must have been rich in organic matter.

This is news
to me
I. Karcz
12.5.68

(sic!)

Of the various hypotheses for the genesis of the "pipes", only two seem adequate to explain the following factors:
1) Simple tube (never branching), 2) several feet in length,
3) diameter 1/8 to 1/4 in, 4) no preserved hard parts,
5) very clean sediment with or without close-packed pipes,
6) occasional downward deflection of bedding laminae adjacent

to a central tube, 7) common truncation of sediment laminae by the tube structure.

Either suspension-feeding annelids or the stalk-borne phoronids would appear capable of fulfilling all of these conditions by their burrowing habits and this author can find no basis for a preference.

D. DIAGENESIS

The most striking and conspicuous diagenetic alteration within the Eribol sandstones is the absence of pore space (Plate 2-E). J.M. Taylor (1950) has divided the mechanisms of pore space reduction in sandstones into three categories: 1) simple pore filling, 2) solid flow, and 3) pressure solution and redeposition. Within the Eribol sandstones, the third mechanism appears to have been the principal method by which the pore space was reduced. Simple pore filling may have been important early in the diagenetic history but if so, its effects have been obscured by subsequent pressure solution. Such a sequence is unlikely since pore filling by quartz overgrowths on detrital grains would be expected to produce many planar contacts between the developing crystals but, in the Eribol sandstones, such planar contacts are rarely observed. Also rare are the "ghost" outlines of detrital grains commonly found in pore-cemented orthoquartzites. The solid-flow mechanism proposed by Taylor was applied mainly to rock fragments or particles having a relatively high plasticity (e.g. glauconite). Solid, plastic flow of quartz grains has not been demonstrated and seems unlikely to have made any significant contribution to the reduction of pore space in these quartzose sandstones.

The presence of clay or iron minerals is taken by some writers (Thompson, 1959; Lerbekmo & Platt, 1962; Weyl, 1959) as a prerequisite of or at least catalytic to the pressure-solution mechanism. The extremely low clay content of some of the orthoquartzites of this study would not appear to support this view for the loss of pore space has been universal regardless of (recognizable) clay content. In certain instances, the evidence for pressure-solution is manifest as sutured grain contacts (Plate 2-F). More commonly, however, the grains have the appearance of having interpenetrated one another without any significant suturing although the interpenetration must have been accomplished by solution processes (Plate 3-A). Many of the rocks also fail to show any perceptible evidence of redeposition. This suggests that pressure solution rather than solution and redeposition has been the dominant mechanism for pore space reduction. Where the removed silica may have migrated to presents a perplexing problem if solution processes alone are proposed and is a serious obstacle to this hypothesis.

The lack of displacement of either horizontal structures (bedding) or vertical structures (the pipes) indicates that

PLATE 2

A. Vertical, plan-view of bedding-plane truncation of the concentric ring structures of the "trumpet pipes".

B. Bedding plane exposure of "exceedingly large pipes". Painted intervals are 1 foot each.

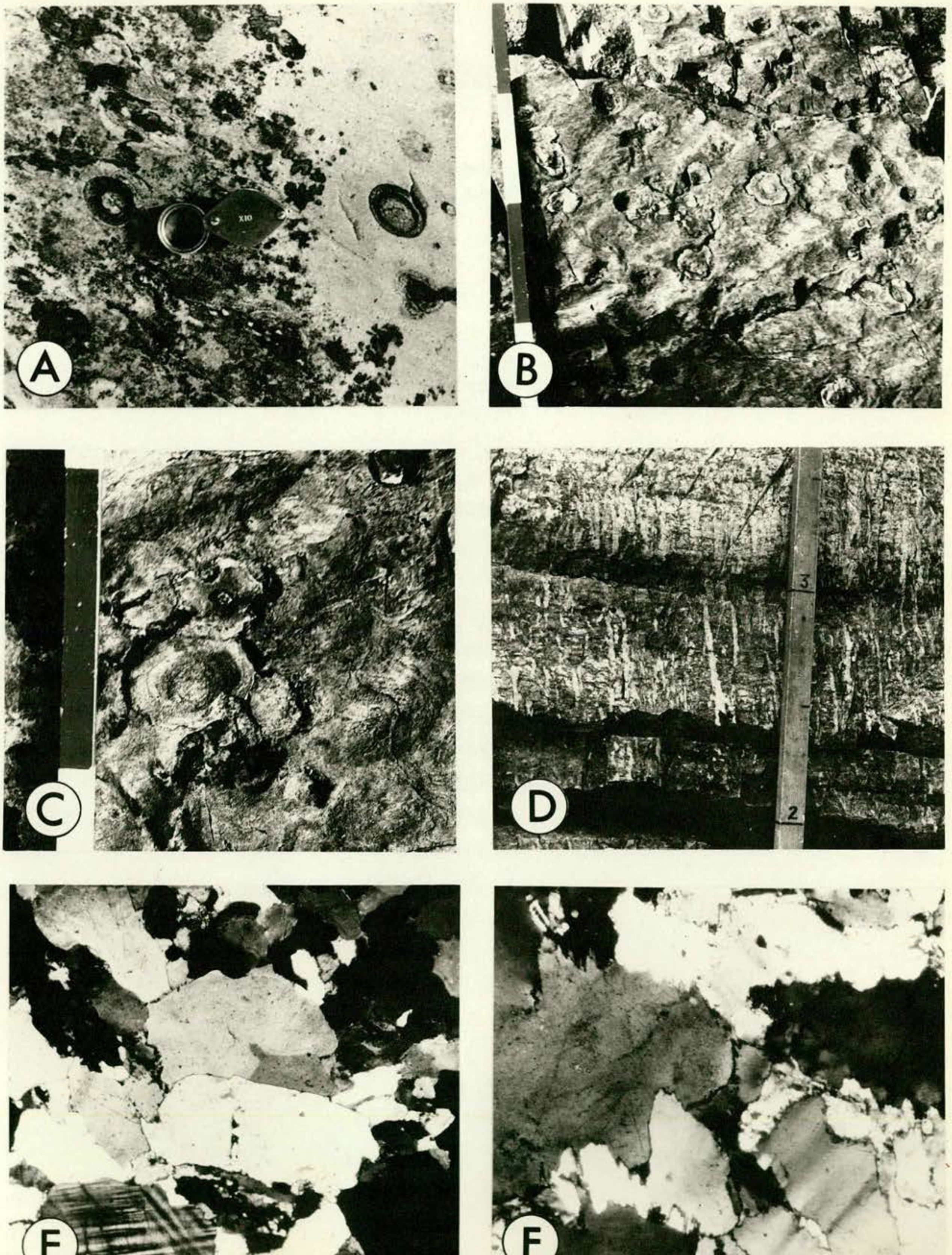
C. An "exceedingly large pipe" showing the concentric ring structure caused by downwarping of sediment laminae toward a central core.

D. "Ordinary pipes" showing their nearly straight character and their consistent orientation normal to bedding. Numbered intervals on staff are 1 foot each.

E. (Q-245) Photomicrograph of Lower sandstone specimen showing the total absence of pore space. Dark areas are grains at extinction and not porosity. Nicols crossed. X 30.

F. (Q-90) Photomicrograph of highly sutured quartz grain contacts. Nicols crossed. X 40.

PLATE 2



the pressure responsible for the solution processes was probably load pressure or vertical compression exerted on strata while their attitude was horizontal. Overburden or "load" pressures are sometimes mistakenly considered as hydrostatic or "lithostatic" in character. The concept of lithostatic pressure is perhaps valid in an overall sense but microscopically, in clastic rocks, this concept remains invalid while there is open pore space or while pore fluids with a lower density than the clastic detritus may escape from the sediment. The pore space in a clastic sediment represents potential gravitational energy and it is the energy of this disequilibrium that has been proposed as the driving force for pressure-solution reactions (Weyl, 1959). An energy equilibrium will be attained only when the pore space has been reduced to zero and only then can the pressures within the rocks be considered lithostatic.

The pressure exerted at the points of contact between grains significantly reduces their solubility along these surfaces of contact (Siever, 1962). The removal of mineral matter is presumably accomplished by diffusion in a solution film between the mineral grains. Weyl (1959) concludes that the rate of transport depends upon the grain size, the amount of pressure exerted upon the grains, the diffusion constant in the solution film, the film thickness, and the stress coefficient of solubility for a particular mineral species. The theory predicts an inverse relationship between grain size and the degree or rate of pressure solution. This theory is supported by emperical data in the Eribol sandstones. Weyl, (1959) also subscribes to the catalytic role of clay minerals in pressure solution but does not regard their presence as a prerequisite to all pressure solution.

As previously indicated, textures demonstrating solution phenomena are much more apparent in the Eribol sandstones than are textures of redeposition (e.g. precipitation of silica as overgrowths or simple pore filler). Even in those horizons which do exhibit authigenic overgrowths, it is doubtful if this precipitation occurred contemporaneously with pressure solution. If this were the case, one might expect to find early overgrowths sutured by later pressure solution. From studies on the sandstones of West Virginia, Heald (1950) concluded that the lack of suturing of overgrowths indicated that the authigenic silica was not derived from pressure solution. Within the Eribol sandstones, this author would conclude that the two processes probably occurred simultaneously, although perhaps not in juxtaposition, and that probably little silica was added to or lost from the system. The evidence in support of this conclusion is three-fold: 1) There are no detectable facies variations in the processes of pore space reduction. This detracts from any hypothesis of lateral migration and precipitation of dissolved material. 2) The lower 800 feet of the overlying carbonates are virtually free of authigenic silica either as quartz euhedra or as chert nodules. This reduces the probability that the silica has migrated upward.

3) The impermeability of the Lewisian metamorphic rocks and their lack of silicification phenomena does not favour the hypothesis that silica-bearing solutions have migrated downward. Therefore, although textural evidence indicative of redeposition is lacking, it seems, nevertheless, probable that the silica must have been redistributed and redeposited within the sandstones. Perhaps it is only the obscurity or lack of "ghost" detrital grain boundaries that have disguised the redepositional aspects of pore space reduction. The lack of strained quartz optics in the Torridonian rocks where immediately underlying the Eribol sandstones suggests that these rocks were already cemented before the compaction and pressure solution processes occurred in the Eribol sandstones.

Nearly all of the quartz grains within the Eribol sandstones display highly strained or undulatory extinction patterns although fracture of the grains is rare and inconsequential in the reduction of pore space within the rocks. The patterns of strain can sometimes be related to pressure "points" with adjacent grains (Plate 3-B). Inclusions within quartz grains have often produced deflections in the strain patterns (Plate 3-C). Where authigenic overgrowths are present on quartz grains, they invariably share the strained extinction pattern of the host grain (Plate 3-D). Petrologists have long assumed that this relationship could be interpreted as indicating pre-strain development of the overgrowths (Lowry, 1956). Recent experimental work by Ernst and

Blatt (1964) has shown however that synthetic overgrowths may be produced simultaneously on undulatory and non-undulatory quartz grains and that in all cases, the secondary enlargements are syntaxial with their nuclei. These experimental results make the interpretation of the observed textures within the Eribol sandstones much easier. It would be difficult indeed to explain why the straining of the quartz grains had been deferred until so late in the process of compaction when it can be easily demonstrated that grains within the overlying carbonates and obviously under considerably less contact pressure have developed strained extinction patterns (Plate 3-E). The optical deformation of the quartz grains probably occurred relatively early in the compaction process before significant pressure solution or pore-filling had occurred. Any overgrowths developed subsequent to this straining would assume the extinction patterns of the host grains (Ernst and Blatt, 1964).

The ubiquitous loss of pore space, the pressure solution features (e.g. suturing) and the strained extinction patterns in the quartz grains might easily mislead an investigator into assuming that quite phenomenal pressures and forces were involved in the compaction of these sandstones. The feldspar grains within the rocks bring the situation into a more realistic perspective by their near-total resistance to the forces of compaction which have so altered the quartz detritus (Plate 3-F). The lack of deformation, solution, or fracturing

PLATE 3

A. (Q-6) Photomicrograph showing the interpenetration of quartz grains without any perceptible suturing effects. Nicols crossed. X 30.

B. (Q-95) Photomicrograph of quartzose sandstone showing a probable relation of undulatory extinction patterns to pressure points. Nicols crossed. X 80.

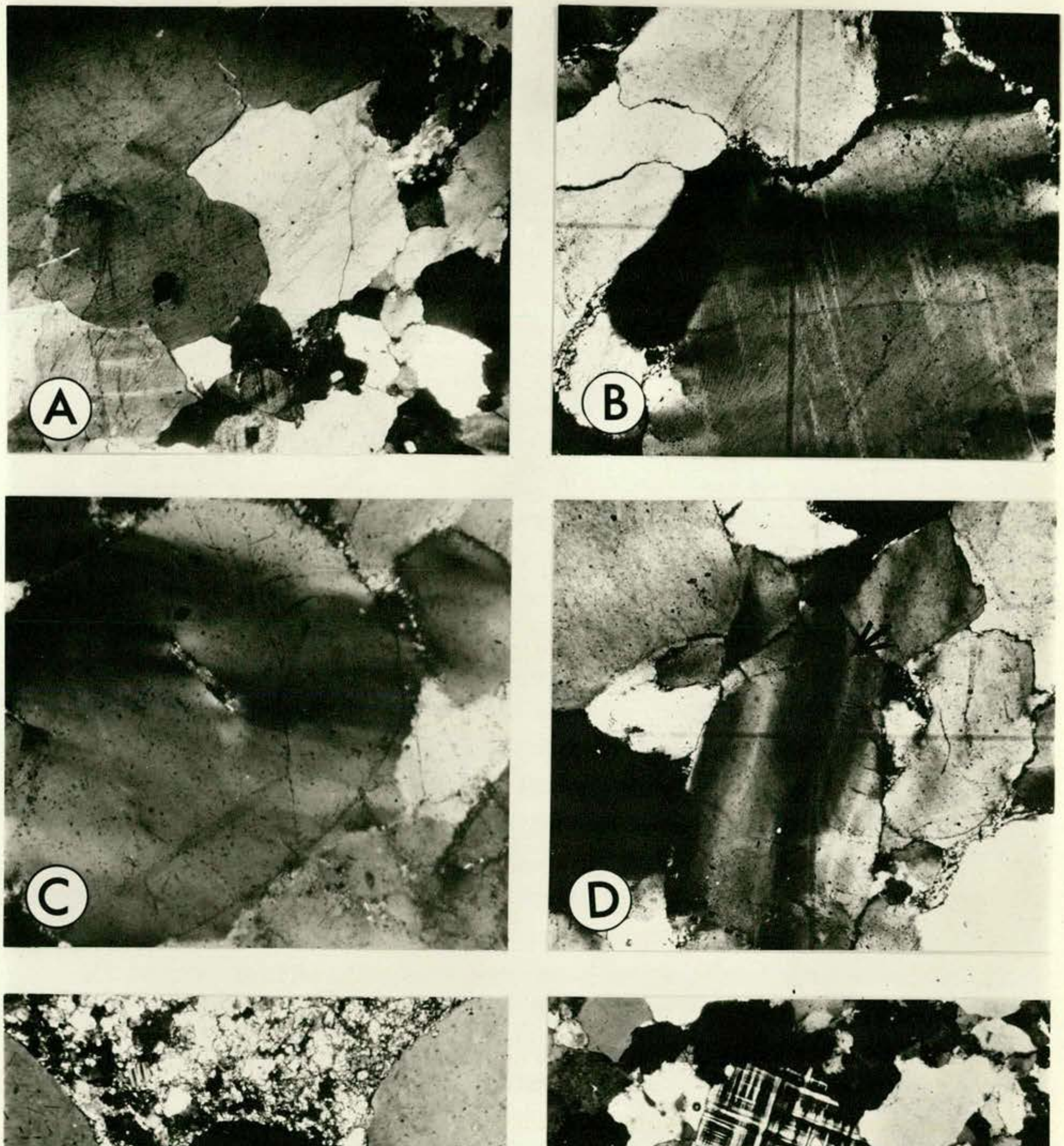
C. (P-70/A) Photomicrograph of quartz grain with a mica inclusion which has deflected the undulatory extinction pattern. Overgrowth shares undulatory pattern. Nicols crossed. X 100.

D. (P-70) Photomicrograph of strained detrital quartz grain with a syntaxial overgrowth sharing the undulatory extinction pattern. Nicols crossed. X 80.

E. (D-616) Photomicrograph of quartzose carbonates showing a quartz grain squeezed between two larger grains and demonstrating clearly the relationship of straining to pressure. Very few of the quartz grains in the carbonate exhibit strained extinction patterns. Nicols crossed. X 50.

F. (Q-2) Photomicrograph of rounded microcline grain in quartzose sandstone. Pore space in the sandstone has been reduced to negligible proportions by pressure and solution processes yet the detrital feldspar retains its abraded shape. Nicols crossed. X 25.

PLATE 3



in the feldspars is in agreement with the work of Heald (1955) on the relative stabilities of various minerals under stress but is contrary to the views of some other workers (Carozzi, 1960). This greater competence of the feldspars over quartz grains in the stress environment affords the only available clue regarding the original "roundness" or textural maturity of the sands. The abrasion of orthoclase in fluvial experiments (Kuenen, 1959) is approximately twice as rapid as quartz abrasion under the same conditions. The limited amount of alteration of the detrital feldspars suggests that abrasion rather than weathering was chiefly responsible for the observed roundness in the Eribol sandstones. Quantitatively, the roundness of the feldspar grains in the Eribol sandstones is approximately 0.7 on the roundness scale of Krumbein and Sloss (1951). If the quartz grains lost half as much volume by abrasion (Kuenen, 1959) they would classify approximately in the 0.3 to 0.5 categories on the same roundness scale. Pressure solution processes have reduced the present roundness factor of the quartz grains to 0.1 or less. In the arkosic sandstones, where the roundness of the feldspars may be determined, the mineralogy supports the textural immaturity of the sediment based on a projection of the probable original degree of roundness. Any conclusions regarding distance of transport or proximity to provenance area based on the textural or mineralogical criteria would be little better than idle speculation and shall not be ventured.

Zircon, tourmaline, muscovite, biotite, magnetite (or ilmenite?), rutile, and feldspars are occasionally observed as inclusions within the detrital quartz grains. Some of the tourmaline and zircon grains observed as inclusions in quartz grains appear to be rounded. This suggests that these inclusions were perhaps abraded before inclusion or that they were partially fused during inclusion. The latter possibility may be excluded on the basis of relative melting temperatures. Pursuing the implications of the first possibility, the problem of a source of quartz grains containing rounded inclusions is encountered. The absence of secondary quartz of this crystal size within the Torridonian arkosic sediments eliminates that source and leaves only the Lewisian complex from which the grains would need be derived. This deduction leads to the requirement that the Lewisian rocks would need to be, at least in part, meta-sediments. This conclusion is in accordance with the work of Sutton and Watson (1950) but they do not note in their arguments, the occurrence of rounded inclusions. Before concluding such a complex history for these grains with inclusions, a thorough investigation of the character and relationships of the zircon and tourmaline grains in the Lewisian rocks would be well advised and a second problem concerning the provenance of occasional metamorphic quartzite pebbles found in the Torridonian and in the Lower member of the Eribol sandstones might be simultaneously solved. Such an investigation would require detailed knowledge of the

Lewisian complex and its distribution and is clearly beyond the scope of the present study.

In addition to secondary enlargements on quartz grains, several other authigenic minerals are found in the Eribol sandstones. Both the potash and plagioclase feldspar grains very commonly possess euhedral, authigenic overgrowths (Plate 4-A). Contrary to the secondary enlargements on feldspars discussed by Irving and Van Hise (1884) and Heald (1950), the overgrowths on microcline, orthoclase and plagioclase within these sandstones do not appear optically to share the twinning phenomena of the host grains (Plate 4-B). They appear only to be syntaxial with the general optical orientation of the detrital host. The overgrowth nearly always exhibits a lower birefringence and a higher susceptibility to alteration than the host grains. The euhedral faces of the authigenic fedlspar overgrowth are often clearly cross-cutting in their relationships to the sutured and undulose patterns of adjacent quartz grains (Plate 4-C). This suggests that the feldspar overgrowths have developed relatively late with respect to the reduction of pore space in the Eribol sandstones.

Another authigenic mineral of interest is leucoxene which occurs as well-rounded ?pseudomorph grains, as alteration rims on ilmenite grains, and as finely divided inter-granular cement between quartz grains. The rounded leucoxene grains are most probably an alteration of ilmenite rather than original detrital constituents of the sediment. An extraordinary aspect of the leucoxene in the Eribol sandstones is its occasional appearance as an authigenic replacement of detrital quartz and feldspar grains (Plate 4-D). The source of the leucoxene is probably largely accounted for by the alteration of ilmenite but a possible second source seems worthy of consideration. It has been previously established that pressure solution processes have profoundly affected these sandstones. Pressure solution of quartz grains which contained fine rutile needles (Plate 4-E) could conceivably contribute as a source of leucoxene.

Authigenic pyrite appears in the upper few feet of the Pipe Rock sandstones (Plate 4-F). The same horizons are also generally dolomitic and the pyrite euhedra replace both the detrital quartz grains and the carbonate cement simultaneously (Plate 5-A). Very often the secondary pyrite has itself developed a secondary oxidation rim of hematite (Plate 5-B). The dolomite of these horizons is also demonstrably authigenic and occurs as pore filler and as an authigenic replacement of detrital quartz grains. A sequence of diagenetic replacements within these upper sand-stones is easily established as 1) dolomitization, 2) pyritization, 3) Oxidation or hematitization. Within the samples which were examined, this sequence of diagenesis is never contradicted.

At some localities, the top 6 inches to 1 foot of the sandstones contain detrital glauconite pellets. The pore space of this uppermost unit has been reduced mostly through pore filling by syntaxial overgrowths on the quartz grains but

PLATE 4

A. (Q-6) Abraded potash feldspar grain which has developed a euhedral authigenic overgrowth. Nicols crossed. X 30.

B. (Q-75) Abraded microcline grain with an authigenic overgrowth which does not share the twinning phenomena of the host grain. Nicols crossed. X 30.

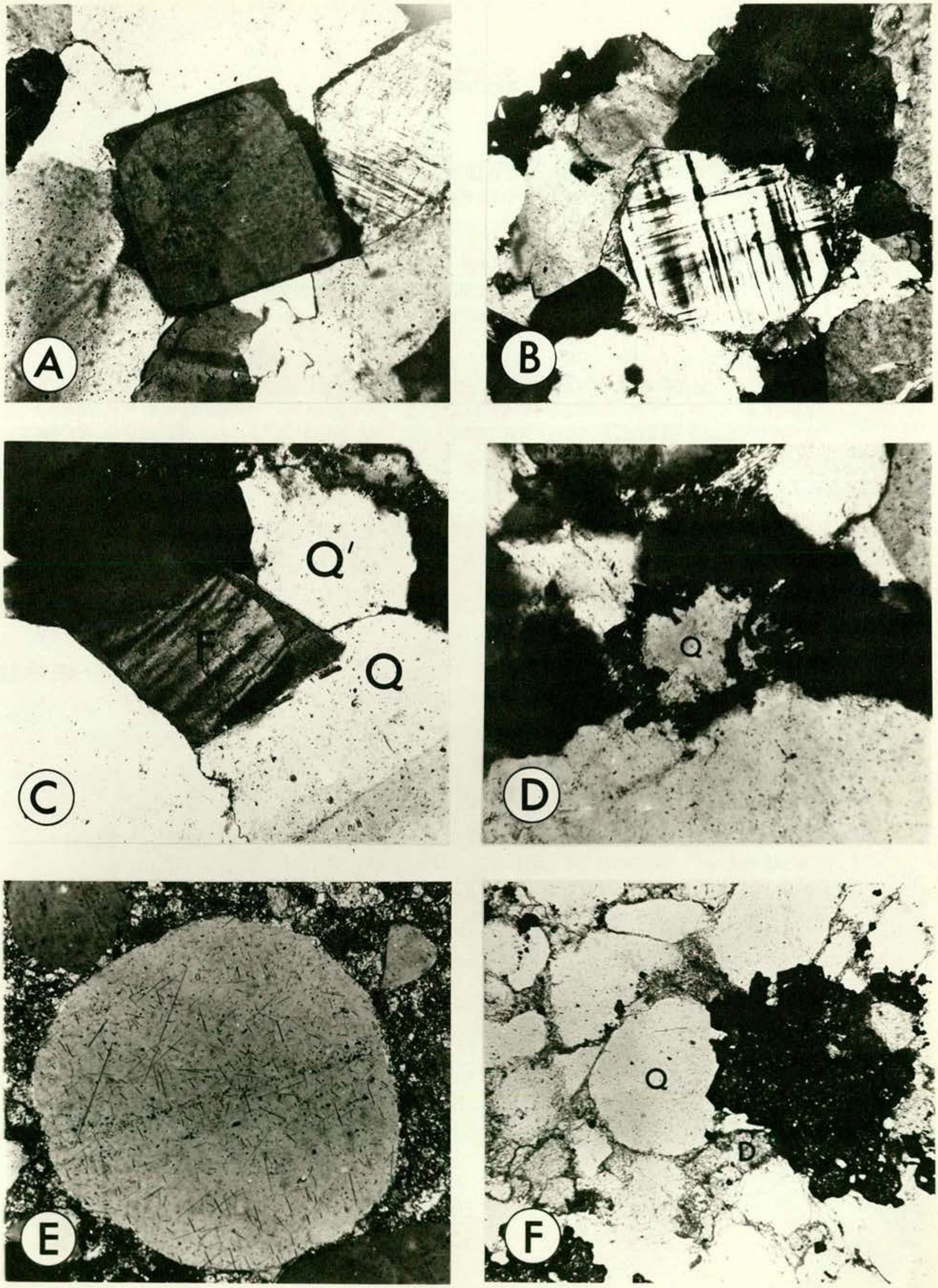
C. (Q-25) Rounded feldspar grain (F) with a euhedral overgrowth clearly cross-cutting a pressure solution contact between adjacent quartz grains (Q and Q'). Nicols crossed. X 80.

D. (Q-245) Photomicrograph of authigenic leucoxene (black) cutting into and replacing a detrital quartz grain (Q). Nicols crossed. X 30.

E. (D-616) Nearly spherical quartz grain in quartzose carbonate horizon showing abundant rutile inclusions within the grain. The grain also demonstrates peripheral replacement by dolomite. Nicols crossed. X 30.

F. (P-313) Photomicrograph showing pyrite (Dark area with planar faces) authigenically replacing dolomite (D) and quartz (Q). The pyrite is surrounded by an oxidation coating of hematite 20 to 50 microns thick. Plane polarized, transmitted light plus obliquely reflected light. X 30.

PLATE 4



evidence of compaction is shown by the solid flow deformation of the glauconite pellets (Plate 5-C) and by strain patterns in the detrital and secondary quartz. Glauconite pellets, where they are in contact with the quartz grains, occasionally show evidence of a microscopic replacement of the quartz (Plate 5-D). In these instances, the glauconite has sometimes replaced both the detrital quartz grains and their secondary overgrowths suggesting that, in a stress environment, glauconite was chemically more stable than either the detrital quartz grain or its early (pre-strain, pre-solid-flow) secondary overgrowths.

Glauconite replacements of calcite, biotite, organic matter, clay minerals, and siliceous sponge spicules are well established diagenetic phenomena (Carozzi, 1960) but to this author's knowledge, the replacement of detrital quartz grains or their secondary enlargements by glauconite is unreported. A very similar occurrence involving glauconite, dolomite, and detrital quartz (Swett, 1964, p. 624, fig. 4F) is now believed to have been incorrectly interpreted as dolomite replacement of quartz followed by glauconite replacement of the dolomite. This instance, within the Eribol sandstones, of the direct replacement of detrital quartz by glauconite leads to the conclusion that the earlier interpretation of the quartz - dolomite - glauconite sequence in the Manitou formation of Colorado should more probably have been interpreted as a quartz - glauconite - dolomite sequence. Within the Eribol sandstones, there is no evidence for a third mineral having entered the replacement mechanics of the glauconite and quartz. Some of the glauconite grains do show a secondary alteration to sericite but this alteration does not enter into the quartz-glauconite replacement and is probably a much later diagenetic event.

E. PROVENANCE

The clear overstep relations of the basal unconformity of the Eribol sandstones over both Lewisian and Torridonian rocks suggest the probability that the detritus of the Eribol sandstones was derived from both sedimentary and metamorphic source rocks. No rock fragments from either source were noted in the Eribol sandstones above the basal 5 to 10 feet with the exception of some problematical quartzite pebbles which are not directly attributable to either the Torridonian or the Lewisian. No rounded overgrowths were noted and the only suggestion of multiple cycles of sedimentation and lithification is the presence of the ?rounded inclusions within some quartz grains.

The thickness of the Torridonian succession exceeds 9000 feet in some localities (Phemister, 1960). This thickness is perhaps greater than the average pre-erosion accumulation along the entire length of the Cambrian unconformity but, at least locally, this volume of sedimentary rocks and perhaps an even greater thickness were totally removed by a period of erosion. The detritus from this erosion of Torridonian strata must certainly have contributed significantly to the Eribol sandstones. The Eribol sandstones have been affected by too many variables, however, to permit even speculative evaluation of the quantitative relative contributions of detritus from the Torridonian and Lewisian sources.

The provenance for the authigenic minerals is possibly within the original sediments with the authigenesis merely reflecting an internal redistribution of the original chemical constituents into a more stable mineralogical pattern. General thoughts regarding the causes and mechanisms of authigenesis will be developed more thoroughly in conjunction with and after discussions of the remaining lithologic units.

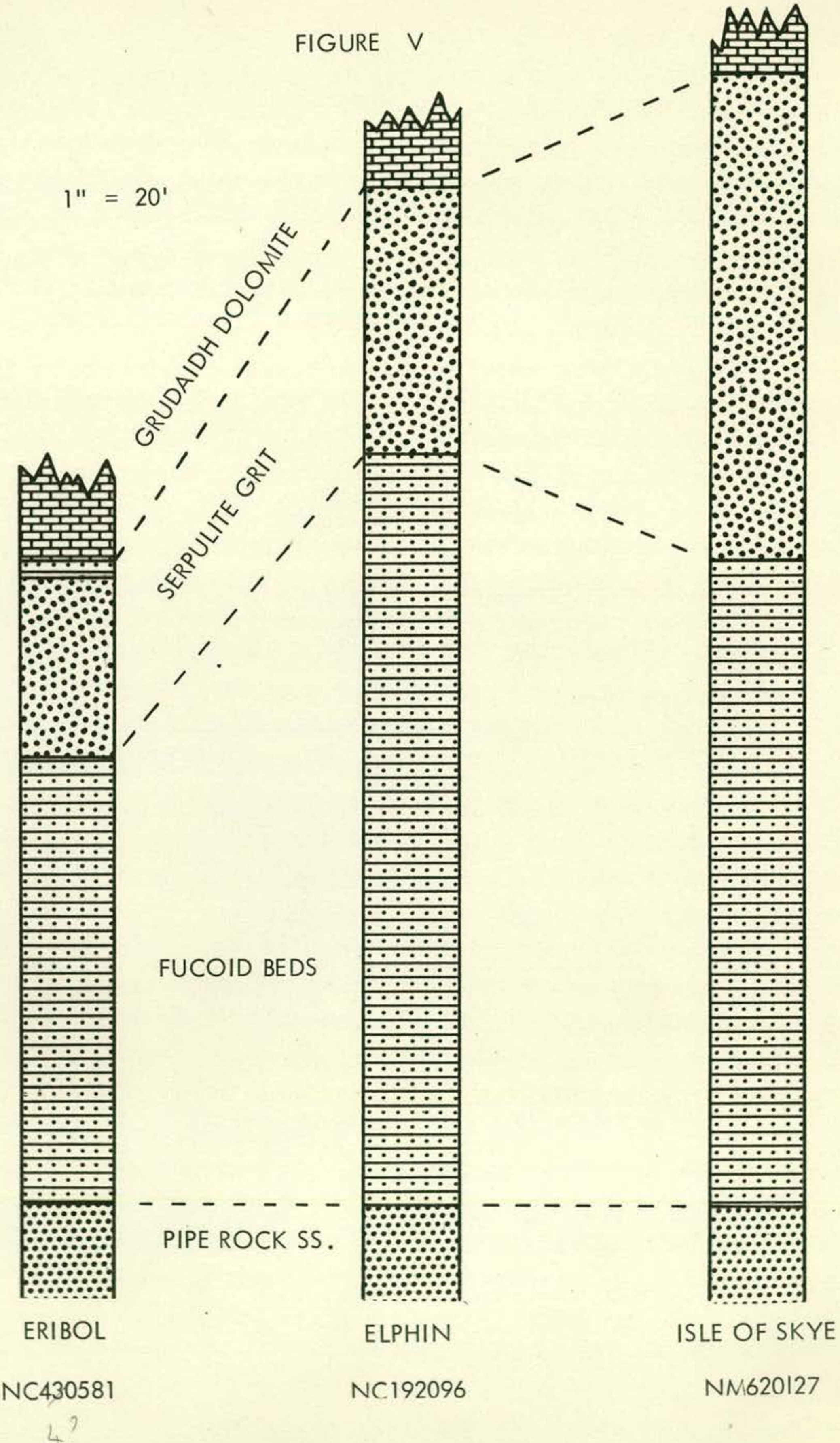
III. AN T-SRON FORMATION

A. GENERAL STATEMENT

The dolomitic shales, dolomitic siltstones, shaly dolostones, silty dolostones and the band of orthoquartzite above the finer-grained units have been variously referred to as the Intermediate Series, the Passage Series, or the Middle Series. In this thesis, this portion of the Cambro-Ordovician succession is renamed the An t-Sron formation after a "type" locality on the west shore of Loch Eribol where a complete sequence of these strata is continuously exposed. Most workers have agreed with the two-fold subdivision of the "Series" into a lower unit, the "Fucoid Beds" (based on the early erroneous interpretation of Planolites trace fossils as plant fossils) and the "Serpulite (or Salterella) Grit", an orthoquartzitic sandstone. This two-fold subdivision is followed within the An t-Sron formation with the "Fucoid Beds" and the "Serpulite Grit" as member units..

The "Fucoid Beds" are relatively unresistant to the forces of weathering and erosion and hence good surface exposures are seldom encountered. The "Serpulite Grit" is considerably more resistant and commonly forms a small escarpment. During the field researches for this study, three locations were visited at An t-Sron, Elphin, and Ord, where the exposures of both members of the formation are virtually complete (precise locations are given with the stratigraphic sections - fig. V).

FIGURE V


1" = 20'
GRUDAIDH DOLOMITE
SERPULITE GRIT
FUCOID BEDS
PIPE ROCK SS.
ERIBOL
NC430581
ELPHIN
NC192096
ISLE OF SKYE
NM620127

4?

At An t-Sron, the continuously exposed stratigraphic thickness of the formation is roughly 80 feet, while at Elphin and Ord, the same units are nearly twice as thick (fig. V). This seems, nevertheless, a remarkably constant unit both lithologically and in thickness over its 100 mile length of outcrop.

The discovery by Professor Lapworth in 1885 of fragments of *Olenellus* within the "Fucoid Beds" (published 1888) and the confirmation of this discovery through additional specimens located by the Geological Survey in 1891 (Geikie, 1891; Peach and Horne, 1892; Peach, 1894) have provided the only true index fossils within the Cambro-Ordovician sequence. This important discovery has, however, permitted the correlation of these rocks - and, by position, the underlying sandstones - with the Lower Cambrian of Northwestern Europe, the British Islands, and Central and Southwestern Europe by Peach et. al. (1907), to Greenland (Poulsen, 1951; Cowie, 1960) and to North America (Peach, 1912).

Peach et. al., (1907) list 43 different fossils from the *Olenellus* zone including echinoderms, annelids, brachiopods, trilobites, phyllocarids, and gasteropods. Despite careful searching in these horizons, this author is unable to make any new paleontological contributions to their list or even to confirm more that the presence of annelid? trails (*Planolites*) (Plate 5-E), echinoderm fragments (Plate 5-F), ?*Volborthella* (Plate 6-A), and Salterella.

PLATE 5

A. (P-319) Photomicrograph demonstrating pyrite (dark area) replacement of quartz (Q) and dolomite (D). Plane polarized, transmitted light plus obliquely reflected light. X 30.

B. (P-319) Authigenic pyrite in a silty dolostone demonstrating the peripheral oxidation of the pyrite to hematite (black rim). Plane polarized, transmitted light plus obliquely reflected light. X 50.

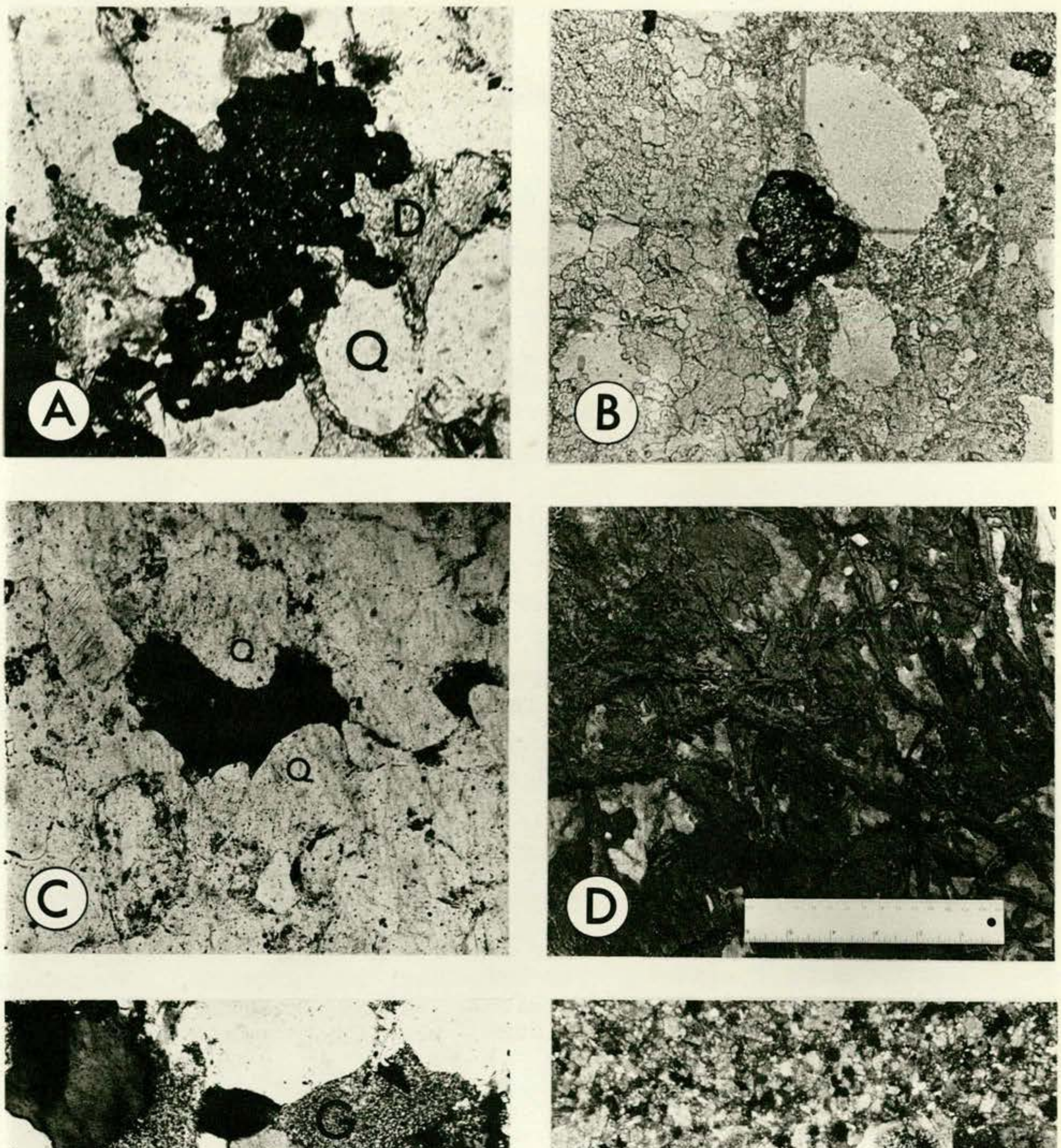
C. (P-319/Skye) Glauconite (dark) occupying the interstitial area between quartz grains (Q). The glauconite has been deformed plastically to fill the pore space. Plane polarized, transmitted light. X 30.

D. (P-319/Skye) Photomicrograph demonstrating glauconite (G) texturally cross-cutting and apparently replacing two adjacent quartz grains (Q). Nicols crossed. X 50.

E. Bedding-plane exposure of Planolites trace fossils (misinterpreted as Fucoid impressions by Peach et. al., 1907).

F. (I-53) Photomicrograph of silty dolostone containing echinoderm fragments with syntaxial overgrowths. Hematite filled sterome structure of the optically continuous grains is the basis for identification as echinoderm debris. Nicols crossed. X 30.

PLATE 5



B. PETROGRAPHY

The "Fucoid Beds" are a diverse mixture of lithologies including dolomitic siltstones, silty dolostones, dolomitic shales, argillaceous dolostones, dolomitic sandstones, and sandy dolostones. In general, the small grain size of all, or at least part, of the clastic portions of these rocks precludes modal analyses and the preceding rock names are based only on visual estimates of the percentage compositions. Weathered exposures of the "Fucoid Beds" are characteristically brown coloured on the weathered surfaces but are dark grey on freshly broken surfaces. These units commonly display small scale current-bedding or ripple features and evidence of reworking of the sediment, generally parallel to bedding, by the action of burrowing organisms (Planolites) (Plate 5-E). In thin section, the rocks appear to be composed mainly of angular detrital quartz grains set in a finer groundmass of unidentifiable clay-size matrix material. Dolomite or ferroandolomite is a conspicuous to predominant constituent of nearly all specimens.

The "Serpulite Grit" member of the An-t-Sron formation is mineralogically an orthoquartzite and texturally, a coarse, well-sorted sandstone. Adjacent to both the upper and lower boundaries of this member, there is dolomite cementation of the sandstones and the occasional occurrence of detrital carbonate grains (probably echinoderm fragments) (Plate 6-B).

Feldspars comprise less than 1 percent of the rocks and only trace quantities of the following detrital constituents were noted: siltstone rock fragments, ilmenite, zircon, rutile, tourmaline, pyrite, hematite, sericite, and chitin fragments. The sandstones are conspicuously current-bedded but exposures are inadequate to obtain reliable current direction measurements.

C. DIAGENESIS

The reduction of pore space in the "Serpulite Grit" has been accomplished largely by simple pore filling, partly by dolomite and partly by quartz overgrowths on detrital grains. Pressure solution has also contributed to pore space reduction but to a much more limited extent than in the Eribol sandstones. Sutured and interpenetrating contacts between adjacent detrital quartz grains are uncommon and the optical deformation or straining of quartz is far less pronounced. This possibly permits a significant conclusion regarding the processes of pore space reduction to be drawn from the relationships between the "Serpulite Grit" and the "Pipe Rock" sandstones. Since the "Serpulite Grit" is separated from the "Pipe Rock" member of the Eribol sandstones by only 60 to 80 feet of dolomitic siltstones and silty dolomites, it is inconceivable that the compaction pressures in these two horizons would have been significantly different from one another. Therefore, it seems necessary to invoke some cause other than differences in pressure to explain the differences in degree of quartz deformation and in the mechanisms of pore space reduction. The close association between the presence of pore filler and the lack of quartz suturing and deformation suggests that the pore filler may have reduced the susceptibility of the sandstones to processes of compaction and pressure solution. This does not necessarily support the conclusion of Heald (1950) that the lack of suturing in the secondary overgrowths indicates that pressure solution and pore filling were not simultaneously occurring side by side. The compaction and induration responsible for the conspicuous deformation of the quartz grains and for the suturing of the detrital grains seems likely to have post-dated the pore filling in the "Serpulite Grit". The presence of the pore filler may explain the lack of these pressure features in that unit.

It is also perhaps further possible to conclude that the silica-bearing solutions responsible for pore filling in the "Serpulite Grit" moved downward from the carbonate rocks above rather than upward from the sandstones below. This source direction of the silica is suggested by the lack of pore cementation in the "Pipe Rock" sandstones and the possibility that the silts and clays of the "Fucoid Beds" might have acted as a barrier to the migrating solutions and account for the pore cementation of the "Serpulite Grit" sandstones. A downward movement of siliceous diagenetic solutions is also consistent with a second hypothesis to be presented in subsequent discussions of the "Fucoid Beds".

It is probable that the carbonate cement, now dolomite, was originally precipitated in these units partly as calcite which was later dolomitized. The main evidence in support of this conclusion is the presence of syntaxial "rim cementation" dolomite (Bathurst, 1958) on probably echinoderm fragments (Plate 6-C). These abraded fragments which exhibit a stereome structure typical of echinoderm plates, were almost certainly originally calcitic. It is also probable that the syntaxial rim cementation on these bioclasts was precipitated prior to dolomitization.

Peripheral replacement of detrital quartz grains by dolomite also indicates an authigenic origin for the dolomite (Plate 6-D). It is further possible to demonstrate with certain quartz grains that the dolomite post-dates authigenic quartz overgrowths (Plate 6-E). It might be argued that the secondary overgrowths were developed during a previous period of lithification and were transported with their detrital hosts to their present positions. This argument can be easily countered by the improbable geometry of cemented grains, the degree of sorting of the grains exclusive of the cemented grains, and the lack of rounded or abraded overgrowths or rock fragments. It is most probable, therefore, that the dolomitization is a later process than silicification within these rocks.

Authigenic pyrite which was noted in the upper few feet of the Eribol sandstones becomes somewhat more abundant in the An t-Sron formation. The pyrite maintains the same paragenetic relationships to the detrital quartz and to the carbonate cement (Plate 6-F) but within the An t-Sron formation, the additional information is available that the pyrite is later than the secondary silica in the diagenetic sequence. Oxidation rims of hematite again commonly appear on the authigenic pyrite euhedra, thus revealing a slightly modified sequence of diagenesis as 1) silicification (authigenic overgrowths on detrital quartz grains), 2) Dolomitization (replaces both detrital and authigenic silica), 3) Pyritization (replaces quartz and dolomite) and 4) Oxidation of pyrite to hematite (hematitization).

D. GEOCHEMICAL STUDIES

In the director's report in the Summary of Progress of the Geological Survey of Great Britain for the Year 1962, C.J. Stubblefield (1963, p.6) states:

> "Geochemical studies of British black shales were continued and further examples of their enrichment in uranium were found. This work also revealed that shales of Cambrian age in the Northwest Highlands were remarkable for their content of potash: four samples representative of large tonnages of rock

PLATE 6

A. (I-40/Kionard) Dolomitic siltstone containing a probable organic structure (c.f. Volborthella sp). Plane polarized, transmitted light. X 30.

B. (I-59) Sandstone from the lower part of the Serpulite Grit member of the An t-Sron formation showing carbonate grains (probably echinoderm fragments) (E) with overgrowths. Nicols crossed. X 50.

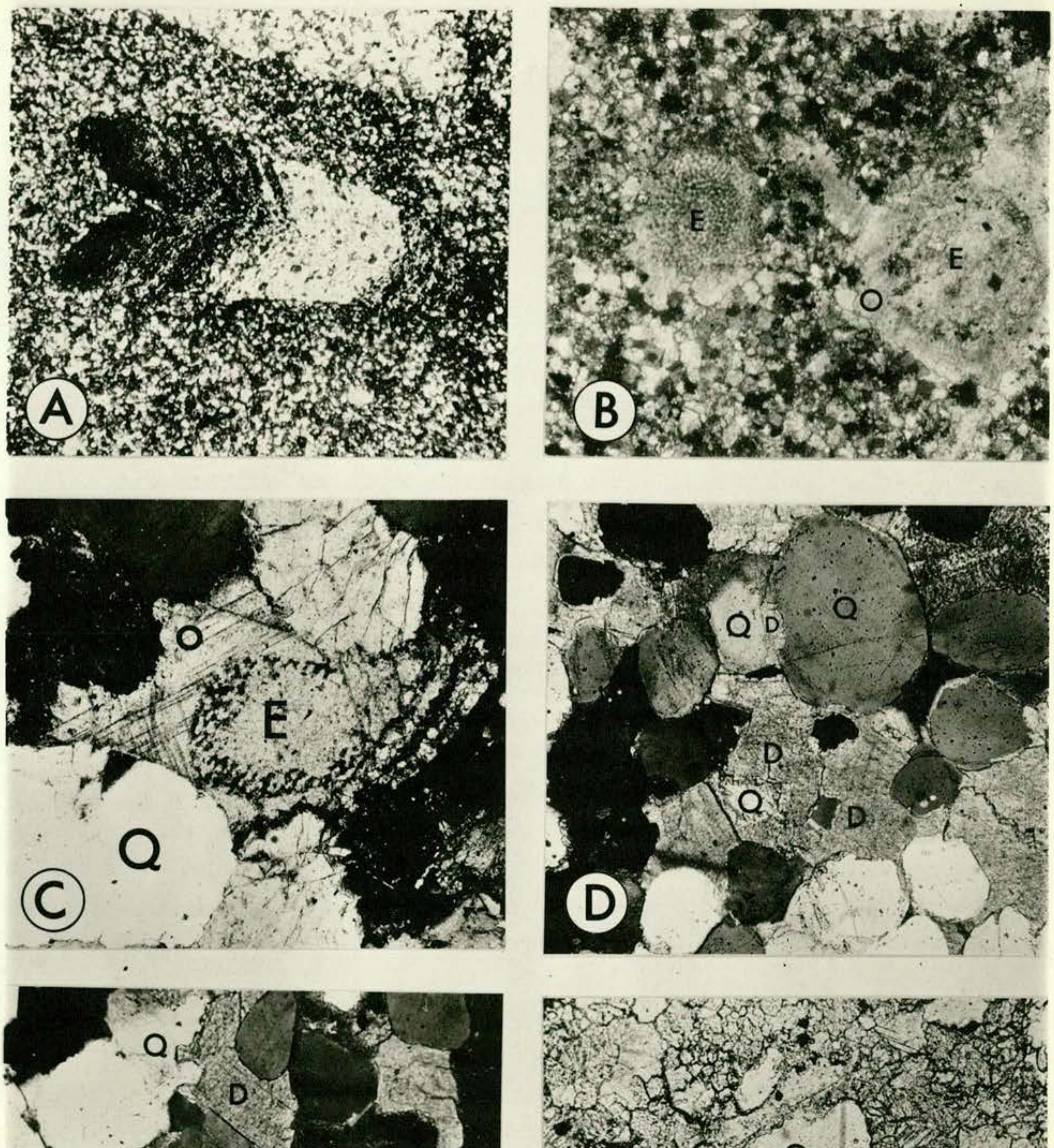
C. (I-59) Serpulite Grit sandstone with an echinoderm fragment (E) with a large syntaxial overgrowth (O). The stereome structure of the carbonate grain is filled with hematite. Nicols crossed. X 100.

D. (I-76) Dolomitic sandstone showing dolomite (D) replacement of detrital quartz grains (Q). Nicols crossed. X 50.

E. (I-76) Dolomitic sandstone showing dolomite (D) replacement of the detrital quartz grains (Q) and their syntaxial authigenic overgrowths. Nicols crossed. X 50.

F. (I-78) Sandstone from the upper portion of the Serpulite Grit showing pyrite (black) replacement of detrital quartz (Q) and dolomite (D). Plane polarized, transmitted light. X 50.

PLATE 6



assayed 11.4 percent K_2O, as compared with the British average of 3.7 percent or with the world average of 3.2 percent for shales of widely varying ages."

Petrographic studies reveal no suggestion of these high potash values. Their abnormality, in terms of quantitative values, is further magnified by the unlikely stratigraphic situation in which they are found. Briefly reviewing the sedimentary succession, arkosic, subarkosic and orthoquartzitic sandstones overlying a planar unconformity grade upward into mostly orthoquartzitic sandstones. Above these sandstones are the shales and siltstones which contain the potash. Overlying these units is another thin orthoquartzitic horizon and finally, this clastic sequence is succeeded by a thick carbonate sequence. This progression from arkose to sub-arkose to orthoquartzite to shale to carbonates suggests a slowly transgressive marine depositional environment with an ever increasing textural and mineralogical maturity of the sediments. It is improbable that any geologist would predict that the shales and siltstones within this sequence would contain greater than 10 percent potash and the discovery of these anomalous values would be unlikely by any method other than the "shotgun" approach of chemically analysing any and all British black shales.

However, having been led to the problem by the Survey, it seemed worthwhile to investigate this aspect of the "Fucoid

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Beds" in conjunction with the present study. Chemical analyses were performed on samples of these horizons through the kind cooperation of Dr. E.L.P. Mercy and his staff. The analytical results provided overwhelming support for the earlier geochemical investigations of the Geological Survey. Table III presents the results of partial analyses performed on five samples taken as representative of the vertical sequence and one further sample to examine the lateral extent. Table IV presents the norms calculated from these analyses and Table V gives the norms recalculated for only the detrital fraction of the samples (subtracting the calcium, magnesium, and carbon dioxide which, in thin section, is obviously dolomite cement). Table VI quotes published analyses of some sedimentary rocks (Pettijohn, 1963) and shows the considerable disparity between these shales and "normal" sediments. A second series of analyses of only K_2O and Na_2O was performed to determine the vertical extent of the high potash values and to ascertain if the soda content of the sandstones and carbonates was proportionally high. The results of these further analyses are tabulated in Table VII.

The Atomic Energy Division of the Geological Survey also extended their investigations of the shales. In the <u>Summary of Progress Report for 1963</u> (Bowie, 1964), they reported that the potash-rich mineral was mainly adularia but partly muscovite and glauconite. An inquiry directed to the

TABLE III

Laboratory Report No. 60

October 6th., 1964

Partial analyses of samples from Fucoid Beds, Sutherland

	805/I-2	806/I-12	807/I-25	808/I-35	809/I-53	810/KI-20
SiO_2	27.56	45.15	23.15	57.64	62.03	39.53
TiO_2	0.36	0.49	0.22	0.60	0.28	0.49
Al_2O_3	6.72	10.20	4.19	12.05	5.63	9.62
Fe_2O_3	2.39	1.79	2.92	1.93	3.51	2.55
MgO	11.12	6.25	12.86	2.50	4.79	8.22
CaO	17.88	10.67	20.43	5.54	7.79	12.24
Na_2O	0.12	0.14	0.10	0.18	0.11	0.15
K_2O	5.08	7.09	3.44	9.77	3.39	7.30
CO_2	27.01	15.47	30.52	6.49	10.49	17.44
Tot.	98.24	97.25	97.83	96.70	98.02	97.54

E.L.P. Mercy

TABLE IV

Norms based on partial analyses of samples from Fucoid Beds

Sample numbers	I-2	I-12	I-25	I-35	I-53	KI-20
Quartz	6.19	14.60	9.06	17.73	46.28	9.25
Orthoclase	30.08	41.77	20.05	57.93	20.05	42.89
Albite	1.05	1.05	1.05	1.57	1.05	1.05
Anorthite	2.79	6.42	0.84	3.07	4.74	4.18
Magnesite	23.18	13.02	26.79	5.21	10.33	17.14
Calcite	30.80	16.66	36.06	8.69	12.12	20.20
Hematite	2.40	1.76	2.88	1.92	3.52	2.55
Rutile	0.32	0.48	0.24	0.56	0.28	0.49
CO_2	1.45	1.41	0.79	-	-	-
TOTALS	98.26%	97.17%	97.76%	96.68%	98.67%	97.75%

TABLE V

Partial analyses of samples from the Fucoid Beds

Recalculation of wt. % and norms of only the detrital components.

Weight Percents

	I-2	I-12	I-25	I-35	I-53	KI-20	Av.
SiO_2	65.26	69.61	60.08	70.15	82.76	66.28	70.35
TiO_2	0.85	0.75	0.65	0.73	0.37	0.82	0.69
Al_2O_3	15.91	15.73	12.32	14.66	7.51	16.13	13.71
Fe_2O_3	5.65	2.76	8.58	2.35	4.68	4.28	4.72
Na_2O	0.28	0.22	0.29	0.22	0.15	0.25	0.23
K_2O	12.03	10.93	10.11	11.88	4.52	12.24	10.28
	99.98%	100.00%	100.00%	99.99%	99.99%	100.00%	99.98%

Norms

	I-2	I-12	I-25	I-35	I-53	KI-20
Quartz	14.45	22.09	26.55	21.54	60.96	15.31
Orth.	70.23	63.21	58.76	69.98	26.41	71.00
Alb.	2.45	1.59	3.08	1.90	1.38	1.74
Anor.	6.51	9.72	2.46	3.71	6.24	6.92
Hem.	5.60	2.66	8.44	2.32	4.64	4.22
Rut.	0.75	0.73	0.70	0.68	0.37	0.81
	99.99%	100.00%	99.99%	100.01%	100.00%	100.00%

TABLE VI

PETTIJOHN, F.J., 1963. Chemical Composition of Sandstones - Excluding Carbonate and Volcanic Sands. Data of Geochem.; 6th Ed.: U.S. Geol. Surv. Prof. Pap. 440-S, 21 p.

	Mean composition of principal sandstone classes				Chem composition of principal sizes		
	Ortho-quartzite	Lithic ss.	Grey-wacke	Arkose	Sandst.	Silts.	Shale
SiO_2	95.4	66.1	66.7	77.1	78.66	69.96	58.38
TiO_2	0.2	0.3	0.6	0.3	0.25	0.59	0.65
Al_2O_3	1.1	8.1	13.5	8.7	4.78	10.52	15.47
Fe_2O_3	1.5	5.2	5.1	2.2	1.38	3.47	6.49
Na_2O	0.1	0.9	2.9	1.5	0.45	1.51	1.31
K_2O	0.2	1.3	2.0	2.8	1.32	2.30	3.25

TABLE VII

Laboratory Report No. 61

October 30th., 1964

		% Na_2O	% K_2O
820/P-315	Sandst.	0.05	1.36
821/P-105	Sandst.	0.03	0.84
822/Q-245	Sandst.	0.03	1.47
823/Q-75	Sandst.	0.06	3.22
824/O-I-133	Sandst.	0.13	6.78
825/O-I-76	Sandst.	0.05	0.61
826/K-I-40	Siltst.	0.17	8.70
827/O-I-52	Siltst.	0.18	7.83
828/I-81	Siltst.	0.13	6.16
829/I-70	Siltst.	0.07	1.00
830/D-10	Carb.	0.09	0.24
831/D-100	Carb.	0.08	0.05
832/D-331	Carb.	0.08	0.04

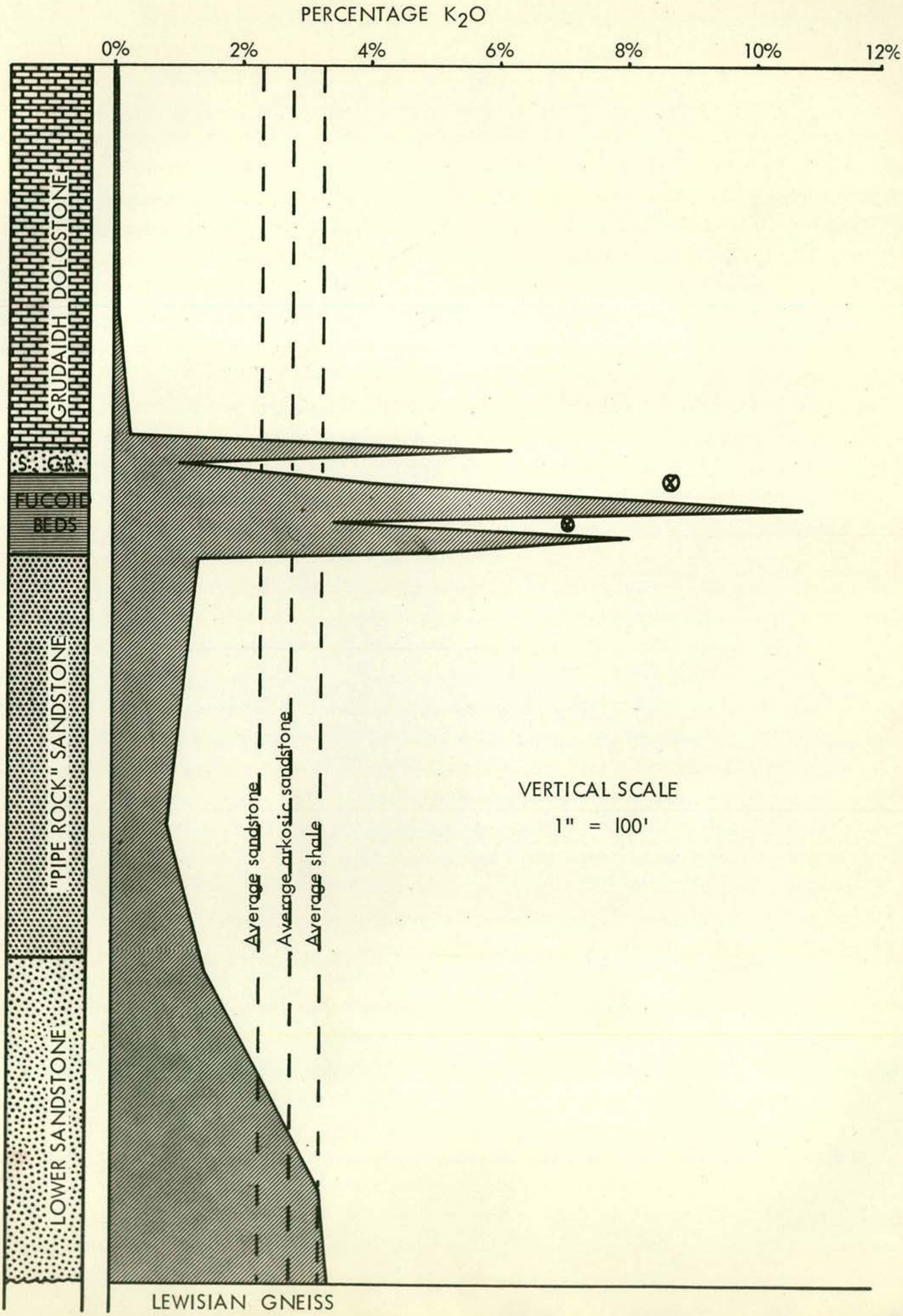
E.L.P. Mercy.

Geological Survey elicited only the reply that their information was soon to be published and could not be divulged prior to publication........

Figure VI shows graphically the percentages of K_2O plotted against the stratigraphic column at An t-Sron. The abrupt and remarkable increase in K_2O content at the base of the "Fucoid Beds" is conspicuously revealed by this diagram. The fluctuations of potash values within the "Fucoid Beds" merely reflect the fact that only six analyses are plotted. The results of two analyses of samples from a more southern locality are shown as circled crosses (Fig. VI) and suggest that further analyses would confirm the higher trend of the K_2O values. These various tabulations and graphical plots indicate the problems involved without, unfortunately, offering any clues to the possible causes of the anomalous potash content.

A tentative hypothesis to explain the K_2O values was derived largely through a process of eliminating unlikely explanations. To review briefly the possibilities: the potash must be either a primary detrital constituent of the sediment, a primary orthochemical constituent of the sediment, or an authigenic enrichment in potash. If the first possibility is considered, an insurmountable problem of provenance and/or mineralogical sorting is encountered for, as shown, the potash content is much higher than "normal" arkosic or clay-rich rocks. Some of the detrital fraction

FIGURE VI


PERCENTAGE K$_2$O
0%
2%
4%
6%
8%
10%
12%
GRUDAIDH DOLOSTONE
S. GR.
FUCOID BEDS
"PIPE ROCK" SANDSTONE
LOWER SANDSTONE
Average sandstone
Average arkosic sandstone
Average shale
VERTICAL SCALE
1" = 100'
LEWISIAN GNEISS

modes, in fact, approach the K_2O content of pure orthoclase.

The second possibility - an orthochemical precipitate - (possibly sylvite ?) would require an evaporitic environment. This is clearly incompatible with the stratigraphic associations and with the faunal content of the rocks.

If these arguments are accepted, they exclude a primary genesis for the potash content of these horizons.

The alternative to a primary genesis is, obviously, a secondary or authigenic enrichment of the potash content. Two mechanisms for a secondary enrichment seem possible: 1) selective replacement of non-potash minerals by a third mineral, or 2) introduction of a potash-rich mineral into the sediment either as a cement or as an authigenic replacement mineral. Considering the first possibility, the only apparent authigenic replacement which might conceivably accomplish a potash enrichment would be the preferential replacement of quartz (and not feldspar) by dolomitization. This explanation encounters impossible volumetric difficulties and the chemical analyses clearly indicate an inverse relationship between the amount of dolomite and the amount of potash.

The remaining possibility - that of potash enrichment through the introduction of some authigenic minerals - fortunately offers a tentative explanation. The recognition of this mechanism of enrichment is not, however, as simple as might be presupposed for there are no perceptible authigenic overgrowths on detrital feldspar grains to account for the high potash values. Staining of the thin sections with sodium cobaltinitrate (Rosenblum, 1956) reveals that the potash is contained in the fine-grained matrix.

Gruner and Thiel (1937) recorded similar high potash values in the Glenwood shale of Minnesota and concluded that since much of the feldspar they studies was extremely fine-grained, ($<$1/256 mm), it was doubtful that it could have withstood the vicissitudes of weathering to which primary detrital material is subjected. They also investigated some fine-grained silts (recent) derived from igneous and metamorphic source rocks and found them to be mainly quartzose.

> "It is much more likely that clayey material of colloidal dimensions and properties abstract potassium from solutions and form orthoclase under favourable conditions." (Gruner and Thiel, 1937)

Y. Baskin (1956) presents abundant data concerning authigenic feldspars and a noteworthy review of the subject. Most of the finely crystalline authigenic feldspars were noted by Baskin to possess the monoclinic symmetry of adularia (rather than the triclinic symmetry of microcline or orthoclase). Compositionally, the authigenic potash feldspars contain an average of 16.4 percent K_2O as compared with an average of 11.7 percent K_2O for potash feldspars from granites and syenites. Baskin estimates that authigenic feldspars rarely make up more than a few percent of the total sediment.

Daly (1917) describes a notable exception to this in rocks containing approximately 40 percent authigenic feldspar. Weiss (1954) recorded authigenic potash feldspars constituting nearly 70 percent of the total composition of Ordovician shales. The lack of evidence to support a hydrothermal or metamorphic origin led Baskin to conclude that the authigenic feldspars are probably a replacement or epigenetic phenomenon developed in argillaceous sediments during consolidation. Some of the evidence observed during the present study does not support Baskin's conclusion.

If the authigenic feldspars of the "Fucoid Beds" were introduced into the shales and siltstones contemporaneously with the authigenic overgrowths noted on the feldspar grains of the Eribol sandstones, a post-compaction genesis is indicated by the cross-cutting relationships of the overgrowths to strained and sutured quartz grains. A post-compaction genesis for the potash creates a source and transportation problem for the potash but this is perhaps not insurmountable. A possibly significant clue to the solution of this problem is perhaps unwittingly offered by Baskin (1956, p. 152) ....."All of the carbonate rocks studies

> that contained authigenic feldspars were partially or completely dolomitized or recrystallized. It is possible that the formation of the authigenic feldspars may have been contemporaneous with the dolomitization or recrystallization."

In itself, this statement is not particularly enlightening, for virtually every carbonate rock in existence has been partially or completely dolomitized or recrystallized. The association of dolomitization with authigenic feldspars, however, it also recorded by Tester and Atwater (1934) and in at least one of the environments discussed by Gruner and Thiel (1937).

The established replacive ability of dolomite leads to the possibility that some potash-rich mineral originally contained in the overlying limestone could perhaps have released potassium when replaced by dolomitization. The paragenetic timing of this is in harmony with the post-compaction, authigenic overgrowths on the feldspars of the Eribol sandstones. A simultaneous release of silica from a dolomitized clay mineral would also lend support to the hypothesis concerning the source of silica for pore filling in the "Serpulite Grit" sandstones.

The prime candidate for a potash-rich mineral which might reasonably have been contained in the limestones is illite (Grim, 1953). Analyses of 15 samples of illite (Deer, Howie, and Zussman, 1962) show an average potash content of 7.16 percent. The hypothesis would receive further support if a kaolinitic residue were left in the limestones but X-ray diffractometer analyses of the clay mineralogy show no kaolinite. This detracts somewhat from the hypothesis that the potash was derived through dolomitization of illitic limestones but possibly indicates only that a kaolinitic residue was not formed. The present clay content of the dolostones is mainly sericite with subordinate chlorite. It is not unlikely that the limestones may have originally contained a higher clay content and illite is the most probable of the clay minerals (Grim, 1953). Whether dolomitization of an illitic limestone did, or could, have produced the high potash values of the "Fucoid Beds" is largely speculative but, to this author, it offers the most acceptable tentative explanation for an extraordinary and perplexing geochemical problem. It also could, perhaps, offer a source of potash for other authigenic feldspars.

TUB SIZED - AIR DRIED

IV. DURNESS CARBONATES

A. GENERAL STATEMENT

The Durness carbonate rocks comprise the bulk of the Cambro-Ordovician sequence of north-west Scotland (see appendix). These carbonates rest with apparent conformity on the An t-Sron formation below and vary in exposed thickness from a few feet to somewhat in excess of 4000 feet, their exposed thickness being dependent on the stratigraphic level of the ubiquitous truncation by the Moine Thrust plane or other Caledonian structures. At no place can a depositional top to these carbonates be observed. The most complete stratigraphic sequence is available in the Eribol and Durness regions and a major portion of the present study is based upon a composite stratigraphic sequence from these two areas.

From the early writings by MacCulloch (1814) to the recent review by Phemister (1960), a casual student of the literature would, with few exceptions, conclude that the bulk of the sequence was composed of limestone. The common application of the term "Durness limestone" to the rocks, stemming from the general ill-preparedness of field geologists to distinguish between limestone and dolostone, is undoubtedly responsible for this misconception. To dismiss this compositional difference as insignificant is inexcusable. The sequence includes limestones, dolomitic limestones, calcareous dolostones, and pure dolostones. The contact relationships between these compositional variations may be either gradational, sharply defined horizons, or as interpenetrating mottled networks.

Peach.et. al. (1907) introduced seven stratigraphic subdivisions of the carbonate succession, divided on the basis of both primary and diagenetic features. Some of their subdivisions are insufficiently distinct or well defined to permit positive recognition and separation in the field. This is especially true of the upper units but attempting to alter deeply ingrained stratigraphic nomenclature is considered impractical. In this thesis, the term "member" is, however, substituted for "groups". Most of the descriptions of the various "groups" offered by Peach et. al., do not afford precise demarcation of the boundaries of the units. Also in their descriptions, many of the more conspicuous features of the rocks are unmentioned and one often has cause to wonder if their (Peach et. al., 1898) mapped "groups" coincide with their described "Groups" (Peach et. al. 1907). Closely following the six-inch geological map, the present writer found in several instances that contacts between subdivisions were drawn at localities exhibiting no significant lithologic change and found that an arbitrary boundary selection was necessary in order to utilize their "group" names. Nearly all of this difficulty, however, arises in the upper three members. The lower four members, the Grudadidh member, the Eilean Dubh member, the Sailmhor member, and the

Sangamore member, are all reasonably distinct lithologic units although different investigators might justifiably select slightly different boundaries without more precise definitions. The Balnakiel, Croisphuill and Durine members while possessing certain distinctive horizons, either merge imperceptibly with one another or their boundaries are concealed. The Croisphuill and Durine members are exposed only in the Durness outlier and their distinction there is rendered difficult by fracturing, contortion, and poor exposures.

B. PETROGRAPHY

Three major petrographic end members may be visualized for the rocks of the Durness carbonates: 1) Dolostone, 2) Limestone, and 3) Chert. These three rock types and their admixtures such as dolomitic limestones, calcareous dolostones, cherty limestones, cherty dolostones etc. account for nearly the total of the sequence. The only remaining litholgies are very minor horizons of arenaceous (generally quartzose) or silty limestones and dolostones. Clay size detrital and or organic impurities might be considered as a fourth compositional end member but is not a lithologic endpoint since no rocks are found with clays as a predominant constituent. Semi-micro quantitative determinations of the insoluble residues (in dilute (25%) hydrochloric acid) were made on

21 chert-free samples. The results of these determinations (Table IX) show the range in the insoluble content of the carbonates from 0.61 percent to 6.87 percent. The average insoluble residue of these representative samples comprises 2.12 percent of the rock but the stratigraphic distribution of the variations in the insoluble content is extremely erratic and no consistency of pattern evolves from which paleo-environmental interpretations might be drawn.

Attempting to describe or discuss the more detailed aspects of the petrography of the carbonate sequence is, in a sense, rather frustrating because many of the primary features and textures of the rocks have been obscured through diagenesis and often only ghost structures remain (Plate 7-A). The interpretation of the "ghost" structures is often highly subjective and yet, these structures represent an important aspect of the history of the rocks.

Unfortunately, no satisfactory petrographic classification scheme has yet been devised for diagenetically altered carbonate rocks. There are such a multitude of variables in the degrees and patterns of recrystallization and authigenesis that any classification permitting a meaningful description and grouping of the multitude of lithologies would necessarily be either so general or so cumbersome as to be of negligible value. It is, however, possible to consider the petrographic classifications of the unaltered limestones (and their genetic implications) and then to discuss the range

TABLE IX

SEMI-MICRO QUANTITATIVE ANALYSES OF INSOLUBLE RESIDUES

Sample *	Wt. % insol. residue
D-10	5.51
D-50	1.79
D-100	2.92
D-150	1.75
D-331	0.61
D-372	1.84
D-441	0.86
D-446	4.76
D-1002	5.31
D-1356	1.25
D-1480	3.28
D-1582	3.77
D-1808	2.02
D-2212	2.21
D-2647	1.58
D-2827	0.63
D-3019	0.90
D-3297	6.87
D-3491	1.90
D-3947	2.10
D-4247	2.72

* All samples are taken from the Durness carbonates. Sample numbers refer to stratigraphic level (in feet) above the base of the carbonate sequence.

and extent of the diagenetic alterations.

1. Limestones

Of the several classifications which have been proposed (Ham, 1962), R.L. Folk's spectral subdivision of limestone types (1962) proves most useful for the limestones of the Durness sequence. A few categories of minor importance based on detrital constituents have been added to Folk's system through the use of modifying terms. The limestone types of the Durness sequence are, in order of decreasing abundance: micrites, intramicrites, microsparites (recrystallized micrites), intramicrosparites, silty micrites, silty intramicrites, quartzose arenaceous micrites, quartzose arenaceous intramicrites, oomicrites, oosparites, and, very rarely, pelmicrites. Breccias are found locally throughout the sequence but are the result of much later tectonics and are not considered here as separate petrographic types.

Many of the micrites are finely laminated and a few contain body fossils or trace fossils. The faunal content of the limestones is never sufficient to warrant the term biomicrite except possibly with the carefully selected hand-size sample. Some of the thinly laminated carbonates are almost certainly algal in origin. These might be classified as biomicrites but are herein included under the term micrites since microscopically they do not appear to contain any recognizable biogenic particles.

The vast majority of the carbonate sequence is now and/or probably was microcrystalline at the time of deposition and would fall in the category of the Type III limestones (the Microcrystalline rocks) (after Folk, 1962). In terms of their genesis, R.L. Folk has proposed that the Type III limestones indicate ....

> ".......both a rapid rate of precipitation of micro-crystalline ooze, together with lack of persistent strong currents. Texturally, they correspond with the claystones among the terrigenous rocks; most form in very shallow, sheltered lagoon areas, or on broad submerged shelves of little relief and moderate depth where wave action is cut off by the very width of the shelf. Some may also form in deeper offshore areas."

The presence of the stromatolitic algal structures common throughout much of the sequence supports the concept of deposition on broad, shallowly submerged shelves of low relief.

2. Dolostones

The effects of dolomitization can be demonstrated in all degrees from the occasional authigenic dolomite euhedra speckled through limestones to totally dolomitized rocks. This complete gradational sequence strongly supports an authigenic interpretation for all of the dolostones but does not preclude the possibility of some of them being primary dolomite deposits.

Dolomitization generally obscures the details of original textures by its tendency to increase the crystal "grain size". Many of the finely laminated dolostones when examined microscopically exhibit a relatively homogeneous medium-crystalline texture and the laminae, conspicuous on weathered exposures are barely, if at all, discernible. This suggests that, although generally increasing the textural coarseness, dolomitization was possibly influenced sufficiently by origianl laminations of impurities to have caused subtle variations in dolomite textures that are now revealed by weathering. It is perhaps of interest that etching of cut surfaces of the finely laminated dolostones with hydrochloric acid fails to expose the laminae so obvious on weathered surfaces. The apparent lack of laminar structure in thin section compared with similarly laminated limestones is caused by the transection of the original laminae by dolomite crystals and supports an authigenic or recrystallization rather than primary origin for the dolostones. A third argument favouring a secondary origin for the dolostones will be presented in detail in subsequent discussions of diagenesis but previewed briefly, it is based on textural evidence within chert nodules in the dolostones which indicates that the chert has been replaced by the dolomite. Primary bedding structures which can be traced across the boundaries of the chert nodules (Plate 7-B) eliminate the possibility of the cherts being primary. Therefore, if the dolomite is replacive to authigenic cherts, it must either be authigenic itself or the replacement could possibly have resulted through late recrystallization of primary dolomite under altered geochemical conditions that were unfavourable to the authigenic chert nodules. Silicification followed by dolomitization seems a more likely explanation of the relationships.

In addition to the thinly-bedded or massive dolostones, there are also dolostones exhibiting distinct similarities to many of the previously noted limestone types such as silty and arenaceous dolostones, oolitic dolostones, and dolostones showing ghosts of intraclasts and fossil debris and trace fossils. Fossiliferous dolostones are a rare phenomenon unless the algal structures in dolostones are considered in this category.

A lithological feature peculiar to the dolomitized units of the succession is mottling. Mottled textures are displayed by both the dolomitic limestones and the dolostones (see stratigraphic descriptions in appendix). The mottled dolomitic limestones of the Croisphuill member are characterized by an anastomosing network of interpenetrating dolomitic and calcitic lithologies arranged in irregular streaks, patches, and bands (Plate 7-C). The two lithologies are usually sharply distinct in their contact relationships. The dolomitic lithology is generally darker in colour and more coarsely crystalline than the calcitic portions. The two separated lithologies give the mottled rock a striking appearance in outcrop. Their relative difference in resistance to chemical weathering produces an extremely rough surface with the dolomitic portions standing out in positive relief as much as 3/4 inches beyond the calcareous portions (Plate 7-D). This pattern of mottling was attributed by Peach et. al. (1907) to the actions of burrowing organisms. A very careful analysis of the patterns reveals however three dimensional relationships that cannot be attributed to burrowing animals. The vertical gradation of the mottled lithologies into unmottled totally dolomitic (Plate 7-E) or totally calcareous units further discourages the burrow hypothesis. Dolomitization in these mottled limestones appears to have followed no fixed pattern but to have proceeded randomly from centers of indeterminate origin.

Mottling is also evident in the totally dolomitic rocks of the Grudaidh and Sailmhor members, the latter being the lithology described as the "Leopard Stone" by Peach et. al. (1907). The mottling of these dolostones expresses variations in grain size and colour (lighter areas being more coarsely crystalline) and is perceived as an extension of the partial dolomitization of the mottled dolomitic limestones. The pattern of mottling in the dolostones is consistent in size, distribution and frequency with the mottled pattern in the dolomitic limestones. Although nowhere in the Durness carbonates was the gradation from mottled dolomitic limestones to mottled dolostones observed, similar patterns of mottling and the complete gradation between the two lithologies have been reported from the Beekmantown rocks of Maryland (Sando, 1957).

The mottling of the dolomitic limestones and dolostones, although almost certainly not caused by the trails and burrows of organisms, do apparently follow some primary feature within the original sediment. //The mottled dolomitic limestones of the Croisphuill member at Durness contain a characteristic molluscan fauna which can be correlated with near certainty over the hundred mile distance between Durness and the Ben Suardal mottled dolomitic limestones on the Isle of Skye. This strongly supports a primary control over the cause of the mottling in this mollusk-bearing horizon.

The argument for primary controls influencing the patterns of mottling is further strengthened by a conspicuous cyclic or rhythmic pattern displayed in the mottling of the Sailmhor dolostones (the "Leopard Stone"). The cycle which is repeated roughly half a dozen times in the Sailmhor member consists of a chert, thinly bedded, light grey, finely crystalline dolostone unit 1 - 3 feet thick, overlain by a very dark, highly mottled, massive; coarsely crystalline dolostone unit 2 - 4 feet thick, in turn overlain by a medium grey, moderately mottled, coarsely crystalline dolostone unit 6 - 10 feet thick. It is very doubtful that this cyclic pattern which is reasonably consistent in its lithologies and their relative stratigraphic order could emerge from the diagenesis of texturally and mineralogically homogeneous sediments or rocks. The cyclic pattern of mottling, colour, and bedding features must surely reflect some primary variation in the carbonate sediments that has later been exploited by diagenesis.

Sujkowski (1958) suggested that the cause of diagnetic reactions is the chemical instability of heterogeneous sediments which tend toward an "unmixing" to more stable mineral suites. If this hypothesis is applied to the patterns of dolomitization in the Durness carbonates, the mottling of the dolomitic limestones might be conceived as the haphazard segregation or "unmixing" of magnesium from magnesian limestones which possessed a relatively low diffusion porosity. If only a limited migration of the ions was possible, "unmixing" could explain the mottled pattern of the dolomitic limestones. Extending this hypothesis slightly leads to the possible interpretation that dolomitization which is expressed as a uniform speckling of limestones with microscopic dolomite euhedra is the result of an "unmixing" in a sediment or rock with a very low diffusion porosity. If the opposite extreme is envisaged (with a high diffusion porosity) it might explain the common segregation of limestone and dolostone into discrete interbedded units. This hypothesis would seem to offer a more reasonable and tenable explanation for the observable gradations between mottled and unmottled horizons than to invoke algal growth patterns or patterns produced by burrowing animals.

The "unmixing" hypothesis while offering a possible explanation for the mottling of the dolomitic limestones is obviously inadequate to explain two periods of dolomitization necessary to produce the mottled pattern of the pure dolostones of the Sailmhor and Grudaidh members. Either a separate mechanism must be invoked for the second stage of dolomitization or a modification of the "unmixing" concept is required. Beales (1953) has proposed that a similar pattern of mottling in the Palliser limestones of Alberta has resulted from dolomitization of magnesian limestones which probably commenced from "more susceptible centers", possibly resulting in ...."a trigger effect in that the

> effective diffusion porosity of these areas was increased, thus facilitating the diffusion of the magnesian solutions which first affected the centers and then spread into the surrounding rock."

Although phrased as a circular argument, Beales' proposed mechanism suggests a possible mechanism for two phases of dolomitization. Substitution of Sujkowski's "unmixing" for Beales' "more susceptible centers" would perhaps improve the hypothesis regarding the initial phase of dolomitization. If, however, the initial dolomitization occurs as a volume for volume replacement and does not increase the effective porosity, Beales' suggestions collapses completely.

The mottling of the dolostones may also be explained simply as two independent phases of dolomitization occurring at different times or at widely different rates. If the initial phase occurred due to the local "unmixing" of metastable sedimentary mixtures and limited by a low diffusion porosity, it could be followed by a later slow permeation of the rocks by magnesium-bearing solutions. Separate periods of dolomitization under differing conditions and rates of diagenesis could easily produce the variations in colour and crystal textures that now express the mottling. This hypothesis is favoured by the lack of any evidence in the partly dolomitized limestones that the early dolomitization has increased the porosity of the rocks.

Oolitic dolostones or partly dolomitized oolitic limestones occur at five separate horizons in three of the stratigraphic members of the Durness carbonates. The oolitic textures have been invaluable in establishing multiple phases of dolomitization during the diagenetic history of the carbonates. To avoid duplication of discussion, the detailed description and interpretations of the oolitic dolostones will be deferred until later discussions of diagenesis in order that the dolomitization in these rocks may be considered relative to the other diagenetic events.

3. Cherts

Silicification features in the Durness carbonates assume a wide variety of forms ranging from sparsely distributed microscopic authigenic quartz crystals to beds of nearly pure chert up to six feet thick. Most of the chert occurs in the upper five members of the formation as thin beds or randomly distributed subspherical to discoidal nodules. The cherts may be white (novaculite), black (flint), or reddish (jasperoid) in colour.

Throughout the following discussions, the term chert is used, unless qualified, as a collective term to describe all of the cryptocrystalline, microcrystalline and occasionally macrocyrstalline varieties of authigenic silica which occur throughout the carbonate sequence.

The cherts of the Durness carbonates may be divided into three distinct textural end members of: 1) microcrystalline quartz, 2) fibrous, chalcedonic quartz, and 3) "drusy" quartz mosaics. Folk and Weaver (1952) have interpreted the probable genesis of these chert types as follows:

> "Microcrystalline quartz results when crystal growth begins at very numerous, closely spaced centers

arranged in a three dimensional array. ..... The individual microcrystals, in random orientation, grow in all directions until they meet the advancing edges of adjacent microcrystals. The rate of formation presumably governs the closeness of the spacing, and this in turn determines the grain size of the chert"

"Chalcedonic quartz, on the other hand, starts formation from only a few centers of crystallization, and because of the limited interference, gives rise to optically continuous fibrous structures. The most favorable conditions for the formation of chalcedonic quartz occurs when crystallization begins at a few centers spaced along a surface, in which case outward growth is unhampered."

"In addition to the micro-forms of quartz, chert nodules may also contain ordinary quartz of larger crystal size, showing straight extinction and normal refractive index. This "drusy" quartz usually forms the final deposit in the center of cavity fillings, while chalcedonic quartz lines the walls. This is probably due to a decreasing rate of precipitation because of the diminishing rate of supply of solution as consolidation proceeds."

The literature regarding the "chert problem" is extensive and a comprehensive survey of this topic alone would undoubtedly exceed the total volume of this thesis and certainly exceed the intentions of this writer. Pettijohn (1957) has offered a perceptive, although somewhat oversimplified, review of the literature prior to 1955. However, many of the significant advances in the understanding of the chemistry and physics of silica diagenesis have appeared since 1955. Recent reviews of the topic are afforded by the S.E.P.M. Symposium volume - Silica in Sediments (Ireland, 1959) and by Siever (1962). The following brief review of the problems and of some of the more recent advances is not intended to be comprehensive but is presented merely to establish some recent views upon which subsequent discussions can be based.

The "chert problem" may be summarized as the inadequate understanding of: 1) the sources of silica, 2) the mechanisms and distances of transportation, and 3) the causes and mechanisms responsible for the precipitation of authigenic silica. Recapitulating some of the proposed answers to the first question, sources which have been suggested for authigenic silica are: 1) Inorganic precipitation from sea water, 2) Metasomatic fluids, 3) Recrystallization of silica-rich volcanic deposits, 4) Biogenic precipitation of amorphous silica from sea water (by radiolaria, diatoms, and sponges) which may either form cherts directly or dissolve and reprecipitate within sediments to produce chert, 5) Pressure solution of detrital quartz, 6) Carbonate replacement of detrital quartz, 7) Clay mineral diagenesis (illite or montmorillonite conversion to kaolinite + silica), or 8) Dolomite replacement of clay minerals.

Sound arguments have been advanced which exclude the first three of these proposed origins for the majority of cherts (Siever, 1962). It is likely, however, that cherts are polygenetic and that the other sources have perhaps all contributed silica in varying proportions to various cherts depending upon the local conditions.

The last suggested source - dolomite replacement of clay minerals - is, to the knowledge of this writer, a previously unexplored possibility. There is poor agreement concerning the role of clay minerals in diagenetic reactions. Sujkowski (1958) maintained that clay minerals retard the diagenetic processes while Heald (1955) believed that the presence of clay minerals was essential as a catalyst to certain diagenetic pressure solution reactions. Illite (perhaps the most likely clay mineral to be found in carbonates) could conceivably serve as a source of both magnesium for dolomitization and of silica for the formation of cherts (and possibly potash for authigenic feldspars). Dolomite replacement of clay minerals during diagenesis of limestones is certainly within the realm of possibility and is perhaps rather probable as a replacement reaction. Dolomite replacement of clay minerals is not proposed as the only source of silica, for in many instances, as in the Durness sequence, silicification can be demonstrated to antedate dolomitization. There are also innumerable examples of chert nodules within undolomitized argillaceous limestones. Illite may be more important as a source of magnesium in dolomitization and incidently contribute silica for the formation of cherts and potassium for the formation of authigenic feldspars.

Although not universally accepted, the problems regarding the movement of silica have been rather well resolved as true solution transport in the form of H_4SiO_4 (Krauskopf, 1959). Silica solubility is essentially independent of pH and Eh in the normal geological environments (between pH 2 - 9) but is markedly influenced by both temperature and pressure. Crystalline quartz, when subjected to pressure solution processes, may become as soluble as amorphous silica (Siever, 1962). The solubility of amorphous silica (100 - 140 ppm. in sea water at 25°C) far exceeds the normal silica concentration of sea water (1 - 4 ppm.) probably largely due to the removal of silica from sea water by organisms (Krauskopf, 1959).

The distance which silica may move before being precipitated as chert nodules or chert beds is still an unknown factor. Local sources within the sediment (as by Sujkowski's unmixing) are totally inadequate to explain some bedded cherts but may be important in nodular cherts. The abundant petrographic evidence indicating the replacement nature of many cherts necessitates the concept of considerable mobility of authigenic silica through the sediments or rocks.

> "Most chert nodules associated with marine limestones are probably post-consolidation, epigene replacement deposits". (Pettijohn, 1957).

In certain instances, as in the Durness carbonates, it may, however, be demonstrated that the chert bodies have affected the compaction of the surrounding sediments and therefore their emplacement and consolidation must have occurred sufficiently early to precede the total loss of pore space by the sediment.

The cause and mechanics of silica precipitation are even less well understood than the source and transfer problems. The microcrystalline character of cherts, the occurrence of water-filled cavities (Folk and Weaver, 1952), and the suggestion that amorphous silica may exist between the fibres of chalcedony (Correns and Nagelschmidt, 1933) all suggest that cherts are originally precipitated as amorphous silica rather than quartz (Siever, 1962). Alternatively, the silica may precipitate directly as microcrystalline quartz due to numerous separate nuclei from which growth commenced. Where cherts occur within carbonate rocks, it is impossible to separate the problems regarding silica precipitation from the allied problem of the simultaneous removal of carbonate. The solution and precipitation of carbonates are controlled by pH, the activity of CO_2, Ca^{++}, and CO_3^{--}, pressure, and temperature. The factors controlling silica precipitation and solution are quite different and the interplay of the two systems is poorly understood.

The early concept expounded by Correns (1950) of silica precipitation and carbonate solution at low pH values or silica solution and carbonate precipitation at higher pH values is no longer considered valid (Alexander, Heston, and Iler, 1954; Krauskopf, 1959; Baas-Becking et. al., 1960; Siever, 1962; and others). A further complexity of the carbonate-silica equilibrium has been observed during the present study wherein petrographic textural evidence suggests that calcite and chert can be replaced simultaneously and with apparent equal facility by dolomite (Plate 7-F). The implication of this evidence is that a physico-chemical environment is possible wherein calcite and chert are simultaneously and equally susceptible to dolomitization. This writer is unable to speculate as to the conditions of temperature, pressure, pH, etc, that could permit this replacement to occur. The replacement of one mineral by another must, in all instances, indicate either a chemical or a physical disequilibrium of some sort, but considerable further work is needed before the conditions of disequilibrium, metastability, and stability between silica and the carbonates will be understood. The multiple reversibility of

PLATE 7

A. (D-1582) Photomicrograph of a medium crystalline dolostone showing a "ghost" structure of a replaced gastropod fragment. Plane polarized, transmitted light. X 30.

B. Primary bedding and intraclast structures in dolostone (lower left of photograph) which may be traced across the dolostone-chert boundary, (the chert is light coloured) thus supporting an authigenic origin for the chert.

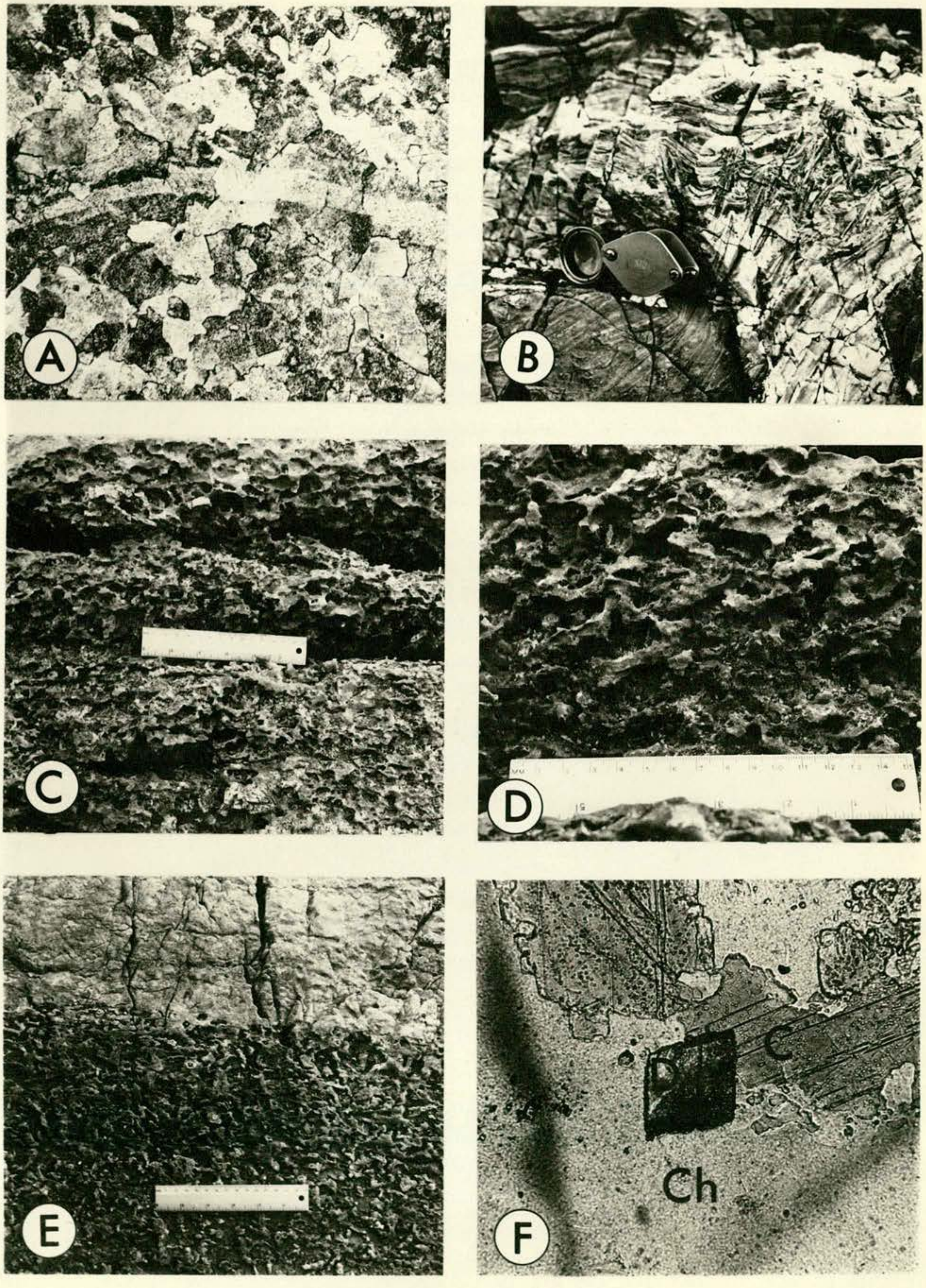
C. Mottled dolomitic limestone showing the extremely rough weathered surface and crude bedding of the mottled pattern.

D. Mottled dolomitic limestone demonstrating the high surface relief produced by differential weathering of the dolomitic lithology (the raised portions) relative to the calcareous lithology (the sunken portions).

E. Mottled dolomitic limestone passing upward into an unmottled dolostone.

F. (D-1703) Photomicrograph showing the cross-cutting textural relationships in some cherts wherein calcite (C) and chert (Ch) are cut across by a later euhedral dolomite crystal (D). Plane polarized transmitted light. X 100.

PLATE 7



replacements in the chert-carbonate system (Walker, 1962; Swett, 1964) suggests that the controls for replacements are probably rather delicately balanced near equilibrium rather than being widely separated extremes of physico-chemical environments. The multiple replacement reversals also detract from the credibility of the "unmixing" hypothesis (Sujkowski, 1958), not only in the chert-carbonate system but also in the calcite-dolomite system (Shearman et. al., 1961; Swett, 1964). Chemical reactions do not possess inertia and thus never extend beyond the equilibrium state. Hence, no pendulum effect is possible and the reversibility of replacements must reflect some controls other than simple chemical unmixing of unstable mixtures.

A parameter of cherts which investigators have almost completely neglected is the external geometry of chert bodies. Although cherts assume a wide variety of shapes, there are certain common denominators and the geometry of chert bodies may perhaps be considered in terms of three end members: 1) nodular cherts, 2) irregular cherts, and 3) chert beds. "Nodular cherts" are not necessarily spherical or elliptical but are distinguished by their "smooth" external forms and a distinct boundary. "Irregular cherts" are rough and angular in their external geometry, often with branching irregular appendages. "Chert beds" are characterized chiefly by their lateral continuity and relatively distinct upper and lower contact boundaries. Nodular cherts grade imperceptibly into chert beds but irregular cherts seldom occur as continuous units.

If the conclusions of Folk and Weaver (1952) regarding the rate of precipitation relative to grain size and textures are valid, perhaps some genetic interpretations for various chert bodies can be extrapolated from their textures. It seems worthwhile to consider the shape of chert bodies not only with respect to their textures (and rate of precipitation) but also with respect to the "unmixing" hypothesis of Sujkowski. Certainly the pattern displayed by nodular cherts is very reminiscent of a three-dimensional unmixing of two highly viscous, immiscible fluids. The geometry of nodular cherts, although often taking weird forms, is almost invariably economical in terms of space occupied per volume of silica. It would appear as though the silica had a high "surface tension" relative to the carbonates and was being excluded or rejected by the carbonates. This concept of an exclusion or rejection of the silica by its carbonate host perhaps explains the random but rather evenly spaced distribution of chert nodules throughout some limestones better than an hypothesis of silica attraction to indeterminate "nuclei" to form nodules. If a random and even distribution of chert nodules is to be explained as the attraction of silica to "centers" of precipitation, the problem of why the "centers" should so often be randomly and evenly distributed is encountered. An exclusion or rejection of silica from the host rock solves this distribution problem. (The exclusion might be compared to the random segregation of water droplets on an oiled surface.)

Nodular cherts, in general, which exhibit the most space-saving geometry are commonly the most finely crystalline, at least within this writer's observations. This observation applies both to thin beds of chert which are more sharply separated from their carbonate host in proportion to decreasing crystal size, and to chert nodules which tend to be more rounded and sharply bounded in proportion to a decrease in crystal size. If the genetic interpretations proposed by Folk and Weaver (1952) are applied to these geometric and textural variations, it suggests that the more sharply defined beds or the more rounded chert nodules have grown more rapidly thereby producing their more finely crystalline textures. More rapid growth suggests either a more rapid accumulation or precipitation of silica and thus perhaps reflects a greater permeability of the host rocks. Conversely, more coarsely crystalline cherts tend to be more irregularly shaped and may reflect a slower accumulation and precipitation of silica, perhaps due to a low permeability of the host rocks.

Some of the other geometrical features of chert bodies are far more perplexing. There are "shell structure" nodules (Plate 8-A) and "balloon-shaped" nodules (Plate 8-B). The shell structure nodules are generally subspherical to ellipsoidal "shells" of chert with or without a nucleus but with a carbonate-filled interior. These shell structure nodules must in some way reflect the precipitation of silica at one or more geochemical "boundaries" as the silica-bearing solutions migrated toward a "center" or alternatively were percolating through the carbonates. Very rarely a "shell structure" nodule may possess two or three concentric shells. Bedding structures which can often be traced through both the concentric chert and carbonate layers confirm their post-depositional origin. These "shell structures" of chert are extremely difficult to reconcile with a silica "exlusion" or "unmixing" hypothesis.

While creating certain problems, the "shell structure" cherts may help to explain why some chert nodules appear to "grow" inward and display an increasing crystal size and convergent chalcedonic pattern toward their centers. Other cherts (generally more irregularly shaped) appear to "grow" outward with crystal size increasing toward their margins. The "shell structure" nodules may represent the initial phase of an incomplete "inward growing" nodule.

Two separate horizons (D-891, D-940) in the Sailmhor dolostone member exhibit unusual "balloon" shaped chert nodules (Plate 8-C & 8-D). The balloon shaped nodules generally measure 1 - 1½ feet both in diameter and in height. They characteristically possess bedding structure continuous with the adjacent dolomite which internally is deflected downward in the centers of the nodules. In some instances the "balloon-shaped" nodules are arranged in a vertical stacking pattern with several nodules resting one atop another. Their shape, distribution arrangement, and internal characteristics are unique to these two horizons. A problem concerning the genesis of these unusual "balloon shaped" nodules arises as to whether the downwarped or collapse structures in their centers are the cause or the result of silicification at these locations. A tentative hypothesis to explain these unusual nodules depends upon the presence of cavernous horizons which occur immediately beneath each of the nodular chert horizons. It is suggested that silicification has selectively replaced zones of ?permeability or ?impermeability caused by the structures having collapsed downward into the underlying cavernous horizons. The siliceous solutions responsible for the nodules may have been migrating laterally through the bedded dolostones or could perhaps have migrated upward from the cavernous horizon. The occurrence, albeit rare, of non-silicified collapse structures argues against the probability that the silicification is responsible, in any way, for the downward deflection of the bedding. Early formation of both the collapse structures and the chert nodules is indicated by the effect which the nodules have exerted on the compaction of the overlying and underlying strata (Plate 8-E).

Another intriguing problem concerning the cherts is the occurrence of mottled chert nodules within the Sailmhor mottled dolostones. This, at first, appears inconsistent with the previously interpreted trends of diagenesis by its apparent contradiction of petrographic evidence which suggests that silicification has preceded dolomitization. If the mottling of the dolostones is to be attributed to dolomitization, it would seem impossible for the cherts to retain this mottling pattern unless the dolomitization had antedated silicification or had followed a pre-existing mottled pattern in the limestones. A possible solution to this dilemma may be that the mottled cherts which occur in the mottled dolostones reflect silicification that occured after dolomitization had produced mottled dolomitic limestones but prior to a second stage of dolomitization which converted the mottled dolomitic limestones to mottled dolostones. Field and petrographic evidence indicates that the dark portions of the mottled cherts were most probably the early dolomitized portions of the pre-silicification, mottled dolomitic limestones. The darker coloured portions contain a considerable quantity of relic dolomite inclusions. The lighter coloured chert is more finely crystalline and very pure and possibly indicates a greater susceptibility of the calcareous portions to silicification both by its lack of incluisions and by its finer texture. This would perhaps be consistent with a prediction of the relative ease of silicification based on the greater susceptibility of calcite (vs. dolomite) to solution and to pH variations.

Occasionally, in the Durness carbonates, chert nodules or beds are encountered which exhibit a very lacy or perforated appearing surface in outcrop. Close examination (with a hand lens) reveals that the holes in the weathered chert surface commonly show a nearly perfect rhombic shape. This distinctive shape persuasively argues that these cherts have suffered partial dolomitization. Removal of the dolomite rhombohedra by preferential weathering has left the lacy appearing cherts.

C. GEOCHEMICAL STUDIES

Chemical analyses of ancient carbonate rocks contribute only marginally to the understanding of their origin, history, or textures and their chief value lies in commercial evaluations of carbonate deposits. Diagenetic alterations in irregularly-distributed patterns profoundly affect the compositions of the Durness rocks and render statistically reliable and meaningful compositional sampling extremely difficult. For these reasons, no analytical work has been contemplated in connection with this study.

For the reader's convenience, the published analytical data on the Durness carbonates compiled by the Mineral Resources Panel and the Chemical Subcommittee of the Scottish Council (A.S. Butler, et. al., 1954) are reviewed in Table X. All of the values shown are well within the ranges of published

TABLE X

Chemical analyses of the Durness carbonates

(after Butler et. al., 1954)

Sample location and number of analyses included.	SiO_2	$Al_2O_3 + Fe_2O_3 + Fe_3O_4$	CaO	MgO	Loss on ignition
Durness outlier					
Gp. II (9)	15.24	4.79	25.41	16.06	37.51
Gp. III (7)	2.57	0.74	30.27	20.63	45.78
Drochaid Mhor (8)	2.32	2.40	30.56	19.29	45.22
Sarsgrum (24)	2.12	0.99	31.06	20.43	45.75
Loch Eribol					
Inbhirean (23)	2.27	1.64	31.58	19.15	45.47
An Druim (10)	1.34	1.47	30.93	20.45	46.03
Inchnadamph					
Calda House (20)	2.27	1.20	30.27	20.52	45.67
Stronechrubie (5)	2.25	1.17	30.84	20.00	45.59
Elphin (8)	3.49	3.11	29.58	19.12	44.32
Loch Kishorn (60)	2.69	1.62	29.77	20.31	45.35

"average" compositions of limestones and dolostones (Pettijohn, 1957; Graf, 1960).

From a petrographic viewpoint, the only figure of interest in Table X is the 15.24 percent silica content of the "Group II" dolostones at Balnakiel. This is a remarkably high silica content relative to other horizons since the Group II (Eilean Dubh member) dolostones exhibit little or no conspicuous chert. This silica content is quite inconsistent with the insoluble residue analyses performed with the present study. This slightly anomalous result can possibly be explained as a localized finely disseminated distribution of detrital or authigenic silica in the thinly bedded (?impermeable) dolostones which was impossible for the sampling geologist to avoid. In other, more permeable horizons where the authigenic silica could migrate and form chert nodules, the sampling geologist would probably avoid these concentrations of silica if the aim was to analyse the carbonates. Thus the differences in the published compositions may not be as significant as they appear.

D. SEDIMENTARY STRUCTURES

Several distinctive types of sedimentary structures may be observed in the Durness carbonates, some of which have been mentioned in the foregoing passages. The sedimentary structures afford many valuable clues to the environment in which the carbonates were deposited. Peach et. al., (1907) in their discussion of the conditions of deposition state:

> "After the deposition of the Serpulite Grit, the stratigraphical chronicle shows that the depression of the sea-floor became more persistent and quicker in pace, and that subsidence was considerably more pronounced than deposition."

An evaluation of the sedimentary structures of the Durness carbonates does not support this conclusion and, in fact, the evidence suggests that the carbonates were deposited in water depths possibly as shallow as, if not more shallow than, the "Fucoid Beds" and "Serpulite Grit" members.

Bedding or stratification is quite pronounced throughout most of the carbonate sequence. The bedding may be characterized by very fine bedding laminae as in the Eilean Dubh member which exhibits bedding ranging from a few millimeters to a few centimeters in thickness, flaggy, as in the Balnakiel member with units 1 to 6 inches thick, "unit bedded" in distinct unit beds 1 foot to 5 feet thick, or finally, some horizons are relatively massive with individual sedimentation units difficult to distinguish as displayed in some of the mottled dolostones of the "Grudaidh" member. Seldom is it possible to ascertain the exact nature of the lithologic break that has produced the bedding units. Occassionally, it is possible to detect variations in crystalline textures, thin clay horizons or in some instances, stylolitic boundaries which mark the bed contacts. ✓ The stylolotic sutures are, of course, secondary solution features but their close parallelism to other bedding features suggests that they have been controlled by a primary factor, perhaps a clay layer or some other primary catalyst to stylolitic solution. The lateral persistence of the stylolitic contacts parallel to the other bedding features also suggests that the pressure solution may have been promoted or caused by an original sedimentary ?compositional boundary.

At some horizons in the thinly bedded carbonates, small scale cross laminations are apparent (Plate 8-F). These small-scale cross-laminae are very probably the result of migrating current-induced ripples as indicated by the displacement of the crests of successively higher layers (Plate 9-A). These ripple features are mostly found in the very pure dolostones but occasionally variable contents of silt-size detrital components have made the laminations more obvious.

Thin (1 - 4 inches) sandy horizons are found at several stratigraphic levels in the carbonate sequence. These thin arenaceous beds often exhibit a surprising lateral persistence - sometimes up to several hundred feet. The sandy horizons are sometimes mixed clastic deposits with the sand grains mixed with micrite and carbonate intraclasts to form

PLATE 8

A. "Shell-structure" chert nodule showing a relatively thin outer chert band containing dolostone and a cherty nucleus within the outer layer.

B. "Balloon-shaped" chert nodule (light coloured) within the Sailmhor member of the Durness carbonates. The centrally downwarped structures within the nodule are sedimentary laminae which may be traced continuously across the lateral margins of the nodules.

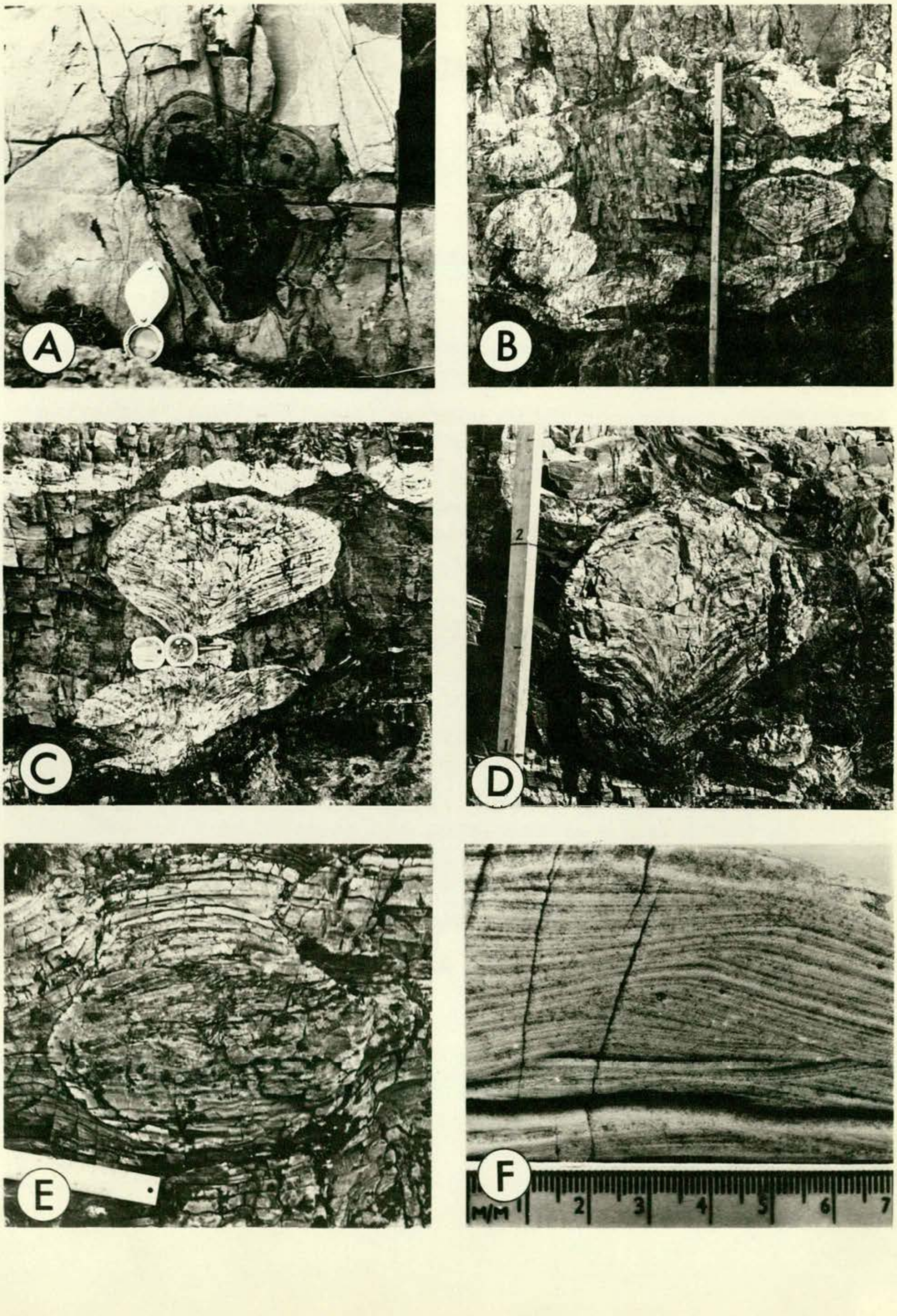
C. "Balloon-shaped" chert nodules (D-891' - D-895') showing details of the internal, centrally downwarped, laminated structures. The photograph also demonstrates the arching of the sedimentary layering in the dolostones overlying the chert nodule. This effect on overlying strata suggests that silicification preceded total compaction of the carbonates.

D. "Balloon-shaped" chert nodules showing the vertical stacking arrangement exhibited by some groups of the nodules.

E. A subspherical chert nodule in dolostone. Bedding structures may be traced across the lateral margins but early silicification and lithification of the nodules is indicated by the deformation of the sedimentary laminae in the dolostones overlying and underlying the chert nodule.

F. Small-scale cross-laminations in fine-grained dolostones of the Eilean Dubh member of the Durness carbonates. Note the lateral off-set of ripple crests in successively higher laminations.

PLATE 8



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arenaceous intramicrites. Other horizons are simply arenaceous micrites showing no evidence of reworking of the carbonates. In all of the arenaceous layers, the quartzose sand grains are extremely spherical and well rounded (Plate 9-B). The beds containing the sand grains generally exhibit sharp upper and lower boundaries with no gradation of the sand into the carbonates above and below either in grain size or grain frequency. The probable genesis of these discrete arenaceous beds is perceived as the result of eolian sands carried seaward during off-shore wind storms.

Nearly all of the sand grains within the carbonates show a pronounced "frosting" effect but it is probable that this surface texture has resulted as much from post-depositional frosting by carbonate replacement (Walker, 1957) as from eolian percussion pitting. Authigenic pitting is indicated by abundant textural evidence as shown in plate 9-C where dolomite may be seen clearly cross-cutting both the detrital quartz grains and their authigenic over-growths.

Intraformational conglomerates also occur very occasionally in the carbonates. The conglomeratic clasts may be either well-rounded fragments of carbonate mud (Plate 9-D) or occasionally, they are distinctly silty and are tabloid in form producing "flake" or "shingle" conglomerates (Plate 9-E). The "flakes" generally contain a greater concentration of silt-size detrital (largely quartzose) material. One "flake" conglomerate horizon in the Eilean Dubh member can be traced laterally to a rather abrupt boundary suggestive of an erosional trough-shaped channel with a flat bottom and several yards wide. This conglomeratic deposit possibly represents the infilling of an east-west oriented tidal channel.

The silty detrital content of the "flake" clasts raises an interesting diagenetic or lithification problem as to why carbonates containing a slightly higher detrital content should have been more completely lithified as they must have been in order that the delicately shaped "flakes" might have been sufficiently coherent to be preserved as "flake" conglomerates. Their preservation suggests that the detrital component has possibly exerted a catalytic effect on the processes of carbonate lithification.

At least five oolitic horizons occur within the Durness carbonates (D-140, D-1363, D-1700, D-1808, D-1958). Additional oolitic horizons are undoubtedly present but have been so obscured by recrystallization or other diagenetic alterations as to escape recognition. Two of the five known horizons were first recognized from thin section studies. The oolitic carbonates have been the focal point of considerable interest during this study for they have afforded the best means of disentangling the diagenetic history of the Durness carbonates as a whole. They have been demonstrably affected by recrystallization, dolomitization, silicification, calcitization, and pyritization. Of the five oolitic horizons, only one (D-1700) has sufficiently retained its original characters to permit classification within the various carbonate classification schemes. This relatively unaltered horizon is a packed oomicrite (after Folk, 1962) or an oolitic packstone (after Dunham, 1962). The oolites occupy roughly 65 percent of the rock with the interstices filled with micrite and sparry calcite at an approximate ratio of 3 : 1 respectively. Throughout these discussions, the term oolite is used only to designate the individual spheroids (= ooid or oolith of some authors). Rocks containing or composed of oolites are desribed as oolitic limestones, cherts and dolostones, or, according to Folk's classification system as oomicrites or oosparites.

In thin section, the oolite diameters range from 0.3 mm to 1.5 mm. Three distinct types of calcareous oolites may be recognized: 1) concentric oolites, 2) radial oolites, and 3) "sparry" oolites.

The genesis of the concentric oolitic structures (Plate 9-F) seems most easily explained as accretionary growth. The concentric oolites are generally large with an equatorial diameter of 1.2 mm. They exhibit a micro-lamination expressive of variation in crystal size and/or impurity content of the laminae.

Oolites with conspicuous radial structures (Plate 10-A and 10-B) are nearly always smaller than the concentric oolites

PLATE 9

A. Small-scale current bedding structures in the fine-grained dolostones of the Eilean Dubh member of the Durness carbonates. Note the lateral off-set of ripple crests in successively higher laminae.

B. (D-793) Detrital quartz grains in dolostone showing a well-developed roundness and sphericity of the grains. The degree of roundness suggests an eolian origin for these quartz grains. Nicols crossed. X 30.

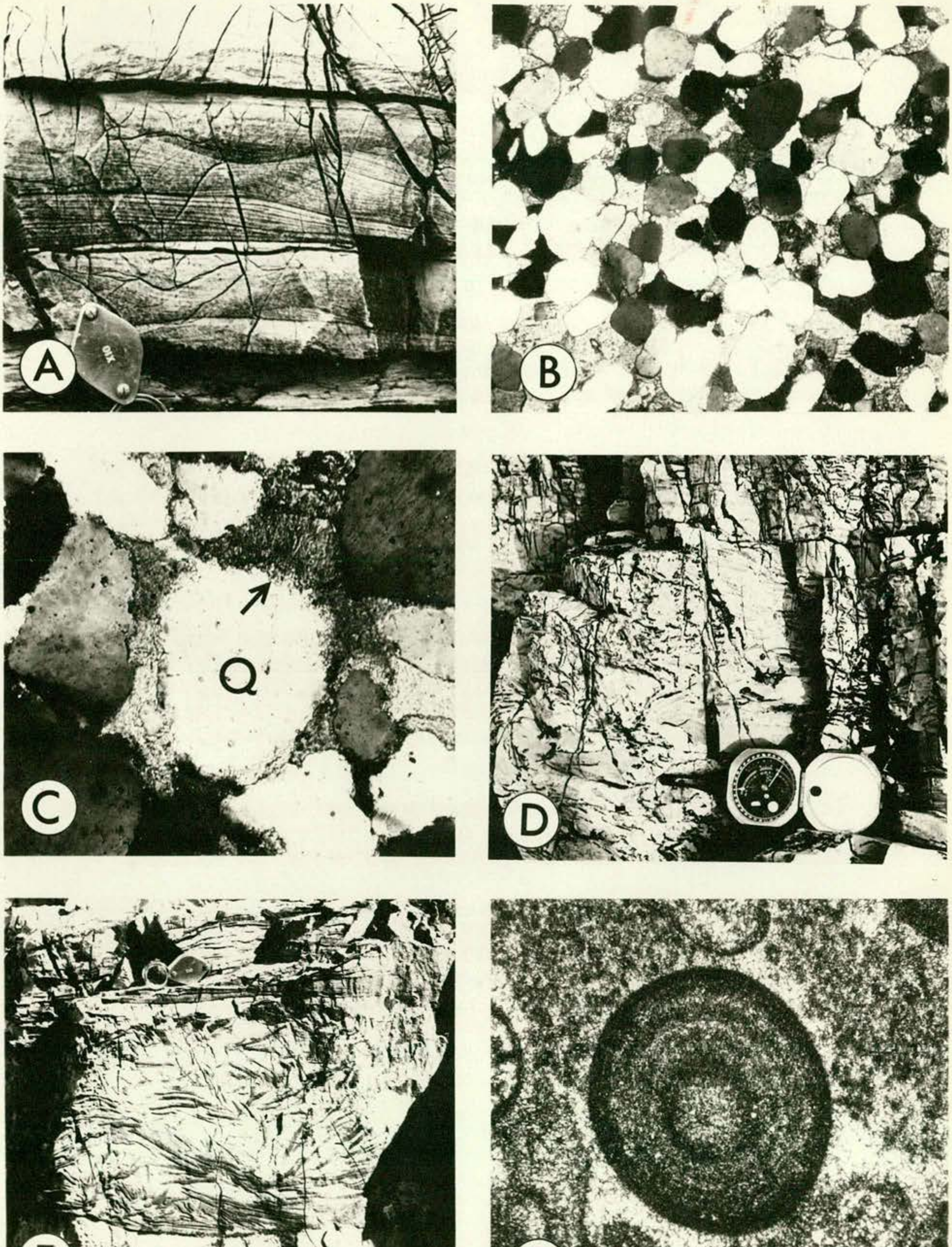
C. (D-793) Detrital quartz grains in dolostone with the grains showing conspicuous peripheral replacement by dolomite. Quartz grains, if dissolved out of the carbonate, are strongly pitted and "frosted". Nicols crossed. X 100.

D. Intraformational conglomerate in dolostone with the clasts and the matrix having a very similar microcrystalline lithology although slightly different in colour. Some of the clasts show a marginal zone of alteration which may reflect a period of subaerial weathering.

E. Intraformational "flake conglomerate" within the Eilean Dubh member. The flakes are slightly silty (quartzose) relative to the microcrystalline carbonate matrix.

F. (D-1700) Concentric "accretionary oolite" surrounded by partly recrystallized micrite. Plane polarized, transmitted light. X 50.

PLATE 9



and have an average diameter of only 0.85 mm. They are divided into 8 to 16 alternating spokes of finely crystalline calcite (0.004 mm - 0.015 mm) and microcrystalline calcite (less than 0.004 mm). The preservation of these radial structures through total silicification also indicates that the microcrystalline sectors contain a greater amount of impurity than the finely crystalline sectors. There is no preferred orientation of the calcite in either portion of these textural alternations and a nucleus may or may not be present. The radial structures never extend to the periphery but are coated by a "concentric" layer 0.01 mm to 0.05 mm thick. The radial pattern may be the result of a recrystallized original calcitic and/or aragonitic growth pattern (cf. Graf and Lamar, 1950) but if so, none of the original radial crystal orientation is retained nor are there any recognizable "bundles" or contraction phenomena as in the "cerebroid oolites" of Carozzi (1962). The absence of a gradation between the radial and concentric types (i.e. there are no partly radial and partly concentric oolites) and their consistent difference in size oppose the interpretation that the radial structures of these Durness oolites have developed through recrystallization. Illing (1954 p. 42-44) attributed similar radial structures to the action of boring organisms but this explanation is discouraged here by the ubiquitous presence of the coating layers which are never penetrated by the radial structures. The boring genesis is opposed by the unlikelihood that all of the bored oolites should have developed an accretionary coating layer or that the accretionary layer should never itself have been bored. Calcareous algae may have been responsible in some way for the radial structures but once again, the omnipresence of the outer concentric layer offers a serious problem. It is perhaps possible that the radial structures represent an algal growth pattern of algae occupying radially arranged tubules (possibly comparable to Girvanella tubules) and that the oolite developed by this algal growth mechanism until it reached a critical size whereupon it became subject to accretionary growth. The accretionary growth induced by wave or current movements could have overwhelmed the algae and have formed the ubiquitous coating layer.

The third type or "sparry oolite" appears to have resulted from a partial recrystallization of the concentric and radial oolites. Although diagenetic, these oolites are sufficiently common and problematical to merit consideration as a separate type. The typical "sparry oolites" possess an internal crystalline texture of "drusy" calcite (Bathurst, 1958) which may occupy greater than 90 percent of the oolitic cross-sectional area. The drusy texture is characterized by a progressive increase in crystal size of the sparry calcite toward the center of the oolite (Plate 10-C and 10-D). The common presence of a centered relic micritic "nucleus" (Plate 10-E) opposes a void-filling interpretation for the drusy textures since the nuclei would have settled to the bottom during any void stage (Carozzi, 1963). Occasionally (roughly 10 percent) the oolites possess nuclei that are not centered and which would appear to have dropped or settled (Plate 10-F). An examination of the relative orientation of these off-center nuclei within the oolites of a single thin section discloses no preferred settling direction. Binocular microscopic examination of polished, etched surfaces of these rocks establishes beyond question that the majority of the nuclei are centered and unsupported within the sparry cores.

These conflicting lines of evidence concerning the void-filling vs. a recrystallization genesis for the drusy textures lead ultimately to the conclusion that either there are separate origins for those sparry oolites with nuclei and those without or that drusy textures in calcite are not always a reliable indication of a pre-existing void. The common denominators of the nucleated sparry oolites and the non-nucleated sparry oolites oppose separate origins for these obviously related sparry textures.

Very recent work on Pleistocene and modern oolites (Bathurst, 1964; Wood, 1964; Friedman, 1964) would seem to support a leaching and void-filling concept for such sparry textures with aragonite solution followed by calcite precipitation in the voids or fluid-filled cavities.

PLATE 10

A. (D-1700) "Radial oolite". The darker spokes are more finely crystalline and contain a greater percentage of impurities than the lighter areas. Note the elongate nucleus of this particular oolite and the accretionary coating layer at the periphery. Plane polarized, transmitted light. X 85.

B. (D-1701) "Radial oolite". The radial spokes of this oolite appear to extend inward to varying levels. The outer accretionary layer on this oolite is unusually thin. Plane polarized, transmitted light. X 80.

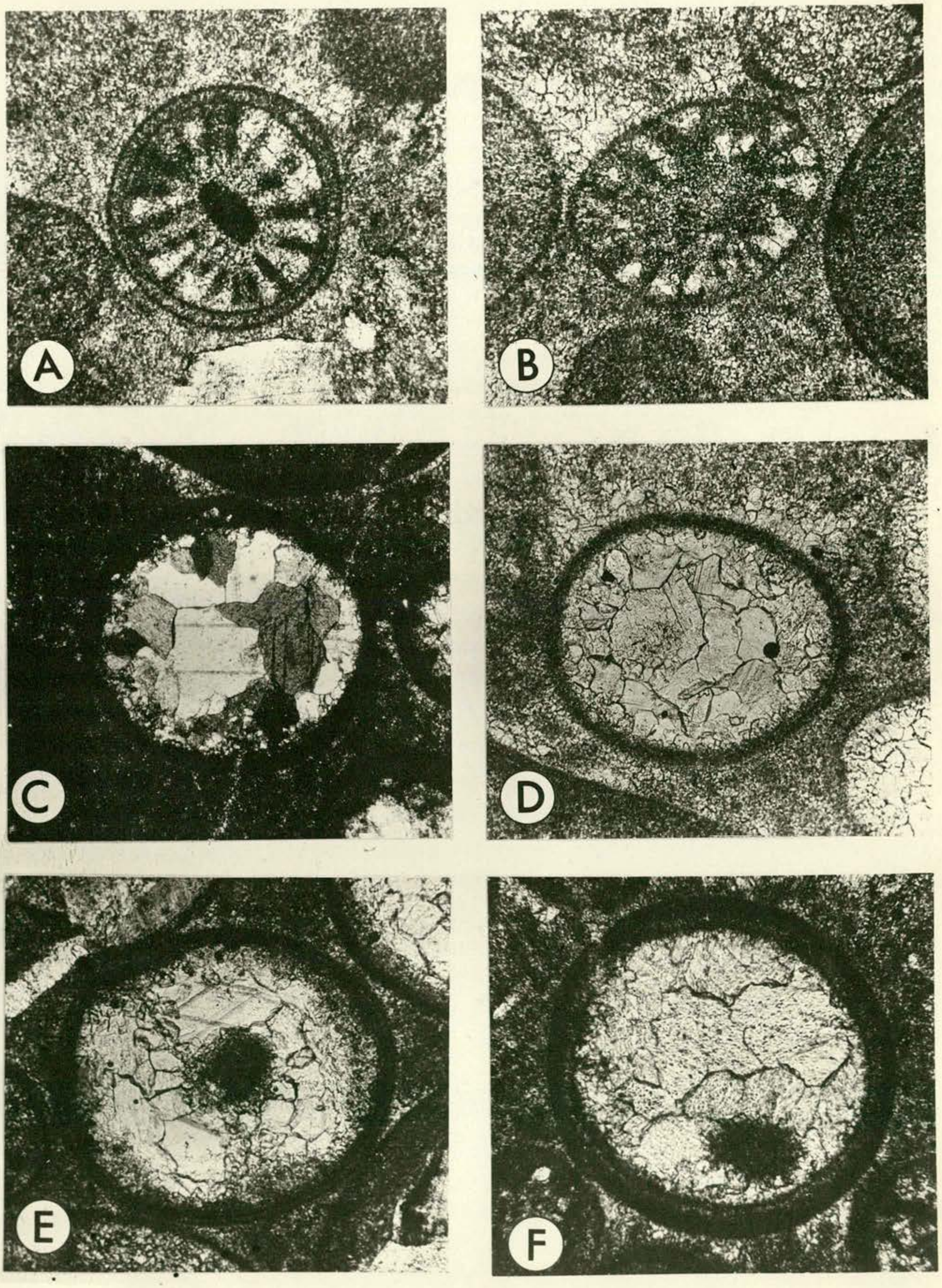
C. (D-1700) "Sparry oolite". Note the "drusy" or void-filling appearance of the sparry calcite centre. Nicols crossed. X 50.

D. (D-1700) "Sparry oolite". The sparry calcite occupies nearly the total cross-sectional area of the oolite. Note the progressive increase in crystal size toward the centre of the oolite. Plane polarized, transmitted light. X 50.

E. (D-1700) Sparry oolite with a relic micritic nucleus. Plane polarized, transmitted light. X 50.

F. (D-1701) Sparry oolite with a relic micritic nucleus which appears to have settled to the margin. Examination of the orientation of marginal nuclei within a single thin section discloses a random orientation with no preferred "settling" direction. Plane polarized, transmitted light. X 60.

PLATE 10



However, in order to apply this leaching mechanism to the Durness oolites, some means of supporting the centrally oriented nuclei during the leached or void stage would need be conceived. It is also somewhat incongruous, at least for the Durness oolites, that if leaching has occurred, it has never transected the oolite boundaries (this might be explained by the presence of micritic or organic "envelopes" (Bathurst, 1964)) and that collapse structures should be so uncommon in these sparry oolites if a void stage has existed (Carozzi, 1961).

The nearly complete gradation from concentric or radial oolites (Plate 9-F and 10-A) to partly sparry oolites (Plate 11-A, 11-B, 11-C and 11-D) to totally sparry oolites (Plate 10-C and 10-D) lends support to an interpretation that these sparry oolites have developed through a process of recrystallization. Yet, if the drusy textures of these oolites have originated by recrystallization, and this seems very probable, the implications of this interpretation must also be accepted - namely that the presence of "drusy" sparry calcite textures need not indicate the former existence of void space. Extending this thought to the "drusy" textures in other rocks could require a serious re-examination of the petrologic interpretations of sparites in general.

A detailed consideration of the effects of diagenesis on these various oolitic horizons and textures will be offered in later discussions under the heading of Diagenesis.

> "Recognizable algal stromatolites are more validly treated as organosedimentary structures than as fossil organisms." (Logan et. al., 1964).

At least fourteen separate stromatolitic algal horizons occur throughout approximately 1500 feet of the Eilean Dubh, Sailmhor, Sangamore, and Balnakiel members. Cryptozoon structures, Collenia structures, and compound structures resulting from mixtures of these two forms are all recognizable within the Durness carbonates. Unfortunately, considerable ambiguity has resulted from over-use of these generic names. The following recent definitions proposed by Logan et. al., (1964) are implied in the above used names:

> "Collenia: discrete or laterally linked hemispheroidal bodies, sometimes spheroidal, composed of concave-convex laminae."

> "Cryptozoon: discrete, club-shaped or columnar structures composed of vertically stacked hemispheroids, expanding upward from a base; the laminae are usually domed toward the top and incurved around the edges."

The lowest stratigraphic occurrence of recognizable algal stromatolites is a Collenia type structure (Plate 11-E) which is found in the Eilean Dubh member dolostones only a few feet above tidal high water mark along the coastal exposures west of Balnakiel (National grid ref. - NC 374686).

This horizon corresponds to the measured stratigraphic interval of D-481 in the composite measured section (see appendix). The rocks of this horizon are totally dolomitic and the dolomitization has undoubtedly obscured the algal stromatolite structures to a considerable extent. The faintly recognizable laminated structures are mainly expressive of variations in the crystalline textures (Plate 11-F). The laminae are arched into heads or hemispherical domes that range from 6 inches to 1 foot in diameter (Plate 12-A). In certain instances the laminae can be traced through a concave arch between adjacent domes but often there is no discernible structure between the domed structures. In some of the more finely crystalline laminae within the Collenia domes, there occurs a distinct microscopic structure which is very like the flexuous tube structures of the algal genus Girvanella (J.H. Johnson, 1961, p. 194, pl. 84 & 85.) (Plate 12-B). This apparent occurrence of a non-stromatolitic genus within these structures (if the binomial nomenclature for stromatolitic algae is accepted) lends support to the arguments advanced by Logan et. al. (1964) that the structural forms of stromatolitic algae are dependent upon environmental conditions rather than being the result of different species. The existence of these Girvanella type tubes within the Collenia stromatolitic structures raises the distinct possibility that Girvanella type Schizophytes, although not commonly associated with stromatolitic structures,

PLATE 11

A. (D-1701) Concentric oolite showing a very slight recrystallization of the central portion. Plane polarized, transmitted light. X 50.

B. (D-1700) Concentric oolite showing partial recrystallization to a sparry oolite. Nicols crossed. X 50.

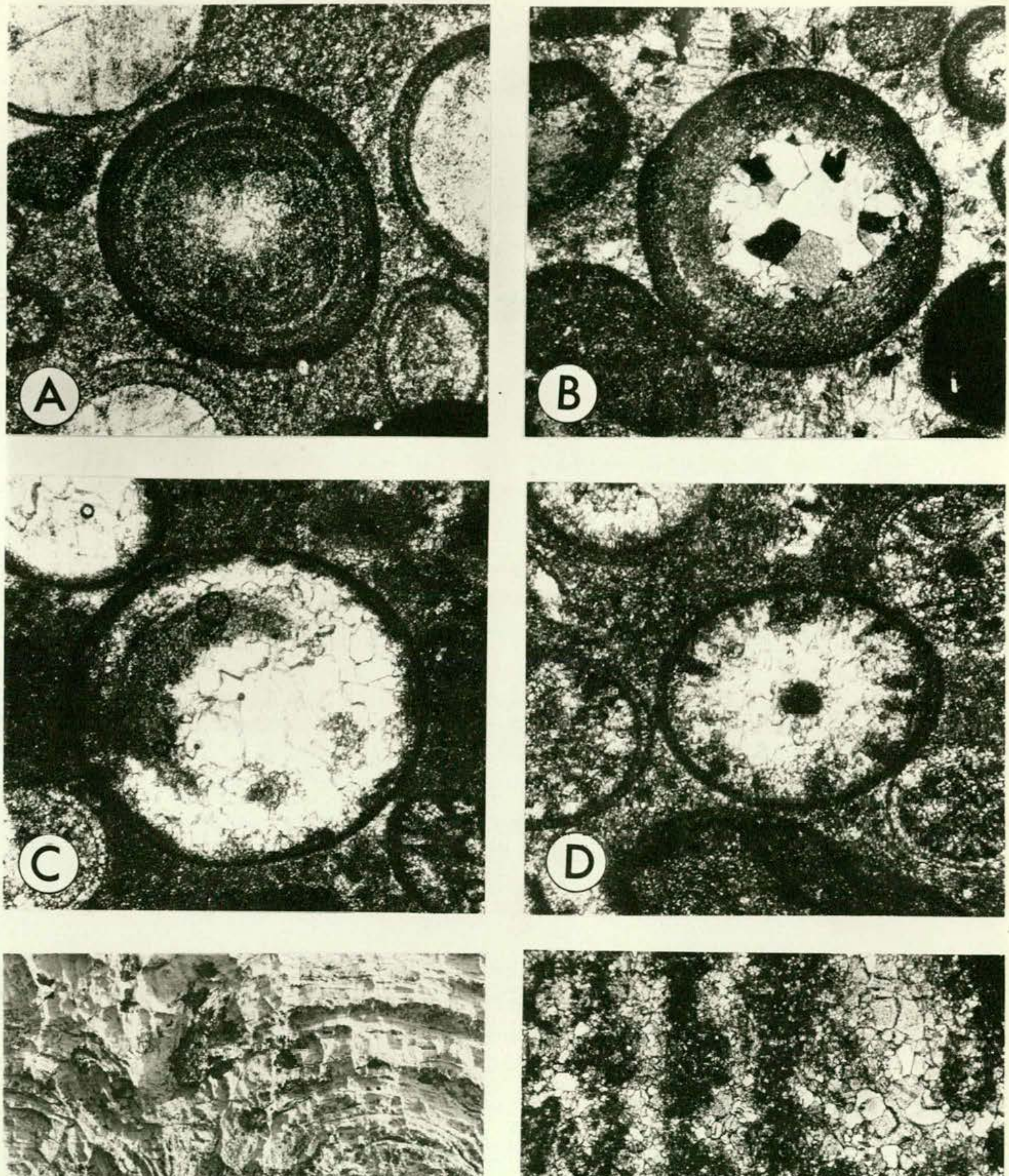
C. (D-1700) Concentric oolite showing a further progression of recrystallization to a sparry oolite. Note the detached micritic relics. Plane polarized, transmitted light. X 60.

D. (D-1701) Radial oolite showing partial recrystallization to a sparry oolite. Plane polarized, transmitted light, X 50.

E. Domed concentric layers of stromatolitic algal "heads" of a Collenia type within the Eilean Dubh member of the Durness carbonates at D-481'.

F. (D-481) Photomicrograph of the laminated structure within the Collenia stromatolite horizon. The laminated appearance in outcrop is mainly due to the conspicuous banded microscopic variations in crystal size. X 30.

PLATE 11



may contribute to their formation under certain environmental conditions.

Logan et. al. (1964) postulate that the laterally linked hemispheroid structure of Collenia is characteristic of a "marine, intertidal mud-flat environment, mainly in the protected locations of re-entrant bays and behind barrier islands and ridges where wave action is usually slight (Black, 1933; Ginsburg et. al., 1954)." If this interpretation is accepted, it agrees with Folk's (1962) interpretation of genesis of the Type III limestones but it contradicts the conditions of deposition suggested by Peach et. al. (1907). Rather than indicating a rapidly subsiding and continually deepening basin, these algal limestones suggest water conditions less deep than indicated by the underlying "Fucoid Beds" or at least extremely shallow.

Roughly 200 feet stratigraphically above the Collenia type structures occurs quite a distinctly different type of algal stromatolitic structure (D-578, D-662, D-740 etc.). These higher algal horizons exhibit the cryptozoon type digitate processes which envelope nodules or structureless hemispherical mounds (Plate 12-C). The subsidiary digitate structures are discrete, vertically stacked hemispheroids consisting of close-linked hemispheroidal microscopic laminae ($\frac{\text{SH - V}}{\text{LLH - C}}$ type of Logan et. al., 1964). They attain diameters up to 2.5 cm (Plate 12-D) and lengths up to 8 cm.

(Plate 12-E). No internal structures other than laminated variations in dolomite or calcite crystal size are discernible within the cryptozoon digits (Plate 12-F).

The composite structures built by these "colonies" of cryptozoon stromatolitic algae are close lateral-linked hemispheroids (LLH-C type of Logan et. al. 1964). The composite structures achieve dimensions up to 5 feet in height and 8 feet in maximum horizontal diameters. Several other higher horizons display a very similar pattern of the close laterally-linked hemispheroids, some with and some without the laminar or stromatolitic microstructure. In those instances where the stromatolitic structure is absent from the hemispheroids, they are generally more coarsely crystalline and the laminations have very probably been lost through recrystallization or authigenesis.

Finally, before considering the secondary structures and authigenic features of the carbonates, some additional comments are pertinent regarding the nature and significance of the previously discussed cyclic or rhythmic patterns of the Sailmhor dolostones (D-900 to D-1386). Reviewing briefly, the rhythm consists of three basic lithologies arranged in a 1 - 2 - 3 - 1 - 2 - 3 sequence. The lithologies are: a very dark coloured and prominently mottled, coarsely crystalline dolostone roughly 4 feet thick which grades upward into 8 to 10 feet of medium grey, coarsely crystalline dolostone, considerably less mottled than the darker unit below. The

PLATE 12

A. Stromatolitic algal "head" structures of the *Collenia* type at D-481'. Note the continuation of the laminae from one "head" to another in the upper layers.

B. (D-481) Photomicrograph of one of the finely crystalline layers within the *Collenia* stromatolites showing light coloured flexuous tubules similar to *Girvanella*.

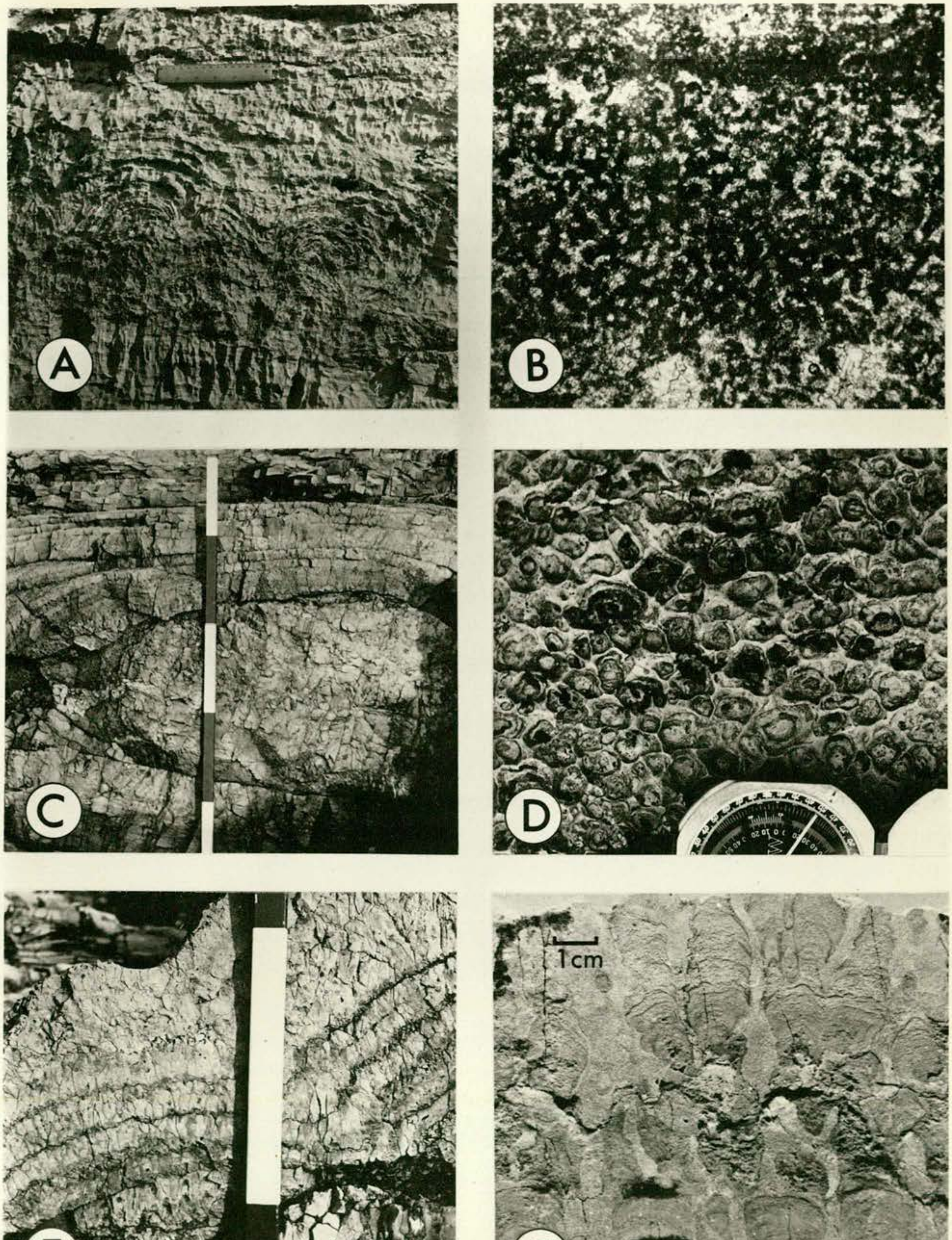
C. *Cryptozoon* type stromatolitic algal laminae coating a slightly siliceous, structureless nodule within the Eilean Dubh dolostones. Painted intervals = 1 foot.

D. Top view of a bedding plane exposure demonstrating the close-linked "colonial" character of the *Cryptozoon* stromatolites.

E. Transverse section on exposed fracture surface at D-663' showing the close-linked digitate structures in the Cryptozoon stromatolites. Painted interval = 1 foot.

F. Surface cut and etched to show the digitate structure of the *Cryptozoon* stromatolites.

PLATE 12



medium grey unit is overlain in turn by a light grey, thinly bedded, medium crystalline, cherty dolostone, 1 to 2 feet thick. The complete cycle is thus roughly 15 feet thick and although now mainly expressive of different diagenetic mottling patterns, must reflect some pre-authigenic sedimentary controls. The term cycle or rhythm is used here according to the definition proposed by Duff and Walton (1962) as ...... "a group of rocks units which tend to occur

> in a certain order and which contains one unit which is repeated frequently through the succession".

The primary rhythmic controls that have produced the diagenetic patterns may have been either compositional or textural variations in the original sediments. The frequency of repetition, the relatively minor thicknessess involved, and the consistent order displayed by the cyclic units perhaps suggest local controlling factors rather than eustatic sea-level changes or orogenic movements.

E. SECONDARY STRUCTURES

Disregarding authigenic replacement structures (e.g. chert nodules or dolomitic mottling) there are certain other secondary structures of interest in the Durness carbonates which have resulted from compaction and/or solution effects. One of the most intriguing of the compaction phenomena is the occurrence of some extraordinary load structures in the Eilean Dubh member dolostones (Plate 13-A). These load structures are arranged along definite planes in thinly bedded lutite-grade dolostones and display a pronounced deformation of the stratification underlying the load plane and, to a lesser extent, a deformation of the stratification overlying the load plane. The structures are approximately 2 to 3 inches in height with some few showing as much as 6 inches of relief on the plane of deformation. In plan, they are not subcircular as are many load structures but are long ridges and troughs (Plate 13-B) somewhat similar in their geometry to the load-casted ripple marks described by Kelling and Walton (1957). They differ from the load-casted ripples in that they display only very slight variations in geometry along their axes as though a slippage movement parallel to the axes had removed any irregularities. Minor irregularities on the ridge and trough structures are also parallel to the major structures. The lithologies above and below the plane of load deformation are virtually identical which is unusual for loading features. The elongation and crude parallelism is also somewhat atypical and is a perplexing aspect of these structures. It is possible that the pattern of loading was influenced by parallel ripple marks (Kelling and Walton, 1957; Dzulynski and Kotlarczyk, 1962) and, in fact, the Eilean Dubh dolostones occasionally show very small-scale cross-stratification (Plate 8-F and 9-A) which would perhaps lend support to this hypothesis.

The roughly parallel elongation might also have resulted from movement of the overlying material (?downslope) during the formation of the structures. The "flame structures" (Walton, 1956) shown in Plate 13-C and the deformation of both the underlying and overlying stratification preclude any possibility of the structures being primary scouring or depositional features. In order for the "flame structures" to exist, the units immediately above and immediately below the plane of load deformation must both have antedated the structures. However, the limited upward influence of the deformation in the units overlying the load structures suggests that only a very few centimeters of material had been deposited above the eventual plane of load deformation before the load structures were formed. This phenomenon of a very limited upward vertical influence of the load deformation is apparently rather commonplace in clastic deposits (E.K. Walton, personal communication) but usually an obvious textural and hence density difference is available to provide a "loading" mechanism for the production of load structures with such limited "overburden" (Kelling and Walton, 1957; Dzulynski and Walton, 1962; and others). In these carbonates, no textural or compositional differences between the layers above and below the load plane are apparent. Thus it becomes difficult to explain how a few centimeters of apparently homogeneous overburden could cause load structures with 10 to 15 cm. of relief. The following hypothesis is offered as a possible genesis for these extraordinary structures in homogeneous carbonate sediments and as a possible factor in the genesis of other load structures.

A homogeneous carbonate mud deposited as a bed perhaps measuring a few centimeters in thickness could be expected to possess initially a fluid-filled porosity of as much as 50 percent of its total volume (Pettijohn, 1957). If such a micritic deposit were capped by a relatively impermeable (?clay) layer, the porosity and entrapped connate water of the unit might have been metastably retained during the deposition of a thin overlying similar carbonate mud layer. In this metastable condition, any slight disturbance or simply critical overloading might have caused a rather abrupt repacking and reduction of the porosity in the lower unit. This would have had the effect of expelling the less dense water up to the relatively impermeable capping layer. The concentration of water (? and clay) at this horizon would very probably produce a highly fluid and well lubricated slurry which would be highly responsive to any differential loading in the thin overlying layer. Sedimentary ripples might cause and be accentuated by differential loading or, if the sediments were resting on a slightly inclined slope, the overlying unit might have been sufficiently well lubricated to slide downslope on the slurry layer and produce, by this movement, the crudely parallel troughs and ridges. The "flame structures" could result as the more fluid and less dense muds attempted to escape upward to the surface or, perhaps more accurately, were being displaced upward by the heavier carbonate materials. This "entrapped water" hypothesis is promoted somewhat by the presence of a thin layer of undetermined composition at the plane of load deformation. It was found impossible to remove from the outcrop a sample which transected this load-plane boundary and, while not necessarily indicating an impermeable layer, it does indicate some compositional inhomogeneity.

Another possible factor in the loading mechanism might be the effect of dilatancy (Mead, 1925). After the initial compaction of the unit underlying the plane of eventual load deformation, any subsequent deformation might increase its porosity and help to "freeze" the structures by decreasing the fluidity.

Most of the remaining secondary structures of the Durness carbonates are either the direct or indirect result of solution or diagenetic phenomena. The secondary structures resulting directly from solution processes include stylolitic sutures and cavernous horizons both of which have been briefly mentioned in the foregoing discussions. Stylolites are only moderately common and are mostly very small-scale solution features with a relief on the stylolitic boundaries ranging from a fraction of a millimeter (microstylolites) up to 2.0 cm. The combined loss of stratigraphic thickness through stylolitic solution processes has probably been negligible.

In thin section, the stylolitic surfaces are marked by thin deposits of relatively insoluble matter, often hematitic, (Plate 13-D), but in outcrop they are commonly detected by their increased susceptibility to weathering which causes them to appear in negative relief.

In partly dolomitized limestones, the dolomitization and lithification has, in every case, preceded the formation of the stylolites. That is to say, the stylolites cut through the dolomitized and undolomitized portions of the rocks without regard to the variations in mineralogy (Plate 13-E). It is also significant in Plate 13-F, that the stylolites cut through the oolites and interstitial matrix without regard to the textural variations.

Perhaps somewhat more surprising is the occurrence of microstylolites within partly dolomitized chert nodules (Plate 14-A) where again, the stylolites appear to disregard the compositional differences. The most widely accepted theory for the origin of stylolites is that they are the result of pressure solution. The removal of material from either side of the stylolitic contact is undeniable. Within the Durness carbonates, this apparent disregard for pre-existing compositional boundaries would seem to imply that the pressure factor is more critical than a chemical or compositional control. It seems unlikely that the solubility potential for chert, dolomite, and calcite could be coincident by any combination of purely chemical factors.

PLATE 13

A. Load structures at D-639'. Note the deformation of the laminae beneath the load plane and the very limited upward extent of the structures in the overlying layers.

B. Oblique view of the under-surface of the plane of load deformation at D-639' demonstrating the crudely parallel, ridge and trough character of the load structures.

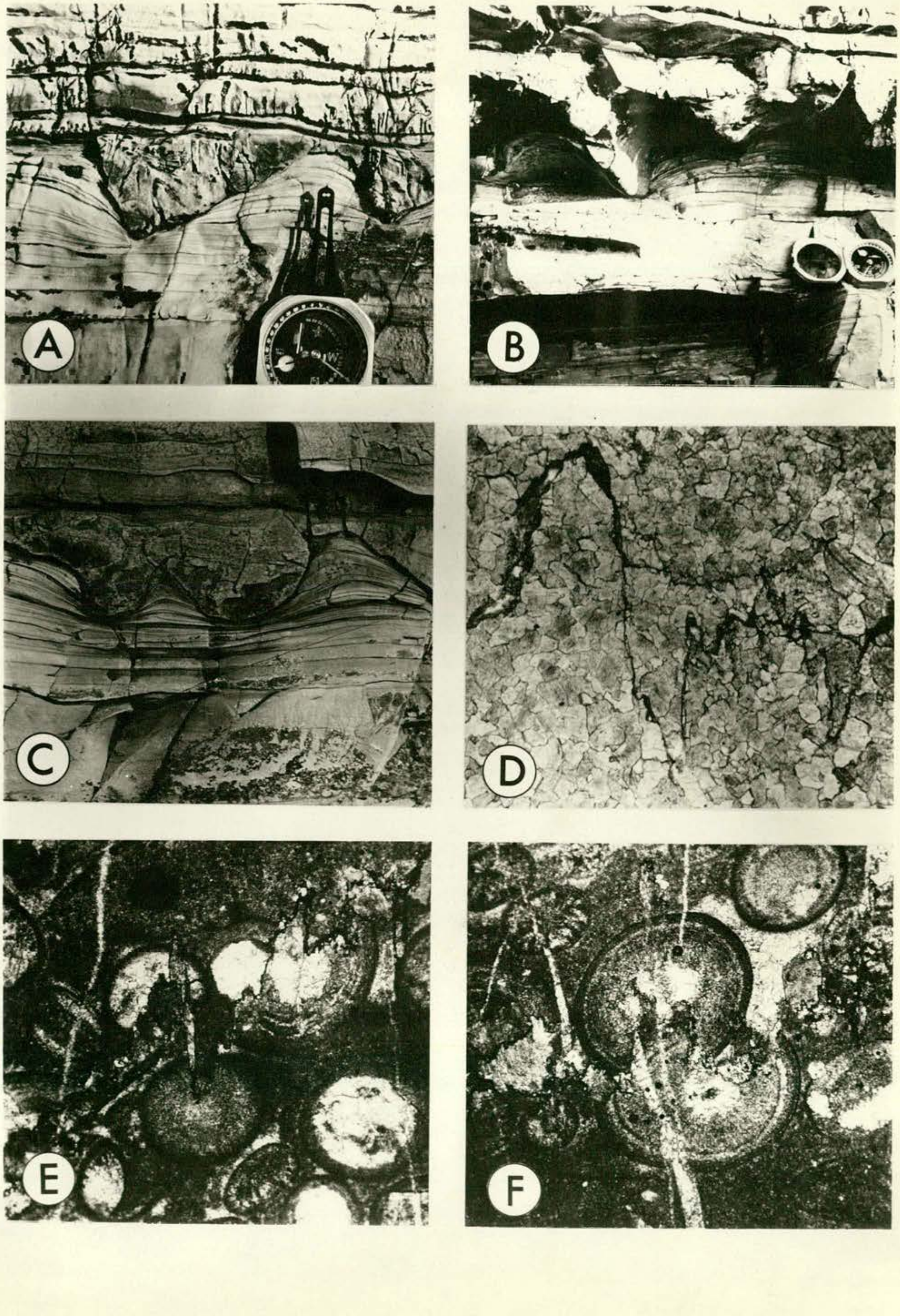
C. Load structures at D-639' showing the development of "flame structures". Note the very limited upward effect of the load structures on the overlying strata. Also note the arching of the sediment laminae in the "escape channel" of the flame structure on the right side of the photograph.

D. (D-351) Photomicrograph showing the characteristic insoluble residue collected along a sutured stylolitic contact in dolostone. Plane polarized, transmitted light. X 50.

E. (D-1700) Stylolitic suture passing through partly dolomitized oolitic limestone. The stylolitic pressure solution appears to disregard the mineralogical variations. Plane polarized, transmitted light. X 30.

F. (D-1701) Photomicrograph showing partial solution of two adjacent, partly recrystallized, concentric oolites. The stylolitic suture continues laterally into the micritic matrix. Plane polarized, transmitted light. X 30.

PLATE 13



A pressure controlled mechanism for stylolite formation in limestones is in harmony with the proposals of Stockdale (1926, pp. 402):

> ......"since pressure is important in increasing the solubility of certain solids in liquids, the rock opposite the top and bottom surfaces of adjacent undulations will succumb to a greater rate of solution, producing (a) a deepening of the interpenetrating parts, (b) a further decrease in pressure (and consequent dissolving of the rocks) on the sides of the undulations, and (c) a possible final development of vertical columns with a decided concentration of the pressure and accompanying solution at the ends".

Prokopovich (1952) presents evidence which seems to contradict the pressure-solution mechanism for certain stylolites and he suggests they could not have formed in a solid rock but probably developed through solution processes in relatively soft sediments. From the character of the stylolitic contacts in the oolitic rocks of the Durness carbonates (Plate 13-F) as in the oolitic limestones studied by Bastin (1951), it seems necessary to conclude that not only the formation of the oolites, but also the lithification of the sediment into a coherent rock must have antedated the stylolitic solution.

The deformation of the stylolitic sutures by probable

Moine cleavages in some of the Durness carbonates permits an upper time limit to be drawn and it is possible to conclude that the stylolites in the Durness carbonates are post-consolidation, post-silicification, post-dolomitization, and pre- ?Moine thrusting.

Cavernous horizons were noted at three stratigraphic levels in the carbonates, two of which were previously mentioned in connection with the "balloon-shaped" chert nodules with the internal "collapse" structures at D-890 and D-920. The vugs of these horizons are approximately 1 to 1½ feet in diameter and are roughly elliptical in shape. Many of the vugs have been partly or totally filled with very coarsely crystalline calcite. If the proposed hypothesis for the genesis of the "balloon shaped" chert nodules and their silicified central collapse structures is accepted, it poses certain difficulties regarding the time of formation of these solution features. The chert nodules containing the "collapse" structures have demonstrably affected the compaction of the sediments immediately above them. Thus, if the pattern of silicification has been controlled by structures collapsing into the cavernous horizons, it implies that the formation of these solution features must also have preceded total compaction of the sediment. Early solution vugs might conceivably have resulted from the differential leaching of ?aragonitic nodules from a calcitic matrix. It is extremely difficult to study these solution features

Stromatolitic heads, not 'collapse' structures.

because of local structural complications presumably caused by the solution and the evidence is by no means certain for the proposed origin of the "collapse" structures. It is equally possible that the downwarped structures which produced the "balloon shaped" nodules were formed by tectonic or slumping processes and that their relationship to the cavernous horizons is completely fortuitous.

The macroscopic diagenetic structures observable within the Durness carbonates are mainly the nodular or bedded cherts and the dolomitic mottling. Although undeniably secondary structures, these authigenic features do not merit separate discussion here.

F. MICROSCOPIC CARBONATE PETROLOGY

1. Fabrics

The most common fabrics (textures and structures) exhibited by the limestones of the Durness carbonates are interlocking, very finely crystalline or microcrystalline textures. Detectable pore space is absent from all of the limestones and hence recrystallization or cementation has necessarily affected all the rocks. This tight interlocking crystalline texture may account for the lack of dolomitization of these units by virtue of its impermeability. Within the microcrystalline textures are occasional irregular patches of more coarsely crystalline fabrics, in some instances suggestive of void-filling "drusy" calcite mosaics, and in other instances more suggestive of "grain growth" mosaics (Bathurst, 1958). The difficulties of applying the genetic interpretations, as proposed by Bathurst (1958), to some of these limestones have been discussed relative to the "sparry oolites". Generally, the limestone textures are too finely crystalline to apply the test of interfacial angles (Bathurst, 1964) to distinguish "grain growth" fabrics from "granular cements" and "drusy mosaic" fabrics.

The dolostones display a wide range of crystalline fabrics from microcrystalline textures to extremely coarsely crystalline mosaic textures. Within the more coarsely crystalline dolostones, the contact relationships between adjacent crystals are also inconstant and vary from planar contacts (Plate 14-B) to irregular contacts (Plate 14-C) and occasionally to highly sutured contacts (Plate 14-D).

Karcz (1964) has mistakenly applied the criteria for distinguishing recrystallization fabrics in limestones (Bathurst, 1958; 1959; 1964) or in metals (Stanton, 1964) to the authigenic replacement fabrics of the Durness dolostones from the Isle of Skye. If the arguments favouring the authigenic nature of these dolostones are accepted, it is clearly inconsistent to treat their fabrics as recrystallization textures. Much further work is necessary before the crystal mechanics of authigenic textures will be sufficiently well understood to permit genetic interpretations to be generally attached to the interfacial contact relationships between authigenic mineral crystals. To apply genetic interpretations to authigenic fabrics would presently require proof that recrystallization had post-dated authigenesis.

The interpretation of a void-filling genesis for certain "drusy" dolomite textures is sometimes supported by the zonation of dolomite crystals which exhibit zones of impurities contained within a closing pattern of adjacent dolomite crystals that must have surrounded a void at the time the impurities were introduced (Plate 14-E). It is probable, although not positive, that these nearly continuous zones of impurities were "primary" to the "drusy" dolomite rather that relic from zoned "drusy" calcite textures. This is but one example of the difficulties encountered in attempting to interpret the genesis of authigenic textures. Another example was previously demonstrated from the Serpulite Grit member of the An t-Sron formation where carbonate "clasts" which exhibit the characteristic stereome structure of echinoderm plates, and possessing "rim cementation" syntaxial overgrowths on the clasts, have been completely dolomitized (Plate 6-C). The resulting interpretive problem is whether the "rim cementation" antedated, post-dated, or coincided with the dolomitization of the clast, and whether, under any of these circumstances, it is appropriate to now consider the syntaxial dolomite overgrowth as "rim cement" as defined by Bathurst (1958).

All of the fabric types described by Bathurst (1958) (granular cements and drusy growth, rim cementation, pressure solution, and grain growth) may be found in comparable fabrics within the Durness dolostones but whether these are primary, recrystallized, or authigenic textures is uncertain.

There are occasional eccentric patterns of dolomite fabrics (e.g. Plate 14-F) whose genesis remain beyond explanation or speculation. Equally perplexing are certain interpenetrating fabrics of dolomite and calcite (Plate 15-A) and of dolomite and chert (Plate 15-B). In most of these problems involving two or more minerals, it is possible, from textural evidence, to determine the host and guest mineral relationships but this often does not explain the fabric relationships.

Chert fabrics display a variety of patterns dependent upon both the chert textures, as discussed earlier, and upon the pre-silicification fabrics. When silicification has acted upon pre-existing clastic textures, it is sometimes possible to observe, within a single thin section, all three of the chert textures (microcrystalline, chalcedonic, and drusy quartz) (Plate 15-C).

2. Diagenesis

For the author, the most intriguing and fascinating aspects of this thesis project have centered upon the petrologic investigations of the carbonates. The pre-

PLATE 14

A. (D-1394) Photomicrograph showing a stylolitic suture cutting through a partly dolomitized chert nodule. Since the stylolite cuts both the chert and the euhedral authigenic dolomite, the period of pressure solution must post-date the silicification and the dolomitization. Plane polarized, transmitted light. X 50.

B. Photomicrograph demonstrating the occurrence of planar intercrystalline contacts in a coarsely crystalline dolostone. Plane polarized, transmitted light. X 30.

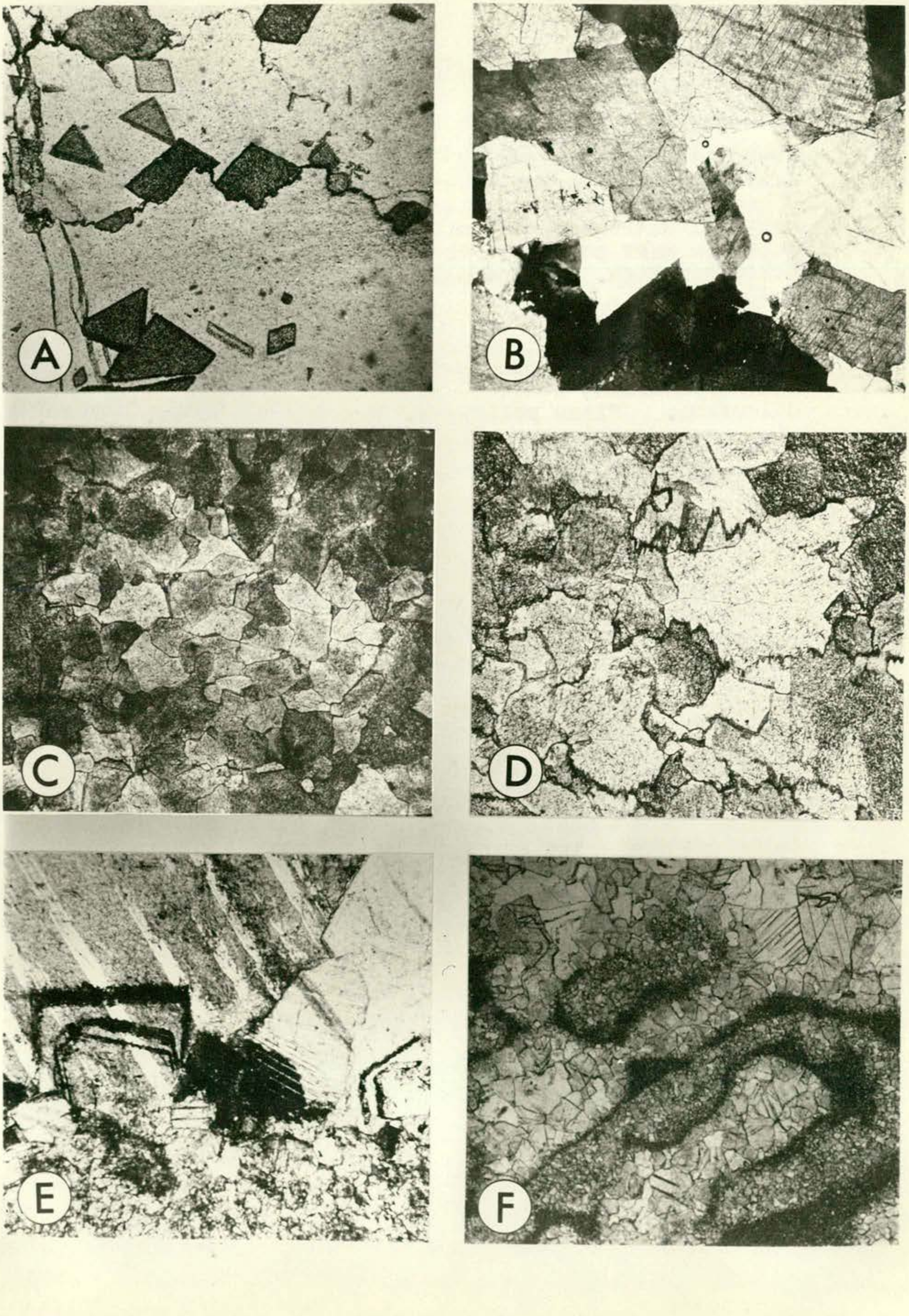
C. Photomicrograph showing the occurrence of irregular intercrystalline contacts in a coarsely crystalline dolostone. Plane polarized, transmitted light. X 30.

D. (D-863) Photomicrograph showing conspicuously sutured intercrystalline grain contacts in coarsely crystalline dolostone. Plane polarized transmitted light. X 50.

E. (D-476) Hematite zonation of a dolomite crystal probably representing an early growth stage while the crystal was marginal to a former void. Nicols crossed. X 50.

F. (D-314) Weird patterns in crystalline textures of dolostone for which no cause is apparent. Plane polarized, transmitted light. X 30.

PLATE 14



occupation with the details of this aspect of the study undoubtedly reflect the writer's predilection, both during the sampling and by an emphasis of interest during the microscopic investigations. With or without a bias toward diagenetic studies, only the rocks themselves could have offered the remarkably clear and consistent, although complex, history of diagenesis which has emerged.

As previously mentioned, five oolitic horizons within the carbonate sequence have been an invaluable asset to the interpretation of the diagenesis (D-140, D-1363, D-1700, D-1808, and D-1958). Authigenic replacement phenomena including silicification, dolomitization, calcitization, and pyritization have transected, obscured or obliterated the oolitic textures. As subsequent photomicrographs will reveal, the oolitic textures have greatly facilitated the interpretation of the diagenetic textures by offering a means for the determination of the guest-host relationships of the authigenic minerals. It will be most economical to consider first the diagenesis of these oolitic horizons and then to project those interpretations to other situations where such consistent primary fabrics are unavailable.

The criteria employed to establish the guest-host relationships of the diagenetic minerals were: 1) Cross-cutting relationships of minerals to structural features (e.g. fossils, bedding, etc.) or textural features (e.g. chalcedonic banding), 2) Relic syntaxial inclusions,

3) Diminution or loss of structural details (producing "ghost structures"), 4) Euhedral crystal faces in an otherwise clastic texture, 5) Mineral incompatibility with structure (e.g. silicified fossils), 6) Pseudomorphic mineral euhedra, and 7) Syntaxial crystal overgrowths on clastic or biogenic particles.

The interpretations of the dolomite-calcite relationships were made considerably easier through the application of an Alizarin Red "S" and potassium ferricyanide stain (Evamy, 1963) to all thin sections. The reliability of the staining procedure was established by X-ray identifications of the carbonate minerals.

The three types of calcitic oolites previously described (concentric, radial, and sparry) plus their variations in thin section resulting from non-equatorial sections, show evidence of subsequent modification by partial to complete dolomitization and/or silicification. Some also show evidence which indicates a calcitization stage post-dating the silicification. It is assumed in the following interpretations that all of the oolites were originally calcitic or aragonitic. The validity of this assumption is enhanced by the consistency in size and form of the calcitic and non-calcitic oolites and by the abundant evidence of the progressive replacement of calcitic oolites by dolomite and silica.

The reversibility of replacements between chert and the carbonate minerals (Walker, 1962; Swett, 1964) and between calcite and dolomite (Shearman et. al., 1961; Swett, 1964) complicates the interpretation of the diagenetic sequence.

a. Silicification

Silicification ranges in degree from a speckling of the oomicrites and oosparites with small authigenic quartz euhedra to complete replacement producing a dense chert. Where the silica is microcrystalline (possibly due to rapid replacement (Folk and Weaver, 1952)), many details of the original fabrics are preserved in slight textural variations or in relic impurities (Plate 15-D). Where silicification has produced a fine to medium crystalline textured chert (perhaps due to a slower replacement) as adjacent to a silicification "front" (Plate 15-E), the structural details of the original fabric are obscured or lost. Partial silicification has, in certain instances, caused an exclusion of iron which has been reprecipitated in the adjacent carbonate and appears as small cubic euhedra of hematite (probably pseudomorphic after pyrite).

Silica has replaced both calcite (Plate 15-F) and dolomite (Plate 16-A) although evidence for the latter is uncommon.

PLATE 15

A. (D-192) Interpenetrating crystalline network of coarsely crystalline calcite (C) and medium crystalline dolomite (D). The contact relationships indicate that the calcite is later than and replacing the dolomite but the cause for this unusual pattern of calcitization is uncertain. Plane polarized, transmitted light. X 30.

B. (D-160) Interpenetrating patterns of chert (Ch) and dolomite (D). Contact relationships indicate that the chert is replacing the dolomite but the cause of the strangely mottled pattern of replacement is uncertain. Plane polarized, transmitted light. X 30.

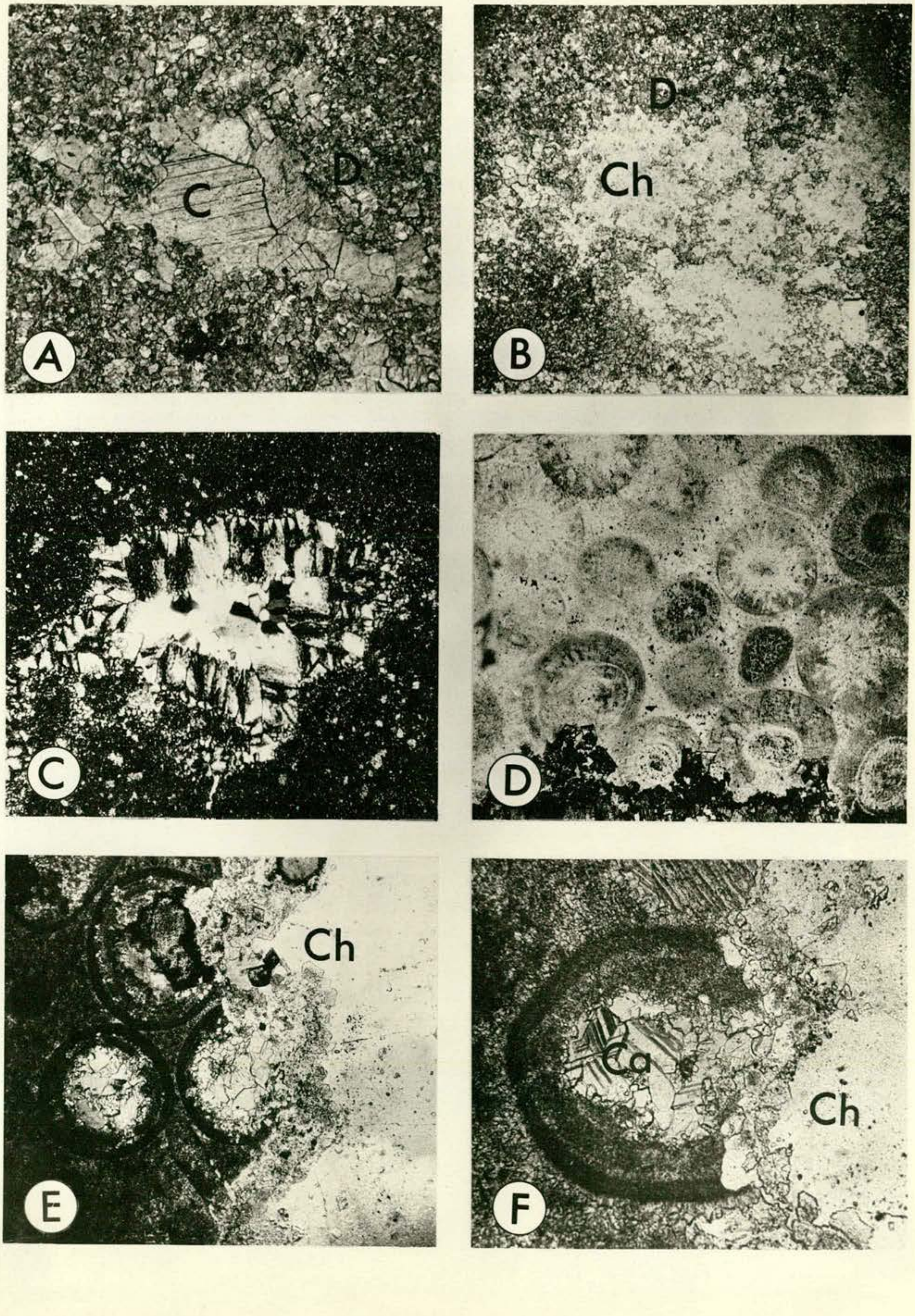
C. (D-1460) Photomicrograph of a completely silicified specimen demonstrating within a single field of view - microcrystalline chert around the margins, chalcedonic (fibrous) chert surrounding the former void space, and drusy chert filling the central cavity. Nicols crossed. X 30.

D. (D-1958) Photomicrograph of a silicified oolitic specimen. In the microcrystalline chert, the structures of the oolites are well preserved in concentrations of relic impurities. Plane polarized, transmitted light. X 20.

E. (D-1702) Partial silicification of an oolitic limestone by rather coarsely crystalline chert (Ch) along a "front" of silicification. Nearly all of the oolitic structures are obscured or lost with this mode of silicification. Plane polarized, transmitted light. X 30.

F. (D-1702) Partial silicification (Ch) of a recrystallized calcitic oolite along a "front" of silicification. Plane polarized, transmitted light. X 50.

PLATE 15



b. Dolomitization

Dolomitization is nearly ubiquitous in the Durness carbonates and, although difficult to prove, it is likely that many, if not all, of the fine-grained dolostones of the lower part of the sequence are authigenic. It can be established with reasonable certainty that the dolomite of the algal horizons and the dolomite of these oolitic horizons is authigenic.

Dolomitization, like silicification, ranges in degree from partial replacement by dolomite euhedra to total alteration of the original limestone. Authigenic dolomite, more than authigenic silica, has a strong tendency toward macrocrystallinity which tends to obscure or destroy original textures (Plate 16-B). It is dubious whether this tendency to increase the coarseness of the crystalline textures during authigenesis should be considered "grain growth" as did Karcz (1964).

Incomplete dolomitization in the oolitic limestones commonly displays the unusual pattern of selective replacement of the oolites rather than the interstitial micrite or sparry calcite (Plate 16-C). Differential weathering of these selectively dolomitized oolites produces a "sandy" appearance on outcrop surfaces (Plate 16-D). An entire oolite is very often replaced by a single crystal of dolomite (Plate 16-E). An interesting phenomenon associated with these single-crystal replacements is the appearance in optic axis sections, under low magnification and with nicols crossed, of a pseudo-interference figure apparently caused by slippage or by overstep growth patterns of the dolomite lattice within the sphere (Plate 16-F). Dolomite crystals outside the oolites or transecting the oolite boundaries do not show this pseudo-interference figure.

The cross-cutting relationships of the authigenic single dolomite crystals to the "drusy" (grain growth?) textures in partly dolomitized sparry oolites establishes that recrystallization of concentric and radial oolites to sparry oolites preceded dolomitization (Plates 17-A and 17-B). The transgression of the oolite boundaries by dolomite rhombohedra (Plate 17-C) and relic concentric and radial structures within single-crystal replacements of the oolites (Plate 17-D and 17-E) confirms both the post-depositional and the authigenic origin of the dolomite.

Dolomite has also replaced previously silicified oolitic rocks (Plates 17-F and 18-A) and generally this second replacement has almost completely destroyed the oolitic textures. Throughout the Durness sequence, dolomite replacement of silica appears as the dominant interaction wherever these two minerals are observed in juxtaposition. Pre-dolomitization chert was probably volumetrically only very slightly more extensive than at present with the later

PLATE 16

A. (D-1702) Partial silicification (Ch) of a previously dolomitized concentric oolite (D). Note the loss of structural detail in the coarsely crystalline chert and the preservation of relic "concentric" ghosts within the dolomitized oolite. Plane polarized, transmitted light. X 30.

B. (D-140) Oolite "ghosts" in a totally dolomitized horizon. The dolomite of the oolites is more finely crystalline than in the inter-oolitic areas possibly reflecting a more rapid replacement of the oolitic material. Plane polarized, transmitted light. X 30.

C. (D-1700) Three adjacent oolites all demonstrating near-total replacement of their oolitic cores by single crystals of dolomite. Planar crystal faces are developed on some of the boundaries within the oolites. Plane polarized, transmitted light. X 30.

D. Weathered surface of dolomitic, oolitic limestone where the selectively dolomitized oolites are raised on the differentially weathered surface and impart a "sandy" appearance.

E. (D-1701) Single-crystal replacement of an oolite (D) and to the left, a sparry calcitic oolite (C). Plane polarized, transmitted light. X 50.

F. (D-1701) Dolomitized oolite (single crystal) displaying a pseudo-interference figure under low magnification and with the nicols crossed. The extinction pattern behaves as a biaxial positive interference figure. (Dolomite = uniaxial negative). Nicols crossed. X 50.

PLATE 16

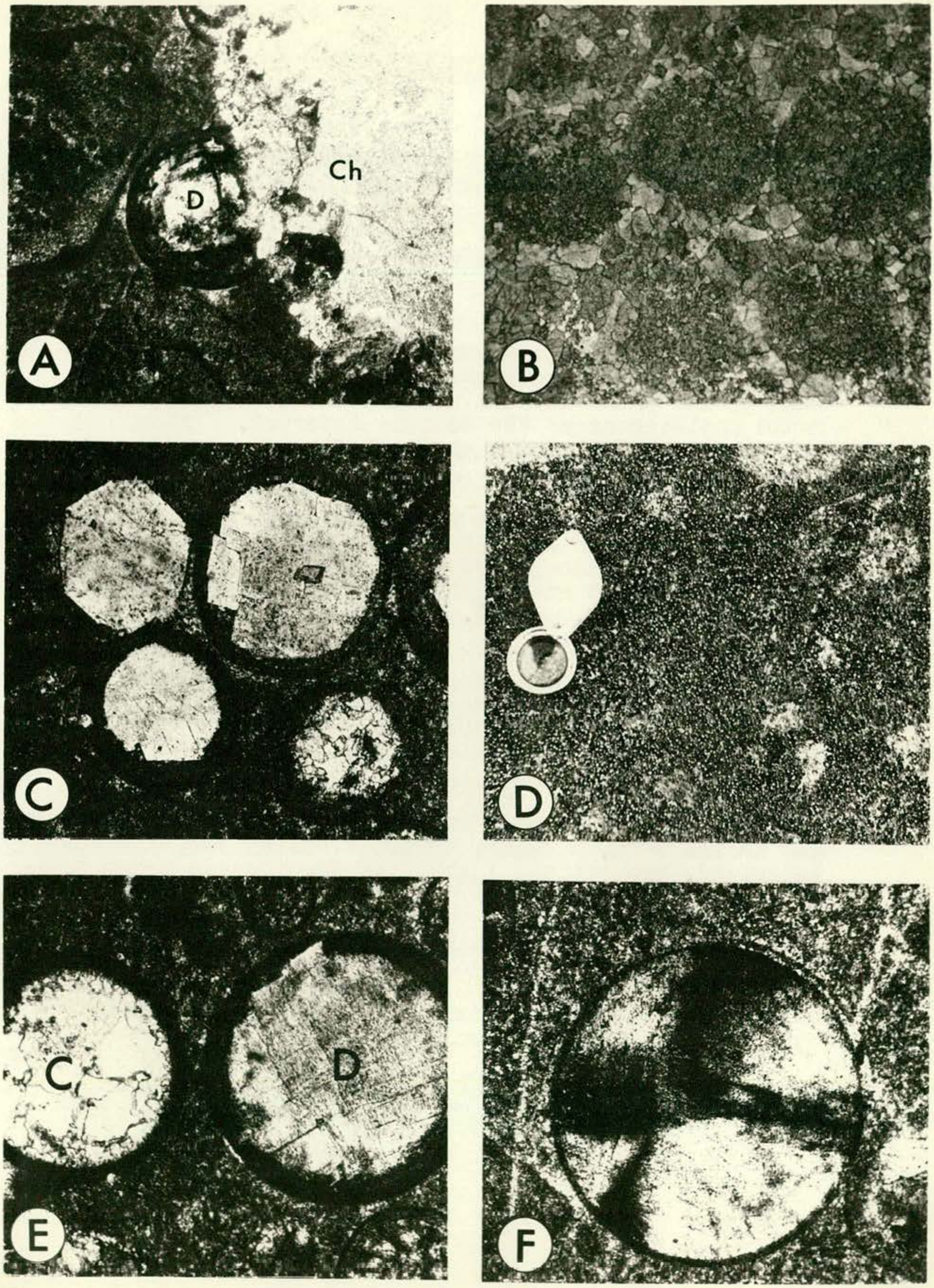


PLATE 17

A. (D-1700) Partial dolomitization (arrow) of a previously recrystallized oolite. Plane polarized, transmitted light. X 60.

B. (D-1701) "Half and half oolite" resulting from partial dolomitization (D) of a sparry oolite (C). Relic micritic nucleus has been preserved through recrystallization and dolomitization. Plane polarized, transmitted light. X 50.

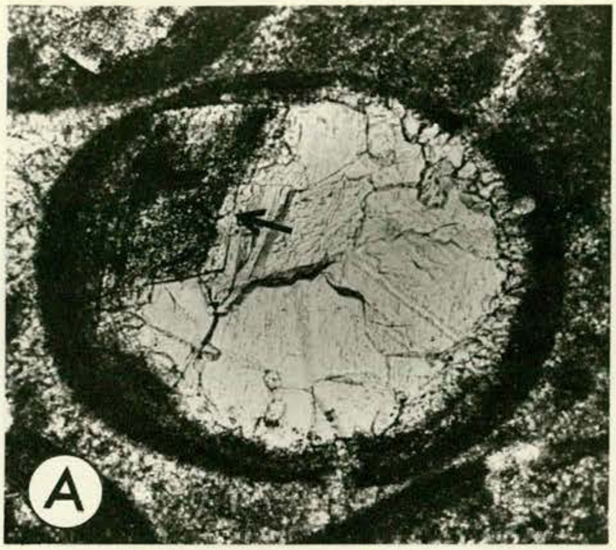
C. (D-1701) Partly dolomitized radial oolite showing a transgression of the calcitic oolite boundary by a dolomite rhomb (D). Plane polarized, transmitted light. X 50.

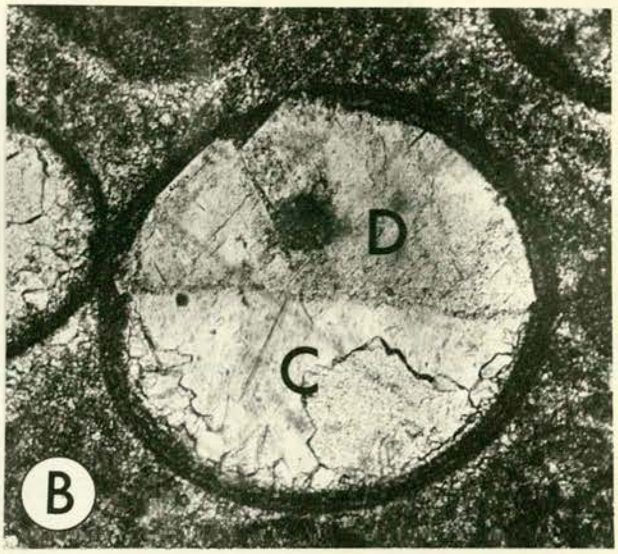
D. (D-1702) An oolite showing relic concentric structures preserved in the single crystal of dolomite which has replaced nearly the entire oolite. Plane polarized, transmitted light. X 60.

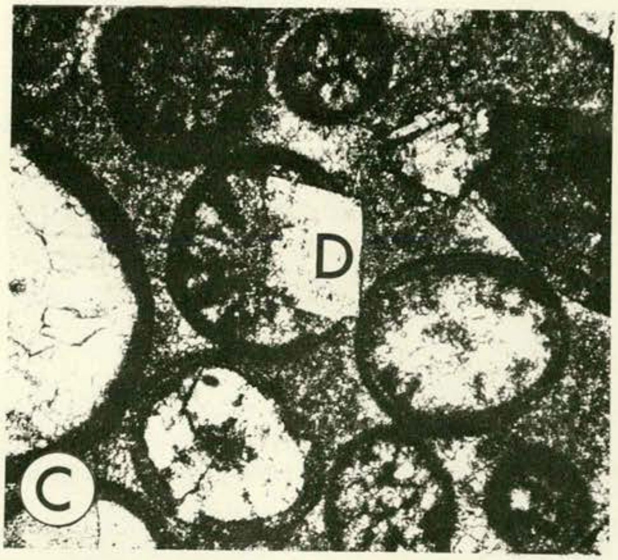
E. (D-1700) An oolite showing a "ghost" radial structure preserved in a single crystal of dolomite which has replaced the oolite. Plane polarized, transmitted light. X 50.

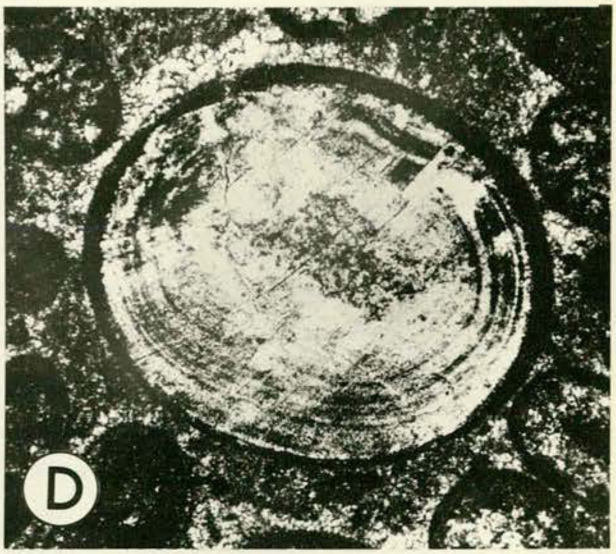
F. (D-1958) Partial replacement of previously silicified oolites by euhedral dolomite rhombs (D). Nicols crossed. X 50.

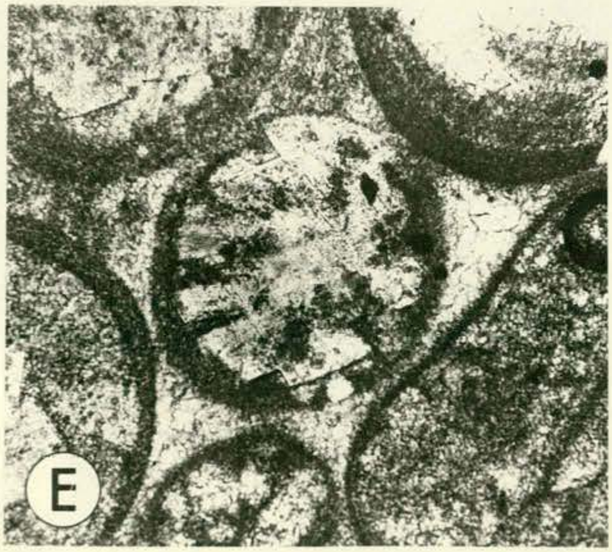
PLATE 17

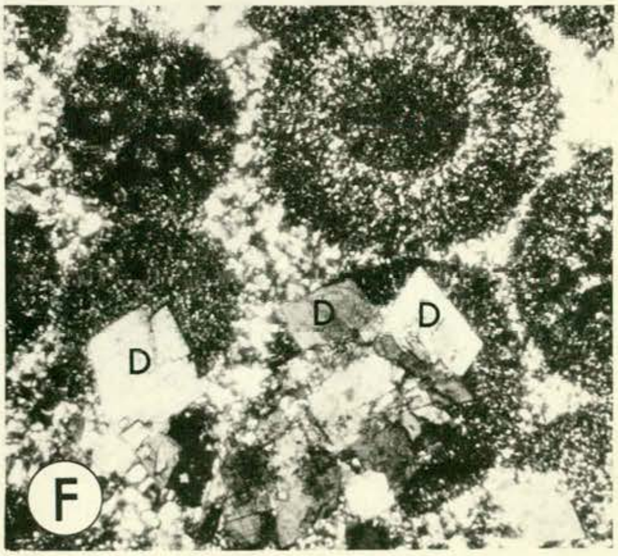












dolomitization of chert being mainly a superficial replacement.

c. Calcitization

Calcitization, as separate from recrystallization, implies an introduction of calcium carbonate as a replacement of another mineral (the inversion of aragonite to calcite being a possible exception (J.H. Taylor, 1964). In the Durness oolitic carbonates, calcite occurs as an authigenic replacement of dolomite (Plate 18-B) and of chert (Plate 18-C).

This writer believes that calcite replacement of both dolomite and silica within these and many other rocks provides a convincing argument for abandoning the term "dedolomitization". The origin and use of the term "dedolomitization" is traced by Shearman et. al. (1961) but despite its ambiguity, they, and several other authors, have used it freely in recent papers.

This author suggests that use of the term "dedolomitization" be discontinued for the following reasons: 1) it requires multiple names for a single replacement process, 2) it is geochemically misleading, 3) it is inconsistent with accepted nomenclature for other replacement processes (e.g. dolomitization, silicification, pyritization, glauconitization, feldspathization, etc. (A.G.I. Glossary of Geology and Related Sciences)), and 4) it is an ambiguous term because several processes including replacement by a horde of authigenic minerals and even solution processes could conceivably be treated as "dedolomitization".

Calcitization of previously silicified oolites is established by the textural and cross-cutting relationships as shown in Plate 18-C. Similar evidence (Plates 17-F and 18-A) has been previously introduced to establish dolomite replacement of previously silicified oolites. The paragenetic or time relationships between dolomitization and calcitization are shown in Plates 18-D and 18-E where it may be seen that the dolomite crystals transect the calcite - chert boundary and must therefore be later than both the silicification and the calcitization. This situation also concisely affords the paragenetic order of three stages of diagenesis as: 1) silicification, 2) calcitization, and 3) dolomitization respectively. However, attempting to propose or devise a geochemical environment which could account for the equal susceptibility of calcite and chert to simultaneous dolomitization, presents a baffling problem for which no solution can be suggested here.

d. Pyritization

Authigenesis involving pyrite is another of the diagenetic phenomena which has left its imprint of the oolitic rocks. Pyritization in the carbonates is not a common phenomenon and the only instance in oolitic rocks was found in the "Grudaidh" member dolostones from the quarry face at Ullapool. There, the authigenic pyrite occurs as small (0.05 mm - 0.25 mm) euhedral crystals that replace previously silicified oolites (Plate 18-F). The pyrite of this horizon can also be established as later than the calcitization stage of diagenesis that has also replaced the chert. Unfortunately, no replacement textures embracing pyrite and dolomite were observed and thus the paragenetic relationships between pyritization and dolomitization remain unsolved. It is somewhat tempting to project the paragenetic relationships of dolomite and pyrite as observed in the "Fucoid Beds" to these carbonates and thereby conclude that the pyrite post-dates the dolomite. This might be true or it might be only partly true, for evidence within the oolitic carbonates at Durness for two periods of dolomitization is clear. It can be safely assumed only that the pyritization has post-dated the first period of dolomitization but not necessarily the second.

3. Sequence of diagenesis

The oolitic horizons of the Durness carbonates clearly demonstrate evidence for several diagenetic events in the history of these rocks including silica replacement of calcite and dolomite, dolomite replacement of calcite and silica, calcite replacement of dolomite and silica and, finally,

PLATE 18

A. (D-1958) Replacement of a previously silicified radial oolite by a euhedral dolomite rhomb. Nicols crossed. X 50.

B. (D-1701) A dolomitized oolite showing post-dolomite calcitization ("dog-tooth crystal (C)). Small relic micritic nucleus precludes the possibility of calcite crystal growth into a pre-dolomite (D) void. Plane polarized, transmitted light. X 50.

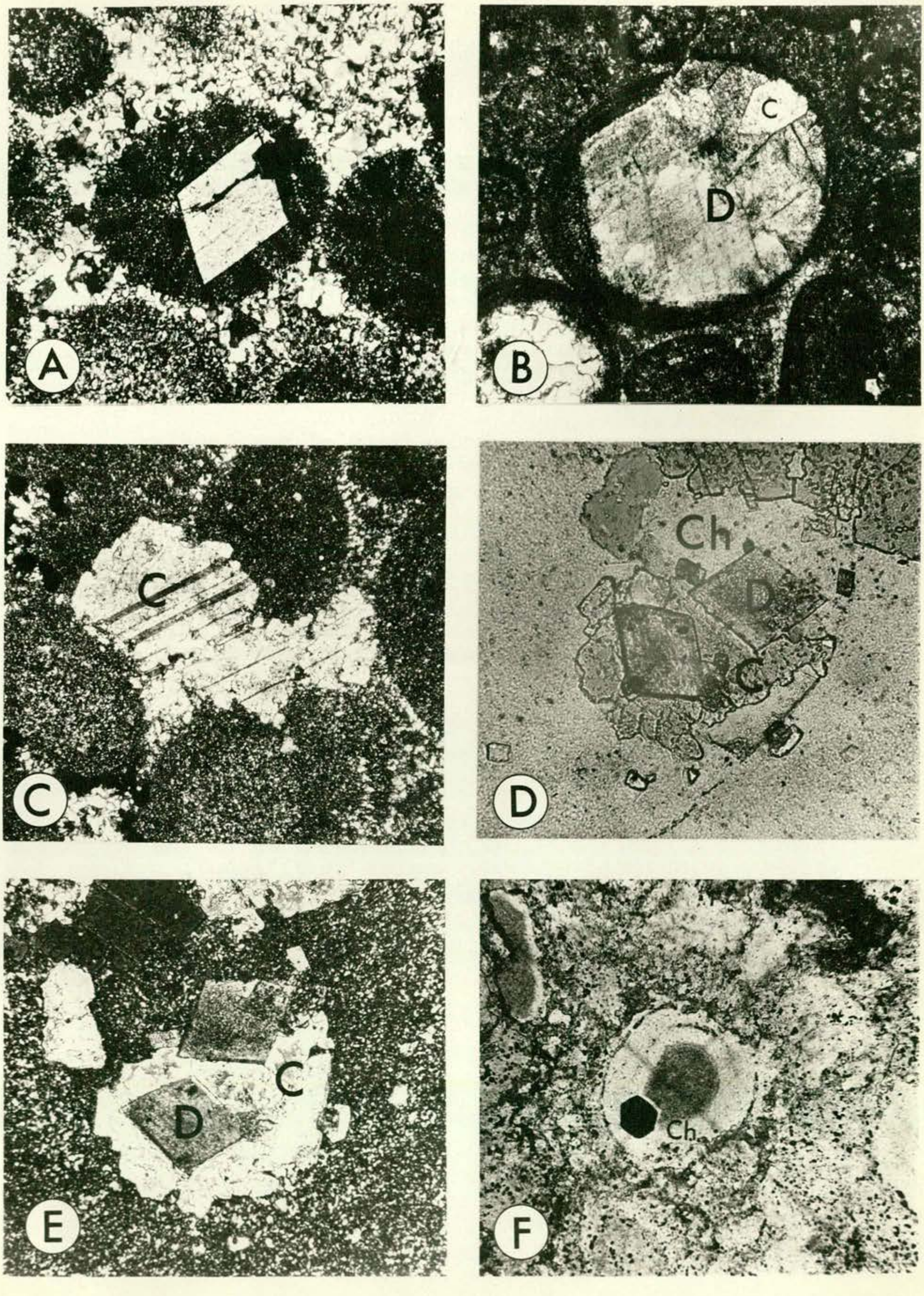
C. (1703) Sparry calcite replacement of previously silicified oolites. Note the cross-cutting relationship of the twinned calcite to ghost oolitic texture. Nicols crossed. X 50.

D. (D-1702) Sparry calcite (C) replacement of previously silicified oolites (Ch) followed paragenetically by dolomite replacement (D) of both the sparry calcite and the chert. Plane polarized, transmitted light. X 100.

E. (D-1702) Paired photomicrograph with preceding description (Plate 18-D) but with nicols crossed. X 100.

F. Euhedral pyrite crystal (black) cutting across the textures of previously silicified oolites in the Gruadaidh member dolostones from the quarry at Ullapool. Plane polarized, transmitted light. X 30.

PLATE 18



pyrite replacement of calcite and silica. In short, nearly every mineral present appears to replace nearly every other mineral with the exception of pyrite. The problem thus faced is to attempt to fit these diverse replacement phenomena into some pattern or scheme which will account for the observed relationships in a meaningful manner. The presence of the oolitic fabrics has considerably simplified, if not made possible, the solution of this problem.

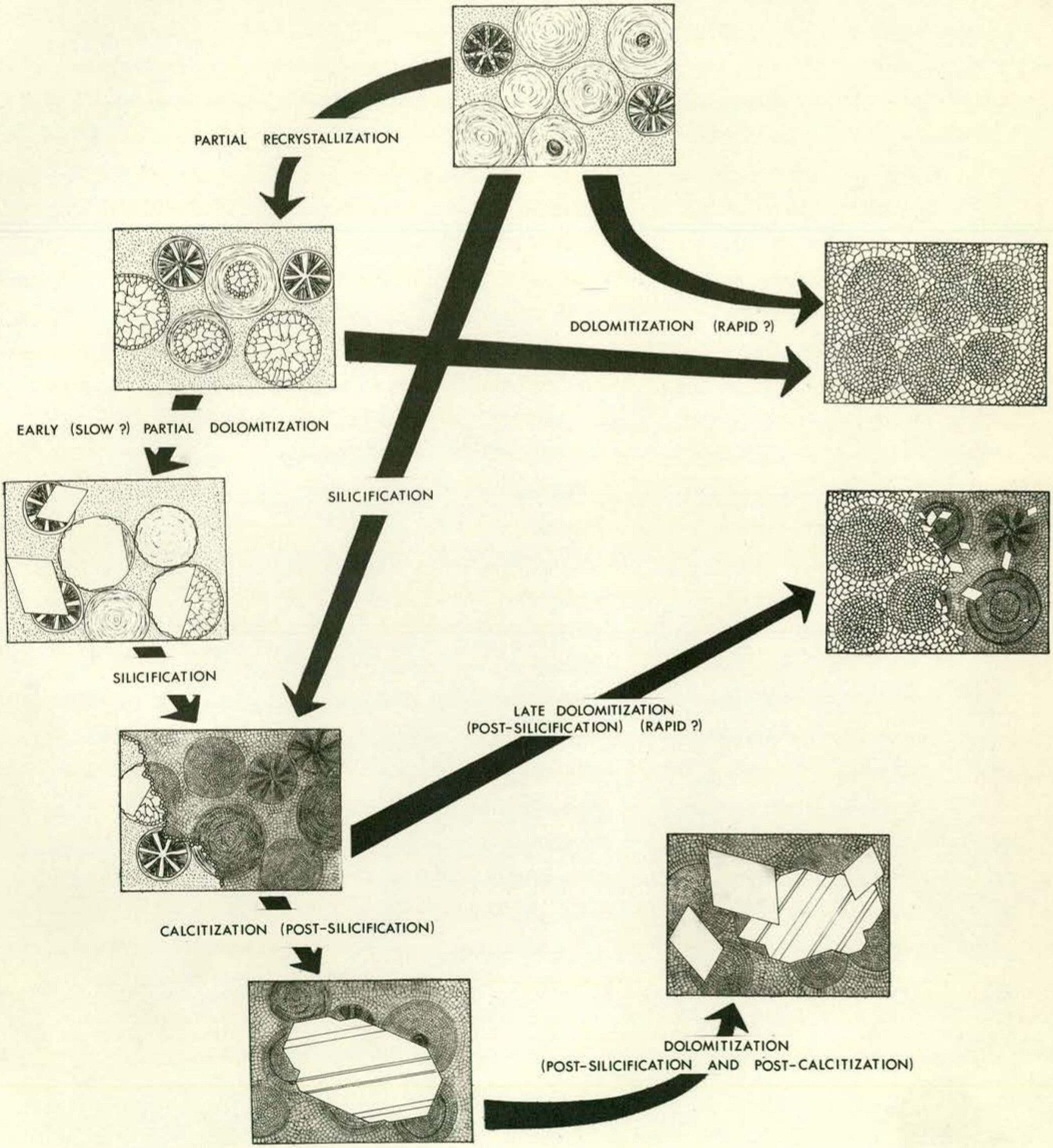
Two assumptions have been made in the following interpretations. First, that a sequence of diagenesis involving the fewest possible replacements is the most probable to have occurred. No defence for this assumption other than a preference for simplicity can be offered. A second, previously defended assumption which seems reasonable, is that all of the oolites were originally deposited as aragonitic (?) bodies (Friedman, 1964) with or without a microcrystalline calcareous matrix (oomicrites or oosparites). This is a necessary and valuable assumption for it often permits the interpretation, by inference, of a stage of diagenesis without direct textural evidence. An example of its value is found in the oolitic cherts where the textural evidence nearly always indicates that silica is the host mineral in its relationships with the carbonates. That is to say, where chert and calcite or dolomite occur in juxtaposition, the textural evidence frequently indicates that the calcite or dolomite have replaced the chert (Plates 18-A and 18-C). From the assumption that the oolites were originally calcitic or aragonitic, it may be inferred that the chert is itself authigenic and that silicification has been frequently succeeded by calcitization or dolomitization.

The assumption of an original calcareous lithology resolves the dilemma with respect to the calcite-silica relationships but does not account for the unusual instance where silica may be shown to have authigenically replaced dolomite (Plate 16-A). This requires that dolomitization has antedated silicification. The only conceivable method by which two diagenetic processes can each antedate the other is for one of the processes to be repeated. Such a repetition has already been suggested as a means of solving the dilemma of the mottled chert nodules. In actuality, the two stages of dolomitization were first recognized in these oolitic horizons and later projected as an explanation for the mottled cherts. The infrequency of textural evidence showing silica replacement of dolomite may perhaps be explained by relics of the early partial dolomitization having served as "centres" for later dolomitization, thus obscuring the majority of circumstances where dolomite might have appeared as the host mineral.

Textural evidence invariably indicates that recrystallization, or whatever other process might have produced the "sparry oolites", preceded both periods of dolomitization and, therefore, also preceded silicification (Plate 17-A). This period of calcite recrystallization or aragonite inversion to calcite is demonstrably separate from the authigenic replacement process of calcitization which has post-dated silicification (Plate 18-C). The cross-cutting relationships of Plate 18-D and 18-E further establish that the later stage of dolomitization which has replaced chert, has also replaced, and is therefore later than, the calcitization calcite.

Except for pyritization, which is a very restricted and localized phenomenon, this completes a possible time sequence which will account for all of the diagenetic fabrics found in the oolitic rocks. A more readily grasped picture of the interrelationships of these various diagenetic events is offered in the schematic diagram of figure VII. In this schematic representation of the paragenesis, it may be seen that several possible short-cuts have been inferred. This inference is only to admit that not all of the oolitic limestones have been subjected to the same or a complete progression involving all of the diagenetic alterations. In fact, no single horizon successfully demonstrates the total diagenetic sequence. The total sequence is approached in the oolitic limestone horizon of the Balnakiel member (D-1700 to D-1702+). This single horizon is, by itself, a remarkable chronicle of the diagenesis for within this unit, it is

FIGURE VII


PARTIAL RECRYSTALLIZATION
DOLOMITIZATION (RAPID ?)
EARLY (SLOW ?) PARTIAL DOLOMITIZATION
SILICIFICATION
SILICIFICATION
LATE DOLOMITIZATION
(POST-SILICIFICATION) (RAPID ?)
CALCITIZATION (POST-SILICIFICATION)
DOLOMITIZATION
(POST-SILICIFICATION AND POST-CALCITIZATION)

possible to observe the results of recrystallization, early partial dolomitization, silicification, calcitization, and late dolomitization processes.

Pyritization, as previously noted, was observed only in an oolitic horizon at Ullapool. Its associations with other authigenic textures establish its time relationships as post-silicification and post-calcitization. No direct evidence can be cited to establish its position with respect to the late stage of dolomitization. It is perhaps likely, from indirect evidence, that pyritization post-dates all of the carbonate - silica diagenetic interactions.

Having established some of the possible interrelationships of the diagenetic processes within the oolitic horizons, it is much easier to approach the interpretive problems in the remaining bulk of the carbonate sequence. Many isolated examples of interpretive problems in the carbonate fabrics can be cited. One common example of a difficult guest-host problem is found where "drusy" or chalcedonic chert textures interpenetrate with sparry dolomite or sparry calcite fabrics (Plate 19-A and 19-B). In these circumstances, it is often difficult to determine if drusy or chalcedonic chert growth has replaced the sparry carbonates or if the drusy and chalcedonic cherts partly filled voids that were later filled by the carbonates. Evidence for the latter possibility is sometimes present where sparry calcite and dolomite void fillers have partly replaced the marginal drusy or chalcedonic cherts

(Plate 19-C and 19-D). The knowledge that periods of calcitization and dolomitization have post-dated silicification makes the interpretation of such textures rather easy but, without this knowledge, they can be extremely puzzling textures.

The significant point concerning the diagenetic phenomena throughout the succession is that no textural or mineralogical situations have emerged that cannot be explained by the proposed sequence of diagenesis or some portion thereof as established in the oolitic horizons. This is certainly not to claim that the distribution and fabrics of the diagenetic phenomena or the causes of the diagenesis can be explained but it is possible to explain all of the observed textural patterns in terms of the paragenetic scheme.

PLATE 19

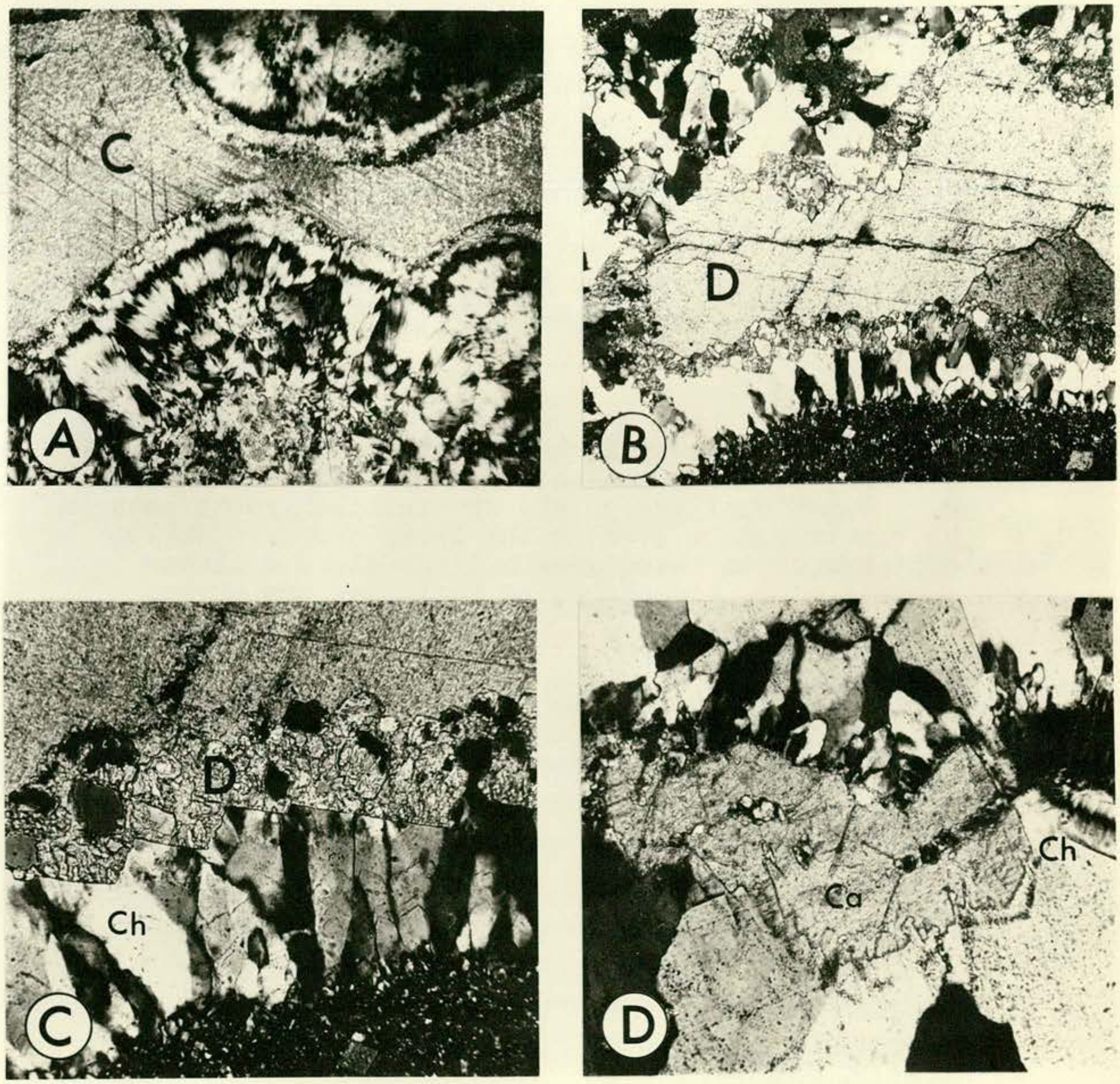
A. (D-2350) Chalcedonic chert in a pattern suggesting a pre-existing void. The chalcedony appears to have only partly filled the void which was later completely filled by sparry calcite (C). Nicols crossed. X 50.

B. (D-1426) Drusy chert in a pattern suggesting development peripheral to a pre-existing void. The drusy chert appears to have only partly filled the void which was later completely filled by sparry dolomite (D). Nicols crossed. X 30.

C. (D-1426) Same specimen as shown in Plate 19 B but under higher magnification to demonstrate the cross-cutting relationship of the dolomite "void filler" to the drusy chert. Faint outlines of the crystalline growth pattern of the quartz can be seen in the drusy chert. The dolomite truncates these growth lines along a planar contact and must, therefore, be replacing the chert. Nicols crossed. X 80.

D. (D-2350) Zone of impurities in chalcedonic chert (Ch) can be traced continuously through calcite (Ca) that has later replaced the chalcedony. Nicols crossed. X 80.

PLATE 19



V. SUMMARY

A. CONDITIONS OF DEPOSITION

It seems most probable that the Cambro-Ordovician succession of north-west Scotland represents the deposits of a slowly transgressive marine environment. The erosional unconformity at the base of the sequence is perceived as the product of marine erosional processes which truncated the pre-existing Lewisian metamorphic complex and the gently folded Torridonain arkosic sediments. The Lower sandstone member of the Eribol sandstones, immediately above the unconformity, is interpreted as a near-shore, sand facies deposit with conspicuous current bedding features resulting largely from the long-shore migration of current-induced megaripples.

The "Pipe Rock" member of the Eribol sandstones was probably deposited under conditions very similar to the Lower sandstone member but with slightly less pronounced current movements. The deposit is distinguished by the occurrence of numerous burrows of suspension-feeding organisms (possibly annelids or phoronids) within the sediments. The mineralogical and textural maturity of the "Pipe Rock" member and the unaltered character of feldspar grains indicates sufficient energy to accomplish considerable further reworking and abrasion of the sands in order to account for their increase in the proportion of quartz/feldspar. The lack of current-bedding may represent only a slight energy decrease in the environment and indicate only that sand waves were no longer migrating with a sufficient sand supply to produce current-bedded deposits.

The "Fucoid Beds" member of the An t-Sron formation perhaps reflects a deepening of water or may simply indicate deposition on an ever broadening shelf and at a greater distance from the detrital sources, thus causing a proportionate decrease in grain size and in the ratio of detrital components to cement. The common occurrence of Planolites trace fossils and ripple-induced, small-scale, current-bedding within the "Fucoid Beds" suggests conditions of deposition involving shallow to moderate depths of water.

The "Serpulite Grit" member of the An t-Sron formation displays a very high index of sorting and of mineralogical maturity. The increase in grain size and the conspicuous current bedding possibly indicates a brief marine regression. The persistence of this relatively thin sandstone over the total length of outcrops from Durness to the Isle of Skye does not support a freak or local change of the environment as an explanation of this orthoquartzitic sandstone. Whatever caused the environmental shift and produced the "Serpulite Grit" sandstone, seems to have been again abruptly reversed and, in the succeeding sediments, to have resumed the transgressive depositional progression.

The thick succession of carbonates which follows the An t-Sron formation appears to reflect biochemical and possibly chemical carbonate deposits in a continually subsiding depositional area (basin or shelf ?). The rate of deposition seems to have roughly paralleled the rate of subsidence with the result that most of the carbonates were probably deposited in an environment as shallow if not more shallow than the clastic sediments below. Shallow conditions of deposition for the carbonates are indicated by the occurrence of oolitic and algal stromatolite deposits interspersed throughout most of the sequence.

Although the bulk of the carbonate succession is now dolostone or dolomitic, the following arguments favour an original calcitic or aragonitic composition for all of the Durness carbonates: 1) considerable, although subordinate, amounts of limestone are scattered in variable patterns throughout the upper two thirds of the carbonates, 2) The limestones show abundant textural evidence for authigenic replacement by dolomite, 3) Oolitic, algal, or macrofossiliferous horizons occur throughout virtually the total carbonate sequence (Modern oolites, even in hypersaline conditions, are calcitic or aragonitic (Eardley, 1938). 4) There is very little indisputable evidence in favour of a primary origin for most of the dolostones common throughout the geological column (Fairbridge, 1957) although primary dolomite is apparently possible under "unusual" conditions (Alderman and Skinner, 1957; Wells, 1962; Curtis, 1963). 5) Microscopic textural evidence within the thick units of dolostone of the Durness carbonates suggest replacement or recrystallization (e.g. dolomite replacement of detrital quartz grains), 6) Chert nodules throughout the carbonates show textural evidence indicating dolomite replacement of the silica. In as much as the cherts are considered authigenic, so must be the dolomite that replaces them.

B. DIAGENESIS

Making reasonable allowances for unresolved "breaks" or missing gaps from the composite stratigraphic measurements, the Cambro-Ordovician succession measures slightly more than 4500 feet thick. Throughout the succession, pore space has been reduced to negligible proportions suggesting that the original deposits may have been 20 to 50 percent thicker than the present sequence. Pore space has been lost principally by pressure solution, solution and redeposition, and recrystallization processes.

Authigenesis in the sandstones consists mainly of ?post-compaction, syntaxial overgrowths on quartz and feldspar grains. Minor episodes of authigenic replacements by dolomite, pyrite, leucoxene, and glauconite were also noted (mainly in the uppermost units of the sandstones).

The "Fucoid Beds" and "Serpulite Grit" members show textural evidence of authigenic dolomite, pyrite and quartz. Geochemical studies have revealed an extraordinary ?authigenic potash enrichment of the siltstones and shales of the "Fucoid Beds" member which yield analytical results indicating K_2O contents as high as 10 - 11 percent of the total rock or as high as 12 percent of the detrital fraction of the rocks.

The Durness carbonates have been subjected to a complex history of authigenesis involving one or more episodes of dolomitization, silicification, calcitization, and minor pyritization. If the arguments for the original calcareous composition of the carbonates is accepted, they establish by implication that dolomitization has been the most widespread of the diagenetic alterations. Two episodes of dolomitization may be demonstrated but the relative importance of each stage is problematical. The later episode of dolomitization was probably responsible for much of the dolomitization of the mottled dolostones and was apparently responsible for the replacement of cherts. Other than these restricted instances, it is impossible to ascertain which of the episodes of dolomitization has been responsible for the dolostones.

Arguments similar to those advanced for the authigenic nature of dolomite may also be advanced to support a secondary origin for the cherts. 1) The textural relationships of the carbonates and silica, 2) sedimentary structures that may be traced through chert nodules, 3) silicified fossils, 4) silicified oolites, 5) silicified algal structures, and 6) the lack of direct evidence for primary cherts.

Post-silicification episodes of calcitization and dolomitization have probably reduced the volume of chert only very slightly and of these two processes, the dolomitization has certainly been the most significant. Post-silicification calcitization is not uncommon but, in terms of the bulk composition of the carbonates, its effects have probably been negligible.

The paragenetic relationships of the recrystallization and authigenic phenomena are remarkably constant and may be summarized as the following sequence.

1. Recrystallization
2. Dolomitization
3. Silicification
4. Calcitization
5. Dolomitization

? 4.-6. Pyritization

Any one or more of these stages of diagenesis may be by-passed in a particular lithology but the evidence accumulated from extensive searching of numerous thin sections never contradicts the proposed order or some portion thereof.

The author feels, perhaps understandably, that the study of the Cambro-Ordovician sequence of the northwest Highlands of Scotland has been extremely rewarding by its considerable extension of his knowledge of the primary and diagenetic features of ancient sediments. The study has also hopefully been productive in its contributions to the interpretations and distinctions of the sedimentary and diagenetic aspects of these rocks in particular and of sedimentary rocks in general.

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VII. APPENDIX

Eden Grove
Bond

COMPOSITE MEASURED SECTION
OF
THE ERIBOL SANDSTONES
AND
THE DURNESS CARBONATES

(see also graphic section in pocket)

The vertical stratigraphic column of the Cambro-Ordovician sequence of the Durness region as described on the following pages is a composite of measured sections from seven related, although geographically separated, areas. In certain instances, the correlation between areas is virtually certain and in other cases no evidence of a direct correlation can be offered. Where there is doubt, an arbitrary stratigraphic interval has been assigned to the ?missing section and the problems recorded. These arbitrary assignments of footage to "missing sections" have enabled the author to consider the sequence as a continuous section in terms of description and sampling procedures. It perhaps has the slightly misleading consequence of making the composite thickness appear some 400 feet thicker than the actual measured rock units. These missing sections do not however account for the discrepency between this author's total thickness and the thickness attributed to the sequence by Peach et. al. (1907).

The methods of measurement are reviewed in the introductory statement of the thesis on Methods of Study.

The lower cross-bedded sandstones were measured and described from an area approximately 1.25 miles southwest of the extreme southwest end of Loch Eribol. The described units start at NC 376541 and the section follows up the stream channel to the base of the cliff, then is measured directly up over the cliff face to the lower limit of the

Pipe Rock member at NC 375537. The base of this portion of the section rests nonconformably on the Lewisian gneiss and the contact is well exposed in the stream bed.

The described section of the Pipe Rocks member is not in direct continuation with the basal cross-bedded sandstones for a more complete and easily traced sequence is found approximately one mile to the south of the lower sandstone section. The "Pipe Rock" section is started at the contact with the cross-bedded units at NC 379522. The section then passes southeast up over a hill and ends at NC 374519. There is some folding and deformation of the strata in these units but the sequence is virtually complete.

The best exposed section of the An t-Sron formation was measured approximately five miles to the northeast of the "Pipe Rock" section. The sequence is exposed along the coast at An-t-Sron on the east side of Loch Eribol. The contact between the Pipe-Rock sandstones and the "Fucoid beds" member is well exposed near the crest of an anticlinal fold at NC 440581. A continuous section passes from there upward stratigraphically along the shore to the southwest. This section includes all of the An t-Sron formation and the Grudaidh member or the lower 196' of the carbonates.

There is no known locality where a continuous section can be positively measured or correlated in the units immediately above the Grudaidh dolostones. An arbitrary "missing section" unit of 100' is introduced here to account for this discontinuity and the measurement of the carbonate sequence is resumed on the extreme western margin of the outer island of Eilean Dubh (NC 373691). The section then follows continuously along the south shoreline of Balnakiel Bay where nearly 2000' of carbonates are continuously exposed.

Above the Balnakiel Bay section there is, once again, some question concerning the correlation with stratigraphically higher units and another 100' "missing section" unit is invoked before proceding with the measurement near the northeast end of Loch Borralaidh (NC 385677). From this point, the section passes stratigraphically upward to the southeast to NC 390671, then moves across the valley to NC 392674 and hence on to NC 398676. Between NC 392674 and NC 398676, the exposures of the sequence become progressively poorer until another missing section and a questionable correlation is required.

The highest parts of the sequence are measured and described from the sequence starting in the small quarry on the west side of the highway at NC 394672 and passing across the highway to outcrops on the northern shoreline of Loch Caladail.

Exposures terminate in a field at the northeast end of Loch Caladail and are superposed there by Moine schists.

The sample labels have been divided into four separate groups on the basis of the measured sections. The "Q" series include all of the lower sandstones. The "P" series are the pipe-rock sandstones, the "I" series represent the An t-Sron formation ("Fucoid beds" and "Serpulite grit") lithologies and the "D" series includes all of the carbonate sequence. As outlined in the introduction, the sample numbers in each series indicate the measured stratigraphic interval between the sample and the base of that particular series (e.g. sample, thin section, or photomicrograph D-1405 refers to a sample taken 1405' stratigraphically above the base of the carbonate section).

COMPOSITE MEASURED SECTION

DURNESS CARBONATES

DURINE MEMBER

4302'	Upper limit of exposures. Beyond this is a field at the northeast end of Loch Caladail. Along the east margin of this field are the metamorphic Moine schists that have overthrust the carbonates. There might conceivably be another 200 to 300 feet of carbonates beneath the field but it is impossible to measure or sample.
4287' - 4302'	Dolostone, medium to light grey. Highly fractured.
4247' - 4287'	Dolostone, dark grey, highly fractured.
4241' - 4247'	Dolostone and limestone, thinly interbedded, highly contorted and folded, some brecciation, some chert, occasional distinct units of limestone or dolostone up to 5" thick.
4236' - 4241'	Dolostone, medium grey color, medium to finely crystalline.
4230' - 4236'	Limestone, medium - dark grey, black chert nodules and white drusy quartz nodules, thin bedded, dolomitic in part.
4208' - 4230'	Dolostone, mainly light grey, some bands of medium grey dolostone, medium crystalline, occasional siliceous bands. Upper 5' very thinly bedded.
4207' - 4208'	A band of chert nodules (mainly drusy quartz) generally less than 1" diameter. The centers of these tiny drusy nodules are often filled with sparry calcite.
4132' - 4207'	Dolostone, light grey color, some units of medium grey color, medium crystalline, rather thinly bedded, highly fractured, some cherty bands. At 4172' there is an 8" band of tiny (1" or less) drusy quartz nodules, the crystal growth patterns may be seen on the weathered surfaces of these nodules.

4054' - 4132'	Covered interval occupying a rather prominent valley extending for some distance northeast from Loch Caladail. Traversing the valley there is a change in dip from 41° on the west to 31° on the east. The combination of the straight valley form and the dip change suggest a fault at this point in the section. The lithologies also show a change from dolostones on the west and limestones on the east (stratigraphically higher) which indicates that no part of the section immediately below is being repeated but omission of some section is very possible. Although a fault is likely, this covered interval was traversed using a dip angle of 36° to calculate the unit thickness.
4037' - 4054'	Mostly covered, occasional outcrop of medium grey dolostone near top.
4023' - 4037'	Dolostone, medium dark grey, medium crystalline, bands of nodular cherts becoming less frequent toward the top of the unit. Dolomite somewhat lighter colored toward top.
4017' - 4023'	Covered interval.
4000' - 4017'	Dolostone, medium grey, medium crystalline, with bands of nodular cherts. (small 1" nodules).
3986' - 4000'	Dolostone, in interbedded but separate units of medium grey color and units of a light grey color. Brecciated with considerable calcite in the brecciated zones. Slightly cherty towards top.
3978' - 3986'	Limestone, pink, limpy, dolomitic, numerous vugs filled with very coarsely crystalline calcite make up a large portion of the rock.
3953' - 3986' FAULT	At 3953' at the edge of the Loch, a highly brecciated and fractured zone with considerable calcite, pink and coarsely crystalline is encountered. This is almost certainly a zone of some displacement. An arbitrary thickness is assigned to the undetermined displacement.
3810' - 3953'	Dolostone, light grey, occasional fine flecks of chert less than 1/2" in diameter. Cherty bed at 3884'.

3790' - 3810'	Dolostone, light grey and medium grey, finely crystalline, rather cherty.
3762' - 3790'	Covered interval.
3765' - 3762'	Dolostone, medium to dark grey, slightly mottled, cherty.
3736' - 3756'	Covered interval.
3734' - 3736'	Dolostone, dark grey, thinly bedded.
3731' - 3734'	Dolostone, medium-light grey, thinly bedded.
3728' - 3731'	Dolostone, dark grey mottled with light grey dolomite, stylolitic, coarsely crystalline, highly fractured.
3716' - 3728'	Dolostone, medium grey bands and light grey bands 2' - 4' thick, medium to finely crystalline, thinly bedded, highly fractured.
3692' - 3728'	Dolostone, interbedded light and medium grey lithologies, thinly bedded, highly fractured but showing no development of cleavages. A slightly cherty horizon occurs between 3712' and 3716'.

DURINE MEMBER

3592' - 3692'	Missing section. Above 3592' in continuation with the section below, the outcrops become so sparce that they are not worth while to include. A more reasonable section is available starting in the quarry on the west side of the highway just opposite the north end of Loch Caladail. In the quarry, different lithologies from those below are encountered and there is no possible method by which the amount of missing section can be computed with any degree of accuracy. An arbitrary thickness of 100' is assigned in order to record the missing section and yet allow continuation of the sampling and stratigraphic sequence. The 100' figure has no significance and the actual amount of missing section may be much more or far less.

CROISPHUILL MEMBER

3537' - 3592'	Dolostone, medium to dark grey, highly fractured, medium to finely crystalline, mottled in a blotchy pattern in some of the darker portions.
3532' - 3537'	Dolostone, medium light grey, unit contains abundant tan chert, dolomite textures are medium crystalline.
3485' - 3532'	Limestone, dolomitic or dolostone, calcareous, light grey, thin bedded, occasional bands of nodular chert up to three inches thick, fine to medium crystalline.
3385' - 3485'	<u>Missing section</u>. At 3385', a brecciated fault zone is encountered with the rocks beyond the fractured zone possessing quite a different lithologic character. The rocks above this horizon are more characteristic of the dolomitic limestones designated as the Group VII units of Peach and Horne. These outcrops are the only known location where the two lithologies may be observed in juxtaposition and thus it is impossible to estimate the amount of section missing at this point. The arbitrary unit thickness of 100' is again assigned here in order to maintain the continuity of the measurement and sampling procedures. The 100' figure has no significance.
3314' - 3385'	Mottled dolomitic limestones with the two mineralogies distinctly separate. The limestone portions are generally light to medium grey and the dolostone portions are buff-pink. Within this unit the dolomitic portions are sometimes a dark grey contrasted with a lighter grey calcareous portion.
3292' - 3314'	Mottled dolomitic limestone, limestone is light grey and the dolomitic portions are buff-pink, strongly cleaved.
3257' - 3292'	Covered interval.
3205' - 3257'	Mostly covered. Very sparse outcrops are consistent in lithology with the unit below.

3147' - 3205'	Dolostone, medium-dark grey, medium to finely crystalline, considerable chert in some portions. Some of the chert nodules are very minute (less than 1/2").
3120' - 3147'	Covered interval.
3117' - 3120'	Dolostone, medium to finely crystalline, medium-dark grey, numerous small chert nodules with some light grey microcrystalline cherts and some white drusy cherts.

CROISPHUILL MEMBER

BALNAKIEL MEMBER

3030' - 3117'	Mottled dolomitic limestone with the two mineralogies remaining separated in the mottled pattern. Vaguely bedded in units 4" to 1' thick. Bands of nodular cherts up to 2" in diameter in several horizons.
3025' - 3030'	Covered interval.
3020' - 3025'	Mottled dolomitic limestone with the separate lithologies responsible for the mottled pattern. Somewhat cherty.
3007' - 3020'	Limestone, pinkish light-grey, somewhat cherty. Highly cleaved.
2987' - 3007'	Covered interval.
2967' - 2987'	Limestone, light grey, somewhat cherty, dolomitic, separation mottling with dolomite in some horizons, some thinly bedded dark grey dolomites which are laterally discontinuous. Sparse outcrops only.
2963' - 2967	Covered interval.
2959' - 2963'	Mottled dolomitic limestones with separation of the lithologies causing the mottled pattern. Abundant jasperoid chert in the upper 2' arranged in thin nodular bands spaced 1' to 6" apart.

2955' - 2959'	Limestone, thinly bedded, pinkish light grey color.
2950' - 2955'	Mottled dolomitic limestone showing separation mottle pattern as above. Somewhat cherty.
2842' - 2850'	Dolostone, medium grey, medium crystalline, slightly mottled.
2938' - 2942'	Mottled dolomitic limestone as above units except perhaps a slightly lighter grey color.
2936' - 2938'	Limestone, medium light grey color, finely crystalline.
2920' - 2936'	Covered interval.
2918' - 2920'	Dolostone, medium dark grey color, some fine streaks of white dolomite. Thin nodular chert bands (nodules less than 1" diam.).
2902' - 2918'	Covered interval.
2847' - 2902'	Dolostone, medium dark grey, numerous thin streaks of white dolomite coursing through the darker matrix. Thin nodular and bedded cherts, generally white are common. A thin bed of jasperoid chert occurs at 2982'.
2834' - 2847'	Covered interval.
2813' - 2834'	Dolostone, medium grey color, mottled, medium crystalline texture.
2976' - 2813'	Covered interval.
2784' - 2796'	Dolostone, medium grey color, mottled with medium to light grey dolostone.
2777' - 2784'	Covered interval.
2773' - 2777'	Dolostone, medium grey color, in 6" beds with showing slight color variations, medium crystalline texture.
2767' - 2773'	Dolostone, slight grey color, thinly bedded, a 1" brownish colored chert band at the top.
2764' - 2767'	Dolostone, medium grey color, mottled.

2760' - 2764'	Dolostone, light grey color, thinly bedded.
2754' - 2760'	Dolostone, dark grey color, mottled.
2752½' - 2754'	Dolostone, light grey color, thinly bedded.
2750' - 2752½'	Dolostone, dark grey, mottled.
2744' - 2750'	Dolostone, light grey, thinly bedded, some nodular cherts present.
2742' - 2744'	Dolostone, dark colored faintly mottled, no chert.
2733' - 2742'	Dolostone, medium-dark grey, mottled, a cherty unit in which the cherts also exhibit the mottling of the host carbonates. This would suggest that the mottled pattern must antedate silicification ??
2723' - 2733'	Dolostone, medium-dark grey, mottled. Very little chert.
2722' - 2723'	Dolostone, medium-light grey color, faintly mottled.
2687' - 2722'	Dolostone, medium dark grey, faintly mottled. Several lighter colored thinly bedded units interbedded with mottled units. Small white drusy chert nodules 2' above base, often with hollow centers. A nodular cherty zone between 2712' and 2714'.
2684' - 2687'	Dolostone, medium light grey, thinly bedded, finely crystalline.
2667' - 2684'	Dolostone, medium dark grey, some nodular chert present, faintly mottled, and exhibiting thin irregular streaks of white dolomite which traverse the mottled patterns.
2655' - 2667'	Mottled dolomitic limestone showing the sharp separation of the two mineralogies to produce the mottling. The limestone is light grey in color while the dolostone is somewhat darker.
2652' - 2655'	Limestone, light grey, finely crystalline, ? dolomitic.

2625' - 2652'	Dolostone, medium grey in part, mottled with lighter grey dolostone. Some chert present in the lower 10'.
2602' - 2625'	Mottled dolomitic limestone showing the distinct separation of the two lithologies. The lower two feet are locally unmottled limestones.
2601' - 2602'	Dolostone, buff-pink, medium crystalline texture.
2569' - 2601'	Mottled limestone, light grey and dolostone, medium grey mixed in the "lumpy" irregular pattern of mottling. Some chert present near the top of this unit. Strongly cleaved in the lower part of the unit.
2568' - 2569'	Dolostone, light grey, finely crystalline texture, thinly bedded.
2561' - 2568'	Mottled dolomitic limestone with the mottling due to the separation of the two lithologies. This horizon exhibits a particularly high relief on the weathered surface with the dolomitic portions extending as much as 3/4" beyond the calcareous portions.
2552' - 2561'	Limestone, light grey color, dolomitic in part but in a rather lacy intergrowth rather than the mottled patterns as above.
2549½' - 2552'	Dolostone, light grey, thinly bedded, coarsely crystalline.
2548' - 2549½'	Dolostones, medium grey color, slightly mottled, some small flecks of chert.
2545' - 2548'	Dolostone, light grey color, thinly bedded, coarsely crystalline.
2521' - 2545'	Dolostone, medium dark grey, mottled, coarsely crystalline, some chert at 2533'.
2507' - 2521'	Dolostone, medium-dark grey, conspicuously mottled. Several bands of white nodular cherts. (Several faults are crossed at this level but the correlation based on these cherty bands seems quite reasonable.)

2470' - 2507'	Dolostone, coarsely crystalline, medium to light grey becoming darker in color and more strongly mottled towards the top. The mottling here is very similar to that in the "Leopard Stone" below but more subdued in its color variation.
2461' - 2470'	Dolostone, coarsely crystalline, medium grey, small irregularly shaped white chert nodules in bands spaced approximately 6" apart.
2430' - 2461'	Dolostone, medium grey color, coarsely crystalline, some horizons of thin nodular grey-white cherts. Intraclast horizon at 2433' (very coarse sand sized intraclasts), with both the clasts and the matrix or cement totally dolomitic.
2417' - 2430'	Covered interval.
2404' - 2417'	Mottled dolomitic limestone with mottling being expressive of the irregular separation of the limestone and dolostone lithologies.
2397' - 2407'	Covered interval.
2387' - 2397'	Dolostone, coarsely crystalline, dark grey color, massive, occasional white chert nodules in slightly favored horizons.
2377' - 2387'	Covered interval.
2370' - 2377'	Mottled dolomitic limestone as above and below.
2369' - 2370'	Dolostone, dark grey color, massive unit with a nodular cherty band near its center.
2330' - 2369'	Mottled dolomitic limestone with the mottling being an expression of the irregular patchy distribution of the separated dolomitic and calcareous lithologies. Sparsely fossiliferous.
2307' - 2330'	Covered interval.
2227' - 2307'	Sparse outcrops but a consistent lithology. All are the mottled dolomitic limestones as described above. Occasional chert nodules are present. A northeast-southwest trending wall is crossed at the top of the hill at the northeast end of Loch Borralaidh at the measured interval of 2262'.

2151' - 2227'	Mottled dolomitic limestone. The limestone and dolostone portions of this rock type are distinctly separated in irregular mottled patterns bordered by sharp boundaries. Weathering of this rock produces an extremely rough surface with the dolomitic portions raised in positive relief as much as 1" but more commonly ½". The mottling, although very irregular, does display a distinctly bedded appearance when viewed from a distance.
2146' - 2151'	Limestone, partly dolomitic, light grey color, slightly mottled appearance and faintly lumpy on weathered surface.
2136' - 2146'	Limestone, dolomitic, light and medium grey mottled with very numerous jasperoid chert nodules in 1" to 2" thick bands spaced ½" to 6" apart. The cherts produce a very strongly bedded appearance in the rock.
2112' - 2136'	Limestone, light grey and medium grey and a pinkish grey, dolomitic slightly mottled. This unit is very brecciated near its base with most of the pinkish lithologies in this zone of brecciation. Carbonates appear moderately cleaved.

BALNAKIEL MEMBER

2102' - 2112' MISSING SECTION 10'?	At 2102' a fault is encountered which cuts obliquely across the stike of the strata. the rocks on either side of this fault zone do not correlate by any perceivable displacement. It is quite certain that none of the stratigraphy measured below is being repeated above the fault for the nodular cherty horizon above the fault (2136' - 2146') is absent from the rocks below the fault. The section as followed, takes the lower stratigraphic horizons to their highest exposed limit on the north side of the fault then drops obliquely back to the lowest stratigraphic units exposed south of the fault. A completely arbitrary * figure of 100' is assigned here to represent the missing section.

SANGAMORE MEMBER

1987' - 2012'	Dolostone, dark grey, coarsely crystalline, some brecciated zones.
1986' - 1987'	Chert, black, dense, microcrystalline, some relics of grey dolomite remain. Laterally very continuous.
1978' - 1985'	Dolostone, medium dark grey color, numerous black chert nodules which occasionally exhibit white centers. Thinly bedded but in massive 1' to 6' units. Some of the chert nodules appear very lacy due to dolomite replacement as indicated by the rhombic holes in the weathered surface.
1924' - 1978'	Covered interval. A parallel section of slightly dubius correlation indicates the interval is composed of dolostones, mottled dark and medium grey, coarsely crystalline, some horizons with chert nodules up to 2" thick ranging in shape from nearly round to extremely elongate parallel to bedding. An oolitic horizon at 1958' in the substitute section.
1902' - 1924'	Dolostone, some brownish tan chert, partly massive, partly thin-bedded, mostly a dark grey color. (This unit marks the extreme upper limit of the coastal exposures at Balnakiel Bay.)
1892' - 1902'	Covered interval (beach).
1890' - 1892'	Limestone, dark grey color, cherty.
1887' - 1890'	Dolostone, reddish dark grey color, thinly bedded.
1879' - 1887'	Limestone, dark grey, cherty, humpy, algal mounds on upper surface.
1870' - 1879'	Dolostone, reddish dark grey, thinly bedded, distinct units 1/8" to 6" thick.

1838' - 1870'	Limestone, dolomitic, dark grey, highly cleaved, only slightly cherty, occasional hemispherical algal structures. The rocks occur as bedded units 4" to 1' thick, some of the units are mottled with dolomite. An horizon of intraclasts was sampled at 1863'.
1836' - 1838'	Limestone, finely crystalline, pronounced algal horizon.
1830' - 1836'	Limestone, dolomitic, dark grey, slightly cherty. Bedded units 4" to 1' thick, some showing slight mottling.
1826' - 1830'	Dolostone, mottled, and limestone, dark grey, in thinly interbedded units containing discreet beds of each lithology.
1823' - 1826'	Dolostone, dark grey, thinly bedded.
1810' - 1823'	Limestone, dark grey, partly dolomitic, cherty, highly cleaved.
1809½' - 1810'	Chert, black , capping the oolitic horizons below.
1787' - 1809½'	Limestone, dark grey, cherty in some units, partly dolomitic, finely crystalline, oolitic in upper 2'. Highly cleaved.
1782' - 1787'	Dolostone, thinly bedded, medium grey color, calcareous near top.
1774' - 1782'	Limestone, medium-dark grey, very humpy, nodular, irregular beds, highly cleaved, cherty in the basal 2'.
1767' - 1774'	Dolostone, buff-grey, in massive 8" to 14" beds, finely crystalline texture. Humpy ?algal structures on upper surface.
1766 - 1767'	Dolostone and limestone, interbedded in ½" beds, cherty, two sizes of ?worm trails on upper surface.
1762' - 1766'	Limestone, medium grey color, finely crystalline, cherty.

1760' - 1762'	Dolostone, medium grey, and limestone, dark grey, interbedded in ½" beds. Cherty. Very small worm trails (2-3 mm diam.) on upper surface.
1750' - 1760'	Limestone, medium grey color, cherty, finely crystalline, occurs in beds ranging from 3" to 1' thick.
1749' - 1750'	Dolostone and limestone in interbedded ½" bands. Cherty. ?Worm trails on upper bedding surface of this unit are 3-5 mm diam. and in an extremely meandering, self crossing pattern.
1739' - 1749'	Limestone, medium grey color, cherty, in beds 3" to 1' thick.
1736' - 1739'	Dolostone, light grey, thinly bedded, cherty near base.
1730' - 1736'	Limestone, medium grey color, occurring in beds 4" to 6" thick. Nodules of pink chert common. Some algal structures (humps) with a relief of more than 1'. Mottled with dolomite in upper 2'.
1729' - 1730'	Chert, dense, grey-buff color.
1728' - 1729'	Limestone, pinkish grey color, lower 6" dolomitic, upper 6" cherty.
1720' - 1728'	Limestone, grey-pink-buff color, numerous intraclasts in basal 1'. Above the intraclasts is a 1' algal horizon with heads composed of digitate structures. The algal horizon is overlain by a cherty cream colored unit which is in turn overlain by a grey limestone with pinkish chert nodules. Barbed-wire fence comes down to the small sea cliff at this point.
1709' - 1720'	Dolostone, light grey color, thin bedded, grouped in units 3" to 8" thick. Very flat and evenly bedded.
1708' - 1709'	Dolostone, calcareous, thinly bedded, some nodular cherts.
1704' - 1708'	Limestone, dolomitic, becomes more thinly bedded and more dolomitic toward top.

1702' - 1704'	Limestone, dolomitic, pinkish-grey, thinly bedded, "humpy" algal structures on upper surface.
1695' - 1702'	Dolostone, thinly bedded, grey, and vaguely mottled limey dolostone. Locally oolitic in the top 2'. Dark cherts show oolitic structure and weathering of dolomitized oolitic limestone portion of this unit produces a very remarkable sandy appearance on the outcrop surface.
1691' - 1695'	Dolostone, grey and buff units in an irregular nodular distribution with layers of finely crystalline, thinly bedded dolostone. Some chert near the top of this unit.
1684' - 1691'	Dolostone, thinly bedded, and mottled dolomitic limestone. Locally brecciated.
1666' - 1684'	Dolostone, thinly bedded, buff color, occurring in units 1" to 4" thick. Some pink colouration along certain bedding planes. Humpy algal structures near top. Some cherty layers.
1616' - 1666'	Missing section. At 1616' a fault is encountered in the coastal stratigraphic exposures and a thorough search has revealed no method by which a correlation can be made. It would not appear that this is a major structural complication and an arbitrary thickness of 50' is assigned to the displacement with the hope that this figure may roughly approximate the actual displacement. The strata above the fault are lithologically quite different than those below and it is therefore certain that section is omitted rather than repeated by the fault.
1611' - 1616'	Limestone, dark grey, dolomitic, strongly cleaved in some layers. Nodular cherts common near top.
1607' - 1611'	Limestone, medium dark grey, mottled with dolomite and chert, occurs in 1' to 2' beds. Fossiliferous near base.
1602' - 1607'	Dolostone, light grey, thinly bedded, mottled with limestone in the upper 6". Some pebbly cherts near base. Carbonates grouped in 2" to 6" beds.

1601' - 1602'	Limestone medium dark grey, some chert, blotchy dolomitization near top.
1599' - 1601'	Dolostone, light grey colored, thinly bedded.
1590' - 1599'	Limestone, dolomitic or dolostone, clacareous cherty, occurring in beds 1" - 3" thick. Medium grey colored, some current structures, cherts are nodular.
1588' - 1590'	Limestone, light grey color, thinly bedded, nodular cherts.
1586' - 1588'	Dolostone, light grey, thinly bedded, nodular cherts.
1580' - 1586'	Dolostone, dark grey, grading upward into limestone, light grey, fossiliferous.
1577' - 1580'	Dolostone, medium dark grey, thinly bedded, unit contains two brownish chert bands the thicker of which is 6" thick.
1575' - 1577'	Dolostone, medium grey, mottled.
1574' - 1575'	Chert, dark grey, very little dolomite remains as relic inclusions but much evidence for dolomitization post-dating the silicification.
1562' - 1574'	Dolostone, medium light grey color, lacy mottled appearance in dolostone, brownish chert nodules in the upper 2' which is more thinly bedded.
1560' - 1562'	Limestone, medium grey, finely crystalline.
1555' - 1560'	Dolostone, grey, medium crystalline grading upward to rather coarsely crystalline dolostone.
1547' - 1555'	Limestone, grey, ginely crystalline, partly dolomitized, considerable nodular chert throughout this unit, Gastropod collected at 1234'.
1529' - 1547'	Limestone, grey - pinkish, bedded in units 2' to 5' thick, chert bands spaced 1' to 5' apart, limestone finely crystalline.

1527' - 1529'	Dolostone, grey color, massive, some suggestion of the algal domes, thickens laterally, contains much chert.
1524' - 1527'	Limestone and dolostone in a patchy distribution but much larger than the mottled pattern of the dolomitic limestones.
1513' - 1524'	Limestone, pinkish-grey color, with nodular cherts in the lower 3'. Some of the cherts are very sharply divided from the host carbonate and are nearly perfect ovals, limestones are highly stylolitic, this unit thins and thickens laterally and a central 3' dolostone unit is locally present. The dolostone is medium grey, and coarsely crystalline.
1482' - 1513'	Dolostone, medium grey, coarsely crystalline, locally highly brecciated with the breccia fragments a finely crystalline, buff-colored dolomite. Jasperoid chert and "lacy chert". Thinly bedded and more siliceous in upper 4'.
1471' - 1482'	Dolostone, light grey, 8" to 2' beds, slightly banded and mottled near the top. Some brecciation, very coarsely crystalline.
1464' - 1471'	Dolostone, light grey, massive, highly brecciated, cavernous, some chert included in breccia fragments.
1461' - 1464'	Dolostone, light grey color, in massive beds 1' thick.
1459' - 1461'	Dolostone, light grey, thinly bedded, contains considerable nodular chert which is highly brecciated.
1455' - 1459'	Dolostone, medium light grey, massive, occurs in 8" beds.
1448' - 1455'	Chert, white and dark grey, dense, perhaps as much as 10 percent dolomite remaining. Although a strikingly different lithology, this unit does not make a conspicuous horizon because of its similarity in color to the units above and below.

1444' - 1448'	Dolostone, medium grey, thinly bedded, with considerable nodular chert, bedding is highly contorted, occasionally there is a "nodule" of coarsely crystalline dolomite with contortion of the adjacent beds.
1440' - 1444'	Dolostone, massive, slightly mottled, dark grey at base grading upward to a medium grey color. There is a cavernous horizon at the base of this unit. The vugs are often filled with coarsely crystalline dolomite.
1432' - 1440'	Dolostone, medium grey, massive, and light grey, thinly bedded, both coarsely crystalline. The two types are interbedded with a rather irregular distribution.
1425' - 1432'	Dolostone, buff colored, contains considerable nodular white chert.
1420' - 1425'	Dolostone, buff and grey mottled, thinly bedded toward the top of this unit.
1418' - 1420'	Dolostone, buff colored, coarsely crystalline granular texture.
1410' - 1418'	Dolostone, grey and pink, the pink color is largely due to fracture filling in breccias.
1405' - 1410'	Chert, pink, highly fractured, some relic inclusions of medium grey dolostone within the chert.

SANGAMORE MEMBER

SAILMHOR MEMBER 562' ✓

1396' - 1405'	Dolostone, pink and grey mottled, medium to coarsely crystalline.
1394' - 1396'	Dolostone, pink and grey mottled with the chert exhibiting a very lacy appearance due to the numerous rhombic holes presumably left by weathering of dolomite from the chert.

1381' - 1394'	Dolostone, pink and grey mottled.
1374' - 1381'	Dolostone, pink and grey mottled becoming increasingly cherty toward the top.
1369' - 1374'	Dolostone, medium to light grey color, finely crystalline, considerable pink chert in nodules and in 1" beds.
1362' - 1369'	Dolostone, dark grey color, mottled, cherty in the lower 8".
1360' - 1362'	Dolostone, light grey, thinly bedded, some pink dolomite, some pink nodular cherts.
1356' - 1360'	Dolostone, medium grey, mottled.
1354' - 1356'	Dolostone, dark grey, mottled locally cherty in 3" band at the base of the unit.
1352' - 1354'	Dolostone, finely crystalline, light grey, thinly bedded.
1333' - 1352'	Dolostone, coarsely crystalline, medium grey color.
1332' - 1333'	Dolostone, light grey, thinly bedded.
1328' - 1332'	Dolostone, medium grey, mottled.
1321' - 1328'	Dolostone, medium - light grey, thinly bedded, cherty in lower 1½'.
1303' - 1321'	Dolostone, medium grey, medium to coarsely crystalline, mottled.
1301' - 1303'	Dolostone, dark grey, thin bedded with a 4" band of chert at the base of the unit.
1297' - 1301'	Dolostone, medium grey, mottling - most pronounced at the base and becoming less conspicuous toward the top.
1295' - 1297'	Chert bed. Chert exhibits the lacy appearance caused by the differential weathering of the dolomite rhombohedra.
1292' - 1295'	Dolostone, medium grey color, mottled.
1291' - 1292'	Dolostone, dark grey, thinly bedded.

1284' - 1291'	Dolostone, medium grey, mottling most pronounced at the base and becoming less conspicuous toward the top.
1283' - 1284'	Dolostone, medium-dark grey, thinly bedded.
1274' - 1283'	Dolostone, medium grey color, coarsely crystalline, mottled.
1270' - 1274'	Dolostone, thinly bedded, medium crystalline texture, somewhat lighter grey than unit below, occurs in 4" beds.
1240' - 1270'	Dolostone, medium-dark grey, mottled coarsely crystalline, beds 3' to 6' thick, very little chert present.
1238' - 1240' *	Nodular chert bed in mottled dolostones. The chert nodules assume the mottled effect of the carbonates. * A questionable correlation across a major fault is made on the basis of this chert unit and although not entirely satisfactory, it is the best correlation available. No strata are repeated but some may be omitted.
1214' - 1238'	Dolostone, medium grey, extremely mottled, coarsely crystalline, occasional bands of nodular cherts are present.
1208' - 1214'	Dolostone, medium grey color, coarsely crystalline, numerous bands of tiny white nodular cherts.
1196' - 1208'	Dolostone, medium grey, coarsely crystalline, same as units above and below except for total absence of chert.
1190' - 1196'	Dolostone, medium grey, coarsely crystalline, several bands of tiny chert nodules.
1188' - 1190'	Dolostone, dark grey color, extremely mottled with a medium grey dolostone.
1185' - 1188'	Dolostone, dark grey color, unmottled, rather massive.
1175' - 1185'	Dolostone, coarsely crystalline, medium grey, conspicuously mottled.
1169' - 1175'	Dolostone, medium to light grey, massive capped by a 2" chert band.

1167½' - 1169'	Dolostone, medium grey color, thinly bedded.
1167' - 1167½'	Chert bed, dark color, dense.
1155' - 1167'	Dolostone, extremely mottled, occasionally there are bands of small, white, nodular cherts.
1150' - 1155'	Dolostone, thinly bedded, medium to light grey color, considerable chert in nodules.
1136' - 1150'	Dolostone, medium to dark grey color, coarsely crystalline, mottled, some pink dolomite.
1122' - 1136'	Dolostone, medium grey, coarsely crystalline, mottled, occasional chert nodules.
1117' - 1122'	Dolostone, thinly bedded, light grey color, finely crystalline, numerous bands of chert nodules 1" to 6" thick.
1100' - 1117'	Dolostone, medium grey, coarsely crystalline, mottled, occasional bands of tiny nodules of white cherts 1' to 3' apart.
1097' - 1100'	Dolostone, thinly bedded, medium grey, finely crystalline, on thin chert bed at base of unit.
1064' - 1097'	Dolostone, medium grey, coarsely crystalline, bands of chert nodules up to 3" in diameter at 1083' and 1085'. A 6" thinly bedded cherty dolomite occurs at 1074'.
1062' - 1064'	Dolostone, medium grey, thinly bedded, with nodular chert bands.
1057' - 1062'	Dolostone, thinly bedded, medium to light grey, interbedded with thin chert lenses ½" thick. This horizon is locally brecciated with coarsely crystalline calcite cement.
1050' - 1057'	Dolostone, medium grey, coarsely crystalline, occasional thin bands of tiny ½" chert nodules.
1045' - 1050'	Dolomite, thinly bedded, numerous chert bands (nodular) and chert beds. Some of the cherts are jasperoid.

1026' - 1045'	Dolostone, dark grey, mottled, coarsely crystalline, chert nodules at 1039' and at 1031'. Those at 1031' are unusual in their occurrence in that the discontinuous elongate nodules about 5" thick arch over the tops of ?algal dome structures. The dolostones are slightly mottled. At the cherty horizons, where the chert is absent, the dolostones are thinly bedded and arch over the domal structures.
1008' - 1026'	Dolostone, medium grey color, mottled. This unit contains numerous bands of nodular cherts ½" to 1" thick and spaced 4" to 3' apart. Between 1021' and 1022' there are three 4" chert beds that are laterally continuous.
1004' - 1008'	Dolostone, thinly bedded, charged with highly irregular shaped nodules of chert. Disarranged bedding structures may be detected within the chert nodules. The nodules here are suggestive of the "inverted tears" but are much less regularly shaped. Some exceed 1' in diameter.
989' - 1004'	Dolostone, medium grey, mottled, tiny nodular cherts (½" to 1" diameter) spaced 4" to 3' apart (horizontally) in rather continuous bands. There are occasionally larger nodules. An horizon of extremely mottled dolostone occurs at 1002' - 1003'
986' - 989'	Dolostone, thinly bedded, medium grey, numerous chert nodules, some are "inverted tear" shapes. No cavernous horizon occurs below this cherty layer.
967' - 986'	Dolostone, dark colored, mottled, coarsely crystalline.
959' - 967'	Dolostone, thinly bedded, dark colored, with numerous chert nodules as below. Some wave or current structures detectable in the dolomite and also preserved in the chert although perhaps somewhat distorted. The chert nodules are large but generally sub-circular.
949' - 959'	Dolostone, mottled dark grey and pinkish grey.

946' - 949'	Dolostone, medium grey, with large nodular cherts, some of which exhibit the inverted tear shape. Occasional vugs present in lower part.
943' - 946'	Dolostone, mottled dark grey and pinkish grey.
939' - 943'	Dolostone, thinly bedded, no mottling, many large chert nodules as below. Some of the "inverted tear" nodules are as much as 1½' in diameter. In this horizon some unsilicified undulations of the dolomitic bedding (?collapse structures or fold structures) may be quite clearly seen. There are some vugs partly or totally filled with pink calcite.
935' - 939'	Dolostone, dark grey color, mottled with medium grey dolostone.
932' - 935'	Dolostone, dark grey, extremely mottled.
925' - 932'	Dolostone, dark grey mottled with medium grey.
919' - 925'	Dolostone, dark grey colored, in thin highly undulatory beds charged with chert nodules. These nodules are generally sub-elliptical and some show only partial silicification of the nodular area with considerable relic dolomite contained within the nodule. Slightly cavernous at base of this unit.
910' - 919'	Dolostone, dark grey color, mottled, no chert present.
904' - 910'	Dolostone, dark colored, coarsely crystalline, mottled, occurs in massive 2' to 3' beds with small 1" nodular cherts in bands parallel to bedding and about 1' apart.
895' - 904'	Dolostone, dark grey, coarsely crystalline, mottled, occurs in two massive beds, the lower 6' thick and the upper 2' thick.

891' - 895'	Dolostone, dark grey colored, thinly bedded and with undulations in the bedding planes common. Numerous chert nodules some sub-elliptical and many "inverted tear" shapes. The silicification appears to either control or be controlled by the undulatory structures in the carbonates. Some unsilicified structures may be present. These large chert nodules have affected to a limited extent the compaction of the overlying strata and therefore must have antedated total compaction.
889' - 891'	Dolostone, extremely cavernous or vuggy, dark grey color except for the cavities which are filled with very coarsely crystalline pink calcite.
866' - 889'	Dolostone, dark grey, mottled, coarsely crystalline, some signs of recrystallization, massive units 4' to 6' thick.
854' - 866'	Dolostone, coarsely crystalline, dark grey colored, considerable chert both in beds and in nodules. Some nodules are up to 6" thick.
852' - 854'	Dolostone, dark grey color, very frequent nodular cherts.
843' - 852'	Dolostone, dark lead color, occurs in massive 1' to 4' beds separated by thin 1" to 4" nodular or bedded chert horizons.

SAILMHOR MEMBER 562'

EILEAN DUBH MEMBER 547'

828' - 843' *	Dolostone, light to medium grey colored, finely crystalline, flat bedded in units 1" to 2' thick. * 843' would be this author's guess for the top of Peach and Horne's Eilean Dubh (Group II) dolomites.
823' - 828'	Dolostone, medium grey, slightly domed, a 1" chert band at base and considerable chert throughout the dolostone. A 4" to 6" chert bed marks the top of this unit. Some of the chert is slightly jasperoid.

818' - 823'	Dolostone, buff and grey color, thinly bedded, brecciated horizon near base.
816' - 818'	Dolostone, medium grey, brecciated, mottled, possible worm borings, cherty, the top of this unit is marked by a 3" cherty band.
810' - 816'	Dolostone, light grey color, massive in the lower 3½' grading upward into thinly bedded dolostones toward top.
802' - 810'	Dolostone, coarsely crystalline, in an unusual globular mixture with a more finely crystalline dolostone. Coarser portions are dark grey in color while the finer portions are pink to buff colored. Some drusy chert present.
800' - 802'	Dolostone, light grey, massive, slightly undulatory or domal on upper surface - ?algal.
795' - 800'	Dolostone, thinly bedded, buff-grey color, finely crystalline.
794+' - 795'	Dolostone, finely crystalline, thinly bedded, weathers to a distinctive orange brown hue.
790' - 794'	Dolostone, grey to buff colored, finely crystalline, contains some sandy lenses up to 4" thick and 3' in length.
769' - 790'	Dolostone, grey, finely crystalline, algal dome structures on upper surface.
765' - 769'	Dolostone, pink, grey and buff colored, very thinly bedded and very finely crystalline.
745' - 765'	Dolostone, finely crystalline, buff to grey to sometimes pinkish hues, occurs in somewhat massive beds up to 1' thick but becoming more thinly bedded toward the top.
740' - 745'	Dolostone, medium grey color, finely crystalline, strongly developed hemispherical algal structures with approximately 2' of vertical relief on the upper surface. Upper-most layers of the algal domes exhibit the cylindrical digitate structures close-packed in a slightly radial pattern. The digitate cylinders measure 1 to 1.5 mm in diameter and exhibit an internal structure of arched microlaminations (convex upward).

729' - 740'	Dolostone, thinly bedded, interbedded grey and pinkish buff units.
722' - 729'	Dolostone, thinly bedded, in units 1" to 3" thick.
712' - 722'	Dolostone, thinly bedded, interbedded pinkish and grey colored units. A 4" brecciated zone occurs at 720'.
707' - 712'	Dolostone, grey to buff colored, massive, finely crystalline.
694' - 707'	Dolostone, interbedded thin bedded units and algal hemispherical units. At 704' is an individual dome with over 4' of relief. The algal horizons are slightly siliceous especially on the upper surfaces. These domal structures do not appear to contain any internal structures (not even bedding). The thinly bedded units associated with the algal heads show considerable brecciation locally.
691' - 694'	Dolostone, medium grey color, conspicuous hemispherical structures with concentric layers of digitate tube-like structures as above and below capping the hemispheres.
688' - 691'	Dolostone, finely crystalline, grey color, occurs in massive 1½' beds.
671' - 688'	Dolostone, buff and pinkish colored, thinly bedded in units which differ slightly in color and range in thickness from a few millimeters up to 6" thick. Occasional sandy horizons up to 3" in thickness are present. The sand seldom appears to constitute more than 50 percent of the rock. Some of the sandy layers are very persistent laterally.
669' - 671'	Dolostone, medium grey color, in one massive 1'+ bed.
667½' - 669'	Dolostone, current bedded, capped by a 4" brecciated horizon.
667' - 667½'	Dolostone, grey, cherty, slightly "humpy", capped by a 1" sandy dolostone layer.

665' - 667'	Dolostone, thinly bedded, grey color, some brecciation locally.
662' - 665'	Dolostone, grey, containing chert nodules over which large algal heads or mounds are built. The "heads" are capped by concentric layers of close packed tube-like structures arranged in a slightly radial fashion. The tubes have arched laminations within each individual structure. There is a rather continuous brecciated layer 6" thick at the base of this unit.
653' - 662'	Dolostone in a nodular, "humpy" ?algal unit with some chert around the peripheries of the nodules. Beds both above and below the nodules appear to be arched around the nodular cores.
647' - 653'	Dolostone, thinly bedded, buff to grey color, finely crystalline.
643' - 647'	Dolostone, finely crystalline, massive, medium grey, hemispherical algal structures but with no internal structures discernable.
616' - 643' *Load structures	Dolostone, buff to grey, thinly bedded, finely crystalline. At 639' there is an horizon which has some very unusual elongate and crudely parallel load structures. These structures are developed only along a very definite single horizon that can be traced more than 100' laterally.
578' - 616'	Dolostone, thinly bedded, color varies from buff to light grey to pinkish grey. Thin hematitic layers on the surface mark some bedding surfaces. At 586' is a buff-colored dolostone unit in which the bedding units intermittently appear to arch up over some structure. The exposures are too poor to study the details of the "humps" but they might be algal "head" structures as above or they may simply represent nodules of early lithification that might have resisted compaction more than the surrounding beds. The underlying strata show no compression by the "nodules". The bedding units all appear to be continuous over the "humps" with some thinning over the humps. Similar structures are also present at 591', 606' and 608'.

576' - 578'	Strongly brecciated zone. May be an erosion surface or (more likely) a plane of thrusting.
541' - 576'	Dolostone, thinly bedded, buff to flesh colored with some grey units, numerous hematitic bands marking bedding surfaces are probably the result of osidation of original iron rich layers in the rock. Finely crystalline. In beds ½" to 2" thick.
521' - 541'	Covered interval (beach). Probably very similar to the thinly bedded dolostones above and below.
482' - 521'	Dolostone, light grey color, thinly bedded, often with surficial hematitic staining along bedding planes as above. Very finely crystalline.
481' - 482'	Dolostone, light grey color, finely crystalline, with very pronounced algal head structures (the lowest algal horizon). These algal "heads' exhibit the typical <u>Collenia</u> -type structure. The heads range in size from 6" to 1' in diameter, and as much as a foot high. No structure other than the concentric layering can be seen.
430' - 481'	Dolostone, light grey, in beds ranging in thickness from 1' to 6'. Generally medium crystalline but with some patches of coarsely crystalline dolostone.
418' - 430'	Dolostones, flesh to lilac colored, thinly bedded, fine to medium crystalline, occurring in beds 1" to 6" thick.
366' - 418'	Dolostone, light tan to pinkish grey, very difficult to study because of covering of seaweed and barnacles. This unit is exposed on the very outer island of Eilean Dubh. (rather between the two outer islands)
336' - 366'	Dolostone, buff to tan color, some pinkish grey units, thinly bedded, finely crystalline. Much of the exposure is obscured by sea weed.
326' - 336'	Dolostone, flesh to nearly white colored, in massive 3' to 5' beds, very finely crystalline.

296' - 326'	Dolostone, light buff to tan color, finely crystalline, thinly bedded. This unit is the very lowest exposed portion of the section at the Balnakiel section. These lowest few units are accessible only at the lowest tides.

EILEAN DUBH MEMBER 547'

196' - 296'	Missing section. The arbitrary thickness of 100' is assigned here to represent an unknown thickness of missing section. The units below this missing section are measured along the east side of Loch Eribol and although the Eilean Dubh dolomites are encountered there, there is no criterion by which the two sections can be correlated. The lowest recognizable algal horizon in the section above (at 481') is absent in the Eribol exposures. It is not certain that any section is missing here and it is possible that far in excess of the allowed 100' may be absent. There is no known location where this problem can be resolved.

GRUDAIDH MEMBER 196'

187' - 196'	Dolostone, dark grey, rather coarsely crystalline, extremely mottled.
179' - 187'	Dolostone, light grey, thinly bedded, with occasional thin hematitic bands separating the beds. Intraformational pebble conglomerate at 180'.
129' - 179'	Dolostone, medium grey, mottled with a slightly lighter grey dolomite. There are some cherty zones where the cherts appear to be oolitic. Some of the darker dolostones have fine fractures filled with white dolomite.
117' - 129'	Dolostone, light grey, finely crystalline, thinly bedded.

110' - 117'	Dolostone, nearly black, rather coarsely crystalline. At the base of this unit is a 2" brecciated zone.
108' - 110'	Dolostone, light grey and medium grey mottled, medium crystalline.
12' - 108'	Dolostone, dark grey, mottled with medium grey dolostone to give a "lumpy" appearance. Bedded in units ranging from 1" to 4" thick. In the lower part of this unit there are occasional vugs that are filled with coarsely crystalline calcite. The unit tends to a lighter color grey toward the top. Within the lower part of this unit (25' - 30') there are some minor structural complications but the lithology remains constant and probably the accuracy of the section is lightly if at all impaired.
10' - 12'	Dolostone, sandy, medium grey, medium crystalline.
4' - 10'	Dolostone, dark grey, mottled with a somewhat lighter colored dolostone to give a "lumpy" mottled appearance. Beds 1" to 4" thick.
0' - 4'	Dolostone, sandy, dark grey, in beds 2" to 4' thick, medium crystalline, slightly mottled, strongly stylolitic.

Below this is a black slaty dolomitic, pyritic shale that is included in the desriptions of the intermediate series.

AN T-SRON FORMATION

79' - 86'	Shale, dolomitic, slaty, pyritic, very black color.
77' - 79'	Sandstone, dolomitic, or dolostone, sandy?. Buff-grey color. Sand fraction is fine-grained.

76' - 77'	Sandstone, dolomitic, buff or tan colored, medium to fine grained, very carious. Possible boring trails weathered out as small tubes in the sandstone.
58' - 76'	Sandstone, quartzitic, medium grained, flesh colored, highly cross-bedded. Bedded in units ranging in thickness from 2' to 5'. This is the "Serpulite or Salterella Grit" of Peach et. al. This sandstone is considerably more resistant to weathering and erosion than the strata above and below and thus often forms a slight escarpment.

"SERPULITE GRIT" MEMBER. 28'

"FUCOID BEDS" MEMBER. 58'

57' - 58'	Shale, very dark grey-brown, strongly cleaved.
0' - 57'	Siltstone, mudstones, fine sandstones, nearly all dolomitic, bedded in units ranging in thickness from ½" to 1½'. Very small current structures are sometimes visible in the siltstones and fine sandstones. Bedding planes sometimes are traversed by numerous trails of some organism. There may also be resting trails of trilobites. These units are the "Fucoid Beds" of Peach et. al.

Below these siltstones are the rather pure quartzitic sandstones of the "pipe rock" series. These are described and numbered separately from the intermediate series.

ERIBOL SANDSTONES

PIPE ROCK MEMBER 319'

312' - 319'	Sandstone, quartzitic, coarse grained, poorly cemented in part, carious, buff-colored, weathers darker than fresh surface. Contains some very large pipes (up to 2" diameter). There is a 3' dolomitic zone in the center of this unit

	in the exposures at An - t-Sron (east side of Loch Eribol)
206' - 312'	Sandstone, quartizitic, fine to medium grained, flesh colored to grey with some reddish streaks. The pipes in this unit are not as conspicuous as in the unit below because they are perhaps not quite as numerous and also because many of the pipes are slightly smaller.
205½' - 206	Sandstone, medium grained, 8" horizon of extremely large diameter pipes (up to 4").
180' - 205½'	Sandstone, quartzitic, fine to medium grained, flesh colored to grey with bands of red hematite staining that often produce a rather striking contrast between the "pipes" and the host sandstones. The pipes often are very white relative to these red rocks. Zones of the "exceedingly numerous pipes" are the most distinctive feature of this unit but many of the units within this group do not contain the "exceedingly numerous pipes". The pipes often touch each other where they are closely packed.
165' - 180'	Sandstone, quartzitic, grey colored, thinly bedded, medium grained, no apparent pipes.
164' - 165'	Thin horizon of the "exceedingly numerous" small pipe structures. Same lithology as above.
155' - 164'	Sandstone, quartzitic, very white colored, The larger trumpet pipes predominate but there are also many of the "ordinary" pipes. The sandstones are bedded in units about 8" thick.
153' - 155'	Sandstone, quartizitic, white, thinly bedded, some large "pipes".
138' - 153'	Sandstone, quartzitic, somewhat whiter than the units below. The large pipes predominate in sandstones bedded in 8" units as above. Some of the smaller pipes in this unit show very barely discernible concentric structures around them and may in fact be the central "cores" of trumpet pipes.

127' - 138'	Sandstone, quartzitic, medium grained. In this horizon the trumpet pipes appear, sometimes by themselves and sometimes with the more "ordinary" pipes. There is very slight cross-bedding in this unit.
99' - 127'	Sandstone, medium grained, flesh colored. A slight amount of argillaceous material in the interpipe areas renders the pipes of the "ordinary" type much more visible. The pipes of this unit often weather out in positive relief relative to the interpipe areas presumeably due to a higher quartz content. The easy visibility and sharp separation may account in part for the somewhat larger appearance of these pipes.
96' - 99'	Sandstone, quartzitic, very coarse grained, dark maroon color. No pipes obvious.
0' - 96'	Sandstone, quartzitic, fine to medium grained, Flesh colored units interbedded with darker maroon colored units with all variations of colors between the two extremes. The reddish maroon color appears to be due to hematite staining and the colors change considerably in both the vertical and horizontal directions. These sandstones are relatively flat bedded except for a very few isolated horizons. The bedding units range in thickness from 6" up to 2' with 1' units being very common. The pipes of this lower unit are very difficult to recognize and, in fact, appear to be absent from some of the units. In the hematite-stained units the pipes are somewhat easier to distinguish. Occasionally a "pipe" will be stained dark, perhaps due to the animal dying with the tube?? One of these stained pipes could be continuously traced down a joint surface for 18". The weathered bedding surfaces offer perhaps the best means for determining the presence of pipes because the surface assumes a strange nubbly effect only when the pipes are present. The pipes of this unit are not at all visible on freshly broken surfaces. Because of the faintness of the pipe structures it is impossible to determine the diameters with any degree of accuracy but 0.2" would be a close approximation.

Below these sandstones are the highly cross-bedded sandstones of the Lower Series. The cross-bedded units are described and numbered separately.

LOWER MEMBER X_q: 255'

175' - 255'	Sandstones, quartzitic, slightly feldspathic, flesh colored, occurring in cross-bedded units 8" to 3' in thickness. Occasional pebbly bands occur at the base of cross-bedded units. These pebbly horizons are always found on the erosion or truncation surface and are never involved in the cross-bedding. Some horizons exhibit a maroon color banding which is almost certainly secondary since the bands transgress units.
118' - 175'	Sandstone, quartzitic, feldspathic, maroon and flesh colored. Occurs in cross-bedded units 6" to 3' thick. The lithology and sedimentary features of this unit remain remarkably constant throughout the unit with the exception of the secondary dark staining which is less conspicuous in the higher horizons. The sands are medium grained and very tightly compacted and hard rocks. No porosity is visible in the hand samples.
102' - 118'	Sandstones, quartzitic, orange-fleshy color, occuring in 1' - 2' units often separated by thin 1" to ½" green chloritic bands. These sandstones are rather coarse grained and very conspicuously cross-bedded.
86' - 102'	Sandstone, quartzitic, feldspathic, maroon colored, conspicuously cross-bedded, banded with fleshy and milky appearing sandstones. These rocks are very hard nearly meta-quartzites.
0' - 86'	Sandstones, quartzitic, feldspathic, and pebbly feldspathic conglomerates. The quartzitic sandstones are mostly coarse grained but some horizons are medium grained or even rarely fine-grained. The thickness of the cross-bedded units varies from 6" to 2', with 8" to 12" units being the most common. The basal few feet of this unit are distinctly

more conglomeratic than the higher horizons although in many locations the sand-sized material rests directly on the unconformity. The conglomeratic material is seldom very coarse and pebbles exceeding ½" diameter are rare. There is no distinct zone that can be designated as a basal conglomerate for the pebbles gradually decrease in frequency and some pebbles are found as much as 200' above the nonconformity.

These sandstone and pebbly conglomerates rest nonconformably on Lewisian Gneiss at this location. The gneiss is a coarse migmatitic banded rock with bands of micaceous gneiss separating the more feldspathic and quartzitic layers.