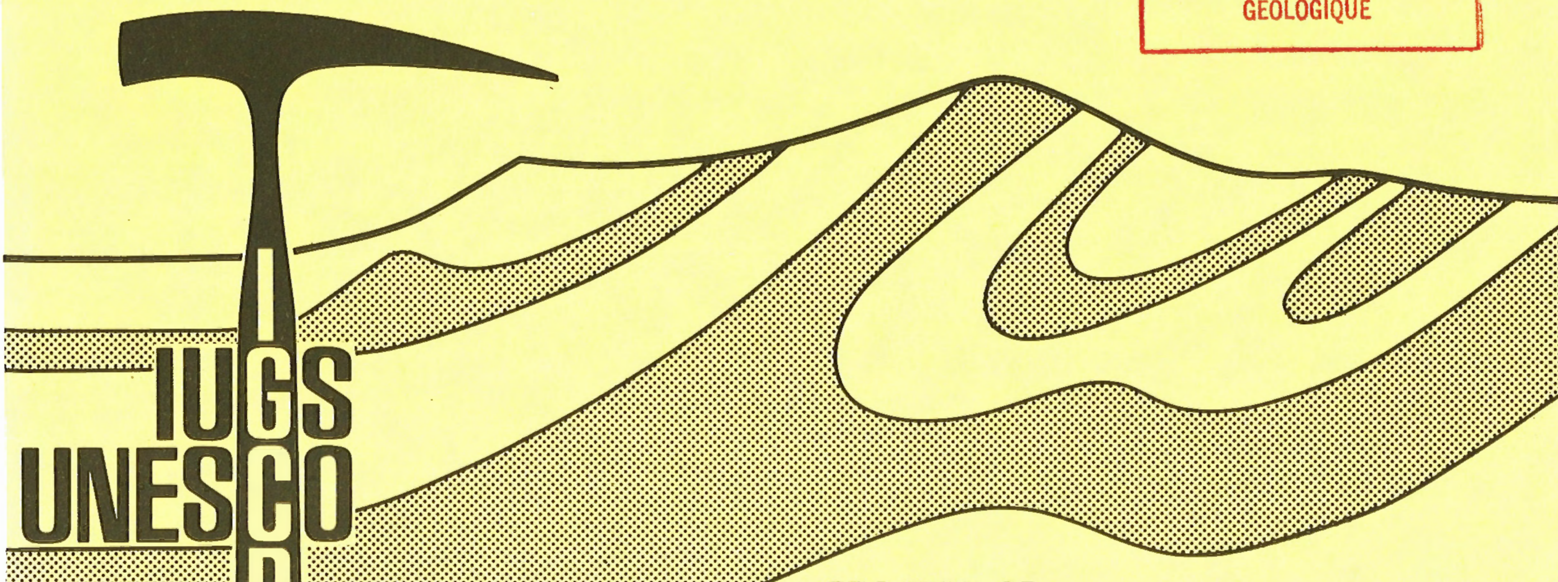
GEOLOGICAL SURVEY OF CANADA
PAPER 78-13

CALEDONIAN - APPALACHIAN OROGEN OF THE NORTH ATLANTIC REGION



PROJECT 27

CALEDONIDE OROGEN 1978

GEOLOGICAL SURVEY OF CANADA
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CALEDONIAN-APPALACHIAN OROGEN OF THE NORTH ATLANTIC REGION

CONTRIBUTIONS TO
INTERNATIONAL GEOLOGICAL CORRELATION PROGRAMME
PROJECT 27 - CALEDONIDE OROGEN

Energy, Mines and Resources Canada
Énergie, Mines et Ressources Canada

OTTAWA 1978

available by mail from

Printing and Publishing
Supply and Services Canada,
Hull, Québec, Canada K1A 0S9,

and

The Geological Survey of Canada
601 Booth St., Ottawa, K1A 0E8

or

Through your bookseller.

Catalogue No. M44-78/13	Price: Canada: $5.00
ISBN - 0-660-01534-X	Other Countries: $6.00

Price subject to change without notice

Preface

The publication of this volume is a Canadian contribution to the International Geological Correlation Programme (IGCP) which is a joint undertaking of UNESCO and IUGS established for the promotion of geological research projects of international scope. At present some 62 projects are being undertaken involving about 100 countries with participation of scientists from government organizations, universities and industry. Since 1971 the government of Canada has endorsed and supported IGCP objectives through a National Committee with members drawn from Canadian universities, industries and government.

The 25 papers published in this volume were prepared as part of one of the 62 projects of the programme, Project 27-The Caledonian Orogen. This project is concerned with the great Paleozoic orogenic belt exposed in the lands bordering both sides of the Atlantic Ocean and which forms such a significant feature of the geology of the Appalachian region of Eastern Canada and the United States. The papers cover a wide variety of related problems including geological comparisons between Newfoundland and Ireland, and between Nova Scotia and Morocco. Such comparisons are unquestionably relevant to the elucidation of the significance of the Caledonian orogeny which cannot be attempted from studies within any one country.

The publication of this material by the Geological Survey of Canada provides additional support to the International Geological Correlation Programme and also ensures that the material will be widely distributed and readily available. Part of the cost of preparation was borne by National Committee funds, the Survey assumed responsibility for printing and distribution. Dr. E.T. Tozer of the Geological Survey who is also Secretary of the IGCP National Committee undertook the not inconsiderable task of preparing the papers for publication. With contributions from many nations, in a variety of styles it was inevitable that there must be some modification of the normal format and editorial standards of the Geological Survey Paper Series, nevertheless by publishing this volume the Survey is making a significant contribution to the progress of the IGCP and such progress is of benefit to Canada and to all the other countries participating in the Programme.

D.J. McLaren
Director General
Geological Survey of Canada

Ottawa
March 31, 1978

Scientific editors

E.T. Tozer
Paul E. Schenk

Technical editing

E.T. Tozer
Lorna A. Firth
Leona R. Mahoney

Text typed by

Angelica F.V. Koops
Suzanne M. Lalonde

Illustrations

Executed or directed by authors

Cover and Figure 19.11

Kitty Vincent

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References to this Volume

It is recommended that reference to contributions in this volume be made in the following form:

Harland, W.B.
1978: The Caledonides of Svalbard; *in* IGCP Project 27, Caledonian-Appalachian Orogen of the North Atlantic Region; Geol. Surv. Canada, Paper 78-13, p. 3-11.

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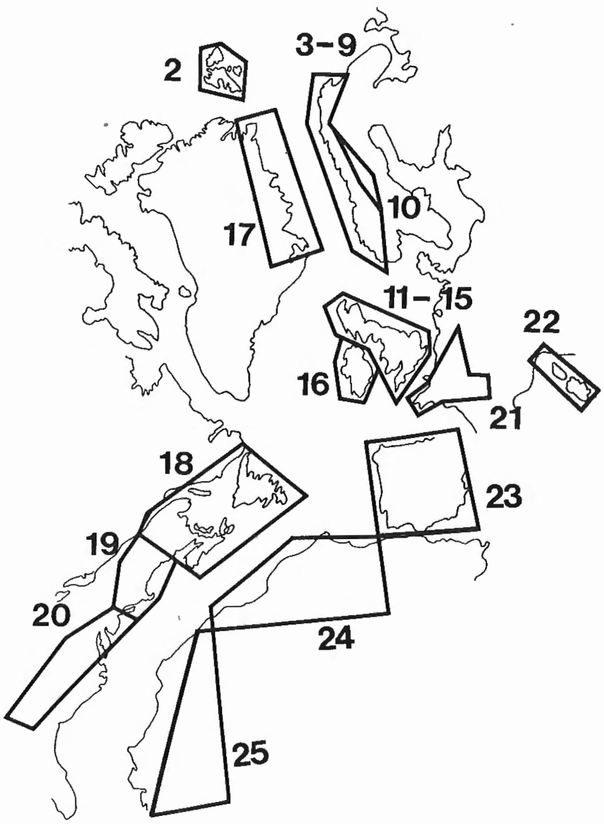
CALEDONIAN-APPALACHIAN OROGEN OF THE NORTH ATLANTIC REGION - FOREWORD

Brian A. Sturt

Geologisk Institutt
Avd. A.
Universitetet i Bergen
5014 Bergen, Norway

This volume in an outcome of Project 27 - the Caledonide Orogen - of the International Geological Correlation Programme (IGCP). This project was proposed by the Norwegian IGCP National Committee in 1973. It was accepted by the International IGCP Board in April 1974 under Division II of the Programme ("Major geological events in time and space their implications in environmental processes"). In November 1974 an organization meeting was held in Oslo under the writer's chairmanship. This led to the organization of the project with the writer as Project Leader and David L. Bruton (Paleontologisk Museum, Sarsgt. 1, Oslo N5) as Project Secretary. The First International Working Group meeting took place at Bergen, in November 1975. At this meeting it was decided to ask participating countries to provide short national syntheses of the Caledonide Orogen. These were presented in preliminary form at the Second International Working Group meeting, Oak Island, Nova Scotia, September - October 1976, organized by Paul Schenk, the Canadian Project Leader. At this meeting it was decided that the syntheses of the Atlantic Caledonides should be published. Svalbard was not dealt with at the Oak Island meeting but in order to complete the coverage of the North Atlantic region W.B. Harland kindly contributed an article on that area. Through the kind offices of the Geological Survey of Canada it has been possible to produce these contributions in their present form. It is hoped, in spite of the varied styles of presentation, that these syntheses will give a quick and useful overview of the present status of knowledge concerning the Caledonian geology in the Atlantic region.

Figure 1.1, prepared by Paul Schenk, indicates the scope of the contributions.



2 - Svalbard
3-9 - Norway
10 - Sweden
11-15 - British Isles
16 - Ireland
17 - East Greenland
18 - Canadian Appalachians
19 - Northern U.S.A. Appalachians
20 - Southern and Central U.S.A. Appalachians
21 - France
22 - Southern France, Corsica and Sardinia
23 - Iberian Peninsula
24 - Morocco
25 - West Africa

Continental arrangement from Figure 9 in: Le Pichon, X., Sibuet, J.-C., and Francheteau, J.; "The Fit of the Continents around the North Atlantic Ocean"; Tectonophysics, v. 38, p. 169-209. (1977). (Paul E. Schenk).

Figure 1.1. Index map, Caledonian-Appalachian Orogen of the North Atlantic Region. Numbers are those of articles.

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Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

THE CALEDONIDES OF SVALBARD

PROJECT 27 CALEDONIDE OROGEN

W.B. Harland

Department of Geology
University of Cambridge
Sedgwick Museum, Downing Street
Cambridge CB2 3EQ, England

INTRODUCTION

Spitsbergen (the principal island), Nordaustlandet, Prins Karls Forland, Kvitøya and Bjørnøya are the larger islands within the Svalbard Archipelago that expose pre-Carboniferous rocks (Fig. 2.1). These older rocks are generally deformed and often metamorphosed and were first described as the Hecla Hoek Formation by Nordenskiöld (1863) who included the Old Red Sandstone in that division. Later usage came to define Hecla Hoek rocks as of pre-Devonian age and the recognition of Spitsbergen as belonging to the Caledonides led to the following simple first approximation of the sequence:

Carboniferous and later strata

Unconformity, representing the SVALBARDIAN FOLDING

Old Red Sandstone (earliest Devonian to latest Middle Devonian)

Unconformity, representing the main CALEDONIAN OROGENY, with tectogenesis, metamorphism and plutonism

Hecla Hoek rocks (Precambrian to Early Palaeozoic).

A close analogy with Scotland was pointed out by Holtedahl (1925). The late Devonian, early Carboniferous peneplain clearly divided older rocks with geosynclinal marine and continental sequences from contrasting Carboniferous through Palaeogene platform facies. The relationship between the highly metamorphosed and the relatively unaltered rocks within the Hecla Hoek complexes long troubled geologists. The schists and gneisses could be: (a) Archaean strata on which rested the relatively unaltered rocks (e.g. Tyrrell, 1922; Odell, 1927; Sandford, 1956); (b) they could all be of similar Palaeozoic age but of differing degrees of metamorphism (e.g. Blomstrand, 1864; Holtedahl, 1929); or (c) they could be a single geosynclinal sequence in which the older strata have suffered the deepest alteration (e.g. Major and Winsnes 1955; Harland and Wilson, 1956). While the last view had prevailed for most areas, the other possibilities have not been eliminated everywhere. A further complication is that the west coast of Spitsbergen and Prins Karls Forland underwent tectogenesis during the mid-Tertiary West Spitsbergen Orogeny, so that the older rocks have been deformed at least twice, and in opposite senses, thrusting to the west prevailing in Palaeozoic time, to the east in Tertiary time.

The resulting complexity delayed elucidation of the stratigraphic sequence in the west. The easternmost sequence is still relatively unknown because of its difficulty of access. Nevertheless, a simple scheme (e.g. with Lower, Middle and Upper Hecla Hoek of Harland and Wilson, 1956) served well while the details of the succession were worked out in different areas.

However, when looked at in detail, there are marked differences even between neighbouring areas though their full significance is debatable. In an attempt to separate fact from speculation, the main body of this paper will summarize the stratigraphic and tectonic sequence in the areas where these are best known; some interpretation will follow. It is a matter of judgement as to how many areas to distinguish or unify. In the past the tendency has been to unify stratigraphic schemes with a great economy in nomenclature but this may well have obscured significant contrasts. However, in this short paper there is no space to go to the other extreme so a compromise is effected.

LITHO-TECTONIC SEQUENCES

To save space the litho-tectonic sequence is summarized in semi-tabular form (youngest events at the top) in ten selected areas shown by number on Figure 2.1.

Abbreviations used are as follows:

Unit names:

Fm - Formation
Gp - Group
SG - Supergroup

For lithology:

am - amphibolite	ps - psammite
cgl - conglomerate	qi - quartzite
dst - dolostone	sh - shale
fe - feldspathite	slt - siltstone
gw - greywacke	spe - semipelite
lst - limestone	sst - sandstone
ma - marble	till - tilloid or tillite
pe - pelite	volc - volcanics

For age:

E - Early, M - Middle, L - Late, with commonly understood period abbreviations.

All radiometric age determinations (in Ma without qualification) can be found in Gayer *et al.*, 1966. Only a few references have been selected for each area.

(1) EASTERN NORDAUSTLANDET AND KVITØYA

See Sandford, 1926, 1954; Flood *et al.*, 1969; Krasil'shchikov, 1973.

No stratal sequence has been elucidated in a complex of high-grade metamorphic rocks with feldspathite and amphibolite. This was originally regarded as Archaean basement or alternatively as equivalent to

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Lower Hecla Hoek rocks in Ny Friesland. Radiometric ages generally show Caledonian recrystallization (393-430 Ma) but some are enigmatic (581-618 Ma). The area is generally icebound.

(2) WESTERN NORDAUSTLANDET

See de Geer, 1923; Kulling, 1934; Sandford, 1950, 1956; Flood *et al.*, 1969.

Carboniferous and Permian strata rest unconformably on the

Murchisonfjorden SG

Kapp Sparre Fm 1 km, lst, dst with ?E Camb. brachiopods, possibly also Ordovician.

Gotia Gp 0.6 km, sh and till (Sveanor Fm) of Vendian age.

Raoldtoppen Gp 1.3 km, dst, lst, chert and sh with stromatolites and pisolites of L Riphean (Karatavian) aspect.

Celsiusbreen Gp 2.5 km, dst, sst, sh.

Franklinsundet Gp 1.8 km, sst, slt and sh.

The above appear in normal sequence and overlie unconformably

Botniahalvøya Gp 6 km, qi, sh and volc, cgl (Kap Hansteen Fm).

The sequence might simply continue downwards towards the east (*see* area (1) above). The Murchisonfjorden Supergroup is folded but not metamorphosed. The older rocks give Caledonian ages (380-420 Ma).

(3) NY FRIESLAND

See Blomstrand, 1864; Tyrrell, 1922; Fairbairn, 1933; Fleming and Edmonds, 1941; Harland and Wilson, 1956; Bayly, 1957; Harland, 1959; Harland *et al.*, 1966; Gayer, 1969; Wallis, 1969; Fortey and Bruton, 1973.

This is the type area of the Hecla Hoek sequence from the eponymous mountain (now Heclahuken). It is the largest area of continuous exposure of these older rocks which were peneplained and overlain by Early Carboniferous (Tournaisian).

The tectogenesis in this area has been named the NY FRIESLAND OROGENY. The geosyncline of 18 km or more thickness has been intensely deformed and metamorphosed in the lower part (365-434 Ma); the upper rocks are folded but little altered. The whole structure shows an approximately NS strike and in the west shows considerable NS elongation and thinning of strata. It is cut by two late tectonic granite batholiths (385-406 Ma).

Hinlopenstretet SG (Upper Hecla Hoek)

Oslobreen Gp 1-2 km, lst and dst with E Camb. and rich E to M Ord. faunas.

Polarisbreen Gp 0.8 km, sh and till. Vendian microfossils.

Lomfjorden SG (Middle Hecla Hoek)

Akadamikerbreen Gp 2 km, dst, lst with stromatolites and pisolites at top. Late Riphean (Karatavian) and possibly partly Vendian.

Veteranen Gp 3.8 km, qi, gw, sh.

Stubendorffbreen SG (Lower Hecla Hoek)

Planetfjella Gp 4.8 km, qi, ps, spe and acid volc seds.

Harkerbreen Gp 4.1 km, qi, ps, spe, volc; fe and am.

Finnlandveggen Gp 2.7 km, spe, ma, volc; fe and am.

The Stubendorffbreen Supergroup is distinguished by a high proportion of volcanics. Those of the Planetfjella Group comprise only acid pyroclastics. The two lower groups include acid and basic pyroclastics and lavas; they were united by Krasil'shchikov (1973) as the Atomfjella phase. No significant breaks have been demonstrated in the sequence.

(4) NORTH CENTRAL SPITSBERGEN

See Holtedahl, 1914 and 1926; Friend, 1961; Gee and Moody-Stuart, 1966; Friend, 1968; Friend and Moody-Stuart, 1972; Harland *et al.*, 1974.

Underlying Early Carboniferous (Tournaisian) strata is the classic Old Red Sandstone "graben" which appears not to have been initially formed as a graben (Harland, 1969). The Old Red Sandstone has been folded, thrust and locally cleaved especially along the sinistral strike-slip Billefjorden Fault Zone. These structures have been referred to as SVALBARDIAN (Vogt, 1938).

Andree Land Gp 5 km, red sst and slt, mainly continental facies ages ranging from E Dev (Gedinnian) to latest M Dev.

Red Bay Gp 2 km, red sst and cgl Earliest Devonian.

The thick Red Bay conglomerate rests on and truncates both the deformed Siktefjellet Group strata and the older metamorphic rocks. This tectonic phase has been referred to as HAAKONIAN (Gee, 1972).

Siktefjellet Gp 1.5 km, grey sst and cgl probably latest Sil.

The above Old Red Sandstone (Red Bay and Siktefjellet) conglomerates rest on tectonised high grade metamorphic rocks.

"Heclahook" (Holtedahl, 1914) pe, ps, eclogites, agmatites and migmatites whose age has only tentatively been suggested as "Lower Hecla Hoek".

Age determinations give a few enigmatic dates (1430-1541 Ma) and more CALEDONIAN *s.s.* figures (365-431 Ma).

(5) NORTHWESTERN SPITSBERGEN

See Holtedahl, 1914 and 1926; Gee and Hjelle, 1966; Hjelle and Ohta, 1974.

In this area, bounded by Raudfjorden and Kongsfjorden, the solid rocks exposed are all metamorphic rocks (375-450 Ma), except for the large Horneman granite batholith (318-340 Ma).

Generalfjella Fm 2 km, ma, pe and qi.

Signehamna Fm 2 - 2.5 km, pe and ps.

Nissenfjella Fm 3 km, pe and am.

The lowest formation passes downwards and eastwards into widespread migmatites. These may be continuous with the migmatites of area (4).

(6) CENTRAL WESTERN SPITSBERGEN

See Holtedahl, 1913; Orvin, 1934; Harland, 1960; Horsfield, 1972; Harland and Horsfield, 1974; Scrutton *et al.*, 1976; Harland *et al.*, in press.

This area extends from Kongsfjorden to Isfjorden. The pre-Carboniferous rocks are referred to as the Western Complex. This complex was deformed first by at least two mid-Palaeozoic orogenic episodes. The most intense of these involved westward overthrusting, nappe formation and some metamorphism. Then in the mid-Cenozoic WEST SPITSBERGEN OROGENY there was eastward overthrusting but little metamorphism. In consequence the sequence of the Western Complex occurs in tectonically separated structures, five of which are numbered below and of which sequences 3 and 4 represent the main outcrop. They are tabulated below in a postulated stratigraphic sequence.

(1) Bullbreen Gp 0.7 km, pe, cgl, lst with M or L Sil. fossils. This occurs in a large klippe or klippen.

(2) Sarsøyra Fm 0.5 km, ma, cgl., pe and volc. pe. with a probable Silurian coral. This is bounded by Tertiary graben faulting.

(3) Comfortlessbreen Gp 2-4 km, lst, till, qi, volc, dst.
St Jonsfjorden Gp 3.8 km, lst, qi, volc, dst.

(4) Kongsvegen Gp 3.1 km, pe, ma, ps.
(3) and (4) occupy the main outcrop area and are separated by a fault of unknown displacement.

(5) Vestgotabreen Fm blue schist meta volc.
This gives a Caledonian radiometric age and is associated with the Cenozoic thrust that bounds the Bullbreen Group.

(7) PRINS KARLS FORLAND

See Tyrrell, 1924; Atkinson, 1956; Atkinson, 1960; Harland *et al.*, in press.

This island, which parallels the coast of area (6) has a similar Tertiary and mid-Palaeozoic tectonic history but exposes a less broken sequence above the tillite formations compared with that of the mainland.

Forland Complex (part of Western Complex)

Grampian Gp 3.6 km, turbidite pe, cgl, qi.

Scotia Gp 1 km, sh, pe, lst.

Peachflya Gp 1.3 km, pe, sst and volc.

Ferrier Gp 0.73 m, till, gw, pe, ps, cgl.

This apparently continuous sequence is truncated downwards as it occurs in a large nappe rooted in the east (A.P. Morris, private communication).

Pinkie Fm 0.2 km, meta volc. is a tectonically isolated unit.

(8) SOUTHWESTERN SPITSBERGEN

See Garwood and Gregory, 1898; Hjelle, 1962, 1969; Flood *et al.*, 1971; Harland and Wright, in press.

From Isfjorden to south of Bellsund:

Kapp Linné Gp c.5 km, cg., till, qi and pe.

Bellsund Gp 1.4 km, pe, volc and lst.

The best sequence is seen west of Recherchefjorden. East of Recherchefjorden is another sequence of more metamorphosed rocks with feldspathites and amphibolites that could be a continuation northwards of the feldspathites of the Eimfjellet Gp (of area 9).

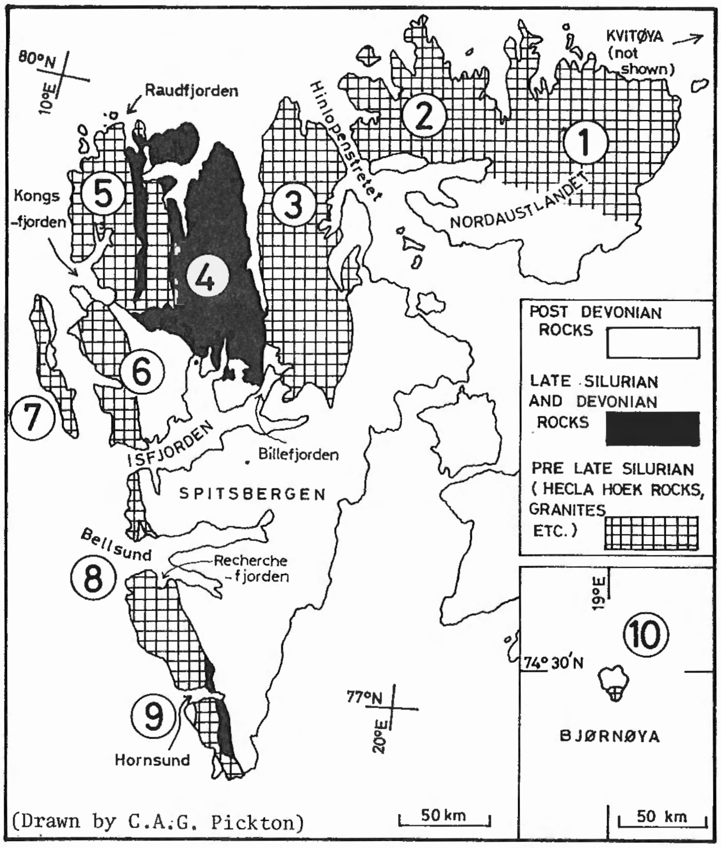


Figure 2.1. Map of Svalbard to show outcrop of older rocks involved in Caledonian diastrophism. The numbers refer to the ten areas distinguished in the text (where the place names for these areas are given) and in Figure 2.2. Other place names mentioned in the text are shown on the map. Kvitøya is about 80 km E of Nordaustlandet in latitude 80° N.

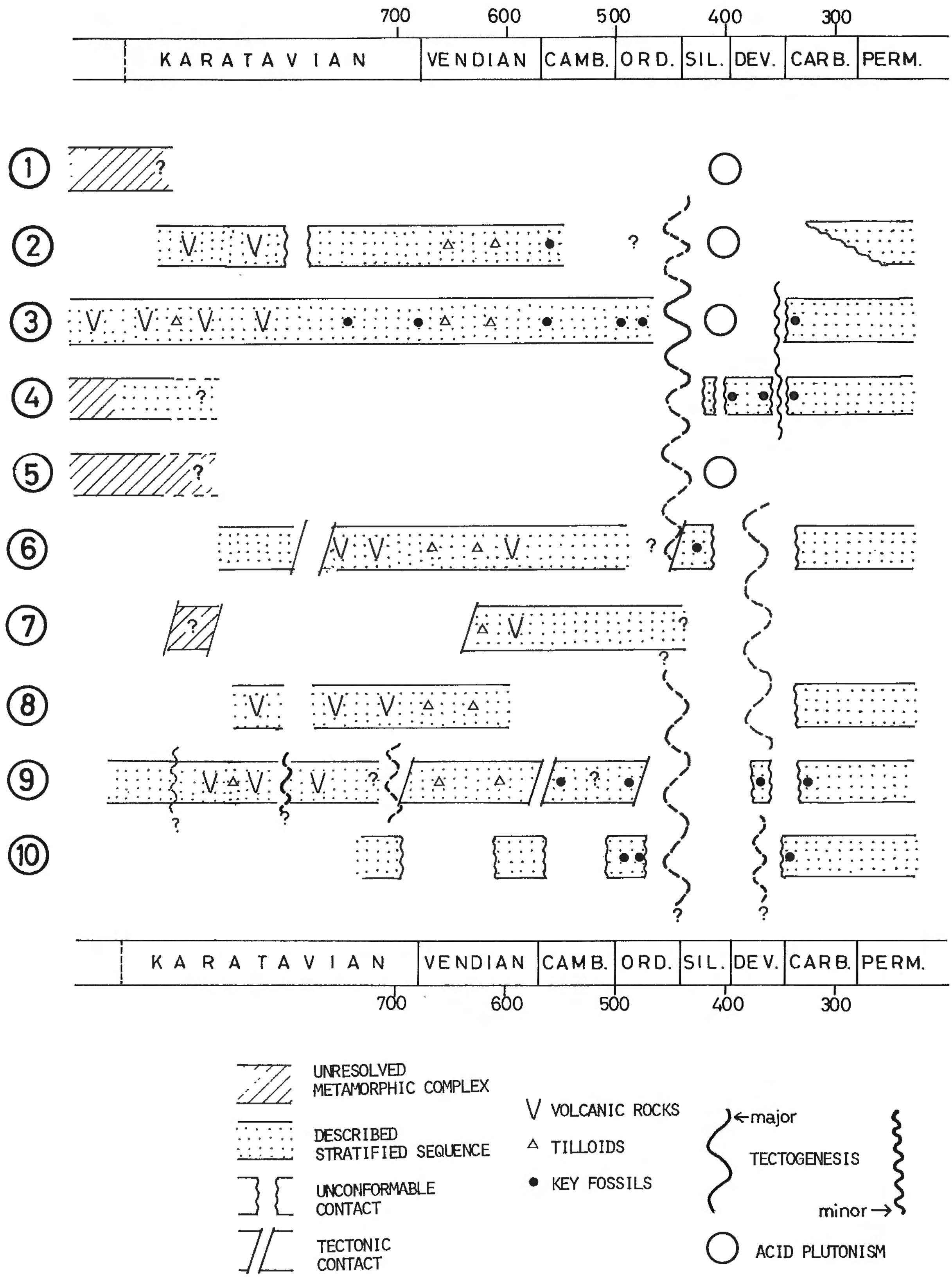


Figure 2.2. Diagrammatic plot of litho-tectonic sequences in the ten areas referred to in the text and the map (Figure 2.1). The plot is schematic and no precise ages of events should be inferred from it. It is based on a more detailed plot with stratal thicknesses in Harland and Wright (in press).

(Drawn by N.J.R. Wright)

(9) SOUTH SPITSBERGEN

See Winsnes, 1955, Birkenmajer, 1958, 1959, 1960, 1972, 1975; Flood *et al.*, 1971.

The sections north and south of Hornsund are the best known. The rocks of this area have been folded and overthrust to the E in the mid-Tertiary WEST SPITSBERGEN OROGENY and the sequence, broken largely by later faulting, is listed here in what might originally have been the unbroken sequence.

Unconformably beneath Early Carboniferous rocks occurs an Old Red Sandstone unit:

Marietoppen Gp 1 km, red sst and slt of Devonian age (Emsian-Eifelian).

This rests unconformably above the older rocks

Sørkapp Land Gp 1.4 km, lst and dst, E Ord. (Candian).

HORNSUNDIAN ? tectonic phase

Sofiekammen Gp 0.8 km, dst, lst and sh. E Camb.

JARLSBERGIAN ? tectonic phase

Sofiebogen Gp 4.7 km, pe, till, dst and lst.

Tertiary tectonic discontinuity may also represent contemporaneous tectonic break referred to as TORELLIAN tectonic phase.

Deilegga Gp 3.5 km, green pe and qi.

WERENSKIOLD ? tectonic phase

Eimfjellet Gp 1.5 km, am, fe, till and qi.

Isbjornhamna Gp 1.5 km, pe and ma.

(10) BJØRNØYA (BEAR ISLAND)

See Holtedahl, 1920a; Horn and Orvin, 1928; Krasil'shchikov and Livshits, 1974; Worsley and Edwards, 1976.

On this small island the outcrop area is small so the record is likely to be partial.

Carboniferous strata pass continuously down into the:

Røedvika Fm sst of late Fammenian to Tournaisian age.

This sets the youngest limit to the mid-Palaeozoic movements for Svalbard and suggests that the Svalbardian phase was entirely Late Devonian, but the older rocks are only slightly tilted.

Ymerdalen Fm 0.64 km lst, dst. M Ord. (Chazy)

separated by a phase of minor folding from the

Sørhamna Fm 0.175 km, pe and qi, ?Vendian.

Russhamna Fm 0.4 km, dst, L Riphean (Karatavian).

DISCUSSION

Many publications have attempted an overview of the Svalbard Caledonides with varying degrees of speculation and simplifications. *See* Nordenskiöld, 1875-1876; Nathorst, 1910; Holtedahl, 1920b; Kulling, 1932; Frebold, 1935; Bailey and Holtedahl, 1938; Orvin, 1940; Harland and Wilson, 1956; Harland, 1959, 1960; Winsnes, 1965, 1975; Harland, Wallis and Gayer, 1966; Sokolov *et al.*, 1968; Harland, 1971a and b; Harland and Gayer, 1972; Krasil'shchikov, 1973; Harland *et al.*, 1974; Birkenmajer, 1975; Harland and Wright, in press.

CORRELATION

Time-correlation of the fossiliferous Palaeozoic rocks is generally fair. The problem here is that the western sequences (especially area (7)) are of highly mobile, often turbiditic, facies and have yielded few fossils even though it seems probable that much of the rock is of Early Palaeozoic age.

One or two of the two principal levels of Varangian (Early Vendian) tillites can be traced in sequences 2, 3, 6, 7, 8 and 9 so that correlation at this horizon is reasonably secure even between different tectonic environments. Vendian microfossils from associated shales confirm the Varangian age of some of the above tillites.

The commonly stromatolitic, pisolitic, oolitic facies that underlie the tillite formations are of Vendian and later Riphean (Karatavian) age but cannot be correlated with the same precision. There is no means yet of correlating older rocks other than by similarity of neighbouring facies. These conclusions are represented in outline in Figure 2.2.

TECTONIC PROVINCES

If Svalbard were treated as one tectonic province a unified scheme, such as appears in the first page of this paper or as elaborated by Krasil'shchikov (1973) or Birkenmajer (1975), would combine all sequences (1) to (10) and so set rather rigid constraints on the stratigraphic breaks that would indicate tectonic phases. The outline of the ten sequences above presents evidence on a comparable basis with little interpretation. While acknowledging that each area may have been separated by minor faults if not major ones (cf. Harland, 1972) some groupings are suggested, namely, 1, 2 and 3; 4 and 5; 6, 7 and 8; 9 and 10. The Billefjorden Fault Zone (Harland *et al.*, 1974) was suggested as a major boundary due to substantial sinistral transcurrent Svalbardian motion so separating areas 1, 2 and 3 from the rest. A further Central West Spitsbergen Fault Zone was postulated in 1975 (Harland and Wright, in press) to separate the western sequence 6, 7 and 8 - together referred to as the Holtedahl Geosyncline (Harland *et al.*, in press) - from the central sequence (4 and 5).

Field work in 1977 confirmed the abundance of volcanic facies at least into Vendian time if not later in areas (8) as well as (6) and (7) and to Vendian time in area (9); but in areas (1), (2) and (3) volcanic facies were not evident above Lower Hecla Hoek rocks.

In conclusion three, four or even five major provinces widely separated in pre-Devonian time may have been brought together in late Devonian time by major transcurrent faults.

TECTONIC PHASES

Although plate motions tend to produce broadly synchronous deformation at plate margins, deformation need not be synchronous and the search for widespread tectonic phases may not be worthwhile.

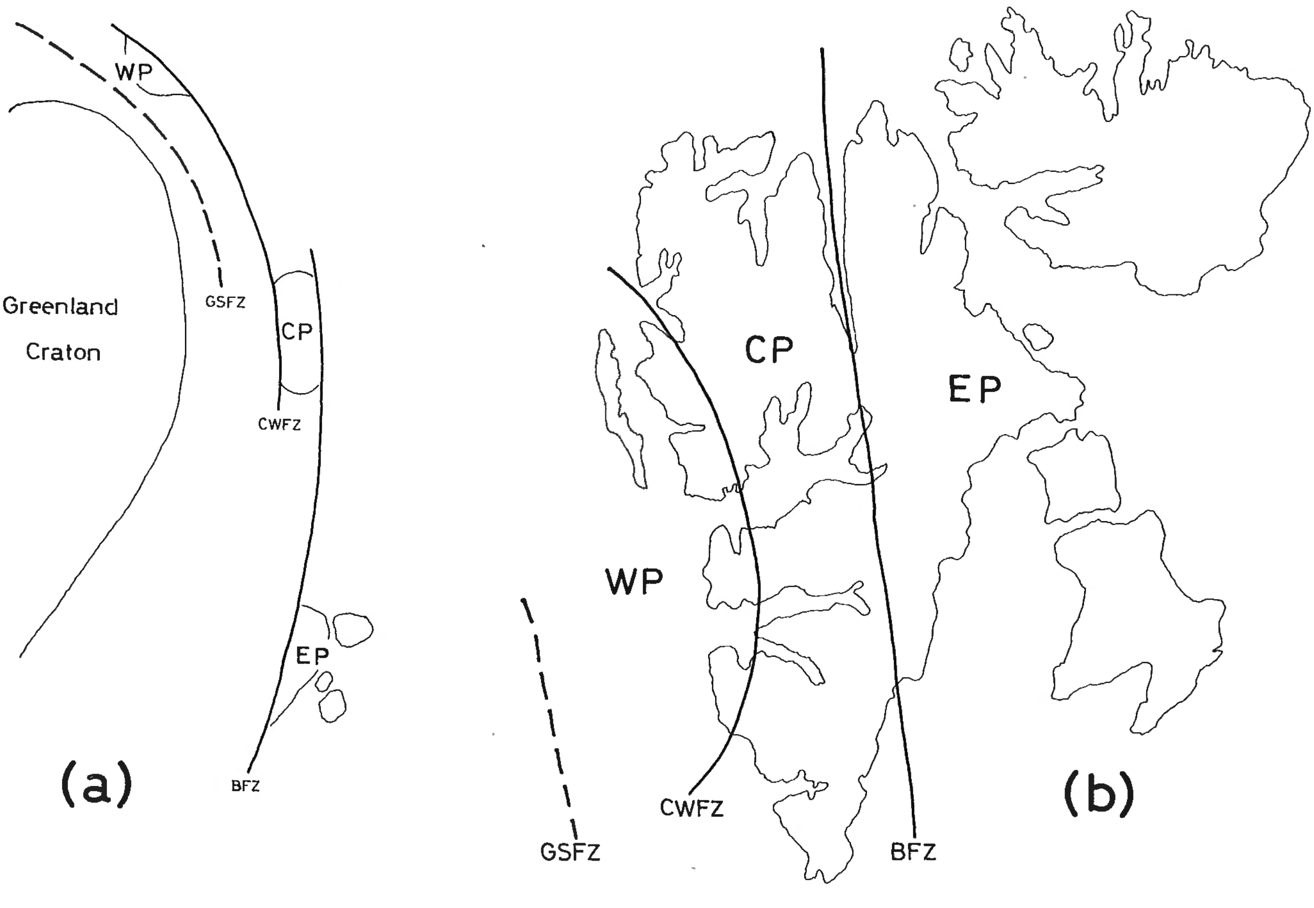
Precambrian phases are difficult to correlate in Svalbard where there has been general mid-Palaeozoic over-printing of radiometric ages. In any case Precambrian events may be regarded as pre-Caledonian. Areas 1, 4 and 5 may well conceal major Precambrian phases; in area (9) the Precambrian Werenskioldian and Torellian phases represent possible but not well-established Precambrian events.

Of Palaeozoic phases the postulated Cambrian breaks: Jarlsbergian and Hornsundian claimed only from area 9, appear to be only minor episodes. The major episode - or the main Caledonian tectogenesis in the east (areas 1, 2 and 3) and centre (areas 4 and 5), at least, corresponds to the Ny Friesland Orogeny which was post-Llanvirn and pre-Devonian - possibly pre-Wenlockian. A later phase particularly in the west (areas 6 - 9) would appear to be very late or post-Silurian. This could correspond to the Haakonian in the centre-north (area 4) but might have been more intense and of greater duration in the west, being analagous to the Ellesmerian of North America and Greenland. Certainly the latest Caledonian event was very widespread in its effect, namely the Late Devonian transcurrent faulting with transpression.

Post-Caledonian movements are evident but decrease in Carboniferous time. There is no major Variscan movement in Svalbard. These movements appear to be the final adjustments of the Caledonian *s.l.* regime.

MID-PALAEOZOIC PLATE MOTIONS

In a sequence of papers, e.g. Harland, 1959, 1965, 1969, 1971a, 1971b, 1972; Harland & Gayer, 1972; Harland *et al.*, 1974; Harland, 1975; Harland & Wright, in press,



Figures 2.3a, b. Diagrammatic maps to illustrate hypothesis of Harland & Wright (in press), in which at least three pre-Devonian provinces in Svalbard, distinguished by their litho-tectonic history, were widely separated until they were brought together by late Devonian sinistral transcurrent faulting. Figure 2.3a is a schematic pre-Devonian arrangement from Harland and Wright Figure 4; Figure 2.3b is much simplified from their Figure 3. Symbols are as follows:

Provinces	Fault Zones
WP = Western Province	GSFZ (postulated) = Greenland-Svalbard Fault Zone
CP = Central Province	CWFZ (postulated) = Central-Western Fault Zone
EP = Eastern Province	BFZ (established) = Billefjorden Fault Zone

I have attempted to define significant fault zones and develop a plate tectonic model for Svalbard through time. Earlier interpretations, (e.g. Harland *et al.*, 1974) invoked only one major fault zone within the Svalbard archipelago. The interpretation presented in this paper (Figures 2.3a, 2.3b) is essentially as put forward in Harland & Wright (in press). In this interpretation four provinces are distinguished: Eastern (areas 1, 2 and 3); Central (areas 4, 5 and 9); Western (areas 6, 7 and 8) and Southern (area 10). The Eastern, Central and Western Provinces are shown on Figures 2.3a, b; the Southern Province (Bear Island) is south of the map area of these figures. The relationship of the Southern Province to the others is questionable. A fault zone defines the boundary between the Eastern and Central provinces (Billefjorden Fault Zone); another between the Central and Western Provinces (Central-Western Fault Zone). Also relevant to this discussion is the hypothetical Greenland-Svalbard Fault Zone, believed to lie west of Svalbard and first postulated in 1965. Dividing Svalbard in this way provides a model to account for the differences in facies and timing of events in the different areas.

The Caledonian events may be summarized as EW compression in the initial Ny Friesland Orogeny in, say, later Ordovician to Silurian time. The effects of this orogeny are clearly displayed in the Eastern, and probably the Central province. This orogenic activity culminated in sinistral transpression with NS elongation in Silurian time. Sinistral strike-slip then took over, involving movement along what are now recognized as the Billefjorden and Central-Western Fault Zones.

The western sequence might have a more complex history. A possible Late Ordovician or Early Silurian episode preceded deposition of the Bullbreen Group rocks which were then deformed in late Silurian or Devonian time. Devonian strata have not been identified in this province. The facies, ages of deposition and deformation all contrast with what is known of the Caledonian sequence in the east, and in East Greenland, and have more similarity with the North Greenland and Ellesmerian orogenic provinces.

The model with late Devonian sinistral motion along the main NS faults serves to account for the differences in facies and timing of events and to juxtapose these provinces as they remained from Carboniferous through Paleocene time.

The typical Caledonian events of the Eastern Province culminated in the final closure of Iapetus. The succeeding sinistral motion along the suture (and along all the faults mentioned) probably coincided with Acadian and/or Ellesmerian orogenic episodes. These contrasts represent a change in relative plate motions that need further investigation.

Until such an interpretation be rejected it would be well to restrict the name Hecla Hoek to the eastern sequences. Harland *et al.*, in press, proposed the name Holtedahl Geosyncline for the rocks of the Western province. The southwestern and southern sequences are still in need of further elucidation.

Many regard the above views as too mobilistic. A more fixistic view, as, for example, outlined by Birkenmajer (1975) should be compared as an alternative model. An extreme fixistic model is developed by Krasil'shchikov (1973).

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3. THE NORWEGIAN CALEDONIDES - INTRODUCTION

Brian A. Sturt

Geologisk Institutt
Avd. A
Universitetet i Bergen
5014 Bergen, Norway

In preparing the synthesis of the Norwegian Caledonides we have divided the orogenic belt into five segments: 1 - northernmost Norway (Finnmark); 2 - north central Norway; 3 - south central Norway; 4 - southern Norway; 5 - the eastern marginal zone (Fig. 3.1). Articles 4-8 of this volume describe these segments. Regional coverage of the Scandinavian Caledonides is completed with the article by Gee (10), on the Swedish Caledonides. The account of the Norwegian Caledonides in this volume also includes an article by Steel (9) on the late orogenic Devonian rocks (Old Red Sandstone) which occur in the southern segment, and, less extensively in the south central segment. The segments are essentially regional subdivisions but they are also divisions which show regionally distinct patterns and which involve unlike geological elements.

The last major synthetic studies of the Norwegian Caledonides (Nicholson, 1974; Strand and Kulling, 1972) contain a wealth of information, but significant advances have been made in recent years which clarify many of the relationships. These recent advances have either not been published before, or appear in scattered publications.

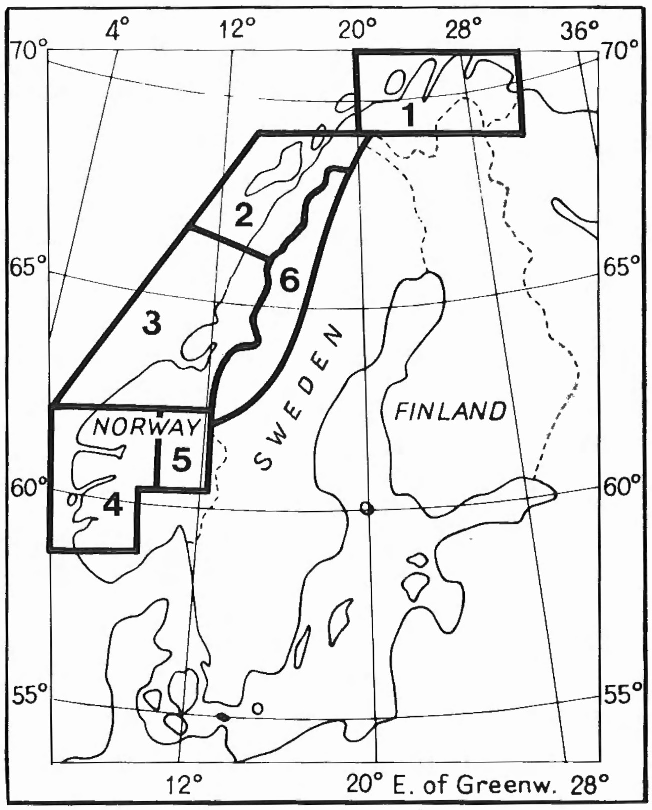
The Norwegian Caledonides are developed upon a framework of the Baltic Shield which shows a complex pattern of Precambrian orogenesis. This is reflected in the several now firmly established age provinces (Figs. 3.2, 3.3). Above the basement occurs an autochthonous cover sequence of Vendian-Silurian deposits which is now only preserved in part beneath the thrust nappes of the Caledonides. A complete Lower Palaeozoic succession is found in the Oslo region, but elsewhere the sequence is incomplete.

The nappes of Caledonian development range from little travelled parautochthon to far-travelled allochthonous units, many of which root westward of the present day geographical confines of Norway. Work spanning the last decade has shown that the nappe sequences are much more complex than previously considered, and attempts to express this in this synthesis will be found in the form of tectonostratigraphic tables. The last decade has also shown how our assessment of Caledonian elements in these nappe sequences and indeed the identification of Caledonian elements elsewhere in Norway has to be substantially modified. Many areas of gneissic and migmatitic rocks previously regarded as an integral part of Caledonian evolution have been demonstrated both by geochronological and by stratigraphical techniques to represent Precambrian crystallines. The great area of gneisses in western Norway known as the North-West Gneiss Region, previously regarded by many as Caledonian in age, has been shown to represent a window of the Precambrian crystalline foundation and to represent an integral part of the Baltic Shield. The same has been shown for windows in the north-central part of the country. In the nappe sequences many gneissic rocks, previously considered to be of Caledonian development, have also been demonstrated to represent allochthonous elements of the former crystalline basement to the Caledonian supracrustal sequences.

The timing of Caledonian orogenesis in Norway (Fig. 3.3) now appears, without any reasonable doubt, to have occurred in two main stages:

1. Late Cambrian-Early Ordovician (Finnmarkian/Grampian).
2. Middle to Late Silurian (Main Scandinavian).

The northernmost segment contains a succession of nappes which not only went through their main tectonothermal evolution during the first of these periods



1 - northernmost Norway (Fig. 4.1);
2 - north central Norway (Fig. 5.1);
3 - south central Norway (Fig. 6.1);
4 - southern Norway (Fig. 7.1);
5 - eastern marginal zone (Fig. 8.1);
6 - Swedish Caledonides (Fig. 10.2).

Figure 3.1. Index map, Scandinavian Caledonides.

IGCP Project 27, Norwegian Contribution No. 5A.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

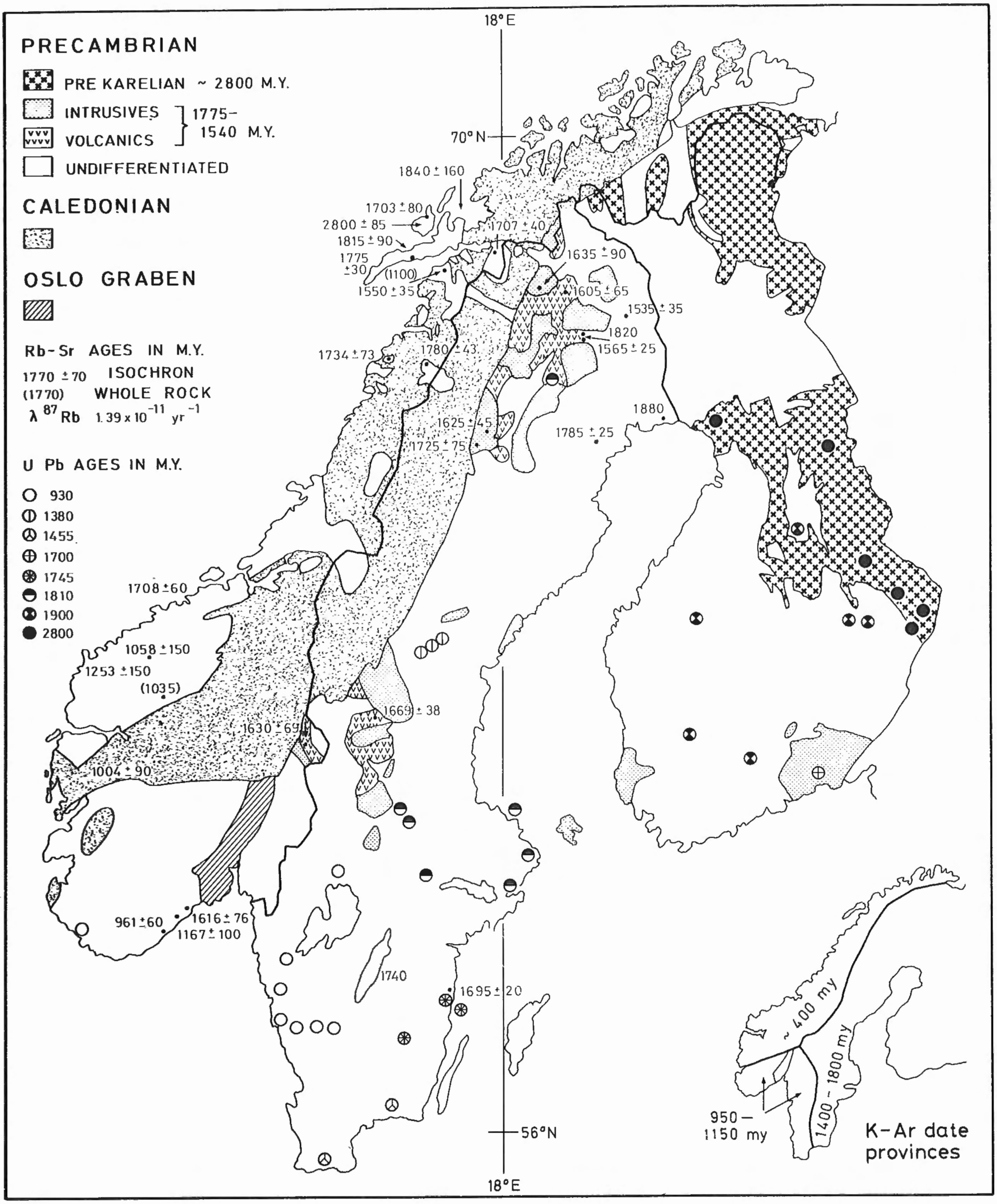


Figure 3.2. Generalized Rb-Sr and U-Pb age determinations from the Baltic Shield (from Wilson and Nicholson, 1973).

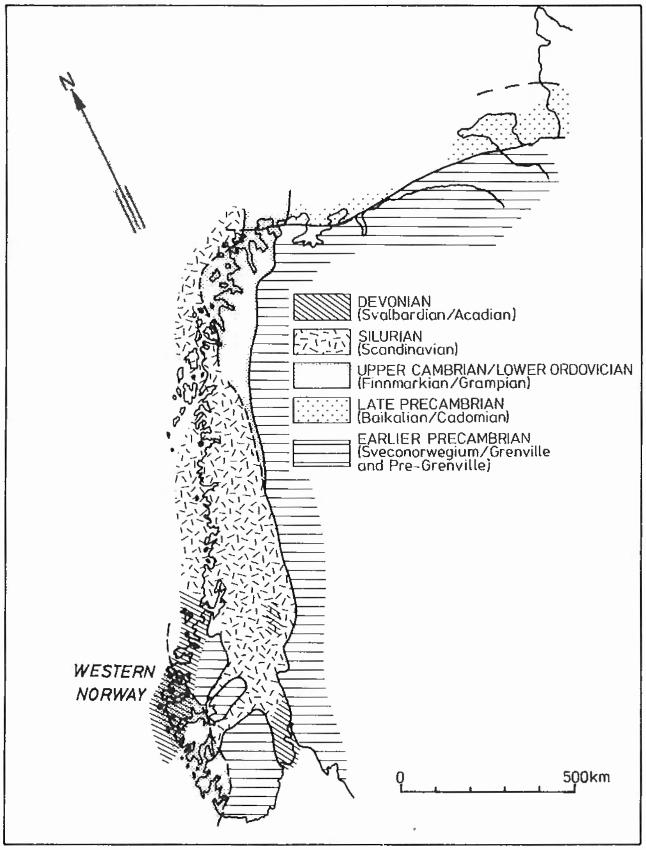


Figure 3.3. Generalized map showing approximate timing of principal zones of tectono-metamorphic evolution in the Norwegian area (modified from Roberts and Gale, in press).

(Finnmarkian) but also suffered their major "mise-en-place" at this early stage. The northernmost segment is also characterized by the unusual pattern of syn-orogenic magmatism developed during this time period. Clear evidence of major deformation and metamorphism at about this time is also seen in Western Norway, although there the substantial reworking during Late Silurian orogenesis has produced a complex structural pattern. With the exception of the northernmost segment the emplacement of the Caledonian nappes, and also much of the structural and metamorphic development within these, is related to the second of the main Caledonian phases (Scandinavian). One of the features which is emerging from current work concerns the increasingly complex relations between folding and thrusting where even major thrusts may be polyphasally refolded. The intensity of such deformation appears to increase progressively westwards.

In models of Caledonian reconstruction, significant advances have been made in relation to the identification of distinctive faunal provinces, which show decidedly different characteristics for the foreland region when compared with the allochthonous units. Within the allochthonous units geochemical work has demonstrated that many of the basic volcanic sequences bear the characteristics of ocean floor and island arc basalts, features which are important in any geotectonic and palaeogeographic reconstructions. One of the outstanding problems for such reconstructions is the presence of major overthrust units of Precambrian crystallines which occupy the highest position in many of the tectono-stratigraphic sequences.

Recent work on the Devonian of Norway has shown the substantial development of molasse sedimentation in relation to the denudation of the Caledonian mountain chain produced during the main Scandinavian phase of the Caledonian orogeny (Mid-Late Silurian). It is also apparent that the Devonian rocks are now in part in allochthonous position, and that strong folding has affected certain areas. Low-grade metamorphism is also prevalent in some of these assemblages. This phase of orogenic activity is regarded as Svalbardian (at the Middle to Upper Devonian boundary) and may well represent in the NE Atlantic region an expression of the Acadian Orogeny so strongly developed in the Appalachians.

In writing these short summaries, although we have tried to take a neutral and impartial line, the views expressed are naturally those of the individual authors; at the same time we are grateful to our many colleagues who have kindly made available to us a considerable volume of unpublished information.

Similarly we express our gratitude to Sissel Dahle who has typed all manuscripts and to Jan Lien who has drawn most of the illustrations.

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4. CALEDONIDES OF NORTHERNMOST NORWAY (FINNMARK)

Brian A. Sturt

Geologisk Institutt
Avd. A
Universitetet i Bergen
5014 Bergen, Norway

and

David Roberts

Norges geologiske undersøkelse
Postboks 3006
7001 Trondheim, Norway

INTRODUCTION

The nappes of the northernmost segment of the Norwegian Caledonides were initially emplaced during the Finnmarkian phase of the orogeny which occurred during the time interval approx. 550 - 500 m.y. B.P. (Sturt *et al.*, 1975), that is during a major period of orogenesis beginning in the Upper Cambrian and extending into Lower Ordovician time (Table 4.1, Fig. 4.1). In the southwest part of this segment, in the Lyngen-Balsfjord area, the Finnmarkian nappe complex is overlain by a younger sequence of nappes, parts of which are composed of Ordovician-Silurian strata. Similarly, a klippe of such rocks is found on the island of Magerøy in the far north. The northeastern section of the Finnmarkian nappe sequence is bounded, on Varanger Peninsula, by a major transcurrent fault - the Trollfjord Komagelv Fault Zone - to the north of which is a thick, low-grade metamorphic sequence of Riphean sediments, the Barents Sea Group (Siedlecka and Siedlecki, 1967; Roberts, 1972). Above these, with either a tectonic or a sedimentary contact, are sediments of the Raggo Group of probable latest Riphean to Vendian age; these show certain similarities with parts of the Finnmarkian nappe complex (Fig. 4.1, Table 4.1).

The foreland to the south consists of a basement of Karelian and pre-Karelian rocks which varies from high-grade gneiss terrains to greenschist facies supracrustals. These are overlain by a succession of late Precambrian and early Paleozoic rocks which comprise two sequences separated by a shallow angular unconformity. In east Finnmark the older of these two sequences - the Older Sandstone Series - is composed of two units, the Vadsø Group and the overlying Tanafjord Group. Shales from the Vadsø Group have been dated at 825 Ma (Pringle, 1973). The younger sequence, comprising in ascending order the Vestertana and Digermul Groups, ranges from basal Vendian tillites, dated at 668 ± 7 Ma (Pringle, 1973) continuously up into Tremadocian strata (Reading, 1965). To the southwest, along the thrust front, the autochthon is represented only by a thin condensed sequence, the Dividal Group or Hyolithus Zone, which passes down into northern Sweden (Fig. 4.1, Table 4.1).

Two major windows of basement show through the Finnmarkian nappes; the Komagfjord and Alta-Kvaenangen (designated the Raipas windows on Fig. 4.1). These expose the so-called Raipas Suite of presumed Karelian age which comprises sandstones, dolomites, quartzites, shales and conglomerates with thick developments of pillow lavas (Barth *et al.*, 1963). Autochthonous late Precambrian-Cambrian rocks containing tillites rest unconformably upon these older Precambrian rocks, and it can be demonstrated that the strong cleavage of the window assemblage is cut by the unconformity. In areas where the Caledonian reworking of the basement/cover assemblage is advanced this unconformity is difficult to detect.

THE FINNMARK NAPPE SEQUENCE

This consists of a succession of nappes with a range of basement/cover relationships (Figs. 4.1 and 4.2). The nappes also vary from little transported parautochthonous units of low metamorphic grade to far-travelled allochthonous nappes in which high-grade Caledonian metamorphism is widespread. The main nappe sequence is as follows:

Magerøy Nappe	- of Silurian emplacement
Kalak Nappe Complex Laksefjord Nappe Gaissa Nappe	- of Finnmarkian emplacement

In the region of the Raipas windows the two lowest units are missing and instead parautochthonous nappes of Raipas rocks may be present; these carry Caledonian metamorphic fabrics of middle to upper greenschist facies. The Magerøy Nappe contains Ordovician-Silurian rocks with metamorphism ranging from greenschist through to high-rank amphibolite facies and was emplaced during late Silurian time (Ramsay and Sturt, 1977). The nappes beneath this unit constitute the Finnmarkian tectonic assemblage (Table 4.1).

The two lowest units, the Gaissa and Laksefjord Nappes, contain a lithostratigraphic sequence similar to the foreland assemblage, of late Precambrian age (Laird, 1972; Gayer and Roberts, 1973). The Gaissa Nappe is only very weakly metamorphosed, while the Laksefjord Nappe lithologies show evidence of middle greenschist facies metamorphism. South of Alta the lowest tectonic unit, the so-called Jerta Nappe, is composed of imbricated, weakly metamorphic lithologies of Dividal Group or Bossekop Group aspect (Zwann, pers. comm.). This lies directly upon autochthonous Dividal Group (Table 4.1), and may be tentatively correlated with the Laksefjord Nappe.

The uppermost of the Finnmarkian units, the Kalak Nappe Complex (also called the Reisa Nappe Complex in north Troms), is by far the most complicated, tectonically, of all the units. This is composed of a number of discrete nappes which contain a variety of gneissic rocks of Precambrian age and a distinctive Vendian to Cambrian lithostratigraphic sequence. Examples are the Hammerfest, Kvaløy and Kvalsund Nappes (Fig. 4.2). Gneisses are locally associated with the sedimentary

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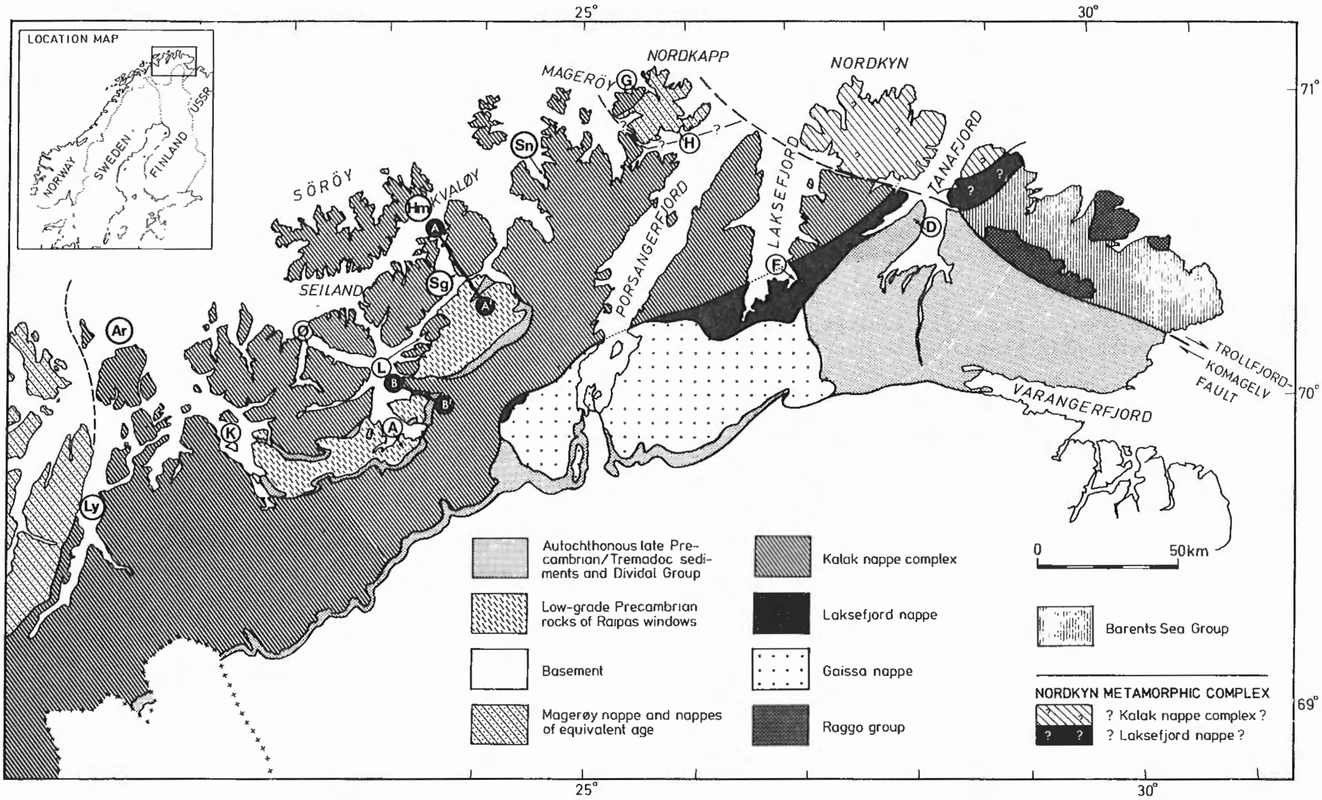
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Table 4.1

THE MAIN FEATURES OF THE FINNMARK NAPPE SEQUENCE

TECTONO-STRATIGRAPHIC UNIT	STRATIGRAPHIC AGE OF SEQUENCE AND MAIN LITHOLOGIES	PALAEONTOLOGICAL EVIDENCE	DEFORMATION/METAMORPHISM AND IGNEOUS ACTIVITY	RADIOMETRIC AGE DETERMINATIONS
MAGERØY NAPPE	Ordovician - Silurian Greywackes, shales, phyllites, schists, limestones, conglomerates.	Llandoverian Fauna Monograptids Corals Brachiopods Crinoids Trace fossils	Greenschist - amphibolite facies regional metamorphism. Prograde metamorphism and thrust fabric. Mafic, ultramafic and granitic rocks.	Not yet dated
THRUST - CONTACT / PROGRADE METAMORPHISM				
KALAK NAPPE COMPLEX — KALAK METASEDIMENTS	Late Precambrian - Cambrian Psammites, schists, limestones - with turbidites at top of succession No glaciogene sediments recorded.	Archaeocyathids of a Lr-mid Cambrian age in Falkenes Marble Group (Sørøya) Trace fossils in Klubben Psammite Group (Sørøya)	Complex deformation history. High-grade regional metamorphism (mainly amphibolite facies) with local development of migmatites. Plutonic mafic, ultramafic, dioritic, granitic, and alkaline rocks in W. Finnmark and N. Troms.	402Ma. K-Ar whole rock. (Date of late metamorphic event) 384-420Ma. K-Ar micas (Date of late metamorphic event) 480-491Ma. K-Ar nephelines (Date of main high grade metamorphism) 501-552Ma. Rb-Sr isochron ages. (Date of medium-high grade metamorphism)
	IN SOME CASES THRUST-CONTACT; IN OTHERS PRIMARY UNCONFORMITY			
KALAK NAPPE COMPLEX — PRECAMBRIAN ELEMENTS	Precambrian Two different types of basement involved in Kalak Nappe Complex. (i) Rocks of Raipas Suite in parautochthonous nappes (ii) Para- and ortho-gneisses in internal thrust sheets within Kalak Nappe Complex, either as gneiss units or together with the Kalak sediments.	None	Polyphase Precambrian deformation and metamorphism (amphibolite facies with relics of granulite facies) Polyphase Caledonian reworking. Metamorphism from greenschist to high amphibolite facies. Igneous rocks: Precambrian granites, gabbros, dolerites, etc. Caledonian - as in Kalak metasediments	1500 Ma. Rb-Sr, on granite dyke cutting gneisses (preliminary isochron from Lerrisfjord area)
THRUST CONTACT/DIAPHTHORITIC METAMORPHISM AND CATACLASIS				
LAKSEFJORD NAPPE	Late Precambrian - Cambrian (?) Sandstones, schists, phyllites, limestones, glaciogene sediments etc.	No fossils recorded	Strongly deformed low greenschist facies metasediments. Local basic dykes.	Rb-Sr isochron ages 503Ma. Slates - Friarfjord Formation (age of cleavage)
	UNCONFORMITY AT BASE OF LAKSEFJORD GROUP			
	Basement gneisses			Not yet dated
THRUST - CONTACT/DIAPHTHORITIC METAMORPHISM AND CATACLASIS				
GAISSA NAPPE	Late Precambrian - Cambrian(?) Sandstones, shales, limestones, glaciogene sediments etc.	Stromatolites in Porsanger dolomite	Little metamorphosed - slightly recrystallized sediments. Variably folded or cleaved.	Not yet dated
THRUST - CONTACT/CATACLASIS				
AUTOCHTHONOUS TREMADOC - LATE PRECAMBRIAN	Late Precambrian - Tremadoc Continuous sequence of sandstones, shales, glaciogene sediments etc.	Cambrian and Tremadocian macro- and trace-fossil faunas.	Little metamorphosed - slightly recrystallized sediments. Variably folded and cleaved.	Rb-Sr isochron ages 515Ma. Shales - Stappogeidde Formation (age of cleavage) 668Ma. Shales - Nyborg Formation (age of diagenesis)
	UNCONFORMITY AT BASE OF VESTERTANA GROUP			
	Older Sandstone Series Sandstones, shales etc.	No fossils recorded	Little metamorphosed - slightly recrystallized sediments. Variably folded and cleaved.	825Ma. Shales - Vadsø Group (age of diagenesis)
REGIONAL UNCONFORMITY				
KARELIAN AND OLDER CRYSTALLINE BASEMENT				



D - Digermul	F - Friarfjord	H - Honningsvag	G - Gjesvaer
Sn - Snøfjord	Hm - Hammerfest	Sg - Stagenes	L - Lerrisfjord
Ø - Øksfjord	A - Alta	K - Kvaenangen	Ar - Arnøy
			Ly - Lyngen

A-A', B-B', lines of sections, Figure 4.2.

Figure 4.1. Tectono-stratigraphic units, northernmost Norwegian Caledonides.

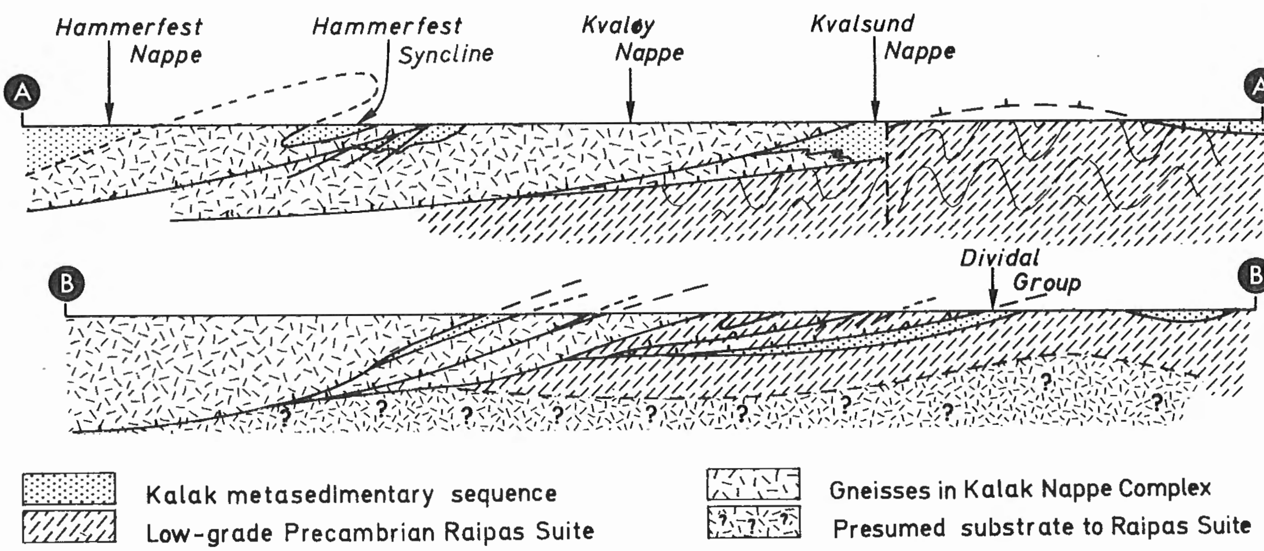


Figure 4.2. Geological cross sections A-A' and B-B' (Figure 4.1).

succession and in many cases it is not clear whether these represent Caledonian gneisses or tectonically incorporated Precambrian elements. The stratigraphy of the Caledonian cover sequence is very distinctive and can be traced throughout the segment. The type succession for west Finnmark and north Troms is found on the island of Sørøy and is as follows (Ramsay, 1971):

5. Hellefjord Schist Group - alternating, thin-bedded metagreywackes and pelitic schists - a turbidite sequence.

4. Aafjord Pelite Group - graphitic schists, mica schists with thin granoblastic quartzites.

3. Falkenes Marble Group - pure and impure marbles with calc-silicates; contain Archeocyathids of Lower Middle Cambrian age.

2. Storelv Schist Group - thick sequence of mica schists with distinctive upper and lower formations.

1. Klubben Psammite Group - thick development of cross-bedded psammites with extensive horizons of semi-pelitic schists.

This succession shows polyphasal structural and metamorphic development with two major deformational episodes: in detail, more complex, multiphase deformational sequences may be established. Fold structures range from major recumbent folds with extensive zones of stratal inversion to more simple structures. The metamorphism of the Kalak Nappe Complex was essentially within the amphibolite facies, though in the lower units greenschist facies may obtain (Sturt *et al.*, 1975). The Caledonian metamorphism of the region appears to be entirely of the medium-pressure Barrovian type. Local areas of highest amphibolite facies are present, and here migmatite complexes are developed.

One of the most significant features of the Kalak Nappe Complex is its igneous history, which is probably unique within the Caledonian-Appalachian orogen. This igneous activity is concentrated in the western part of the Kalak Nappe Complex in the Seiland Petrographic Province, within which occurred a great plethora of igneous activity showing a considerable petrological variation (Sturt and Ramsay, 1965; Robins and Gardner, 1974; Bennett, 1974). Major layered gabbro intrusions with parent magmas ranging from quartz-normative tholeiite to alkali olivine basalt were emplaced, these showing many features characteristic of such intrusions in stable tectonic environments. Major ultramafic complexes with compositions ranging from dunites through

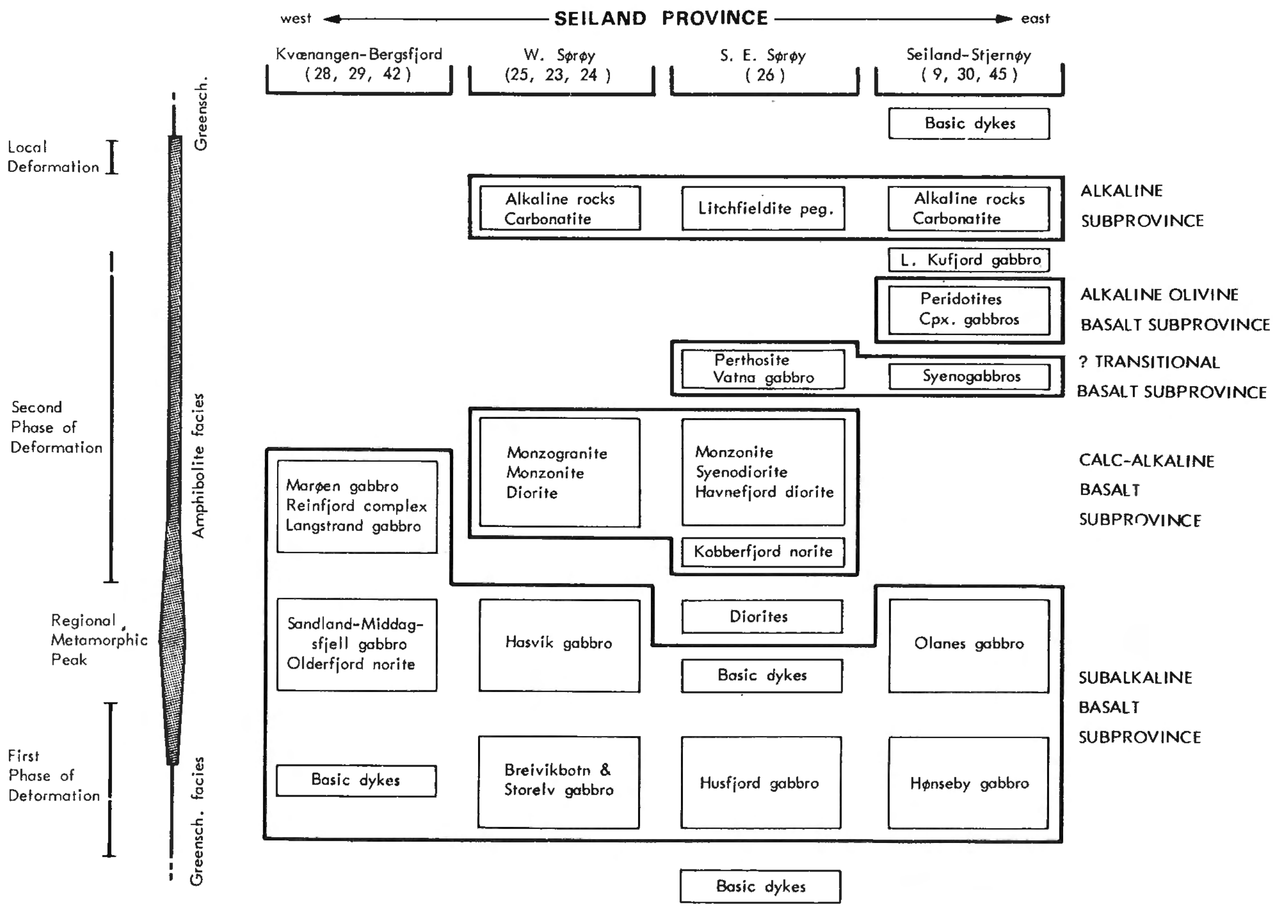


Figure 4.3. Magmatic sequence of the Seiland petrographic province (from Robins and Gardner, 1974). Numbered references are: 9 - Roberts, 1971. 23 - Sturt and Taylor, 1972; Ramsay, 1971; Sturt and Ramsay, 1965. 26 - Speedyman, 1967. 28 - Gronow, 1967. 29 - Hooper, 1971. 30 - Robins and Gardner, 1975. 42 - Bennett, 1974. 45 - Robins, 1974.

wehrlites to lherzolites occur, as do extensive alkaline complexes containing carbonatites, nepheline syenites, alkaline pyroxenites and a range of other syenitic types. A deformed alkaline pyroxenite ring-dyke complex is present on Stjernøy (Robins, pers. comm.), and elsewhere there are numerous small intrusions of granite, monzonite and diorite (Fig. 4.3).

The emplacement of the igneous complexes spans the sequence of polyphasal tectono-metamorphic evolution of the region, and all rank as syn-tectonic intrusions (Figs. 4.3, 4.4). As a result they show variable effects of tectonic and regional metamorphic processes, and in turn provide a complex sequence of contact metamorphic recrystallizations and overprinting on the envelope rocks. The general evolutionary pattern of the basic rocks is of a progressively developing alkaline trend (Sturt and Ramsay, 1965; Robins and Gardner, 1974), though exceptions do occur. The entire plutonic complex is allochthonous, deriving from a northwesterly source probably above a mantle diapir (Ramsay, 1973).

Geochronological studies in the Kalak Nappe Complex show from the Rb/Sr whole-rock isochron method (Rb^{87} = 1.39 x $10^{-11}yr^{-1}$) that the oldest members of the Seiland province were emplaced approximately 550 Ma ago (for intrusions emplaced during the D_1 sequence of strains), and that the youngest major intrusives (alkaline) date to ca. 500 Ma, these latter having been emplaced late in the protracted D_2 deformation episode (Sturt *et al.*, 1975). Attempts to date the peak of the regional metamorphism indicate ages of around 535 Ma. Studies of the slaty cleavage in the foreland and in the Laksefjord Nappe indicate that this structure was formed around 510-500 Ma ago and thus corresponds to the D_2 deformation in the Kalak Nappe Complex (Table 4.1). The mineral ages are all considered to represent cooling/uplift ages and can be divided into two groups:

(i) Nephelines, 480-494 Ma, taken to represent cooling/uplift ages for the Finnmarkian event.

(ii) Micas, 385-420 Ma, taken to represent cooling/uplift ages in relation to the subsequent main Scandinavian event.

BASEMENT/COVER RELATIONS WITHIN THE FINNMARK NAPPE COMPLEX

Primary stratigraphic contacts are seen between cover sediments and basement in the foreland and around parts of the Raipas windows. Within the nappe complexes a primary unconformity is documented from the Laksefjord Nappe where the metasediments rest on gabbroic rocks of assumed Precambrian age (Laird, 1972). In the Kalak Nappe Complex areas of gneissic rocks have long been known, though until recently they have been considered as representing gneisses generated in response to Caledonian orogenic processes. Recently a primary stratigraphic unconformity has been discovered on the island of Kvaløy (Fig. 4.1) beneath the Klubben Psammite Group (Sturt, unpub. data). The subjacent gneisses represent a series of foliated rocks of complex evolution with several phases of reworking and injection prior to deposition of the Klubben Psammite. Further south in a nappe unit probably equivalent to the gneissic rocks on Kvaløy Rb/Sr isochron work shows that late granitic dykes cutting gneisses are approximately 1500 Ma (Sturt, unpub. data). In a higher tectonic unit at Øksfjord gneisses have been dated to 1030 Ma. The implications of this are manyfold:

1. The thesis of a Caledonian origin for all the gneissic rocks of the region must be reconsidered, and except for migmatite complexes derived from specifically recognized members of the Kalak succession the status of such rocks is in doubt.

2. The unconformity on Kvaløy beneath the Klubben Psammite implies that Baltic continental crust extended for a considerable distance to the west beneath the depositional basin for the Kalak sediments, and that this crust is of high metamorphic grade and considerable antiquity.

3. As such gneisses also occur sandwiched between thrust sheets of parautochthonous rocks of the Precambrian Raipas Series, they in part represent a crystalline basement to these rocks.

4. The highest nappe yet recognized in the Finnmarkian sequence, on the island of Arnøy, truncates the full Sørøy sequence and is composed of a complex development of gneisses and granulite facies rocks. If these also turn out to be Precambrian crystallines the implication would be that virtually the entire sedimentary basin in which the west Finnmark sequence accumulated was floored by continental crust (Fig. 4.4).

The major thrust planes in the region are clearly demarcated and particularly where gneisses occur at such contacts a range of mylonitic and blastomylonitic rocks are developed. Where metasediments make contact with thrust planes quartz-mylonites are extensively developed from the psammitic rocks and a range of phyllonitic rocks from the micaceous members of the sequence. Towards thrust planes marked extensional and flattening fabrics are characteristic. Major overthrusting occurred in two distinct episodes (Sturt *et al.*, 1975):

(i) In relation to Finnmarkian orogenic development, i.e. in late Cambrian to early Ordovician time.

(ii) In relation to late Silurian (main Scandinavian) orogenic development.

In the earlier of these episodes (Finnmarkian) it can be seen that thrusting was polyphasal, and episodes of thrusting can be distinguished both early and late in the sequence of tectono-thermal development. The latest thrust movements are characterized by diaphthoretic metamorphism.

The later of these major episodes, i.e. late Silurian (main Scandinavian), is clearly seen in relation to the Magerøy Nappe on the island of Magerøy. Here, thrust movements were coeval with high-rank amphibolite facies metamorphism (Ramsay and Sturt, 1977). At the southern margin of the Finnmarkian segment the younger nappes, however, have diaphthoretic fabrics at their contacts which post-date the metamorphic maximum of the nappe lithologies. This indicates that the history of thrusting during the main Scandinavian phase of the Caledonian orogeny was probably also polyphasal.

THE MAGERØY NAPPE

The rocks of the Magerøy Nappe are found only on the island of Magerøy, which is separated from the mainland by a major fault. This unit contains a sequence dominated by metagreywackes and mica schists with subordinate limestones and conglomerates. Sporadic finds of fossils demonstrate an Ordovician-Silurian age for much of the succession (Føyn, 1966). The rocks were affected by polyphasal deformation and metamorphism presumably during late Silurian orogenesis (Ramsay and Sturt, 1971). Metamorphism is entirely of the medium-pressure Barrovian type and shows a marked increase in grade from east to west. Igneous rocks present include gabbros, peridotites, diorites and granites. As yet no geochronological information is available.

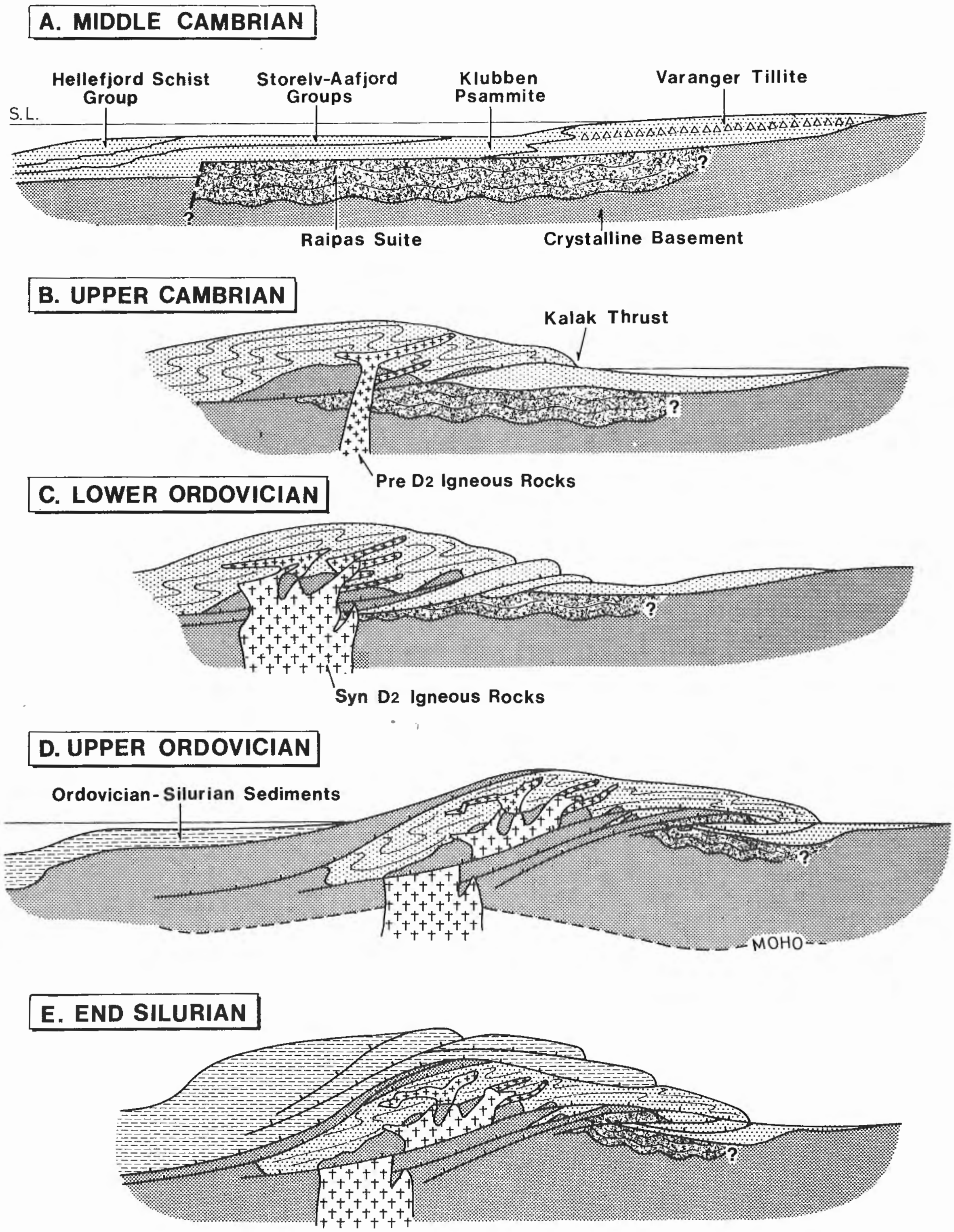


Figure 4.4. Diagrammatic profiles depicting the tectono-stratigraphic and igneous evolution of the Caledonides of Finnmark and North Troms, northern Norway (from Sturt *et al.*, 1977).

In general terms the gross lithofacies of the Magerøy Nappe resemble those of the younger nappes south of the Finnmarkian segment, west and southwest of the Lyngen Peninsula. In particular there is a striking similarity between certain tilloidal intraformational conglomerates on Magerøy and conglomerates in the lowest nappe unit west of the Lyngen gabbro. Recent fossil finds south of Balsfjord in fact indicate a Lower Silurian age for at least part of the succession (Bjorlykke, pers. comm.), thus providing an important link between the Magerøy Nappe and the Silurian-deformed nappe pile of Nordland and Trøndelag (Roberts and Gale, 1977) (Fig. 4.4).

The sequence in the Magerøy Nappe and its southerly equivalents was presumably developed in a basin sited to the west of the Finnmarkian fold belt (Fig. 4.4).

ROCKS NORTH OF THE TROLLFJORD KOMAGELV FAULT ZONE

On Varanger Peninsula two lithostratigraphical sequences have been described from north of the complex dextral-transcurrent Trollfjord-Komagelv Fault Zone; the Barents Sea Group and the Raggo Group (Siedlecka and Siedlecki, 1967; Siedlecka, 1975). These are quite dissimilar to the late Precambrian rocks of the Vadsø and Tanafjord Groups occurring in the autochthon southwest of the fault zone.

The 9 km-thick Barents Sea Group consists of proximal and distal turbidites superseded by a variable sequence of siltstones, mudstones, sandstones and stromatolite-bearing carbonates. By analogy with a comparable succession on the Ribachiy Peninsula, U.S.S.R., the Barents Sea Group has been considered as Riphean in age (Siedlecka, 1975). Some confirmation of this has been provided by K/Ar age determinations on dolerite dykes from the Båtsfjord area which indicate that the sedimentary pile was folded and weakly metamorphosed sometime prior to 640-650 Ma (Beckinsale *et al.*, 1976), and possibly before ca. 740 Ma (Beckinsale, pers. comm.). As the cleavage in the autochthon south of the fault is 515 Ma (Sturt *et al.*, 1975; Pringle, 1973) and as such dykes are completely absent in this area, the implication is that the Trollfjord-Komagelv Fault Zone separates two regions or blocks with quite different early geological histories.

Studies on the palaeomagnetism of the Båtsfjord dykes also favour the notion that the Barents Sea Group is a foreign element in the Caledonides of northern Norway and suggest that the sediments were deposited, folded and intruded by dykes before being laterally displaced, by dextral transcurrent movement, along the Trollfjord-Komagelv Fault Zone over a distance of several hundred kilometres (Kjode *et al.*, unpub. data). Most of this movement occurred between 640 to ca 530 Ma.

Until recently the Raggo Group has been considered to be composed of two formations, the Løkvikfjell and the Berlevåg Formations. It is now known that the contact between these is of thrust-fault character with the Berlevåg Formation apparently correlating with part of the Kalak Nappe Complex sequence occurring on nearby Nordkyn Peninsula (Levell and Roberts, in press). Rocks of the Løkvikfjell Formation may be equivalents of part of the Laksefjord Nappe succession. As the metamorphic fabric in the Berlevåg and Løkvikfjell Formations is thus almost certainly of Finnmarkian age, and since comparatively little displacement has occurred along the Trollfjord-Komagelv Fault Zone subsequent to the emplacement of the Kalak Nappe, this is further confirmation that the major component of transcurrence is pre Finnmarkian.

A few dolerite dykes of late Devonian age, 355 Ma, are also present on Varanger Peninsula (Beckinsale *et al.*, 1976). These occur on both sides of the Trollfjord-Komagelv Fault Zone, indicating that significant lateral translative movements had ceased by the very end of the Caledonian orogeny.

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5. CALEDONIDES OF NORTH-CENTRAL NORWAY

PROJECT 27 CALEDONIDE OROGEN

Magne Gustavson

Norges geologiske undersøkelse
Postboks 3006
7001 Trondheim, Norway

INTRODUCTION

The geology of north-central Norway (Fig. 5.1) is dominated by nappes of relatively high-grade psammitic, pelitic and calcareous metamorphic rocks with subordinate metavolcanics and with some intrusive massifs of Caledonian age. The generally granitoid Precambrian basement is exposed in a number of tectonic windows. The depositional age of the metasediments of the nappe sequence has for a long time been regarded as most probably Cambro-Silurian (Strand, 1972), but recent age determinations are indicating that parts of certain successions may be of late Precambrian age (P. Taylor, pers. comm.; A. Raheim, pers. comm.). So far the Caledonian age of the deformation and main metamorphic episodes has not been doubted.

Geochronological studies from the central part of the area indicate that the main metamorphism in this area occurred in the Silurian (Wilson, 1972). Some late-tectonic intrusives, including the largest granite massif in the Norwegian Caledonides, the Bindal massif, west of Hattfjelldal, were emplaced at the very end of Silurian time (Priem *et al.*, 1975). The great strike extension and relatively constant thicknesses of many sedimentary units over large areas of Nordland and Troms are suggestive of fairly stable depositional conditions. It is also probable that the trend of larger fold structures and the mountain chain in general, N-S to NE-SW, is approximately parallel to the longer dimension of the original sedimentary basins (Gustavson, 1966; 1972). The complex tectonics and high-grade metamorphism have so far precluded any attempt at geotectonic reconstruction for this region. It can be suggested, however, that an eventual Caledonian subduction zone was most probably situated somewhere to the west of the present coast line.

THE AUTOCHTHON: PRECAMBRIAN BASEMENT AND COVER

Continuous areas of Precambrian basement east of the Caledonian nappes are mostly confined to the Swedish side of the national border in this region, but basement rocks enter Norwegian territory in Dividalen, about 100 km E of Narvik at about 68°30'N (Fig. 5.1). It is here covered, to the west, by an approximately 100 m-thick autochthonous sequence of Lower Cambrian sandstone and shale, dated by fossils (Vogt, 1967). Large areas of Precambrian basement are exposed in Lofoten-Vesterålen and in adjacent coastal districts. The Precambrian west of the Ofoten Synform is continuous with basement within the Caledonian nappe area in Tysfjord. To the east, in Sweden, this basement is overlain by allochthonous Precambrian rocks of the "Syenite Nappe".

A number of tectonic windows are exposed within the Caledonian nappe region, the largest being the Rombak Window (Fig. 5.1) where the Precambrian is overlain by a thin sequence of supposed Lower Cambrian conglomerate and sandstone. The autochthonous sediments are also preserved in small areas of "down folded" rocks within the window. West of the Ofoten Synform conglomerate and sandstone in a small area on Hinnøy may represent the original cover sediments, but these are now metamorphosed and probably thrust some distance along the surface of the Precambrian granite below (Gustavson, 1972).

Sandstones of Dividal Group (Hyolithes Zone) type are also present around Precambrian windows in Troms (e.g. the Mauken Window) (Gustavson, 1966; Berthelsen, 1967). Tectonic windows in Nordland, other than the Rombak Window, are: Rishaugfjell, Nasafjäll and Børgefjell in an eastern zone and Heggmovatn, Svartisen, Sjona and Høgtuva in the west, to mention the largest ones (*see* Fig. 5.1). Autochthonous sediments have been described from the eastern margin of the Nasafjäll Window (which lies in Sweden) and the Børgefjell Window (Greiling, 1974). Around the coastal windows no autochthonous cover has yet been found.

In a very general way, the Precambrian windows along the national border and those along the coast of Nordland-Troms can be regarded as forming elongate, anticlinal structures, between which the Caledonian nappes lie folded in a major synclinorial depression.

Apart from the Lofoten-Vesterålen area, where a wide range of radiometric ages have been reported from about 3500 Ma to about 1100 Ma (Taylor, 1974) absolute age determinations so far seem to indicate that the largely granitoid basement of Nordland is about 1700 - 1800 Ma old, that is Late Svecofennian. Some indications of a younger Precambrian thermal event of about 1100 - 1200 Ma have also been reported.

TECTONO-STRATIGRAPHIC UNITS IN THE NAPPE SEQUENCE

The accompanying table (5.1) outlines the main features of the individual nappe units and their relative positions in the tectono-stratigraphic column. The areal distribution of the different units is shown on the sketch map (Fig. 5.1).

The lower nappes

The lowermost thrust unit in Eastern Troms is composed of phyllonitic and mylonitic rocks, including Precambrian granite slices and psammitic metasediments with some dolomite lenses (Gustavson, 1972; Kalsbeek and Olesen, 1967). It can probably be correlated with the Abisko Nappe of the neighbouring Torneträsk area (Sweden). This implies that the lowermost thrust nappe, the Rautas Nappe Complex, is absent in eastern Troms (Mortensen, 1972). Equivalents of the Abisko Nappe have also been inferred west of the Rombak Window, on both sides of the Ofoten Synform. Correlations to the south are conjectural. Correlation has been proposed between the psammitic/semi-pelitic Meløy Group above the western basement window of Svartisen and the Juron Quartzite of the Seve-Köli Nappe Complex (Nicholson and Rutland, 1969).

IGCP Project 27, Norwegian Contribution No. 5C.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Another possibility is that the Meløy psammites, together with similar rocks above other basement culminations, are structural equivalents of the Abisko Nappe. At the Børgefjell Window the Fjällfjäll arkoses (Zachrisson, 1969) form a basal nappe in the allochthonous sequence.

The Seve-Köli Nappe Complex

This major structural unit covers Norwegian ground in the following areas:

(a) The Hattfjelldal area,
(b) South of the Nasafjäll Window,
(c) The Sulitjelma area, and
(d) Eastern Troms-Ofoten.

The rocks present in the first three areas belong to the generally greenschist facies Köli part of the nappe complex. Fossils found in area (c) are of probable Middle Ordovician age. In the Hattfjelldal area correlations with Swedish areas indicate an Ordovician and Silurian age for the Köli rocks. The metasedimentary sequences above the Abisko Nappe in East Troms are metamorphosed to amphibolite facies or high greenschist facies and Seve-Köli distinction is not easy, but it would seem that the lowermost parts with large amphibolite intrusives correspond directly with Seve rocks in the Torneträsk area (Mortensen, 1972). A larger part of the sequence may, however, belong to the Köli, indicating a considerably higher metamorphic grade than usual for this part of the nappe complex.

In the far north, recent fossil findings in the Balsfjord area of Central Troms indicate a Llandovery age for the alternating calcareous and pelitic sequence of that area (Bjorlykke, pers. comm.). Relations of these rocks to the succession in South Troms are not clear. A relatively thin schist and psammite sequence separates the fossiliferous rocks from the Precambrian basement of the Mauken window. The sedimentary sequence is clearly allochthonous above the basement but there may be more than one major thrust plane between the Precambrian and the fossil-bearing succession. In Table 5.1 the Silurian strata have been tentatively placed in the Seve-Köli Nappe Complex (Köli part), but this position may be revised as mapping proceeds in the area.

The Rödingfjäll and Gasak Nappes

In recent literature on the Caledonian sequences these two nappe units have been regarded as equivalents (Nicholson and Rutland, 1969). New information from the area west of the Nasafjäll Window shows that this opinion can no longer be maintained, as marble horizons (Fauske Marble Group) belonging to the uppermost Gasak sequence are here overlain by the Rödingfjäll Nappe (Gjelle, 1974). As the two nappes have their main distribution in different areas where they are in the same structural position relative to the Seve-Köli rocks below, they are here treated together.

The Rödingfjäll Nappe (Fig. 5.1) has its main distribution in the southern part of the area. Its amphibolite facies mica gneisses, marbles and amphibolites show a distinct metamorphic and tectonic boundary against the Köli rocks below. To the west the nappe is overlain by another high-grade nappe unit, the Helgeland Nappe Complex, of similar lithologies (Ramberg, 1967). In addition to the metasedimentary rocks a large gabbro body, the Umbukta gabbro, is situated within the Rödingfjäll Nappe.

The Gasak Nappe (Kautsky, 1953) of the Sulitjelma area includes the Sulitjelma Schist Sequence and the Sulitjelma Gabbro (Nicholson and Rutland, 1969). As in the Rödingfjäll Nappe staurolite and kyanite are not uncommon in schists of appropriate composition. In central Troms an upper nappe with marbles and mica gneisses, partly with kyanite and staurolite, overlies the Seve-Köli Nappe Complex. In the central part of the Ofoten Synform the Niingen Group of mica gneisses constitutes an upper unit, possibly forming a separate nappe. Another possibility is that an inferred slide at the base of the Narvik Group defines the base of a separate nappe complex which would then include the gneissic Narvik Group with the Råna norite massif and the iron ore-bearing Salangen Group of pelitic and calcareous sediments (Gustavson, 1972). Lithologies are similar in several respects to those of the Gasak and Rödingfjäll Nappes. (Figure 5.1 is drawn in accordance with this alternative).

In the areas south of Tromsø the uppermost thrust unit may represent still another Rödingfjäll nappe equivalent, but here again our knowledge about intervening areas is too scanty (Olesen, 1971; Landmark, 1973).

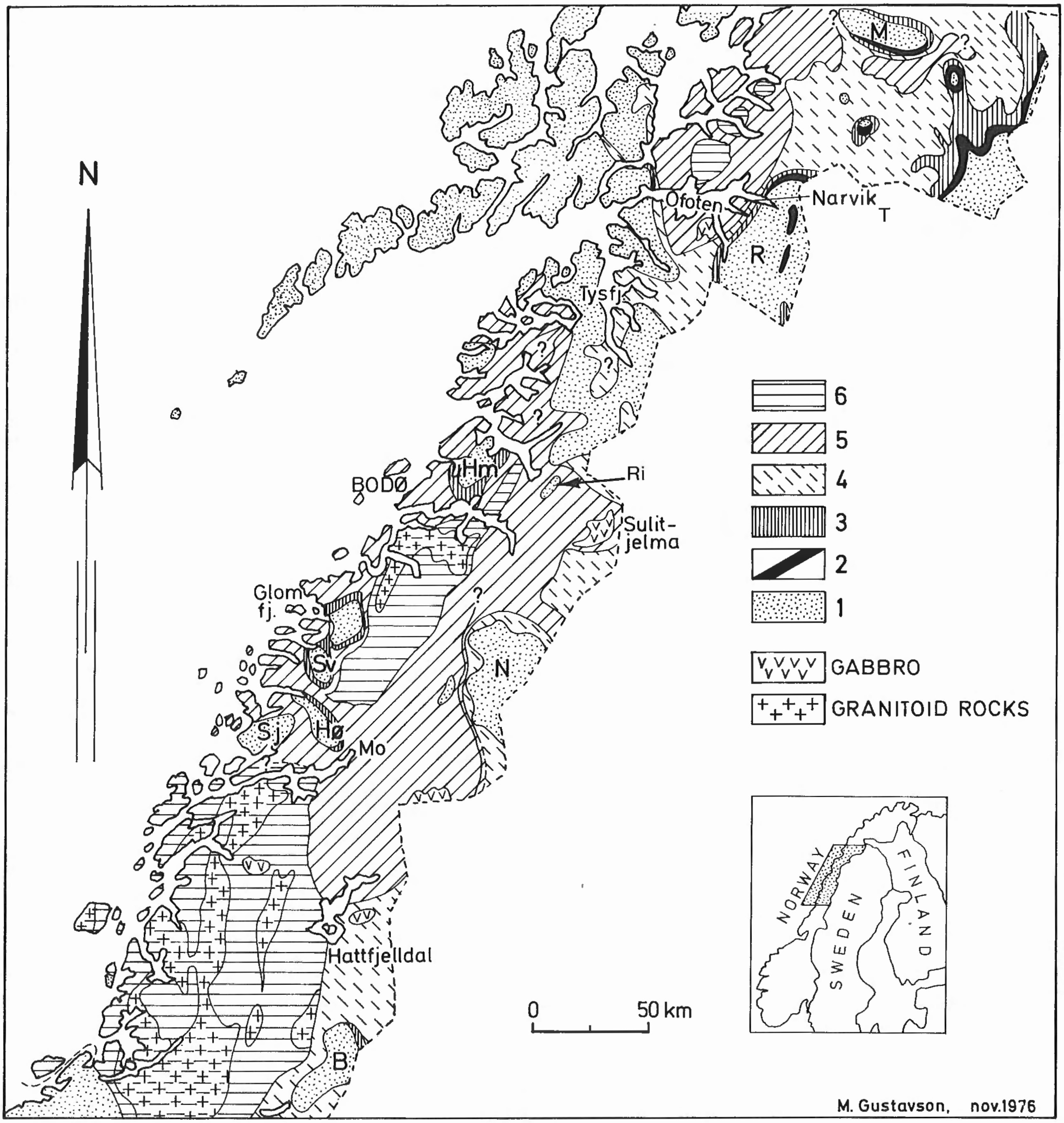
The upper thrust units

Relations between the Dunderland Group with its iron ores, marbles and pelitic formations and the Rödingfjäll Nappe lithologies to the east are not known. So far, no major tectonic break has been discovered. If a major part of the Råna sequence (Bugge, 1948) is to be included within the Rödingfjäll Nappe, and accepting that the thrust zone below the Narvik Group in Ofoten is equivalent to the Rödingfjäll thrust, this would lead to the structural and stratigraphic correlation of the Råna and Ofoten iron ore-bearing formations. The status of the Fauske Marble Group is conjectural (Nicholson and Rutland, 1969). If it is considered as forming the upper part of the Gasak Nappe, it cannot be correlated in a structural sense with the marbles of the Dunderland Group for reasons discussed in connection with the Gasak/Rödingfjäll Nappe relationship.

Both the Fauske Marble Group and the Dunderland Group are, with some intervening schist sequences, overlain to the west and northwest, respectively, by the Beiarn Nappe, the highest of the major structural units in central Nordland (Nicholson and Rutland, 1969; Wells and Bradshaw, 1970; Bennett, 1970). In the southern area the Helgeland Nappe Complex is in a similar position (Ramberg, 1967; Gustavson, 1972). Although embracing similar rock types (mica genisses, marbles and igneous rocks ranging from ultramafic to granitoid types), a direct correlation of the Beiarn and Helgeland Nappes is contradicted by their apparently different age of emplacement. While the thrust plane below the Helgeland Nappe Complex is a typical late structure, post-dating the two major fold episodes and the main amphibolite facies metamorphism, the Beiarn Nappe transport has been described as of F_1 age.

Correlations with the Ofoten - South Troms area are uncertain, but one possibility already mentioned requires that the Niingen Group forms a separate nappe in a position similar to that of the Beiarn Nappe.

The age of the rocks in these uppermost nappes is, as for the other higher-grade nappes, unknown. Some low-grade (greenschist facies) rocks of Ordovician age in the coastal area have been tentatively correlated by the author with parts of the Köli Sequence (Gustavson, 1975), As the low-grade rocks seem to be transitional into higher-grade ones in the Helgeland Nappe Complex it has been concluded that at least some parts of the Complex are Lower Palaeozoic rocks.



1 - Precambrian basement.
2 - Dividal Group (Lower Cambrian autochthon) and possible equivalents.
3 - Lower nappes, partly sparagmitic, in eastern area and psammitic-semipelitic rocks above basement culminations in the western areas.
4 - Seve-Köli Nappe Complex
5 - Rödingfjäll and Gasak Nappes and possible equivalents, correlations uncertain north of Bodø-Fauske.
6 - Upper nappe units (Helgeland Nappe Complex, Beiarn Nappe, Niingen Nappe (?)).

Abbreviations, Precambrian tectonic windows:

M - Mauken	R - Rombak	Hm - Heggmovatn	N - Nasafjäll
Sv - Svartisen + Glomfjord	Hø - Høgtuva	Sj - Sjona	B - Børgefjell
	T - Tornetrask	Ri - Rishaugfjell	

Figure 5.1. Tectono-stratigraphic units of the north-central Norwegian Caledonides.

Table 5.1

The Nordland-South Troms Nappe Sequence (North-Central Norway).

	TECTONO-STRATIGRAPHIC UNITS		MAIN LITHOLOGIES	FOSSIL EVIDENCE	IGNEOUS ROCKS	CHARACTER OF DEFORMATION & METAMORPHIC GRADE	RADIOMETRIC AGES
ALLOCHTHONOUS UNITS	UPPER NAPPES	Beiarn Nappe Helgeland Nappe C. Niingen Group (Ofoten)	Pelitic and semipelitic rocks, calcareous limestones, local dolomite or limestone conglomerates.	None	Basic, intermediate and granitoid intrusives (large). Ultramafics. Some horizons of volcanic (basaltic) origin.	Deformation strong 3 or 4 episodes. Metamorphism amphibolite f. with migmatisation, local greenschist facies.	(2) 865-560 m.y. (Rb/Sr whole-rock, Beiarn nappe), deposition age. (5) 424 ± 26 m.y. Bindal granite, post-or late F_2 emplacement age (Rb-Sr whole-rock.
	THRUST CONTACT (F_1 BEIARN NAPPE, POST-F_2 HELGELAND NAPPE						
	RØDINGFJELL AND GASAK NAPPES	Rødingfjell Nappe (Helgeland) Gasak Nappe (Sulitjelma) Narvik & Salangen Groups (Ofoten)	Dolomitic and calcareous limestone, pelitic and calcareous pelitic rocks, sedimentary iron ores, minor sandstones and conglomerates. Amphibolites (various origins).	None	Gabbros with associated ultramafics. Small isolated ultramafics. Granite and trondhjemite bodies. Some amphibolites may be intrusives, some of volcanic origin.	Deformation strong, (4 episodes), metamorphism generally amphibolite facies, migmatites common in some areas (mainly basal part)	Preliminary results from Rb-Sr determinations indicate a Later Precambrium age for the deposition of iron/ore-bearing sequence (2,3). (4) 409-423 m.y. (K-Ar biotites, Sulitjelma), latest metamorphism.
	THRUST CONTACT (PRE - F_2) ?						
	SEVE-KØLI NAPPE COMPLEX	Only Køli part in Nordland, also Seve, E. Troms	Pelitic and semipelitic rocks, greywackes, volcanics and sedim. with volc. debris, conglomerates. Gneisses, amphibolites and marbles in Seve part.	Mid-Ordovician fossils, Sulitjelma (Køli) (Various fossils Sagelv-vatn, Balsfjord. Llandovery Relations of sequence to Seve-Køli Complex somewhat uncertain)	Ultramafic rocks common. Amphibolite/metagabbro, large in E. Troms. Granite bodies (small) and veins, are common in Seve part.	Deformation strong, 3 or 4 def. episodes Metamorphism: In Seve, amphibolite facies, in Køli greenschist facies.	(4) 430-360 m.y., (K-Ar and Rb-Sr mica dates), latest metamorph. 420 m.y. (30) (Pb model age, galena, Balsfjord, Troms).
	THRUST CONTACT (F_1, AT LEAST PRE-F_2)						
	LOWER NAPPES	Abisko Nappes (Troms) Fjellfjell arkose (S. Helgeland)	Psammitic and semipelitic rocks common (partly sparagmites)	None	None	Planar foliation developed, folding locally strong. Rel. low-grade (mostly greenschist facies, biotite zone)	None
	THRUST CONTACT (POST- METAMORPHIC, POST-F_2 ?)						
AUTOCHTHONOUS UNITS	AUTOCHTON COVER	Dividal Group (Troms)	Shale, siltstone, sandstone, conglomerate	Lower Cambrian fossils (E. Troms)	None	Slight thrusting upper part, low- or non-metamorphic	None
	UNCONFORMITY						
	LATE SVECOFENNIAN BASEMENT, LOCALLY YOUNGER (LØDINGEN GRANITE 1415 ± 80 m.y. (2))						.707 - 1780 m.y. (Rb-Sr WHOLEROCK). 1100 - 1200 m.y. REGENERATION

References for radiometric age determinations:

2 - Taylor, 1975
4 - Wilson, 1972
5 - Priem *et al.*, 1975
See also Moorbath *et al.*, 1963.

COMMENTS ON STRUCTURAL HISTORY

In those parts of the region where detailed structural work has been carried out, at least three episodes of folding have been described. All fold episodes have been ascribed to the Caledonian orogeny and, as mentioned earlier, indications so far point to a Silurian age of deformation.

At least two of the fold episodes (generally F_1 and F_2) seem to have been rather intense and resulted in patterns of isoclinal to tight folds in most areas. In the coastal districts of Nordland the granitoid Precambrian basement became involved in the F_1/F_2 folding, such that it now forms cores to the large fold nappes of the Svartisen - Glomfjord region (Rutland and Nicholson, 1965). In the other areas, involvement of basal massifs was less intensive, but clearly of an early age in the deformation history. Folding in the metasedimentary sequences into isoclinal or subisoclinal folds on a major or minor scale seems to have been much the same in all the allochthonous units although variations due to the competency contrasts of the various lithological units are the rule. Our knowledge of the structural development is still too poor, however, to allow the conclusion that all nappe units share a common deformational history. The partly different ages of thrusting of the various nappe complexes indicate that differences in the tectonic history are present.

The final thrust movements on the basal thrust planes above non-metamorphic autochthon in the east are clearly post-metamorphic, i.e., they post-date the peak of the regional metamorphism. In some areas the main metamorphism is referred to F_1, but frequently this must have continued into the F_2, or even post-F_2, stage. The thrusting of the Seve-Köli Nappe Complex has been regarded as of F_1 age. In some areas, e.g. the South Troms area, movements occurred on the Seva thrust plane also in later episodes (Gustavson, 1972; Mortensen, 1972).

It has been shown that the Rödingfjäll Nappe thrust plane was deformed by regional folds on N-S axes (Ramberg, 1967). It is possible that these folds correspond to F_2 folds in the Helgeland Nappe Complex and it may be therefore, that thrusting of the Rödingfjäll Nappe is pre-F_2 in this southern district. In the Sulitjelma area thrusting of the Gasak Nappe occurred between the D_1 and D_2 deformations of that segment of the orogen (Henley, 1970). Like the Seve Köli Nappe, thrusting of the Beiarn Nappe has been related to F_1. It has already been mentioned that the Helgeland Nappe, lying in a similar structural position, is a late, post-F_2, structure. In East Troms similarly, the Upper Nappe has been emplaced during late thrust movements.

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6. CALEDONIDES OF SOUTH CENTRAL NORWAY

David Roberts

Norges geologiske undersøkelse
Postboks 3006
7001 Trondheim, Norway

INTRODUCTION

From Dombås in the south (62°N) to the Hattfjelldal area of southern Nordland (65°30'N) the Caledonides are extensively exposed in two major districts: (1) the Trondheim region, including Smøla, and (2) the Grong-Bindal-Hattfjelldal region. An additional though no less important element in the picture is that of the Old Red Sandstone districts of coastal Trøndelag (the county containing Trondheim) and at Røragen, east of Røros. The Trondheim and Grong regions are separated by the Grong-Olden Culmination of peripherally foliated Precambrian crystallines passing from Sweden across to the western basement area of Caledonized Svecofennian supracrustals (Fig. 6.1). On its east and south sides the Trondheim region is bordered with a tectonic contact by the Late Precambrian 'sparagmites' and by a chain of small windows of basement porphyry and granite exposed in an antiformal structure.

Palaeontological evidence from different parts of the Trondheim region denotes that the Silurian-deformed and metamorphosed stratigraphical sequences extend from at least the Tremadocian up into the Llandoverian (Vogt, 1945). Biostratigraphic control is unfortunately lacking from the metasedimentary succession north of the Grong-Olden culmination, but there is no reason to doubt the gross correlation with the lithostratigraphy just to the south. Geochronological data, in conjunction with the pre-orogenic and late-orogenic biostratigraphy, place the climactic Caledonian deformative event in this segment of the orogen in the mid-Silurian, in late Llandoverian to Wenlockian time (Wilson *et al.*, 1973). In coastal Trøndelag the oldest, post-Wenlock, Old Red Sandstone rudites are certainly Downtonian and possibly of Ludlovian age (Siedlecka and Siedlecki, 1972).

THE TRONDHEIM REGION: TECTONOSTRATIGRAPHY

Mapping and research over the past decade have reaffirmed the long-held though not undisputed notion that the Lower Palaeozoic succession in this area is essentially allochthonous, deriving from a primary location to the northwest of the present Atlantic seaboard and constituting a part of the Seve-Köli Nappe Complex (Table 6.1). The principal tectonic unit, the Trondheim Nappe (Wolff, 1967), covers the greater part of the region from Trondheimsfjord east to the sparagmite-basement window border zone, and from Dombås in the south to the Grong culmination in the north. This nappe displays a complex, polyphase internal deformation. It is also increasingly clear that this metamorphic allochthon is capable of further dissection into separate disjunctive and conjunctive units (Gale and Roberts, 1974), and local thinner nappes of these and older rocks have been demonstrated in eastern, northwestern and southern areas (Wolff, 1976; Gee, 1975; Guézou *et al.*, 1972).

The regional stratigraphy within the Trondheim Nappe, established in the classical Hølonda-Horg area in the west (Vogt, 1945), comprises five groups:

1) Gula Group - age uncertain, possibly mainly Precambrian;
2) Støren Group - pre-Arenig to Arenig;
3) Lower Hovin Group - Lower to Middle Ordovician, possibly into Upper Ordovician;
4) Upper Hovin Group - Upper Ordovician to ? Lower Silurian; and
5) Horg Group - age in doubt, assumed Lower Llandoverian.

In other parts of the region local names have been adopted for correlatives of these groups. Of these one can mention the stratigraphy in eastern districts: Fundsjø Group (Støren), Sulåmo (Lower Hovin), Kjølhaug (Upper Hovin) and Slågån Groups (Horg) (Wolff, 1967). The full stratigraphical succession within the nappe has been termed the Trondheim Supergroup (Gale and Roberts, 1974).

The Gula Group underlies the central part of the Trondheim region in the core of an antiformal structure (Wolff, 1967; Roberts, 1967; Roberts *et al.*, 1970; Olesen *et al.*, 1973) and reappears in western districts extending down into Surnadal, and in the northwest flanking the Tømmerås window. In these marginal districts the Gula lies with tectonic contact upon discontinuous Vendian meta-arkoses of Särv Nappe (sub-Seve) character. The Gula comprises a complex sequence of mostly middle to upper amphibolite facies pelitic to psammitic schists with subordinate amphibolites, marbles and a variety of intrusives including occasional serpentinized ultrabasics. Migmatization is locally pervasive. The contact zone with the suprajacent Støren or younger groups is of tectonic character (Gale and Roberts, 1974; Guezon *et al.*, 1972). Thus the Gula appears to constitute an allochthonous unit in its own right. If this is confirmed, then it would be reasonable to refer to this as the Gula Nappe.

The Støren to Horg Group succession and equivalents is a broadly continuous, mostly low-grade assemblage of sedimentary and volcanic rocks of Köli affinity. In recent years much attention has been paid to the 2.5 - 3 km-thick sequence of basaltic lavas (greenstones) of the Støren Group which have revealed remarkably consistent ocean-floor tholeiite characteristics, with incipient island-arc type lavas appearing higher up as well as in the Lower Hovin Group (Gale and Roberts, 1974). The succeeding Lower Hovin, Upper Hovin and Horg Groups and their eastern correlatives constitute a heterogeneous assemblage of, firstly, conglomerates, fossiliferous sediments and carbonates, and local serpentinite lenses and marginal-basin to island-arc lavas and pyroclastics; and, later, of flysch-type metagreywacke-phyllite alternations with thick, lensoid, deep-marine conglomerates in which strike-parallel and transverse facies variations are frequently very rapid (Siedlecka, 1967; Roberts, 1972).

In keeping with the tectonic and geochemical evidence, this volcanic arc/back arc basinal assemblage and underlying Støren oceanic crust is considered to have been translated, or obducted, eastwards upon the

IGCP Project 27, Norwegian Contribution No. 5D.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Table 6.1

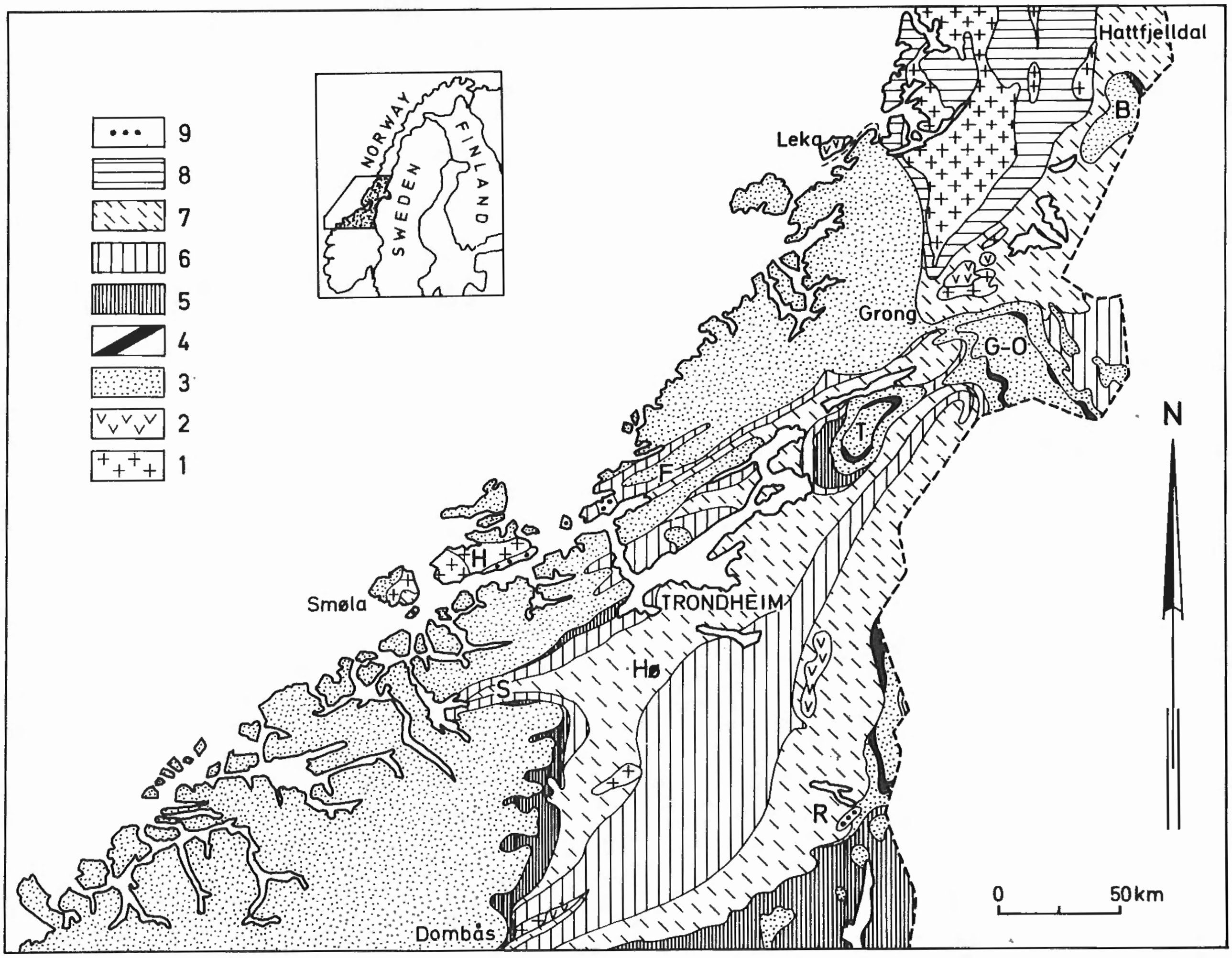
Principal features of the south-central Norwegian Caledonides.

TECTONOSTRATIGRAPHIC UNITS		MAIN LITHOLOGIES AND STRATIGRAPHIC AGE	PALAEONTOLOGICAL EVIDENCE	DEFORMATION, METAMORPHISM, PLUTONIC ACTIVITY	RADIOMETRIC AGE DETERMINATIONS
	OLD RED SANDSTONE	Downtonian (? or Ludlovian) – M. Devonian Breccias, conglomerates, sandstones, arkoses, mudstones	Plant fossils: L. to M. Devonian Dictyocaris and eurypterids: ? Ludlovian – L. Downtonian	West coast – open folding, faulting Røtangen – Two fold episodes, very low-grade metamorphism, faulting	None directly; but 363 ± 15 m.y. K/Ar biotite from lamprophyre dyke, Ytterøy [37]
	UNCONFORMITY				
ALLOCHTHONOUS UNITS	**HELGELAND NAPPE COMPLEX**	Assumed Vendian to Silurian Psammitic to pelitic schists and gneisses, carbonates, conglomerates, local amphibolites, basic to acidic lavas and tuffs.	None	Polyphase deformation, 3 or 4 episodes. Metamorphism in amphibolite facies, local migmatitisation; greenschist facies in west. Basal thrust post-F_2. Serpentinites, gabbros, granites.	424 ± 26 m.y. Rb/Sr isochron from late phase differentiates of the Bindal granite.[25]
	THRUST CONTACT				
SEVE-KØLI NAPPE COMPLEX (Including Trondheim Nappe)	**GRONG KØLI, STØREN NAPPE (TRONDHEIM KØLI), SMØLA SEQUENCE**	Pre-Arenig to L. Silurian Basaltic lavas, conglomerates, pelitic to psammitic metasediments, metalimestones, acid to basic tuffs, flysch sequences. On Smøla – andesitic to basaltic lavas and limestones.	Rich fauna of graptolites, trilobites and brachiopods in parts of the Støren Nappe. Also gastropods, corals, conodonts, cystoids and molluscs. On Smøla – gastropods	Polyphase deformation, 3 main episodes. Metamorphism middle greenschist facies locally low amphibolite facies. Basal thrust pre-D_1 (to syn-D_1). Gabbros, diorites, trondhjemites, granodiorites, lensoid serpentinites	438 – 402 m.y. K/Ar micas and phyllites from the Støren Nappe.[2] 504 m.y. Rb/Sr biotite, trondhjemite, Grong.[38] 420 m.y. Pb model age, galena, ore zone in greenstone, Grong district.[22]
	THRUST CONTACT				
SEVE-KØLI NAPPE COMPLEX (Including Trondheim Nappe)	**GULA GROUP ALLOCHTHON, SEVE NAPPE**	? Precambrian to ? Cambrian Various schists, gneisses, psammites, amphibolites, migmatites, local conglomerates and marbles.	None: but Dictyonema (Tremadocian) of controversial position at one locality (Gula Group or Støren Group).	Complex polyphase deformation. Mostly middle to upper amphibolite facies metamorphism ± migmatisation locally lower grade. Various ultramafic to granodioritic intrusives.	418 – 415 m.y. K/Ar whole-rock[2] 412 m.y. K/Ar muscovites, pegmatites[2] 468 – 418 m.y. K/Ar hornblendes[2] 407 – 403 m.y. K/Ar bioties, trondhjemite[2] 473 m.y. Rb/Sr isochron, trondhjemite[39]
	THRUST CONTACT				
ALLOCHTHONOUS UNITS	**LOWER NAPPE(S), SÄRV NAPPE EQUIVALENS**	Vendian – ? Cambrian Feldspathic sandstones, arkoses, siltstones.	None	2–3 fold episodes	Not yet dated
	THRUST CONTACT				
AUTOCHTHONOUS UNITS	**AUTOCHTHONOUS SEDIMENTS OLDEN & TØMMERÅS**	Vendian – Cambrian Basal conglomerate, quartzites, phyllites, limestone.	None	At least 2 deformation episodes. Greenschist facies metamorphism.	Not yet dated
AUTOCHTHONOUS UNITS	**"SPARAGMITE" SEQUENCE IN SOUTHEAST**	Late Precambrian – Cambrian Arkoses, phyllites, conglomerates, local limestones, tillites.	None	2 fold episodes; weak, lowest greenschist facies metamorphism	Not yet dated
	UNCONFORMITY				
AUTOCHTHONOUS UNITS	PRECAMBRIAN CRYSTALLINE BASEMENT			Caledonian	380 – 430 m.y. Caledonian overprinting (Rb/Sr, K/Ar)[15,40]
				Sveconorwegian	1000 – 1100 m.y. Regeneration[15,40]
				Gothian	1468 m.y. Grong-Olden porphyries (Rb/Sr isochron)[38]
				Svecofennian	1708 – 1880 m.y. W. basal gneiss region (Rb/Sr isochrons)[40,41] 1800 – 1900 m.y. Namsos gneiss region (Rb/Sr whole rocks).[38] 1750 m.y. Børgefjell (Rb/Sr whole rock)[42]

References for radiometric age determinations:

2 - Wilson *et al.*, 1973.
15 - Råheim and Sundvoll, 1976.
22 - Moorbath and Vokes, 1963.
25 - Priem *et al.*, 1975.
28 - Oftedahl, 1964.
37 - Priem *et al.*, 1968.
38 - Z.W.O. Lab. Isotopen-geologie, Amsterdam, 2nd Progress Report, 1968.
39 - Berthomier *et al.*, 1972.
40 - Brueckner, 1972.
41 - Pidgeon and Råheim, 1972.
42 - Z.W.O. Lab. Isotopen-geologie, Amsterdam, 1st Progress Report, 1967.

Gula at an early stage in the mid-Silurian orogenesis. The designation Støren Nappe has been adopted for this far-travelled allochthonous unit (Gale and Roberts, 1974). The age of the Gula Group is of renewed interest here. Recent geochronological studies from the Surnadal district suggest that it may be older than hitherto assumed, perhaps largely Precambrian (A. Råheim and B. Sundvoll, pers. comm.); and, in addition, it may possibly carry traces of a late Cambrian or early Ordovician thermal event (Wilson *et al.*, 1973; Berthomier *et al.*, 1972; Guesou, 1975). If this is so, and in view of its allochthonous nature, then the Gula Group should be divorced from the Trondheim Supergroup. The Gula does, in fact, appear to show close similarity to the Seve in Sweden and its tectonic position beneath the Trondheim Köli is comparable. Certain Swedish authors have virtually denied the surface extension of Seve into Norway (Zachrisson, 1973), although the tectonic and lithologic arguments for its occurrence, as the Gula, are strong.



1 - Granite, trondhjemite, quartz diorite.
2 - Gabbro (+ Leka ultramafics).
3 - Precambrian basement.
4 - Autochthonous sediments in window areas.
5 - Autochthonous "sparagmites" in southeast; allochthonous meta-arkoses of Särv type in west and northwest.
6 - Gula Group allochthon and equivalents; Seve in Grong-Olden district.
7 - Köli sequences: Grong Köli, Trondheim Supergroup (minus Gula), Smøla succession.
8 - Helgeland Nappe Complex.
9 - Old Red Sandstone.

Abbreviations: B - Børgefjell, H - Hitra, G-O - Grong-Olden culmination, S - Surnadal, T - Tømmerås, R - Røragen, F - Fosen, Hø - Hølonda

Map compiled from data in several references cited in text; for the southwest Trondheim area also from Rohr-Torp (1972) and Wolff (1976). The map differs slightly from Table 6.1 in that the allochthonous Särv-type meta-arkoses are here included with the sparagmite sequence. The Trondheim Nappe embraces units 6 and 7 within the confines of the Trondheim region.

Figure 6.1. Principal tectonostratigraphic units of the south-central Norwegian Caledonides.

A peripheral Caledonian element in the Trondheim region, sensu lato, is that of the supracrustal sequence on the island of Smøla. This comprises a succession of low-grade basaltic to andesitic volcanics with subordinate acidic extrusives, conglomerates and a fossiliferous limestone formation of Middle Ordovician age (Fediuk and Siedlecki, in press). Contact relations with a gneiss complex of assumed Precambrian age are nowhere seen. Following a folding and low-grade metamorphic event the Ordovician rocks were extensively intruded by quartz diorites (Fig. 6.1); the latter occur as clasts in nearby Downtonian conglomerates. Preliminary radiometric work indicates that the diorite intrusion on Smøla dates to Early Ordovician time (Råheim and Sundvoll, pers. comm.)

THE GRONG-BINDAL-HATTFJELLDAL REGION: TECTONOSTRATIGRAPHY

Three principal lithotectonic units can be recognized north of the Grong-Olden culmination up to the latitude of Hattfjelldal:

1) the Seve supracrustal complex;
2) a low-grade metasedimentary and metavolcanic Køli succession;
3) the Helgeland Nappe Complex (Table 6.1).

The Seve, represented mainly by schists, gneisses and amphibolites, is exposed only in the southeasternmost part of this district. The greenschist facies Köli sequence is widely developed above this, from Grong northeastwards to beyond the Børgefjell basement window. Lithologies can be traced without much difficulty across the Grong culmination into the Snåsa district and further into the low-grade Trondheim Supergroup complex. Relationships around the Børgefjell massif are complicated. Quartzites and phyllites believed to be autochthonous, are present locally below a thin 'lower nappe' of arkoses, but further work is required.

The stratigraphy and tectonics of the Grong Köli succession are the subject of current detailed investigations and the full geological picture is not yet clear. Northeast of Grong a major island-arc volcanic unit (Gale and Roberts, 1974) composed largely of greenstone lavas with a variety of keratophyric to rhyodacitic extrusives is associated with a plutonic infrastructure of gabbros and later diorites and trondhjemites (Halls *et al.*, 1977). The trondhjemites intrude and engulf the greenstones and both these rock-types occur in an overlying polymict conglomerate. This forms the base of a thick sequence of calcareous psammites, phyllites and conglomerates, the Liming Group (Foslie, 1957, 1958); and a succeeding mixed phyllite, quartzite, keratophyre and greenstone assemblage with limestone near the top. Earlier the basal volcanics were referred to a separate nappe, the Gjersvik Nappe (Foslie, 1957, 1958; Oftedahl, 1956), and considered as Støren Group correlatives, but on stratigraphic and geochemical grounds they are currently thought to be younger than the Støren (Roberts, 1975). The overlying Liming and later sediments are devoid of fossils, but comparison with neighbouring areas in Sweden denotes a probable Upper Ordovician to Lower Silurian age. The main Caledonian deformation and metamorphism occurred in Middle to Upper Silurian time; as yet the only radiometric age determination from the Grong Köli is that of a Pb isotope model age of 420 ± 70 Ma on galena from an ore zone in the lower greenstones (Moorbath *et al.*, 1963). Preliminary Rb-Sr dating of a major trondhjemite massif suggests a late Middle Ordovician intrusive age (Råheim and Sundvoll, pers. comm.).

The Køli rocks are tectonically overlain to the northwest by the Helgeland Nappe Complex (Gustavson, 1975) (in the present area the Rødingfjäll Nappe, which overlies Köli north of Hattfjelldal, is absent). The Helgeland Nappe Complex is mostly an amphibolite facies complex of mica schists, gneisses, marbles and calc-schists, local amphibolites and a variety of igneous rock-types including the prominent Bindal granitic massif west of Hattfjelldal. Somewhat lower grade metasediments and metavolcanics have been described from some western coastal districts; these are also considered to belong to the Helgeland Nappe Complex. On Leka (Fig. 6.1), similar low-grade volcano-sedimentary supracrustals include basal ultramafic rocks, gabbros and hawaiitic volcanics and a younger formation of island-arc greenstones and quartz-keratophyres within a calc-greywacke, phyllite and conglomerate succession (Prestvik, 1974). Whether these rocks belong to the low-grade part of the Helgeland Nappe Complex or to the Køli is undecided.

The climactic Caledonian, thermotectonic event in the Helgeland Nappe Complex is thought to date to middle to late Silurian time. Xenoliths of foliated metasediments, some carrying regional F_2 folds, occur in the Bindal massif of dioritic to granitic rocks, and the later differentiates of this massif have provided a Rb-Sr isochron emplacement age of 424 ± 26 Ma - uppermost Silurian (Priem *et al.*, 1975).

BASEMENT-COVER RELATIONSHIPS AND REGIONAL TECTONICS

Extensive modern investigations and mapping have shown that the concept of a regionally established Vendian to Cambrian autochthon in this part of the Norwegian Caledonides required serious revision, and that only a very small proportion of these supracrustals can be regarded as *in situ* sediments. In the north a thin metasediment veneer considered to be autochthonous is present locally around the Børgefjell granite gneiss basement (Fig. 6.1), but in most places either a "lower nappe" of arkoses or the Köli sequence rides allochthonously upon the crystallines. To the south a thin crust of graphitic phyllite, quartzite and limestone occurs in autochthonous position upon the Precambrian Olden granite and porphyry massif (within the Grong-Olden culmination) (Gee, 1975) and is succeeded with tectonic contact by similar but foliated basement crystallines. The Caledonian remobilization of this gneissic basement, recently mistakenly interpreted as constituting the Offerdal Nappe (Gee, 1974, 1975), increases markedly westwards. Interpretation of the basement-cover situation around the Tømmerås window and beneath western and southwestern margins of the Trondheim Nappe has been somewhat contentious over the years, opinion varying from autochthonous (Peacey, 1964; Oftedahl, 1964) to almost totally allochthonous interpretations for the older cover sediments (Wolff, 1976; Gee 1974, 1975). Regional studies in the Tømmerås district (Wolff, 1976) and excursion investigations (Gee, 1974) favour the allochthon hypothesis. Only thin sediments of Olden autochthon affinity appear to be *in situ* above this basement massif. Further west the allochthonous cover-on-basement relationship is camouflaged by tectonometamorphic overprinting and has to be inferred. On Smøla the contact is nowhere seen but the fold style in the Ordovician cover is simpler and the metamorphism of low grade (Fediuk and Siedlecki, 1977).

The Silurian-transported and deformed nappe pile above the "Caledonized" basement is constructed from a variety of lithologies. These range from Precambrian crystallines though Vendian arkosic psammites to Silurian pelites. The lowermost allochthon equivalents of the Swedish Offerdal and Särv Nappes are of limited or uncertain occurrence in this part of Norway, and the extent of the Seve, Grong and Trondheim Köli, and Helgeland Nappe Complex, have been dealt with earlier in this synopsis. Westward thinning of nappe units reported from Nordland (Nicholson and Rutland, 1969) and central Sweden (Zachrisson, 1969) is less evident in this region though not denied. This nappe geometry is disturbed by locally complex internal deformation of certain nappe units, by difficulties encountered in the west in distinguishing between Seve, higher grade Köli and Caledonized supracrustals of possible Svecofennian age, and by superposed components of strain relating to heterogeneous stretching associated with diapiric doming and gravity collapse processes.

A polyphase, tectonometamorphic development of the several nappe units has been documented from all parts of the region with three or four principal fold episodes the first two of which are responsible for the macroscopic structures and main thrusting (Guézou *et al.*, 1972; Roberts *et al.*, 1970; Olesen *et al.*, 1973; Rui, 1972). The regional metamorphic fabrics were largely imposed early in this history with a minimum age of 438 ± 12 Ma for the last metamorphic phase in the Trondheim region (Wilson *et al.*, 1973). A contact metamorphic mineralogy around gabbros in this same region is overprinted by regional metamorphic assemblages (Olesen *et al.*, 1973; Birkeland and Nilsen, 1972). Trondhjemites were intruded in at least three phases; the larger bodies are mostly of Ordovician age while the youngest emplacement is syn- to post-tectonic and mid-Silurian.

Major fold structures are of variable development and trend. Open to tight antiforms and synforms of Caledonoid axial orientation, c. NE-SW, deform the flattened and stretched early isoclines and regional foliation, with early lineations mostly approximating E-W to SE-NW (Roberts, 1967). Large areas of inverted sequences relate either to syn-metamorphic or post-metamorphic structures. In the Trondheim region, a central mushroom-antiform dominates the picture, exposing Gula in its core, with overturned synclines of the part of the Trondheim Supergroup known as the Støren Nappe (see above) to the west and to the east (Wolff, 1967; Roberts, 1967; Olesen *et al.*, 1973). East of Tømmerås, in the Hegsjøfjell area, the major antiform is itself deformed by the more open eastern syncline. In the south of the region the picture is less straightforward and other solutions have been proposed (Gee, 1975). North of the Grong-Olden culmination partial inversion relates to an overturned synclinal structure; this is transected by the Helgeland Nappe Complex. Thrusting developed at various stages during the construction of the nappe pile from pre-D_1 in the case of the Støren Nappe (Gale and Roberts, 1974) to post-D_2 for the Helgeland Nappe Complex (Gustavson, 1975). Cumulative allochthon translations, the sum total of a variety of strain increments, may be calculated in terms of hundreds of kilometres (Gale and Roberts, 1974; Gee, 1975).

Following the construction of the main nappe pile there is widespread evidence for gravitational sagging or collapse which produced a younger episode of mesoscopic folds with flat-lying axial surfaces, the folds varying in style and character depending on the pre-collapse attitude of the layering (Roberts, 1967, 1971; Ramberg, 1967). This also produced boudinage and extensional features in appropriate locations, thus adding to the regional flattening and thrust translation, and is an interpretation which has received support in recent regional syntheses (Gee, 1975). An interesting late structure, observed by the author in many areas from the Trondheim region up to Børgefjell, is that of a near-vertical kink fold set of remarkably consistent approximately N-S trend. This denotes a reimposed regional horizontal shortening, albeit weak, following the vertical compression relating to gravitational collapse (Roberts, 1971).

DOWNTONIAN-DEVONIAN SEDIMENTATION AND TECTONICS

Sedimentary sequences classified as late-orogenic with respect to the Caledonian cycle are those of the Old Red Sandstone areas of coastal Trøndelag and Nordmøre - Hitra, the Ørlandet area of Outer Fosen and the archipelago southeast of Smøla - and of Røragen along the southeast margin of the Trondheim region (Fig. 6.1, Table 6.1).

The Old Red sediments of the coastal districts have been described by Siedlecka and Siedlecki (1972), Fediuk and Siedlecki (in press), and Siedlecka (1975). The chronostratigraphy of the Old Red molasse of the coastal districts is based on finds of *Dictyocaris* and eurypterids on Hitra and islands southeast of Smøla and of plant fossils in the Outer Fosen area. Sediments range in age from Lower Downtonian, possibly uppermost Ludlovian, to Middle Devonian. Sedimentological studies have shown that the deposits are alluvial and fluvial, braided river accumulations. Deposition occurred in what was evidently a tectonically active zone, probably in faulted intramontane basins, with rapid facies variation reflecting the crustal movements. Southeast of Smøla and on Hitra a basal breccia contains blocks of local quartz diorite up to 1.5 m, rarely 3 m, across. Conglomerates, sandstones, arkoses and mudstones constitute the bulk of the successions; southeast of Smøla the sequence is over 3.7 km thick, on Hitra 1.3 km. In the Ørlandet area a similar Old Red succession totals at least 2.5 km in thickness. The coastal Old Red sediments were folded and faulted subsequent to their deposition. The main fold structure is that of a gentle to close syncline of NE-SW trend with the southeastern limb locally faulted. The folding is considered to relate to the Svalbardian parorogenic movements of lowermost Upper Devonian age, with the faulting partly Devonian and partly Mesozoic or Cenozoic (Oftedahl, 1975).

The Devonian of Røragen is a unique element in the geology of southeast Norway (Holmsen, 1973; Roberts, 1974). What little remains of the original succession comprises conglomerates, breccias, sandstones and shales going over to phyllites and is about 1.2 km thick. Plant fossils and their spores indicate a Lower Devonian age. Again, the depositional evironment was one of bajadas with irregular fluvial interludes in a fault-controlled intramontane basin, with periodic crustal movements. The succession was later folded along two trends and subjected to very low-grade metamorphism with the pelitic lithologies in particular exhibiting distinct lepidoblastic textures. The early folding followed E-W axes, and could be related to similar trending folds in the Devonian of west Norway. Later minor folds deform the phyllite fabric along NE-SW axes, and local minor thrusting and faulting completed the diastrophism. The deformation of the Røragen Devonian is considered to be Svalbardian (Frasnian) and it has been suggested that the main folding of the Oslo region Lower Palaeozoic foreland succession may date to this period rather than Gedinnian (Roberts, 1974).

No igneous rocks have been reported from the south-central Norwegian Devonian districts, but an ENE-WSW trending lamprophyre dyke on the island of Ytterøy in Trondheimsfjord has yielded a K-Ar biotite age of 363 ± 15 Ma (Priem *et al.*, 1968). This was considered to be an intrusion age, around the Middle to Upper Devonian boundary.

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CALEDONIDES OF SOUTHERN NORWAY

Brian A. Sturt and Asbjørn Thon

Geologisk Institutt
Avd. A
Universitetet i Bergen
5014 Bergen, Norway

INTRODUCTION

The southern Norwegian Caledonides show a complex pattern of evolution, dominated by a series of far-travelled thrust nappes which contain rocks of Precambrian age as well as those generated during the Caledonian cycle. The autochthonous cover to the underlying Precambrian basement can in many cases be followed for a considerable distance westwards and has generally very low metamorphic grade, implying that the main zone of Caledonian metamorphism lay west of the present distribution of allochthonous Caledonian metamorphics. In the northwestern section a Caledonian development of metasediments and metavolcanics occurs in the coastal regions north of Sognefjord, though their precise relations to the basement substrate have not been clarified. Subsequent to the major Caledonian diastrophism and nappe transport, a series of basins accumulated coarse Devonian clastic assemblages. These were subjected to deformation (Svalbardian Phase) and very low grade metamorphism. The basins are now partly in allochthonous positions.

THE NAPPE AREA OF THE SOUTHERN NORWEGIAN CALEDONIDES

This contribution will deal essentially with the complex of metamorphic nappes. The eastern nappes with their low metamorphic state will be covered by K. Bjørlykke in the description the Eastern Marginal Zone (Article 8).

The Hardangervidda-Ryfylke area

This classical nappe area has been subjected to considerable restudy in recent years, and a reasonably adequate model is now available (Naterstad *et al.*, 1973; Andresen, 1974b). The nappe sequence is emplaced above an autochthonous basement containing both Precambrian crystallines and a Cambro-Ordovician cover sequence (Table 7.1). The basement rocks appear to form two distinct units and a primary stratigraphic unconformity is assumed between them. The oldest complex is one of crystalline gneisses (Table 7.1), and it is assumed that the major metamorphism and dominant foliation of these rocks was completed prior to the deposition of the Telemark Suite. The supracrustal sequence of the Telemark Suite includes metasediments and metavolcanics; emplaced into these rocks is a series of granitic - dioritic plutons. These rocks were subjected to Precambrian folding and metamorphism (Andresen *et al.*, 1974), the latter being essentially of greenschist facies, though locally higher grade metamorphism occurs.

The autochthonous cover sequence, although thin, is extensively developed beneath the nappe sequence. Palaeontological evidence shows that the sequence ranges from Lower Cambrian to Llanvirn-Llandeilo age (Andresen, 1974a). A well-preserved primary stratigraphical unconformity is observed in a number of localities. The metamorphism of this sequence appears to be essentially of low greenschist facies, implying that the Caledonian thermal maximum must have lain considerably further west.

The nappe sequence (Table 7.1) shows a succession of distinctive tectono-stratigraphic units of which only two of the five identified units (Holmasjø and Revsvegg units) represent rocks of the Caledonian depositional cycle and were affected uniquely by Caledonian metamorphism. The other nappes clearly represent pre-Caledonian basement and had variable histories of sedimentation, deformation and metamorphism in Precambrian times (Anderson *et al.*, 1974; Heier *et al.*, 1972; Andresen and Heier, 1975). In the Stavanger area, similar Precambrian nappe units have recently been reported with Rb/Sr whole rock isochron ages in the range 1534-1160 Ma (*see* Table 7.1). It is considered that these may represent the allochthonous equivalents of the two major units in the autochthonous basement, i.e. the Telemark Suite and the Pre-Telemark gneiss complex. If this be the case, it is apparent that the Telemark Suite of the Dyrskard unit has undergone higher grade metamorphism during Precambrian times than the equivalent rocks of the substrate to the nappe pile. The Holmasjø unit apparently has a lower metamorphic grade than the uppermost Revsvegg unit, implying an original (pre-thrusting) increase in metamorphic grade. This pattern is also indicated by the apparent increase in the metamorphic grade of Caledonian overprint metamorphism in the three thrust nappes of Precambrian rocks. It is suggested that the Revsvegg unit may represent an original Caledonian cover sequence to the underlying crystallines of the Kvitenut Complex, though movements have apparently occurred at the contact. Post-thrusting open folding and faulting of the thrust planes is clearly demonstrable in the area.

The nappe sequence of the eastern and central Jotun complex

Three major nappe units are recognized in this area (Table 7.2), the highest of which is the Upper Jotun Nappe comprising Precambrian crystallines below which is, first, the Valdres Nappe, then the Quartz Sandstone Nappe. The lower nappes consist almost entirely of low-grade Eo-Cambrian to Middle Ordovician metasediments. The underlying Precambrian basement is essentially composed of the rocks of the Telemark Formation and pre-Telemark crystallines. Lower Palaeozoic autochthon is only present as a thin skin over the basement and below the nappe complex (Strand, 1972).

The lowest of the nappes is the Quartz-Sandstone Nappe which contains sediments ranging from Late Precambrian to Middle Ordovician, (Oslo 4a) (Strand, 1972). The rocks of this nappe show strong internal folding; the main features are summarized by Bjørlykke (Article 8). Above this occurs the Lower Jotun or Valdres Nappe (Kulling, 1961) which is a unit containing

IGCP Project 27, Norwegian Contribution No. 5E.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Table 7.1

Summary of Nappe Sequence - Hardangervidda Ryfylke, Southern Norway.

	TECTONOSTRATIGRAPHIC UNIT	MAIN LITHOLOGIES	FOSSIL EVIDENCE	IGNEOUS ROCKS	CHARACTER OF DEFORMATION AND METAMORPHIC GRADE	RADIOMETRIC AGES
ALLOCHTHONOUS UNITS	**REVSVEGG UNIT** **(Formation)**	Essentially micaceous gneisses of unknown age.	None	Minor amphibolites and granodiorites	Polyphasal Caledonian deformation Caledonian amphibolite facies metamorphism with greenschist facies retrogression.	(6) 430 ± 250 m.y.
	PRIMARY STRATIGRAPHIC UNCONFORMITY OR THRUST CONTACT (Mylonites described at interface)					
	KVITENUT COMPLEX	Gneisses and migmatites of granitic and dioritic composition, augen gneisses, amphibolites, and thin sporadic calc-silicate bands.	None	Granites, diorites, monzonites. Minor ultrabasics.	Polyphasal deformation both during Precambrian and Caledonian. Metamorphism: relics of older Pre-cambrian granulite facies within upper amphibolite facies complex of probable Precambrian age. Upper greenschist - lower amphibolite facies in Caledonian	(6) 1643 ± 88 m.y. (9) 1639 ± 100 m.y. (8) 1534 ± 125 m.a. x (8) 1243 ± 160 m.y. x (Intrusive Granite) x From Stavanger Area but correlated with this complex
	THRUST CONTACT WITH MYLONITE DEVELOPMENT					
	DYRSKARD UNIT **(Group)**	Upper: Quartzites, meta-arkoses, quartz-rich black schists. Middle: Interbanded quartzites and amphibolites. Lower: Banded supracrustal gneisses, quartzites and quartzofeld-spathic gneisses. (The rocks of this unit are correlated with the Telemark Suite)	None	The amphibolites in the sequence are regarded as volcanics.	Polyphasal deformation both during Precambrian and Caledonian. Amphibolite facies metamorphism of Precambrian age. Caledonian overprint metamorphism of greenschist facies.	(6) 1289 ± 80 m.y. (7) 1160 ± 24 m.y. (from Stavanger area but correlated with Dyrskard Unit).
	THRUST CONTACT WITH MYLONITE DEVELOPMENT					
	NUPSFONN COMPLEX	Rock assemblage resembles that of Kvitenut Complex. Granitic and granodioritic gneisses which are variably migmatitic. Lower part comprises quartzites, quartz-schists, amphibolites and minor develop-ments of calcareous rocks. (Question as to true status of this lower unit).	None	Granites, granodiorites and quartz-diorites. Amphibolites with minor bodies of metagabbro.	Polyphasal deformation in both Precambrian (?) and Caledonian. Amphibolite facies metamorphism in Precambrian (?), and/or Caledonian. Late Caledonian greenschist facies overprint metamorphism.	No information available.
	THRUST CONTACT WITH CATACLASIS					
	HOLMASJØ UNIT **(Formation)**	Strongly deformed quartz-schists with layers of feldspathic quartz-sandstone and quartzite.	None	None recorded	Caledonian middle greenschist facies retrograde metamorphism. Relics of garnet and epidote clouded albite indicate former grade conditions.	(6) 409 ± 15 (taken to indicate age of Caledonian metamorphism).
THRUST CONTACT WITH CATACLASIS						
AUTOCHTHONOUS UNITS	**CAMBRO-ORDOVICIAN**	Stratigraphic range Cambrian up to Llanvirn/Llandeilo. Thin sequence variably preserved beneath the over-lying thrust plane, and is sometimes cut-out. Comprises conglomerates, sandstones, siltstones, mudstones, black shales and limestones	Poorly preserved fauna includes: - Dictyonema flabbeliforme, Obulus sp., Ptychopyge sp., Orthis ss., Ormoceras (?) sp., Torellella (?) sp. An early middle Cambrian fauna containing Paradoxides oelandicus has been recorded from Jøsenfjord.	None recorded	Maximum prograde metamorphism in lowest greenschist facies - Late Caledonian	No information available
	PRIMARY STRATIGRAPHIC UNCONFORMITY					
	PRECAMBRIAN BASEMENT	Rocks of Telemark suite: Metasediments (quartzites, quartz schists, conglomerates, minor calcareous schists and limestones.	None	Volcanics: - acid lavas, ignimbrites, agglomerates & tuffs. Basaltic lavas and tuffs. Plutonics: - granites, granodiorites, monzodiorites and metadolerites.	At least one major Precambrian deformation phase (though probably polyphasal). Greenschist facies metamorphism of Precambrian age.	(6) 943 ± 100 m.y. (assumed to be a metamorphic age)
		PRIMARY STRATIGRAPHIC UNCONFORMITY (ASSUMED)				
		Granitic and granodioritic gneisses with basic enclaves.	None	Not recorded	Complex Precambrian deformation High amphibolite facies Precambrian metamorphism	No information available.

along part of its lower contact Precambrian crystallines resembling those of the Upper Jotun Nappe (Strand, 1972; Nickelsen, 1967, 1974; Loeschke and Nickelsen, 1968; Hossack, 1972). These crystallines are succeeded, in part with preserved stratigraphic unconformity, by a thick sequence of the Eo-Cambrian sparagmite sequence of Valdres. The sparagmites in turn pass up through a Cambrian sequence and reach their highest stratigraphic level in the upper part of the Mellsenn Group which has a Middle Ordovician graptolite-shelly fauna (Oslo 4a). Strong internal folding is shown in the rocks of this nappe with major overturns. The folding was apparently developed prior to the "mise-en-place" of the Valdres Nappe. The metamorphic grade is low-greenschist facies. Rocks of the Valdres Nappe continue westward and have a marked development at the Tyin window beneath the Jotun Nappe.

Table 7.2

Nappe Sequence Eastern Jotunheimen, Southern Norway.

	TECTONOSTRATIGRAPHIC UNIT	MAIN LITHOLOGIES	FOSSIL EVIDENCE	IGNEOUS ROCKS	CHARACTER OF DEFORMATION & METAMORPHIC GRADE	RADIOMETRIC AGES
ALLOCHTHONOUS UNITS	**ØLJUVATN UNIT (Formation)**	Sandstone polymict conglomerate	None		Polyphasal deformation greenschist facies metamorphism (probably Caledonian)	None
		PROBABLY PRIMARY UNCONFORMITY				
	UPPER JOTUN NAPPE	Anorthosites. Peridotites. Various pyroxene-granulites and gneisses in the core. In peripheral areas gabbros and granitic rocks.	None	Anorthosites, peridotites, gabbros, syenites, granites	Precambrian granulite facies, corona forming in anorthosites. Amphibolite facies (probably precamb.) Strongly developed cataclastic rocks during Caledonian thrusting, with greenschist facies metamorphic assemblages produced.	K/Ar mineral ages in the range 428 – 1280 m.y.
		THRUST CONTACT: WITH MAJOR DEVELOPMENT OF BLASTOMYLONITES				
	LOWER JOTUN NAPPE: **VALDRES OG MELLSENN GROUPS**	Shales, sandstones, tillites (correlated with the thick Moelv tillite) Coarse conglomerates.	4 a	None	Polyphasal deformation with strong pebble elongation (Bygdin) cleavage formation. Recumbent folds in the Mellene area. Caledonian low greenschist facies met.	None
		THRUST CONTACT, IN PART SLIGHTLY DISTURBED PRIMARY UNCONFORMITY				
	LOWER JOTUN NAPPE: **ESPEDAL RØSSJØKOLLAN POSSIBLY GRØNNESEN-KNIPA**	Varied gneisses and other rocks resembling those of the Upper Jotun Nappe.	None	Granite, metagabbro, anorthosite	Amphibolite facies of probable Precambrian age. Relict Precambrian granulite facies suggest by the presence of jotunperthites Caledonian cataclastic with accompanying greenschist facies metamorphism.	None
		THRUST CONTACT				
	QUARTZ-SANDSTONE NAPPE	Phyllites { Limestones locally, schist, shales and sandstones passing up into "Quartz sandstone" (Fine grained sandstone, coarse arkoses and quartzites) (eocambrian)	3 a-b 2c 2a 1c	None	Slaty cleavage and folding. Caledonian low greenschist metamorphism	None
		THRUST PLANE				
AUTOCHTHON	**CAMBRIAN**	Thickness < 100m. Alunschist. Dark sandy shale Thin basal conglomerate locally	Middle Cambrium (1c)[1] in Arudal (Strand (1954)).	None	Very low grade Caledonian metamorphism. Locally tectonic thickening and imbrication.	None
		PRIMARY STRATIGRAPHIC UNCONFORMITY				
	PRECAMBRIAN BASEMENT	Telemark supracrustals (metadacite, metabasalt and quartzites. Various gneisses (both para- and orthogneisses).	None	Granites, gabbros, dolerites	Polyphasal deformation. Mainly Precambrian amphibolite facies, also Precambrian greenschist-facies in Telemark supracrustals. Precambrian granulite facies in Arendal area.	Mainly Sveconorwegian dates 900-1200 m.y. but some older (Svecofennian) 1616 ± 38 m.y. 1700 ± 100 m.y.

Note[1] (Numbers 1 c etc. refer to standard stratigraphic subdivision of the Lower Palaeozoic sequence in the Oslo area as summarized by Henningsmoen , 1960 (p. 130 – 131, Pl. 7). Geology of Norway, Norges Geol. Unders. 208, ed. Olaf Holtedahl.

The Upper Jotun Nappe is the highest structural unit in the area, and occupies the central part of a major northeasterly trending depression (Faltungsgraben). The nappe contains a variety of rock types of the Bergen-Jotun kindred. In the interior parts granulite facies assemblages dominate (anorthosites, mangerites, jotunites and pyroxene granulites) (Griffin, 1971; Battey and McRitchie, 1975). The peripheral areas, however, are mainly amphibolite facies rocks with essentially gabbroic and granitic rocks which may retain relics of primary igneous textures. The Upper Jotun Nappe evidently includes several tectono-stratographic units but at present insufficient work has been done to provide details. The thrust-planes are always marked by extensive zones of cataclastic rocks and near such zones Caledonian greenschist facies overprint metamorphism

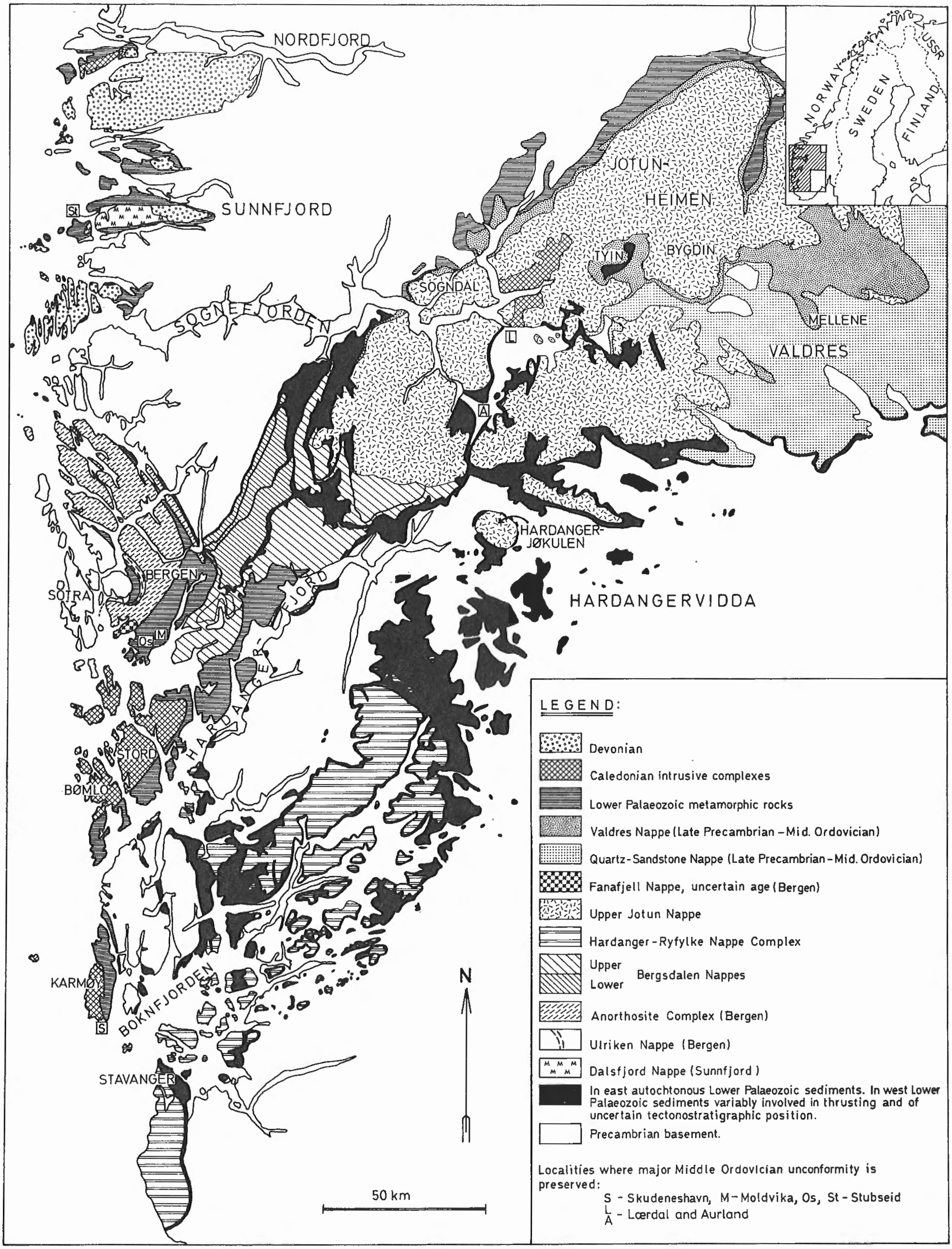

Figure 7.1. Principal tectono-stratigraphic units of the southern Norwegian Caledonides.

Table 7.3

Bergen Nappe Sequence, Southern Norway.

	TECTONOSTRATIGRAPHIC UNITS	MAIN LITHOLOGIES	FOSSIL EVIDENCE	IGNEOUS ROCKS	CHARACTER OF DEFORMATION AND GRADE OF METAMORPHISM	RADIOMETRIC AGES
ALLOCHTHONOUS UNITS	**FANAFJELL NAPPE**	Ortho- and paragneisses: Granites, micaschists, quartzites, greenschists, minor marbles. Age uncertain.	None	Granites, Porphyries, Metadolerites.	Polyphasal deformation probably both Precambrian and Caledonian. Amphibolite facies metamorphism (probably Precambrian), Caledonian greenschist facies metamorphism.	453 ± 50 m.y. (variably sheared granites).
	THRUST CONTACT CATACLASIS AND VARIABLE BLASTOMYLONITES DEVELOPMENT THRUST PLANE FOLDED AND CUTS ALREADY FOLDED THRUST PLANE IN SUBSTRATE.					
	ANORTHOSITE COMPLEX	Anorthosites, gabbros, mangerites. Ortho- and paragneisses with quartzites. Possibility that these rocks represent several tectono-stratigraphic units, but this awaits confirmation.	None	Anorthosites, Gabbros, minor ultramafics Mangerites, Metadolerites, Trondhjemites.	Precambrian polyphasal deformation. Caledonian polyphasal deformation. Precambrian polyphasal granulite facies metamorphism - local eclogite development. Amphibolite facies metamorphism of Precambrian/Caledonian age. Caledonian greenschist facies overprint metamorphism.	1064 ± 24 m.y. (Mangerites) 1775 m.y. (Approx. isochron for granulite facies paragneisses, partially downgraded to amphibolite facies).
	THRUST CONTACT - EXTENSIVE BLASTOMYLONITE DEVELOPMENT COMPLEX REFOLDING OF THRUST PLANE					
	ULRIKEN GNEISS NAPPE	Quartzites, conglomerates, micaschists. Assumed Lr. Palaeozoic age.	None	None recorded.	Polyphasal Caledonian deformation. Caledonian middle greenschist facies metamorphism.	None available.
		IN PART THRUST CONTACT - IN PART PRIMARY STRATIGRAPHIC UNCONFORMITY				
		Migmatitic para- and orthogneisses. Intrusive granitic complexes. The Ulrikens Nappe is multiple unit containing at least three thrust sheets, which in themselves exhibit a variety of basement/cover relations involving the above units.	None	Granites, Pegmatites, Metadolerites.	Polyphasal Precambrian deformation. Polyphasal Caledonian reworking. Precambrian high amphibolite facies polyphasal metamorphism with extensive anatexis. Caledonian greenschist facies overprint metamorphism.	1440 ± 100 m.y. (age of late anatexis).
	THRUST CONTACT - EXTENSIVE BLASTOMYLONITE DEVELOPMENT COMPLEX REFOLDING OF THRUST PLANE					
	MAJOR AND MINOR BERGEN SCHIST ARCS — HOLDHUS GROUP	Conglomerates, phyllites, mica-schists, limestones, metagreywacks	Llandoverian: graptolite/brachiopod fauna. Ashgill: coral/shelly fauna	Lamprophyres, Trondhjemites.	Polyphasal Caledonian (late) deformation. Caledonian metamorphism of middle to upper Greenschist facies.	None available
		HIGHLY DEFORMED PRIMARY STRATIGRAPHIC UNCONFORMITY				
	SAMNANGER COMPLEX	Mica schists, quartz schists, conglomerates, marbles, greenschists, amphibolites. The relations between the two complexes are often obscured by the pervasive late Caledonian polyphase deformation and metamorphism. Within these sequences occur slices of variably Caledonized basement gneisses.	None	Quartz diorites, Minor granodiorites. Gabbros, Serpentinites, Peridotites. Extensive basic lavas, in part pillowed.	Polyphasal early Caledonian deformation. Polyphasal late Caledonian reworking. Early Caledonian upper greenschist-lower amphibolite facies metamorphism. Late Caledonian greenschist facies overprint metamorphism - often pervasive.	None available
	THRUST CONTACT - EXTENSIVE BLASTOMYLONITE AND PHYLLONITE DEVELOPMENT COMPLEX REFOLDING OF THRUST PLANE					
? AUTOCHTHON ?	**ØYGARDENS GNEISS COMPLEX**	Para- and Orthogneisses with amphibolites It is not possible to say whether this is true basement or if thrust planes occur beneath sea-level.	None	Granites, Gabbros, Metadolerites, Amphibolites, Diorites, Quartz Syenites, Pegmatites	Polyphasal Precambrian deformation. Polyphasal Caledonian reworking. Precambrian high amphibolite facies, metamorphism with relicts of older granulite facies metamorphism. Caledonian greenschist facies overprint metamorphism, may range up into amphibolite facies.	473 ± 31 m.y., 890 ± 150 m.y., 800 ± 14 m.y. — Younger igneous intrusions 1024 ± 85 m.y., 1042 ± 92 m.y. — Presumed Grenville reworking 1750 ± 60 m.y. — Older gneisses

dominates. An important recent discovery is the presence of a supracrustal cover sequence to the Upper Jotun Nappe (Askvik, 1976). This sequence - the Øljuvatn Formation - rests with assumed stratigraphic unconformity on mangerites of the Nappe, and has obviously been transported as part of this structure. These supracrustals show only greenschist facies metamorphism, which would indicate that the main metamorphism of the Upper Jotun Nappe was probably Precambrian. Little geochronological data is available from the Nappe and only K-Ar mineral ages in the range 426-1280 Ma come from the main outcrop (Battey and McRitchie, 1975).

From the Kaupanger area a Rb-Sr whole rock age of 448 ± 30 Ma is reported for trondhjemites intruding partly retrograded anorthosites (Berthomier *et al.*, 1972). The Hardangerjøkul massif, regarded as a klippe of the Jotun Nappe, has yielded an Rb-Sr whole rock isochron of 1639 ± 89 Ma (Priem, 1968).

The status of the Upper Jotun Nappe has long been a controversial subject in Norwegian geology and explanations vary from hypotheses that it is a far-travelled thrust nappe (Strand, 1972), that it is a fold nappe, or that it has a local root-zone and is upthrust from below (Smithson *et al.*, 1974). The latter hypothesis has been considered to have support in the pattern of regional gravity anomalies. However, rocks of the basement windows of Laerdal and Aurland show little signs of Caledonian reworking and cover sequences are preserved. In all places at the contact of the Upper Jotun Nappe, marked developments of mylonitic rocks are found and in the section from Valdres to Sogndal variable developments of the Valdres Nappe intervene between the Upper Jotun Nappe and the autochthon (Strand, 1972). The strong stretching fabric with general NW-SE trend show little variation around the central zone, indicating a northwesterly source for the whole mass. The authors thus prefer the hypothesis of a distant nappe transport and consider that the provenance of the Upper Jotun Nappe must have lain at some considerable distance westwards of the present Norwegian coast. When this great nappe is seen together with those of the Hardanger-Ryfylke area, which occupy a similar tectonic position, one cannot but marvel at the enormous extent of old Precambrian crust which has been thrust into place from a distant northwestward source.

The nappe sequence of the western Jotun complex - Bergsdalen

This complex shows surprising differences to the patterns observed in the eastern section of the Jotun region. The thick development of the Valdres (Lower Jotun) Nappe and its underlying Quartz Sandstone Nappe is but little represented in the western region. The rocks of the Bergsdalen Nappe underlie the Upper Jotun Nappe in the southern part of this area, and have a somewhat enigmatic position (Kvale, 1960). They have been subdivided essentially as the Upper and Lower nappes separated by a thick development of Lower Palaeozoic phyllites. The Bergsdalen nappes are separated from the basement gneisses of the North-West gneiss region by similar phyllites. To the west they are overridden by the Bergen Nappe sequence, though the contact is complicated by subsequent folding. To the southwest the Bergsdalen complex borders against the structurally underlying Caledonian schist sequence of Norheimsund. The Bergsdalen nappes contain an assemblage of Precambrian genisses and supracrustals. The latter includes a sequence of quartzites, conglomerates and metavolcanics which have been correlated with the Telemark Suite (Kvale, 1960). This assemblage is intruded by a series of plutons of varied composition which were thought to be of Caledonian age. Recent Rb-Sr isochron studies, however, clearly demonstrate their Precambrian age: Hernes (1274 ± 48 Ma, Pringle *et al.*, 1975); Hodnaberg (1004 ± 100 Ma, Brueckner, 1972); Berge (950 Ma, Pringle, pers. comm.); and Fosse (969 ± 17 Ma, J. Gray, pers. comm.). Recent studies by the authors indicate that certain of the rocks of the Bergsdalen previously identified as metavolcanics are in fact retrograded and variably mylonitic gneisses, though the extent of these rocks is not yet known. This reworking is, in part, of Caledonian development. In the Upper Bergsdalen Nappe a thick sequence of low metamorphic grade supracrustals occurs, and whether these are Precambrian or Lower Palaeozoic is not clear. In the southern part of this nappe are localized developments of rocks of the anorthosite kindred (Kvale, 1960) though there is no precise information concerning their tectono-stratigraphic position. The metamorphism of the Bergsdalen nappes is generally regarded as being of Caledonian age and ranges high up in the amphibolite facies (Kvale, 1960). We consider, however, that a constraint is put on the extent of Caledonian metamorphism by the maximum grade attained by the Lower Palaeozoic sediments. These latter never reach higher than greenschist facies, implying that the amphibolite facies metamorphism in the rocks of Precambrian age is also of pre-Caledonian development. There are also abundant indications of greenschist facies overprint in these rocks which probably represents the effects of Caledonian metamorphism related to thrusting.

To the north of the Bergsdalen Complex, the phyllite units continue across Sognefjord. Beneath the Upper Jotun Nappe in the Sogndal region a nappe of Caledonian supracrustal rocks rests with thrust contact above the basement (Skerlie, 1957; Lacour, 1969). The rocks of this nappe closely resemble the older sequence of the Bergen schist arcs both in terms of lithology and of metamorphic grade. We consider that this unit probably represent a klippe of Major Bergen Schist Arc. Rocks of Valdres type have been identified in a thrust nappe between this and the Upper Jotun Nappe. If this is correct it implies a very substantial lateral translation of the Valdres Nappe, i.e. from a position west of the provenance for the Major Bergen Arc Unit.

The Bergen Nappe Complex

This is the classical area of the Bergen Arcs (Kolderup and Kolderup, 1940), and involves a sequence of nappes of both supracrustals and gneisses thrust over a gneissic basement (Table 7.3). At the eastern boundary this complex transgresses the Bergsdalen Nappe Complex and the basement gneisses of the northwest gneiss region. The western boundary is with the gneisses of the Øygarden Gneiss Complex which, although it forms on Sotra the local basement of the Bergen nappes, may possibly represent an allochthonous unit though no thrust planes are seen. The characteristics of the Bergsdalen Nappe Complex are indicated in the preceding section. The northwest gneiss area contains a great variety of para- and ortho-gneisses together with supracrustals of assumed Precambrian age. Rb-Sr isochron ages are generally Svecofenninan (1600-1800 Ma) (Brueckner, 1972; Sturt *et al.*, 1975), though Grenville ages (950-1200 Ma) are also recorded (Bryhni *et al.*, 1971). The western gneisses of Øygarden show an older complex of Svecofennian age with the evidence of subsequent Grenville reworking (Sturt *et al.*, 1975). The Caledonian rocks of the schist arcs (mapped as Lower Palaeozoic metamorphic rocks on Figure 7.1) fall into two sequences separated by a major stratigraphic unconformity above which occurs a major development of polymict conglomerate (Moberg Conglomerate) (Sturt and Thon, 1976). The older or Samnanger complex (Table 7.3) had a complex orogenic development in early Caledonian times which involved polyphasal deformation, metamorphism and igneous intrusion (Foerseth *et al.*, 1977). The metamorphism ranges into the amphibolite facies, though the pattern is complicated by late Caledonian reworking and overprint metamorphism. The sequence is considered to be of probable Cambrian-early Ordovician age as the result of regional correlations, though no fauna has yet been recorded from the rocks. Conspicuous in this sequence is a thick development of submarine lavas which have many of the geochemical characteristics of ocean-floor basalts (Inderhang, 1975).

The younger sequence (Holdhus Group (Foerseth *et al.*, 1977) commences with the Moberg conglomerate which passes upwards into Ashgillian (Oslo-5a) limestones with a coral-shelly fauna (Kolderup and Kolderup, 1940). Higher in the sequence occur Lower Llandovery rocks which contain a monograptid assemblage. These rocks have been involved in polyphasal deformation and metamorphism ranging from low to high greenschist facies. It would appear that the first phase of major folding pre-dates the "mise-en-place" of the thrust sheets. A number of highly mylonitic sheets of basement gneisses have been imbricated up into the sequence, which in fact can be subdivided into a series of minor tectonic units. Near thrust planes very pronounced zones of mylonitic and phyllonitic rocks are present. Structurally above occur the rocks of the Ulriken Gneiss Nappe (Sturt *et al.*, 1975). This is a complex unit comprising a series of thin thrust sheets of gneiss, each capped by a cover sequence of presumed Lower Palaeozoic metasediments. In some instances undisturbed primary stratigraphic unconformities can be seen at the base of the sediments, though in zones of high strain the rocks have a mylonitic aspect. The gneisses are both of para- and of ortho-type and have had a complex pattern of tectono-metamorphic evolution involving multiple deformation and several periods of migmatitization. The latest anatexis has been dated at 1440 ± 100 Ma (Sturt *et al.*, 1975). The cover sequence is metamorphosed only to middle greenschist facies, and is considered to represent a former western margin to a lower Palaeozoic basin.

The Anorthosite Complex sits with disjunctive thrust contact above the Ulriken Gneiss Nappe and comprises rocks of Precambrian age (Sturt *et al.*, 1975). The assemblage in this complex is varied and contains a range of para- and orthogneisses together with plutonic rocks of the anorthosite-mangerite kindred. Recent investigations have shown that the anorthosites have a protracted history of deformation and metamorphic recrystallization in the granulite facies (Griffin, 1972) prior to the intrusion of the rocks of the mangerite suite (Austrheim, pers. comm.). These latter have also been subjected to granulite facies metamorphism (Griffin, 1972), and this implies that the anorthosites with their considerable pre-mangerite emplacement history may well represent a more ancient granulite facies complex. An Rb-Sr isochron age is available for the mangerites at 1064 ± 24 Ma and the initial Sr^{87}/Sr^{86} ratio of 0.7030 implies that this must be close to the age of intrusion (Sturt *et al.*, 1975). The rocks of the Anorthosite Complex show effects of reworking under both amphibolite and greenschist facies which to a large extent is of Caledonian age. The paragneisses of this complex record this particularly well and only relics of the earlier granulite facies are seen in a general amphibolite facies surround. Partially downgraded granulites have yielded an approximate isochron at 1775 Ma (Sturt *et al.*, 1975). The thrust planes beneath the nappe units described above have been polyphasally refolded which produces complex geometries.

Work in progress by the authors shows that the highest unit in the Bergen Nappe sequence is the Fanafjell nappe (Table 7.3). This nappe transgresses the already folded thrust plane between the Anorthosite Complex and the Major Bergen Arc. The thrust plane at the base of the Fanafjell nappe is also strongly folded by westward facing folds attesting to the complex pattern of fold and thrust tectonics in this western section of the southern Caledonian segment.

To the south of the Bergen region in the island complex towards Stavanger occurs a succession of Lower Palaeozoic rocks both of older and younger Caledonian development in allochthonous position. Their exact relations to the Bergen Nappes and Hardanger - Ryfylke Nappes are not yet clear. The major unconformity observed in the Major Bergen Arc is also again seen at Karmøy. There a thick development of polymict-conglomerates capped by calcareous sandstones, with an Ashgill-Llandoverian fauna (Isachsen, 1940), rests with profound stratigraphic unconformity on a substrate of polyphasally deformed and metamorphosed (to upper greenschist facies) early Caledonian rocks. These latter include a variety of mica-schists, greenstones, metagabbro and serpentinites which had undergone considerable uplift and erosion prior to the deposition of the conglomerate.

In Stord an assemblage of Upper Ordovician and Lower Silurian rocks has been identified, though the base of the sequence is obscured by faulting (Foerseth *et al.*, 1975). The northern part of Stord and much of Bømlo contains a considerable development of Caledonian volcanic rocks which are to a large extent of sub-aerial eruption and show certain geochemical affinities with those of continental rift zones (Furnes and Foerseth, 1975). The age of these rocks is uncertain, but an Rb-Sr isochron of 455 ± 5 Ma for rhyolites indicates a Llandeilo/Caradoc age (Priem and Torske, 1973). These rocks have a quite complex structural history with variable, though low metamorphic grade. The exact position in the sequence of Caledonian development has not yet been ascertained.

The Caledonian rocks of the coastal area north of Sognefjord

Again this area involves both older and younger Caledonian complexes, separated by the polymict Middle Ordovician Stubseid conglomerate (Skjerlie, 1969). Polyphasal deformation and metamorphism apparently occurred prior to the deposition of this conglomerate and again in the late Silurian. The metamorphism does not appear to have exceeded greenschist facies at either stage. The relationships of the lower structural part of this complex to the gneissic basement have not yet been clearly defined. In the Sunnfjord area a sequence of nappes occurs containing elements of both Caledonian supracrustals and Precambrian crystallines (mainly mangerites). Above these occurs the Devonian Old Red Sandstone clastic sequences which occupy an essentially allochthonous position. The pre-Stubseid Caledonian succession contains a major development of submarine basaltic lavas which have many of the geochemical characteristics of ocean-floor basalts (Furnes *et al.*, 1976). Volcanic rocks are comparatively rare in the younger succession.

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PROJECT 27 CALEDONIDE OROGEN

THE EASTERN MARGINAL ZONE OF THE CALEDONIDE OROGEN IN NORWAY

Knut Bjørlykke

Geologisk Institutt
Avd. A
Universitetet i Bergen
5014 Bergen, Norway

INTRODUCTION

Along the parts of the eastern margin of the Caledonian orogenic belt that lie in Norway a late Precambrian or Lower Palaeozoic sedimentary cover overlies a crystalline Precambrian basement. The part that lies in Sweden is described by Gee (Article 10). The stratigraphic relationships in these sediments show that the late Precambrian sedimentary rocks are restricted to well-defined cratonic basins and that a series of transgressions finally led to an almost total submergence of the Baltic Shield in Middle Cambrian time.

Thick sequences of late Precambrian sedimentary rocks of low metamorphic grade are found in two main areas (Table 8.1):

1. South Norway (Hedmark and Oppland counties)
2. North Norway (Finnmark county).

SOUTH NORWAY (HEDMARK AND OPPLAND)

In South Norway late Precambrian sediments (Hedmark Group) were deposited in N- to NNW-trending grabens that were 30-60 km wide (Fig. 8.1) (Bjørlykke *et al.*, 1976). These sediments are in Scandinavia traditionally referred to as sparagmites and they occur in the area north of Lake Mjøsa (Gudbrandsdalen) and in Østerdalen. Similar sediments are also found in the adjoining parts of Sweden (Gee *et al.*, 1974). The base of this sedimentary sequence is nowhere exposed, but estimates of sedimentary sequences made by Åm (1975) suggest a 3-4 km thickness in the deeper parts of the basin.

During the deposition of the Brøttum Formation (Table 8.1) the initial subsidence led to deep water conditions in large parts of the basin and deposition of turbidites (Englund, 1966, 1972). Both the Biskopås Conglomerate and the Ring Formation can be mapped out

Table 8.1

Late Precambrian and Early Cambrian Stratigraphy of Norway.

S. Norway (Moelv - Ringsaker)	Thickness in metres	Formations	Lithology	Fossils	N. Norway (Tanafjord - Digermulen)	Thickness in metres	?	Formations	Lithology	Fossils
Middle Cambr.	30	Paradoxides	S.		Middle Cambr.	700	Digermul Group	Kistedal F.	Shale, sandst.	(Paradoxides)
	1	Strenuella	Limest.	(strenuella linnarsoni)		500		Duolbas-gaissa F.	quartzites	(Holmia sp.)
Lower Cambrian	30	Holmia	Shale	(Holmia kierulfi)	Lower Cambrian					
	1-5	Bråstad	Shale	(Volbertella)		600	Vestertana Group	Breivik F.	sandst. siltst.	(Platysole-nites)
	0-4	Bråstad	Sandst.	(Mobergella holsti)						
	0-2	Breensæther	Limest.	(Holmia et. mickwitzi)						
Precambrian, Hedmark Group	100-200	Vangsås F.	Sandst.	(Scolithus)	Precambrian	450		Stappogiedde F.	sandst., shale	
	20	Ekre	Shale			10-50		Mortensnes Tillite		
	5-20	Moelv	Tillite			2-400		Nyborg F.	sandst., shale, limest.	
	0-300	Ring F.	Arkose			5-50		Smalfjord Tillite		
	50-100	Biri F.	Shale, limest.							1-2° unconformity
	0-100	Biskopås	Conglome-rate	(Papillo-membrana)		1200		Tanafjord Group	dolomite, shale, sandst.	
	0-20	Biri	Limest. (often eroded)			580		Vadsø Group		
	1500	Brøttum	Shale, graywackes arkoses congl.			? 9000		Barents Sea Group		

K. BJØRLYKKE

IGCP Project 27, Norwegian Contribution No. 5F.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Table 8.2

Summary of stratigraphy and sedimentation in late Precambrian and Lower Palaeozoic rocks in the Sparagmite and Oslo Regions.

STRATIGRAPHIC UNIT	THICKNESS	LITHOLOGY	DEPOSITONAL ENVIRONMENT	BASIN DEVELOPMENT	GEOTECTONIC MOVEMENTS
LUDLOVIAN	500 M	SANDSTONE	DELTAIC	REGRESSION PROGRADATION	OROGENIC UPLIFT
SILURIAN (LLANDOVERIAN-WENLOCKIAN)	600 M	CARB/SHALE	EPICONT. SEA	TRANSGRESSION MINOR UNCONFORM.	EMERGENCE
ORDOVICIAN (ARENIGAN-AȘHGILLIAN)	550 M	LIMEST./ SHALE	EPICONT. SEA	TRANSGRESSION REGRESSION	
ALUM SHALE (M.CAMBR.-TREMADOCIAN	100 M	BLACK SHALE	STAGNANT EPICONT.SEA	TRANSGRESSION	CRATONIC SUBMERGENCE
HOLMIA SERIES (L. CAMBRIAN)	50 M	SHALE/ SANDST.	SHALLOW MARINE		
VANGSAS FORM.	4o M	QUARTZITE	SHALLOW MARINE	MINOR UNCONFORM. TRANSGRESSION	
VANGSÅS FORM.	200 M	SANDST.	FLUVIAL/BRAIDED RIVER	REGRESSION	
EKRE SHALE	20-50 M	SHALE/ SANDST.	PRODELTA/DELTA SLOPE	TRANSGRESSION	
MOELV TILLITE	1-20 M	DIAMICTITE	GLACIAL	REGRESSION	
RING FORMATION	0-200 M	ARKOSE/ CONGL.	FAN DELTA	REGRESSION PROGRADATION	RIFTING BLOCK
BIRI FORMATION	100 M	SHALE/ LIMEST.	PLATFORM/BASIN	TRANSGRESSION	FAULTING
BISKOPÅS CONGL.	0-200 M	BOULDER AND PEBBLE CONGL.	FAN DELTA	REGRESSION PROGRAD.	
BIRI, FORMATION	0-10	CARBONATE AND PHOS-PHATES	PLATFORM BASIN	TRANSGRESSION	
BRØTTUM FORMATION	2000 M	SANDST. CONGL. ARKOSES & GREY WACKE	TURBIDITE AND DELTA SEDIMENTATION	INITIAL BASIONAL SUBSIDENCE	BLOCK FAULTING

as clastic wedges (fan deltas) disappearing into a shale or finer sandstone facies (Bjørlykke *et al.*, 1976). Also carbonate sedimentation is found to be restricted to the marginal shelf facies of the basin. Carbonate sedimentation took place during two periods of transgression onto the Baltic Shield surrounding the basins, holding back clastic sediment supply. Evidence of the first transgressive episode is found as limestone beds below the Biskopås Conglomerate at Moelv and Biri and as clasts in the basal part of the conglomerate. Microfossils have been described from phosphorite pebbles in this conglomerate (Manum, 1967; Spjeldnaes, 1967).

The Moelv Tillite is a diamictite showing positive evidence of deposition by glacial processes including striated pebbles, drop-stones and ice-wedges (Bjørlykke, 1974a; Nystuen, 1976a, b). Two main glacial facies can be recognized in the Moelv Tillite:

1) massive diamictite morainal facies;
2) scattered clasts in a laminated matrix (dropstone facies).

The morainal facies can be shown to represent the shallower parts of the basin on top of the fan deltas where the ice was grounded. Dropstone facies were deposited in the deeper parts where the ice sheets were floating (Bjorlykke *et al.*, 1976).

The Moelv Tillite is conformably overlain by the Ekre Shale which passes upwards into the Vangsås Formation. The lower part of the Vangsås Formation (Vardal Member), which is a red feldspathic sandstone, was probably deposited by braided rivers. The upper part of the Vangsås, the Ringsaker Quartzite is an orthoquartzite member which contains low-angle cross-bedding and Scolithus burrows in the topmost part, and represents a shallow marine evironment. Lead mineralization (galena) in the Vangsås Formation has been the object of economic prospecting (Bjørlykke *et al.*, 1973).

The Lower Cambrian Holmia Series represents a transgressive fining-upwards sequence from marine sandstone, shales and limestone and is succeeded by the Middle Cambrian black bituminous alum shale. The black shales are succeeded by more calcareous sediments representing an upwards shallowing sequence to the reef and coral-algal facies of Upper Caradocian age. In the Mjøsa district the Upper Ordovician and lowermost Silurian is missing. A thin transgressive quartzite and carbonate-shale was deposited following the transgression in Middle Llandovery times.

In Wenlockian times the Caledonian orogenic uplift started to produce clastic deltaic sediments prograding southwards over marine sediments, reaching the Oslo-Asker district in Ludlovian time (Skjeseth, 1963; Bjørlykke *et al.*, 1976). In the Oslo-Asker district of the Oslo Region a continuous highly fossiliferous sedimentary sequence is found ranging in age from Middle Cambrian to the Silurian interrupted by only minor breaks (Table 8.2) (Strand and Henningsmoen, 1960).

Between Engerdalen near the Swedish border and Østerdalen are sparagmites which were deposited in an eastern basin more or less separated from the central basin by a horst structure, which was plunging to the north. The oldest recognized stratigraphic unit in the eastern basin is the Biri Formation and we find here evidence of locally developed depositional environments which differ in several respects from that of

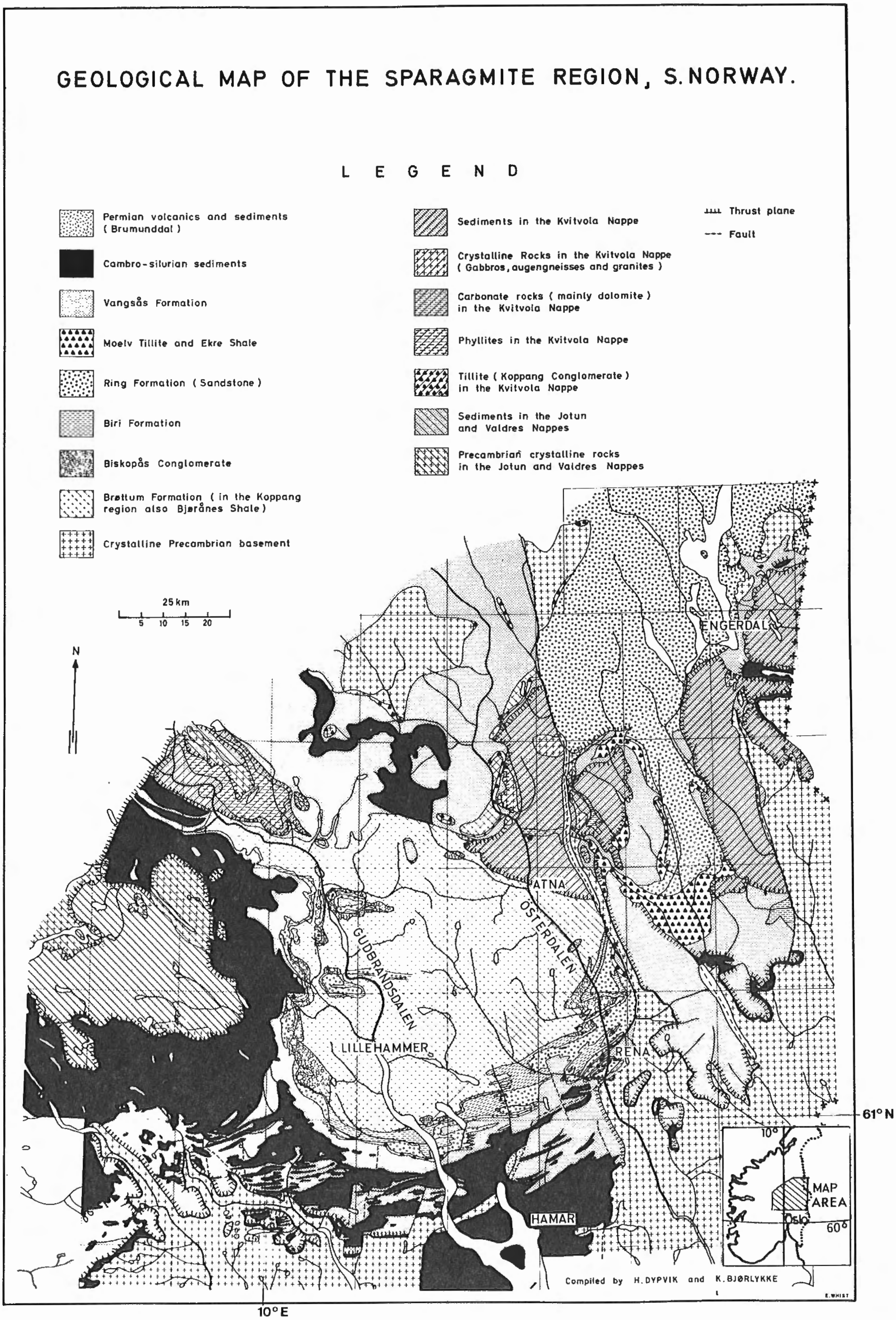


Figure 8.1. Geological map of the Sparagmite Region, South Norway, compiled by Dypvik and Bjørlykke. In the eastern part unpublished data from Nystuen is included.

the central basin (Nystuen, 1976a, b). It seems that the eastern side of the sparagmite basin had a more well-defined fault as well as the largest fault displacement, such that the basins may be regarded partly as half-grabens. There are several lines of evidence which indicate that syn-sedimentary faulting was accompanied by volcanicity. Alkali-basaltic rocks associated with sparagmite yield a preliminary Rb-Sr isochron with a latest Precambrian (Vendian) age (Holmsen and Oftedahl, 1956; Bjørlykke, 1969; Bjørlykke *et al.*, 1976; Raheim, pers. comm., 1976).

Sparagmites in the Valdres Nappe (Figure 8.1) to the west of the central basin were earlier thought to be of Ordovician age because they apparently overlie shales with a Middle Ordovician fauna (Strand, 1959), but Kulling (1961) and later Nickelsen (1967) and Loeschke (1967) showed that the section on which this conclusion was based (Mellene) was inverted. The latter authors worked out a late Precambrian stratigraphy and recognized a glacial conglomerate which could be correlated with the Moelv Tillite. The basal part of the sparagmite in the Valdres Nappe in the Dokkvann area contains frequent charnockitic rocks from the Jotun Nappe (Goldschmidt, 1916; Nickelsen, 1974). It is therefore clear that the Valdres sparagmite must be allochthonous, transported from approximately the same root zone as the Jotun Nappe (*see* also Sturt and Thon, Article 7).

The Kvitvola Nappe (Fig. 8.1) is found in the Østerdalen and Engerdalen areas and consists of impact quartzites, dolomites and pelitic rocks which have been subjected to a higher degree of regional metamorphism and deformation than the underlying sparagmites (Holmsen and Oftedahl, 1956, Bjørlykke, 1965). Basement rocks such as gabbros similar to those of the Jotun Nappe Complex, granitic rocks and augen-gneisses are also included in the Kvitvola Nappe. This suggests that it may be part of the Jotun Nappe Complex. Mainly because of its carbonate and diamictite horizon, the Kvitvola Nappe has been correlated with the late Precambrian sparagmites, but there are no definite age determinations to support this. Lithologically the Kvitvola rocks show closer resemblance to the late Precambrian sequence in Finnmark than to the sparagmites of south Norway. The extensive dolomites which contain magnesite (Nystuen, 1969), and the lateral continuity of facies suggest that the Kvitvola sequence was deposited along the western stable (passive) margin of the Baltic Shield (Bjørlykke *et al.*, 1976).

The late Precambrian and Lower Palaeozoic sediments have been subjected to a variable degree of Caledonian thrusting. The following tectonic units can be recognized.

1. Strictly autochthonous sedimentary rocks resting with primary contact on crystalline basement.

2. Parautochthonous rocks subjected to folding and smaller scale thrusting within the cratonic basin in which they were deposited. In the central sparagmite basin this unit consists of the sparagmite sequence from the Brøttum to the Ekre Formations.

3. Allochthonous rocks thrust above Cambrian shale. This nappe is called the Osen Nappe (Quartz Sandstone Nappe). It is the easternmost nappe which has been thrust over Lower Palaeozoic rocks. It can be seen on the map (Figure 8.1) where the Vangsas Formation is separated from Cambro-Silurian sediments by a thrust plane. The Osen Nappe consists mainly of sandstones of the Vangsas Formation which lie on the Ekre Shale. The nappe has been thrusted, with an imbricated structure, above a thin autochthonous sequence which includes the Alum Shale.

4. Lower Palaeozoic rocks (Middle Ordovician - Upper Silurian) thrust over underlying Alum Shale in the Oslo Region.

5. Strongly allochthonous and tectonically deformed rocks, usually with a well-developed lineation (Kvitvola Nappe), comprising quartzite, dolomite, gabbros, anorthosites and augen gneisses.

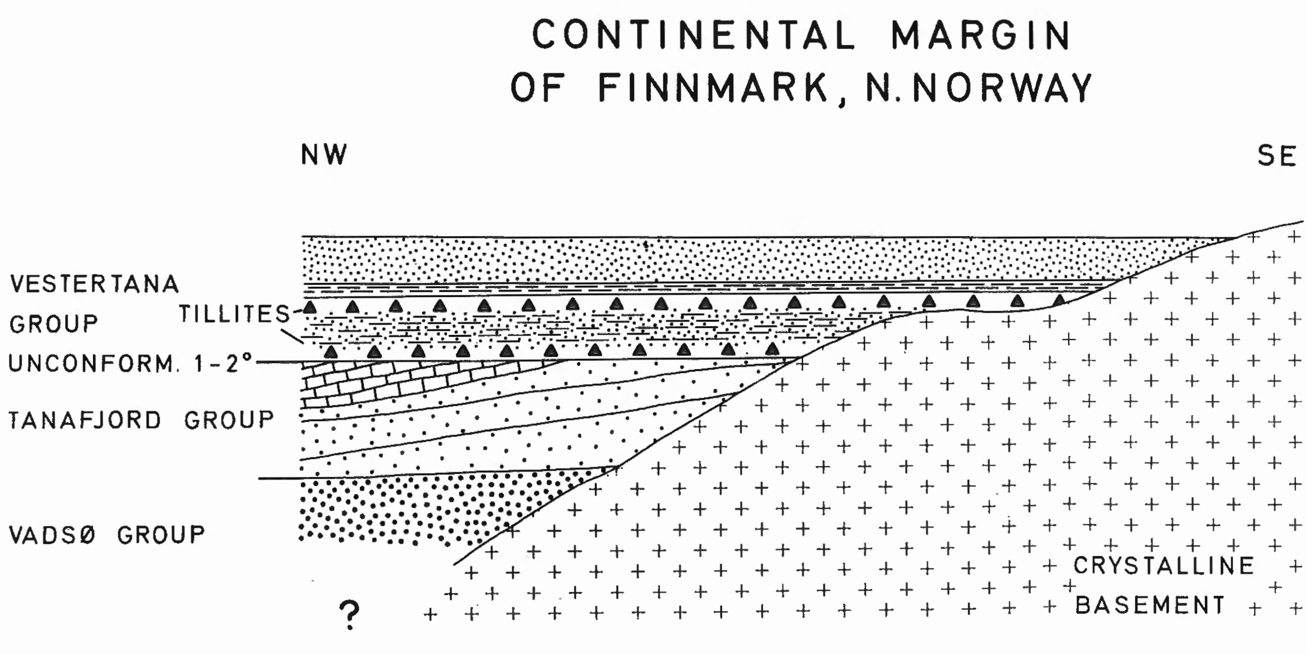
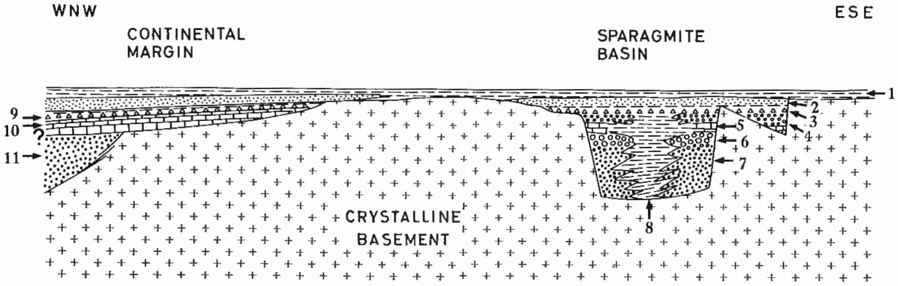


Figure 8.2. Diagrammatic representation of late Precambrian sediments lapping onto the crystalline basement in Finnmark, North Norway.



1 - Lower Cambrian shale
2 - Vangsås Formation (sandstone)
3 - Moelv Tillite
4 - Ring Formation
5 - Biri Formation (limestone)
6 - Biskopås Conglomerate
7 - Brøttum Formation (sandstone)
8 - Shale
9 - Tillite
10 - Carbonate (dolomite)
11 - Sandstone

Figure 8.3. Reconstructed cross section through the Baltic Shield in early Palaeozoic time showing a rift valley basin (Sparagmite Basin) and an onlapping continental margin sequence to the west corresponding to the Finnmark sequence and possibly also the Kvitvola sequence.

Table 8.3

Comparison between the Rift Valley basins of South Norway and the continental margin facies of North Norway

	Sparagmite Region South Norway	Finnmark North Norway
Dimensions of the basin	50-60 km width	Hundreds of km
Thickness of latest Precambrian sediment	3000-4000 m	10-20 km ?
Lithostratigraphic Units	Hedmark Group	Vestertana Group Tanafjord Group Vadsø Group
Lithology		
Arkoses	common	rare
Coarse conglomerates exclusive of glacial conglomerates	common	some
Glacial conglomerates	One well-defined unit (Moelv Tillite)	Two well-defined tillite (Smalfjord and Mortensnes Tillite)
Carbonate rocks	Marginal limestone facies passing into basinal mud facies	Extensive dolomites, supratidal facies
Facies variation	Rapid lateral facies variation from margin to central basin	Persistent lithology over long distances
Turbidites	present in the basal part	common
Alluvial fan deposits	widespread along the basin margin	not reported
Type of basin subsidence	contemporaneous faulting	subsiding continental margin without clear evidence of contemporaneous faulting
Volcanic activity	Lavas of probable latest Precambrian age	No evidence of volcanic activity
Plate tectonic situation	Rift Valley	Passive margin

NORTH NORWAY (FINNMARK)

In the central and eastern part of Finnmark late Precambrian rocks are found lapping on to the Precambrian crystalline basement (Fig. 8.2). To the north they are overlain by a sequence of nappes of successively higher metamorphic grade and tectonic deformation (*see* Sturt and Roberts, Article 4). The Riphean Barents Sea Group is found in the northern part of the Varanger Peninsula (Siedlecka and Siedlecki, 1972) bounded by the Komagelv-Trollfjord fault (Fig. 4.1). Some details are also contained in the Sturt and Roberts contribution. To the south of this major fault, around the Varanger Fjord, the Vadsø Group is found in direct contact with the Precambrian basement (Banks *et al.*, 1974). These sediments, which are mainly fluviatile, are overlain by the Tanafjord Group which contains a series of mostly marine sandstones and shales with supratidal dolomites in the uppermost part (Holtedahl, 1918; Føyn, 1937). The unconformably overlying Vestertana Group (Reading, 1965; Banks *et al.*, 1971) contains two tillite horizons separated by carbonates, shales and sandstones of the Nyborg Formation, which have been radiometrically dated at 668 ± 27 Ma (Pringle, 1973).

The glacial conglomerates represents two distinct climatic episodes. Around the inner part of the Varangerfjord at Biggjojavre they rest unconformably on a glacially striated pavement of the underlying sandstone of the Vadsø Group (Reusch, 1891; Reading and Walker, 1966; Bjørlykke, 1967). The upper part of the Vestertana Group is exposed at Digermulen (Reading, 1965) and equivalent rocks (Dividal Group or Hyolithus Zone) are found directly above the Precambrian basement to the south (Foyn, 1967). The Stappogiedde Formation of the Tanafjord section can be correlated with the Hyolithus Zone, through Sweden and corresponds to the Ekre Shale and Vangsas Formation of South Norway (Føyn, 1967; Bjørlykke *et al.*, 1967). The sediments overlying the Stappogiedde Formation (Breivik Formation and Duolbasgaissa Formation) contain Lower and Middle Cambrian trace fossils (Banks *et al.*, 1971; Banks, 1973) and body fossils (Henningsmoen, 1960).

Further west around the inner part of Laksefjord and Porsangerfjord occurs a sequence of stromatolitic dolomites, sandstones and tillites that can be correlated with the section of eastern Finnmark (Tanafjord) (White, 1969; Roberts, 1974). These sediments have been thrust to the south over the Dividal Group. Also as far west as Alta the same tillite horizons can be found in tectonic windows (Føyn, 1963).

SUMMARY

The late Precambrian sediments in North and South Norway show a number of distinctly different characteristics (Fig. 8.3, Table 8.3). While sedimentation in the sparagmite region in South Norway is characterized by rapid lateral facies variations and clastic wedges (fan deltas) the broad features of stratigraphy in Finnmark can be correlated over long distances (100-300 km) along the depositional strike.

Also the development of extensive supratidal dolomites in North Norway provides evidence of sedimentation on a stable craton subsiding at variable rates. The sparagmites have formed a response to block faulting in a series of rift valleys (Bjørlykke *et al.*, 1976), or an aulacogen (Roberts and Gale, 1977), probably associated with volcanicity. The block faulting died out in early Cambrian time as the Baltic Shield was being covered by an epicontinental sea.

In late Precambrian time the continental passive margin facies was represented by the Finnmark sequence and possibly also by the Kvitvola nappe.

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9. LATE-OROGENIC DEVONIAN BASIN FORMATION IN THE WESTERN NORWEGIAN CALEDONIDES

Ron J. Steel

Geologisk Institutt
Avd. A
Universitetet i Bergen
5014 Bergen, Norway

INTRODUCTION

Non-marine sediments known as the Old Red Sandstone, of Devonian and possibly also latest Silurian age, presently occur in two coastal districts of western Norway (Nordfjord-Sognefjord and Trondheim areas) and in a small area (Røragen) of eastern Norway. They occur in relatively small, fault-controlled basins. Distribution of these rocks in the Nordfjord Sognefjord area is shown on Figure 9.1 and also on Figure 7.1 (this volume, Article 7). The occurrences in the Trondheim area and at Røragen are shown on Figure 6.1 (this volume, Article 6). Details for the coastal areas are provided by Bryhni (1975), Nilsen (1973) and Steel (in press); for Røragen, by Roberts (1974). The basins presently vary in size from approximately 2000 km^2 to less than 12 km^2, in their type and style of sediment infill from dominantly laterally dispersed fanglomerates to dominantly longitudinally dispersed sandstones, in their age from Lower to Middle Devonian (Nilsen, 1973) and in the stratigraphic thickness of their sediment infill from less than 1 km to apparently 25 km.

Despite this considerable variation there are very many points of similarity between the basins. They are often elongate parallel to the structural trend in adjacent Caledonian or "caledonized" Precambrian basement, their sediment infill often shows a cyclic organization whose scale and style reflects varying syndepositional dip-slip or strike-slip-dominated faulting, and many of the basins show evidence of having been thrust eastwards after infilling (Bryhni, 1975; Steel, in press). Because of these similarities only one of these basins, the Hornelen Basin, the most intensely studied one, is described in some detail here. In addition, the Hornelen Basin is briefly contrasted with the Solund Basin (Fig. 9.1) to emphasize the varying tectonic context of Devonian basin formation within relatively short distances.

HORNELEN BASIN

Hornelen Basin (Fig. 9.1) is largely filled with sandstones, 25 km in stratigraphic thickness, which are spectacularly organized into basin-wide, coarsening-upwards cycles of the order of 100-200 m in thickness (Figs. 9.3, 9.4) (Brynhi, 1964; Steel *et al.*, 1977).

Most of the cycles can be traced from a marginal conglomeratic facies, by vigorous interfingering, into an axial sandstone facies (Fig. 9.2). The coarsening-upwards in each conglomerate cycle (Fig. 9.4), represents the basinwards progradation of a marginal alluvial fan body, and it is believed that smaller segments (of the order of 10-20 m) within each fan are of direct tectonic significance (Larsen and Steel, in press). Of particular importance in the Hornelen Basin is the fact that the fans radiating from the southern margin are relatively large, relatively thin and stream-dominated while those along the northern margin are small, thick and dominated by debris flow deposits (Fig. 9.2) (Steel, in press). This essentially reflects the overall asymmetry of basin subsidence, as outlined below in the tectonic model (Fig. 9.5).

The axially developed sandstones have been dispersed largely westwards (longitudinally), but their cyclicity is in notable continuity with the marginal fanglomerates and also is dominated by a coarsening-upwards motif (Fig. 9.3). The sandstones represent a wide variety of environments within an extensive alluvial plain system. Braided river channel sediments pass into flood-basin and lacustrine deposits via a variety of levee, crevasse splay and crevasse channel sub-environments. The prominent axial cyclicity (Fig. 9.3) has resulted from repeated westwards progradation of the alluvial plain system in response to an equivalent number of basin-floor subsidence episodes.

It is likely that the apparent 25 km thickness has accumulated by progressive eastwards overlap of the basement. Estimation of the amount of overlap per cycle suggests that the locus of sedimentation steadily shifted eastwards by the order of 0.25 km between the westwards progradational episodes constructing each cycle. It has also been proposed that the simplest method to produce this overlap of successive basin-fill increments is periodic right-lateral fault movement, with a strike-slip component about twice as large as the dip-slip component, along a "restrictive" double bend on the northern margin of the basin as shown in Figure 9.5. This proposal satisfies the sedimentary evidence from the geometry of the basin cycles, from the palaeocurrents (Fig. 9.5b) and is not inconsistent with the present tectonic features of the basin (Fig. 9.5a). An additional notable feature is the asymmetry of the basin with respect to fan type, size and thickness on opposite basin margins, as noted above. This can be accounted for in the proposed model since it would necessarily involve a preferential subsidence along the main northern fault zone (Fig. 9.5c), causing the observed facies contrasts. It has not previously been proposed, from independent evidence, that Nordfjord presently occupies a major strike-slip fault, nor is there evidence of the continued presence of such faulting farther east. Similar types of basins presently forming along known strike-slip faults, however, for example along the San Andreas Fault (Crowell, 1974), are frequently associated with considerable variation in throw laterally, as the strike-slip component is often "taken up" along associated minor fault lines.

THE CONTRAST BETWEEN HORNELEN AND SOLUND BASINS

Solund Basin is of the same order of size as Hornelen Basin, presently lies less than 50 km to the south, but is notably in contrast because it is filled

IGCP Project 27, Norwegian Contribution No. 5G.
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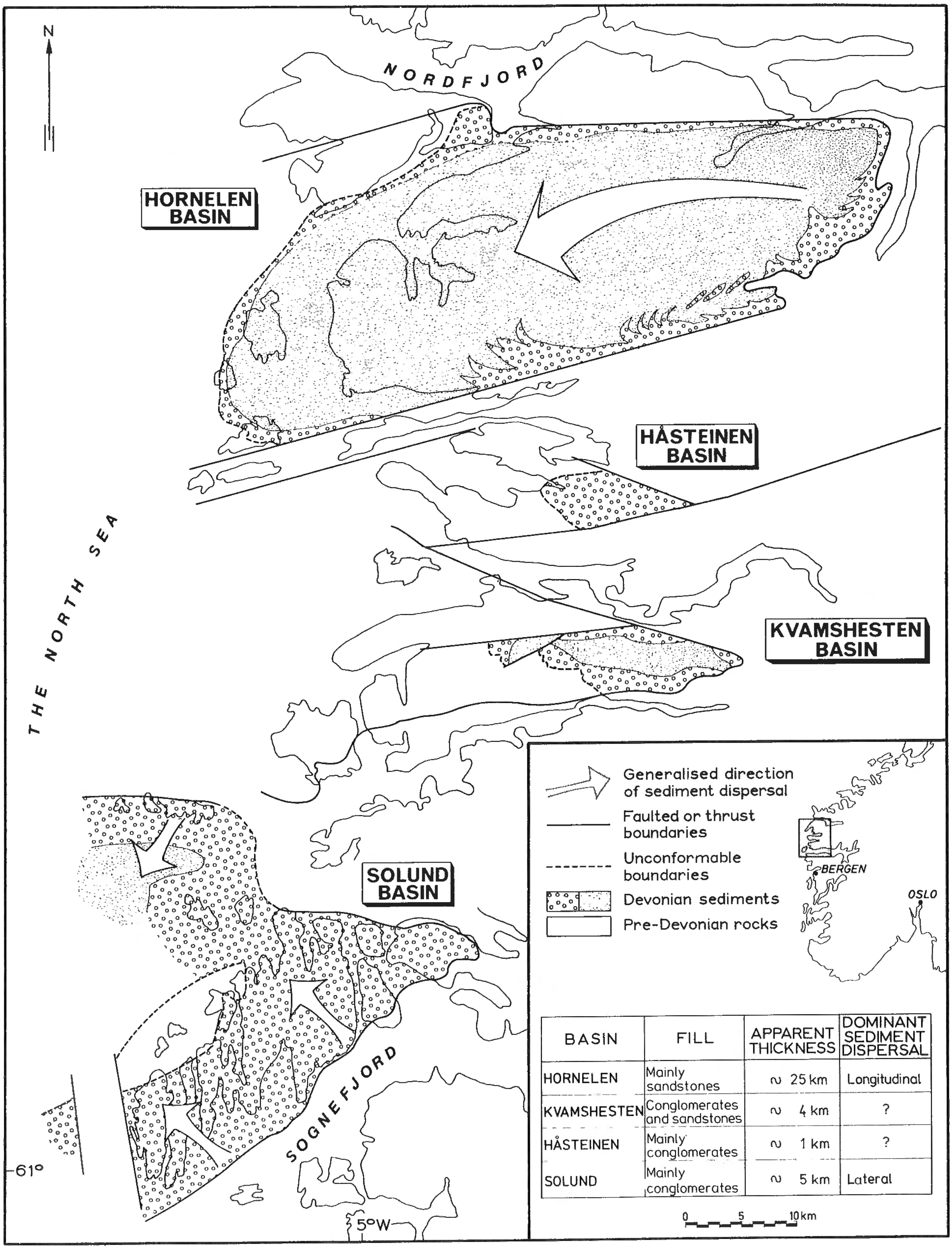

BASIN	FILL	APPARENT THICKNESS	DOMINANT SEDIMENT DISPERSAL
HORNELEN	Mainly sandstones	∾ 25 km	Longitudinal
KVAMSHESTEN	Conglomerates and sandstones	∾ 4 km	?
HÅSTEINEN	Mainly conglomerates	∾ 1 km	?
SOLUND	Mainly conglomerates	∾ 5 km	Lateral

Figure 9.1. The Devonian basins of western Norway (partly after Nilsen (1968, 1973). Note the contrasting dispersal directions, grain sizes and thicknesses of the infilling sediments in Hornelen and Solund Basins.

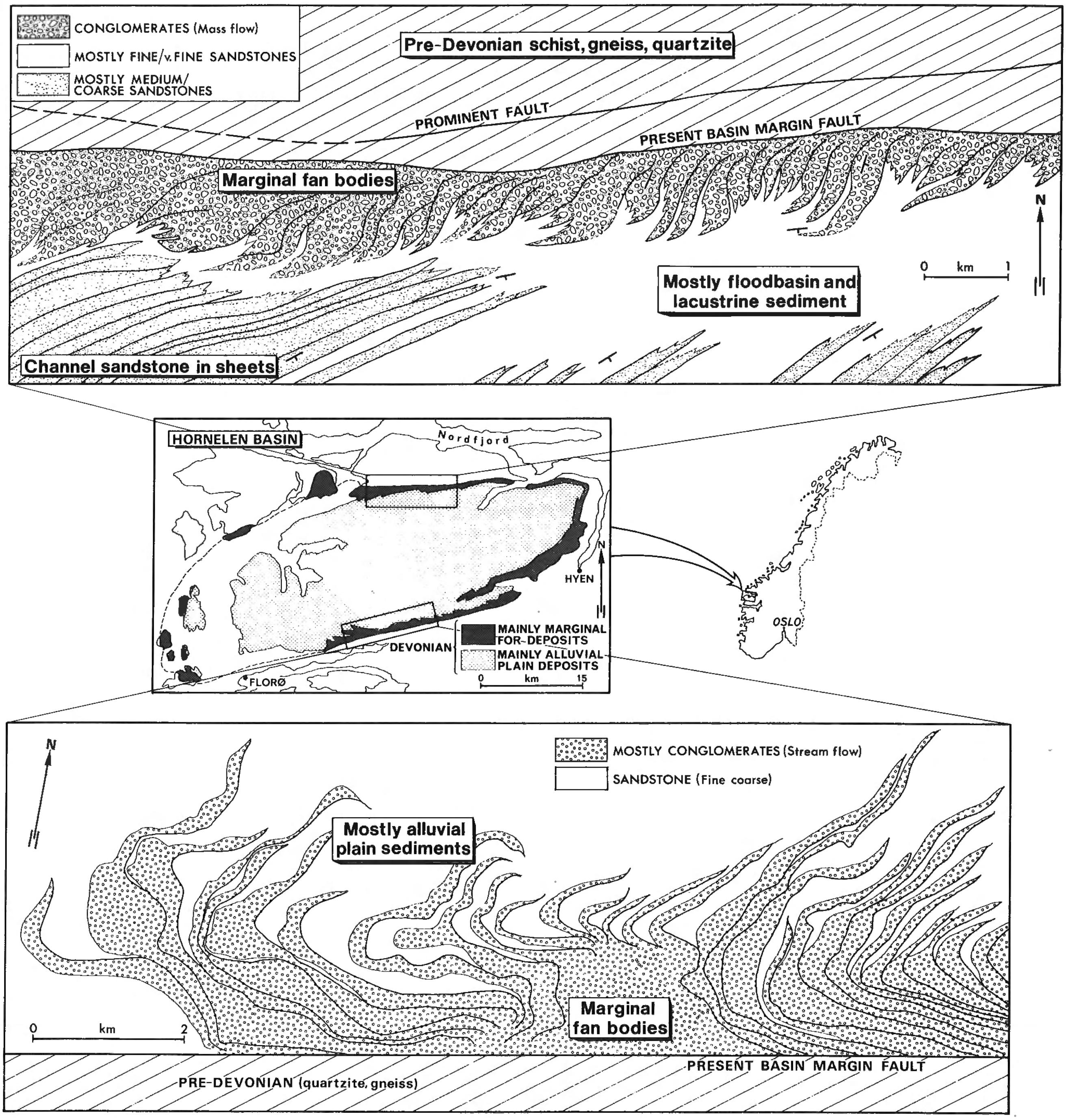

Figure 9.2. Details of the lithofacies in Hornelen Basin and, in particular, the contrast in type and size of fan body along the north and south margins: Along the northern margin the succession dips steeply (70) southeastwards, and in the south there is an open syncline, plunging 20-30 degrees eastwards.

largely with 5 km of conglomerates which have been dispersed laterally from the basin margins (Fig. 9.1) (Nilsen, 1968). A traverse through 3 kilometres of this basin sequence has shown clearly that here also coarsening-upwards conglomerate sequences, though less uniform and less frequent than in Hornelen, commonly dominate the succession (Steel, in press). The great thickness of coarse-grained conglomerate in this basin may imply repeated vertical uplift of drainage areas. This is likely to have occurred in a context of dip-slip dominated faulting. The faulting here, as in the Hornelen Basin, has produced a coarsening upwards "motif" as the basic sedimentary response to tectonism but the dip-slip movement here, in contrast to strike-slip dominated movement in Hornelen, has caused a sub-vertical stacking of prograding conglomerate bodies in contrast to an overlapped stacking of sandstone bodies in Hornelen. These two basins, therefore, demonstrate clearly the strong influence of varying tectonic context on the sedimentation history of late-orogenic basin formation in this western region of the Norwegian Caledonides.

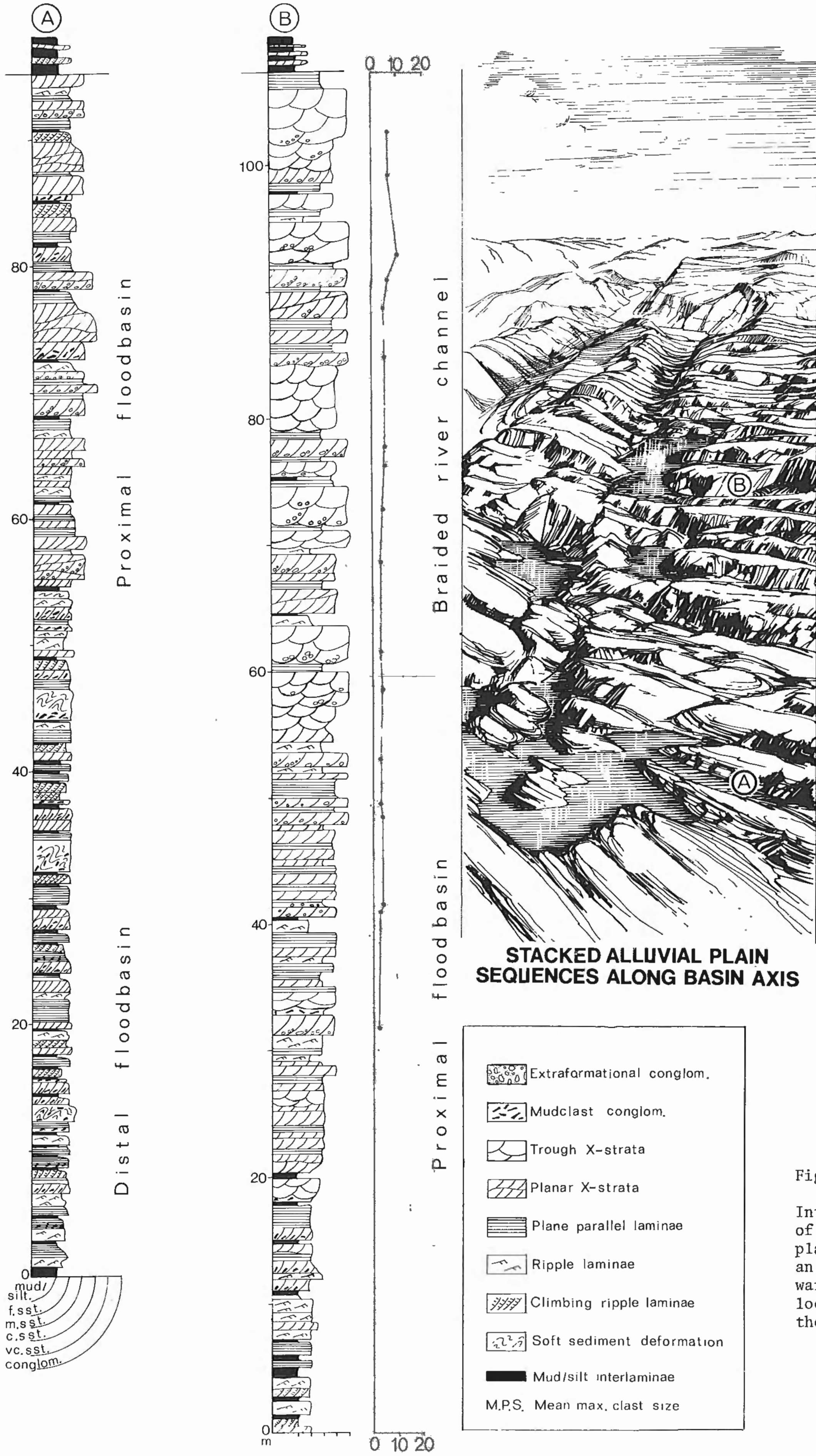

Figure 9.3.

Internal details of two types of coarsening-upwards alluvial plain sequence. The sketch, an oblique aerial view eastwards along the basin axis, locates the profiles within the markedly cyclic succession.

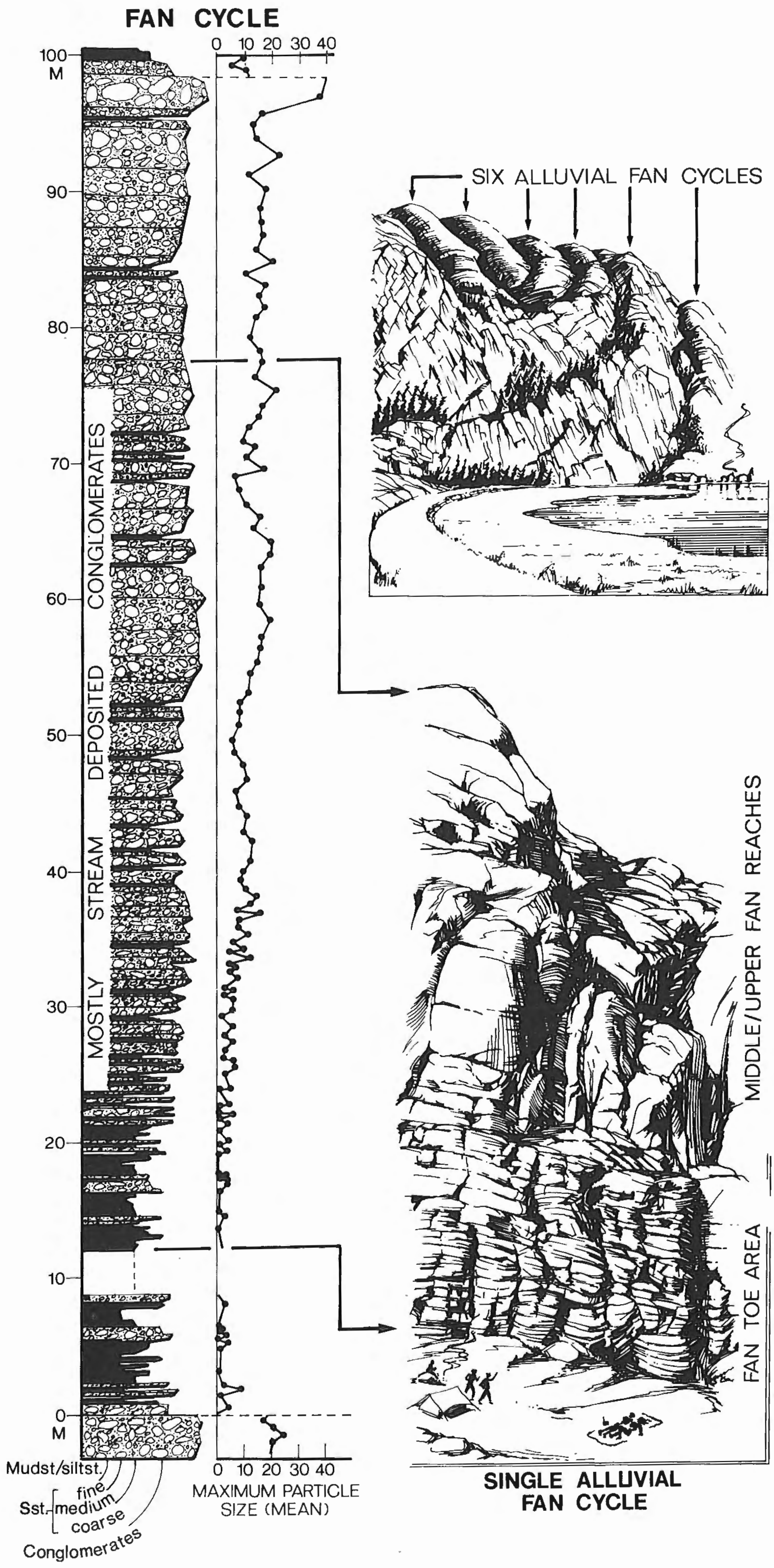

Figure 9.4.

Internal details of a single coarsening-upwards sequence from the southern margin of Hornelen Basin. The sketch shows (upper right) the typical multiple development of conglomeratic fan bodies as the marginal facies of the cyclic basin fill.

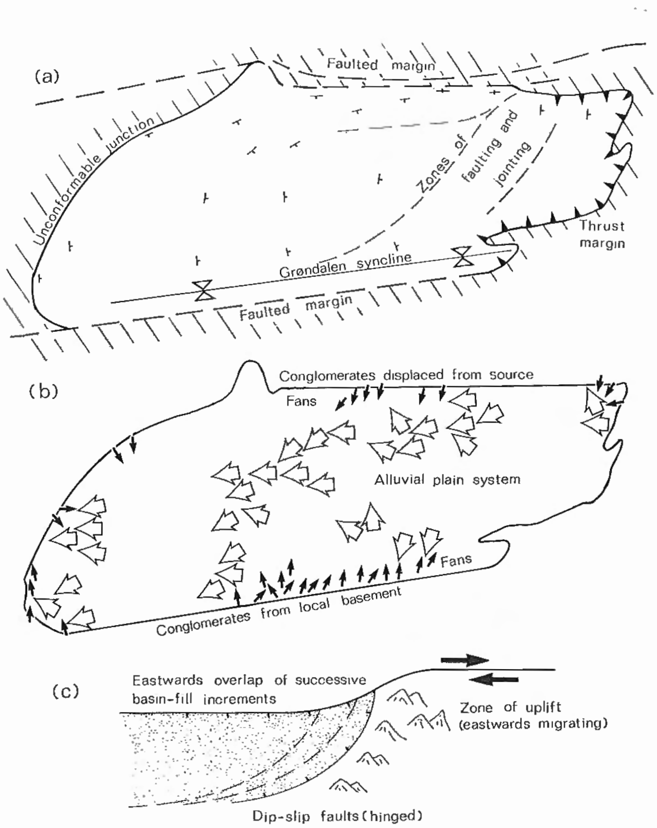

Figure 9.5.

(a) Some of the main structural features of Hornelen Basin,
(b) palaeocurrent distribution through time, and
(c) suggested origin of the basin by dextral wrench faulting. Non-continuous wrench movement caused the cyclic development of the basin succession. The basin sequence was later folded with thrusting along the eastern margin.

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THE SWEDISH CALEDONIDES - A SHORT SYNTHESIS

D.G. Gee

Sveriges geologiska undersökning
104 05 Stockholm 50, Sweden

INTRODUCTION

Reconstructions of the northern Atlantic (the Greenland and Norwegian Seas) and adjacent areas for periods prior to the Mesozoic suggest that the Caledonide Orogen occupied a linear belt up to about 1000 km wide and extending from Spitsbergen in the north to the United Kingdom and Ireland in the south. Within the mountain range of western Scandinavia (the Scandes) the Caledonides attain a maximum width of about 300 km; thus less than half of a section through the orogen is exposed on land. The structure of the Scandes is dominated by nappe tectonics with displacements from west to east across the Baltoscandian basement (Fig. 10.1). Basement and cover, both allochthonous and autochthonous, together are folded on axes parallel to the length of the orogen. These large-scale folds change in style from open and upright in the east becoming tighter further west; in the western coastal zone (Norway) they are, at least locally, recumbent. General summaries of the Scandinavian Caledonides are to be found in Strand (1961) and Nicholson (1974).

The Swedish Caledonides make up a subordinate part of the orogen in the Scandes (Fig. 10.2). Kulling (*in* Strand and Kulling, 1972) and Lindström (*in* Lundegårdh, Lundqvist and Lindström, 1974) have described the development of the orogen in Sweden. The Caledonian Front is exposed over a distance of about 900 km and the width of the belt varies from about 150 km in the south to a few tens of kilometres in the north. Four counties in Sweden contain Caledonian units in their western parts: Norrbotten, Västerbotten, Jämtland and Kopparberg (Fig. 10.3). Descriptions of the geology of Norrbotten are to be found in Kulling (1964 and in press); of Västerbotten by Kulling (*in* Gavelin and Kulling, 1955); of Jämtland by Högbom (1920) and Strömberg (in prep.); and of Kopparberg by Hjelmqvist (1966). Asklund (1960) and Kulling (1960) independently summarized the southern and northern parts of the Swedish Caledonides in a description accompanying the 1:1 000 000 geological map of Sweden (Sveriges geologiska undersökning, 1958).

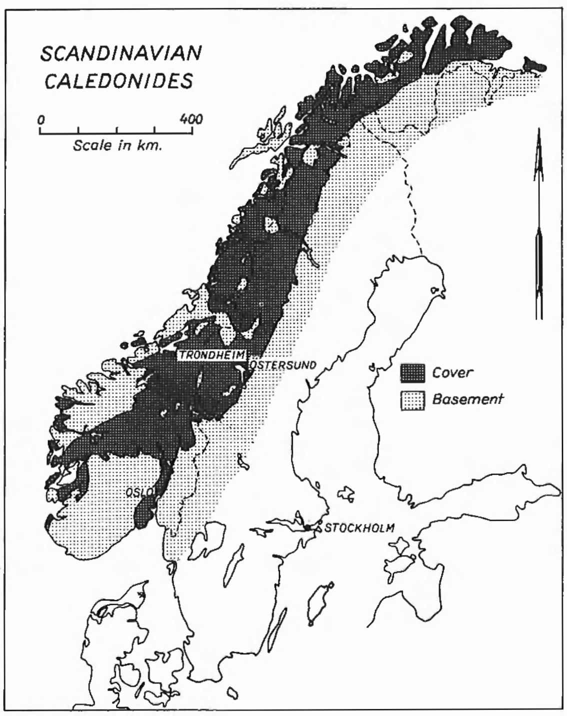

Figure 10.1. Basement-cover relationships in the Scandinavian Caledonides.

GENERAL STRUCTURE

During the middle and latter parts of the 19th Century, the theory of nappe tectonics was applied successfully in several of the major mountain belts of the world. In the Scandinavian Caledonides, Törnebohm (1888, 1896) demonstrated displacements in the order of 100 km, drawing attention to the superposition of a high grade metamorphic suite (the Seve) on the largely unmetamorphosed Lower Palaeozoic sediments and underlying granitic and rhyolitic basement of the Baltoscandian Platform. Some forty years later, Holtedahl (1936) inferred derivation of the Jotun Nappe from west of the Norwegian Coast, and in 1938 Asklund presented a basis for requiring that the displacement of the Great Seve Nappe was at least twice the distance established by Törnebohm (1888).

The allochthonous character of the major thrust sheets in the eastern part of the Scandes is, in general, easily recognizable, thick mylonite zones separating distinctive tectonic units, the latter often showing markedly different metamorphic parageneses. The basement beneath the nappes is unambiguously of lower grade than the allochthon; it is in part imbricated and gently folded by the major Caledonide (NNE) open antiforms and synforms. Further west, along the Norwegian coast, basement and cover, often of similar metamorphic grade, are intimately folded together and "consequently, nappe interpretations of relationships are less than pressing unless urged by wider considerations" (Nicholson, 1974, p. 163). Hence, until the 1970's, those accepting derivation of the major allochthon from west of the Norwegian coast have largely derived their experience of the orogen in eastern parts, Kautsky (1946, 1953) in the Sulitelma area, Kulling (1960) in Norrbotten and Västerbotten, and Zachrisson (1969) in northern Jämtland and southern Västerbotten. Oftedahl (1966) drew attention to the importance of Kautsky's interpretation of the northern Scandes. Gustavsson (1966) likewise inferred the allochthonous nature of the cover sequences in northern Norway. The alternative hypothesis preferred by many authors, particularly during the 1960's, required accumulation of eugeosynclinal pile in western Norway followed by uplift, gravitational collapse and displacement eastwards of the

IGCP Project 27, Swedish Contribution.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

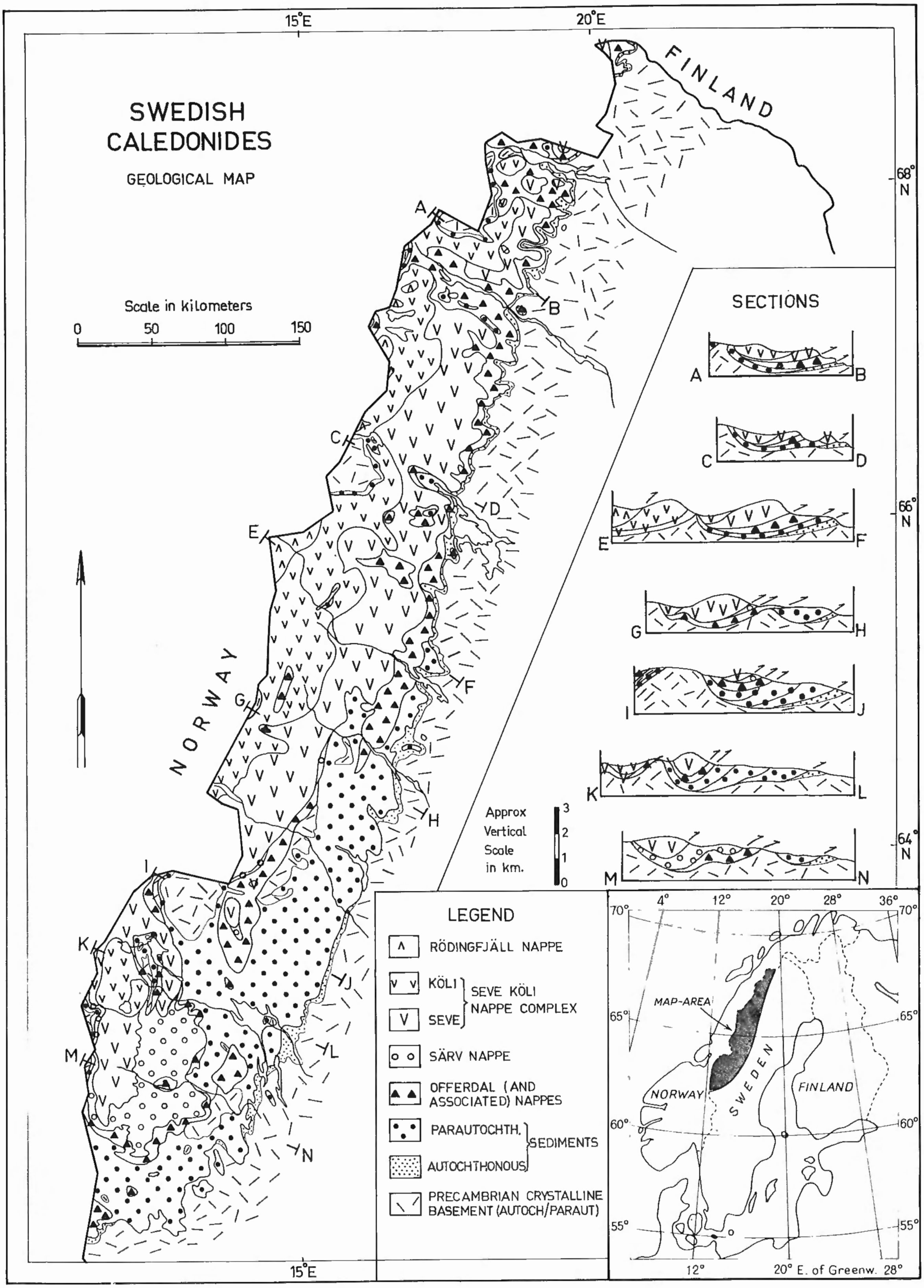

Figure 10.2. Geological map of the Swedish Caledonides.

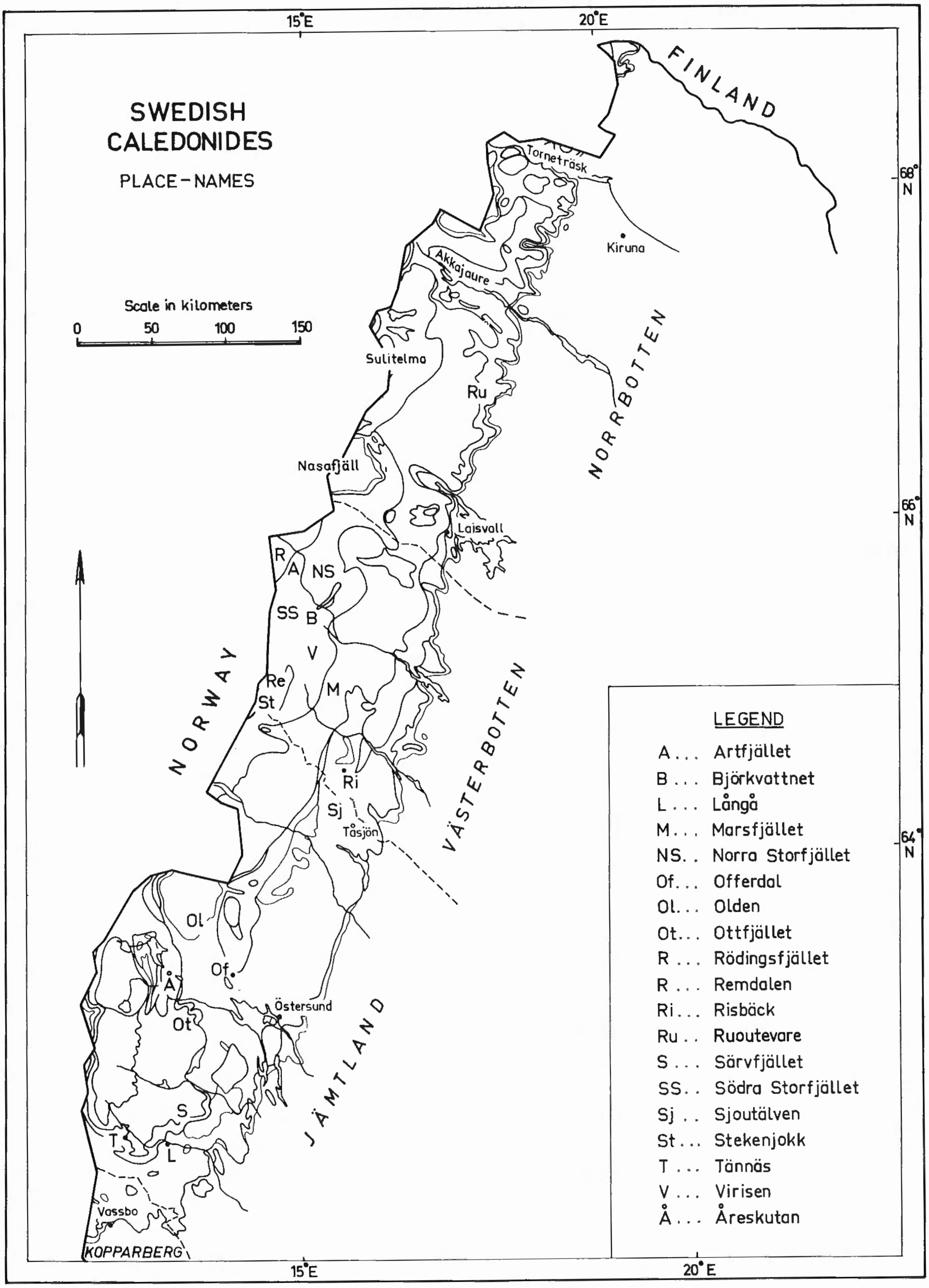

Figure 10.3. Location of places referred to in the text.

nappe units largely by stretching. This hypothesis required the displacement distances to decrease from over a hundred kilometres in the east to zero in the west. Analysis of the geological relationships renders this hypothesis untenable. Nevertheless, collapse and stretching clearly play an important role in the movement of the nappes (Ramberg, 1966).

Within the Swedish part of the Scandinavian Caledonides six main tectonic units are recognized and described below - the Autochthon, Parautochthon, Offerdal and associated nappes, Särv Nappe, Seve-Köli Nappe Complex and Rödingsfjäll Nappe (Fig. 10.4). The term Parautochthon is preferred for all the nappe units that occur between the Offerdal Nappe and the Autochthon, units that show unambiguous stratigraphic affinities with the autochthonous sediments. Recent work suggests that the distinction between the highest parautochthonous units and the lowest units in the Offerdal Nappe is insecure in some areas. Basement beneath the parautochthonous nappes, appearing in the deeper parts of the windows, may be more or less autochthonous, influenced only by vertical movements, or parautochthonous and displaced a short distance eastwards. In that the basement is clearly not passive in these areas, it is treated here as parautochthonous. Likewise the thin sedimentary veneer on this basement, which at least locally, can be demonstrated to have been deposited on the Precambrian crystalline rocks, is treated here as parautochthonous.

Westward thinning is the geometrical characteristic of all the nappe units as is apparent from the sections (Fig. 10.2). Zachrisson (1969, 1973) and Nicholson and Rutland (1969) drew attention to this phenomenon for the higher tectonic units. However, the thinning is far from regular, some tectonic units "pinching-and-swelling" both along and perpendicular to the axis of the orogen. Recently, very large scale pinch-and-swell structures have been shown (Gee, 1976 and 1977) to be composite phenomena composed of more than one nappe unit, implying build up of the nappe pile prior to collapse and stretching. This is treated further below.

AUTOCHTHON

Basement

The autochthonous basement is largely composed of crystalline rocks of Svecofennian (c. 1700-1800 Ma) age. In the northernmost area (north of Kiruna) an older gneissic basement is inferred on the basis of U/Pb age determinations on zircon and sphene (Kuovo; *in* Welin *et al.*, 1971). Throughout much of the Caledonian Front granites and gneisses dominate, intruded by occasional gabbros and dolerites. The latter locally have been dated by the K/Ar method to c. 1250 Ma (Welin and Lundqvist, 1975). East of the southernmost part of the Front, granites and related porphyritic rhyolites dominate the basement (Hjelmquist, 1966) and have been shown to be late or post-Svecofennian in age (Welin and Lundqvist, 1970). Overlying sandstones (Dala) are closely associated with the rhyolites.

Precambrian sandstones (Sjöfall sandstone, Ödman, 1957) of similar or younger age than the Dala sandstones (Witschard, pers. comm.) occur locally in Norrbotten in the area southeast of Akkajaure, unconformably beneath the autochthonous Cambrian sediments.

Caledonian sedimentary sequence

The autochthonous sediments of the Caledonian Front in Sweden are usually thin (less than 150 m) and often in the order of a few tens of metres. They may locally be absent, the lowest thrust sheets of the parautochthon riding directly on the basement surface. In general, the décollement surface occurs in the black alum shales of Middle and Upper Cambrian age, a unit usually not exceeding 50 m in thickness and underlain by a few tens of metres of shales and sandstones. Lower Cambrian body fossils have been recorded from a variety of localities in these shales underlying the alum shales and trace fossils are common in some of the sandstones (Kulling; *in* Strand and Kulling, 1972). In the Torneträsk area of northern Sweden (Fig. 10.3), about 100 m of shales and sandstones (Torneträsk Group, Kulling, in press) are preserved below the alum shales. Kulling has described a breccia (Vakkejokk) of possible glacial origin (Varangian) and a sandstone unit with Precambrian fossils including *Spriggia* sp., the relationship between these two units being uncertain. Locally (e.g. Laisvall and Vassbo) these autochthonous sandstones contain important lead-zinc mineralizations (Grip, 1967).

In central Jämtland the autochthonous sequences reach into the Upper Ordovician. Local "highs" occur (Thorslund, 1943) with Middle Ordovician limestones deposited directly on the Precambrian crystalline basement.

PARAUTOCHTHON

Above the décollement surface in Sweden, the lowermost nappe units usually comprise thin slices of sediments comparable with the autochthon and dominated by quartzites, black shales and greywackes. Throughout much of Norrbotten and northern Västerbotten these are generally too thin to show on Figure 10.2. In southern Västerbotten and Jämtland their development below the Offerdal Nappe is more complete and a series of thrust units contains sediments (the Jämtland Supergroup) ranging in age from late Precambrian to Middle Silurian. The general stratigraphy is illustrated in Figure 10.4 and has been described by Thorslund (1960). A few hundred metres of pre-Varangian fluviatile and shallow marine sandstones, the Risbäck Group in northern Jämtland and southern Västerbotten (Kulling; *in* Strand and Kulling, 1972 and Gee *et al.*, 1974) have been referred to as the Långå Group in southern Jämtland (Strömberg, 1974). They are overlain by a few metres to tens of metres of glaciogenic sediments (tillites and laminated shales with drop-stones) and then sandstones (quartzites) and shales up to 250 m thick, together composing the Sjoutälven Group (Gee *et al.*, 1974). These sandstones and shales are correlatable with the autochthonous sandstones and pass up through grey-green siltstones into the alum shale formation of Middle and Upper Cambrian and locally of Tremadocian age. The black alum shales and overlying shales and greywackes are referred to as the Tåsjön Group. The greywackes occupy large areas of central Jämtland and are probably in the order of about 500 m thick. They range in age from Arenig to Caradoc, in the Ashgill passing up into dark shales. Laterally, the greywacke facies passes eastwards into platform limestones in the lowermost Jämtland nappes and the autochthon. In the easternmost parautochthon there is an uninterrupted passage from Ashgill shales into Llandovery quartzites. (Thorslund, 1943). Further west, uppermost zones of the Ashgill are apparently absent and thin shallow water sandstones occur in the lowermost Llandovery, apparently conformably overlying the deeper water Ordovician sequence, and passing up into coral limestones and then dark shales and greywackes, this entire sequence being called the Ånge Group (Strömberg, 1974). The youngest sediments were thought to be of Llandovery age (Thorslund, 1948, 1960) but are now thought to be younger (possibly of Ludlow age, Karis pers. comm.).

The occurrence of the pre-Varangian rocks in the northern and southern parts of Jämtland was previously interpreted (Asklund, 1960) as being related to

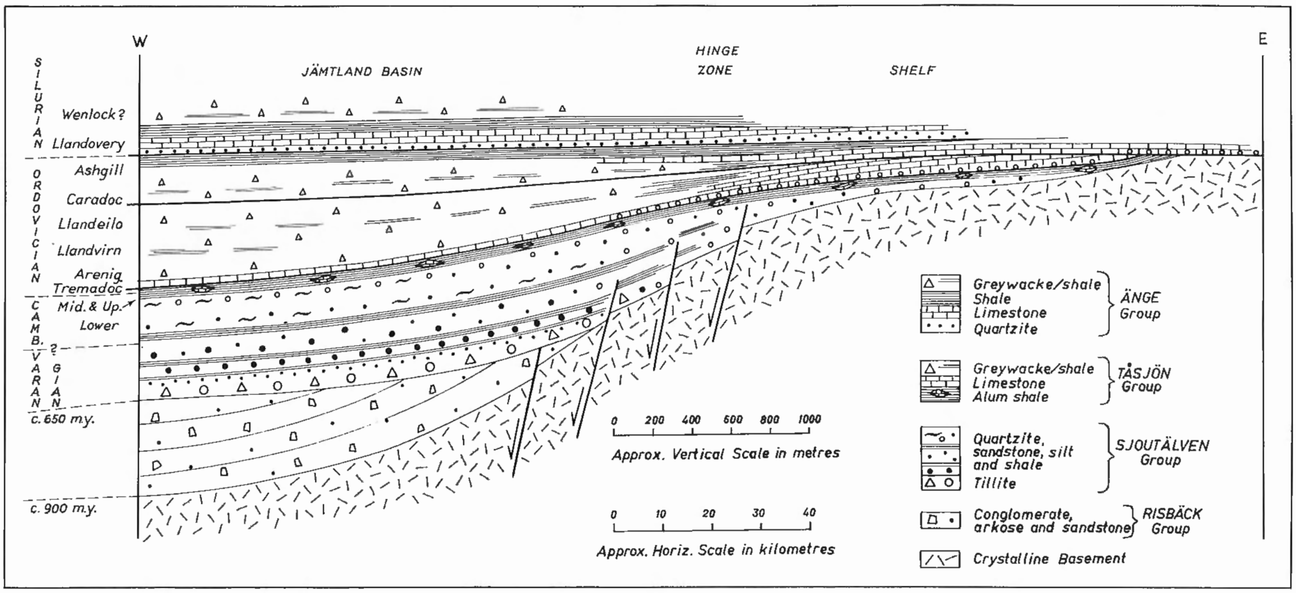

Figure 10.4. General facies relationships of the Jämtland Supergroup (taken with modification from Gee, 1972, 1975b).

deposition in local basins separated by a "high" in central Jämtland. Recent investigations have shown the pre-Varangian sequence in northern Jämtland to be allochthonous and this is also probably the case in southern Jämtland and Kopparberg Counties. Thus the stratigraphic development described from central Jämtland and referred to as the "Jämtland Basin" (Fig. 10.4) is probably representative of areas both to the north and south. There are no major north-south facies changes within the Lower Palaeozoic sequence and the "Jämtland Basin" was probably developed extensively along the Baltoscandian Platform-margin representing a zone of crustal instability along the eastern edge of the Caledonian geosyncline. It is important to note that the displacement of these parautochthonous sequences can be shown (at least in northern Jämtland) to exceed 70 km implying that the "Jämtland Basin", prior to orogenesis, extended far west into Trøndelag.

Along the border between Norway and Sweden there occur a number of windows exposing Precambrian basement crystalline rocks. This "axis" of windows (Fig. 10.1) can be traced both further northeastwards into northernmost Norway and southwards into southern Norway. Of fundamental importance to the interpretation of the derivation of the Scandian nappes is the recognition that the autochthonous sediments in the Caledonian Front are similar to those deposited on the basement of these windows. A number of subordinate windows occur between the Norwegian/Swedish border and the Caledonian Front and they likewise show a similar stratigraphic development, the Lower Cambrian and perhaps the latest Precambrian sediments appearing west of the Front as characteristic white, yellow, and blue quartzites. Locally, they are overlain by black shales, limestones, and greywackes. Varangian tillites appear in some of the western windows and are the oldest sediments deposited on and east of these windows. Further west, a feldspathic sandstone/arkose facies is developed comparable with the Risbäck Group of the northern Jämtland parautochthon.

Minor basement lenses up to a few tens of metres thick and several hundred metres to a few kilometres in other dimensions occur as isolated slices in the nappes of the parautochthon, usually at the base of the quartzites. At several localities their geometry is well established by drilling (e.g. Laisvall, Lilljequist, 1973) and they have been interpreted as sliced-off tops of basement irregularities (monadnocks) on an otherwise regular late Precambrian peneplain. Further west in the vicinity and west of the windows basement imbrication is extensive (e.g. in the Grong-Olden culmination, Gee, 1975b).

OFFERDAL AND ASSOCIATED NAPPES

The Offerdal Nappe has been traditionally treated as the lowermost unit of the major allochthon in Sweden. In the type area in central Jämtland it is composed of unfossiliferous feldspathic sandstones and crystalline basement rocks and has been referred to by some authors (Asklund, 1960) as the granite mylonite nappe. In southern Jämtland (Fig. 10.3) a tectonic unit composed of augen granites and augen gneisses (the Tännäs Augen Gneiss) apparently is developed at the same tectonic level (shown as Offerdal and associated nappes, Fig. 10.2). Northwards from Jämtland, cataclastic granites occur locally between the parautochthon and the overlying Seve-Köli Nappe Complex and are extensively developed in the Akkajaure Complex of Norrbotten. All these units have clearly been derived from west of the windows along the Norwegian-Swedish Border. Kulling (1960) required a derivation zone along the Norwegian west coast for the Akkajaure Complex and Gee (1975a and in press) has outlined the evidence for derivation of the Offerdal Nappe of central Jämtland from westernmost Trøndelag, Norway.

Crystalline basement units carry a penetrative, cataclastic foliation; the meta-sandstones, generally in greenschist facies, are isoclinally folded with development of a penetrative, fine-grained schistosity. These foliations and the extensive mylonites, both within and at the base of the nappe, are generally flatlying and truncate the cleavages, folds and thrusts in the underlying parautochthon. Thus, though the general regional evidence requires displacement of the Caledonian nappe-pile to commence in the west and progress eastwards (Gee, 1975a) the parautochthonous units were apparently deformed prior to being overriden by the Offerdal Nappe.

Late Precambrian feldspathic sandstones (Risbäck Group) in the highest tectonic units of the parautochthon are compositionally similar to the overlying Offerdal Nappe sediments. Distinction between the Offerdal and these lower nappes is locally ambiguous particularly in northern Jämtland. In Norrbotten, feldspathic sandstones and arkoses that are overridden, at least in part, by the Akkajaure Complex may have been deposited on the basement from which the latter was derived. They are transported from west of the border windows and are included here in the Offerdal Nappe.

SÄRV NAPPE

The Särv Nappe (Fig. 10.2) has been identified and defined in Jämtland (Strömbert, 1955, 1961). It is composed essentially of a metasandstone sequence (the Särv Group, Strömberg, 1969) intruded by an olivine-bearing tholeiitic dyke-swarm (the Ottfjället Dolerites). The sedimentary sequence (c. 2000 m+, Kumpulainen, pers. comm.) contains subordinate limestones and a diamictite of probable glacial origin (Röshoff, 1975). Within the main outcrop area of the Särv Nappe the basic sheets dip westwards at moderate to high angles and sediments dip eastwards, a high angle usually being preserved between the intrusions and the country-rock. Prior to nappe emplacement, the sandstones were largely flat-lying intruded by a vertical, approximately north-northeast trending dyke-swarm. There is as yet no unambiguous evidence of regional deformation of the Särv Group prior to intrusion of the dyke-swarm although, locally, dyke-sediment relationships have been interpreted to imply pre-dyke folding (Strömberg, 1961).

The age of the dyke-swarm is controversial. Recent isotopic age-determinations by the K/Ar method have yielded ages ranging from c. 2 500 to 600 Ma (Point *et al.*, 1976; Claesson, 1976). The Rb/Sr method has provided an isochron favouring a late Precambrian age (735 ± 260 Ma, Claesson, 1976).

Internal deformation of the nappe is concentrated to zones of high strain with rotation of the dykes towards parallellism with the bedding in the sandstones, development of a penetrative foliation and general retrogression associated with greenschist/lower amphibolite facies metamorphism. Outside these zones of high strain, sedimentary structures and intrusive relationships between the dykes and the sediments are well-preserved. At the base of the nappe, locally, a blastomylonite zone sharply truncates the dyke swarm; however, generally the sediment-dyke angle decreases towards the base of the nappe and the rocks pass downwards into banded greenschists and feldspathic psammites above the basal mylonites. Where the nappe thins both westwards and northwards, concordance between greenschists, amphibolites and psammites is general and discordant intrusive relationships are recorded only from local "swells".

The Särv Nappe has been identified as far north as northernmost Jämtland in Sweden; it may well be present further north. In northernmost Norway comparable units occur in the Laksefjord Nappe (Gayer, pers. comm.). In Kopparberg County and further southwest in Norway the Kvitvola Nappe (Törnebohm, 1896) has been correlated with the Särv Nappe on the basis of the similarity of the sediments. Dykes are apparently lacking. It is improbable that the dyke swarm originally terminated abruptly southwards; thus correlation of the Kvitvola Nappe with the underlying Offerdal Nappe is preferred.

SEVE-KÖLI NAPPE COMPLEX

The Seve-Köli Nappe Complex (Zachrisson, 1969; Zwart, 1974) is divisible into two major rock units, a lower Seve and an upper Köli division. The Seve is generally of higher metamorphic grade than both the Köli and the underlying Särv Nappe. Both Seve and Köli are complex, heterogeneous, polymetamorphic units composed of several subordinate nappes. The Köli contains Ordovician and Silurian fossiliferous units and the Seve, at least locally, includes slices of Precambrian basement rocks. However, age-relationships between the two units are not fully understood; the contact between them is usually "transitional". Regional discordance and development of local phyllonites have been demonstrated in southern Västerbotten and Jämtland.

Within the Seve-Köli Nappe Complex, there is abundant evidence of small and medium scale folding and refolding and sequences of deformation in relation to metamorphism have been described from various areas (Zachrisson, 1969; Trouw, 1973; Zwart, 1974; Stephens, 1977). However, the basic structure of the complex is dominated by thrust nappes, not by recumbent fold nappes, the thrusting accompanying some of the earlier phases of folding (often recumbent on WNW-axes) and folded by the later NE-trending, upright antiforms and synforms. The earliest isoclinal folding apparently predated the thrusting of the nappe complex and the peak of the regional metamorphism, both of these having clearly occurred prior to displacement of this allochthon into Sweden.

Seve

Aspects of the development of the Seve in Sweden has been treated by Zachrisson (1973) where attention was drawn, in particular, to the gross geometry of the nappe with thinning from a thickness of a few kilometres in the east to zero in the vicinity of the border zone windows. The Seve continues northwards into the Kalak Nappe in northern Norway (this volume, Article 4). Recent evidence (Gee, 1977) indicates that, at least in the central part of the Scandes, the Seve "swells" again in Norway, west of the border zone windows, forming parts of very large scale lenses in the Norwegian allochthon.

Where the Seve has been investigated in some detail (e.g. on Åreskutan, Helfrich, 1967; Marsfjället, Trouw, 1973) it has been shown to be composed of at least three major tectonic units separated by blastomylonite zones. Metamorphic grade ranges from lower amphibolite to upper amphibolite (locally with eclogites) and granulite facies. Some of the high-grade rocks were regarded by Högbom (1909) and Asklund (1938, 1960) as being of pre-Caledonian origin and this view has received recent support from age-determination investigations by Reymer (pers. comm.) in northern Jämtland/southern Västerbotten and Koark (pers. comm.) on Åreskutan. Further north, in Norrbotten, huge lenticular anorthosite units (Ruoutevare anorthosite) with subordinate gabbro/norites and associated titaniferous magnetite ores occur in the base of the Seve-Köli Nappe Complex. The largest anorthosite lens is c. 700-1000 m thick and has a north-south dimension of about 15 km and an east-west axis of at least 20 km. Although subject to Caledonian metamorphism, with development of almandine and well-lineated hornblende, these units are very probably of Precambrian origin. Similar anorthosites, basic rocks and ores occur in the Lofoten basement (Romey, 1971) some 200 km to the northwest, a displacement distance which is compatible with the tectonic position of these Seve units above the Akkajaure Complex. Smaller lenses of anorthosite have been reported from the same tectonic level in southern Västerbotten.

The Seve is dominated by amphibolites, schists and gneisses, their approximate distribution being shown on the 1:1 000 000 geological map of Sweden. Massive concordant basic rocks, completely recrystallized in amphibolite facies, occupy large areas; they may be of intrusive or extrusive origin. Ultrabasic rocks are also locally frequent. Subordinate quartzites and marbles occur in some areas. Schistose meta-arkoses and banded amphibolites are present in the lower parts of the Seve in Jämtland. They are of higher metamorphic grade (containing eclogite in northernmost Jämtland) than the Särv Nappe to which they might otherwise be referred.

Köli

Seve schists and amphibolites pass downgrade upwards into the phyllites and greenschists of the Köli. The contact zone is generally concordant and transitional. In some areas, such as southern Västerbotten, a tectonic contact (Zachrisson, 1969) has been inferred on the basis of the presence of phyllonites, the existence of a low angle regional discordance between the Seve and Köli units and the absence of certain isograds (Trouw, 1973). Within those areas where the Köli has been investigated in some detail (Kautsky, 1953; Zachrisson, 1969) it has been shown to be a composite, generally low-grade unit made up of at least three nappes. The recognition of tectonic boundaries in the Köli successions and the relationship of the latter to the pre-orogenic volcano-sedimentary stratigraphy is controversial (e.g. Henley, 1970). Well-preserved diagnostic fossils are few and assessment of the age of the concordant, attenuated sequences is often problematical.

In the lowest tectonic division of the Koli, Kulling (1933; *in* Gavelin and Kulling, 1955; 1958; *in* Strand and Kulling, 1972) described a succession which included two fossiliferous units. These provided support for his stratigraphic interpretation and established that at least a part of the Köli was of Ordovician and Silurian age. He described from the now classical, Björkvattnet-Virisen area, a part of which has been treated recently by Stephens (1977), a sequence of phyllites and ultrabasic rocks, including detrital serpentinites overlain by greenschists (Seima Formation). The latter pass up into greywackes and conglomerates (Gilliks Formation) overlain by quartzites (Vojtja Formation) and coral limestones (Slätdal Formation) of Ashgill age and then by black phyllites (Broken Formation) of Middle and Upper Llandovery age. These pass up transitionally into calcareous greywackes and phyllites (Lövfjäll Formation) and then into coarse greywackes and sandstones (Vesken Formation) overlain by quartzites (Viris Formation). Subsequent mapping has identified an overlying sequence of calcareous phyllites, conglomerates and greywackes. The thickness of the entire section in the type area is in the order of a few kilometres but the relationship of this figure to the pre-orogenic thickness remains unassessed.

Passing westwards the succession thins dramatically; units at the base are cut out and other lithologies appear above. From areas in southern Västerbotten adjacent to the border with Norway, Zachrisson (1964 and 1969) has described three groups, the Tjopasi, Lasterfjäll and Remdalen Groups, the first including the Gilliks and underlying formations referred to above, the second starting with Vojtja Formation and including all the overlying formations described by Kulling and reaching into overlying greenschists and keratophyres and the third starting with conglomerates (the Remdalen conglomerate) and passing up into mixed phyllites and largely basic, volcanic rocks. This succession is illustrated in Figure 10.5, where it is compared with the Jämtland Supergroup of the parautochthon. This Köli succession in southern Västerbotten/ northern Jämtland, was shown (Zachrisson, 1969) to be overthrust by two nappes, the Gellvernokko and Leipik Nappes. Neither contain fossiliferous units but both are made up of lithologies comparable in composition and metamorphic grade to the underlying Köli Supergroup.

In central and northern Västerbotten, the sequences of Kulling's classical stratigraphic succession are overthrust by a higher grade unit in the Södra Storfjäll (Beskow, 1929) and Norra Storfjäll (Kulling; *in* Gavelin and Kulling, 1955) areas, the Storfjäll Nappe Complex (Kulling; *in* Strand and Kulling 1972). The lithologies are in many respects comparable to the Köli Supergroup and contain poorly preserved fossils of Ordovician or Silurian age. Synorogenic gabbros (Artfjället) and granites (Vilasund) intrude the succession. The metamorphic grade in Södra Storfjället reaches upper amphibolite facies with development of garnet, staurolite schists, whilst in Norra Storfjället kyanite schists pass laterally into migmatites. The inter-relationship of these high grade units in Södra and Norra Storfjället has remained obscure until recently. Häggbom (1976)

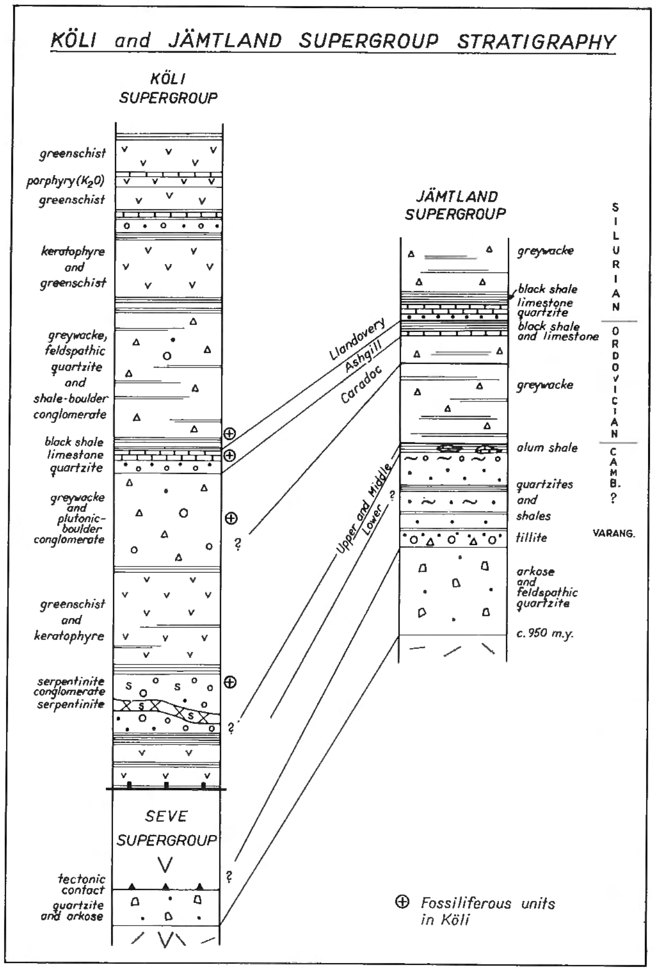

Figure 10.5. Köli Supergroup stratigraphy compared with the Jämtland Supergroup (from Gee, 1975a).

demonstrated that their gross geometry was that of very large-scale lenses. The high grade basal part of the Storfjäll Nappe Complex shows a pinch-and-swell geometry within the Köli comparable to that of the huge amphibolite units in the Seve. In the upper part of the Storfjäll Nappe, metamorphic grade decreases to greenschist facies.

Köli stratigraphy further north in Norrbotten is less well-known. Kulling (*in* Strand and Kulling, 1972) has described fossiliferous units of comparable lithology and similar age to those in Västerbotten. Different Köli nappe units, the Salo, Vasten and Pieske Nappes, have been identified by Kautsky (1953) and disputed by Kulling (*in* Strand and Kulling, 1972).

The Seve-Köli Nappe Complex contains a large number of complex polymetallic sulphide mineralizations, the most important being located in the Köli. One of these ores is in production (at Stekenjokk in southernmost Västerbotten) and has been described by Zachrisson (1971) and Juve (1974).

RÖDINGSFJÄLL NAPPE

Above the phyllites of the Storfjäll Nappe Complex in northern Västerbotten is another higher grade unit, referred to as the Rödingsfjäll Nappe (Kulling; *in* Gavelin and Kulling, 1955). Composed of a migmatite complex in its lower part, it passes up into schists, marbles and amphibolites. At various localities further north in Norrbotten the uppermost units in the Swedish Caledonides (e.g. the Gasak Nappe, Kautsky, 1953) are upper amphibolite facies schists and amphibolites. These have been correlated with the Rödingsfjäll Nappe and this interpretation has been shown in Figure 10.2; nevertheless correlation with the Storfjäll Complex is equally possible.

CONCLUDING REMARKS

Interpretation of the evolution of the Caledonide Orogen in Sweden cannot be treated independently of the Scandes as a whole. The allochthon and part of the parautochthon have been derived from Norway or west of the Norwegian coast. Estimates of displacement distance of the nappes and timing of this movement are fundamentally dependent on data from the Norwegian Caledonides. Late-orogenic sedimentation of molasse character in intramontane basins is restricted to Norwegian territory and critically influences assessment of the timing of the orogenic processes. The mapping of the nappe units westwards towards the Norwegian coast provides a basis for obtaining minimum estimates of displacement distance. During recent years Swedish research in the context of the International Geodynamics Project (Annersten, 1973; Bylund *et al.*, 1976) has concentrated on a profile through the central part of the Scandinavian Caledonides and this has resulted in various attempts (Gee, 1975a and in press) to analyse the evolution of the orogen from the late Precambrian to the Devonian. It is apparent that displacement distances are vast, in the order of 1000 km, movement having occurred during the Silurian (post-Llandovery) and continuing into at least the Early Devonian. Translation can be estimated to a minimum of c. 500 km and involved build up of a nappe-pile composed of basement and cover elements that both have a higher density than the underlying Baltoscandian basement with its veneer of non-volcanic clastic sediments. Collapse of this nappe pile resulted in the vast pinch-and-swell geometry of the major allochthonous units and further displacement eastwards onto the Baltoscandian Platform.

Two major problem areas are central to future Caledonian research in Scandinavia. The first is the establishment in detail of the pre-orogenic relationships of the different tectonic units and the character of the environments from which they were derived. The second is the establishment within the general context of the Caledonide Orogen of a coherent theory for the emplacement of a dense allochthon several hundred kilometres onto the lighter Baltoscandian Platform.

ACKNOWLEDGEMENTS

This account of Swedish part of the Caledonide Orogen has benefitted from discussion with Dr. Ebbe Zachrisson. My thanks are also due to Mrs. Stina Järnefors and Mrs. Ann-Marie Bellström for drawing the diagrams.

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PROJECT 27 IUGS UNESCO CALEDONIDE OROGEN

THE CALEDONIDE OROGEN IN BRITAIN - FOREWORD

B.C. King

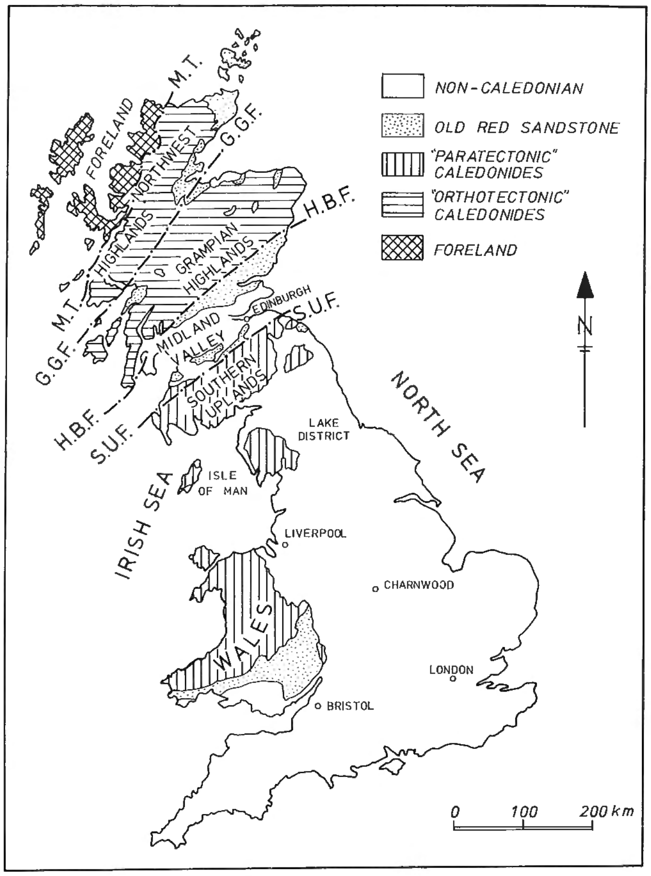
Department of Geology
Bedford College
Regent's Park
London NW1 4NS, England

The following articles are intended to be an informative statement of the state of knowledge of the Caledonides of Britain. Janet Watson (Article 12) deals with the Basement of the Orogen; Anthony L. Harris, Michael R.W. Johnson and Derek Powell (Article 13) with the Orthotectonic Caledonides (Moines and Dalradians) of Scotland; Bernard E. Leake (Article 14) with the Caledonides of the Midland Valley of Scotland; and Gilbert Kelling (Article 15) with the Paratectonic Caledonides of mainland Britain.

The greater part of Britain is a sector of the Caledonide Orogen (Fig. 11.1). Four structural divisions are conveniently recognized divided by major dislocations represented by the Moine Thrust Zone, the Highland Boundary Fault and the Southern Uplands Fault. The formations involved range in age from later Precambrian to Silurian and it is becoming increasingly clear that the earliest orogenic events date back to the Precambrian, possibly as old as 1100 Ma.

The Midland Valley is essentially a graben occupied by post-Caledonide formations, but, since the units to the north and south have totally contrasting structures and metamorphism, it must be significantly related to the Orogen itself. Both here and in the Southern Uplands to the south the nature of the Precambrian basement is a matter of speculation.

In general the Orogen is sharply defined to the northwest where strongly deformed formations override autochthonous basement, but in the south and southeast Caledonian structures weaken and fade beneath the younger rocks of the English Midlands.



G.G.F. - Great Glen Fault
H.B.F. - Highland Boundary Fault
M.T. - Moine Thrust
S.U.F. - Southern Uplands Fault
(Gilbert Kelling)

Figure 11.1 Generalized map of mainland Great Britain, showing outcrop-distribution of rocks assigned to the orthotectonic, paratectonic and molasse zones of the Caledonides.

IGCP Project 27, British Contribution No. 1A.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

THE BASEMENT OF THE CALEDONIDE OROGEN IN BRITAIN

PROJECT 27 CALEDONIDE OROGEN

Janet Watson

Department of Geology
Imperial College of Science and Technology
Prince Consort Road
London SW7 2AZ , England

NATURE AND DISTRIBUTION OF BASEMENT UNITS

Most of the Caledonian orogen is underlain by Precambrian basement of continental character which may be considered on a regional basis in terms of two domains (Fig. 12.1, inset).

1. A northwestern domain, comprising the basement of the Hebridean craton and adjacent parts of the northwestern Caledonides in Scotland and Ireland: mainly gneisses and granulites >2700 Ma.

2. A southeastern domain represented by the basement of the Midland craton and adjacent parts of the southeastern Caledonides in England, Wales and Ireland: mainly low-grade metamorphic and plutonic rocks overlain by late Precambrian volcanics.

A NE-SW zone almost devoid of evidence relative to the nature of the basement separates the domains mentioned above. It has been suggested (e.g. Dewey, 1969) that parts of this zone are underlain by basement of oceanic type marking the site of a Caledonian suture. Although geophysical evidence does not seem to support this proposal the differences in constitution and history which distinguish the basement rocks of the northwestern and southeastern domains would be compatible with the suggestion that they formerly belonged to separate continental plates.

Decisions as to which units should be assigned to the basement depend largely on the views adopted as to the time-span of the Caledonian cycle. In the northwestern domain, the Moinian and Dalradian Supergroups which include metasediments of late Precambrian age and which have undergone polyphase deformation extending over a considerable time-period have most usually been regarded as early units of the cover-succession in contradistinction to the obviously pre-Caledonian Lewisian gneisses on which they rest. Recent isotopic evidence which suggests that portions of the Moinian Supergroup and associated rocks underwent important deformation and metamorphism at or before 1000 Ma, indicates that these rocks might logically be regarded as portions of a Grenville province acting as basement to the Caledonides; because the extent of Grenville activity is not yet known, they are dealt with here along with other early cover-units. In the southeastern domain, volcanic and associated sedimentary formations of very late Precambrian age have traditionally been assigned to the basement because they are considerably more disturbed than the Lower Palaeozoic cover and are penetrated by intrusions which do not enter the cover. This procedure is followed here in spite of the fact that the rocks concerned are essentially coeval with the Dalradian Supergroup of the northwestern Caledonides, which is regarded as part of the cover.

BASEMENT OF THE HEBRIDEAN CRATON AND NORTHWESTERN CALEDONIDES

The cratonic area west of the Caledonian front in Scotland is floored by a basement of gneisses and granulites, known collectively as the Lewisian complex, which forms part of a large province originally extending into Rockall Bank and south Greenland. The principal rock-types are tonalitic to granodioritic gneisses and granulites derived from plutonic rocks and interleaved with supracrustal gneisses and with metamorphosed layered basic, ultrabasic and anorthositic igneous complexes (Watson, 1975).

The majority of the Lewisian rocks are Archaean and underwent deep-seated metamorphism at about 2800-2700 Ma. They subsequently suffered inhomogeneous tectonic and metamorphic modifications which continued intermittently until about 1800 Ma. The emplacement of a suite of tholeiitic dolerite dykes (the Scourie dyke swarm) provided time-markers which have been used to separate early (Scourian) and later (Laxfordian) tectonic episodes (Sutton and Watson, 1951). The principal structures built up during the initial (2800-2700 Ma) Scourian stages define a crude NNE grain. Later Scourian and Laxfordian events led to the formation of a system of NW and NE ductile shear-zones and to the development of several areas heavily veined by potassic pegmatites. No significant activity followed the emplacement of late Laxfordian pegmatites at about 1800 Ma on the craton.

Lewisian gneisses essentially similar to those of the craton, and in some areas yielding isochron ages of 2700 Ma, extend into the Caledonian orogen for 100-150 km from the western orogenic front. They are seen unconformably below the older units of the Moine Supergroup in the Northern Highlands of Scotland and appear elsewhere in tectonic contact with Dalradian rocks (Shetland, western Ireland). Granulites possibly derived from a Lewisian type of basement have been found as inclusions in volcanic vents in the Midland Valley of Scotland (Upton *et al.*, 1976) and central Ireland (Strogen, 1974) and granulites are thought, on geophysical evidence, to underlie much of the Southern Uplands of Scotland (Powell, 1971). These somewhat ambiguous indications suggest that the metamorphic (northwestern) Caledonides are largely floored by a basement of deep-seated Archaean to early Proterozoic rocks. The older members of the Moine Supergroup in the Northern Highlands, which rest unconformably on the basement and which appear to have accumulated before 1000 Ma (Brook, Brewer and Powell, 1976) together with some isolated groups in the Ox Mountains and elsewhere in Ireland may be tentatively regarded as a younger basement unit; as these rocks appear to have been tectonically and metamorphically modified by 1000 Ma, it follows that Grenville activity not recorded in the craton (*see* above) took place in the northwestern part of the future Caledonian orogen.

IGCP Project 27, British Contribution No. 1B.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

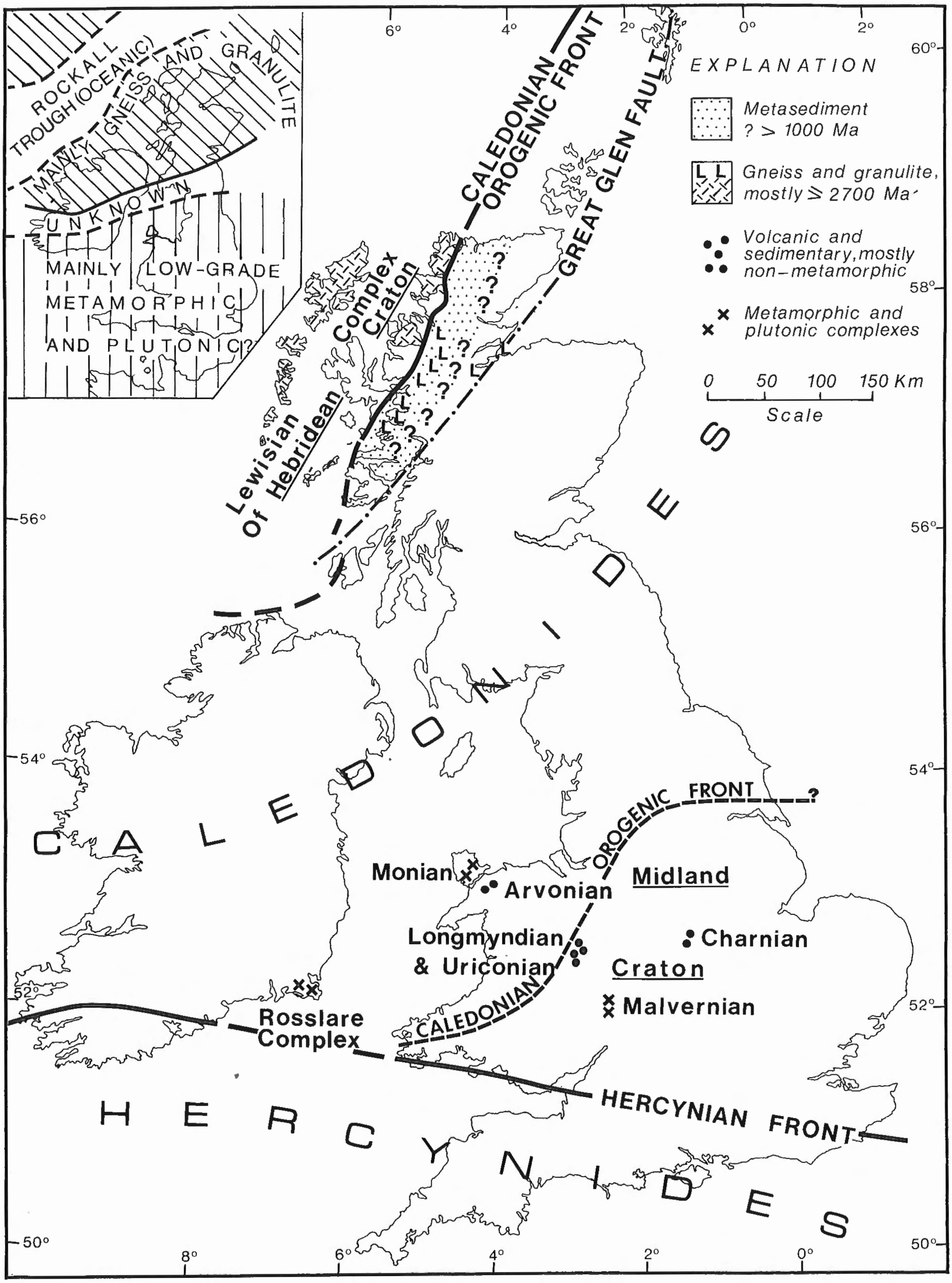


Figure 12.1. Diagrammatic map showing the principal outcrops of basement in the Caledonides and Hebridean craton. Inset, inferred extent of the two contrasted basement domains.

In the present context, the important features of the basement in the northwestern domain are (a) the general predominance of Na_2O over K_2O, the strong depletion in K, Rb, U and other incompatible elements shown by Lewisian granulites and the general absence of mineralization, (b) the dry and refractory nature of the deep-seated gneisses and (c) the occurrence of a grid of deep dislocations. The response of the basement during the Caledonian cycle suggests that old dislocations were reactivated. Most basement units take the form of tectonically defined wedges or slices; mantled gneiss domes produced by the regeneration of buoyant granites are absent.

PRECAMBRIAN COVER ON THE LEWISIAN BASEMENT

The basement of the Hebridean craton is overlain unconformably by two groups of Precambrian arkoses, sandstones and siltstones (the Stoer and Torridon groups, ≈1000 Ma and 800 Ma, known collectively as the Torridonian). These divisions are in turn overlain by transgressive Cambro-Ordovician quartzites, siltstones and dolomites. Although minor unconformities separate successive divisions, none have suffered severe deformation or metamorphism and all can be assigned to the cratonic cover; all are folded and disrupted at the Caledonian orogenic front. In the adjacent part of the orogen, the lowermost Moinian rocks appear as a thick sequence of metapsammites and metapelites interfolded and tectonically interleaved with Lewisian gneisses. The dating of these rocks (*see* this volume, Article 13, p. 79) suggests that the younger Moinian and Dalradian cover-units may rest unconformably on them but no such relationship has yet been confirmed.

THE MIDLAND CRATON AND SOUTHEASTERN CALEDONIDES

The basement of the southeastern domain is poorly exposed and little known. Exposures in a few localities in northern Wales, southeast Ireland, the Channel Islands and the English Midlands, with evidence supplied by geophysical surveys and by the nature of clastic fragments in younger formations suggest that much of the domain is underlain by granitic and associated rocks or by metamorphic complexes of medium grade (Malvernian, Monian, Rosslare Complex etc. cf. Dunning and Max, 1975). There are no indications of the presence of granulites or gneisses formed at depth. Some parts of this basement appear from preliminary isotopic data to be Archaean, but most remain undated. Aeromagnetic and gravity maps suggest that the basement in England is traversed by large northwesterly lineaments.

Assemblages of mainly acid lavas, pyroclastics, sub-volcanic intrusions and clastic sediments which are non-metamorphic and are severely disturbed only in the vicinity of large dislocations appear widely beneath the Palaeozoic cover of the Midland craton and Welsh basin. These rocks (Uriconian, Charnian, Pebidian, Arvonian etc.) have yielded minimum isotopic ages <700 Ma and some carry Vendian or very late Proterozoic fossils (Downie, 1975). They are broadly comparable in character and age with assemblages which underlie the Avalon platform at the southeastern margin of the Appalachians in Newfoundland and which extend thence as far as Massachusetts (e.g. Rast, O'Brien and Wardley, 1976). They are often assigned to late stages of the Cadomian orogeny and form a local basement to the Lower Palaeozoic successions in the eastern Caledonides. Their time of accumulation, however, apparently overlapped with that of the Dalradian Supergroup which represents an early cover-unit in the northwestern part of the Caledonides. The basement-cover junction therefore does not, on a regional scale, correspond to a time-plane.

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IUGS UNESCO IGCP PROJECT 27 CALEDONIDE OROGEN

Anthony L. Harris

Geology Department
The University
Liverpool, England

Michael R.W. Johnson

Grant Institute of Geology
University of Edinburgh
West Mains Road
Edinburgh, Scotland

Derek Powell

Department of Geology
Bedford College
Regent's Park
London NW1 4NS, England

1. MOINE ROCKS OF BRITAIN

Derek Powell

Within Britain Moine rocks crop out only in the Scottish Highlands. They comprise an ill-defined assemblage of Proterozoic metamorphic rocks of largely sedimentary origin. Despite having the largest outcrop extent of all the Precambrian rock groups in northern Britain they are the least well-understood in terms of their age, correlatives and tectono-metamorphic history.

Moine rocks *sensu stricto*, are bounded to the northwest by the Moine Thrust Zone, to the southeast by the Great Glen Fault; they occupy what may be called the Northern Highland Block. The psammitic Central Highland Granulites, which are commonly accepted as equivalents of the Moine *s.s.*, lie to the southeast of the Great Glen Fault beneath the late Proterozoic to Lower Palaeozoic, Dalradian Supergroup.

Summaries of the nature and constitution of the Moine rocks *s.s.* can be found in Johnstone (1975) and Johnstone, Smith & Harris (1969) who divided the Moine *s.s.* into three tectonic-stratigraphic divisions - the Morar, Glenfinnan and Locheil; relationships between these divisions, which are separated by major tectonic discontinuities, are, however, uncertain. Details of the stratigraphy of parts of the Moine *s.s.* are given in Ramsay & Spring (1962), Powell (1964), Dalziel (1966), Tanner, Johnstone, Smith & Harris (1970) and Powell (1974). Information about the Central Highland "Moine" is summarized by Johnstone (1966).

The Moine rocks *s.s.* lie, in places, unconformably upon Lewisian gneisses (Clough; *in* Peach *et al.*, 1910; Ramsay, 1958) which appear to be of Scourian Type (Watson, 1975) and are thus older than c. 2200 Ma. The youngest dates so far, from Lewisian inliers, are K-Ar minimum dates of 2200 - 1500 Ma (Watson, 1975). The Moine *s.s.* is overlain by Middle Old Red Sandstone (ORS) sediments and is intruded by the Carn Chuinneag Granite which has yielded an Rb-Sr isochron suggesting an age of emplacement of 560 ± 10 Ma. Radiometric ages from metamorphic rocks and minerals of the Moine sequence range from 1050 to 320 Ma (Long & Lambert, 1963; Dewey & Pankhurst, 1970; van Breemen, Pigeon & Johnson, 1974; Brook, Powell & Brewer, 1976) a spread which reflects their long and complex history.

As a whole the Moine rocks of Scotland have been considered as an older part of the Dalradian sedimentary succession because the Central Highland Moine passes up without a break into the younger supergroup, and no unconformity has been identified within the Moine. In consequence, all orogenic events affecting the Moine have been regarded as Caledonian (Ramsay, 1963; Johnson & Shepherd, 1970; Dewey & Pankhurst, 1970). An alternative view is that the Moine *s.s.* constitutes a much older sedimentary sequence which suffered Precambrian (Morarian - c. 730 Ma) orogenic activity overprinted by Caledonian (Long & Lambert, 1963; Powell, 1974; van Breemen *et al.*, 1974). Recent work has identified rocks of Grenville age within the Moine rocks *s.s.*, suggesting that the older orogenic events belong to the Grenville orogenic episode (Brook *et al.*, 1976). It thus appears likely that at least part of the Moine *s.s.* belongs to an older sedimentary sequence than the Central Highland Moine (Powell, 1974; Brook *et al.*, 1976) since it would appear unlikely that the Central Highland Moine can be very much older than 668 ± 23 Ma (Johnstone, 1975; Pringle, 1972).

Many workers have assumed correlation of the Torridonian sediments of the NW Caledonian foreland with the Moine (Kennedy, 1951; Sutton & Watson, 1964; Dewey & Pankhurst, 1970). Others have correlated the older, Stoer Group (Stewart, 1975) of the Torridonian, with the Moine *s.s.*, and the overlying Torridon Group (Stewart, 1975) with an undefined Upper Moine (Garson & Plant, 1973). In view of the ages for diagenesis of the Stoer Group at c. 995 Ma and of the Torridon Group at c. 810 Ma (recalculated from Moorbath, 1969) and the suggested age of sedimentation of the Moine *s.s.*, between 1250 and 1050 Ma (Brook *et al.*, 1976) it would appear that such correlations are not justified.

The tectonic history of Moine rocks of the Northern Highland Block has been investigated in a number of isolated regions (Ramsay, 1958, 1963; Brown *et al.*, 1970; Soper, 1971; Tanner, 1971; Tobisch *et al.*, 1970; and Powell 1974). Though there is agreement that polyphase deformation has occurred and attempts have been made to correlate deformation sequences (Tanner, 1971; Tobisch *et al.*, 1970; Powell, 1974) a complete tectonic pattern has not emerged. An Alpine-type major structure has been recognized in part of the southwestern Moine (Ramsay, 1963; Powell, 1974) but elsewhere the regional structure is unknown or not well understood. In some

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Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

places slivers of Lewisian basement appear to have been brought into high levels of the Moine cover along syn-metamorphic slide zones (Tanner, 1971; Tanner *et al.*, 1970).

The metamorphic history of the Moine sequence *s.s.* is polyphasal. In general terms the grade of metamorphism rises from west to east into a belt of migmatitic gneisses (Kennedy, 1949). Sillimanite, staurolite and kyanite have been recorded but there is no evidence that the rocks reached granulite facies. In detail the pattern of metamorphism is complex (Dalziel & Brown, 1965; Johnstone *et al.*, 1969; Soper, 1971; Powell, 1974; Winchester, 1971) but there is some evidence for two overlapping major episodes of metamorphism (Long & Lambert, 1963; Powell, 1974; Winchester, 1974; van Breemen *et al.*, 1974). Granitic and other gneisses are developed some of which appear to be products of the early metamorphism. Pegmatites belonging to both phases of metamorphism form, together with gneisses, a so-called Injection Belt (Phemister, 1960).

Recent isotope studies support the separation of both metamorphic and tectonic events into an earlier Precambrian group, c. 730 Ma (van Breemen *et al.*, 1974) or c. 1050 (Brook *et al.*, 1976), and a Caledonian group c. 560 - 320 Ma. The precise constitution and nature of these two groups is, however, debated (van Breemen *et al.*, 1974; Powell, 1974).

Plate tectonic models of the evolution of the rocks of Scotland during late Precambrian and early Palaeozoic time propose inclusion of the Moine as either an older part of the Dalradian sedimentary prism accumulated on the edge of the North American - Greenland continent, which subsequently became involved in Caledonian orogenic activity (Dewey & Pankhurst, 1970); as pre-Dalradian sediments which were deformed and metamorphosed during the closure of a Precambrian ocean before the rifting which gave rise to Iapetus (Dewey & Kidd, 1974); or as late Precambrian sediments which became involved in late Precambrian subduction during the closure of Iapetus (Garson & Plant, 1973). The first of these models does not account for the antiquity of Moine sedimentation or its early phase of deformation and metamorphism. The second may be correct, but closure of a Precambrian ocean or small ocean basin is likely to have occurred at least 480 Ma earlier than subsequent rifting. The third model depends upon interpretation of a zone of ultrabasic rocks in the Moine of the N. Highlands as a subduction zone, an interpretation which has been strongly criticized (Johnson, 1975; Moorhouse, 1976).

Whether the Moine rocks *s.s.* and the early tectono-metamorphic events which affected them are of Grenville age remains to be proved. Further investigations into the stratigraphy, tectonic pattern and sequence of tectono-metamorphic events are evidently necessary.

2. THE MOINE THRUST BELT

Michael R.W. Johnson

The Moine thrust belt is a narrow zone of thrust faults, more than 250 km long and up to 18 km wide, dividing the Moine metamorphic block from the so-called Caledonian Foreland. The Moine thrust, the most easterly thrust in the belt, carries Moine and Lewisian rocks with small 'parcels' of Torridonian and Cambro-Ordovician rocks. The thickness of the nappe above the Moine thrust is unknown but probably large. Although a post-early Ordovician age for the thrust belt is highly likely (Johnson & Shepherd, 1970) the lower age limit is uncertain. Polyphase deformations (Christie, 1963; Johnson, 1957, 1960; Barber, 1965; Soper & Wilkinson, 1975), involving at least four episodes and repeated displacement along individual thrusts, may indicate prolonged (? until early Devonian times) history. The earliest structural event consists of ductile flow in shear belts, forming mylonite zones up to 800 m thick, and recumbent folds (Johnson, 1960). Later "brittle" clean-cut thrusting has distrupted these early features.

Apart from the time-span perhaps the most controversial aspect of the thrust belt concerns the amount and direction of displacement. Minimum estimates (c. 20 km) for displacement along the major thrusts seem rather low when compared to the estimates made for comparable thrusts in other orogenic belts. As regards the displacement vector, recent work on the deformed Cambrian Pipe Rock (McLeish, 1971; Wilkinson *et al.*, 1975) confirms the widely held view that the sense of translation on thrusts was from east to west. It also has provided useful calculations of the shear strain within the early mylonite belts. Christie (1963) has challenged the orthodox view and has proposed considerable strike-slip movement along the Moint thrust, basing his argument on the transverse linear fabrics which are ubiquitous in the higher structural levels in the thrust belt and in the near-thrust Moines. Alternative interpretations of the transverse structures have been proposed (e.g. Johnson, 1965) and current work is aimed at solving this problem. The transverse structures are controversial in other respects, notably their possible regional extent and correlation with structures in the central part of the Moines. Although many workers have suggested that such correlations are reliable, these are open to question now that the validity of a Precambrian orogenic event in the Moines is gaining acceptance (Powell, 1974) (*see* Powell, this article, above).

3. DALRADIAN ROCKS OF BRITAIN

Anthony L. Harris

The sedimentary passage from the Central Highland Moine to Dalradian is marked by a diversification of rock type consequent on improved sorting of the sediments. Thus the transitional rocks are interbedded quartzites and pelites rather than the uniform micaceous psammites of the Moine. These transitional rocks were formally included with the Dalradian Supergroup by Harris & Pitcher (1975) and are the oldest of several lithostratigraphic subgroups into which the Dalradian has been divided. The subgroups, which are distributed into three groups, span the Cambrian-Precambrian boundary (Downie *et al.*, 1971). The base of the Argyll (Middle Dalradian) Group is marked by the widespread occurrence of the Varangian Portaskaig Tillite, while the limestones and associated volcanic rocks of the Lower Cambrian Tayvallich Subgroup occur at the top of the Argyll Group. Sandstones, shales and volcanic rocks known collectively as the Highland Border Series are included with the Southern Highland (Upper Dalradian) Group as are the fossiliferous Macduff Slates thought to be Lower Ordovician (Downie *et al.*, 1971). The sedimentary history of the supergroup is marked by an increasing maturity of the basin of deposition from shallow-water limestones, pelites and quartzites to turbidites. From early Cambrian times sedimentation was accompanied by widespread basic vulcanicity. It has been inferred that the Cambrian quartzites and limestones of the Durness succession, which are preserved on the Caledonian foreland of NW Scotland, may be related to the upper parts of the Argyll Group. Such shelf deposits may have provided a source of clastic siliceous and calcareous material very early in the

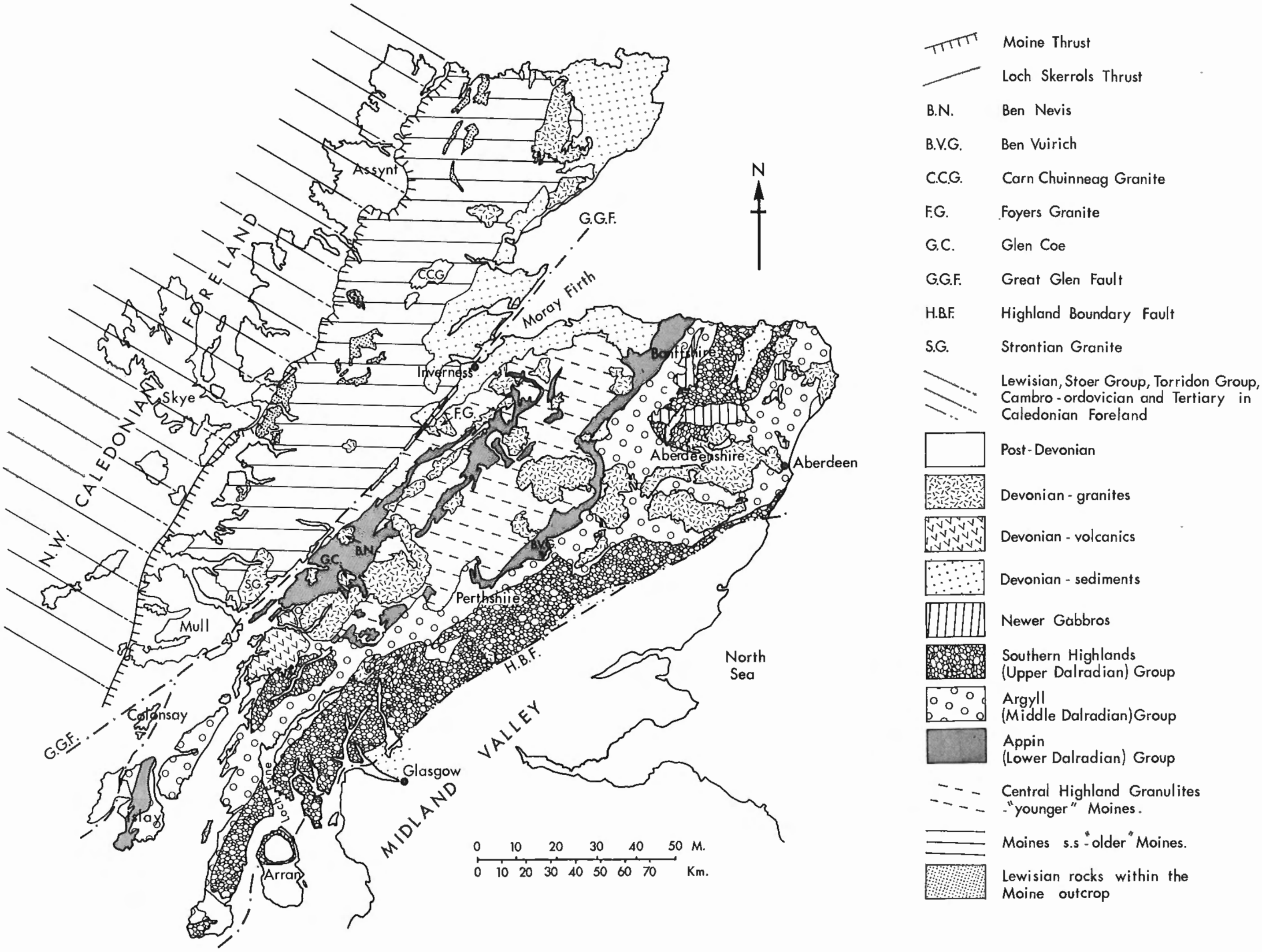

Figure 13.1. Lithostratigraphic map of the Scottish Metamorphic Caledonides

Cambrian, but while deposition of carbonates continued in NW Scotland until the Ordovician, from early Cambrian onwards Dalradian rocks are largely clastic.

Both Central Highland Moine and Dalradian rocks underwent deformation and regional metamorphism together and polyphase deformation has been recorded in many localities (e.g. Roberts, 1974). There is no indication that the Moine rocks of the Central Highlands underwent any Precambrian orogenic activity.

In the SW Highlands the regional structure is dominated by an anticlinal gravity nappe (the Tay Nappe) the upward, SE-facing root of which can be traced as the Ardrishaig Anticline through the Loch Awe-Loch Fyne district of Argyllshire (Roberts, 1974). Its downward-facing closure (Aberfoyle 'Anticline') can be traced along the adjacent Highland Border to the southeast, (Shackleton, 1958). Northwest from the Ardrishaig Anticline the axial planes of major folds of comparable age fan through the vertical (Loch Awe and Tayvallich synclines) and come to dip southeast while facing up to the northwest (Islay Anticline). Traced to the east into Perthshire this 'mushroom' form of the Tay Nappe complex becomes less well-established because a major closure comparable to the Ardrishaig Anticline is absent, although, as in the SW Highlands, large areas of horizontal inverted strata thought to lie on the lower limb of the nappe are recorded (e.g. Loch Tay Inversion). The geometry of the major structure in Perthshire is further complicated by the existence of the Boundary Slide (Rast, 1958) which separates the Dalradian rocks (above) from the Central Highland Moine (below). Harris, Bradbury & McGonigal (1976) have suggested that successively lower levels in the nappe are characterized by increasing bulk strain and complexity of deformation, the lower levels being characterized by at least three major episodes. Bradbury *et al.* (1976) have shown that in Perthshire the 514 $^{+6}/_{-7}$ Ma U-Pb zircon age on the synorogenic Ben Vuirich granite (Pankhurst & Pidgeon, 1976) dates a time in the interval between the second and third deformation episodes, and is broadly coeval with migmatization in that area. Right-way-up Dalradian rocks in NE Scotland probably lie on the upper limb of the continuation of the Tay Nappe into Banffshire and Aberdeenshire. Here the third episode of deformation is preceded by the 501 ± 17 Ma emplacement of the 'Newer' Gabbros (Pankhurst, 1970).

The SE Highlands (Barrow, 1893) constitute the type area for Miyashiro's (1961) kyanite-sillimanite facies series, while the Buchan area (Read, 1952) is the type area of Miyashiro's (*op. cit.*) intermediate low-pressure facies series. In both areas it is probable that the presence of sillimanite and associated migmatites is due to a thermal overprint on rocks already at kyanite or andalusite grade (Chinner, 1966). The thermal overprint is probably related to the emplacement of the 'Newer' Gabbros at 501 ± 17 Ma (Pankhurst, 1970; Ashworth, 1975). Chinner (1966) has traced the boundary between the Barrovian and Buchan metamorphic areas as the kyanite-andalusite isograd. Much of the Dalradian - Central Highland Moine area lies in the epidote amphibolite and amphibolite facies of metamorphism but lower facies are encountered in peripheral areas such as the Highland Border and in major depressions of regional plunge such as Banffshire.

4. THE GREAT GLEN FAULT

Derek Powell

The Great Glen Fault Zone is a northeasterly trending, essentially vertical, feature which traverses some 160 km of the Scottish Highlands. Its submarine extension has been traced northeastwards to the Shetland Islands where it is represented by the Walls Boundary and Nesting Fault Systems (Flinn, 1961, 1969, 1976). To the southwest, the main fault zone appears to pass north of the island of Colonsay (McQuillin & Binns, 1973) whilst a southern branch may pass through Islay to connect with the Leannan Fault in Ireland (Pitcher *et al.*, 1964; Dobson & Evans, 1974). From north of Colonsay the main fault may continue towards the west or southwest for at least 160 km (Bailey *et al.*, 1975). Its total length thus probably exceeds 720 km. Correlation of the fault zone with similar structures in Newfoundland has been discussed by Pitcher (1969), Phillips *et al.* (1969), Kay (1967), Webb (1969) and Wilson (1962).

Kennedy, in 1946, was the first to postulate horizontal movement along the fault zone considering such to have taken place in pre-Carboniferous time. Since Kennedy's suggestion of a 104 km sinistral displacement, sinistral shifts of 133 km (Holgate, 1969), 160 km (Winchester, 1973) and dextral movements of 29 km (Holgate, 1969), 120 km (Garson & Plant, 1972), and between 64 and 96 km (Flinn, 1969, 1976) have been postulated. It is now generally agreed that movement took place in post-Lower Cretaceous time though earlier movement prior to middle Old Red Sandstone deposition has been suggested (Garson & Plant, 1972; *see* also Flinn, 1969). Earth tremors related to the fault zone are recorded in historical time.

The matching of the Strontian and Foyers granites across the fault zone, cited by Kennedy as evidence for lateral slip, has been questioned (Marston, 1970; Munro, 1965). Kennedy's matching of regional metamorphic zones has however been supported, though modified, by Winchester (1973). Assuming the age of metamorphism giving rise to the zonal pattern in the Northern Highland and the Grampian Highland blocks on either side of the Great Glen Fault Zone, to be between 530 and 480 Ma (Dewey & Pankhurst, 1970; Flinn & Pringle, 1976) and accepting the evidence for dextral movements of at least 64 km since Devonian times (Flinn, 1969) it follows that a sinistral shift of ca 224 km occurred some time between 530-480 Ma and 395 Ma. There are, however, doubts as to whether or not the tectono-metamorphic events on either side of the fault are wholly of the same age. Isotopic evidence suggests the influence of Precambrian tectono-metamorphic activity in the Northern Highland Block at between 1050 and 730 Ma overprinted by Caledonian at 560 to 450 Ma (van Breemen *et al.*, 1974; Brook *et al.*, 1976). In contrast orogenic activity in the Grampian Block seems to have occurred between 530 and 480 Ma (Dewey & Pankhurst, 1970; Flinn & Pringle, 1976).

Such differences between the Northern and Grampian blocks together with those of tectonic style may indicate either even greater horizontal displacements than those already proposed, considerable vertical movements, or that the fault zone is a more fundamental structure than just a transcurrent fault.

5. THE HIGHLAND BORDER

Anthony L. Harris

The Highland Border zone is marked by a major commonly reversed fault - the Highland Boundary Fault - the trace of which is locally picked out by bodies of serpentinite and by the cherts, black shales and spilitic rocks of the Highland Border Series. Regionally the fault separates Dalradian rocks from Upper Palaeozoic rocks occupying the Midland Valley of Scotland. Locally the fault is overstepped by Devonian sediments and volcanic rocks and in such areas the course of the fault is commonly marked by a monoclinal change in their dip. The coincidence of the fault with lenticles of Highland Border Series, with downward-facing Dalradian structures and with Dalradian rocks of low metamorphic grade suggests that its history must date back much further than the post-Lower Devonian episodes of displacement. Its earliest movements probably coincided with the downbending of recumbent Dalradian structures into a downward-facing attitude and the downbending of initially shallowly dipping isograd surfaces into steep attitudes. It is possible that, between the earliest traced displacements in the Highland Border zone and the episodes which displaced and influenced the deposition of the Lower Devonian strata, movements on the fault were normal. Thus high grade, structurally low Dalradian rocks may underlie the Midland Valley. Certainly the high-grade Dalradian rocks of Connemara lie to the south of the supposed extension of the fault into western Ireland.

6. POST-OROGENIC FORMATIONS

Michael R.W. Johnson

Apart from the insignificant amounts of Late Palaeozoic, Mesozoic and Tertiary rocks the most important post-metamorphic formations in the Scottish Highland region are Devonian. During the Devonian the crust underwent increasing stabilization linked with the ending of subduction and the closure of Iapetus. The Devonian is represented by clastics, volcanics, granite bodies and minor intrusions. Uplift and erosion of the Moine and Dalradian led to the formation of intermontane basins in which molasse-type Devonian rocks (Old Red Sandstone) accumulated. Calc-alkaline and acid volcanics were poured out from volcanic centres like that at Glen Coe. The volcanics and the post-tectonic granites, like the ring-complex at Ben Nevis and the numerous large plutons, indicate that thermal activity persisted in the Highlands after the uplift.

Several important faults were initiated or active during Devonian times. Downwarping led to formation of the Midland Valley graben and the Highland Boundary Fault (*see* Harris, this article), which now forms the southern margin of the Dalradian, and was active in the Devonian. Major folds have affected the Lower Devonian rocks in the northern part of the Midland Valley.

The Great Glen Fault (*see* Powell, this article) and associated faults represent a wrench fault regime which was active in Mid-Devonian times and subsequently. Controversy continues on whether these are sinistral or dextral faults.

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CALEDONIDES OF THE MIDLAND VALLEY OF SCOTLAND

Bernard E. Leake

Department of Geology
The University, Glasgow
Glasgow, G12 8QQ, Scotland

The Midland Valley lies between the Highland Boundary Fault (HBF) and the Southern Uplands Fault (SUF) (this volume, article 11, Figure 11.1). A summary of the geology is given in MacGregor & MacGregor (1948) and George (1965). Along the HBF is a sliver of serpentine and Ordovician (Arenig or Llanvirn?) black shales and radiolarian cherts associated with basaltic pillow lavas. The age of initiation of the HBF is uncertain; it could be as early as late Cambrian - early Ordovician if the remarkable parallelism and proximity of the hinge line of the Dalradian Tay nappe (the downward facing Aberfoyle anticline) to the HBF is taken as evidence of a genetic connection between the Tay nappe and HBF. The fault is generally considered to be pre-Lower Devonian as the Lower Old Red Sandstone (ORS) is reported (George, 1963) to cross the fault without much change of thickness though the lack of a complete succession to the north of the HBF makes correlation difficult. Whereas the main movement on the HBF was a downthrow as a reversed fault to the southeast, later movement (as a normal fault) with a downthrow to the northwest, has preserved Upper ORS and Carboniferous rocks, including Coal Measures in Kintyre, to the north of the fault. It is critical to note that the conglomerates of the Lower ORS are largely formed of well-rounded quartzite boulders derived from Dalradian formations now removed by erosion and, indeed, in the region of Loch Lomond they are sharply upturned against the HBF.

The Upper ORS and succeeding Carboniferous are nearly horizontal formations and occupy erosional depressions in what is virtually the present Dalradian topography. Moreover, the pebble conglomerates of the Upper ORS contain fragments of the immediately adjacent Dalradians - 'schistose grits' and vein-quartz.

The HBF can be traced with some ambiguity into Ireland (Max and Riddihough, 1975) whereas the SUF is more confidently correlated (Leake, 1963; Max & Ryan, 1975). The evidence from Ireland indicates that the Dalradian rocks underlie the Midland Valley in that Dalradian rocks occur south of the probable continuations of the HBF in Ireland.

The Valley was probably tectonically distinguished from the rocks of the Southern Uplands by Silurian times as Silurian successions differ from those to the south in the Southern Uplands or at Girvan (this volume, article 15) in the SW part of the Midland Valley. The exact age of initiation of the SUF is unknown but late Silurian conglomerate wedges from the south suggest that it existed in the late Silurian, though George (1965) has argued for an early Devonian initiation. The oldest known exposed rocks within the Midland Valley are the Arenig rocks of Girvan that are unconformably overlain by Caradoc rocks but Upton *et al.* (1976) have reported garnetiferous quartz feldspathic gneiss fragments in Carboniferous volcanoes that indicate the presence of a granulite basement; presumably Precambrian.

The Lower ORS within the Valley rests with angular and non-angular unconformity upon the mudstones and shales of Wenlock and Ludlow age being markedly unconformable upon folded Silurian near the marginal faults. The Lower ORS is formed of thick sandstones and conglomerates interbedded with volcanic rocks, olivine basalts, andesites, rhyolites, dacites, agglomerates and tuffs, the whole series being folded and then unconformably overlain by Upper ORS that passes conformably into the basal Carboniferous. On either side of the Valley at least 3000 m of Lower ORS is believed to have been eroded before the deposition of the Upper ORS, the rifting having preserved the Lower ORS preferentially within the valley (George, 1963). The Carboniferous Calciferous Sandstone Group (and oil shales) were followed by the Carboniferous Limestone Group (but with few limestones), 'Millstone Grit' and the Coal Measures. Thick sequences (in places exceeding 1000 m), mainly of alkali basalts occur in the Calciferous Sandstone Group in the west, but extend into Carboniferous Limestone Group in the east - locally volcanics are found in the Millstone Grit (e.g. Ayrshire). Mugearites and trachytes are less common, but plugs of basalt and basaltic agglomerate are very numerous. More strongly alkaline rocks, teschenites, etc. form sills and occasional dykes and there are several sheets and 'laccoliths' of phonolite or trachyte.

The main structure of the Midland Valley is a shallow syncline, which is gently eastward plunging in the west, while an approximately north-south directed arch separates the eastern and western groups of upper Carboniferous basins. In the southwest the Ayrshire basin contains red beds which have been ascribed to the Permian, while in Arran the Carboniferous is succeeded by considerable thicknesses of undoubted Permian and Trias.

An extensive suite of large sills (up to 300 m thick) and broad east-west trending dykes, all of very uniform quartz dolerite (tholeiitic) composition are emplaced within Carboniferous (and older formations) and are inferred to be of early Permian age.

The major Caledonide problems are structural, magmatic and stratigraphical. What was the age and the reason for the initiation of the boundary fractures and how does the Valley fit into any plate tectonic model? Is it for example, an intracratonic basin or a back arc basin? Is the Highland Border Ordovician an ophiolite sequence and if so how can the Dalradian sequence occur south of the HBF in Ireland? What were the nature of the first fault movements and how do they relate to the enormous thickening of the Lower ORS from SW (c. 3000 m) to NE (c. 9000 m) (Armstrong & Peterson, 1970) on the north side of the valley? Differential subsidence within a basin being pulled apart? How are the voluminous Lower ORS volcanic rocks explained - intraplate volcanoes, perhaps connected with rifting, or plate-edge magmatism connected with a subduction zone? Why are there important granitic plutons of Devonian age both to the north and south but none in the Valley? Are there thick lower Palaeozoic and Dalradian sequences under the exposed rocks of the Valley and if not is this due to subsequent removal or were they never deposited? What is the deep structure of this part of Scotland and does it confirm or deny the postulated subduction zone

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Geol. Surv. Canada, Paper 78-13 (1978).

that might have passed through the miniscule blue schist outcrops at Girvan? The results of the recent north-south seismic profile through Britain (LISP) should give most valuable information on these problems. What is the age of the basement revealed in xenolithic fragments; Lewisian or younger and why have no Dalradian fragments been identified?

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15. THE PARATECTONIC CALEDONIDES OF MAINLAND BRITAIN

Gilbert Kelling

Department of Geology
University of Keele
Keele, Staffordshire, ST5 5BG, England

This article deals with the Caledonides of three regions within mainland Britain (this volume, article 11, Fig. 11.1):

1. Southern Uplands of Scotland;
2. The Lake District and the Isle of Man;
3. Wales, the Welsh Borderland and Central England.

1. SOUTHERN UPLANDS OF SCOTLAND

In this region (Fig. 15.1) the Caledonides are represented by Ordovician and Silurian rocks which attain a maximum aggregate thickness of around 9 km (Walton, 1965). These rocks are virtually unmetamorphosed except for local developments of greenschist facies metasediments (Weir, 1974) and the thermal effects associated with later intrusions. The rocks are disposed in a series of structures trending ENE-WSW. With the exception of the Girvan sequences, the region is bounded to the north by the Southern Uplands Fault, a complex fracture which effectively drops down Upper Palaeozoic rocks to the northwest; in the south and east, the region is bounded by Devonian and Carboniferous rocks which are unconformable on the Lower Palaeozoic sequences. Geographically, it is convenient to refer to four areas (Peach and Horne, 1899). From north to south these comprise the Girvan area, the Northern Belt (between the Southern Uplands Fault and the Silurian boundary, the Central Belt (Silurian rocks with inliers of the Moffat Series shales) and the Southern Belt (made up of mainly younger Silurian rocks, south of the Moffat Series) (*see* Fig. 15.1).

The oldest known rocks (Arenig-?Llanvirn) occur in the Girvan area and doubtfully in one small inlier in the Northern Belt. They form a strongly deformed ophiolitic complex including pillow-lavas, serpentinites, cherts and black graptolitic shales, together with glaucophane schists and eclogites which may be pre-Arenig in age. Together with gabbroic and trondjhemitic intrusives, these rocks constitute the Ballantrae Igneous Complex of the Girvan region.

Although originally regarded as Arenig, the thin sequences of black, graptolitic mudstones, cherts and spilitic or andesitic tuffs and lavas which occur near the base of the thick greywacke-turbidite successions of the Northern Belt are now recognized as spanning the time-interval from ?Llanvirn to early Caradoc (Lamont and Lindstrom, 1957; Kelling, 1961; Bergstrom, 1970; Walton & Weir, 1974). However, rapid accumulation of coarse clastics commenced in Caradoc times and continued into the Wenlock. In the northwest (Girvan area) Caradoc-Ashgill platform sediments (including limestones) step northwards over the Ballantrae Igneous Complex (Williams, 1962), on to a landmass which has been termed Cockburnland (Walton, 1963; Dewey, 1971). Southwards, the platform facies passes rapidly into a thick pile of Caradoc-Ashgill greywackes and rudites, straddling the Southern Uplands Fault (Williams, 1962) and constituting the greater part of the Northern Belt successions (Kelling, 1961, 1962). These sediments appear to have been deposited in a deep marine trough or trench (Piper, 1972) and were probably formed in a series of overlapping submarine fans at the base of a southeast-facing submarine slope. Evidence of lateral input from the northwest is common in older sequences but the younger, more distal turbidites display more consistent axial transport, mainly towards the southwest (Walton, 1963, 1965; Kelling, 1962; Weir, 1973). To the southeast, in the Central Belt, the several kilometres of Late Ordovician greywackes are represented by the Moffat Series, a few tens of metres of graptolitic mudstones, with tuffaceous and clay bands, which appear to have accumulated on an 'axial rise' which permitted pelagic sedimentation to continue beyond the influence of the bottom-hugging turbidity currents (Kelling, 1964, p. 86; Walton, 1965).

The Silurian Birkhill Shales of the Moffat Series pass laterally into greywacke sequences both to the northwest and, at least

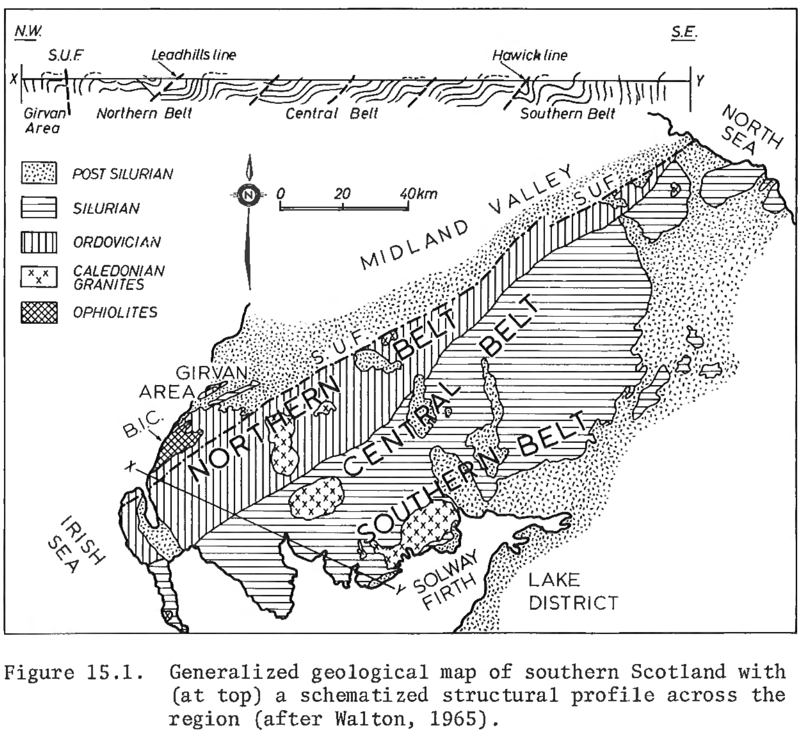

Figure 15.1. Generalized geological map of southern Scotland with (at top) a schematized structural profile across the region (after Walton, 1965).
B.I.C. - Ballantrae Igneous Complex
S.U.F. - Southern Uplands Fault

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Geol. Surv. Canada, Paper 78-13 (1978).

in part, to the southwest where rocks previously thought to be Wenlock have now been identified as Llandovery. This suggests a possible symmetry to the Southern Uplands basin with clastics derived from both northwest and southeast, at least for the Llandovery. Wenlock turbidites occupy most of the Southern Belt, while coeval platform sediments occupy a few small patches in the Girvan region.

Sediments of Ludlow age are absent from the Southern Uplands and are only doubtfully present in a few Lower Palaeozoic inliers in the southern part of the Midland Valley of Scotland, where they occur near the top of fluvio-deltaic redbed sequences of late Wenlock age, which pass down transitionally into early Wenlock-Llandovery turbidites, derived from the south (Walton, 1965; Rolfe, 1961). The implication is clear that, by mid-Silurian times, the Cockburnland landmass had expanded to include much of the former Northern Belt and Central Belt trough and rise and an independent, essentially shallow marine basin had come into existence within the Midland Valley as a precursor to the molassic Old Red Sandstone trough.

Structurally, the rocks of the Southern Uplands are disposed in a series of steep and flat-lying zones, each zone occupying several kilometres across the strike (Fig. 15.1). The steep zones consist of beds dipping at high angles and often overturned where folds tend to be restricted to small anticline-synclinal pairs (Craig & Walton, 1959; Walton, 1961, 1965; Kelling, 1961). The flat-lying zones take their name from the attitude of the 'faltenspiegel' (the tangent to the crests of the folds of one bed) which tends to be simple and near the horizontal, although the beds themselves are deformed into many asymmetrical or isoclinal folds. Axial planes generally dip southeastwards and the tendency to young northwestwards is offset by a series of major thrust faults which bring up older rocks to the north. Thus the inliers of the Moffat "Series" are brought into their present position by the major Ettrick Valley Thrust and the repeated outcrop of the graptolitic shales is brought about by schuppen associated with these major thrusts. Similarly it is thought that the Silurian boundary is a major thrust which brings Ordovician rocks up to the north of dominantly northwards-younging Silurian rocks. The structural configuration of the Lower Palaeozoic rocks at the northeastern end of the Southern Uplands appears to be substantially different to that described above (Shiells and Dearman, 1963) and shows some affinities with the tectonic history of the Upper Dalradian. In most parts of the Southern Uplands at least three phases of post-Caradoc, pre-Lower Old Red Sandstone tectonic deformation can be discerned but only the first two involve significant folding (Williams, 1959; Kelling, 1961; Walton, 1965; Rust, 1965; Weir, 1968).

While the structural style of the Lower Palaeozoic rocks in the Southern Uplands strongly suggests décollement, the nature of the pre-Arenig basement remains uncertain. Geophysical evidence (Powell, 1970, 1971) indicates a Lower Palaeozoic 'cover' depth of some 12 km and an underlying high-velocity basement of crystalline rocks, possibly composed of Lewisian-type metamorphics (Powell, 1971, p. 371). An alternative explanation for these geophysical observations is offered by Mitchell and McKerrow (1975), who invoke either a thick pile of folded turbidites, metamorphosed at depth, or an imbricate series of thrust-slices carrying Lake District sediments and volcanics and tectonically emplaced below the Southern Uplands turbidite complex.

The area has been affected by magmatism at different periods during the Lower Palaeozoic. The Caledonian orogeny was responsible not only for the Arenig 'spilitic' suite but also for some interbedded Caradoc spilites, andesites, keratophyres and ashes in the Northern Belt.

The abundance of fresh andesite clasts in the Late Ordovician (Kelling, 1962) to mid-Silurian (Ziegler and McKerrow, 1975) turbidites, ultimately derived from a northern landmass, demonstrates that andesitic vulcanism within and to the north of the Midland Valley probably was much more widespread and prolonged than is indicated by restricted present outcrops.

Major intrusions include granites such as the Loch Doon and Cairnsmore of Fleet masses as well as many smaller bodies. Minor intrusions comprise the 'porphyrites', which form dykes and small irregular sheets, many of which are affected by later tectonic episodes.

2. THE LAKE DISTRICT AND THE ISLE OF MAN

This region (this volume, article 11, Fig. 11.1) displays a thick sequence of mildly to unmetamorphosed sediments and volcanics spanning the time interval from late Cambrian to late Silurian, and surrounded by Carboniferous or younger rocks. Although no undoubted Precambrian rocks are known here, small outcrops at Ingleton, about 15 km to the southeast of the Lake District, include late Precambrian or Cambrian (O'Nions *et al.*, 1973) sediments strongly deformed, which may represent a northeasterly extension of the Irish Sea Horst.

The oldest rocks in this region are the Manx Group of the Isle of Man, a 7.5 km pile of slates and greywacke-turbidites of probable late Tremadoc or early Arenig age (Downie and Ford, 1966; Simpson, 1963). These sediments may be partly coeval with or transitional up into the Skiddaw Group of the Lake District, a 9.5 km thick sequence of mud-dominated turbidites of Arenig-Llanvirn age. These are succeeded, with apparent discordance (but *see* Moseley, 1975) by nearly 5 km of andesitic lavas and pyroclastics (Borrowdale Volcanic Group) of late Llandeilo - early Caradoc age (Moseley, 1964). The succeeding late Caradoc to Ludlow sequence rests unconformably on the underlying rocks. The late Ordovician sequence is relatively thin and dominantly argillaceous (with subordinate tuffs) but with some important calcareous units near the top. The conformably succeeding Silurian is thick (max. 4 km) and mainly of deep marine aspect, commencing with a group of graptolitic mudstones and shales which pass up into a thicker sequence of Ludlovian turbidite sandstones and calcareous siltstones. Palaeocurrent vectors from these late Silurian turbidites indicate dominant flow towards the south and southeast (Furness *et al.*, 1967; McCabe and Waugh, 1973) and suggest that these sediments may represent the advancing front of the flysch-wedge which had prograded southwards across the Southern Uplands of Scotland during the Silurian (Ziegler, 1970).

The Lower Palaeozoic rocks of this region appear to have been affected by three main phases of fold-deformation (Simpson, 1968):

(i) A late Llanvirn phase which produced a set of large folds trending north-northeast (Jeans, 1972), with accompanying cleavage and, in the Isle of Man, synkinematic chlorite-grade metamorphism;

(ii) A pre-Caradoc episode responsible for warping of the Borrowdale Group and the older sequences into broad, east-west trending folds (Simpson, 1968; Soper and Numan, 1974);

(iii) The end-Silurian deformation phase which established the dominant ENE-WSW trend of the major folds, cleavage and thrusts in these rocks and refolded some of the earlier structures. The emplacement of granites (e.g. the Shap mass) and less commonly, basic complexes (Carrock Fell) represents the final Caledonian event in the Lake District (early Devonian).

3. WALES, THE WELSH BORDERLAND AND CENTRAL ENGLAND

Here a maximum thickness of approximately 13 km of Lower Palaeozoic sedimentary and volcanic rocks, virtually unmetamorphosed, are disposed in a series of major folds of broad, relatively open style and general NE-SW trend, and are cut by major fractures of similar trend, which are best developed near the southeast and northwest borders of the regions.

Precambrian rocks outcrop sporadically on the flanks of these younger rocks: to the east, in Leicestershire; the Malvern Hills and the Borderland (Shropshire); to the southwest in Pembrokeshire (St. Davids); and to the northwest, in Anglesey and Caernarvonshire. In this northwestern area the Precambrian comprises essentially two groups of rocks: an older, thick (? 9 km) series of strongly deformed and differentially metamorphosed flyschoid metasediments and metavolcanics (Monian) which includes serpentinites and glaucophane schists and is unconformably overlain by an acidic lava and pyroclastic sequence (Arvonian) which appears to pass conformably up into the Cambrian sediments in more southeasterly outcrops (Shackleton, 1969; Wood, 1969).

The Precambrian rocks of central England are known from several isolated outcrops, the most important being in Leicestershire (Charnwood Forest, about 100 km northeast of the Malverns), and in the Malvern Hills of Worcestershire. Boreholes indicate that similar rocks are extensively developed at shallow depths beneath the Upper Palaeozoic/Mesozoic cover of the Midlands region (Wright, 1969). The Charnian of Leicestershire is a thick, strongly deformed series of mainly volcaniclastic sediments, with northwest-southeast trending structures, intruded by diorites and syntectonic syenites which yield late Precambrian - early Cambrian dates (Cribb, 1975). In the Malverns, the gneisses of the basement complex may represent the oldest known rocks of southern Britain and are succeeded unconformably by a late Precambrian sequence of acid to intermediate volcanics (Warren House Group).

In Shropshire, the Precambrian consists of a lower sequence of acid volcanics (Uriconian) overlain by nearly 9 km of very slightly metamorphosed flysch and molasse (Longmyndian) with at least two major unconformities represented in this sequence. These rocks were folded prior to early Cambrian times and are covered discordantly by Lower Cambrian orthoquartzites. However, radiometric dates recently obtained from the basal Longmyndian indicate an Eo-Cambrian age - an anomaly still to be resolved (Bath, 1974; Toghill, 1975). In the St. Davids district of southwest Wales only the lower, volcanic, Precambrian succession is preserved, again unconformably covered by Lower Cambrian sediments.

Within the Lower Palaeozoic terrain of Wales three main areas may be distinguished geologically (Fig. 15.2). From southeast to northwest these are:

(a) the Borderland-Pembrokeshire platform,
(b) the Central Wales basin, and
(c) the Harlech Dome-Snowdonia district (cf. Jones, 1938).

Area (a) yields relatively thin sequences of sediments which are consistently of shallow-marine aspect throughout the time-interval from Early Cambrian to Late Silurian and which show numerous internal time-breaks and discordances, the most widespread of which occurs at the base of the Upper Llandovery. Tuffs and lavas of mixed composition (spilites to rhyolites and keratophyres) attain significant thicknesses in the Llanvirn and Caradoc Series of Shropshire (Shelve district) and further to the southwest in the Builth and

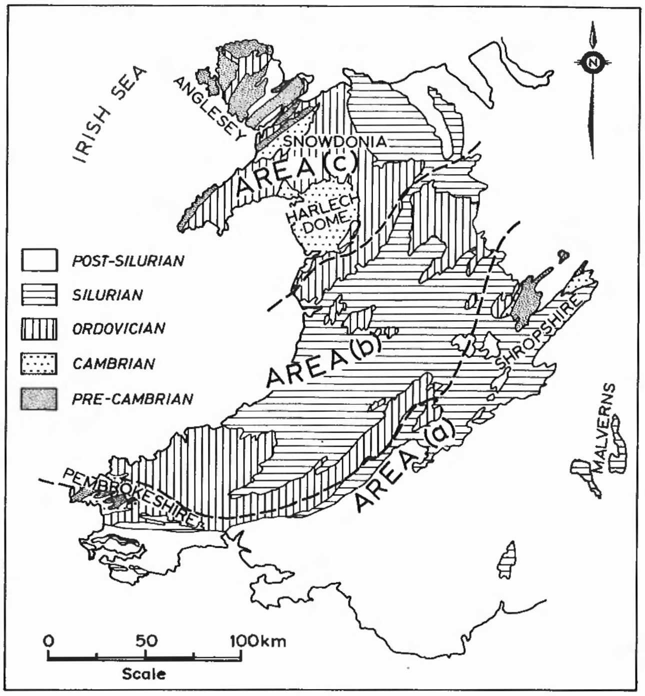

Figure 15.2. Generalized geological map of Wales and the Welsh Borderland indicating the approximate limits of the areas (a, b, c) described in the text.

Llanwrtyd areas of Breconshire. Basic volcanics of Llandovery age occur near Bristol (Tortworth) and in south Pembrokeshire (Ziegler *et al.*, 1969). These "shelf" deposits evidently were formed on the western margins of a stable platform cored by Precambrian crystalline rocks (the "Midlands Block").

Area (b) is characterized by thick (4-6 km) Ordovician and Silurian clastic sequences of "basinal" aspect, with restricted graptolitic faunas. Dark shales and localized thick submarine volcanics are typical of the rather attentuated Ordovician successions whereas the Silurian is marked by substantial thicknesses of turbidite sandstones and conglomerates (e.g. Aberystwyth Grits), mostly transported from the southwest (Wood and Smith, 1958). Deeper marine conditions thus seem to have prevailed in this region through much of Early Palaeozoic time (Ziegler, 1970).

Area (c) includes mainly Cambrian and Ordovician rocks which show marked and rapid lateral variations in facies and thickness which, together with the relative paucity of fossils, render correlation in this structurally complex area a matter of some difficulty, especially in the Lower Cambrian. The Cambrian sequence in the northwest (Caernarvonshire) commences with a group of shallow-marine conglomerates, sandstones and slates interbedded with ignimbrites which are disconformably succeeded by a group of turbidite sandstones which pass up into shallow-marine sandstones and shales with several internal time-breaks. To the southeast, in the Harlech Dome, the Cambrian is represented by a thick (4.5 km) sequence of poorly fossiliferous sandy turbidites passing up into a more argillaceous group of shallow-marine aspect (George, 1961; Crimes, 1970). The base of the Cambrian is not seen here but there do not appear to be any significant breaks in the sequence, in contrast to the Caernarvonshire Cambrian. Throughout area (c) the basal Ordovician (Arenig) is represented by a shallow-marine facies which oversteps northwestwards with increasing discordance across Cambrian and older rocks. This overstep emphasizes the existence of a structurally positive region, known as the Irish Sea Horst, which strongly influenced sedimentation in north Wales throughout much of Early Palaeozoic time. The later Ordovician history of northwest Wales is dominated by post-Llanvirnian uplift and the eruption of the great thicknesses of Caradocian shallow marine and subaerial acid volcanics which form the spectacular mountain scenery of Snowdonia. In northeast Wales (Denbighshire) an east-west trending turbidite trough came into existence during the Silurian (Cummins, 1957, 1959) and the northern flank of this basin is marked by spectacular slumping (Jones, 1937).

The Lower Palaeozoic rocks of Wales provide evidence of polyphase deformation (Dewey, 1969; Bates, 1974), the most important phases being end-Silurian. Mid- or Late Ordovician movements, responsible for widespread warping, particularly in North Wales, may have evoked a cover response from major basement fractures, inducing at least some of the anomalous fold and cleavage trends encountered in this region (Shackleton, 1954; Dewey, 1969). Traditionally, the major structures are a series of large, crenulated folds, trending mainly northeast-southwest and accompanied by subvertical faults of similar trend. However, northwest dipping thrusts are known to be an important element in Anglesey and northwest Wales (Shackleton, 1959; George, 1961) and recent work suggests that these thrust-features may have their counterparts on the southern and southeastern margins of the Welsh basin. In southwest Wales the Caledonian structures swing into an east-west alignment and here, too, sedimentation continued, virtually uninterrupted, into the Old Red Sandstone. Thus, in this region, there is no clear distinction between the belated deformative phases of the Caledonian and the premonitory pulses of the Hercynian orogeny.

Significantly, syn- or post-orogenic plutonism appears to be lacking in the Welsh region, although smaller acid plutons and basic intrusives form important contemporaneous elements in the Caradoc volcanics of north Wales.

A SUMMARY OF DEFORMATION HISTORY IN THE BRITISH PARATECTONIC CALEDONIDES

As stated earlier, the evolution of the British Caledonides involves a long and complex history of stratigraphic and structural events, manifested in the marked variations in rock-facies and thicknesses and the development of unconformities and other discordant junctions, which have been outlined above. Consideration of these effects clearly indicates that the British Caledonides must be considered in terms of a number of zones or regions with significantly differing sequences and developmental histories. Diastrophism occurs in various ways and at greatly varying rates but Dewey (1974) has argued that elucidation of the deformation-history is greatly enhanced if the assumption is adopted that thick terrestrial molasse sequences result from rapid uplift of adjacent "orogenically" deformed terrains. On this basis, it is possible to recognize three major "orogenic" events in the British Caledonides.

The first of these is the late Precambrian "Celtic" (Wright, 1969) or "Cadomian" event, discernible only in the Welsh region, which was responsible for deformation of the Monian and formation of the Upper Longmyndian molasse of Shropshire. The second major event has been termed the "Grampian" (Rast and Crimes, 1969) and corresponds to the mid-Ordovician deformative and metamorphic phases which are manifest in the Scottish Highlands but which also find expression in terms of volcanicity, lithological contrast or actual unconformity within the paratectonic belt. There is persuasive evidence, principally from western Ireland, that this "Grampian" event may involve diachronous structural and metamorphic phases spanning the Arenigian - early Caradocian time-interval. The last major episode of deformation is the late Silurian or "Cymrian" event whose structural effects are widespread throughout south Scotland, the Lake District and Wales, where thin Lower Devonian molasse sequences lap unconformably across strongly deformed older rocks. Along the northern and southern borders of the paratectonic belt (Scottish Midland Valley and South Wales-Welsh Borderland respectively) the thick molasse sequences (previously regarded as entirely of Early Devonian age but almost certainly including Late Silurian redbeds; cf. Walmsley and Bassett, 1976) pass quasi-conformably down into the Silurian and the final Caledonian event is represented by a mid-Devonian episode of faulting and gentle warping.

A PLATE TECTONICS MODEL FOR THE BRITISH CALEDONIDES

The justification for viewing the evolution of the British Caledonides in the context of plate tectonics and ocean-floor spreading involves manifold considerations, but principally (a) the occurrence of ophiolite suites which correspond closely in sequence, structure and composition to present-day ocean crust and upper mantle and are associated with sediments of deep oceanic aspect; (b) the nature of certain lateral changes in sedimentary facies (e.g. the "shelf" to "basin" facies of the Welsh late Ordovician and Silurian) which are analogous to those encountered at existing continent-ocean interfaces; (c) the existence of geochemical polarity in volcanic sequences and of paired metamorphic belts, both of which phenomena are associated with modern island arcs. Additional evidence for the existence of a major Lower Palaeozoic ocean derives from the contrasting nature of Cambro-Ordovician benthonic faunas on opposite flanks of the Caledonian orogen. The totally dissimilar nature and history of the Precambrian basement in the Welsh/English and Scottish/Irish zones also argues for an initial separation of substantially greater magnitude than exists at present, with the corollary that intervening (?oceanic) crust has been destroyed to achieve the current proximity.

Viewed in this context, the history of the British Caledonides may be expressed in terms of the growth and destruction of a major ocean which came into existence in the late Precambrian and persisted in the British region until late Silurian time (Dewey, 1969, 1971, 1974).

The earliest evidence for the existence of such an ocean derives from Anglesey, northwest Wales, where the late Precambrian Monian sequence includes metamorphosed turbidites and ophiolitic equivalents of ocean crust layers 1, 2 and 3 forming a tectonic mélange which probably formed in a subduction zone (but *see* Maltman, 1975). The pre-Arvonian ("Celtic event") deformation of this sequence may have resulted from the arrival of a southern pre-Caledonian continent at this subducting margin with the formation of a paired blueschist-greenschist metamorphic belt in Anglesey. In Shropshire, Uriconian acid volcanism and Longmyndian molasse deposition may represent differing responses within the southern continental plate to the uplift of this Monian orogen, which continued (as the Irish Sea Horst) to play an important role in the Lower Palaeozoic history of the Welsh-Irish-Lake District basins. Further (Eo-Cambrian) but less severe deformation of the Longmyndian and the Arvonian phase of acid volcanism may record additional within-plate movements heralding inception of the Cambrian marine basin of Wales. Although pillow-basalts and dark lutites occur within the Lower Palaeozoic sequence in Wales there is no unequivocal evidence for the presence of true oceanic crust or subduction and plate consumption during this period. This, coupled with the presence of an underlying continental basement, suggests that during Early Palaeozoic time the Welsh area may have been a marginal basin, possibly of Sea of Japan type, lying to the southeast of the Lower Palaeozoic ocean (Ziegler and McKerrow, 1972).

Conversely, in the zone lying north of the Irish Sea Horst and south of the Highland Boundary Fault, the apparent absence of true continental basement and the common occurrence of the ocean crust association of cherts, black shales, pillow-basalts and gabbroic intrusives suggest that this area occupies the site of an early Palaeozoic ocean. Within this central zone of the Caledonides the oldest rocks seen are late Cambrian sediments (Lake District/Isle of Man). These are succeeded by a thick pile of mid to late Ordovician andesites which probably indicates the site of an island arc and suggests the existence of a subduction zone lying beneath the Lake District (Fitton & Hughes, 1970).

The northern margin of the Palaeozoic Ocean in Britian has had a history substantially different from that outlined above. Here, the Torridonian-Moine-Dalradian assemblages appear to record prolonged accumulation of a thick pile of terrestrial and marine sediments, deposited on the southeastern flank of the Lewisian continent during the main phase of Proto-Atlantic opening in the late Precambrian to early Ordovician time-interval. It is probable that the wedge of Dalradian sediments extended into what is now the Scottish Midland Valley and that the continent-ocean suture of early Ordovician time is represented by the lutites, cherts, pillow basalts and serpentinites of the Arenigian Ballantrae Complex. The associated blue-schist metamorphism suggests that this Complex was sited in a deep oceanic trench above a subduction zone at which consumption of the northwestward-spreading ocean floor commenced during the early Ordovician. However the relationship of these rocks to the analogous sequences of the Highland Border Group on the northern margin of the Midland Valley remains in doubt. Moreover, recent geochemical work (Wilkinson and Cann, 1974) on the Arenig-Caradoc Ballantrae spilites suggests that hot-spot basalts and island-arc tholeiites may be present, in addition to 'normal' ocean-floor basalts.

The Mid-Ordovician Grampian deformation and metamorphism of the Scottish Highlands has been linked with the inception of oceanic contraction and was followed by very rapid elevation and unroofing. During the Caradoc and early Silurian, large quantities of detritus (including abundant andesitic volcanics) derived from this uplifted region were supplied across a narrow platform region (Girvan area) and filled the Ballantrae trench, gradually advancing southward across the thin pelagic sequences (Moffat Series) of the ocean floor.

By the mid-Silurian (Wenlock) it appears that the Proto-Atlantic had narrowed considerably, promoting much stronger affinities between the shallow-water benthos on either side of the ocean (cf. Cocks and McKerrow, 1973). Internal source-lands (e.g. Cockburnland, Solwayland) also appeared within the contracting ocean, with resultant reworking of earlier Palaeozoic sediments and volcanics. Complete closure of the ocean in the British region may have ensued from collision of the northern and southern continents in the late Silurian and led to the final deformation of rocks within the paratectonic zone ("Cymrian event"). Subsequent isostatic uplift of the overthickened crust resulting from the collision, and deformation (and the accompanying magmatism) was accompanied by deposition of the Lower Old Red Sandstone (Lower Devonian) molasse in fault-bounded cuvettes within the orogen and on its northern and southern flanks.

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THE CALEDONIDE OROGEN IN IRELAND

PROJECT 27 CALEDONIDE OROGEN

W.E.A. Phillips

Department of Geology,
Trinity College,
Dublin 2, Eire.

REGIONAL GEOLOGY

In Ireland, the Caledonide Orogen can be divided into five stratigraphical and structural zones (Fig. 16.1), which are comparable with Zones C - G of the Canadian Appalachians (Williams *et al.*, 1972). Northwest and southeast foreland zones lie outside Ireland.

Zone 1 (C, Appalachians) in northwestern Ireland is composed of a thick (c. 17 km) succession of clastic sedimentary and minor basic volcanic rocks (Harris and Pitcher, 1975), all affected by late Cambrian-early Ordovician polyphase deformation, up to amphibolite facies metamorphism and syntectonic calc-alkaline intrusions (Pitcher and Berger, 1972). These late Precambrian and Cambrian rocks of the Moine and Dalradian Supergroups (c. 700-500 Ma) are underlain by partly rejuvenated continental basement which in north Mayo was probably deformed in the Grenvillian Orogeny (van Breeman *et al.*, 1976). In the extreme northwest, the Lewisian gneisses of Inishtrahull were deformed before 1500 Ma (MacIntyre *et al.*, 1975). The Dalradian rocks of Connemara lie detached and south of the main part of Zone 1. This detachment can be explained by strike-slip faulting (Phillips *et al.*, 1976). The Dalradian of Connemara has an unconformable cover of gently folded unmetamorphosed Silurian clastic sediments derived from a northern landmass (Laird & McKerrow, 1970; Piper, 1972). In the rest of Zone 1 a few isolated outliers of Lower Middle Devonian form the only pre-Carboniferous cover to the Dalradian. These fluviatile red beds are faulted and gently folded.

Zone 2 (D, Appalachians) consists of greywackes, pelites, and acid and basic volcanic rocks of late Cambrian (?) to early Caradoc age. Though these rocks have been regarded as an unconformable cover sequence to the Dalradian (Dewey *et al.*, 1970), it appears more likely that they represent a slope and shallow water volcanic facies developed along the southern side of the Dalradian basin (Phillips *et al.* , *op. cit.*). These rocks appear to have been affected by the same polyphase deformation, greenschist facies metamorphism and calc-alkaline intrusions as the Dalradian of Zone 1. There is an underlying and partly rejuvenated basement of amphibolite - granulite facies metasedimentary rocks, which, like the "older" Moine of Scotland, may have a Grenvillian orogenic history (Phillips *et al.*, 1975). There is an unconformable cover of folded late Ordovician and Silurian clastic sediments derived from the north (Phillips, 1974). A further angular unconformity lies at the base of Lower Devonian fluviatile red beds with basaltic flows. There are also some small syntectonic and post-tectonic granitic intrusions of late Silurian-Devonian age. The boundary with Zone 1 is a slide or thrust zone, probably of mid-Ordovician age; this is replaced further west by the younger sinistral Leck-Leannan Fault (Pitcher *et al.*, 1964; Phillips *et al.*, 1969). The Orthotectonic or Metamorphic Caledonides are confined to Zone 1 and 2. The rocks of Zone 2 strike eastwards towards the Midland Valley of Scotland.

Zone 3 (E, Appalachians) contains a distinctive suite of Llanvirn-Wenlock pelites, cherts, greywackes and minor basic volcanic rocks. Strike faults separate stratigraphic units which generally young to the north, however the sedimentary pile as a whole youngs in the opposite sense with the oldest rocks outcropping in the north. A quartz-bearing turbidite facies appears first in the north (Caradoc) and arrived progressively later towards the south. The succession is usually overturned, dipping southeastwards, and pre-cleavage overturned folds and/or thrust slices are detected by alternating facing up to the northwest and down to the southeast on cleavage. Large-scale monoclines and cross-folds post-date the cleavage. There are some late-tectonic intrusions of granodiorite probably of Lower Devonian age (O'Connor, 1975). Xenoliths of granulite-amphibolite facies gneisses from Lower Carboniferous agglomerates suggest a basement of continental crust comparable to that of Zone 2 (Strogen, 1974). The boundary between Zone 2 and 3 is concealed both by Carboniferous rocks and probably by the Devonian Galway Granite.

Zone 4 (F, Appalachians) contains at the base a thick sequence of Lower Cambrian-Llanvirn greywackes, quartzites and siltstones with local centres of basaltic volcanism. A basin margin lay to the southeast in Zone 5. These rocks are followed by an extensive suite of tholeiitic and calc-alkaline basalts and andesites followed in the southeast by abundant rhyolites. The main axis of this volcanism extends from Arklow to Dingle (Fig. 16.1). Volcanism ended in the Ashgill in the east but in the Ludlow in the west. The volcanic rocks pass laterally towards the northwest and north into a pelite and greywacke succession with some Llanvirnian-Caradocian local centres of tholeiitic basalt. In the southwest and northwest the sequence is a conformable greywacke succession up into the Ludlow where there is a transition into fluviatile red beds. Caledonian deformation produced upright variably plunging folds with steep dipping cleavage and minor subsequent ductile and brittle deformation. The Leinster Granite plutons were intruded during the later stages of deformation (Brindley, 1969), probably in late Silurian-early Devonian times. The boundary between Zones 3 and 4 is a major strike fault (Navan-Shannon Fault).

Zone 5 (G, Appalachians) forms a small part of southeast Ireland (Max, 1975). It comprises early grey gneisses of the Rosslare Complex, possibly as old as 2400 Ma. Two subsequent cycles of localized granitic intrusion, deformation, amphibolite facies metamorphism and intrusion of basic dykes are as yet undated. These events were followed by the deposition of late Precambrian greywackes and quartz sandstones, correlated with the Mona Complex of Wales. These younger rocks underwent

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Caledonian-Appalachian Orogen of the North Atlantic Region;
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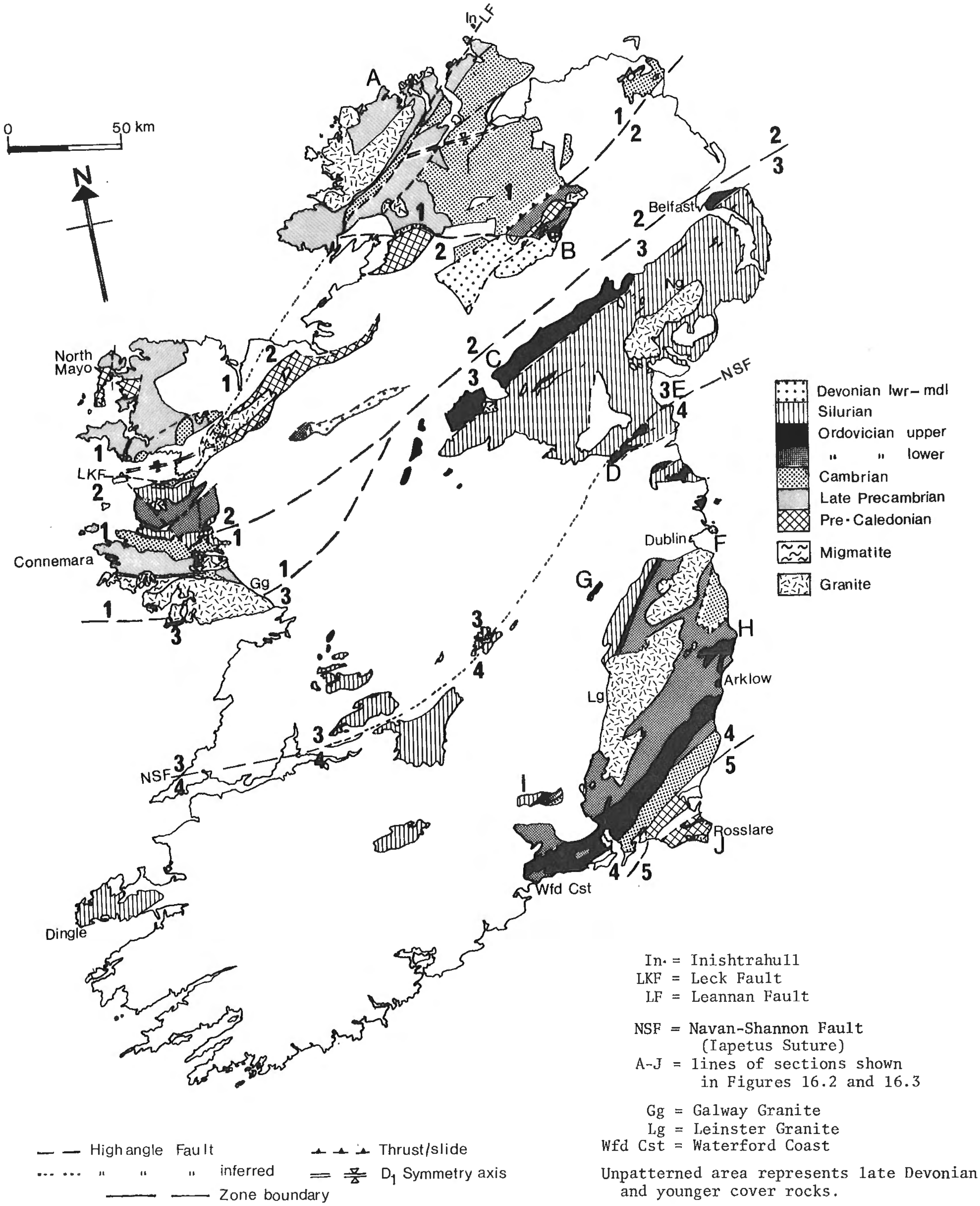

Figure 16.1. The composition of the main structural zones (1-5) of the Caledonide Orogen in Ireland. Pre-Caledonian includes 'old Moine' - probably Grenville and Lewisian - in the northwest. The Pre-Caledonian of the southeast includes the Rosslare Complex (cf. Pentevrian of northwest France) and the Cullenstown Formation (cf. Mona Complex of north Wales). The late Precambrian and Cambrian of the northwest comprises 'young Moine' and the Dalradian Supergroup.

polyphase deformation and greenschist facies metamorphism prior to emplacement of the Carnsore Granite at 551 ± 10 Ma. The earlier gneisses were mylonitized during this deformation. Most of the zone was probably uplifted and eroded during the Cambrian. Locally there is an unconformable cover of gently folded but faulted Arenig conglomerates and shales with a shelly fauna. The boundary between Zones 4 and 5 is faulted where exposed.

SYNTHESIS OF THE LATE PRECAMBRIAN AND EARLY PALAEOZOIC THROUGH DEVONIAN EVOLUTION

Consideration of palaeogeography, structure, geochemistry of igneous rocks, and Ordovician faunal provinciality has led to the hypothesis that a Northwestern and a Southeastern plate can be recognized (Dewey, 1974; Phillips *et al.*, 1976) in the Caledonide Orogen of Ireland. In the Northwestern Plate, comprising Zones 1 - 3, American faunal province graptolites and shelly faunas have been found in the lower Ordovician rocks of Zone 2. In the Southeastern Plate which includes Zones 4 and 5, the lower Ordovician faunas are of European provinciality. It appears, therefore that the boundary, or Iapetus Suture, between the two plates is the strike fault between Zones 3 and 4. Using palaeontological and isotopic dating to define the time of ending of igneous activity related to subduction, it appears that plate collision migrated westwards with time and was followed by several hundred kilometres of dextral strike-slip displacements on the Iapetus Suture.

Northwestern Plate

The Caledonian Orogenic Cycle started about 700 Ma ago in Zone 1 with deposition of Moine and younger late Precambrian-early Ordovician Dalradian sediments. Most of the Precambrian succession is of shallow-marine shelf facies, dominated by arkoses, sandstones, dark pelites and thin limestones. Sedimentation seems to have been related to a landmass of continental crust lying to the northwest. Soon after deposition of the late Precambrian tillite, more basinal sedimentation set in, with turbidites, dark pelites and tholeiitic basaltic volcanicity. This facies is also seen in Zone 2 and extends up into rocks as young as lower Caradocian (J. Sullivan pers. comm., 1976). The earlier Ordovician rocks of the western part of Zone 2 show a southwards transition into a shallow-water volcanic shelf facies with American faunal province shelly and graptolitic faunas (Dewey *et al.*, 1970, Williams, 1972). There is evidence that sedimentation after the tillite was related to a northwestern and a southeastern landmass. The basin as a whole was probably ensialic, formed by rifting associated with opening of the Iapetus Ocean further to the southeast. No unmoved contacts with the underlying basement are preserved. In late Cambrian-early Ordovician times, deformation produced large recumbent folds and thrust sheets which face away from a central upright NE-SW trending synclinal symmetry axis (Phillips, in press.). Further smaller scale coaxial folding and cross-folding was associated with normal (Barrovian) metamorphism usually of the greenschist but sometimes of low amphibolite facies grade. In the Dalradian rocks of Connemara, metamorphism of medium to high amphibolite facies grade was accompanied by unusually profuse calc-alkaline igneous activity (Leake, 1970). Early normal (Barrovian) metamorphism here was followed by low pressure (Buchan) metamorphism (Yardley, 1976). The overall facing direction of D1 structures in Connemara is uncertain, there has been intense refolding by large-scale D2-D4 structures. Isotopic and palaeontological evidence suggests that deformation and metamorphism occurred in late Cambrian-Arenig times in Zone 1 (Leggo and Pidgeon, 1970; Pidgeon, 1969), but in later Llanvirn-early Caradoc times in Zone 2 (Phillips *et al.*, 1976). These processes may well have spread upwards and outwards from the more central parts of the orogen towards its southeastern margin in Zone 2. The anomalous situation of Zone 1 Dalradian rocks of Connemara lying south of Zone 2 may have arisen by sinistral strike-slip movement on the pre-Upper Llandovery fault forming the northern margin of the Connemara Dalradian. In Zone 2 there is a major angular unconformity followed by northern derived clastic sediments ranging in age from upper Caradoc to Wenlock. This marine transgression across the earlier orthotectonic Caledonides, reached the western part of Zone 2 in the Upper Llandovery. Silurian sediments here show a complete transgressive-regressive cycle ending with probable Ludlovian red beds; there is also a transition from a northern shallow-shelf facies to a deeper marine turbidite facies in the south. This lateral facies change extends southwards over the Dalradian rocks of Connemara (Zone 1) and is probably continued by the deeper water facies of Zone 3. Late Silurian-early Devonian deformation produced upright NE-SW to E-W trending folds and cleavage. There is a marked increase in the intensity of this deformation towards the Leck Fault in the western part of Zone 2 (Dewey and McManus, 1964). Gently folded Lower Old Red Sandstone fluviatile sediments with some basaltic flows rest with angular unconformity on folded Silurian rocks in Zone 2. The small granitic intrusions in these Devonian rocks are probably of the same age as the larger post-tectonic granites of Connemara, dated in the range of 400-430 Ma (Leggo *et al.*, 1966). The late, post-Wenlock Caledonian movements also generated sinistral strike-slip movement on the Leck-Leannan Fault of Zone 1. The synclinal symmetry axis of the Dalradian appears to have been displaced sinistrally by about 160 km. There is evidence that this movement may have started in the Wenlock and continued to late Devonian times (Phillips *et al.*, 1969).

The oldest dated rocks of Zone 3 are late Llanvirn pelites lying within a northern belt of Caradocian turbidites. These northern turbidites contain igneous and metamorphic detritus probably derived from the Lower Ordovician volcanic rocks and their metamorphic basement to the north in Zone 2. To the southeast these Ordovician turbidites pass into a black shale and chert facies. The late Llanvirn - early Caradoc (*Nemagraptus gracilis* zone) sediments of Zone 3 must be of the same age as the youngest sediments in the orthotectonic Caledonides, yet the Zone 3 rocks have not been affected by the Caledonian orogenesis seen but 40 km to the north in Tyrone. This contrast enhances the case for an allochthonous origin for the Ordovician and Silurian rocks of Zone 3, argued below. The largest part of Zone 3 is composed of Llandovery-Wenlock turbidites with small inliers of older dark pelite, chert and basic volcanic rocks. Faunas are predominantly graptolitic (Weir, 1973). The southeasterly progression of the turbidite facies is continued, with a pelagic dark shale facies continuing into the Upper Llandovery near the boundary with Zone 4. Extensive slumping is typical of Zone 3, producing large units of structureless sediments and also slump folds and fault blocks in coherent bedded sediment on the scale of several kilometres. Preliminary data suggest slumping was towards a southern quadrant. The sedimentary facies and distinctive structural style of strike-faulted strips of succession facing northwestwards but with younger rocks to the southeast, is consistent with a model of northwestwards obduction of Iapetus ocean-floor sediment onto a basement of continental crust of the Northwestern Plate. Such a basement is inferred from the evidence of the

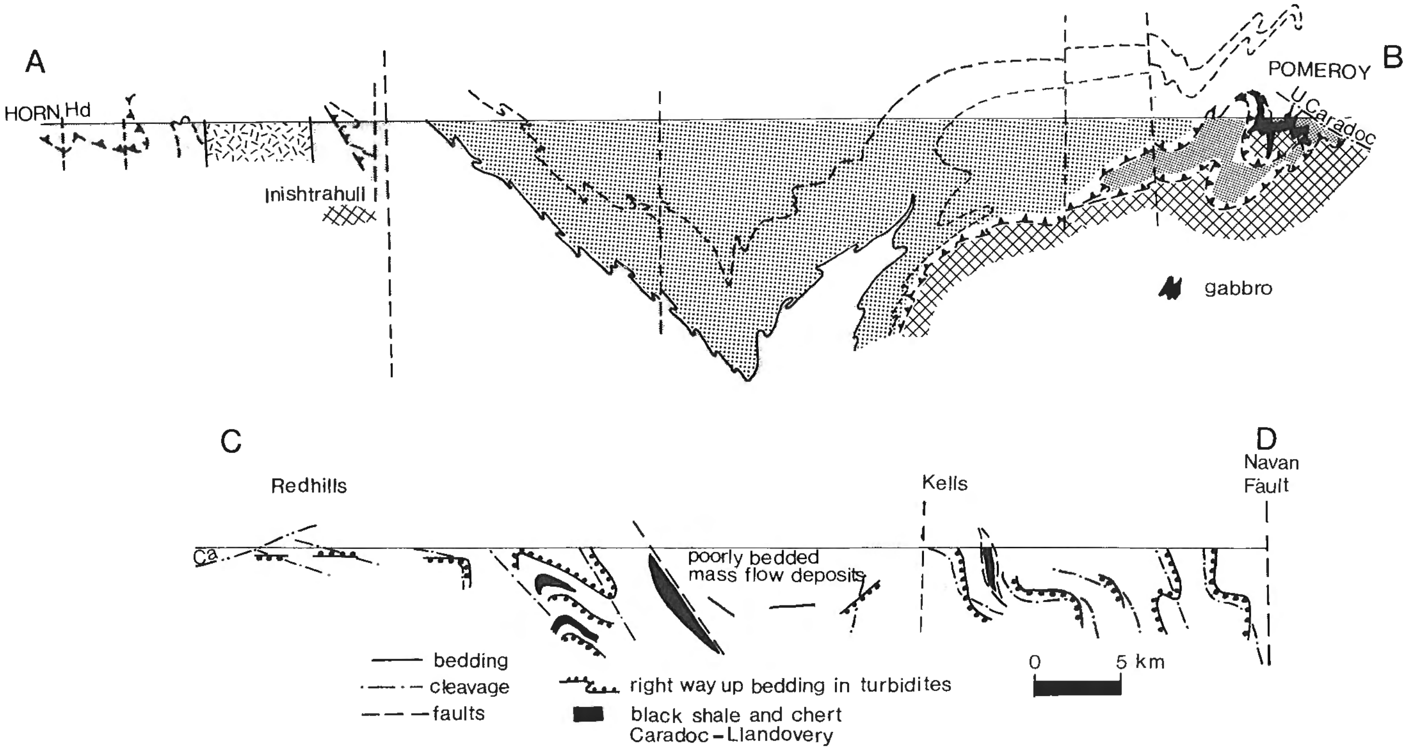

Figure 16.2. Sections through the Caledonide Orogen of the Northwestern Plate, Ireland. Section A-B, symbols as for Figure 16.1. Section C-D, unpatterned area represents Caradoc-Llandovery turbidites and mass flow deposits. C = Lower Carboniferous.

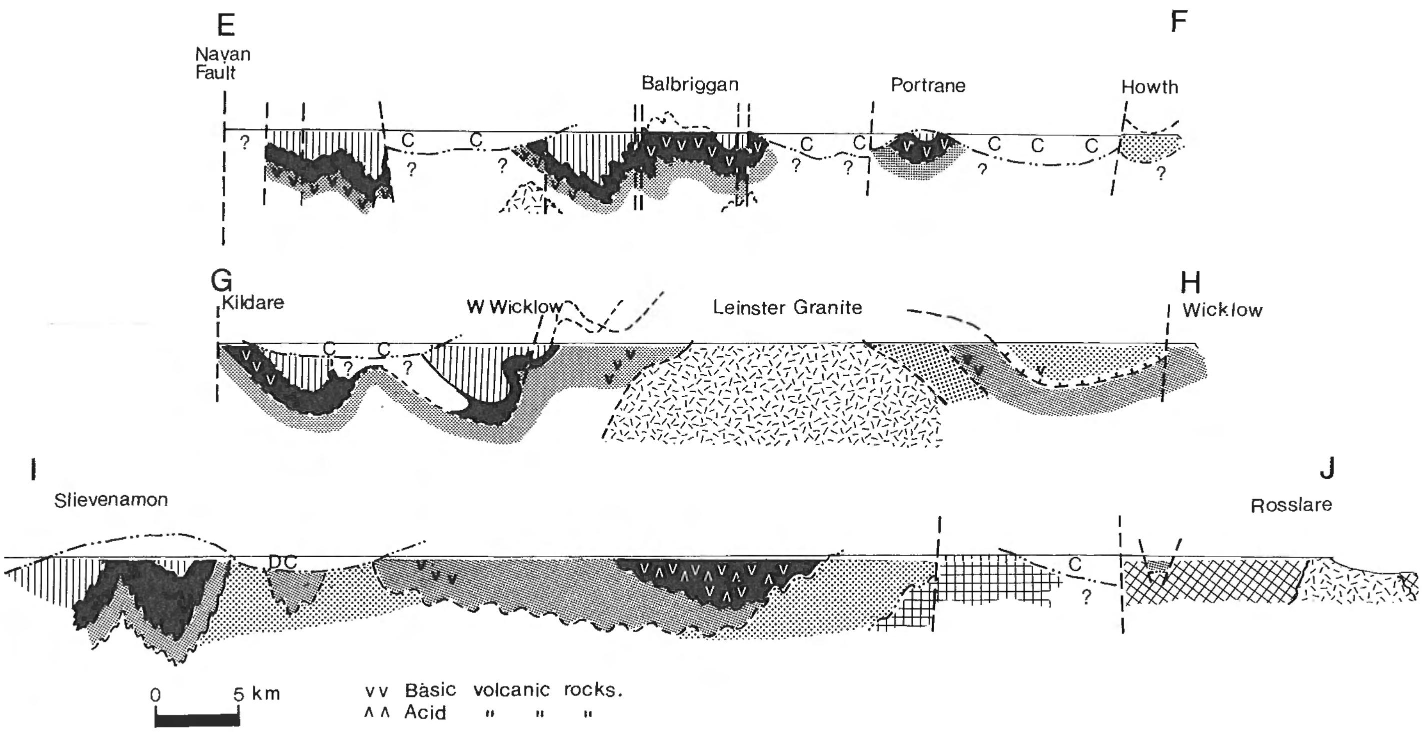

Figure 16.3. Sections through the Caledonide Orogen of the Southeastern Plate, Ireland. Symbols as for Figure 16.1. D = Devonian (Upper). C = Carboniferous.

gneiss xenoliths, regional gravity and magnetic data, and analogy with the Southern Uplands of Scotland where more detailed geophysical data also indicates a continental crustal basement. Northwesterly subduction of Iapetus oceanic crust could readily account for the Ordovician volcanic arc of Zone 2 and the Caledonian thermo-tectonic history of Zones 1 and 2. A further implication of an obduction model for Zone 3 is that thrusting of allochthonous sedimentary slices was a continuous process starting probably in the Arenig and continuing through the Silurian. It is possible therefore that some of the tectonic deformation, which produced the NE-SW to E-W trending cleavage and F_1 folds, was diachronous as a result of cumulative obduction. Emplacement of the Galway Granite dated at 407 ± 6 Ma* (Leggo *et al.*, 1966) and of the Newry Granodiorite - 399 ± 21 Ma - post-dates the local D_1 deformation. The age of later structures is uncertain, NE-SW trending F_2monoclines produced a series of major steps of bedding and cleavage, which ascend northwestwards with flat lying limbs becoming dominant in that direction. NW-SE trending kink bands and F_3 cross-folds with wavelengths up to 1 km are also important. Though some of the strike faults are probably pre-D_1 structures related to obduction, there is evidence for post-cleavage movements on many of these faults.

Southeastern Plate

A major angular unconformity is inferred on structural grounds between the Cambrian of Zone 4 and the Precambrian of Zone 5 (Dhonau, 1972). The two deformation phases and up to greenschist facies metamorphism of the Cullenstown Formation of Zone 5 is best correlated with the Monian orogenesis of Wales and the Cadomian of northwest France (Max, 1975). In comparison to the Northwest Plate, the Caledonian cycle started much later, in early Cambrian times; and there is also a contrast in basement history. The Cambrian sediments of Zone 4 have been dated by acritarchs; the rocks consist of greywackes, siltstones and quartzites. The sediments appear to have been deposited in a basin (Leinster Basin) separated from the Welsh Basin by an Irish Sea Landmass lying within Zone 5. There is considerable evidence for derivation of turbidites and slumped quartzites from this landmass, with more distal basinal turbidites, suspension sediments and local basic volcanism lying further to the northwest (Crimes and Crossley, 1968; Brück and Reeves, 1976). In the Dublin region there is some evidence of derivation from continental crust to the northwest (Brindley *et al.*, 1973; Brück and Reeves, 1976). The Leinster basin may well have been an ensialic graben related to opening of the Iapetus Ocean to the northwest. In Zone 5, shallow-water Arenig conglomerates and siltstones, with a shelly Celtic Province fauna (Williams, 1969), rest unconformably upon the Precambrian basement and indicate the continued survival of the Irish Sea Landmass. To the northwest in Zone 4, the Cambrian passes conformably up into the Ordovician within a basinal siltstone facies, with local development of basic lavas and tuffs (Brück *et al.*, 1974) and then conformably into the Silurian (Brück, 1972). Atlantic Province Arenig graptolites have been found in the southeastern part of Zone 4, and Llanvirn graptolites of the same province have been found in the northern part of the Zone (Skevington, 1974). The major feature of the Ordovician and Silurian rocks of Zone 4 is the development of a southeastern volcanic arc with a sedimentary arc to the northwest. There is a divergence of about 9^o towards the west between the volcanic arc and the Iapetus Suture. In the northern part of Zone 4, Atlantic faunal province graptolites indicate that volcanism started locally in the Llanvirn; it became most widespread in the Caradoc when tholeiitic basic and intermediate eruptives formed islands associated with non-terrigenous sediment and local shelly faunas. There is no geochemical evidence for continental crust playing a role in the generation of these magmas. Volcanism is most extensively developed further to the southeast in Zone 4 in a belt extending from Arklow to the Waterford coast (Stillman *et al.*, 1974). In Waterford pre-upper Llandeilo fault-controlled submarine basins developed probably on continental crust of the type seen now immediately to the east. Volcanism produced large shield volcanoes with mixed tholeiitic and calc-alkaline basalts and andesites. After a hiatus marked by an unconformity further east and then by deposition of late Llandeilo limestones with shelly faunas, renewed fracturing and basin subsidence was accompanied by abundant calc-alkaline basaltic, andesitic and rhyolitic eruptions and intrusion of rhyolitic sheets. Subsequently a series of more alkaline intrusions were emplaced. The late Silurian to early Devonian Leinster Granites were probably genetically related to the volcanic arc. The arc appears to swing westwards and be represented by the acid and basic volcanic rocks of Wenlock age in the Dingle Peninsula. The petrological and geochemical variations of the volcanism as a whole, and the Kuroko-type mineralization associated with it at Avoca are consistent with a model of southeast directed subduction of Iapetus oceanic crust beneath an Andean-type arc built up on continental crust which was probably thicker towards the southeast. The volcanic rocks of the main axis from Arklow to Dingle, pass into a basinal shale and turbidite sequence to the northwest (Cope, 1959; Doran, 1974) which extends up into the Lower Ludlow (Parkin, 1976). The turbidites show evidence of derivation from the volcanic arc; they also arrived at progressively later times towards the northwest, probably as a result of shallowing in this direction. In the Dingle Peninsula the Silurian passes up into a fluviatile facies within the Ludlow (Holland, 1969). Caledonian deformation in Zone 4 produced upright and asymmetrical NE-SW trending folds with a steep axial planar cleavage. The intensity of deformation decreases towards the west where an E-W trend is seen. The Leinster Granites were emplaced after the first Caledonian deformation but are affected by later less intense Caledonian strain. These later Caledonian movements have produced some asymmetrical minor folds and crenulation cleavage. Faulting seems to be the most important form of Caledonian strain in the Arenig rocks of Zone 5.

ACKNOWLEDGMENTS

I am very grateful to Professor C.H. Holland and Dr. P.M. Brück for their helpful comments.

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* (5.0×10^{10} half life is used throughout this paper)

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APPENDIX

Important gaps in present research

1. Seismic refraction line experiments to study crustal structure and thickness.

2. Drilling stratigraphic bore holes in critical areas e.g. Arenig of South Mayo, Cambrian of Dublin area.

3. Detailed petrological studies to investigate Caledonian geothermal gradients.

IUGS UNESCO
PROJECT 27
CALEDONIDE
OROGEN

17. EAST GREENLAND CALEDONIAN FOLD BELT

Niels Henriksen

Grønlands Geologiske Undersøgelse
Øster Voldgade 10
DK-1350 Copenhagen, Denmark

INTRODUCTION

The Caledonian fold belt can be traced for more than 1300 km along the east coast of Greenland between 70° and 82°N (Fig. 17.3). At the partly exposed, western thrust boundary the fold belt comes into contact with unmetamorphosed Cambro-Silurian foreland sediments in the north and with rocks of the Precambrian Greenland shield further south. South of 70°N the fold belt is covered by Tertiary basalts and in the east it is bordered by a belt of late Palaeozoic and Mesozoic sediments and the Greenland Sea.

The fold belt is of a composite nature comprising pre-Caledonian as well as Caledonian elements. In the most comprehensive accounts of the fold belt (Haller, 1970, 1971) the main part of the metamorphic complexes are considered as being the deep-seated, mobile infrastructure of the Caledonian fold belt. The characteristics of the complexes are regarded as being essentially Caledonian in origin and the associated high-grade metasedimentary rocks are viewed as parts of the late Precambrian accumulations, metamorphosed during the Caledonian orogenic event. Recent work in the southern part of the fold belt (Henriksen and Higgins, 1976) and in particular the results of radiometric dating suggest that areas of pre-Caledonian basement rocks are more widespread than earlier supposed. The new work from the southern part of the fold belt indicates that the metamorphic complexes retain characteristics of several pre-Caledonian orogenic events and that some high-grade metasediments were metamorphosed in pre-Caledonian time. North of 76°N the main part of the fold belt is formed by the Caledonian reactivated remnants of the presumed late Precambrian Carolinidian fold belt.

In Figure 17.1 the apparent chronological succession of events in the East Greenland fold belt is summarized according to the present knowledge.

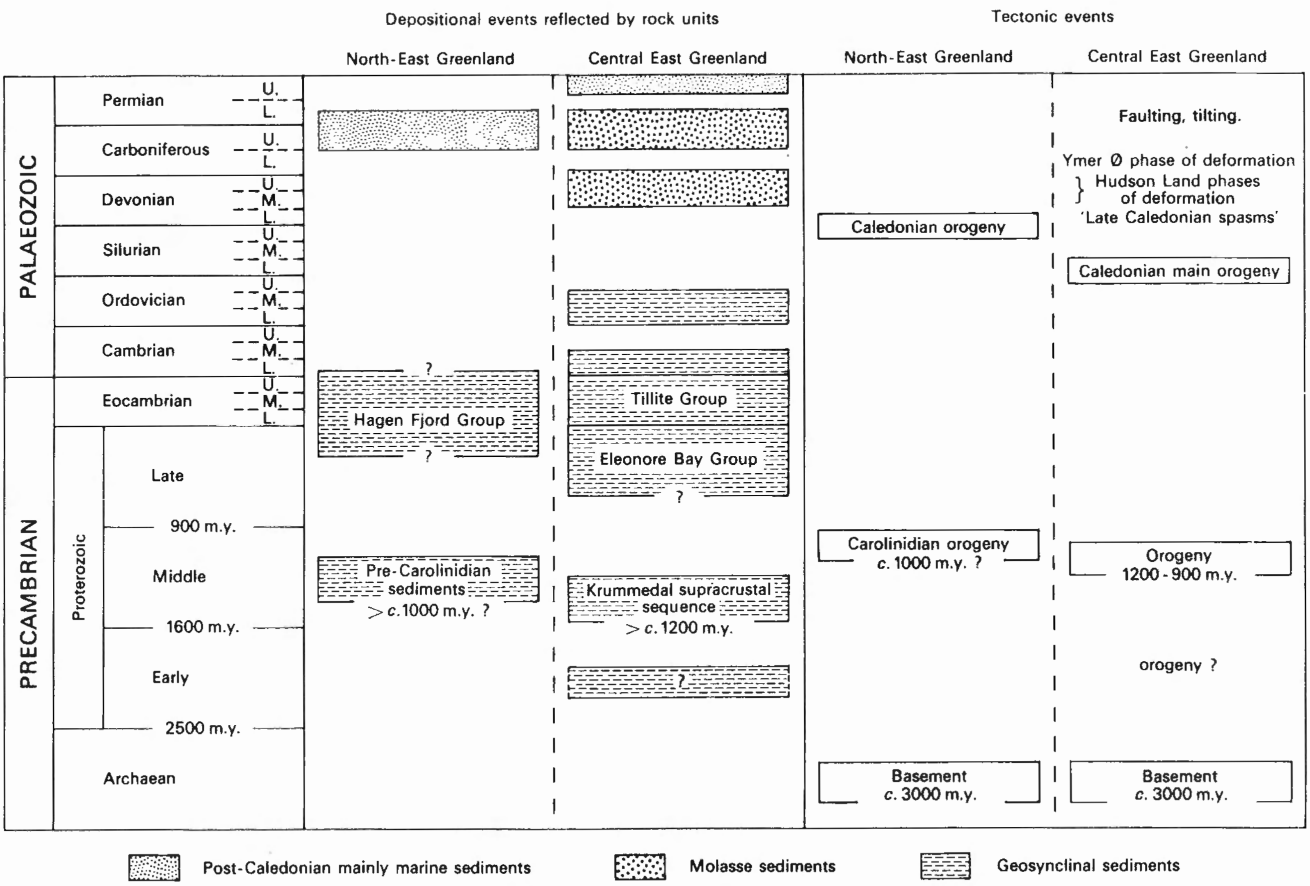


Figure 17.1. Provisional chronological scheme of deposition and orogenic events represented in the East Greenland fold belt.

IGCP Project 27, Danish Contribution No. 1.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

LATE PRECAMBRIAN-LOWER PALAEOZOIC GEOSYNCLINAL DEPOSITS AND THEIR CALEDONIAN DEFORMATION

Major developments of late Precambrian and early Palaeozoic non-metamorphic to slightly metamorphic sediments are found in North-East Greenland and Central East Greenland. The two successions are developed in different ways, but both reflect accumulations of geosynclinal proportions which later became involved in Caledonian deformation. In the south between $71^{\circ}30'$-$76^{\circ}00'$N a miogeosynclinal sequence outcrops which totals more than 17 000 m in thickness, and includes presumed Upper Proterozoic, Eocambrian and Cambro-Ordovician sediments. In the north between $79^{\circ}30'$ and $81^{\circ}30'$N an approximately 6000 m thick miogeosynclinal sequence is preserved in allochthonous Caledonian thrust masses. The non-fossiliferous geosynclinal sequence in the north comprises presumed Upper Proterozoic to Eocambrian or Lower Cambrian rocks. A basal unconformity has not been found in connection with either of the geosynclinal sequences. The division and the development of the deposits are shown in Figure 17.2.

The main outcrops of the geosynclinal sediments in central East Greenland are found in two N-S trending belts on both sides of a "Central Metamorphic Complex" between 72° and 74°N and in a NW-SE trending graben structure south of 76°N (*see* Fig. 17.3). In these areas the sediments are only gently deformed in a series of open folds with usually vertical axial planes.

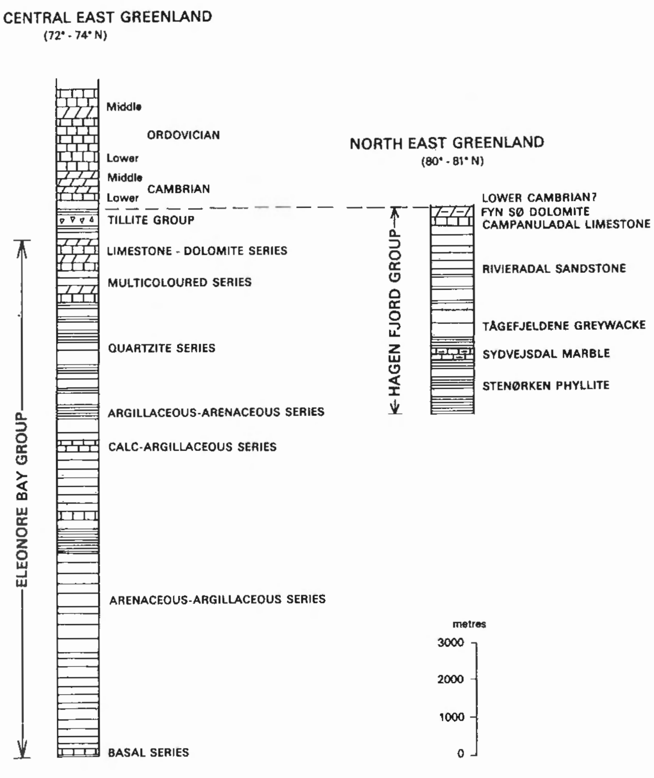


Figure 17.2. Caledonian geosynclinal deposits in the East Greenland fold belt (modified after Haller, 1971).

The deformation of the sedimentary sequence in the northernmost part of the fold belt ($79^{\circ}50'$-$81^{\circ}30'$N) is characterized by considerable lateral shortening, with formation of thrust sheets and gliding nappes. The nappes partially overlap the autochthonous foreland sequence which locally is exposed in tectonic windows.

METAMORPHIC COMPLEXES

Major parts of the fold belt consist of complexes of infracrustal gneisses and migmatites, as well as high-grade schistose and gneissose supracrustals. They include reworked units of pre-Caledonian basement gneisses and schists as well as metamorphosed equivalents to the Caledonian geosynclinal deposits.

"The Central Metamorphic Complex" occupies the inner fjord zone and part of the nunatak zone between 70° and 74°N (Fig. 17.3). It comprises Archean infracrustal complexes (e.g. the Flyverfjord complex) dated at c. 2500-3000 Ma, associated with and overlain by miogeosynclinal supracrustal sequences (e.g. the Krummedal sequence) which have yielded whole rock isochron ages of c. 1000-1200 Ma. The pre-Caledonian infracrustals and supracrustals were strongly folded together during a middle Proterozoic orogeny at the same time as an extensive migmatization took place in some areas and syn- to post-kinematic acid intrusions were formed. During the Caledonian orogeny the complex was reactivated and partly superimposed by Caledonian migmatization and plutonism. In the south the complex is involved in westwards directed, Caledonian thrust nappes.

The region north of 76°N is also characterized by superimposed sets of folds, which are of infrastructural nature south of approximately 79°N and of a more superficial nature further north. The older set of folds is by Haller (1961) referred to a pre-Caledonian event which was termed "the Carolinidian orogeny". The younger NNE-SSW trending pattern is described as formed during the main phase of the Caledonian orogeny. The pre-Caledonian structures were refolded and the earlier rocks reactivated in the deep-seated infrastructural zone of the Caledonian fold belt.

Pre-Carolinidian geosynclinal sediments which are characterized by an ubiquitous distribution of basic sills and dykes are preserved in a low metamorphic form in parautochthonous thrust blocks in the northernmost part of the fold belt. Traced to the south the pre-Carolinidian metamorphic sediments merge into infracrustal rocks. These infracrustal units include not only migmatized equivalents of the pre-Carolinidian sediments, but also remnants of a pre-Carolinidian basement complex yielding an age of c. 3000 Ma. The East Greenland fold belt north of 76°N thus contains elements from at least three orogenic events. The latest of these, the Caledonian event, is generally reflected by the structural and plutonic reactivation of the older rocks units, and only in the region between $79^{\circ}30'$ and $81^{\circ}30'$N have late Precambrian geosynclinal deposits been found.

LATE OROGENIC PLUTONIC ROCKS

South of 76°N the folded Caledonian geosynclinal sediments are cut by a number of granitic intrusions. Most of these are confined to the border zone between the metamorphic complexes and the folded non-metamorphic sediments. The granitic intrusions are mainly post-tectonic stocks and batholiths, or thick sheet-formed bodies, which discordantly cut the surrounding formations.

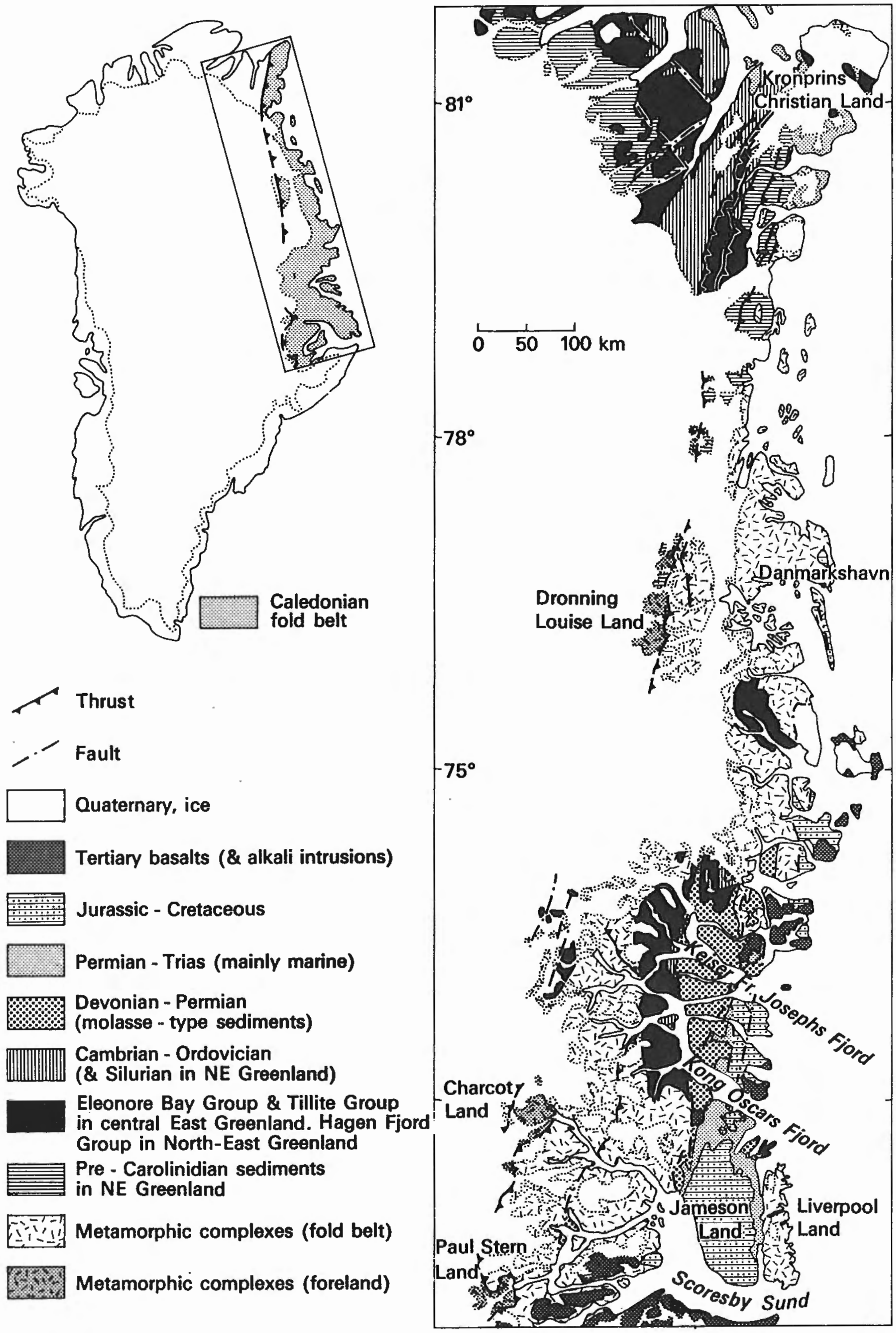
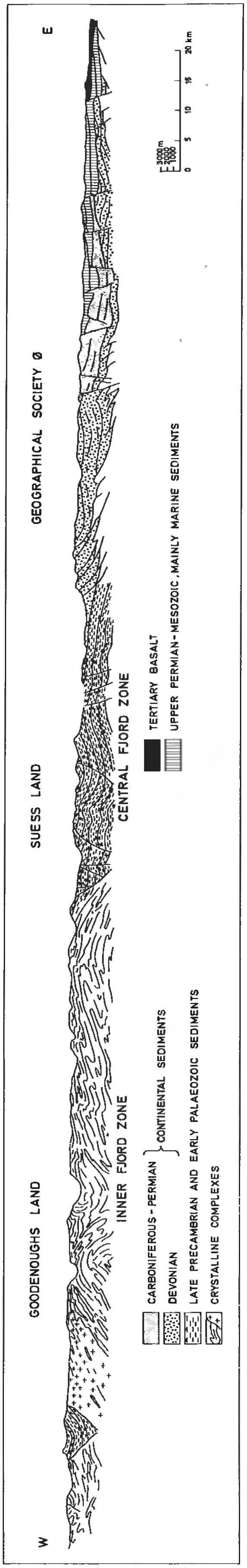


Figure 17.3. Main elements of the Caledonian fold belt in East Greenland. After A.K. Higgins, 1976.



AGE OF CALEDONIAN OROGENESIS IN EAST GREENLAND

The youngest preserved sediments in the Caledonian geosyncline in central East Greenland are of Middle Ordovician or perhaps lowermost Upper Ordovician age. After the Caledonian main orogeny the folded sediments were unconformably overlain by a thick sequence of Middle and Upper Devonian molasse deposits of Old Red Sandstone facies. The stratigraphical evidence for the age of the Caledonian main orogeny in central East Greenland thus places it in the interval between Middle Ordovician and Middle Devonian. A plausible assumption would be, that the main orogeny here occurred shortly after the deposition of the youngest preserved geosynclinal sediments, i.e. in latest Ordovician to Early Silurian time.

In the northernmost part of the fold belt the structures are probably somewhat younger. Here Middle to (?) Upper Silurian sediments are found in the autochthonous foreland sequence, immediately capped by Caledonian thrust sheets.

DEVONIAN-LOWER PERMIAN MOLASSE, AND DEFORMATION

After the Caledonian main orogeny an intermontane molasse basin formed in East Greenland between 71° and 74°N. More than 7000 m of coarse clastic red-coloured sediments of Middle and Upper Devonian age accumulated. A great deal of volcanic material in the form of lava-flows and tuffs is mixed with the sandstones in many places. Folding, thrusting and uplift disturbed the Devonian sedimentation in the northern part of the molasse basin.

Following deposition of the youngest Devonian sediments a break in sedimentation in the Lower Carboniferous was caused by a deformational event, which affected mainly the Devonian molasse deposits. Continental deposition continued thereafter throughout Upper Carboniferous and into the Lower Permian. This succession totals 5000 to 6000 m of mainly clastic sediments which were deposited in intermontane basins.

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MAPS

Region 70°-72°N:	Mapped (1968-72) for publication at scale 1:100 000 by the Geological Survey of Greenland (maps in prep.).
" 72°-76°N:	Mapped by the Danish expeditions to East Greenland (Koch & Haller, 1971) and published at scale 1:250 000. Meddr. Grønland, bd. 183.
" 72°-82°N:	Covered by a tectonic map at scale 1:500 000 compiled by Haller (1970) from mapping by the Danish expeditions to East Greenland, Meddr. Grønland, bd. 171, nr. 5.
Whole fold belt:	Covered by tectonic/geological map at scale 1:2 500 000 by the Geological Survey of Greenland (1970).

APPENDIX

Problems and topics/areas where research is needed.

AREAS: The region 72°-74°N is accessible with charter plane to Mesters Vig airport. Internal transport in this region can be based on inflatable rubber boats. Fjords are ice free in August - mid September. All other regions are more difficult of access and work must mainly be based on helicopter support.

PERMISSION TO WORK: All expeditions must formally apply to the Ministry for Greenland, Hausergade 3, 1128 Copenhagen K. for permission to work in Greenland.

TOPICS:

Metamorphic and plutonic complexes. Detailed structural, stratigraphical, petrological and geochemical investigations are needed in the area from 72°-76°N.

Precambrian-Lower Palaeozoic geosynclinal deposits: Detailed sedimentological, stratigraphical and palaeontological investigations are needed in the area from 72°-76°N.

Carolinidian sector of the fold belt (76°-82°N). Very difficult accessibility. This part of the fold belt is almost unknown and remains to be mapped systematically. Therefore any investigations in this part of NE Greenland are of general interest and important for confirming or rejecting the postulated existence of a special Carolinidian orogeny (1000 m.y. event ?).

SYNTHESIS OF THE CANADIAN APPALACHIANS

PROJECT 27 CALEDONIDE OROGEN

Paul E. Schenk

Department of Geology
Dalhousie University
Halifax, Nova Scotia, B3H 3J5
Canada

INTRODUCTION

REGIONAL SETTING

The Canadian segment (Figs. 18.1, 18.2) of the Appalachian Orogen encompasses an area of approximately 10^6 km^2, of which only one-third is presently above sea level. To the northwest, the Orogen is bounded by the crystalline Canadian Precambrian Shield with fringing, gently folded to flat-lying Paleozoic carbonates. To the southeast, the limit of the Orogen is covered by the Atlantic Ocean and submarine Cenozoic and Mesozoic sediments of the continental shelves and slopes of Nova Scotia and Newfoundland.

The Canadian Appalachians include six main rock assemblages:

1. Older Precambrian rocks which underwent an orogenic event perhaps 1000 Ma ago, and which crop out today both in a median belt and also as Grenvillian inliers along the western margin (Fig. 18.1).

2. Younger Precambrian, Cambrian, and Ordovician rocks that were deformed in some places during the Devonian Acadian Orogeny, and in other places also during the Ordovician Taconian Orogeny (Fig. 18.3).

3. Silurian through Early Devonian rocks that are in some places post-orogenic continental sediments and volcanics, and in other places marine sediments deformed during the Acadian Orogeny (Fig. 18.4).

4. Granitic batholiths of generally Devonian to Carboniferous age (Fig. 18.5).

5. Carboniferous and Permian sedimentary and volcanic rocks deposited generally in intermontane basins, usually fault-bound, and deformed during the Late Carboniferous Maritime Disturbance (Fig. 18.6).

6. Late Triassic sedimentary and volcanic rocks preserved in a rift-zone developed after the Maritime Disturbance (Fig. 18.1).

This paper is basically concerned with assemblages 2 to 4, although both assemblages 1 and 5 are also discussed.

TECTONO-STRATIGRAPHIC ZONES

The Canadian Appalachians are divisible into a number of tectono-stratigraphic zones. Various schemes have been proposed (e.g. Poole, 1967; Belt, 1968; Williams, *et al.*, 1972, 1974; Williams, 1976 and in press; Rast *et al.*, 1976a and b; Church, 1977; Ruitenberg *et al.*, 1977; and Keppie, 1977a). Whereas these zones may have provincial usefulness, extrapolation away from their area of definition has led to controversy (e.g. Rast *et al.*, 1976a; Fyffe, 1977). In part, this is due to the varying widths of each zone. For example, the Dunnage Zone (*see* below) wedges out across Newfoundland because of the southwestward convergence of the Humber and Gander Zones (Fig. 18.3); the Avalon Zone is 550 km wide in Newfoundland but apparently only 25 km wide in southwestern New Brunswick; the Meguma Zone apparently terminates abruptly both to the northeast and to the southwest of Nova Scotia. In part the difficulty in tracing laterally these zones arises from diachroneity of both deformation and metamorphism. The Taconian Orogeny, for example, was the major tectonic event in Québec whereas in Nova Scotia its effect was minimal. Middle Paleozoic zones are very prominent in New Brunswick but absent in Newfoundland (Fig. 18.4). However, generalized tectono-stratigraphic zones at least of Early Paleozoic age may be traced considerable distances if the Orogen is taken as a whole (Fig. 18.1).

The scheme of tectono-stratigraphic zones adopted in this paper on the Canadian Appalachians is an amalgam of Williams (1976) for the Early Paleozoic, Rast *et al.* (1976a and b modified) for the Middle Paleozoic (Siluro-Devonian), and Belt (1968) for the Late Paleozoic. In preview, distinct, continuous zones are most clearly defined by the Late Precambrian through Ordovician succession (Fig. 18.3). In Siluro-Devonian times the zonal pattern had changed (Fig. 18.4). Devonian deformation and plutonism (Fig. 18.5) obliterated this later pattern. During Carboniferous time a simple pattern of one complex, actively subsiding zone was created (Fig. 18.6). Brief descriptions follow of the Late Precambrian-Early Paleozoic (Fig. 18.3), Middle (Fig. 18.4), and Late Paleozoic (Fig. 18.6) frameworks.

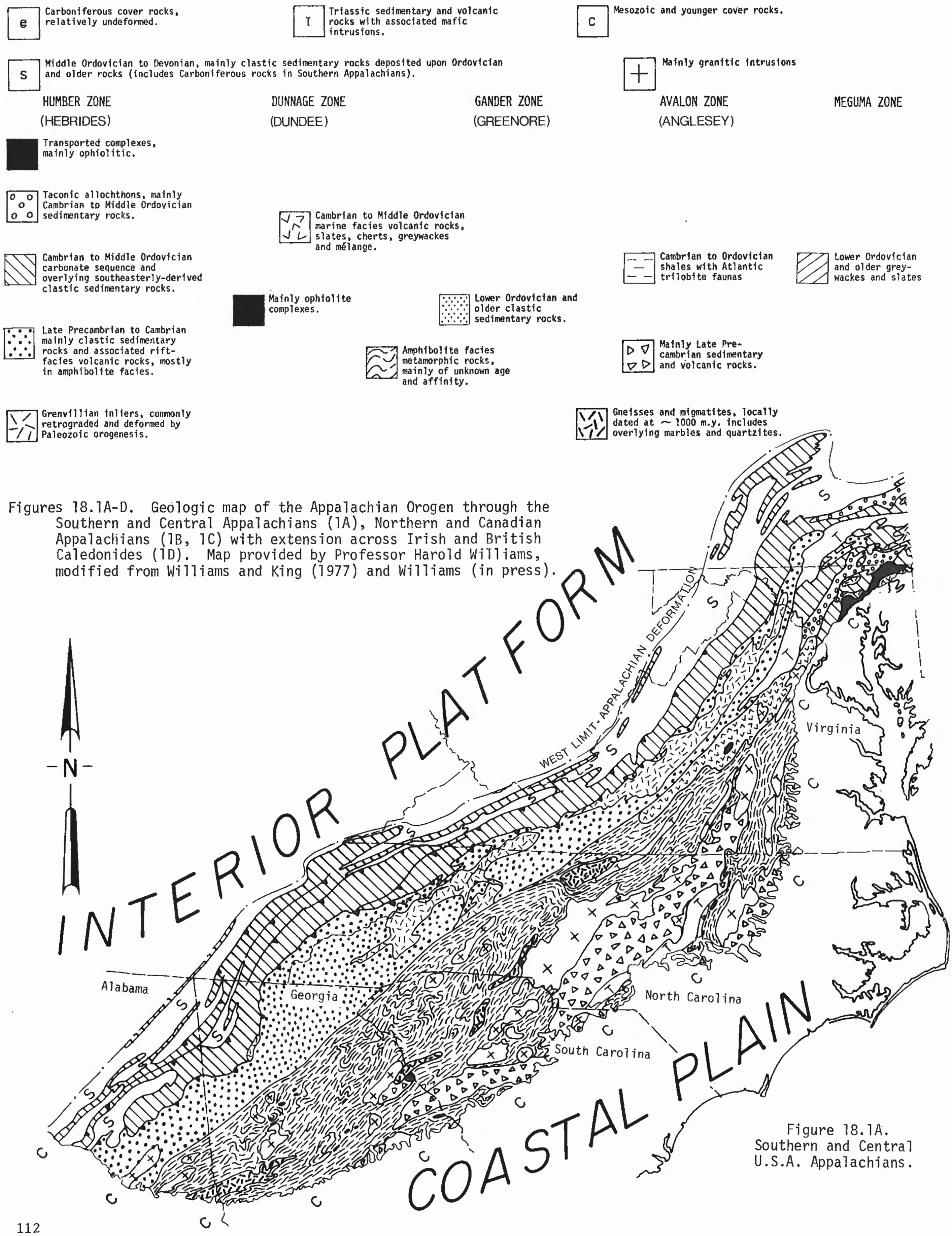
A. Late Precambrian-Early Paleozoic (Cambro-Ordovician) Zones

Five Late Precambrian through Ordovician zones are distinguished. They are listed below with, in parentheses, the letters and names of the equivalent zones as distinguished in the earlier syntheses. The progression is from the northwest to the southeast (Fig. 18.3).

1. *Humber Zone* - The Western Margin of Early Paleozoic North America (= Lomond (A); Hampden (B); western part of the Fleur de Lys (C); Cloridorme).

This zone records the Early Paleozoic construction and destruction of the eastern margin of North America. To the south, the Humber extends into the Valley and Ridge and also the Blue Ridge Provinces of the United States (Williams, 1976). To the northeast, the Humber Zone may continue as the Hebrides Zone of the British Caledonides (Williams, in press) where Lewisian basement of the Scottish Highlands is overlain by siliciclastics (the Torridonian) and then Cambro-Ordovician carbonates. The Moine/Dalradian area of the southeastern Highlands and equivalent rocks in Ireland are similar in age, lithology, metamorphic grade, and structural style to the Fleur de Lys Supergroup of Newfoundland (Williams *et al.*, 1974).

IGCP Project 27, Canadian Contribution No. 6.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).



Figures 18.1A-D. Geologic map of the Appalachian Orogen through the Southern and Central Appalachians (1A), Northern and Canadian Appalachians (1B, 1C) with extension across Irish and British Caledonides (1D). Map provided by Professor Harold Williams, modified from Williams and King (1977) and Williams (in press).

Figure 18.1A. Southern and Central U.S.A. Appalachians.

Figure 18.1B. Northern U.S.A. and Canadian Appalachians (part). Joins Figure 18.1A. Legend on page 112.

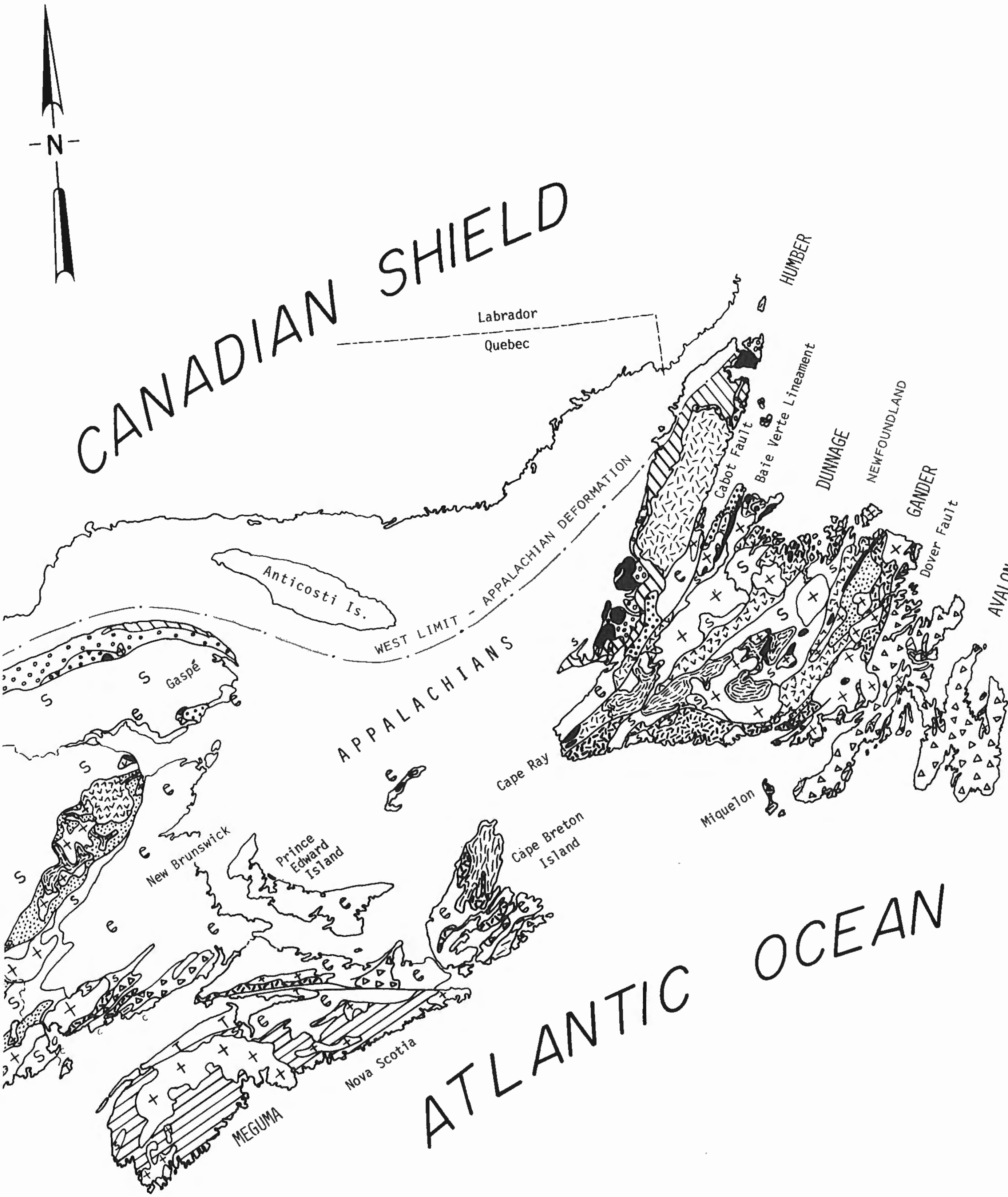


Figure 18.1C. Canadian Appalachians (part). Overlaps Figure 18.1B. Legend on page 112.

Figure 18.1D. Irish and British Caledonides. Legend on page 112.

The Humber Zone is characterized by a thick assemblage of geoclinal sediments with Grenville inliers (Figs. 18.1, 18.3) (Williams and King, 1977). Tectono-stratigraphic units which may be followed from Newfoundland to Georgia are:

1. a miogeoclinal, Cambro-Ordovician carbonate succession (Rodgers, 1968);
2. crystalline inliers of Late Precambrian (Grenville) age;
3. thick Late Precambrian and Early Paleozoic eugeoclinal siliciclastics;
4. transported slices with associated mélanges;
5. ophiolite suites and small ultramafic plutons; and
6. a lineament called the Baie Verte-Brompton Line of Newfoundland and Québec (Williams and St. Julien, 1978), which may reappear as the Brevard Zone of the Southern Appalachians (Williams and King, 1977).

In Newfoundland, the Humber Zone (Figs. 18.1, 18.3) consists of continental basement flanked to the southeast by eugeoclinal, and to the northwest by miogeoclinal sediments and transported units. The basement consists of approximately 1000 Ma Grenvillian rocks cut by northeasterly trending mafic dykes. This complex is locally overlain by plateau basalts and a succession of upwardly maturing siliciclastic sediments. To the southeast, the thick, eugeoclinal sediments (western part of the Fleur de Lys Supergroup) began to be deposited in the Late Precambrian. To the northwest of the basement complex, miogeoclinal, Cambro-Ordovician carbonate-bank complexes are overlain by southeasterly derived siliciclastics, mélange, and a variety of transported slices capped by ophiolite suites. The main structures of the zone are of Early to Middle Ordovician age (Taconian) (Fig. 18.8); however, easterly parts were also involved in the Devonian (Acadian) (Fig. 18.8) and Carboniferous (Alleghanian) deformations. Williams (1977) suggested that ocean-floor ophiolite complexes (e.g. the Bay of Islands Complex - Fig. 18.7) slid westwardly first over the initially undisturbed eugeoclinal Fleur de Lys Supergroup, then the basement complex, and finally the miogeocline to the west. In turn, tectonic slices, conceivably from the eugeocline, were thrust westwardly as far as the miogeocline during the Middle Ordovician (Fig. 18.8) (Williams *et al.*, 1974). In general, the eugeocline experienced intense polyphase deformation, granitic to ultramafic intrusion, and metamorphism to the amphibolite facies. These tectonic events were of Middle Ordovician age in the west, but possibly Siluro-Devonian in the east (Williams, 1977). Bursnall and de Wit (1975) believed that final metamorphism and deformation did not occur until the Devonian. The eastern margin of the Humber Zone in Newfoundland is drawn at the Baie Verte Lineament (Fig. 18.1), a steep structural belt marked by deformed ophiolites and mafic volcanic rocks, and interpreted as the ancient continental margin - ocean interface (Williams and King, 1977).

In Québec, the autochthonous succession of western Newfoundland (i.e. the Grenville basement and Cambro-Ordovician carbonates and siliciclastics) is hidden almost entirely under extensive structural cover of thrust slices transported during the Taconian Orogeny (Fig. 18.9). The Lomond Zone equivalent (the Cloridorme Zone - Fig. 18.3) consists of autochthonous and parautochthonous shelf, flysch, and regressive successions west of Québec City. The Hampden Zone of Newfoundland may appear as imbricated thrusts of shelf and flysch units in the "external domain" of St. Julien and Hubert (1975). The Fleur de Lys equivalent may by Lower Cambrian clastic carbonates, Cambrian siliciclastics, Upper Cambrian to Lower Ordovician shale and carbonate conglomerates, and Middle Ordovician shale and carbonate that now occur in nappe structures south of the St. Lawrence River from Gaspé to beyond Montréal. The emplacement of these nappes started in the early Middle Ordovician, and they were in turn folded by the end of the Middle Ordovician (St. Julien and Hubert, 1975) (Fig. 18.9). These transported sedimentary slices emplaced during the Middle Ordovician Taconian Orogeny are the dominant structural feature of the Québec Appalachians. The southeastern limit of the Humber Zone in Québec is drawn at the Brompton Line - a continuation from Newfoundland of the Baie Verte Lineament (Williams and St. Julien, 1978).

2. Dunnage Zone - A scar of the Early Paleozoic Atlantic (= eastern part of the Fleur de Lys (C); Notre Dame (D); Exploits (E); Botwood (F); Miramichi).

The Dunnage Zone records the development of the Early Paleozoic (Cambrian through Middle Ordovician) oceanic crust and island arcs.

The zone is characterized by its assemblage of relatively weakly deformed and metamorphosed remnants of oceanic plate, island arcs, and subduction-related mélanges. Figure 18.1 shows that the zone can be traced in a discontinuous manner from the southern United States, through Atlantic Canada and Ireland to Britain as the Dundee Zone (Williams, 1976 and in press; Williams and King, 1977).

In Newfoundland, the Dunnage Zone consists of Cambro-Ordovician, mainly mafic, volcanics with cherts, slates, greywackes, and minor carbonates. These are stacked in a shoaling succession from lowermost mafic dykes, gabbros, and pillow lavas upward through marine cherts and turbidites into pyroclastic and volcaniclastics capped by carbonate and subaerial tuffs. The western margin (Notre Dame Zone) is marked by ophiolite suites - e.g. the Betts Cove complex (Fig. 18.8) which presents a slice through the Early Paleozoic oceanic crust. Mattinson (1975) claimed that this ophiolite was formed perhaps 45 Ma (at 463 Ma - Late Ordovician) after the Bay of Islands ophiolite (at 508 Ma - earliest Ordovician). The central area (Exploits Zone) includes the Dunnage Mélange (Kay, 1976) which may represent an ancient oceanic trench-slope deposit to the east of an island arc. This mélange contains Tremodocian matrix and Arenig clasts, which according to Hibbard *et al.* (1977), suggests slumping from a trench slope. The eastern margin (Botwood Zone) is marked by discontinuous mafic-ultramafic plutons. Ages of volcanic rocks in the Dunnage Zone are 440-450 Ma (Ordovician-Silurian boundary) or older which is the same time limit as for the emplacement of the west Newfoundland allochthons. Although the overlying Silurian rocks are markedly different in lithology (as described below), the Acadian Orogeny produced the sole penetrative deformation (Williams *et al.*, 1974; Kennedy, 1976). The Dunnage Zone pinches out in southwestern Newfoundland where only a narrow belt of mylonite (Cape Ray Suture) separates the Humber from the Gander Zone.

In the Québec Appalachians, the Dunnage Zone appears as an assemblage of ophiolite, shale mélange, slate, sandstone, tuff, and calcalkaline volcanics that crop out south of Québec City. The ophiolite may represent the old, pre-Middle Ordovician oceanic crust. This assemblage was thrust westward to be imbricated along major thrust faults in one of the last spasms of the Taconian orogeny (St. Julien and Hubert, 1975) (Fig. 18.9).

In northwestern New Brunswick, Botwood Zone equivalents appear to be present (Rast *et al.*, 1976a). A similar lithologic but more complicated succession includes a basal ophiolite sequence (Pajari *et al.*, 1975) overlain by a mélange-turbidite complex of presumed

Early or Middle Ordovician age, then pillow lavas and more turbidites, capped by Silurian rocks (Fig. 18.10).

3. *The Gander Zone* (= G) - The Continental Rise of the Avalon Microcontinent.

The Gander Zone consists of Precambrian basement overlain by a eugeoclinal wedge whose sediments were derived from the Avalon Zone to the southeast.

The zone is a polydeformed, pre-Middle Ordovician, eugeoclinal mass of sediments resting on Precambrian gneissic basement. The complex was deformed and intruded by distinctive garnetiferous granites at 570 Ma, during the Ganderian (formerly Avalonian) Orogeny (Kennedy, 1976). To the southwest, the eugeoclinal sediments may reappear in the Miramichi Zone of New Brunswick (Fig. 18.3) and the gneissic basement complexes of Maine and possibly Virginia and North Carolina (Charlotte Belt) (Fig. 18.1); the distinctive garnetiferous granites may be followed to Florida (Kennedy, 1976). Transatlantic, the Gander Zone may continue as the Greenore Zone of southern Ireland and England (Fig. 18.1) (Williams, in press).

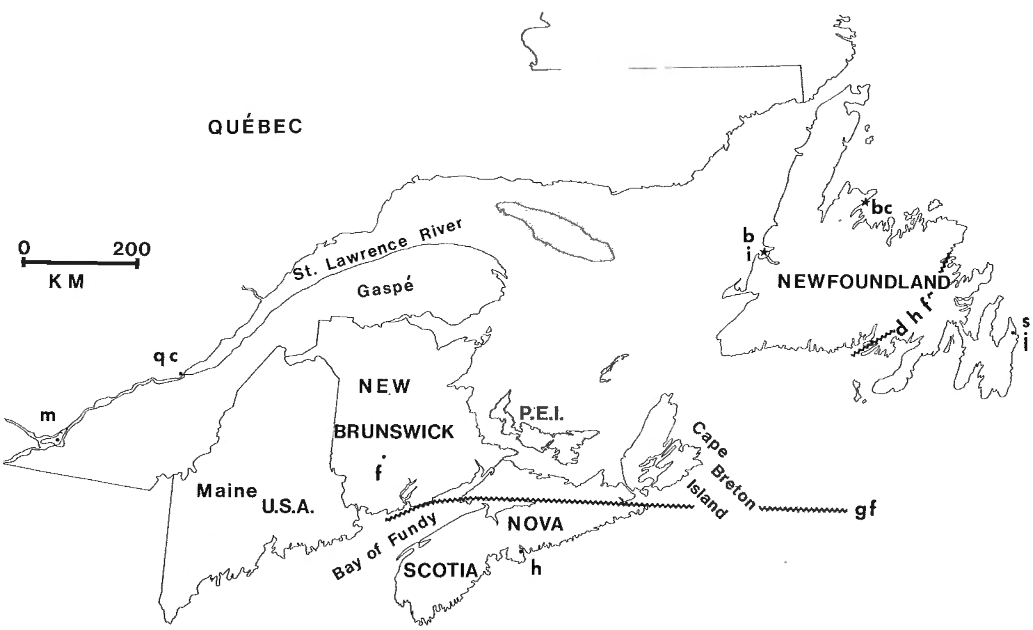
In Newfoundland, the Gander Zone consists predominantly of Late Precambrian siliceous turbidites, the Gander Lake Group, which is similar in many respects, except for time of deformation, to the Fleur de Lys Supergroup - its mirror-image in western Newfoundland. These eugeoclinal sediments of the Gander Zone are intruded by 570 Ma (Early Cambrian) granites which are important in establishing the Ganderian Orogeny (Kennedy, 1976). However, Bell *et al.* (1977) have established ages of 420-355 Ma (Acadian — Early Devonian through Early Carboniferous) for at least some of these granites. They find very little evidence for either Precambrian plutonism or for Precambrian deformation in the Gander Zone. Moreover, Strong *et al.* (1974) interpret deformation of the eugeoclinal succession to metamorphic activity at the plate margin of an easterly dipping subduction zone. To the southeast, the Gander Zone is separated from the Avalon Zone by the Dover-Hermitage fault (Figs. 18.1, 18.2), with movement indicated during the Hadrynian and again in the Devonian (Blackwood and O'Driscoll, 1976). A pluton that straddles the zone boundary sets a minimum age of 345 Ma (Viséan) for both the juxtaposition of the two zones, as well as for the age of metamorphism (Bell *et al.*, 1977).

In northeastern Nova Scotia, the Gander Zone has been extended across northern Cape Breton Island (Neale and Kennedy, 1975). Here, these rocks (Fourchu? Group of Wiebe, 1972) rest on marble, gneiss, and schist of the George River Group (Milligan, 1970). The George River and equivalent in New Brunswick, the Greenhead Group, are the oldest rocks in the area, and are believed to be the base of the Avalon Zone in general.

The Gander Zone may continue into north-central New Brunswick as the Miramichi Zone. Polydeformed, moderately metamorphosed, Ordovician turbidites and andesitic to rhyolitic volcanics overlie quartz arenites with Arenigian fossils (Neuman, 1972), which are above schists, and gneisses. Garnetiferous granites intrude the complex. Rast *et al.* (1976b) interpret the Miramichi to be an ensialic volcanic arc. However, Fyffe (1977) advises caution because the Miramichi Zone differs significantly from the Gander of Newfoundland. Most importantly no certain Precambrian continental basement is exposed, the paragneiss being the result of a Paleozoic metamorphic event, and intruded granites are Early Ordovician.

4. *Avalon Zone* (= H) - The Remnant of a Microcontinent

This was a late Precambrian and Early Paleozoic microcontinent. The Avalon Zone is characterized by Late Precambrian volcanic and sedimentary rocks cut by granites and overlain unconformably by gently dipping



m - Montréal	bi - Bay of Islands
qc - Québec City	bc - Betts Cove
f - Fredericton	gf - Glooscap Fault
h - Halifax	dhf - Dover-Hermitage Fault
P.E.I.- Prince Edward Island	sj - St. John's

Figure 18.2. Geographic map of the Canadian Appalachians showing localities mentioned in text.

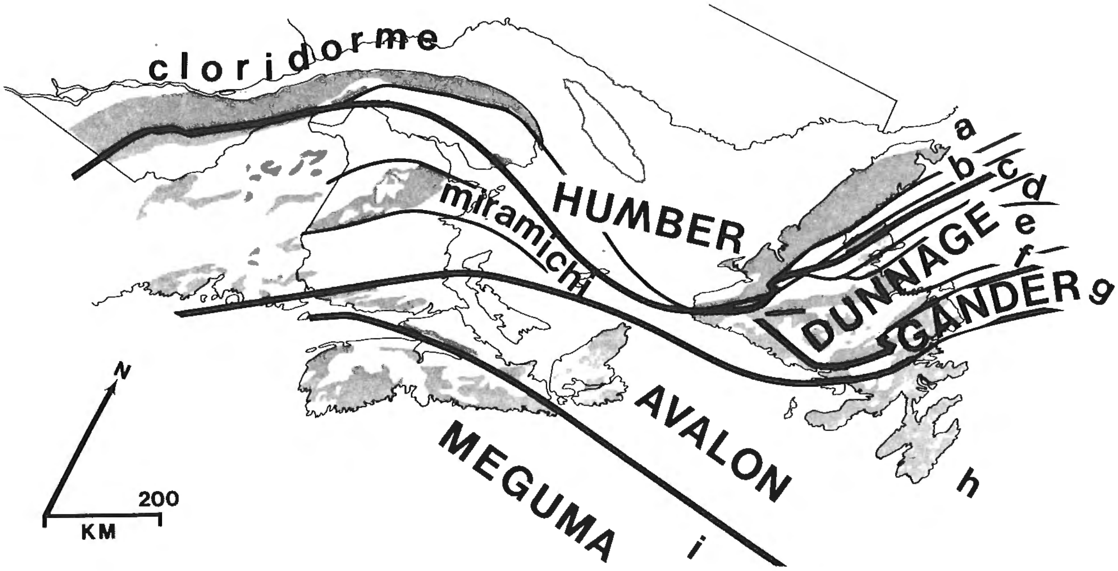


Figure 18.3. Distribution (patterned) of Precambrian, Cambrian, and Ordovician rocks in the Canadian Appalachians (modified from Figure 1 in Williams and King, 1976). Capital letters and thick lines designate the five major zones of Williams (1976). Lower-case letters and thin lines designate Cambro-Ordovician zones in Quebec and New Brunswick (after Rast *et al.*, 1976b), and Newfoundland zones: a - Lomond; b - Hampden; c - Fleur de Lys; d - Notre Dame; e - Exploits, f - Botwood; g - Gander; and h - Avalon, (after Williams *et al.*, 1974). i - Meguma Zone.

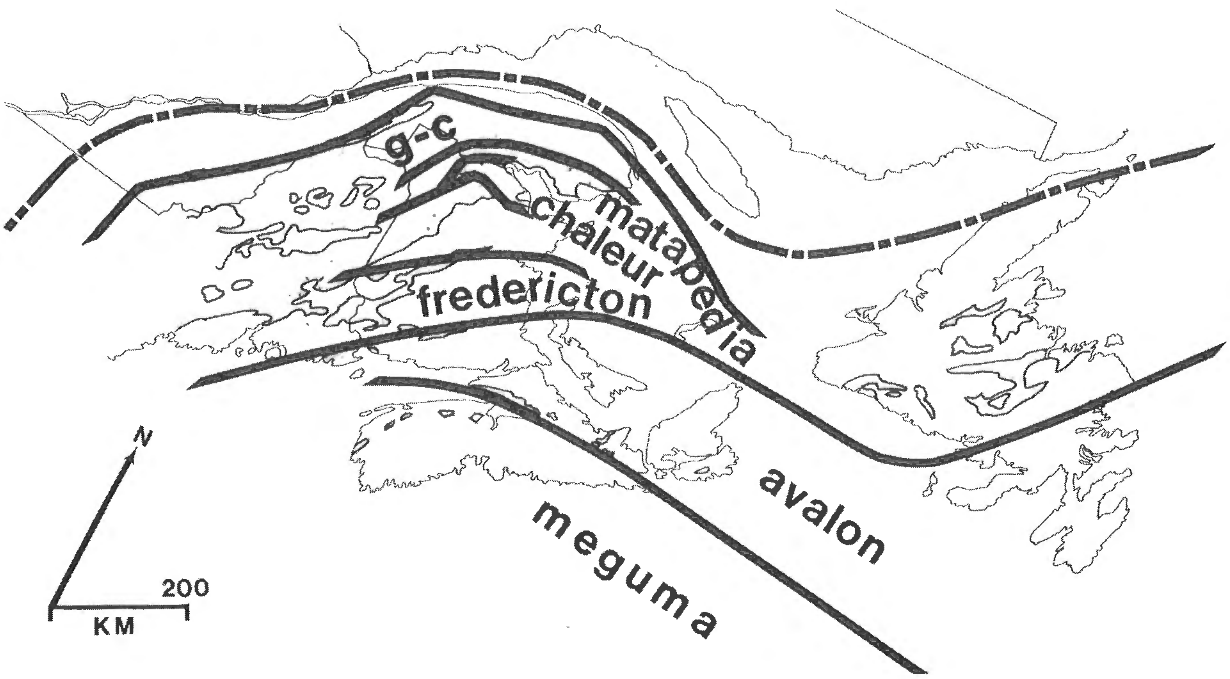


Figure 18.4. Distribution (patterned) of Silurian and Devonian rocks and tectono-stratigraphic zones in the Canadian Appalachians. g - c = Gaspé-Connecticut Zone. The heavy, broken line along the top of the figure marks the northwestern limit of Appalachian deformation. Distribution of rocks is modified from Figure 18.1.

Cambro-Ordovician shallow-water sediments containing Acado-Baltic faunas. Figure 18.1 shows that the zone can be traced southwestward to Massachusetts and even farther south to the Carolina Slate Belt of Virginia and the Carolinas, and northeastward to Wales as the Anglesey Zone (Williams, in press). The northern part of Britanny in northwestern France is a further extension.

In Newfoundland, the Avalon Zone consists of Late Precambrian volcanic and sedimentary rocks overlain by shallow-water, Cambro-Ordovician sediments. The late Precambrian succession starts with Hadrynian subaerial volcanics and minor sediments, and continues with a flyschoid facies including minor volcanics, a tillite layer (Brückner and Anderson, 1971), as well as oceanic tholeiite (Strong, in prep.). There is next an angular unconformity which indicates the Ganderian (formerly Avalonian) Orogeny. Above the unconformity are late Precambrian rocks of molasse-type facies, followed by shallow-water Cambrian shales and Lower Ordovician sandstones and shales with oolitic hematite (Rast *et al.*, 1976a). The basement of the Avalon Zone is unknown in Newfoundland but it may be present in Nova Scotia and New Brunswick (Kennedy, 1976).

The successions in the Avalon Zone of Cape Breton Island and southern New Brunswick are remarkably similar. In both areas, the succession consists of marble, gneisses, and schists overlain by volcanic rocks. In New Brunswick, the marble (Green Head Group) has been dated as Neohelikian or older (Hoffman, 1974) and the overlying volcanics (the Coldbrook Group) at 750 ± 30 Ma (Cormier, 1969). These volcanics are mainly acidic and characterized by felsic flows and subaerial tuffs in the western two-thirds of the area, and intercalations of marine arkose, siltstone, and conglomerate in the eastern third (Giles and Ruitenberg, 1977).

5. *The Meguma Zone* (= I) - Western Margin of Gondwana. The Meguma Zone (Fig. 18.3) is characterized by the Cambro-Ordovician Meguma Group - a eugeoclinal complex with no exposed basement. The group is overlain conformably by relatively thin Silurian and Lower Devonian paralic sediments, and intruded by large Devonian and Early Carboniferous granitic plutons.

This zone is interpreted by the author as a continental embankment, probably of Gondwana. The Meguma cannot be traced either southwestward or northeastward.

B. Middle Paleozoic (Siluro-Devonian) Zones

Tectono-stratigraphic zones evident during Silurian and Devonian times are six in number, two of which (the Avalon and Meguma Zones) continue from the earlier framework. The remaining four zones (the Gaspé-Connecticut, Matapedia, Chaleur, and Fredericton Zones) (Fig. 18.4) are smaller in size than the older ones. Perhaps as a consequence of this, as well as the complexities of the Devonian Acadian Orogeny, the Siluro-Devonian zones can be traced only in part to the southwest of New Brunswick and not at all to the northeast.

C. Late Paleozoic (Carboniferous-Permian) Zones

The Carboniferous-Permian tectonic framework is relatively simple (Fig. 18.6). The Fundy Basin contains a number of horsts that in part supplied large volumes of sediments to intervening, intermontane basins. Relatively thin sedimentary units extended both to the northwest and southeast. As discussed below, the Fundy Basin has been interpreted by various workers as a rift-zone, a shear zone, and an aulacogen. Interestingly, the southern border of the basin was reactivated during the Mesozoic to preserve a Late Triassic succession (Fig. 18.1); today, much of this area is again flooded by the sea.

RADIOMETRIC AGE DATES

The geochronologic scale is being changed to conform with the decay constants selected by the I.U.G.S. Subcommission on Geochronology (Steiger and Jaeger, 1977). The chosen constants are

$$\lambda\,(^{87}Rb) = 1.42 \times 10^{-11} y^{-1},\ \lambda\,(^{40}K_{\beta^-}) = 4.962 \times 10^{-10} y^{-1},$$

$$\text{and } \lambda\,(^{40}K_e) + \lambda^1(^{40}K_e) = 0.581 \times 10^{-10} y^{-1}.$$

The resulting geochronologic scale sets beginning ages (in Ma) for the following units: Hadrynian 900; Cambrian 595; Early Ordovician (Tremadocian) 512; Middle Ordovician (Llanvirnian) 475; Late Ordovician (Caradocian) 465; Silurian (Llandoverian) 444; Early Devonian (Gedinnian) 420; Siegenian 415; Emsian 399; Middle Devonian (Eifelian) 395; Late Devonian (Frasnian) 385; Early Carboniferous (Tournaisian) 370; Late Carboniferous (Namurian) 340; Permian 290 (calculated by Dr. J.D. Keppie, Leader, Deformation Working Group, Canadian IGCP Caledonian Orogen Project). This scale will be used in this paper; however, most references do not give sufficient data to recalculate dates of individual samples. Such a recalculation would certainly be within the error range and well within five per cent of the given date (Dr. M. Zentili, pers. comm.). Certainly dates cited in the older literature should be used with caution, as well as those in this paper which were calculated by constants different than those of Steiger and Jaeger (1977). Where an isotopic date is not given here, the age of the unit has been determined by biostratigraphy and/or field relations.

GEOLOGICAL EVOLUTION

LATE PRECAMBRIAN (>1200-595 Ma)

Highly metamorphosed, presumably Helikian, miogeoclinal assemblages are exposed in the Avalon Zone of eastern Newfoundland, Cape Breton Island, and southern New Brunswick. Deformation was during continental collision approximately 1000 Ma ago, and widespread magmatism may be due to reactivation of an older basement (Dewey and Burke, 1973). At this time, a supercontinent may have been assembled, consisting of Africa, North America (with Grenville Orogen), Fennoscandia (with the Dalslandian Orogen), and Western Europe (with the Pentevrian Orogen of Brittany) (Schenk, 1971, Stewart, 1976). In general, Strong (in prep.) interprets the overlying Hadrynian succession of the North Atlantic region as representing a long-lived, continental extension leading to eventual, large-scaled rupture and the development of the Paleozoic Atlantic.

In both Nova Scotia and southern New Brunswick, Hadrynian rocks, dominantly subaerial volcanics with minor sediments, are lithologically similar and presumably broadly equivalent in age to Hadrynian units of the Avalon Peninsula of Newfoundland. Was the paleoenvironment one of compression with subduction and volcanic arcs, or one of tension with rifting, and ocean-floor spreading? Murphy (1977) interpreted the Nova Scotian equivalent (the Fourchu Group) as a Japan-type island arc, as did Dewey (1969), Helmstaedt and Tella (1973), and Rast *et al.* (1976a). Giles and Ruitenberg (1977) interpreted the New Brunswick equivalent (the Coldbrook Group) to deposition along the margin of an intracratonic basin (or possibly continental margin) over an intracratonic rift zone. Rast *et al.* (1976a) interpreted the Coldbrook and Fourchu, and other Hadrynian volcanics from Carolina to the British Isles, as an ensialic island arc succession along the flank of the Avalon microcontinent. Schenk (1971) considered the Hadrynian volcanics of the Avalon to have formed in a rift zone produced by the initial separation of the microcontinent during post-Grenville rifting of North America and Africa - i.e. during the earliest stage of the development of the Paleozoic Atlantic. In Newfoundland, Papezik (1972) interpreted the volcanics there as due to block faulting related to rifting, although Hughes and Brückner (1971) interpreted the same rock as of island-arc origin.

In southeastern Newfoundland, Hadrynian flyschoid units conformably above the volcanics may be equivalent to the bulk of the Gander Group to the west. The latter may be related to subduction before the proposed Ganderian (Avalonian) Orogeny (Kennedy, 1976). The flyschoid sediments include tillite (Brückner and Anderson, 1971) which may also occur farther west in the Fleur de Lys Group (Church, 1969). King (1977) interpreted the paleogeography of the Avalon as belonging to a partly marine, basin-and-range domain within a wide distentional belt. Marine sedimentation of the flyschoid deltaic units around a volcanic terrane led to eventual infilling of the basin. Intermittent deposition of volcanic ash was followed by alluvial fans issuing from a rising mountain front to the north. The Late Hadrynian and Cambro-Ordovician environments were tidal to offshore. Strong (in prep.) notes the occurrence of continental alkali basalts and oceanic tholeiites within flyschoid sediments of the Avalon Zone. He suggests that the long-lived, Hadrynian continental extension led to local rupture and small seafloor spreading of several presently separated areas of the North Atlantic area - e.g. Anglesey, Lizard Peninsula, southern Brittany, the Iberian Peninsula, and northwest Africa. These areas may have been parts of the late Precambrian supercontinent mentioned above.

In southeastern Newfoundland, evidence for an Early Cambrian orogeny is disputed. An angular unconformity above, and metamorphism of older units below, were interpreted by Kennedy (1976) as caused by the Ganderian Orogeny, whereas Strong attributes them either to activity related to a southeasterly dipping subduction zone (Strong *et al.*, 1974), mild compression, or to foundering and collapse of the thick basaltic pile of underlying units (Strong, in prep.). Foliated granitic plutons that cut the Gander Group are significant in this argument. On field evidence, Kennedy (1976) related them to 573 Ma granites which intruded the eugeoclinal Gander Group and were foliated by the Ganderian Orogeny. Bell *et al.* (1977) have dated some of these foliated granites at 420 to 345 Ma old (i.e. Devonian-Early Carboniferous) and concluded that there is very little evidence for either Precambrian deformation or for Precambrian plutonism, and that the Ganderian Orogeny must be significantly younger than Hadrynian. Rast *et al.* (1972a) agreed that within the Avalon Zone there is little evidence of deformation, but that tectonic activity was intense both northwestward in a behind-the-arc basin, and also southeastward (the Cadomian event of Brittany, and possibly also Pan African, Rifian, and Demaran orogenies). Strong counters that the irregular timings may not indicate any major continental collisions. Overlying redbeds with subaerial, acidic pyroclastics and basalt/rhyolite were attributed by Kennedy (1976) to post-orogenic molasse after the Ganderian Orogeny, and by Strong (in prep.) to continued Hadrynian tension. Diabase dyke swarms dated at 600 Ma (Stukas and Reynolds, 1974) with tholeiitic lavas signal ultimate rifting and the birth of the Paleozoic Atlantic.

CAMBRIAN

In the Québec Appalachians, siliciclastics from the Canadian Shield were carried across a presumed carbonate miogeocline to build a thick eugeocline of turbidites, conglomerate, and periodically, carbonate breccias (Fig. 18.9) (St. Julien and Hubert, 1975). Trilobites in carbonate clasts are of North American affinity. In the Humber Zone of northwestern Newfoundland, a somewhat similar geoclinal assemblage with the addition of the preserved carbonate miogeocline, again has North American trilobites. There, the succession passes southeasterly into a presumed island-arc complex of basic and acidic volcanics and oceanic crust (Strong, 1978). In latest Cambrian or earliest Ordovician time, the Newfoundland section underwent orogenic deformation producing recumbent-fold complexes and moderate-grade metamorphism (Kennedy, 1976). Deformation and metamorphism decrease southeasterly into the island-arc volcanics (Poole, 1976). Small ultramafic bodies emplaced early in the orogeny may be related to subduction (Stevens *et al.*, 1974).

The Avalon Zone was platformal in character through the Cambrian, and contains Acado-Baltic faunas. Thin, diachronous, paralic quartz meta-arenites record flooding by the Cambrian sea. Overlying lithologies are shales and thin stromatolitic carbonates (Poole, 1976).

CAMBRO-ORDOVICIAN: THE MEGUMA GROUP

The Meguma Group (Fig. 18.10) forms the bedrock of almost all of Nova Scotia south of the Glooscap Fault, as well as the foundation of the offshore Scotian Shelf (King and MacLean, 1976) - a total area of approximately 125×10^3 km^2. The thickness of the Group may exceed 14 km, and if folding and strain are considered, the volume of the Meguma may exceed 4×10^6 km^3 (Schenk, 1976; Zentilli and Schenk, in prep.). The group consists of quartz metawacke turbidites interstratified with black slate - the basal half (Cambrian) being mainly thick turbidites, the upper half (Ordovician) mainly black slate; however, the two lithologies are complexly intercalated.

Southeasterly derived quartz metawacke turbidites and slates of Cambrian age are overlain by thin turbidites and Tremadocian slate of a shoaling succession from abyssal plain to continental rise to outer shelf origin (Schenk, 1976). In northwestern Nova Scotia, an abrupt, vertical change in lithology occurs. Slates of the upper Meguma Group are overlain conformably by quartz arenites, paraconglomerates, and thick volcanics, all of Caradocian age or younger. Schenk (1972) interpreted the paraconglomerate as possible tillite related to Saharan glaciation; Lane (1978) traced the paraconglomerate for 230 km and attributed the paralic assemblage of quartz arenites and shales to glacio-eustatic changes in sea-level. If these sediments are glaciogenetic, they indicate not only proximity to northwest Africa, but also that the underlying slates of the upper Meguma Group encompasses the Ordovician System. If the Taconian Orogeny did not affect the Meguma Zone, one explanation could be that during the Late Ordovician, this zone was not yet part of North America. However, the presence of dropstones do not necessarily mean close proximity to an ice sheet - the seafloor off Newfoundland, today 1400 km from the nearest ice-sheet (Greenland), receives vast amounts of glacial erratics from floating icebergs.

Preliminary strain-analysis indicates that rotation of primary sedimentary structures during folding was considerably less than the paleocurrent variability inherent in the sedimentologic model (Zentilli *et al.*, 1977). Slates of the Meguma Group were regionally metamorphosed approximately 412-400 Ma ago (Early Devonian) Reynolds *et al.*, 1973, Reynolds and Muecke, in press), and intruded, after upright folding, by granitic plutons with cooling dates culminating at approximately 370-360 Ma.

Schenk (1970b, 1975) has interpreted the Meguma Group as a eugeoclinal complex of a deep-sea fan and overlying continental rise. The source area was a deeply eroded, metasedimentary-metaigneous terrane located to the present southeast (Schenk, 1970b). This source area was continental in dimension if the great volume of sediment is considered. The environment of deposition is envisioned as a passive, Atlantic-margin type, not unlike that of the present Hatteras fan.

During the Cambrian (? and latest Hadrynian), the Meguma Zone was a continental rise complex of deep-sea, coalescing fans (Schenk, 1975). The turbidites show large-scaled rhythms (Harris and Schenk, 1976). The source area to the southeast was presumably either the Saharan Shield (Schenk, 1972, 1975), Western Europe, or possibly the Guyanan Shield. That part of the Meguma Group not now against North America, as well as its shelf-equivalent, remains either in northwestern Morocco, in the southern or western Iberian Peninsula, in western France, or in western Colombia, as discussed below. One of my models for the entire Cambrian through Lower Devonian stack in the Meguma Zone is a continental embankment as defined by Dickinson (1974). Such a feature forms on a passive, Atlantic-type margin if clastic sediment from the rifted margin is so voluminous that progradation leads to a building-up of the continental rise, and a building-out of the continental slope. Three main depositional phases result (Burke, 1972): a basal phase of sandy turbidites (here, the Cambrian) deposited near the toe of the embankment; a middle phase of mainly shaly rocks (the Ordovician) deposited on the advancing frontal slope of the embankment; and an upper phase of mainly sandy, paralic strata (the Silurian-Lower Devonian) deposited along the prograding outer edge of the top of the embankment. The complex was affixed to the Avalon during the Acadian Orogeny. Oceanic volcanics and ultramafics are not known in the Meguma Zone.

ORDOVICIAN

(a) *Humber Zone*. This period saw the greatest plate activity in the Appalachians. In Québec, flysch deposition continued through the Early Ordovician, but by late Early Ordovician time, subduction resulted in volcanic-arc mélanges (Fig. 18.9). Allochthons were finally emplaced during the early Late Ordovician onto the Humber Zone. This, the Taconian Orogeny, is responsible for almost all of the deformation in Cambro-Ordovician rocks of the Québec Appalachians (St. Julien and Hubert, 1975).

In northwestern Newfoundland, a sub-Middle Ordovician unconformity marks at least 100 metres of uplift of the carbonate bank. The Cambro-Ordovician platform and its karstic unconformity is associated with "Mississippi-Valley type" zinc deposits. These characteristics can be followed from Norway, through Greenland, Scotland, and Newfoundland, to Alabama (Swinden and Strong, 1976). After this upwarp, the miogeocline sank, perhaps to oceanic depths, and was covered by flysch wedges from the east. The advent of flysch is diachronous: Early Ordovician in the east and early Mid-Ordovician in the west of Newfoundland, and late Middle Ordovician in Gaspé and New York State (Williams and Stevens, 1974). In Newfoundland, the flysch was bordered to the southeast by in turn: an ophiolite belt, an island arc, and mélange of a subduction zone (Poole, 1976). Tectonic slices were driven from the southeast, beginning in the Early Ordovician and culminating in the mid-Ordovician so that a completed stack of slices was emplaced on the Humber Zone in the final stages of gravity sliding, in which the rocks on each higher slice are stratigraphically older and more easterly derived than those below (Williams, 1975). Reefal carbonates of Middle Ordovician age cap the succession.

(b) *Dunnage and Gander Zones*. Middle Ordovician volcanics throughout these zones are mainly the result of island-arc activity (Poole, 1976). In northwestern Newfoundland, the 45 Ma difference in the ages of ophiolite or ophiolite-related rocks suggested to Mattinson (1975) that the present exposures of ophiolite did not originally form one continuous ocean-plate. Instead, he pictured a Late Cambrian, narrow rift to form the Bay of Islands ophiolite (Fig. 18.2 and 18.3) which later obducted during ocean-floor closing in the Early Ordovician. The Betts Cove ophiolite (Figs. 18.2 and 18.8) would have formed after Middle Ordovician rifting, and was in turn obducted during the Middle Devonian.

In northwestern New Brunswick, subduction is evident in a mélange-turbidite complex of presumed Early or Middle Ordovician age which is overlain by Upper Ordovician pillow lavas and turbidites (Rast *et al.*, 1976a). During the Late Ordovician, the island-arc complex was deformed and eroded both in New Brunswick and adjacent Maine. A subduction zone may have existed southeast of this deforming belt, and continued northeasterly across Newfoundland. In central New Brunswick and adjacent Maine, thick quartz arenite with a shelly fauna of Early Ordovician age suggests that this area may have been underlain by crystalline basement (Poole, 1967).

In general, by Middle Ordovician time, the Paleozoic Atlantic between the Humber and Avalon Zones was a narrow trough which from that time on became infilled with terrigenous detritus. Much of the deformation and magmatism now observed was the result of final closure which concluded by the end of the Carboniferous.

(c) *Avalon Zone*. The Lower Ordovician of southeastern Newfoundland is characterized by shallow-water siliciclastics. The northeastern part of mainland Nova Scotia exposes the enigmatic Browns Mountain Group. This unit is an Hadrynian pile of volcanic, pyroclastic, and sedimentary rock that is similar in succession to the Middle Cambrian Bourinot Group of southeastern Cape Breton Island and the Late Precambrian Harbour Main and Love Cove Groups of southeastern Newfoundland. The Browns Mountain was intruded by Hadrynian (582 Ma) and younger plutons. Benson (1973) interpreted the Browns Mountain as an Ordovician arc complex.

SILURO-DEVONIAN

In general, the Silurian rocks of Atlantic Canada are considerably different in lithology to those of the underlying Ordovician, and often the two systems are separated by an angular unconformity.

This general change in regime in Atlantic Canada is usually attributed to the Middle to Late Ordovician Taconian Orogeny. In central Newfoundland, Silurian conglomerates and continental redbeds overlie marine Ordovician strata. In the Gaspé-Connecticut Zone of the Québec Appalachians (Fig. 18.4), Siluro-Devonian volcanics and sediments unconformably overlie Ordovician and older rocks. The depositional and structural trend of this younger succession crosses earlier Ordovician structures, including the Brompton Line (Williams and St. Julien, 1978). In Gaspé and New Brunswick, the Matapedia Zone (Fig. 18.4) features oceanic floor overlain by Ordovician and Silurian calcareous turbidites that remained undisturbed until Early Devonian time (Rast *et al.*, 1976b). The Fredericton Zone (Fig. 18.4) has Siluro-Devonian mafic volcanics and turbidites unconformably overlying Ordovician black slates and pillow lavas which were severely deformed during Late

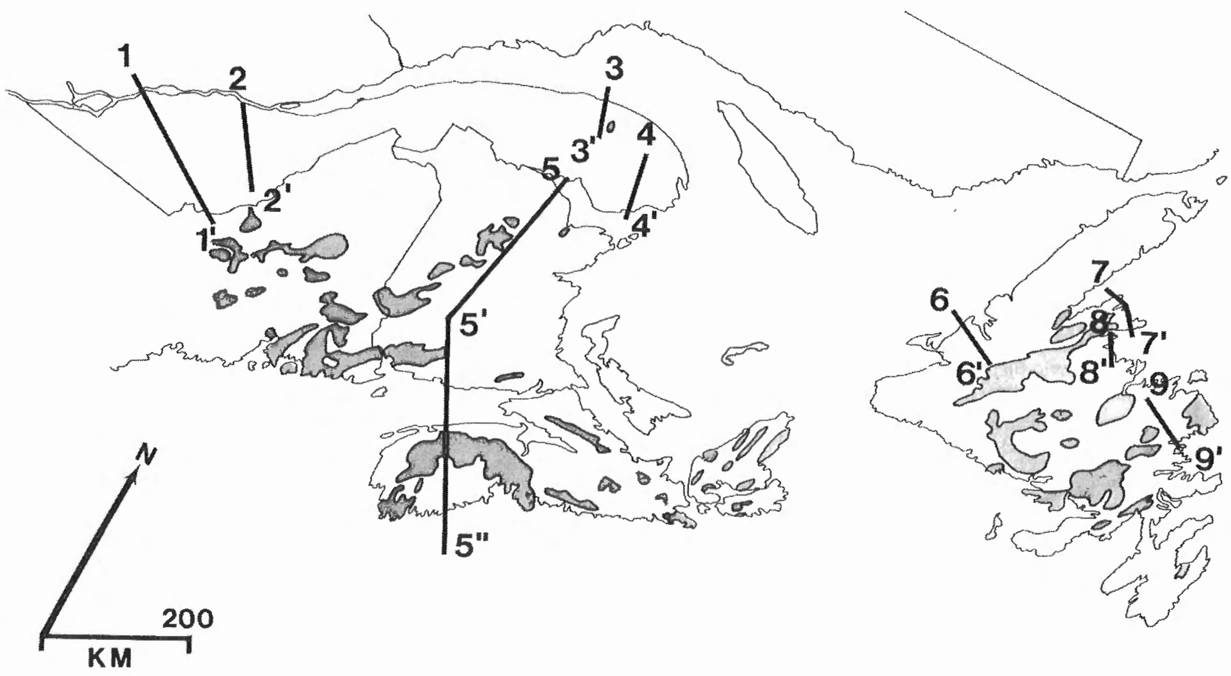


Figure 18.5. Distribution (patterned) of Paleozoic (mainly Devonian) granitic plutons in the Canadian Appalachians. Lines are positions of sections (Figures 18.7 - 18.10).

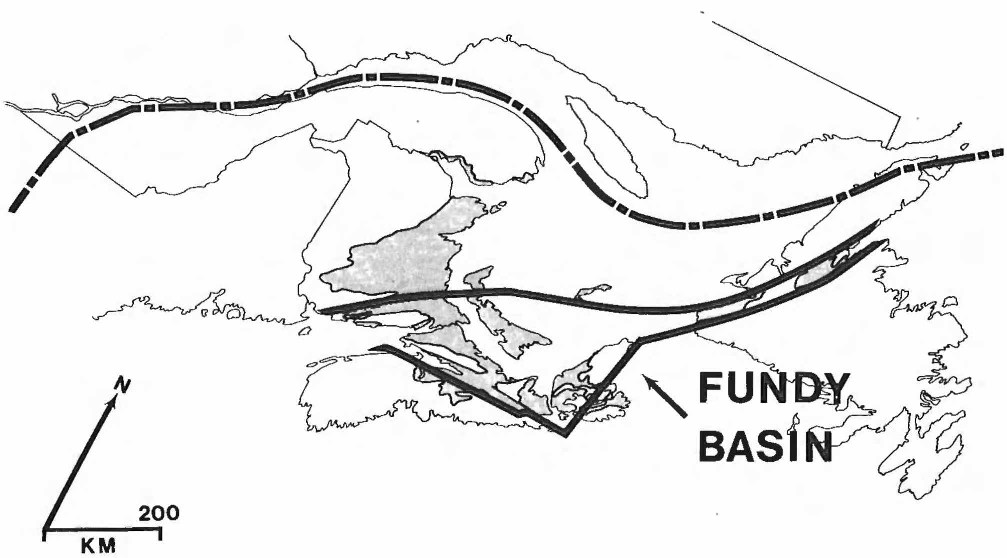


Figure 18.6. Distribution (patterned) of Carboniferous and Permian rocks in the Canadian Appalachians. The heavy broken line along the top of the figure marks the northwestern limit of Appalachian deformation. Extent of the Fundy Basin modified from Belt (1968).

Ordovician time (Helmstaedt and Tella, 1973). In the Avalon Zone of eastern mainland Nova Scotia, paralic sediments of the Arisaig Group, which includes almost the entire Silurian System, rest disconformably on Late Ordovician, in part subaerial, volcanics (Lane and Jensen, 1976). The Arisaig grades upward into Lower Devonian continental redbeds. Granting the existence of Taconian structural deformation, many of the stratigraphic, paleontologic, and sedimentologic changes during the Late Ordovician may be due to non-orogenic events.

In the Meguma Zone structural effects of this orogeny are obscure whereas stratigraphic changes are abrupt and profound (Fig. 18.10). In northwestern Nova Scotia, slates of the Cambro-Ordovician Meguma Group are overlain by latest Ordovician (?) and Silurian, paralic slate, quartz arenites, and paraconglomerates, and differentiated alkaline volcancis and volcaniclastic sediments (White Rock and Kentville Formations) (Lane, 1976 and 1978; Sarkar, 1978). Lane (1978) has traced the polymictic paraconglomerate over a distance of 230 km, usually within a 50 m-thick zone transitional into the underlying Meguma. Schenk (1972) has suggested that dropstone clasts in the paraconglomerate may indicate a glacial regime, possibly related to Saharan glaciation. Schenk (1972) and Lane (1978) have interpreted the units of the White Rock Formation and the overlying black Silurian slates (Kentville Formation) to glacio-eustatic changes in sea-level. The contact between the White Rock and the Meguma is conformable although western Nova Scotia does exhibit a single locality with an apparently angular unconformity which may be an expression of refracted cleavage. The basal paraconglomerate separating deep from shallow-water sediments could be evidence of the Taconian Orogeny; however, the clasts are in matrix-support, some are faceted and polished, and their lithologies are of metamorphosed sediments which would require very deep erosion. The finer-grained sediments in the base of the White Rock Formation are similar to the famous "microconglomerate" of Northwest Africa and Western Europe. In these areas, overlying Silurian black shales are usually attributed to deglaciation. This Ashgillian glacial episode was a very major event and must have had world-wide effect in lowering sea-level. The response in the Meguma Zone was an abrupt change from deep-water graptolitic black slate of the upper part of the Meguma Group to paralic sediments of the White Rock Formation. The presence of dropped pebbles and boulders may indicate some proximity to the Saharan ice sheet. The absence of obvious effects of the North American Taconian Orogeny may indicate that the Meguma Zone was not yet part of North America during the Ordovician.

Conformably above the black graptolitic shales of the Kentville Formation, the Lower Devonian Torbrook Formation consists of paralic sediments containing a very rich Acado-Baltic fauna (Jensen, 1976). The entire Cambrian through Lower Devonian succession was folded and then intruded by Early Devonian and Late Devonian to Early Carboniferous granites. The magma for these batholiths could have been produced either by partial fusion of an underlying crystalline basement (e.g. Avalon Zone), or through mixing of mantle and crystal melts (McKenzie and Clarke, 1975). Keppie (1977a) has noted some evidence for pre-Devonian, post-Tremadocian (probably Silurian) deformation in the Meguma Zone.

Silurian marine siliciclastics with some volcanics onlap the uplifted and eroded Canadian Appalachians from the allochthon belt of Québec and Newfoundland southeastward at least through the Avalon Zone of Nova Scotia. In the Newfoundland Dunnage Zone, sedimentation was continuous over most of the area and ranges from Ordovician flysch to Silurian shallow-marine to fluvial sediments with mainly subaerial calc-alkaline volcanism. Most of Newfoundland was probably land during Late Silurian and Devonian time (Poole, 1976).

The Fredericton Zone of New Brunswick received thick turbidites and volcanics of Silurian through Early Devonian age (Poole, 1976). The Ordovician-Silurian unconformity, both here and in central and northern New Brunswick, indicates the profound effect of the Taconian Orogeny (Rast *et al.*, 1976a). McKerrow and Ziegler (1972b) suggested that the Fredericton Zone was a Silurian—Early Devonian oceanic trench with marginal subduction zones. Andesitic volcanics of this age both along its northern margin (central and southern Gaspé and northern New Brunswick) and southern margin (southwestern New Brunswick, coastal Maine and Massachusetts) were related to subduction along the flanks of the trough. Poole (1976) suggested that the Fredericton is a relict of the proto-Atlantic; Rast *et al.* (1976b) suggested a graben-structure or behind-the-arc basin.

In the Avalon Zone of northern Nova Scotia, almost the entire Silurian System here is represented by 2 km of shallow marine (depth of water less than 6 m) to brackish-water siliciclastics (the Arisaig Group) deposited in a tropical or subtropical setting (Boucot *etal.*, 1974; Bamback, 1969). The section rests on Ordovician subaerial (in part) volcanics and is capped by Lower Devonian redbeds (Lane and Jensen, 1976). The richly fossiliferous, shallow-marine to paralic section on the Avalon contrasts strongly with the all-but unfossiliferous sediments of the same environment and age in the White Rock Formation of the Meguma Zone. A cold, hostile environment is suggested for the latter (Schenk, 1975).

Thick (at least one km), mainly subaerial, volcanic rocks centred in western Nova Scotia are of the alkaline-tholeiitic series (Sarker, 1977). Taylor (1967) interpreted them as products of an island arc, whereas Sarkar concluded that the volcanics were associated with neither spreading-ridge nor subduction-margin regimes, and suggested rifting of a thick, sialic basement. I suggest that the alkaline basalts and associated shoaling are due to hot-spot activity, with additional shoaling due to glacio-eustasy. The Meguma Zone appears to be a continental embankment of Gondwanaland (whether Morocco, Western Europe, or Columbia, as discussed below). During the Mesozoic at least, Gondwana (specifically Africa) was stationary over the mantle, and so hot-spot volcanicity was also stationary. Today Africa is again stationary, and a third of the world's obvious hot-spots are in the Africa plate - mostly on land or adjacent to the coast (Burke and Wilson, 1977). In 1971, Schenk suggested that a modern analogue to the Early Paleozoic Meguma Zone and Avalon microcontinent could be the Moroccan rise-shelf and the sialic microcontinent of the eastern Canary Islands. Today this microcontinent is intimately connected with two stationary hot-spots, in a manner analogous to the White Rock alkaline basalts.

In the Meguma Zone richly fossiliferous, paralic, Lower Devonian sediment accumulated conformably on Silurian graptolitic slate (Jensen, 1976). Intercalated redbeds both here and in the Avalon Zone of northern Nova Scotia may reflect uplift and orogeny immediately to the south. Basic sills of unknown age intrude the Lower Devonian of the Meguma Zone. The Cambro-Lower Devonian sediments were folded and then intruded by Devonian and Early Carboniferous granites. Curiously, neither allochthons or subduction zones have yet been recorded for the Acadian Orogeny, although Keppie (1977a) speculated that a subduction zone dipped southeasterly beneath the Meguma. Poole (1976) suggested that the crust beneath the Appalachian orogen thickened by shortening, and as a result supracrustal rocks were thrown up into upright folds and intruded in the Mid-Devonian by Late Silurian-Early Devonian magmas which produced volcanism of that age.

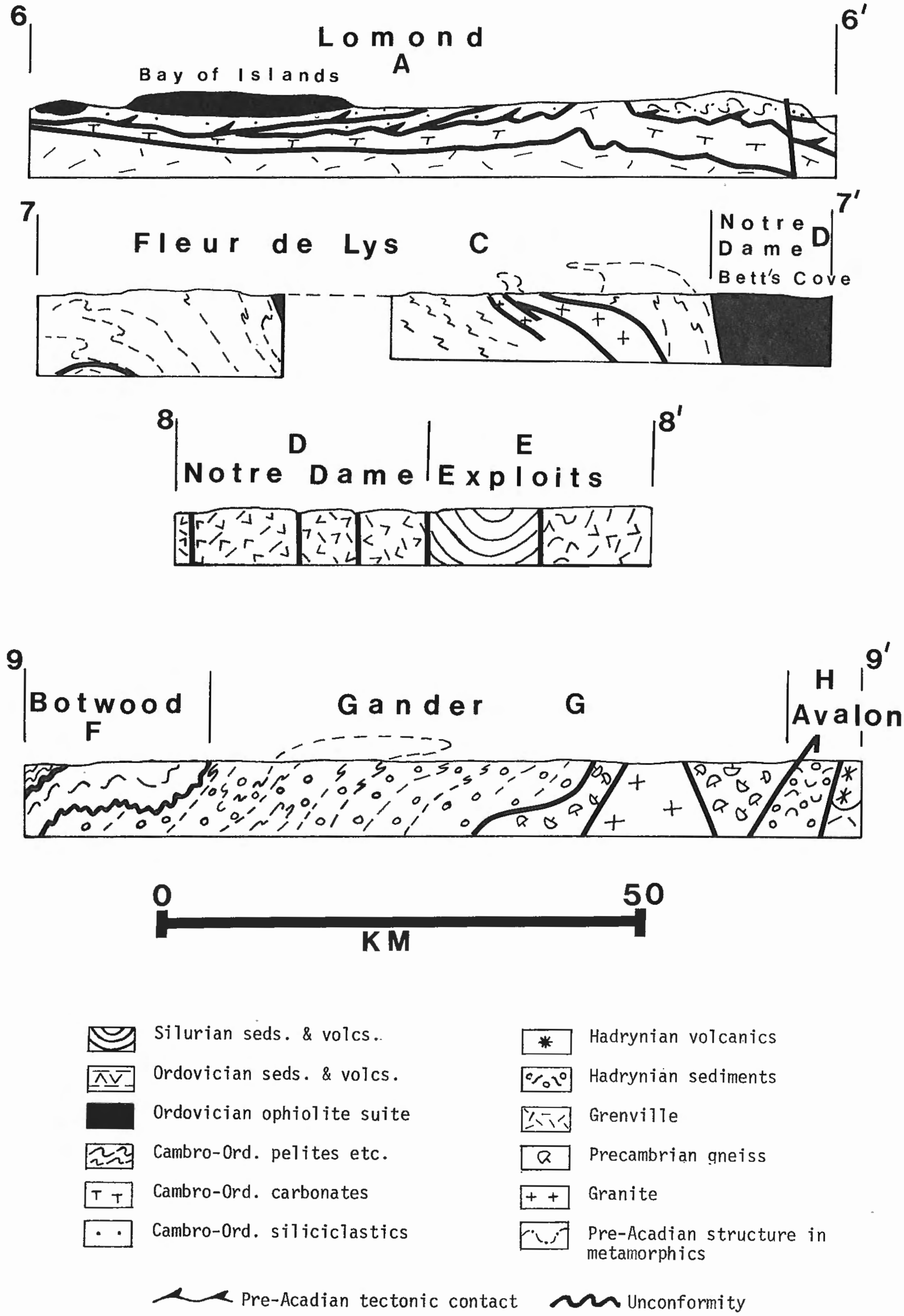


Figure 18.7. Sections 6, 7, 8, and 9 across Newfoundland (modified from Williams *et al.*, 1972). See Figure 18.5 for localities of sections.

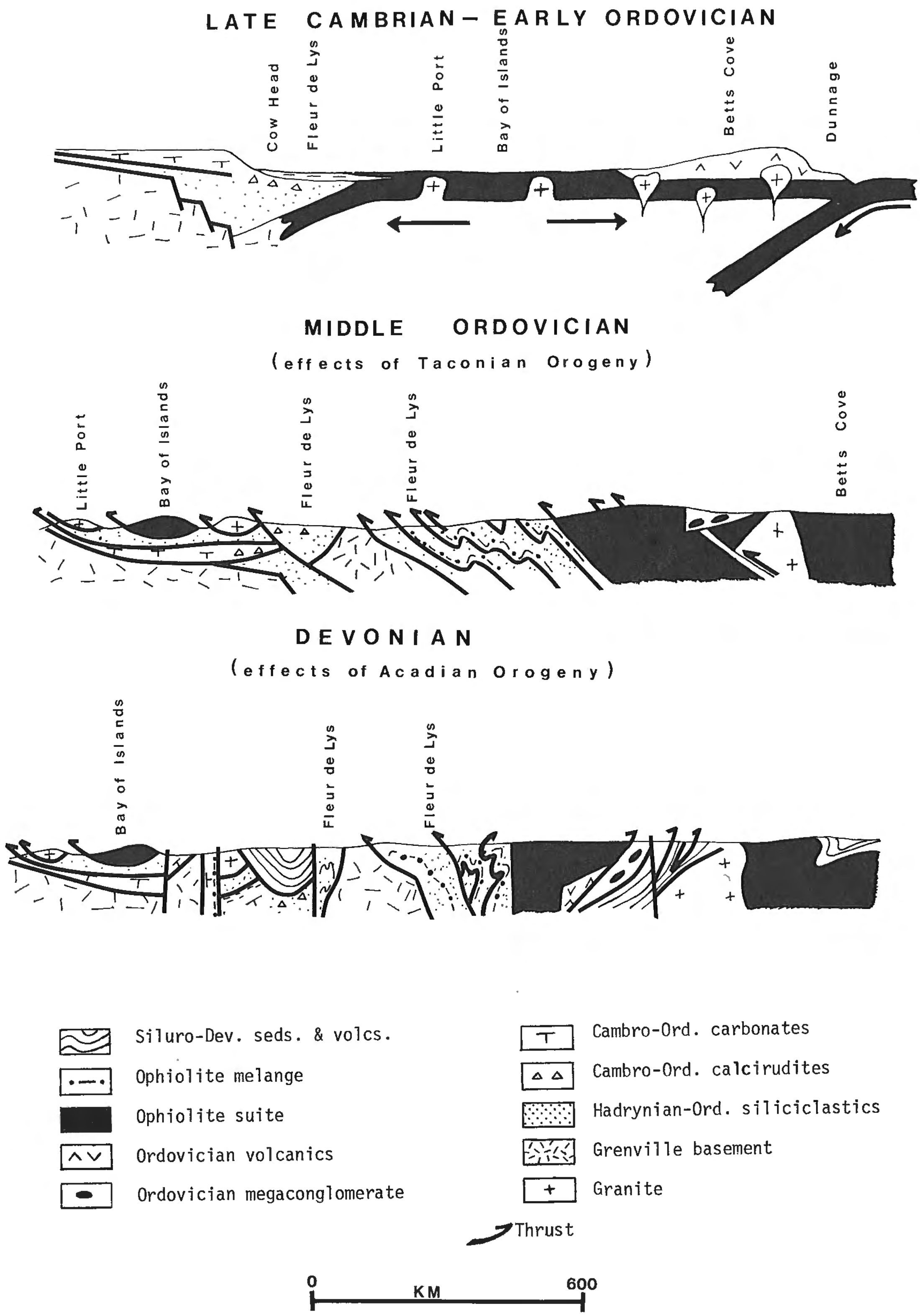


Figure 18.8. Schematic sections across northeastern Newfoundland to show the effects of the Taconian and Acadian Orogenies (courtesy of H. Williams).

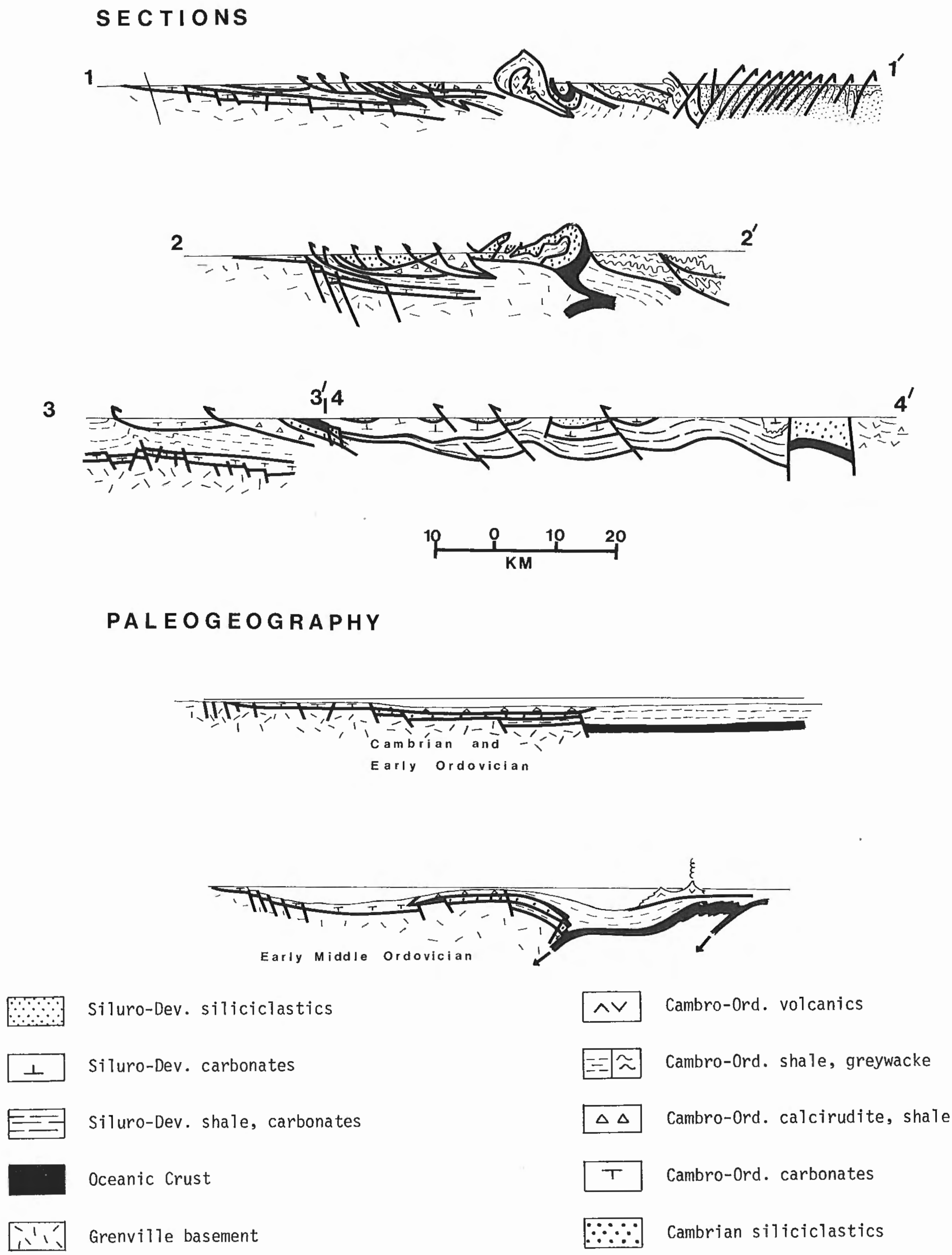


Figure 18.9. Sections 1, 2, 3 and 4 across the Québec Appalachians and interpreted paleogeography from Cambrian through early Middle Ordovician time (modified from St. Julien and Hubert, 1975). *See* Figure 18.5 for localities of sections.

The age given for the Acadian Orogeny varies in the Canadian Appalachians. This is due to several factors: conflicting evidence from different measures of orogenic climax; different sampling techniques and decay constants used in determining radiometric ages; the nature of collision between two irregular continents; and the necessity of closing both the seaway occupied by the former Dunnage Zone, and also that of the Meguma Zone.

In Newfoundland, stratigraphic and structural evidence indicates that the ocean which occupied the old Dunnage Zone was closed at least along its northwestern margin before Silurian time (Williams and St. Julien, 1978). Poole (1976) agreed that this ocean was almost closed during the Late Ordovician, but that complete closure did not occur until Late Silurian time. On paleontological grounds McKerrow and Cocks (1977) suggested that the Late Ordovician ocean was still wide enough to inhibit faunal migration. Boucot *et al.* (1964) maintained that closure occurred in the late Early Devonian. Granitic plutonism continued in Newfoundland through the Early Carboniferous (Bell *et al.*, 1977).

In New Brunswick, Rast and Stringer (1974) suggested a pre-Late Devonian age for the Acadian Orogeny, although the most intense deformation occurred in Early Devonian time (Rast *et al.*, 1976a). Donohoe and Pajari (1973) showed that the orogeny occurred earlier in the southeast and later in the northwest of the province. The last phase of folding, metamorphism, and intrusion in southwestern New Brunswick occurred between Gedinnian and Frasnian time as determined stratigraphically; 402 to 386 Ma (about Givetian) as determined radiometrically with $\lambda Rb = 1.42 \times 10^{-11} y^{-1}$ (Pajari *et al.*, 1974). In general, Naylor (1971) argued that the entire Early Devonian cycle of sedimentation, subsequent folding, igneous intrusion, and regional metamorphism occurred during the interval of 30 Ma.

Whereas closure of the old Dunnage Zone appears to have been during the Early Devonian in New Brunswick and somewhat before this time in Newfoundland, the age of the Acadian Orogeny in the Meguma and adjacent Avalon Zones can be set more precisely. The youngest pre-orogenic rock is the Torbrook Formation of Siegenian to possibly Emsian age (Jensen, 1976) - thus the maximum age of the orogeny in the Meguma Zone is approximately 397 Ma. The oldest post-orogenic rock is the McAdam Lake Formation of earliest Eifelian age (M.S. Barss, pers. comm.). If the intercalibration given above (p. 119) is correct, this suggests that the orogeny took place before 395 Ma and that the Acadian Orogeny occurred during a 2 Ma interval of late Emsian time. However, the date of regional metamorphism in the Meguma Group has been set between 412 and 400 Ma with $^{40}\lambda_{\beta} = 4.963 \times 10^{-10} y^{-1}$ as recommended by Steiger and Jaeger (1977) (Reynolds and Muecke, in press). That is, radiometry indicates that the Acadian Orogeny began in Early Devonian time (Siegenian); granitic intrusion began at this time and culminated with post-deformational Late Devonian plutons (370 to 360 Ma) (Reynolds and Muecke, in press). Cormier and Smith (1973) argued for at least two episodes of post-orogenic granitic intrusions, the earlier at 408 ± 6 Ma (Viséan) with $\lambda Rb = 1.42 \times 10^{-11} y^{-1}$ as recommended by Steiger and Jaeger (1977). These radiometric dates evidently conflict with the stratigraphic ages. Apparently both regional metamorphism and early, post-folding plutons are of Siegenian age despite the fact that (1) Siegenian to perhaps Emsian fossiliferous strata are folded and then intruded by these plutons, and that (2) the stratigraphic gap between pre- and post-orogenic strata is late Emsian.

CARBONIFEROUS AND PERMIAN

From the Middle Devonian until the Early Permian, locally very thick (up to 13 km), almost entirely subaerial, fluvial to lacustrine siliciclastics accumulated in successor basins across southeastern Canada (Fig. 18.6). These generally unmetamorphosed sediments, with fairly common basic alkaline lavas, rest with angular unconformity or nonconformity on underlying, generally metamorphosed rock. Carbonates and sulphates of Carboniferous age are exceptions to the usual fluvial to lacustrine lithologies. The Windsor Group (Viséan) consists of at least a dozen, fill-in cyclothems which follow a model of carbonate, overlain by evaporites, and next, redbeds (Bell, 1958; Schenk, 1970a). The Upper Carboniferous is distinctive in containing important coal deposits with terrestrial cyclothems (Hacquebard, 1972).

The fill of the Fundy Basin of northern Nova Scotia and southern New Brunswick is an exception to the usually undisturbed nature of the strata. Here both the intensity of deformation and thickness of strata are at maxima. Intrabasinal horsts of basement rock are common and were major, proximal sources of sediment during Carboniferous time. Tectonic deformation was severe in places, and is probably mainly Late Carboniferous. Fyson (1967) suggested that folding in the Carboniferous was due to gravity sliding related to normal and strike-slip faulting. Certainly the presence of Lower Carboniferous evaporites would assist such movements. Poole (1967) called this episode of deformation the Maritime Disturbance, which was broadly coeval with the Alleghanian Orogeny of the American Appalachians.

The tectonic setting of the Fundy Basin has been called a rift-valley by Belt (1968), perhaps involving a right-lateral slip of 200 km (Webb, 1969). Schenk (1975) and Keppie (1977a) have suggested that it may be an aulacogen. Rast and Currie (1976) saw the northern border of the Fundy Basin in southern New Brunswick as the passage of the Variscan Front from Britain across northern Nova Scotia and southern New Brunswick. They concluded from two distinct cleavages in Lower and Upper Carboniferous rock that tectonic transport occurred first toward the northwest and then toward the southeast. Keppie (1977a) challenged the assumption that Nova Scotia would be part of the Variscan Orogenic Belt, despite the presence of Lower Carboniferous granites in the Meguma Zone.

The sedimentary pattern through this interval of time in the Canadian Appalachians was one of alluvial fans grading laterally into fluvial plains spotted with lakes, all within a structurally complex, intermontane system. The sources for these very thick, siliciclastic deposits were mainly horsts of the Avalon Zone and to a lesser extent the Meguma and Fredericton Zones in the Maritime Provinces (Schenk, 1976). At this time Atlantic Canada must have been near the centre of a vast megacontinent, the newly created Pangea. The specific location must have been adjacent to northwest Africa, or possibly northwest South America and then northwest Africa. Piqué (1975, 1977) compared the Devono-Carboniferous rift-valleys of Atlantic Canada with similar features of northwestern Morocco. There, the Bou-Regreg Formation (Cambrian ?) of the Rabat-Tiflet Zone (*see* discussion, "Plate-tectonic models", below) was folded and metamorphosed before the Hercynian Orogeny. This zone was the source area for thick, Devonian and Carboniferous (up to Namurian) pro-deltaic turbidite sands and shales deposited rythmically in a basin to the south. Basaltic volcanics issued from the faulted basin margin. Volcanics associated with similar basin faults are common in the Maritime Provinces. Lower Carboniferous carbonate-evaporite cyclothems record repeated floodings of generally hypersaline seas into tortuously interconnected, graben basins. The resulting carbonates show progressive increase in salinity, culminating in intertidal, algal stromatolites and disruptive, diagenetic, supratidal sulfate (Schenk, 1970, 1976). The climate was hot and arid, the setting mid-continental and equatorial. The rift-valley settine could be due to separation of northwestern South America from southeastern Atlantic Canada (McKerrow and Ziegler, 1972a) - possibly as an aulacogen. Alternatively the block-fault topography could be due to alternating tension and compression due to shearing along a sinusoidal fracture zone - the Glooscap Fault (Schenk, 1976).

In Newfoundland, granites of earliest Devonian (400 Ma) to Upper Carboniferous (300 Ma) age intrude rock of the old Avalon and Dunnage Zones (Bell *et al.*, 1977). Bell and Blenkinsop (1977) related the Carboniferous plutons to similar ones in Western Europe and Northwestern Africa. They extended the Hercynian (Variscan) Front westward from southwest England and Wales to Newfoundland along the boundaries of the Hampden and Fleur de Lys Zones in the northeast and the Lomond and Exploits Zones in the southwest of Newfoundland (i.e. along the Old Baie Verte Lineament in Newfoundland) to continue through southern New Brunswick (Rast and Currie, 1976) (Fig. 18.1). Dewey and Burke (1973) considered the widespread, Late Carboniferous plutonism to be due to basement reactivation following mid-Carboniferous collision between North America - northern Europe and Africa - southern Europe. As detailed below, the later Late Carboniferous (Maritime or Alleghanian) disturbance may be due to that part of Africa south of the South Atlas Fault colliding with North America. In the Atlantic Provinces, this Maritime Disturbance squeezed very thick (locally greater than 13 km) Carboniferous successions. Both Milligan (1970) and Currie (1977) reported evidence for very large-scaled, low-angle thrusting in Cape Breton Island. Precambrian crystalline rocks (George River Group) and Lower Carboniferous sedimentary rocks have over-ridden Viséan-Namurian but not Westphalian sediments. This klippe (or klippen) may be very extensive in Cape Breton (Dr. G.C. Milligan, pers. comm.). Currie (1977) attributed the overthrusting to gravity sliding from a large basement block which rose in Tournaisian and Viséan time. Emplacement was in post-Namurian time.

The Permian is poorly developed and both Lower and Middle Triassic are absent on land. During the Triassic, the area was uplifted and gravity faulting coincided with old lines of weakness, mainly along the Meguma-Avalon boundary. Rifting during the Jurassic permitted flooding of the present Atlantic into the area.

PLATE-TECTONIC MODELS

The application of plate-tectonic theory to the northern part of the Appalachian Orogen has spawned a plethora of evolutionary models (*see* Wilson, 1966; Bird and Dewey, 1970; Schenk, 1971; Rodgers, 1972; McKerrow and Zeigler, 1972a; Rast and Grant, 1973; Dewey and Kidd, 1974; St. Julien and Hubert, 1975; Poole, 1967, 1976; Ruitenberg *et al.*, 1977; and Keppie, 1977b). Most workers accept the Dunnage Zone as the scar of a

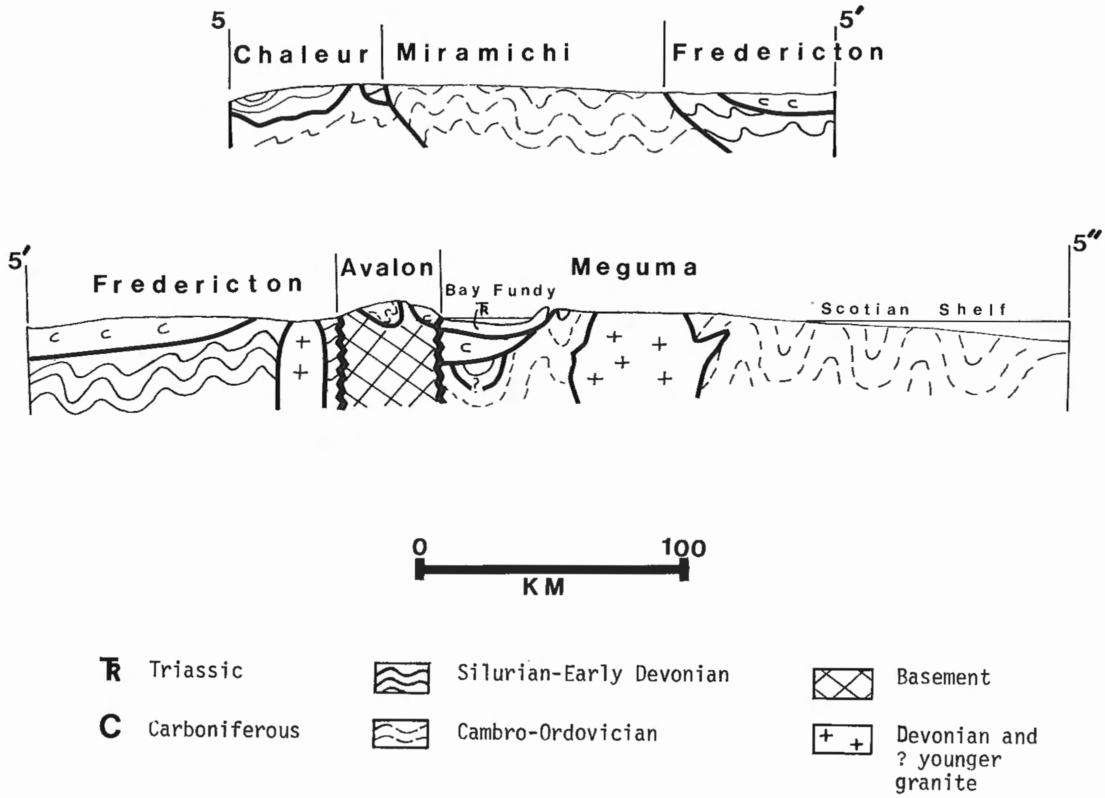


Figure 18.10. Section 5 across New Brunswick and Nova Scotia (modified from Rast *et al.*, 1976b). *See* Figure 18.5 for locality of sections.

A - Avalon microcontinent

col - Columbia

CS - Canadian Shield

E - eugeocline or abyssal plain deposit

F - France

G - Greece

GS - Guyanan Shield

I - Italy

M - miogeocline or stable craton

ma - Morocco

N. Am. - North America

N. Ap - Northern Appalachians

R - Rehamna Massif

R-T - Rabat-Tiflet Zone of Meseta

S - Spain

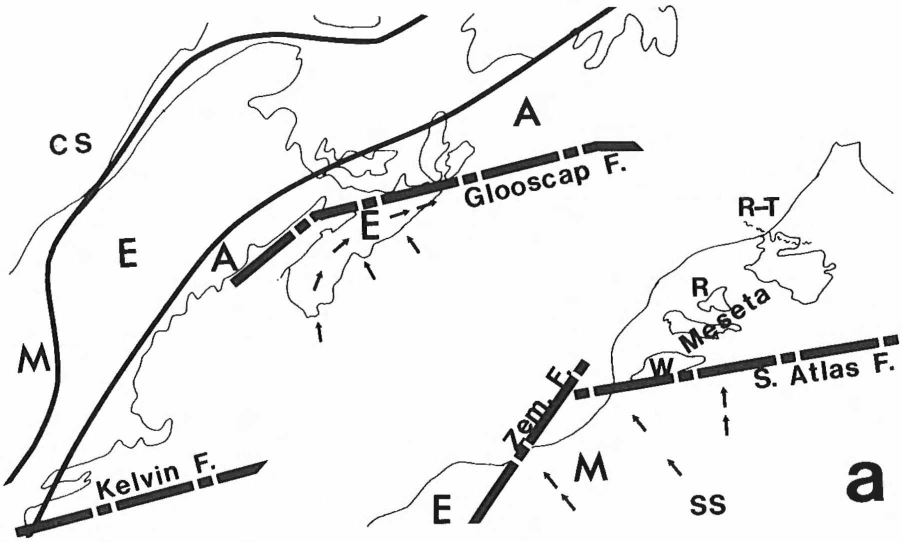
S. Ap - Southern Appalachians

S Atlas F - South Atlas Fault

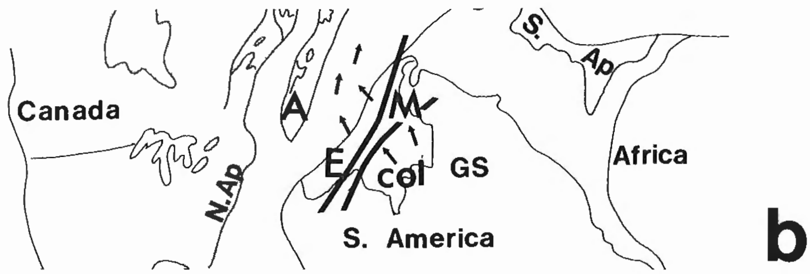
SS - Saharan Shield

W - Western High Atlas

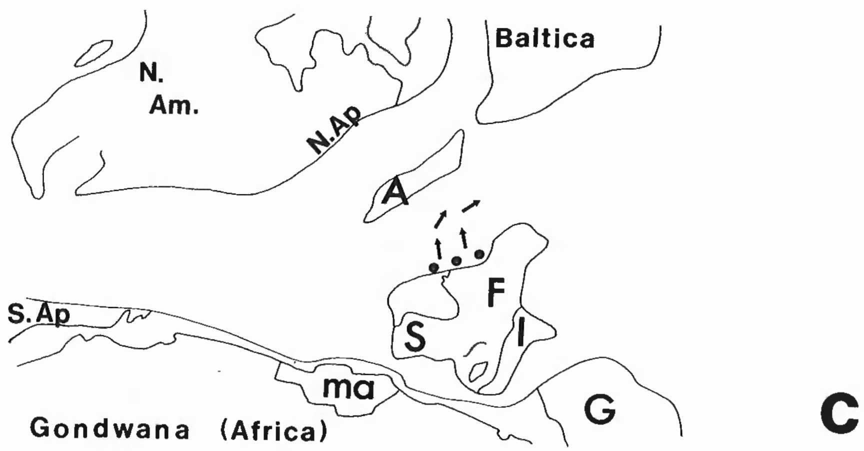
Zem. F - Zemmour Fault



a. Position of Atlantic Canada before Mesozoic rifting (after Le Pichon *et al.*, 1977.



b. Silurian distribution of the continents based on paleomagnetism, after Smith *et al.*, 1973 (modified from McKerrow and Ziegler, 1972a). The large megacontinent comprising South America, Africa and the Southern Appalachians would be part of Gondwana. Columbian patterns after Irving (1975) and Shagam (1975). Meguma dispersal pattern is deflected by the Avalon microcontinent.



c. Early Silurian distribution of the continents, modified from Ziegler *et al.* (1977). The line of dots off Brittany (France) is the proposed subduction zone of Lefort (1977). Arrows here show the Meguma dispersal pattern deflected by the Avalon microcontinent.

Figures 18.11a, b, c. Three hypotheses for the Paleozoic position of Atlantic Canada. (Arrows indicate sediment dispersal patterns.)

Paleozoic Atlantic, separating the continental margin of the Canadian Shield (the Humber Zone) to the northwest from the much smaller Avalon microcontinent to the southeast. The Meguma Zone would be a continental-margin complex stranded against North America (Schenk, 1971, 1976).

In the simplest proposed mechanism, a single "Wilsonian cycle" was employed to open the Paleozoic Atlantic Ocean in the Late Precambrian; start closure in the Ordovician with North America-island-arc collision in the Late Ordovician (Taconian Orogeny); and impact Atlantic Canada against northwestern Africa during the Early Devonian Acadian Orogeny (Fig. 18.11a). In an even broader scale, diachronous orogenies might be caused by oblique collision, with Silurian impact in the north (Caledonian), Devonian (Acadian) in Atlantic Canada, and late Paleozoic (Alleghanian) in the United States (Schenk, 1976). The Late Devonian-Carboniferous rift-zone setting in Atlantic Canada (Belt, 1968) might be due to shearing motion as Africa rotated clockwise against North America to close the Paleozoic Atlantic in the Late Carboniferous (Alleghanian Orogeny-Maritime Disturbance) (Schenk, 1975).

Recent discoveries in Morocco as detailed by Michard (this volume, Article 24) lend further geologic support to the model that the Meguma Zone may be a remnant of an Early Paleozoic Moroccan continental margin (Schenk, 1971, 1975). Contrary to previous opinion, northwest Africa does show very widespread and clear evidence of an Early Devonian Acadian Orogeny (Hollard *et al.*, 1976; Michard, 1976; Michard *et al.*, 1977; J.P. Schaer, pers. comm.). In the Rehamna Massif of the Moroccan Meseta, (Fig. 18.11a) Lower and Middle Devonian conglomerate redbed molasse of Old Red Sandstone facies overlies with angular unconformity rock as old as Cambrian. This boulder to pebble conglomerate is 400 m thick in the west, and becomes both thinner and finer-grained towards the east. Farther southwest in the Western High Atlas (Fig. 18.11a), the Lower Devonian Talmakent Formation is a 200 m-thick, "Old Red Sandstone-facies", red sandstone and conglomerate unit which rests disconformably on rock as old as Cambrian. Farther north in the Rabat-Tiflet Zone of the Meseta (Fig. 18.11a), calcalkaline granite dated at 414 ± 1 Ma (Early Devonian) has intruded marine metasediments presumably of Cambro-Ordovician age. Pebbles of this granite occur in Middle Devonian basal conglomerate. The Rabat-Tiflet Zone also contains the Bou-Regreg Formation (of presumed Cambrian age) that is very similar to the basal part of the Meguma Group. The Bou-Regreg is a thick succession of feldspathic to quartzose, fine-grained greywacke (thick "proximal" turbidites) with slump structures and carbonate concretions, alternating with thinner, shaly strata containing "distal" turbidites. The formation has been folded with axial lines trending west, diverging toward the south (similar to folds in the Meguma Group), and metamorphosed before the intrusion of the nearby 414 Ma granite. The Rabat-Tiflet Zone is in right-lateral fault-contact with the generally shallow-water, cratonic, Cambro-Ordovician sediments of the Meseta to the south, and so paleogeographic relations are unknown (Piqué; *in* Michard *et al.*, 1977). To the southeast (south of Marrakech), the Lower Carboniferous consists of a basal conglomerate with quartzite blocks, overlain by neritic shales with Upper Viséan fossils. This succession rests with angular discordance on Cambro-Ordovician (?) marine sandstones and shales (Huvelin, 1970). In southeastern Morocco, late Early Devonian regression is followed by latest Early Devonian transgression. Middle Devonian siliciclastics thicken markedly towards the west. Throughout Morocco, Early and Middle Devonian sedimentary trends are very distinct from later patterns which are due to Hercynian phases. The above workers concluded that an Acadian mountain range existed along western Morocco during the Early and Middle Devonian. Schenk argued on the grounds of gross stratigraphy, sedimentology, and some paleontology that a Moroccan Lower Paleozoic rise-slope-shelf prism was deformed by Devonian continental collision with North America. A segment of this prism remained against North America as the Meguma Zone after ragged rifting during the Mesozoic. This preliminary model was criticized because of the absence of clear structural and magmatic evidence of the Acadian Orogeny in Morocco (Hollard and Schaer, 1971; McKerrow and Ziegler, 1972a). Reconstructions of Late Ordovician and Early to Middle Devonian paleogeography based on geological, paleomagnetic, and paleoclimatic data place the Meguma and Avalon Zones off northwestern Africa in a configuration similar to that of Figure 18.11a (Zonenshayn and Gorodnitskiy, 1977).

However, more than one cycle of opening and closing has been proposed for the Paleozoic Atlantic. Dickinson (1974) cautioned that according to present rates of seafloor spreading and consumption, ocean basins as large as the present North Atlantic could form, or once formed could disappear, within 50 to 100 Ma. Thus during the 220 Ma time-span extending from Late Precambrian to the Carboniferous, several wide ocean basins could have evolved. Indeed McKerrow and Ziegler (1972a) suggested four continental collisions involving eastern North America. Paleomagnetic evidence (Smith *et al.*, 1973) suggests that the Caledonian (Erian) Orogeny of Late Silurian-Early Devonian age (Ireland to Scandinavia) resulted from collision between North America-Greenland and the Baltic Shield. The Acadian Orogeny of Middle Devonian age (New York to Newfoundland) resulted from sandwiching of the Avalon Zone (a very long prong of the Baltic Shield) between colliding continents of North America (Atlantic Canada) and northwestern South America (Colombia) (Fig. 18.11b). Rifting of Colombia from Atlantic Canada would result in the Carboniferous rift-valley tectonics in the latter (Belt, 1968). Late Carboniferous impact between North America and northwestern Africa south of the South Atlas Fault (Fig. 18.11b) created the Alleghanian Orogeny (south of New York) and the Maritime Disturbance to the north.

This entire scenario is very attractive, especially for the Carboniferous (van Houten, 1976). Furthermore, Viséan formainifera and algae of Atlantic Canada are almost entirely of the North American Realm, not the Tethyan, and so suggest that an oceanic seaway separated the North American and European-African blocks during Early Carboniferous time (Jansa *et al.*, in press). If this reconstruction of events is true, then the Meguma Zone would be a continental embankment of Colombia, not of Morocco. The Lower Paleozoic of Colombia is not well-known but the general stratigraphic section is similar to that of the Meguma Zone, with notable exceptions (Krummenacher, 1973) - neither Nouvelle Ecosse, nor Nouveau Maroc, but Nuevo Colombia? The paleogeography of Colombia during Cambro-Ordovician time was one of three, northerly trending zones (Fig. 18.11b) - epicontinental to the east, miogeoclinal in the central area, and ensimatic ocean to the west (Irving, 1975). Sediment transport was towards the northwest or west-northwest (Shagam, 1975), dovetailing with that of the Meguma (Schenk, 1970b). Specifically, the Quetame Group (Cambro-Ordovician?) appears to be very similar to the Meguma Group. The former is a structurally complex, low-grade, metamorphic succession at least 3 km in thickness. The lower part consists of quartzites with some conglomerate; the upper part of phyllites. The Quetame Group is overlain in angular unconformity with unmetamorphosed Middle Devonian sediments (Campbell and Burgl, 1965). Silurian sediments are unknown. The Cambro-Ordovician rocks were deformed, metamorphosed, and intruded during the Silurian-Early Devonian interval.

Although this orogeny has usually been called Caledonian, radiometric ages range from Late Ordovician through Early Devonian with a mode at Late Ordovician (Irving, 1975). The Early Devonian age of regional metamorphism in the Meguma Zone may be significant. The Venezuelan Andes has a Lower Paleozoic section of, from bottom to top, graptolitic black shales, fossiliferous siliceous limestones, and thick, interbedded marine conglomerates and fossiliferous shales. Major episodes of deformation, regional metamorphism, and/or granitic plutonism occurred in the late Precambrian, Devonian-Mississippian, and Late Permian-Early Traissic (Shagam, 1975). Keppie (1977b) suggested that southern Mexico or northern Central America may offer even closer similarities to Nova Scotia. However, much of the Precambrian and Early Paleozoic history of these areas is very poorly known because of scarcity of exposures and incomplete mapping (Dengo, 1975). The regional setting may have involved two northerly trending geosynclines - the eastern related to the Cordilleran zone, the western to the Ouachitan. Evidence does exist for both Late Ordovician as well as Devonian orogenies, but the data are still incomplete, conflicting, and confusing.

Ruitenberg *et al.* (1977) have presented a clear and attractive model for the evolution of the New Brunswick Appalachians; unfortunately, their paper arrived too late to be incorporated entirely in this synthesis. They divided the province into five Cambro-Ordovician zones, and further divide these on the basis of characteristics of the overlying Siluro-Devonian rocks. In general, the zones and subzones are a combination of those shown in Figures 18.3 and 18.4 with two important exceptions: (1) the Miramichi Zone narrows towards the southwest to approximately the outcrop width of the Siluro-Devonian (Fig. 8.3); and (2) a small but major zone occurs directly to the north of the Miramichi Zone. They postulate that Hadrynian distension caused initial intense volcanism on the Avalon Zone, and eventual rupture during the Cambrian which isolated this zone as a microcontinent. The Avalon was at least twice as wide as its present exposure because it underlies the Silurian Fredericton Zone (Fig. 18.4). This rifting created an open ocean to the south (to receive in part the Meguma Group conceivably from Africa) and a marginal basin to the north. This basin started to close during Early Ordovician time when a south-dipping subduction zone along its southern flank built an island-arc complex (the Miramichi Zone, restricted, Fig. 18.3). The Late Ordovician Taconian Orogeny marked closure of the marginal basin. The deformed Miramichi Zone, now acting as a geanticline, spread Silurian flysch both toward the north and south. To the south, the ensialic Fredericton Zone consisted of three belts: a northern deep-water belt; a medial shallow-water belt (as a back-arc basin); and a southern volcanic belt (as an ensialic volcanic arc). Subduction beneath this arc caused contraction of the Fredericton seaway and complete closure during the Early Devonian Acadian Orogeny. This general scenario does not conflict seriously with earlier schemes and has the advantage of details collated by active field workers.

Morris (1976) proposed an innovative model based on paleomagnetism for the relative paleogeographic positions of the Appalachian and Caledonian orogenic belts. He noted that Ordovician through Lower Devonian paleomagnetic poles of Britain are significantly different to contemporaneous poles of North America, whereas those of the Upper Devonian in both regions are similar. Moreover, the Lower Paleozoic zones of Newfoundland and Nova Scotia (i.e. zones a-i - Fig. 18.3) are usually bounded by major faults along which dominant strike-slip movement is at least suspected. The Early Paleozoic reconstruction of Morris places Britain near the latitude of northern Florida between the Ordovician and the Early Devonian; - i.e. the Appalachian and Caledonian regions would have formed side-by-side, rather than end-to-end as conventionally thought. During the Middle Devonian, major sinistral transcurrent faulting at a rate of approximately 9 ± 4 cm/y would slip the Caledonian region to a position on-trend with the northern Appalachians by Late Devonian time. If true, the zones of the Canadian Appalachians would consist of a fortuitous collection of random fault slices of the Appalachian and Caledonian regions.

Obviously the Early Paleozoic distribution of continents as based on paleomagnetism is not yet dogma. The relative positions of North America and Europe-Africa during the Paleozoic may well have been similar to that of today. The Early Silurian map of Ziegler *et al.* (1977) shows the Avalon Zone as a microcontinent between North America to the northwest, and Western Europe, not northwestern South America (McKerrow and Ziegler, 1972a), to the southeast (Fig. 18.11c). South America is in its accustomed location "below" North America (indeed South China, not Atlantic Canada is offshore western South America!). Could northwestern France, or even western Iberia act as source areas for the Meguma Group?

Recent work has revealed that the major tectonometamorphic event in the southern Armorican Massif (Brittany) and in the western Massif Central of France occurred between Early and late Middle Devonian time - i.e. at the time of the Acadian Orogeny (Autran, this volume, Article 21; Ters, 1976; Bernard-Griffiths *et al.*, 1977). Recall that northern Brittany has striking similarities to the Avalon Zone (or Anglesey Zone of Williams, in press). Compare southern Brittany-Western Massif Central with the Meguma Zone:

1. in France, the youngest, pre-orogenic strata are Early Devonian (Gedinnian); in the Meguma, youngest strata are Gedinnian to Siegenian, possibly Emsian;

2. in France, the oldest, post-orogenic strata are late Middle Devonian (Givetian); in the Meguma, early Middle Devonian (Eifelian);

3. in France, anatexic granites average 375 Ma (range 385 to 300 Ma); in the Meguma, post-folding, possibly anatexic granites average 375 Ma (range 385 to 350 Ma);

4. in France, the metamorphic event occurred between 360 and 420 Ma ($\lambda\ ^{87}Rb = 1.47 \times 10^{-11} y^{-1}$); in the Meguma, between 400 and 412 Ma.

Furthermore, Lefort (1977) has identified a possible Late Proterozoic to Middle Paleozoic subduction zone that dips southeasterly beneath Brittany. His Early Paleozoic reconstruction involves three areas, from northwest to southeast: 1) a thick Paleozoic section in two parts - basal, possibly metamorphosed strata similar to Ordovician metagreywacke outcrops, overlain by Devonian and Carboniferous rock; 2) a mafic rock complex that dips southeasterly and is similar to nearby outcrops of ophiolite; and 3) a broad Precambrian terrane marked by grabens containing thin Paleozoic sediments. Lefort suggested that this eastern continental margin was quite passive with very slow subduction in order to explain the absences of both high pressure—low temperature metamorphics as well as large volumes of plutonic rocks. Conceivably the associated trench would be a minor topographic feature and was buried beneath prograding deep-sea fans. Could the Meguma Group be the southwestward continuation of this Lower Paleozoic, presumably deep-water zone off either southwestern France or even Portugal?

A latest Devonian reconstruction by Cogné (1977) places the Meguma and Avalon Zones offshore southern Portugal. The Southern Portuguese Zone trends east-west across southern Portugal and southwestern Spain (Schermerhorn, 1971, 1975). The oldest rocks dated are latest Devonian (Middle to Upper Famennian), eugeoclinal phyllites and quartzites first deformed during Late Carboniferous (Middle Westphalian) time. Older Paleozoic rocks are exposed farther north in the central part of the Iberian Massif of Spain (Capote *et al.*, 1977) and in central and northern Portugal (Vegas *et al.*, 1977). The general picture is that of an enormous, presumably very late Precambrian, eugeoclinal complex that extends towards the south, and that shoals upward from deep-water turbidites through thin, Lower Cambrian shelf-carbonates, Tremadocian shallow-water redbeds, to the widespread Arenig Armorican Quartzite. The eugeoclinal complex is interpreted as a volcanic ridge to the north, grading southward to deep-water sediments, including thick turbidites and smaller quantities of pebbly mudstone, thin ferruginous carbonates, and olistostromes. The section was folded during three Hercynian events between Late Devonian and Late Carboniferous time. The credulous reader might accept the Meguma Zone in this framework if he/she considered that the age of most of the Meguma Group is unknown, and that both prograding sedimentation at a continental margin as well as orogeny may be very diachronous.

In summary, the Acadian orogenic events in both western Morocco and France, coupled with recent paleomagnetic analyses, strengthens the argument that the Acadian Orogeny resulted from a collision between North America (including the Avalon Zone) and Western Europe-Africa, and that the Meguma Group was a rise-complex built either off Morocco or Western Europe.

ACKNOWLEDGEMENTS

This synthesis concentrates on rock that I have seen and studied. Regional and inter-regional correlations and syntheses are of little value if based only on the literature. I sincerely thank field trip leaders of the 1977 I.G.C.P. Caledonian Orogen Project excursions in Morocco (Drs. A. Michard, A. Piqué, J.P. Schaer, and others), in Spain (Dr. R. Vegas), and in Western France (Drs. M. Ters, A. Autran, P. Cavet, J.P. Floch, M. Robardet, and others). Dr. W.H. Poole and Professor J. Riva are thanked sincerely for their comments and suggestions on a very preliminary draft of the synthesis. Professor Hank Williams contributed previously unpublished diagrams of Newfoundland, the map of the Appalachian Orogen with extension across the British Isles (Fig. 18.1), unpublished manuscripts, and advice. Dr. M. Zentilli suggested the geochronologic scale. The manuscript was circulated to heads of working groups of the Canadian team of the Appalachian-Caledonian Orogen Project. Only I am responsible for errors, ignorance, and bias.

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SYNTHESIS OF THE GEOLOGY OF THE NORTHEASTERN APPALACHIANS, U.S.A.

PROJECT 27 CALEDONIDE OROGEN

Philip H. Osberg

Department of Geological Sciences,
University of Maine,
110 Boardman Hall,
Orono, Maine 04473, U.S.A.

INTRODUCTION

This synthesis covers the Appalachian Mountain belt between the Hudson River in New York and the international border between the United States and Canada. The broadest aspects of stratigraphy, structural geol ogy, plutonism, and metamorphism are described and an attempt is made to organize this material into a history based on the plate tectonic model. For purposes of discussion it is convenient to divide the Appalachian belt into three parts: the western margin, the medial zone, and the eastern zone (Fig. 19.1).

DISTRIBUTION OF ULTRAMAFIC ROCKS

Ultramafic rocks are represented by dunite, peridotite, pyroxenite, serpentinite, and talc-carbonate rock. Their distribution is shown in Figure 19.2. A continuous belt of ultramafic bodies is exposed in Connecticut, Massachusetts, Vermont, and Quebec. Most of these bodies are diapiric and are injected into rocks of Precambrian to Lower Ordovician age (Chidester, 1968); some ultramafics, such as those at Thetford Mines, Quebec, are interpreted as ophiolites (Laurent, 1973). A second, less well-known belt is exposed in northernmost New Hampshire and in western Maine (Fig. 19.2). A single large body in this segment is interpreted as an incomplete ophiolite section (Boudette, oral communication, 1970). This belt is truncated both to the southwest and northeast by Siluro-Devonian rocks (Boucot, 1961; Harwood, 1973). A few ultramafics are found as an extension of this belt in northwestern Maine where they are exposed in pre-Silurian rocks in cores of anticlines that fold the Siluro-Devonian section (Hall, oral communication, 1968; Rabbe, written communication, 1976). Ultramafic bodies are also exposed in coastal Maine (Smith, Bastin, and Brown, 1907; Bickel, 1971) (Fig. 19.2).

The ultramafic rocks of the two western belts are clearly older than the Siluro-Devonian section. In southern Quebec no rocks of Middle Ordovician age or younger are intruded, but elsewhere relationships are less clear. The ultramafics of coastal Maine intrude rocks of various ages, and on the basis of somewhat tenuous evidence have been interpreted to cut Siluro-Devonian rocks (Smith, Bastin and Brown, 1907).

BASEMENT DOMAINS

Rocks that were originally metamorphosed to high grade and underlie stratigraphic sections of Paleozoic age are referred to as basement. Some of these rocks contain retrograde mineral assemblages. Although detailed stratigraphic relationships have not been worked out in basement rocks, they do contain distinctive lithologies and, where dated, they have consistent ranges of isotopic ages. On the basis of lithologic distinctions and on the ranges of age, basement has been divided into domains as shown in Figure 19.3.

Basement A is exposed primarily along the east side of the western margin (Fig. 19.3). To the west it underlies the Craton, with extensive exposures in, for example, the Adirondacks. It consists of a complex of paragneiss, marble, schist, and orthogneisses (Harwood, 1975; Norton, 1976). Radiometric ages of 1100 Ma (Ratcliffe and Zartman, 1971; Brookins and Norton, 1975) are common, although somewhat younger ages have been determined for certain orthogneisses (Norton, 1976).

Basement B is exposed in western Maine (Fig. 19.3) and consists of feldspathic schist and gneiss, quartzite, sulfidic schist, amphibolite, and quartzofeldspathic gneiss containing distinctive clast-like nodules of quartz (Boudette and Boone, 1976). Naylor *et al.* (1973) report a minimum isotopic age of 950 Ma for rocks in this massif. These basement rocks are lithologically different from the rock-types that characterize basement A, but their ages may be similar. These rocks may represent a small crustal fragment or possibly a variant of basement A.

Basement C is exposed in small areas in west-central Massachusetts, along the west edge of the medial zone and in east-central Maine, southeastern New Hampshire and in central Massachusetts, along the east boundary of the medial zone (Fig. 19.3). These rocks consist primarily of a bimodal suite of intercalated tonalites and amphibolites; limesilicate gneisses, sulphidic gneisses, and orthogneisses are less common. Naylor (1973) reports an age of 575 Ma for these rocks in west-central Massachusetts, and Besancon *et al.* (1977) report an age of 600-620 Ma for orthogneisses along the east border of this basement complex. The stratified core rocks of the Oliverian gneisses, giving zircon ages of 450 Ma (Naylor, 1968; Brookins and Methot, 1971), are included in this basement domain.

Basement D is exposed in coastal Maine and over much of the southern part of the eastern zone (Fig. 19.3). Feldspathic gneiss, sillimanite-bearing biotite gneiss, schists, marbles, amphibolite, granite gneisses, and granites characterize this domain (Emerson, 1917). Radiometric ages fall in the range 540-600 Ma (Fairbairn *et al.*, 1967; Zartman and Naylor, 1972).

CAMBRO-ORDOVICIAN STRATIGRAPHY

The regional stratigraphic relationships for the Cambro-Ordovician are shown in Figure 19.4.

Sections of the western margin

Rocks of the western margin include quartz arenite, dolostone, and limestone of Cambrian through Lower Ordovician age (Cady, 1945). These rocks intertongue to the east with a thick section of metashales and metagreywackes (Osberg, 1967; Cady, 1969). Both of these

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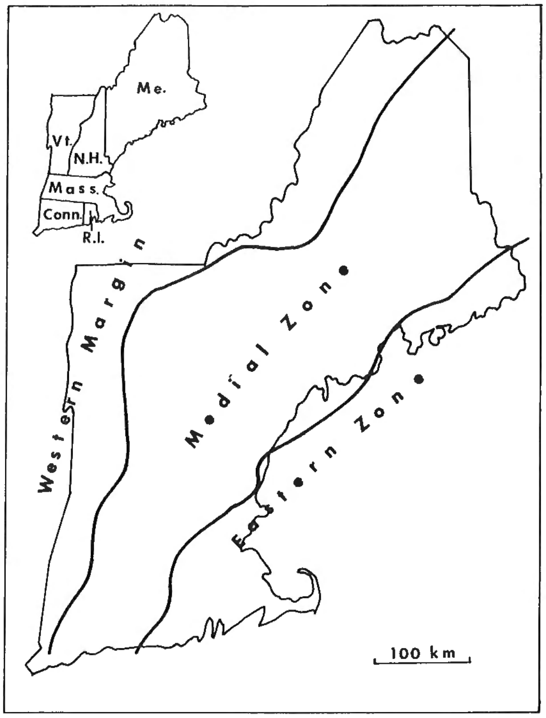

Figure 19.1. Index map and tectonic division of the New England Appalachians.

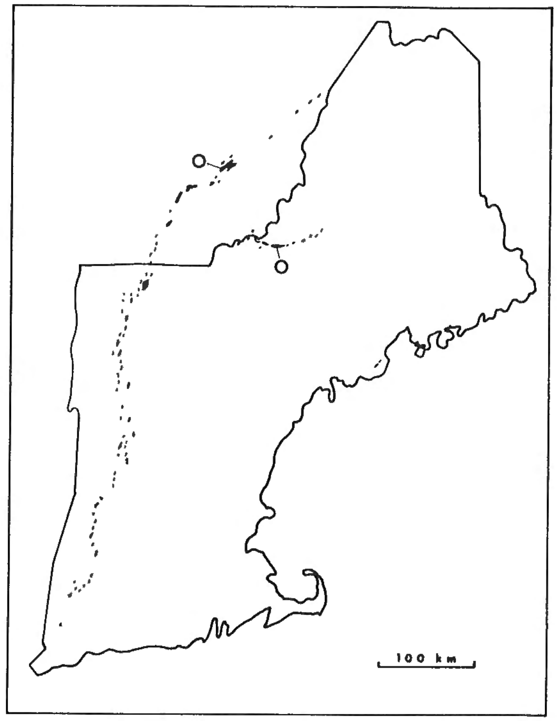

Figure 19.2. Distribution of ultramafic rocks. O indicates ophiolite sections.

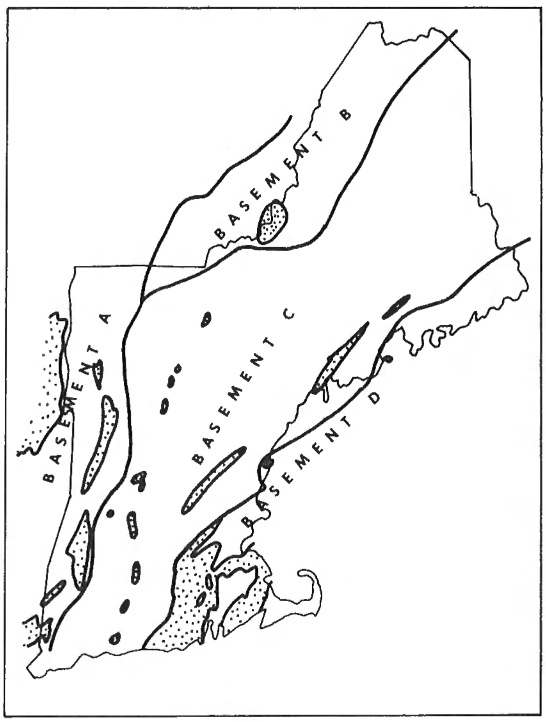

Figure 19.3. Basement domains in New England. Stippled areas indicate areas of exposed basement.

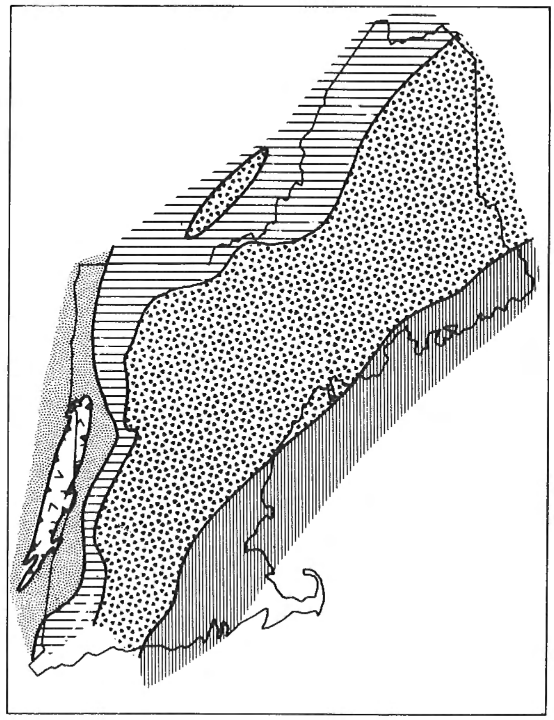

Figure 19.4. Stratigraphic relationships for the Cambro-Ordovician.

Stippled = miogeosynclinal and exogeosynclinal facies
V's = allochthonous section
Horizontal ruling = eugeosynclinal facies
Triangles = volcanic arc section
Vertical ruling = sections of the eastern zone

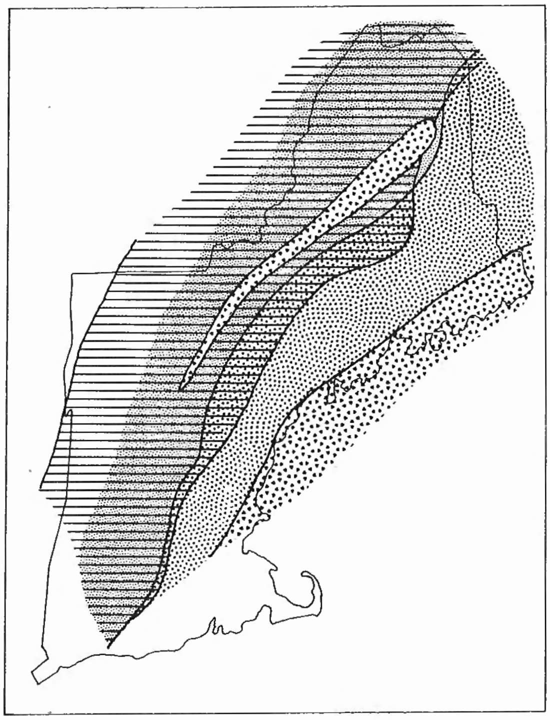

Figure 19.5. Stratigraphic relationships for the Siluro-Devonian.

Fine stippled = Silurian shelf facies
Coarse stippled = Silurian turbidites
Horizontal ruling = Devonian turbidites
Triangles = Siluro-Devonian volcanics

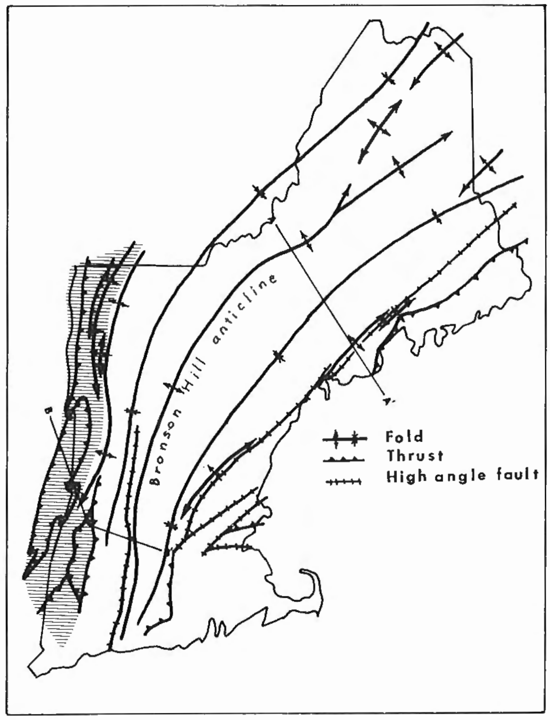

Figure 19.6. Structural features in New England. A-A' and B-B' indicate locations of structure sections (Fig. 19.7). Horizontal ruling = area where structural features of Ordovician age are prevalent.

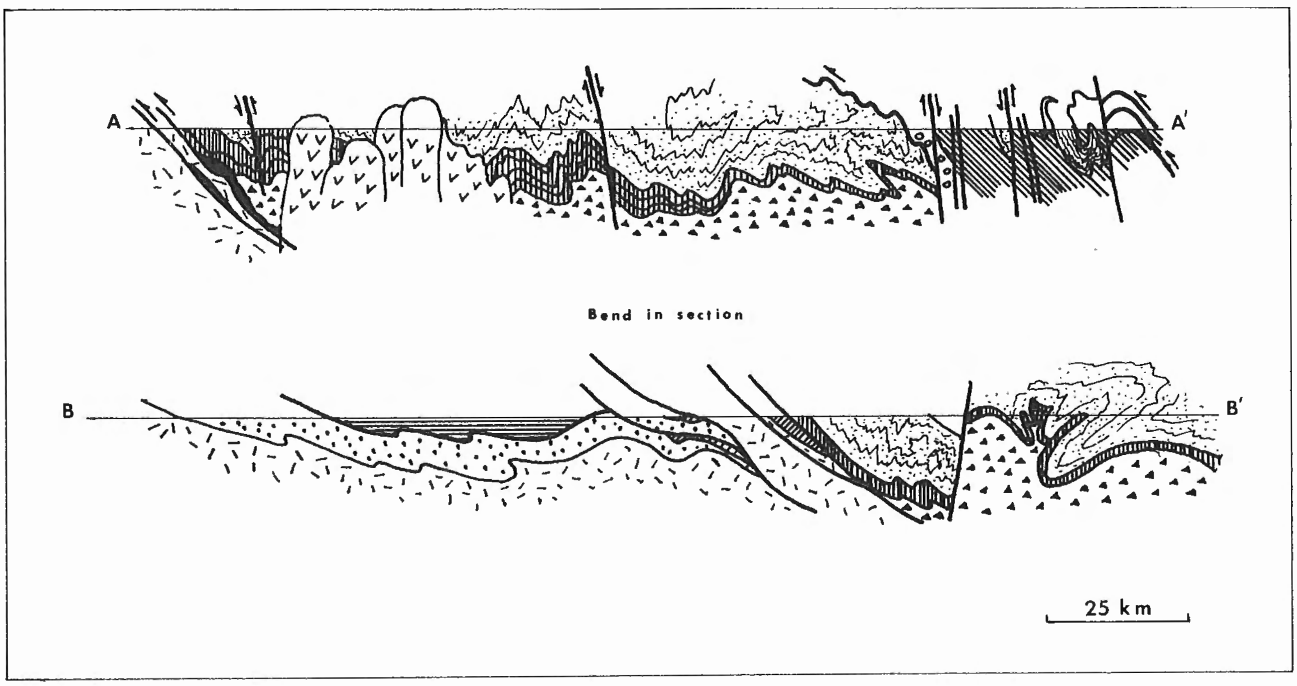

Figure 19.7. Structure sections across the New England Appalachians.

random slashes = basement B (Section A-A'), basement A (Section B-B')
diamonds = basement C
circles = basement D
black = ophiolites and ultramafics
horizontal ruling = Cambro-Ordovician rocks of Taconics
coarse stippled = Cambro-Ordovician miogeosynclinal rocks of the western margin
left-inclined ruling = Cambro-Ordovician eugeosynclinal rocks of the western margin
vertical ruling = Cambro-Ordovician rocks of medial zone
right-inclined ruling = Cambro-Ordovician rocks of eastern zone
fine stippled = Siluro-Devonian rocks
wavy lines = Ordovician rocks of eastern zone
v's = plutons

sections lie unconformably on basement (Brace, 1953; Thompson, 1967). Limestone mélanges at the boundary between the miogeosynclinal and eugeosynclinal facies may represent continental slope slide deposits (Rodgers, 1968).

Middle Ordovician shales, limestone and wacke of exogeosynclinal facies rest unconformably on the Cambrian to Lower Ordovician section of the shelf facies and locally directly on basement (Thompson, 1967; Zen, 1961; Osberg, 1967; Hall, 1966). These sediments had an eastern source and represent a precursor to the Taconic orogeny.

An allochthonous section of Cambro-Ordovician age rests on both the Cambro-Ordovician miogeosynclinal facies and the Middle Ordovician exogeosynclinal facies (Zen, 1967). Its sedimentary character is intermediate between miogeosynclinal and eugeosynclinal facies. The root zone of this section is interpreted to be to the east, but it has not been identified.

Sections of the medial zone

Ordovician rocks exposed along the west edge of the medial zone consist of quartz schists, metagreywacke, conglomerate, mafic volcanics that locally display pillow structure, fragmental volcanics, some felsic volcanics, black sulfidic schist/shale, and chert. This section contains Middle Ordovician fossils and locally rests unconformably on low-grade metamorphic phyllites (Hall, 1970), although in western Maine it is in probable thrust contact with basement B.

Rocks along the east edge of the medial zone consists of quartz schist, thin greenstone, thin black carbonaceous phyllite, and grey and green schist (Hussey, 1971). Part of this section is intruded by pegmatites that yield a minimum Ordovician age (Wones, 1974); the section rests directly on basement of uncertain age.

Sections of the eastern zone

Rocks exposed in the south part of the eastern zone belong to a common stratigraphy: shales, cherts, and interbedded shales and limestone. These sections contain Lower and Middle Cambrian fossils having Acado-Baltic character. Locally, the section lies with unconformity on granites of Precambrian age (Emerson, 1917).

Along coastal Maine the stratigraphic relations are confused by faults. Four different stratigraphic groupings have been established: (1) quartz schist, greywacke, mafic volcanics, minor felsic volcanics, and minor marble, (2) conglomeratic quartzite, mica schist, minor marble, and dark sulfidic mica schist, and quartzite, (3) mica schist, conglomeratic quartzite, marble, conglomeratic marble, polymictic conglomerate, and minor limesilicate quartzite, and (4) biotite schist, greywacke, minor limey quartzite, and biotitic quartzite. Section 4 contains Middle Ordovician fossils. However, the four sections are separated by thrusts or high-angle faults so that correlations between them are difficult.

SILURO-DEVONIAN SEQUENCES

Silurian shelf facies

A section containing basal quartzite/conglomeratic quartzite overlain by marble and limey shale makes up a consistent stratigraphy (Billings, 1956) widely distributed over the western margin and the western part of the medial zone (Fig. 19.5). This section rests with disconformity on older rocks (Harwood, 1973) and is itself overlain unconformably by a Devonian (?) section (Billings, 1956; Hall, 1966). This circumstance produces a patchy pattern of outcrop.

Silurian turbidite sequence

A Silurian sequence, thousands of metres thick, is exposed in northeastern Maine and central Massachusetts (Fig. 19.5). This sequence contains graded quartzwacke, shale, thin ribbon limestone, greywacke, conglomerate, and their metamorphosed equivalents. The facies is proximal toward the west (Osberg *et al.*, 1968) and possibly grades to the east into a volcano-clastic section (Metzger, 1975). The relationship between this turbidite sequence and the Silurian shelf facies to the west is uncertain. It may be transitional (Harwood and Berry, 1967), or it may be that the two facies were on opposite sides of a geanticline and have been positioned in close proximity by subsequent deformations.

Devonian turbidite sequence

A turbidite sequence of Devonian (?) and Devonian age is exposed in the western margin and in the western part of the medial zone (Fig. 19.5). This section contains grey shales and wackes, commonly displaying graded beds. It is thousands of metres thick and proximal to the east (Hall, Pollock, and Dolan, 1976). A prominent limestone facies at its west margin is consistent with an eastern source. The Devonian turbidite basin is offset to the west with respect to the Silurian turbidite basin.

Siluro-Devonian volcanic facies

A Siluro-Devonian volcanic facies is widely scattered along two lines (Fig. 19.5): (1) coastal Maine and northeastern Massachusetts, and (2) north-central Maine, possibly extending southward along the border between New Hampshire and Vermont (Rankin, 1968; Osberg, 1974). The coastal volcanics are predominantly felsite, pyroclastics, and some basalts (Gates, 1961; Brookins, Berdan, and Stewart, 1973). In northern Maine the Siluro-Devonian section is mostly felsic, although some mafic volcanics have been included in the section in New Hampshire and Vermont. In northwestern Maine part of the Siluro-Devonian volcanic section is interbedded with the Devonian turbidite sequence (Boucot, 1961).

Post-orogenic Devonian formations

Red and grey conglomerate and sandstone of Middle Devonian age have restricted outcrop in north-central Maine (Boucot *et al.*, 1964) and in eastern Maine (Smith and White, 1905). These formations are largely continental and lie unconformably on older rocks.

POST-DEVONIAN ROCKS

Rocks of Carboniferous age occupy fault-bounded blocks in eastern Rhode Island, eastern Massachusetts, and in east-central Maine. Representative rocks include red and grey conglomerate, red and grey sandstone, shale and coal beds (Quinn, 1963; Grew, 1974). The rocks in east-central Maine are continuous with Mississippian rocks in New Brunswick.

Red conglomerate, sandstone, shale, and mafic volcanics of Triassic age occupy a down-faulted block in west-central Massachusetts and central Connecticut.

STRUCTURAL GEOLOGY

Features of Ordovician age

Structural features of undoubted Ordovician age are exposed only in the western margin (Fig. 19.6). The earliest of these is high-angle faulting, the extent and importance of which is uncertain (Zen, 1967; Osberg, 1967). Thrusts, nappes, and recumbent folds with westward direction of transport developed from Middle Ordovician through Upper Ordovician time (Zen, 1967; Harwood, 1975). Some allochthons, such as the earlier Taconic slices, contain only Paleozoic rocks (Zen, 1967), but others, higher in the nappe sequence, contain basement rocks (Ratcliffe, 1975); the Berkshire massif may itself be allochthonous. In northernmost Vermont, some overturned folds may have been produced in a later Ordovician episode of deformation.

Features of Devonian age

Most of the structural features in New England were developed in Devonian time (Fig. 19.6). The earliest of these are recumbent folds and associated thrusts, which though poorly exposed are widely distributed (Osberg, 1975). These recumbent folds face west along the west edge of the medial zone and may in part refold the earlier Ordovician folds in the western margin. The thrusts east of the Berkshire massif (Norton, 1976) may also be of Devonian age. In the medial zone recumbent folds also face west, but in central Massachusetts and in Connecticut they are refolded so as to produce younger recumbent folds that at least locally face east (Robinson, 1967; Thompson *et al.*, 1968). The recumbent folds are refolded by intensely developed upright folds. Folds defined by the boundary between pre-Silurian and Siluro-Devonian rocks have modest amplitude-wavelength ratios, and comprise most of the major fold features (Fig. 19.6), but within the Siluro-Devonian sedimentary prisms flattened upright isoclinal folds formed concomitantly (Hatch, 1968). It is not clear whether the domes in the Bronson Hill anticline (Fig. 19.6) predate or postdate this episode of folding. At least two generations of minor structural features are younger than the episode of upright folding (Osberg, 1968). Certain thrusts and high-angle faults throughout the region have been shown to be of Devonian age (Moench, 1970; Lundgren, 1972; Pavlides, 1974; Stewart, 1974), but some can be shown to have had major displacements in Mesozoic time (Emerson, 1917; Larrabee, 1963).

Structure sections across the region are presented in Figure 19.7; their locations are shown on Figure 19.6.

PLUTONS

Plutons of Ordovician age (Fig. 19.8)

Igneous rocks of Ordovician age belong to the Highlandcroft plutonic series (Billings, 1956). Naylor (1968) has suggested that the nonstratified part of the Oliverian plutonic series should also be included in the Highlandcroft plutonic series, and that differences between the two groups of rocks are due to deformational and later thermal effects. The Highlandcroft plutonic series includes quartz diorite, granodiorite, quartz monzonite, granite, and syenite, all showing a weak to strong tectonic fabric (Billings, 1956). Rocks of this series are found principally along the Bronson Hill anticline (Billings, 1956) and its extensions into Maine (Moench and Zartman, 1976; Neuman, 1967; Ekren and Frischknecht, 1967). Similar rocks are found in pre-Siluro-Devonian sections at the east edge of the medial zone.

A gabbroic pluton in north-central Maine (Griscom, 1966) may represent an earlier intrusive phase associated with pillowed mafic volcanics.

Siluro-Devonian plutons

Gabbroic plutons are found principally in Maine and eastern Massachusetts. Many are layered (Gates, 1961; Bickford, 1963), and some have associated granitic-granophyric phases (Chapman, 1968). Their distribution is shown in Figure 19.9.

Coarse to very-coarse grained granodiorite, quartz monzonite, and granite have been assigned to the New Hampshire magma series (Billings, 1956). These rocks have a marked igneous foliation, and those in the central part of the medial zone contain a strong tectonic fabric as well (Billings, 1956; Page, 1968). These plutons form large sheets (Billings, 1956) and block-like bodies (Sweeney, 1972). They are confined to the medial zone (Fig. 19.9).

Peraluminous granites and quartz monzonites (Fig. 19.9) are younger than the foliated plutons and are also found in the medial zone as well as along the west edge of the eastern zone. Biotite granite and biotite/amphibole granodiorite are associated with the peraluminous rocks in the medial zone. These plutons tend to be younger than the peraluminous rocks.

Mildly alkalic granites (Fig. 18.9) are confined to the eastern zone. These include both supersolvus and subsolvus granites, some of which display Rapakivi texture.

Post-Devonian plutons

Plutons of Permian age occur in Rhode Island. They are fine-grained, grey, binary quartz monzonite. Other plutons in New England having similar characteristics could possibly be Permian, but their ages have not been demonstrated. Mesozoic plutons are present in New Hampshire, Vermont, and in Maine.

METAMORPHISM

Paleozoic metamorphism occurred in Ordovician, Devonian, and Permian time. Thompson and Norton (1968) have shown the distribution of isograds for New England and have delineated a hypothetical Al_2SiO_5-triple point isobar. Their paper did not treat the age of metamorphism and they combined data on metamorphism of different ages. The steep gradients on their map in eastern Massachusetts, New Hampshire, and Maine are at post-metamorphic faults.

Metamorphism of Ordovician age

Metamorphism of Ordovician age is best preserved in the western margin (Fig. 19.10). A Barrovian metamorphic event which has produced staurolite-kyanite bearing assemblages is recorded in New York (Ratcliffe, 1968), and in Vermont (Lanphere and Albee, 1974). This same event has produced biotite-chlorite assemblages in west-central Vermont (Laird, 1975), recording an earlier high-pressure metamorphism in that area (Fig. 19.10). Elsewhere, Stewart *et al.* (1974) have postulated an Ordovician metamorphism for parts of coastal Maine.

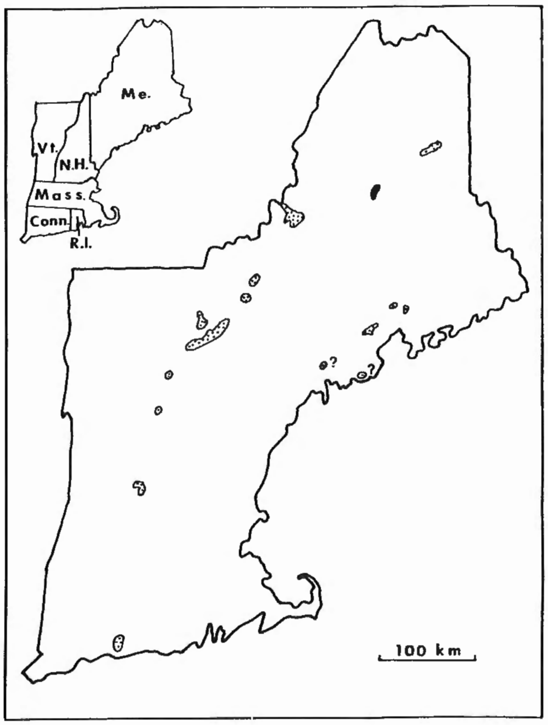

Figure 19.8. Location of Ordovician plutons.

Stippled = granodiorite, granite and syenite
Black = gabbro

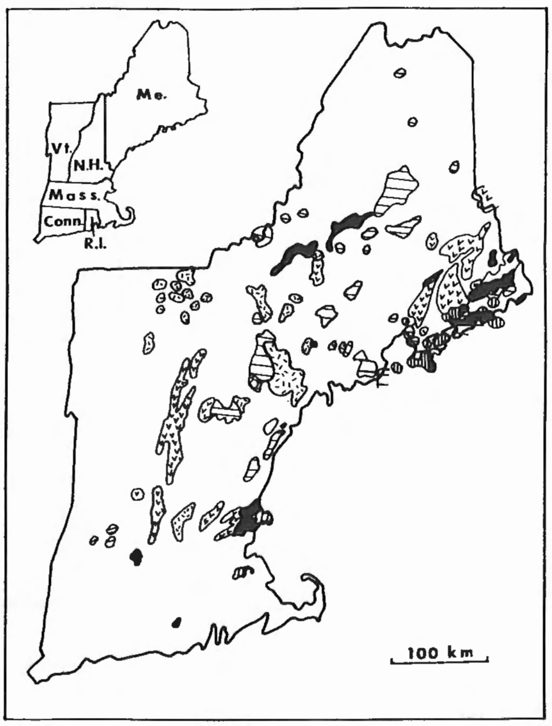

Figure 19.9. Location of Siluro-Devonian plutons

Black = gabbro
V's = coarse-grained granodiorite, quartz monzonite and granite
Random slashes = peraluminous granite and quartz monzonite
Horizontal ruling = biotite and biotite-amphibole granite and granodiorite
Vertical ruling = mildly alkalic granites

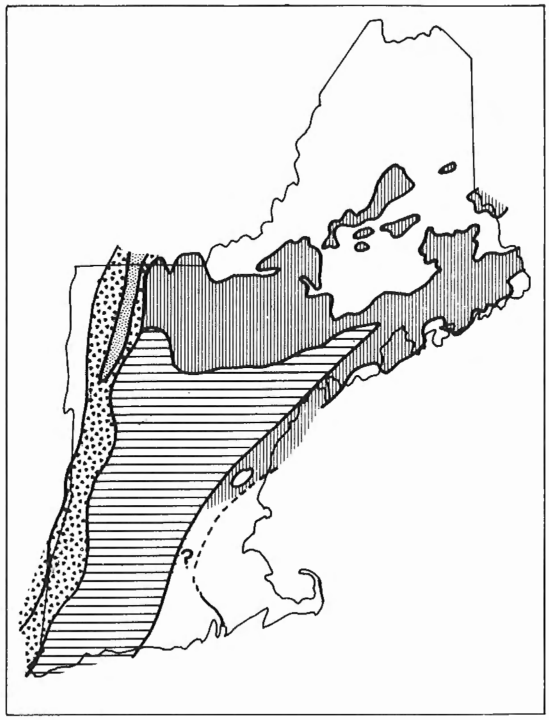

Figure 19.10. Metamorphic relationships.

Stippled = Ordovician high-pressure metamorphism
Triangles = Ordovician Barrovian metamorphism
Vertical ruling = Devonian Buchan metamorphism

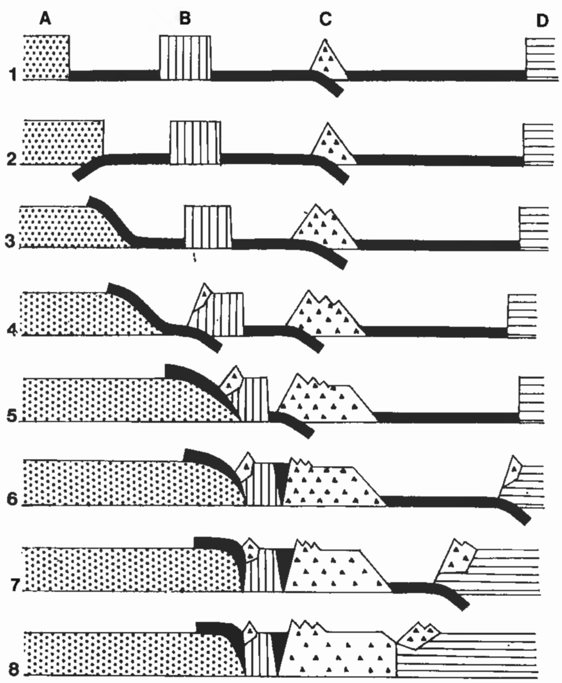

Figure 19.11. Development of Plate margins.

Heavy line = oceanic crust
Stippled = continental crust (Basement A)
Vertical ruling = continental crust (Basement B)
Horizontal ruling = continental crust (Basement D)
Triangles = volcanics and volcanic arc
1 - Late Cambrian
2,3,4 - Lower Ordovician
5,6 - Middle and Upper Ordovician
7 - Silurian
8 - Lower Devonian

Metamorphism of Devonian age

Both Barrovian and Buchan types of metamorphism have been produced in the medial zone and western part of the eastern zone (Fig. 19.10). These metamorphic types are separated by the Al_2SiO_5-triple point isobar which must swing northward to include kyanite localities not reported by Thompson and Norton (1968). The terrane of Barrovian metamorphism is terminated against post-metamorphic faults in the east (Fig. 19.10).

Post-Devonian metamorphism

Carboniferous rocks in Rhode Island and in eastern Massachusetts have been metamorphosed to staurolite grade (Quinn, 1963; Grew, 1974). Although this metamorphic event is locally well documented, its areal extent and its relationship to older episodes of metamorphism has not been thoroughly investigated.

Zeolite assemblages have been described from Lower Paleozoic rocks (Zen, 1974; Richter and Roy, 1976); they also occur in Triassic rocks. Heretofore, these assemblages have been thought to have been produced in separate metamorphic events of Ordovician and Triassic age, but the widespread occurrence of prehnite-bearing veins through the Siluro-Devonian section of New England (including the high-grade terranes) suggests the possibility of a pervasive zeolite metamorphism of Triassic age.

PLATE TECTONIC MODEL

Continental plates

Basements A, B and D (Fig. 19.3) are interpreted as continental material. These basement domains contain Precambrian rocks of different character and average age and dissimilar Paleozoic sections. These relationships indicate that they have not been parts of the same plate throughout decipherable geologic time. The relative positions that they now have are thought to have been obtained in Ordovician to Devonian time.

Island arc

Basement C, for the most part, is thought to be the deeply eroded roots of an island arc, consisting of tonalite and amphibolite intruded by granodiorite and granite. Radiometric ages suggest that its age is 550-450 Ma. If the 600-620 Ma date (Besancon *et al.*, 1977) is validated by additional work, at least the eastern part of this basement domain must contain continental material.

Higher level deposits in this island arc are exposed both along its west margin and locally along its east edge. The western section is highly volcanic; mafic, pillowed basalts grade upward through agglomerates into more felsic volcanics. The thickness of these volcanics varies along strike as do the details of the section. The relationship of the volcanic section to the underlying tonalite and amphibolite is to a large extent gradational (Lundgren, 1962; Robinson, 1967; Naylor, 1968).

The section exposed at the east edge of the island arc is relatively thin and contains little volcanic material.

In this reconstruction the western volcanic section marks the active part of the arc and is taken as the leading edge of the arc. The eastern section, with its paucity of volcanics, is taken as its trailing edge. The western volcanic section can be traced northward from central Connecticut through eastern Vermont to northern Vermont where it bifurcates; one branch traces into southern Quebec where it is known as the Ascot-Weedon volcanics, and the other branch traverses northern New Hampshire into western and north-central Maine and possibly into the Tetagouche volcanics of New Brunswick. The southern section of volcanics and the eastern branch of volcanics are thought to be continuous, and the branch containing the Ascot-Weedon volcanics is interpreted as a distinct and different volcanic section.

Plate margins

Consuming plate margins are thought to have operated in Ordovician time along the northern third of the east edge of basement A, and along the western borders of basements B and C. These relationships are complicated and, therefore, they are schematically shown in Figure 19.11. Activity along these subduction zones was sequential with final closure occurring in Middle to Upper Ordovician time. Initial subduction occurred along the east margin of basement A from north-central Vermont northward into Quebec. The descending plate is thought to have dipped westward beneath basement A, possibly accounting for some of the easterly movement on thrust faults in the Cambrian section of Quebec (Osberg, 1965). Subsequently, the oceanic plate was broken and obducted to the west (St. Julien, 1975). Displacement was of the order of tens of kilometres. The sections at Thetford Mines, Quebec (Laurent, 1973) are transported remnants of these obducted plates. Direct evidence of obduction has not been found to the south in Vermont, but the presence of high pressure minerals (Laird, 1973), which commonly occur beneath obducted plates, suggests that obduction occurred at least as far south as north-central Vermont.

With the welding of oceanic crust to basement A by obduction, a new subduction zone developed at the west edge of basement B. This subduction zone dipped eastward, building a small volcanic edifice (Ascot-Weedon volcanics) on the leading edge of basement B. Operation of this subduction zone closed the oceanic basin between basements A and B.

A third subduction zone developed in oceanic crust east of basements A and B. Operation of this subduction zone was initiated in Late Cambrian time, was in part concurrent with that of the subduction system that juxtaposed basements A and B, and lasted longer, possibly into the Upper Ordovician. A large and mature volcanic arc, recognized as basement C, was built at this converging plate boundary. Since the leading edge of this arc lies toward the west, the subducted oceanic plate dipped eastward. The collision of this arc with basements A and B culminated the Taconic orogeny.

A fourth converging plate margin is interpreted to have existed at the west edge of basement D in Late Silurian and Lower Devonian time. The abundant volcanics along coastal Maine and in Massachusetts represent a volcanic edifice constructed on the leading edge of basement D in a Cordilleran-type plate margin. The presence of these volcanics combined with the few ultramafic bodies that lie just to the west suggest that the subduction zone dipped east. Operation on this subduction zone closed a narrow ocean between basements C and D. The collision of basement D with the accreted basements A, B, and C produced the Acadian Orogeny. The paucity of geologic features related to subduction at the junction between basements D and C suggest that this suture is cryptic.

Turbidite sequences

A turbidite sequence was deposited over much of the medial zone in Silurian time from the erosion of a mountain chain to the west formed in the Taconic orogeny. In Upper Silurian time this turbidite section may also have been supplied by debris from the east as basement D approached basements A, B, and C.

When basement D collided with basements A, B, and C a new mountain system was formed in the region of their junction, and a Devonian turbidite section was built to the west. This sedimentary prism was constructed out over the medial zone and it extended into the western margin.

Deformational effects

Structural features identified as Ordovician indicate a transport direction towards the west. Recumbent folds and associated thrusts were formed as a consequence of the collision of basements A and B with the island arc (basement C). Presumably later adjustments to plate movements produced other structural effects, but these have not been sorted out except possibly in northern Vermont and southern Quebec (Rickard, 1965).

The direction of tectonic transport in the Acadian orogeny was also directed towards the west. Early, major recumbent folds face west and associated thrust plates moved towards the west.

Later compressive adjustments produced the major anticlinoria and synclinoria. The anticlinoria are particularly complicated because they involve older rocks having marked differences in rheological properties, which gave rise to spectacular superimposed folds and the localization of diapiric domes. Locally, such as in parts of the Bronson Hill anticlinorium, original west-facing recumbent folds have been back-folded producing a second generation of recumbent folds that face east (Robinson, 1967). The synclinoria, because they contain thick prisms of rocks with more-or-less uniform rheological character, are structurally simpler than the anticlinoria, even though these sedimentary sections are deformed by isoclinal upright folds.

Continental volcanics

Mostly felsic volcanics of Siluro-Devonian age in north-western Maine were generated by partial melting of the crust as a consequence of continental collision. Melting zones were located where continental crust was abnormally thickened.

Plutonism and metamorphism

The development of granitic and granodioritic magmas and regional metamorphism was synchronous and was consequent on collision of crustal blocks. Compositional peculiarities of plutons may in part be due to compositional differences of the crustal blocks in which they were generated by partial fusion. Regional metamorphism was a Barrovian-type at deeper levels and of Buchan-type at higher levels.

Both the Taconic and Acadian orogenies produced their own plutonic-metamorphic associations. The absence of Buchan-type metamorphic associations of Taconic or Acadian age along the western margin of the orogen suggests that the tectonic transport to the west has carried Barrovian-type terrane completely over the Buchan-type metamorphic terrane. Toward the east the Devonian metamorphism overprints the Ordovician metamorphism. A Buchan-type metamorphic series of Devonian age borders the Acadian metamorphic terrane to the north and east.

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SYNTHESIS OF THE SOUTHERN AND CENTRAL APPALACHIANS, U.S.A.

Robert D. Hatcher, Jr.

Department of Chemistry and Geology,
Clemson University,
Clemson, South Carolina, 29631, U.S.A.

INTRODUCTION

The segment of the Appalachians to be discussed herein is that from the lower reaches of the Hudson River in New York to the Coastal Plain overlap in Alabama (Figs. 20.1, 20.2). That portion from the Hudson River to central Virginia is commonly called the Central Appalachians, while that from Virginia to Alabama is called the southern Appalachians.

REGIONAL GEOLOGY

The southern and central Appalachians may be divided into four major physiographic-structural subdivisions: the Cumberland-Allegheny Plateau, Valley and Ridge, Blue Ridge (and Reading prong) and Piedmont. Several structural-stratigraphic-petrologic subdivisions of the Blue Ridge and Piedmont will be discussed briefly.

The dominant structures of the Cumberland-Allegheny Plateau and Valley and Ridge were formed during the late Paleozoic Alleghanian (Hercynian) orogeny. Therefore little discussion of the structural styles of these zones will be made except to say that most recent workers conclude that the deformation was of a "thin-skinned" nature (Rodgers, 1949, 1964, 1970) and several have concluded that gravity is the driving mechanism by which deformation occurred (Gwinn, 1964, 1970; Dennison, 1974; Milici, 1975). Earlier tectonic events are reflected in the sedimentary record, however.

The eastern part of the Valley and Ridge in Pennsylvania, New Jersey and southeastern New York (Lebanon or Great Valley) consists of Cambrian and Ordovician rocks whose structural style is markedly different from that farther west, more like that in the Piedmont than the Valley and Ridge. Recumbent folds, folded thrusts and other polydeformationl features occur here. Yet the rocks have not been appreciably metamorphosed (Rodgers, 1970; Root, 1970; A.A. Drake, written commun., 1976). Much of this deformation is thought to be Taconic but Alleghanian structures are also present (Root, 1970). The Great Valley is also the site of emplacement of transported sedimentary sequences like those occurring farther north in the Taconic Mountains, Quebec and Newfoundland. None of these klippen are known to occur farther south than Harrisburg, Pennsylvania (A.A. Drake, written commun., 1976).

The Blue Ridge is an anticlinorium in the central Appalachians and its anticlinal character continues southward but in the southern Appalachians it becomes a series of crystalline thrust sheets. The principal line of evidence for the allochthonous character of the Blue Ridge is the Grandfather Mountain window (Fig. 20.1, 3d) in North Carolina.

The oldest rocks of the Blue Ridge belt are ortho- and paragneisses, hereafter called basement, that were metamorphosed 1100 Ma ago. These are overlain by upper Precambrian metasedimentary and metavolcanic rocks which in a few places are overlain by lower Paleozoic sedimentary rocks. However, along the western edge of the Blue Ridge in parts of Tennessee and Virginia, Cambrian sedimentary rocks rest upon basement. Lower Paleozoic metasedimentary rocks are probably represented in the sequences in the Murphy syncline (Fig. 20.1, 3b) and other smaller synclinal structures in the Blue Ridge. The Talladega belt along the western edge of the extension of the Blue Ridge in Alabama contains rocks that may be as young as Mississippian.

Lower Paleozoic granitic plutons are concentrated in a belt along the eastern edge of the Blue Ridge. Ultramafic and mafic bodies emplaced prior to Taconic metamorphism, follow a similar trend in the same area. A few plutons occur outside this eastern belt.

Paleozoic metamorphism in the Blue Ridge has been recognized as classical Barrovian metamorphism (Carpenter, 1970). There was apparently only one metamorphic event affecting the southern Blue Ridge which occurred 450-480 Ma ago (Butler, 1972; Dallmeyer, 1975).

The southern Blue Ridge is divisible into three distinctive belts (Hatcher, in prep.): 1) a western belt of imbricate thrusts, which involves upper Precambrian and lower Paleozoic rocks and some basement rocks; 2) a central belt where most basement rocks occur along with higher metamorphic grade upper Precambrian and possibly lower Paleozoic metasedimentary rocks, a southward extension of the core of the Blue Ridge anticlinorium of the central Appalachians (BLUE-GREEN-LONG AXIS of Rankin (1976)); 3) an eastern belt of high grade upper Precambrian-early Paleozoic (?) metasedimentary and metavolcanic rocks. The eastern belt is characterized by the very restricted occurrence of basement rocks and contains the lower Paleozoic plutons and ultramafic rocks mentioned above. The ultramafic rocks could have been emplaced along the sole of a thrust, or series of thrusts generated during closing of the Late Precambrian Ocoee basin (Hatcher, in prep.). Later deformation and metamorphism served to disperse them and destroy their original character.

The Blue Ridge becomes a narrow northeast-plunging anticlinorium in Virginia (Fig. 20.1, 3e). This trend continues into Maryland and Pennsylvania where the basement rocks plunge beneath the surface and the South Mountain fold is the only remnant of the wide expansive Blue Ridge farther south (Cloos, 1947, 1971; Espenshade, 1970). This structure disappears beneath a Triassic basin in Pennsylvania. The South Mountain fold and central Blue Ridge anticlinorium are thought to have formed during the Acadian orogeny, but the cleavage there and major deformation are thought to be Alleghanian (Root, 1970).

Basement rocks reappear in northeast Pennsylvania in the Reading Prong (Fig. 20.1, 5), which could be a continuation of the Blue Ridge. The basement rocks in the southwest portion of the Reading Prong have been proven allochthonous (Drake, 1970). Those at the northeast end are thought by some to be autochthonous (Dallmeyer, 1974) but geophysical evidence supports

IGCP Project 27, U.S.A. Contribution No. 2.
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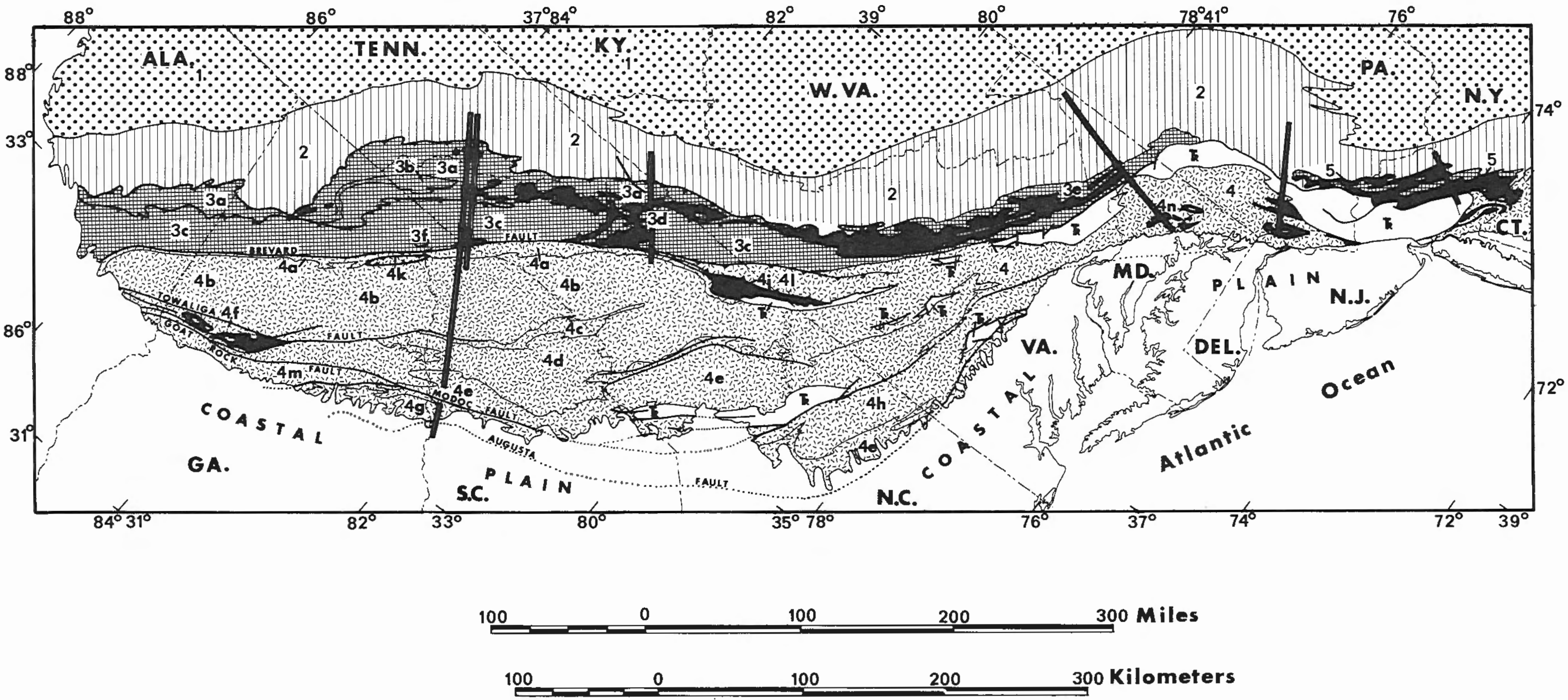

1 - Cumberland-Allegheny Plateau
2 - Valley and Ridge
3 - Blue Ridge
3a - Western and central belts of the Blue Ridge
3b - Murphy syncline
3c - Eastern belt of the Blue Ridge (Fries-Hayesville thrust sheet)
3d - Grandfather Mountain window
3e - Blue Ridge anticlinorium of Virginia and Maryland becoming the South Mountain fold in Maryland and Pennsylvania
3f - Tallulah Falls Dome
4 - Piedmont
4a - Chauga belt
4b - Inner Piedmont
4c - Kings Mountain belt
4d - Charlotte belt
4e - Carolina slate belt
4f - Pine Mountain belt
4g - Kiokee belt
4h - Raleigh belt
4j - Sauratown Mountains anticlinorium
4k - Alto allochthon
4l - Smith River allochthon
4m - Uchee belt
4n - Baltimore Gneiss domes
5 - Reading Prong

TR - Triassic rocks
Basement rocks are shown in black.
The heavy black lines show the locations of cross-sections in Figure 20.2.

Figure 20.1. Geologic sketch map showing the subdivisions of the Appalachians southwest of the Hudson River. Compiled and modified from Rodgers (1970, Plate 1), King and Beikman (1974) and other published sources.

the idea of an allochthonous or para-autochthonous setting for this area as well (A.A. Drake, written comm., 1976).

The recognition of pre- or synmetamorphic faults in the higher grade portions of the southern Blue Ridge is a recent development. The premetamorphic Greenbrier fault of the western Blue Ridge in Tennessee and western North Carolina has been known for some time (Hadley and Goldsmith, 1963; King, 1964). Others in the eastern Blue Ridge have now been recognized by the author. Several other faults, previously interpreted as Alleghanian features, are being reinterpreted as older structures (Hatcher, in prep.).

The boundary between the southern Blue Ridge and Piedmont from Virginia southward is the Brevard fault (Fig. 20.1). Various interpretations have been offered for this structure. It has been interpreted as a strike-slip fault (Reed and Bryant, 1964), a strike-slip fault with a dip-slip component of movement (Reed, Bryant and Myers, 1970), an Alpine root zone (Burchfiel and Livingston, 1967), a thrust (Hatcher, 1971; Rankin and others, 1973), the sole of a large nappe (Clarke, 1952; Bentley and Neathery, 1970), a zone of detachment (*abscherung*) (Griffin, 1971a, 1971b), and a transported suture (Rankin, 1975). It is a very linear feature and certainly a major fault having an early history of ductile deformation and a later history of brittle deformation. It is stratigraphically controlled in part and contains slices of what appear to be unmetamorphosed platform type carbonate rocks derived from the footwall of the Blue Ridge thrust sheet. Structural sequences (including deformational styles and orientations) northwest, within and southeast of the Brevard zone are identical, except for the episodes of faulting within the Brevard zone.

A narrow belt - the Chauga belt (Fig. 20.1, 4a) - extending from central North Carolina to central Georgia, lies southeast of the Brevard Zone. It is a belt of metasedimentary and metavolcanic rocks of lower metamorphic grade than the rocks of the Inner Piedmont and Blue Ridge. This feature is a synclinorium and flanks the higher grade Inner Piedmont along the northwest edge of the latter. The Brevard Zone occupies the western limb of the Chauga belt. The boundary between the Inner Piedmont (Fig, 20.1, 4b) and the Chauga belt is tectonic in a few places but is a metamorphic gradient in others. Metasedimentary and metavolcanic rocks of high grade intruded by granitic to gabbroic plutons of mainly Early Paleozoic age are found in the Inner Piedmont. Inner Piedmont structures consist of west vergent folds of several generations and several large allochthons have been recognized immediately southeast of the Brevard zone, probably derived from nappes farther southeast.

The Kings Mountain belt (Fig. 20.1, 4c) lies southeast of the Inner Piedmont in the Carolinas. It is a belt of metasedimentary rocks of slightly lower grade compared with the Inner Piedmont, and may be synclinal, but its structure is complicated by faulting along one or both boundaries (Butler and Fullagar, 1976).

Southeast of the Inner Piedmont in Alabama and western Georgia is the anticlinal Pine Mountain belt (Fig. 20.1, 4f). It contains basement rocks overlain by moderate grade metasedimentary rocks. Both its northwest and southeast boundaries are faulted.

East of the Kings Mountain belt in the Carolinas is the Charlotte belt (Fig. 20.1, 4d). This is another high grade zone and most of the plutons of the southern Appalachians occur here. Their compositions range from gabbroic to granitic. Their ages range from 415 to 385 Ma (both pre- and post-metamorphic) with another group of granitic plutons whose ages cluster about 300 Ma (Butler and Fullagar, 1976).

The Carolina slate belt (Fig. 20.1, 4e) extends from Virginia to Georgia. It contains an assemblage of mafic to felsic volcanic rocks and some metasedimentary rocks ranging in age from late Precambrian to early Paleozoic. Metamorphic grade is low. These rocks have been intruded by gabbroic to granitic plutons ranging in age from 705 to 510 Ma and by 300 Ma old granitic rocks (Butler and Fullagar, 1976). The boundary between the Carolina slate belt and Charlotte belt is probably metamorphic but may involve an antiformal-synformal relationship between the two belts in which the thermal surfaces have been folded (Hatcher, 1972a, Fig. 3).

East of the Carolina Slate belt are the antiformal Uchee belt in Alabama and Georgia, the Kiokee belt in Georgia and South Carolina, and the Raleigh belt in North Carolina (Fig. 20.1, 4g, h, m). All contain higher grade rocks not unlike those of the Charlotte belt. Carolina slate belt rocks occur east of both the Raleigh and Kiokee belts.

Several large faults exist within the Piedmont (Fig. 20.1). The Towaliga and Goat Rock faults flank the Pine Mountain belt (Fig. 20.1, 4f) in Alabama and Georgia and probably extend eastward into South Carolina and beyond. The Gold Hill-Silver Hill fault zone occurs along the Charlotte belt-Carolina slate belt boundary in North Carolina. The Raleigh belt is probably flanked on both sides by large faults. The precise movement sense of these faults is unknown; like the Brevard fault they possess an extended history of renewed movement. This group of faults may form a major system in the eastern Piedmont (Hatcher *et al.*, 1977).

The Piedmont of Virginia consists of the James River synclinorium on the southeast flank of the Blue Ridge anticlinorium; a broad belt, farther southeast, presumably anticlinal, of metagraywackes, schists and some metavolcanic rocks; then the Arvonia syncline of slates containing Ordovician fossils. To the southwest, in the western Piedmont along the Virginia-North Carolina line, there is an anticlinorium of basement rocks (Sauratown Mountains anticlinorium (Fig. 20.1, 4j). This feature is separated from the Blue Ridge here by the Smith River allochthon. Its relationship to the James River synclinorium and other features to the northeast is presently unknown.

The Piedmont of Maryland and Pennsylvania consists of an assemblage of pelitic and quartzofeldspathic rocks, metavolcanic rocks, carbonate rocks and diamictite of upper Precambrian to lower Paleozoic age known as the Glenarm Series (Hopson, 1964; Higgins, 1972). Earlier Precambrian basement rocks occur in the eastern Piedmont as the Baltimore Gneiss domes (Fig. 20.1, 4n) and the Mine Ridge anticlinorium. The latter is west of the Martic Line, while the Baltimore Gneiss domes are east of it. Ultramafic, gabbroic and granitic rocks of Paleozoic age intrude these rocks. Their exact ages are at present unknown because zircon age determinations made previously are in error because of inherited lead (M.W. Higgins, oral commun., 1976). Most of the gabbros and ultramafic rocks, along with the Baltimore Gneiss bodies may be tectonically transported to their present positions (Rodgers, 1970; A.A. Drake, written commun., 1976).

The Martic Line of Maryland and Pennsylvania separates the less intensely deformed platform assemblage of Cambrian and Ordovician carbonates and clastics to the west from the more extensively deformed and metamorphosed clastic-volcanic assemblage to the east. A major facies change appears to be associated with this structure as well (Wise, 1970; Higgins, 1972). The Martic Line is another feature in the Appalachians that has experienced a long and complex deformational history. Wise (1970) interpreted the line as a product of early imbricate thrusting followed by several events

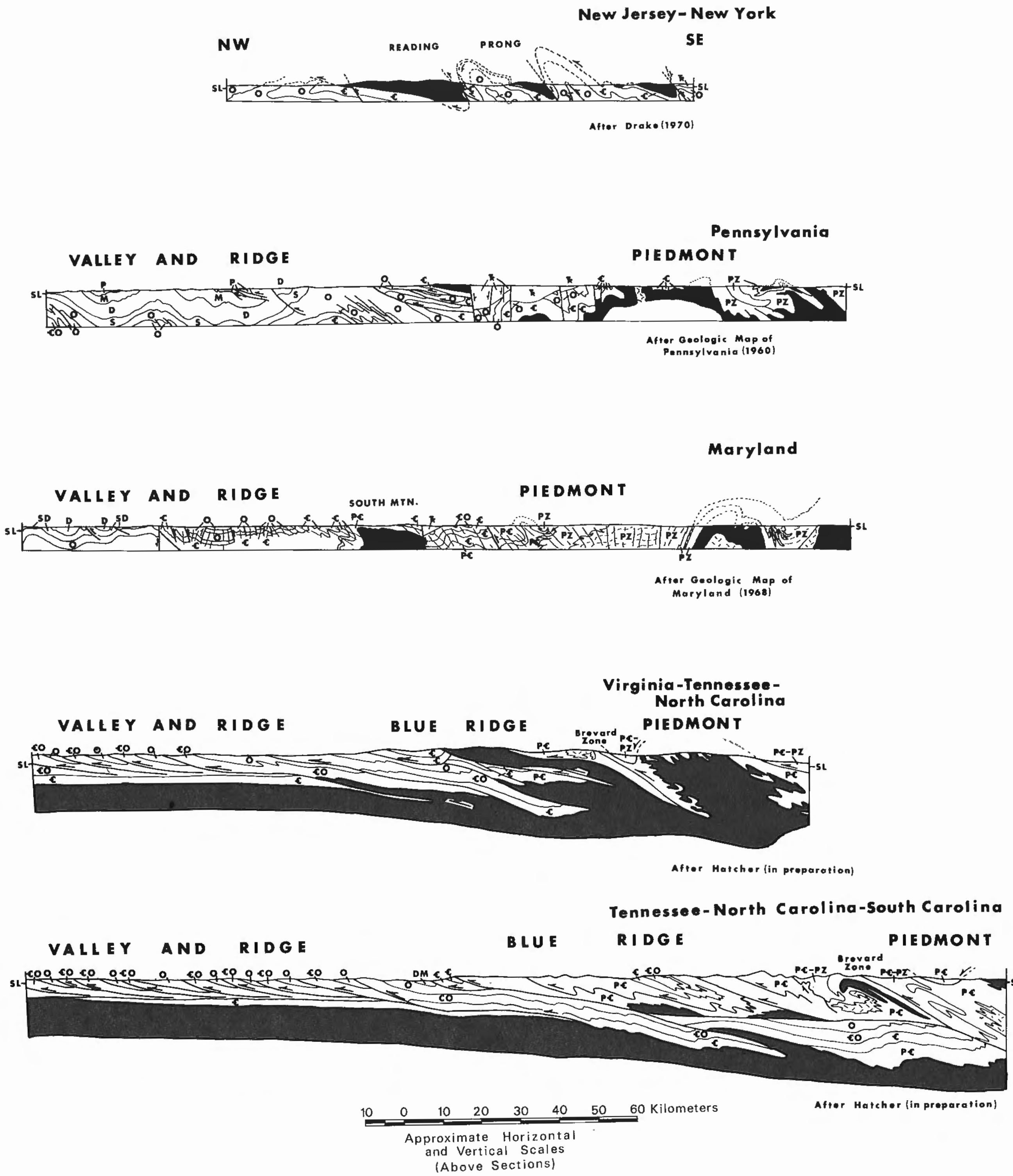

Figure 20.2a. (Caption on facing page)

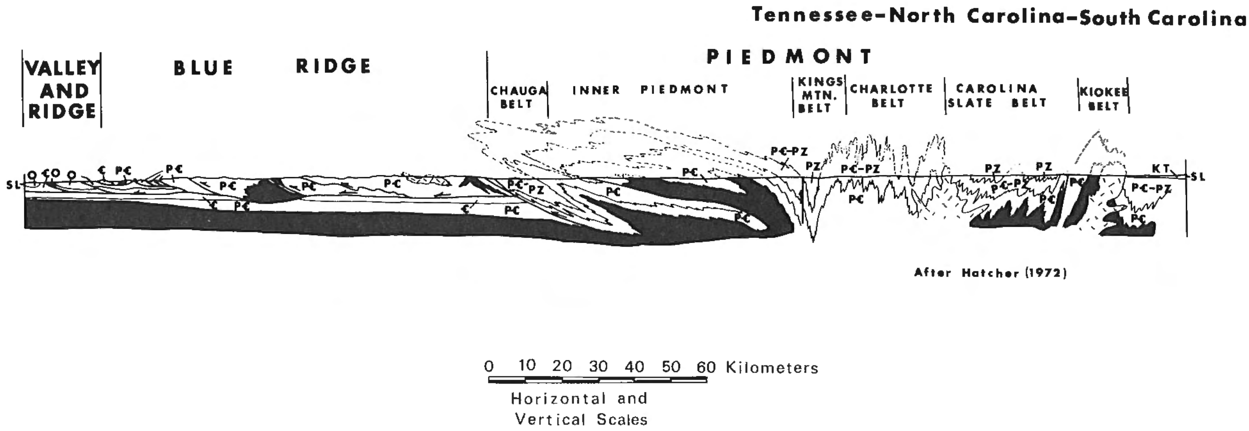

Figure 20.2b.

PЄ - Upper Precambrian sedimentary and/or volcanic rocks
PЄ-PZ - Upper Precambrian - lower Paleozoic metasedimentary and/or metavolcanic rocks
PZ - Paleozoic metasedimentary and/or metavolcanic rocks
Є - Cambrian sedimentary rocks

CO - Cambrian-Ordovician rocks
O - Ordovician sedimentary rocks
S - Silurian sedimentary rocks
D - Devonian sedimentary rocks
DM - Devonian-Mississippian sedimentary rocks
M - Mississippian sedimentary rocks
P - Pennsylvanian sedimentary rocks
TR - Triassic rocks

Basement rocks are black.
Paleozoic plutons are indicated by a cross-hatched pattern.

Figures 20.2a, b. Representative cross-sections through the central and southern Appalachians.

of later refolding. It may have played a more important role as a line along which smaller crustal blocks were rejoined to North America in the Early Paleozoic followed by the deformational history described by Wise (1970) and Higgins (1972).

Metamorphism in the Piedmont is Barrovian and the major thermal peak is probably Taconic (Ordovician). However, there is more than a suggestion of a less intense Acadian metamorphism in the central and southern Piedmont (Higgins, 1973; Butler and Fullagar, in press). An earlier, Precambrian metamorphism has also been clearly identified in basement rocks of the Piedmont (Wagner and Crawford, 1975). A Paleozoic greenschist overprint is present over much of the Piedmont and a later event involving filling of hydrothermal veins by zeolites, prehnite and calcite is just becoming known.

SUMMARY OF LATE PRECAMBRIAN THROUGH DEVONIAN EVOLUTION

The eastern margin of North America was rifted during the late Precambrian (Rankin, 1975, 1976) and several trailing fragments of continental crust were left in the wake of separation and formation of Iapetus. Some of these fragments were isolated (Charlotte belt-Carolina slate belt block) while others were left connected in peninsular fashion to North America (Inner Piedmont block). In the late Precambrian a subduction zone and island arc system formed adjacent to the easternmost of these fragments, which are now in Virginia and the Carolinas. Towards the continent there was at least one other fragment or peninsula of continental material but oceanic crust (small oceans, marginal seas, or back-arc basins) likely separated it from North America and the slate belt island-arc.

The initial collapse of these basins probably took place in the Cambrian, and this produced the initial stages of the Taconic orogeny. Thrusts in the Blue Ridge may have begun to move and the ultramafic bodies now on the eastern side of the Blue Ridge would have been emplaced. Some larger mafic-untramafic complexes, like the Baltimore Gabbro and several others in the North Carolina Blue Ridge may be fragments of oceanic crust and mantle caught up in the deforming mass when the small oceans were collapsing.

The Ordovician was a time of intense deformation, metamorphism and plutonism in the central and southern Appalachians. Tectonic lands rose from the deforming orogen to shed sediments to the west into the miogeocline. The island-arc system which had existed prior to this time may have been destroyed. However, plutons ranging in age from 425 and 385 Ma were probably formed above a west-dipping subduction zone. The earlier subduction zone beneath the island arc probably had an eastward dip (Glover and Sinha, 1973; Glover, 1976). The later west-dipping subduction zone is responsible for the final closure of Iapetus and the collision of North America with Africa. This latter event probably occurred to the north in the Devonian (Bird and Dewey, 1970), but may have happened later in the central and southern Appalachians (Hatcher, 1972a), producing the Alleghanian orogeny.

ACKNOWLEDGMENTS

Critical reviews by A.A. Drake, Jr., and J.R. Butler are very much appreciated.

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APPENDIX

PROBLEMS AND TOPICS FOR RESEARCH

1. Resolution of pre-Alleghanian structure in the Valley and Ridge.

2. Delineation of the areas affected by one and more than one Paleozoic metamorphic event in the Piedmont.

3. Resolution of pre- and synmetamorphic faults in the southern Blue Ridge and Piedmont.

4. Stratigraphic studies in medium to high grade metamorphic rocks to relate these rocks to lower grade equivalents in the southern Blue Ridge (more specifically, the relationships between the Ocoee Series, Ashe Formation, Tallulah Falls Formation, basement rocks and younger rocks).

5. Continuation and expansion of age dating programs to date the plutons of the Piedmont and Blue Ridge and determine the extent of basement rocks.

6. Detailed examination of critical boundaries, such as that between the Kings Mountain and Charlotte belts, to help solve questions related to the existence and locations of cryptic sutures.

7. Investigate the possible reasons for the allochthonous character of the southwestern portions of the Reading Prong and the autochthonous nature of the northeast portion.

8. Determine if the Baltimore gabbro complex is an ophiolite and its relationships to the James Run gneiss. Also, the relationships of Glenarm Series stratigraphy to the plutonic rocks of Pennsylvania, Maryland and northern Virginia need to be better resolved.

9. Chronology of deformation established in a few areas should be worked out in others and correlated throughout the Piedmont and Blue Ridge.

10. Combined geologic-geophysical studies to determine whether the Baltimore Gneiss domes are allochthonous or are in place.

11. Determine the nature of the southwest end of the Smith River allochthon. Its extent and relationships to the James River synclinorium and other features of the central Virginia Piedmont should also be determined.

Most of the problems outlined above could be resolved by additional quality geologic mapping. This is particularly true of the southern Appalachian Blue Ridge and Piedmont where only a few small areas of quality detailed geologic mapping involving integrated structural-stratigraphic-petrologic-geochronologic studies exist. Although these areas are increasing in size, there are very large areas where no modern geologic mapping exists.

SYNTHESE PROVISOIRE DES EVENEMENTS OROGENIQUES CALEDONIENS EN FRANCE

OROGÈNE CALÉDONIEN

Projet PICG n° 27

A. Autran

Bureau de recherches géologiques et minières
Service géologique national - B.P. 6009
Orléans Cedex 45018, France

INTRODUCTION

En dehors de l'Ardenne, terminaison occidentale du "Massif schisteux rhénan", où des déformations calédoniennes d'âge ante-dévonien sont connues et classiques depuis longtemps, différents géologues ont envisagé l'existence d'une véritable orogenèse calédonienne dans le Massif Central français, mais toujours à titre d'hypothèse et sur une argumentation fragile. Ce n'est que depuis 1973-1974 qu'un ensemble de faits nouveaux acquis en stratigraphie, cartographie et analyse structurale, pétrologie des séries métamorphiques et magmatiques et en géochronologie isotopique[1] a permis l'identification indiscutable d'une zone orogénique majeure d'âge Dévonien inférieur à moyen dans le Sud du Massif Armoricain, l'ensemble du Massif Central et probablement les Vosges et les massifs de socle des Alpes occidentales. (fig. 21.1)

On ne connait pas encore partout l'extension exacte du domaine affecté, ni le style des déformations, leur niveau structural, l'âge des formations métamorphisées.

Aussi cette mise au point a pour seule ambition de rassembler et tenter d'ordonner les faits actuellement certains que l'on peut corréler avec les évènements orogéniques calédoniens.

REGION N-E DE LA FRANCE: ARDENNE ET VOSGES

ARDENNE (figures 21.2, 21.3)

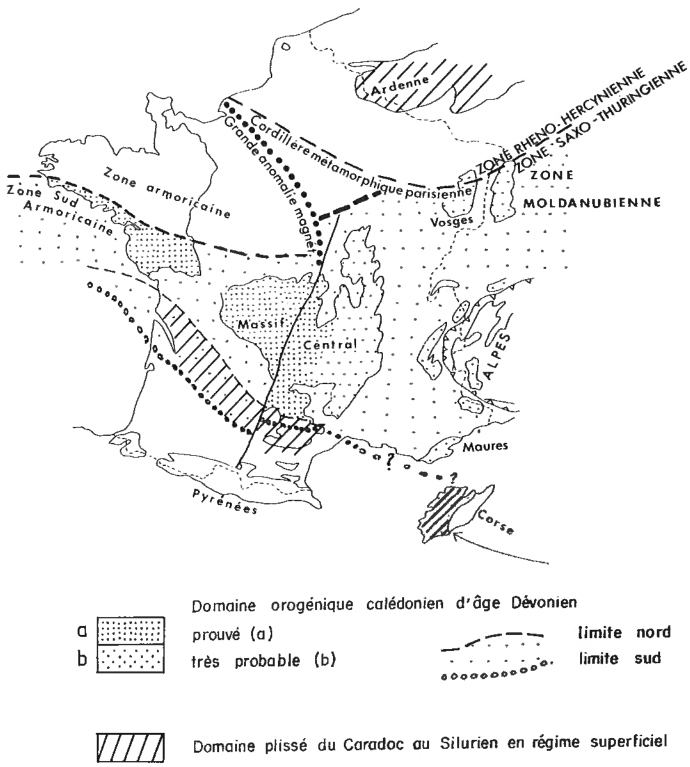
L'Ardenne est sur la bordure Nord de la zone Saxo-Thuringienne elle-même caractérisée par une série paléozoique continue du Cambrien basal au Carbonifère et des plissements uniquement carbonifères.

En Ardenne on observe le déplacement de l'époque des déformations calédoniennes du Sud vers le Nord depuis le Caradocien jusqu'au Silurien supérieur:

Ardenne méridionale: transgression du Dévonien basal (Downtonien) sur le Cambrien (Revinien-Devillien) plissé pendant le Caradoc comme le montre la discordance d'un flysch Caradocien typique sur le Llandelien dans la zone du Brabant au Nord (ride du Condroz).

Ardenne septentrionale et Brabant: discordance du Siegenien inférieur sur le Silurien.

Cette phase de plissement d'âge caradocienne à silurienne (accompagnée de volcanisme acide au Nord), n'a pas produit de schistosité au niveau d'érosion où on peut l'observer. Les plis ont une tendance à être déversés vers le Nord, mais leur reconstitution exacte n'est pas réalisée car la tectonique hercynienne post-westphallienne accompagnée de schistosité et de métamorphisme léger (au maximum zone à staurotide) les a beaucoup modifié. (voir figures 21.2 et 21.3)

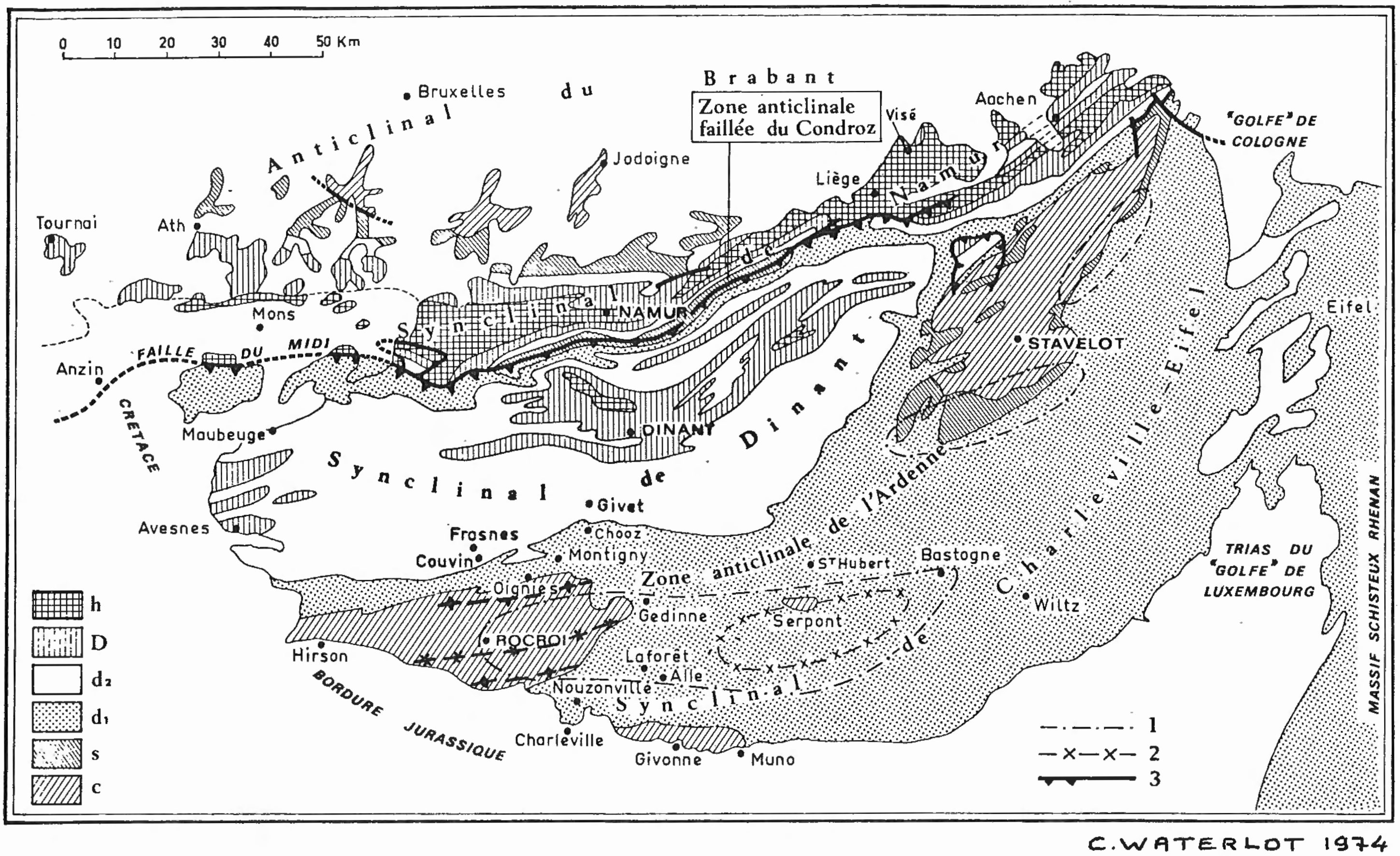


Les limites de la zone orogénique calédonienne du Dévonien ont été prolongées sur le socle caché sous les bassins mésozoiques grâce aux cartes de gravimétrie, aéromagnétisme et aux forages profonds (Autran, Gérard et Weber, 1976 - Bull. Soc. Géol. Fr. no.5).

Figure 21.1 Orogène calédonien de la France.

(1) Les âges isotopiques obtenus par la méthode Rb/Sr et donnés dans ce papier sont toujours des âges d'isochrone de roches totale calculés avec la constante de $1.47.10^{-11}an^{-1}$. Il conviendrait de les vieillir de 3,5% pour tenir compte de la nouvelle valeur proposée pour la constante $1,42.10^{-11}an^{-1}$.

PICG Projet 27, Contribution française n° 3.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).



h. terrain houiller. - D. Dinantien. - d_2. Dévonien supérieur et moyen. - d_1. Dévonien inférieur. - s. Siluro-Ordovicien. - c. Cambrien. - 1. Limite de l'épizone (à chlorite). - 2. Limite de la mésozone (à biotite). - 3. failles. - Plis calédoniens de la clinal de Rocroi: anticlinal synclinal

Figure 21.2 Carte schématique des principaux plis de l'Ardenne

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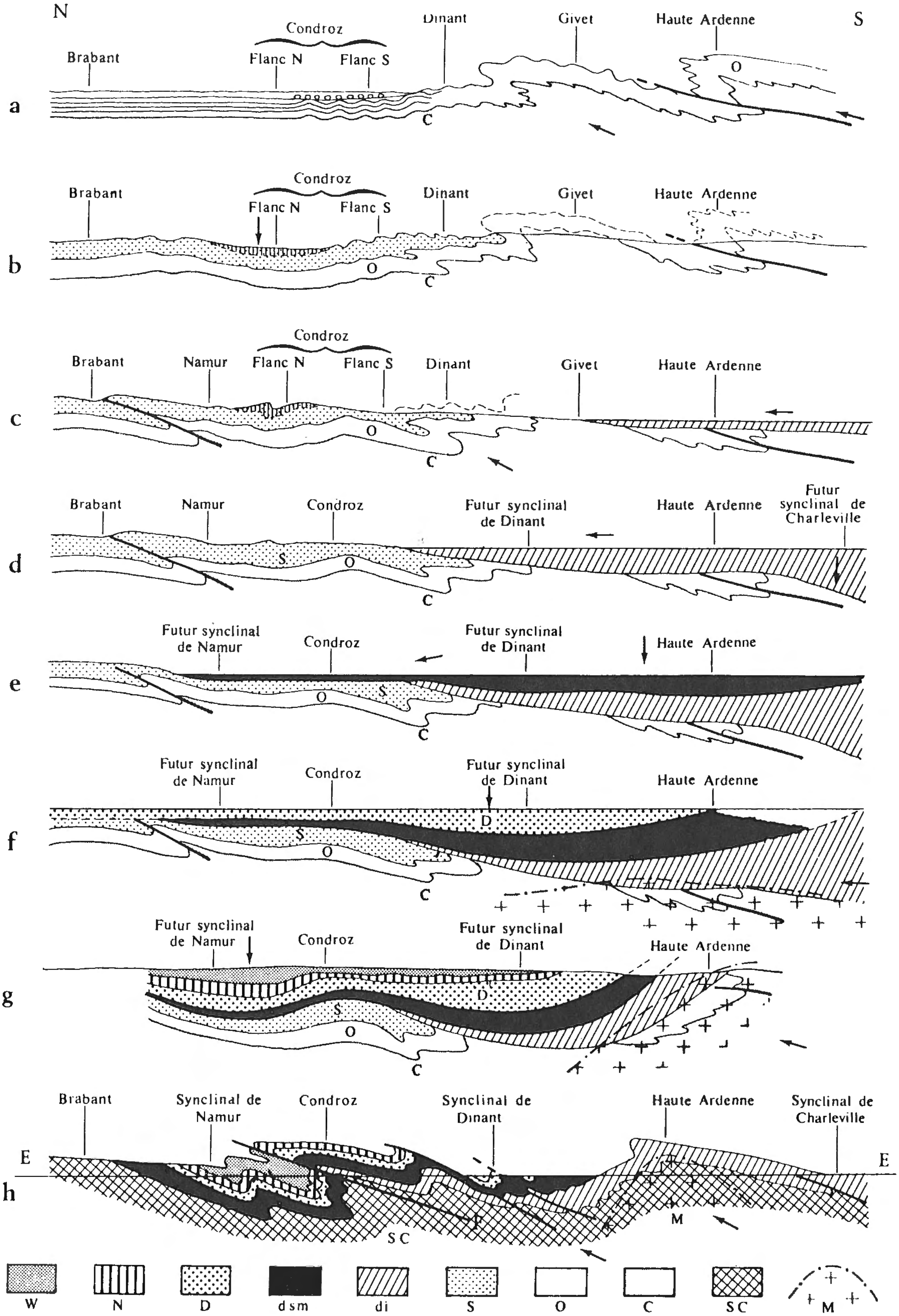
a) Au Caradocien, après la phase ardennaise (à rapprocher de la phase taconique): surrection de la Haute-Ardenne.
b) Au Ludlowien supérieur, après la phase condrusienne. Le Ludlowien supérieur est représenté par des hachures verticales en surimposition sur le figuré du Silurien.
c) Au Gédinnien inférieur (Downtonien), après la phase brabançonne.
d) Au Dévonien inférieur.
e) Au Dévonien moyen et supérieur.
f) Au Dinantien final avec la phase sudète (niveau de la "Grande Brèche").
g) Au cours du Westphalien (déformation du bassin de sédimentation lagunaire).
h) A la fin du Westphalien, avec la phase asturienne (formation des synclinaux de Namur et de Dinant, de la Grande Faille du Midi: exondation générale du pays).

W= Westphalien. - N = Namurien. - D = Dinantien. - dsm = Dévonien supérieur et moyen. - di = Dévonien inférieur. - S = Silurien. - O = Ordovicien. - C = Cambrien. - SC = Siluro-Cambrien indifférencié. - M = zone métamorphique de l'Ardenne. - flèches noires obliques = sens des poussées orogéniques. - flèches noires horizontales = sens des transgressions marines. - flèches noires verticales = zones de subsidence. - F = Grande Faille du Midi. - EE' = surface d'érosion actuelle.

Figure 21.3 L'Ardenne lors des différentes phases orogéniques des plissements calédoniens et hercyniens.



N
S
a
Brabant
Condroz
Flanc N
Flanc S
Dinant
Givet
Haute Ardenne
O
C
b
Brabant
Condroz
Flanc N
Flanc S
Dinant
Givet
Haute Ardenne
O
C
c
Brabant
Namur
Condroz
Flanc N
Flanc S
Dinant
Givet
Haute Ardenne
O
C
d
Brabant
Namur
Condroz
Futur synclinal de Dinant
Haute Ardenne
Futur synclinal de Charleville
S
O
C
e
Futur synclinal de Namur
Condroz
Futur synclinal de Dinant
Haute Ardenne
O
S
C
f
Futur synclinal de Namur
Condroz
Futur synclinal de Dinant
Haute Ardenne
D
S
O
C
g
Futur synclinal de Namur
Condroz
Futur synclinal de Dinant
Haute Ardenne
D
S
O
C
h
Brabant
Synclinal de Namur
Condroz
Synclinal de Dinant
Haute Ardenne
Synclinal de Charleville
E
E
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M
W
N
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di
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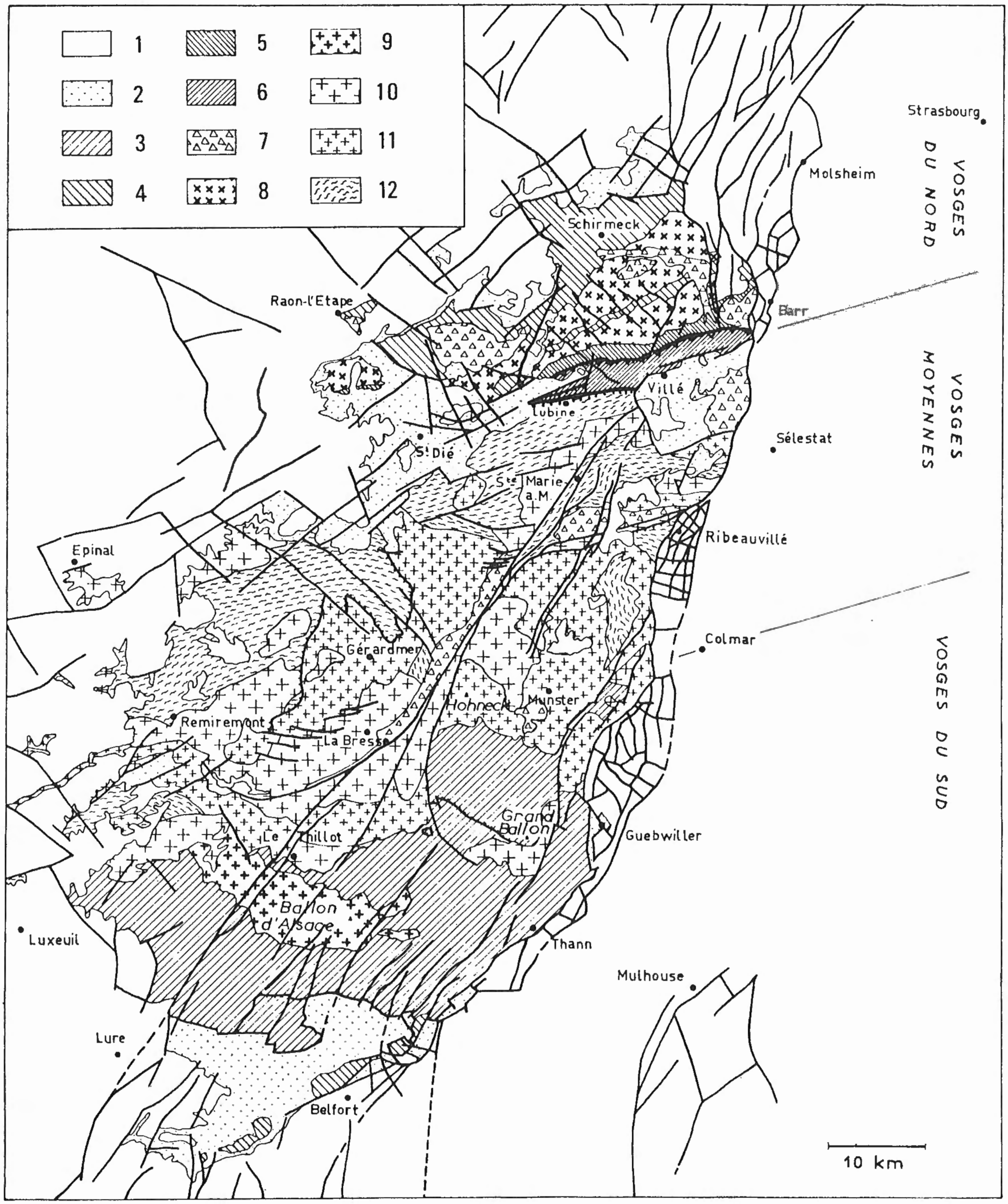
C
SC
M

Figure 21.3.



Les grands ensembles géologiques ont été distingués:

1. Terrains secondaires, tertiaires et quaternaires.
2. Permien.
3. Viséen des Vosges méridionales.
4. Dévonien du Belfortais et Dévono-Dinantien de la Bruche.
5. Siluro-Ordovicien.
6. Série de Villé et formations du Climont. (Protérozoique Sup^2-Eocambrien).
7. Granites intrusifs tardi-hercyniens.
8. Granitoides des Vosges du Nord. } Viséen à Westphalien
9. Granite des Ballons.
10. Granites intrusifs des Vosges moyennes
11. Granite fondamental et granito-gneiss. ~ 400 Ma
12. Gneiss et migmatites.

Figure 21.4. Carte structurale du socle vosgien (d'après J.-P. von Eller et coll., 1972 réduite et complétée par P. Fluck, 1973). On y distingue également les trois parties fondamentales du socle vosgien.

VOSGES (figure 21.4)

Ici on n'a pas décrit de déformations ni de métamorphismes au Cambrien, à l'Ordovicien ou au Dévonien. Mais il est très probable que certaines structurations tectono-métamorphiques majeures attribuées au cycle Cadomien (Protérozoique Supérieur) ou à un Hercynien précoce se sont produites lors d'une orogenèse calédonienne. Les indices pour cette hypothèse sont les suivants:

1) hiatus stratigraphiques: - Cambrien et Ordovicien inférieur
   - Silurien moyen supérieur et Dévonien inférieur.
2) transgression d'un mésodévonien dans les Vosges du Nord et du Sud auquel succède un important volcanisme à spilites et kératophyre jusqu'au Frasnien, dont les caractères géochimiques de ce volcanisme indiquent un site géodynamique de type arc insulaire.
3) âges isotopiques à 407-400 MA et 440 MA sur les granites et les roches anatectiques les plus anciens (Hameurt et Vidal, 1973; Bonhomme et Fluck, 1974).
4) lithologie du socle et évolution métamorphique des Vosges centrales et méridionales, très comparable à celles des autres massifs de la "zone moldanubienne" de l'orogène varisque, où depuis 4 ou 5 ans on découvre l'importance de la structuration calédonienne (Massif de la Forêt Noire en Allemagne).

LE MASSIF ARMORICAIN (figure 21.5)

Cette description est rédigée essentiellement à partir de la mise au point de Cogné (1976).

Il est constitué de deux grandes régions: (fig. 21.5) séparées par un accident majeur la "zone broyée Sud armoricaine".

Au Nord - "Le Massif armoricain" proprement dit, constitué d'un soubassement Précambrien récent (cycle cadomien achevé par la mise en place de granites vers 560 Ma), recouvert de dépôts paléozoiques épicontinentaux et de volcanites. Ce domaine a eu une paléogéographie variée pendant le Paléozoique. En relation avec l'orogenèse calédonienne, on ne peut y déceler aucune phase de plissement ni de métamorphisme régional depuis le Cambrien inférieur jusqu'au Dévonien supérieur-viséen où les premières déformations hercyniennes l'affecte (phase bretonne). Par contre des granites (trondjehmites - granites alcalins) s'y mettent en place à l'Ordovicien le long de la bordure sud, apparemment dans un contexte cratonique stable et des volcanites y sont connues à la limite Cambrien-Trémadoc, à l'Ordovicien supérieur, au Dévonien inférieur et à la limite Dévonien-Viséen.

Au Sud - les régions sud armoricaines sont formées d'épaisses séries métamorphiques épizonales à catazonales où aucune stratigraphie rigoureuse n'est encore connue au nord de la Loire. "Il reste difficile d'y séparer clairement ce qui relève d'une évolution cadomienne relique par rapport à une histoire tectono métamorphiqe polyphasée s'achevant avec les plissements hercyniens". C'est le domaine que J. Cogné propose d'appeler "orogène Ligérien", qui est aussi nommé "zone Vendéo-Sud Armoricaine".

On reviendra plus en détail sur cette région car c'est elle qui montre des indices importants d'une orogenèse calédonienne.

Entre ces deux régions: la zone broyée sud-armoricaine puissant complexe de blastomylonites et de métamorphites, jalonnée de leucogranites en lames syncinématiques et lobes.

Cette zone est le lieu de cisaillements importants d'abord tangentiels puis en faille transcurente, répartis sur plusieurs périodes au cours des temps hercyniens proprement dits.

LA STRUCTURATION CALEDONIENNE MAJEURE DE LA ZONE VENDEO-SUD ARMORICAINE

Les preuves les plus convaincantes se trouvent en Vendée où la stratigraphie et la structure tectonométamorphique montrent des relations remarquables élucidées par le travail cartographique de Ters (1976) (fig. 21.6).

La série est structurée en grands plis couchés pluri-kilométriques, déversés vers le Sud, d'axe EW.[1] Le Paléozoique représenté par un Cambrien moyen localement, et plus généralement par l'Ordovicien et le Silurien tous deux associés à des volcanites surtout acides, occupe le coeur des synclinaux où il est affecté des mêmes déformations que les terrains briovériens (protérozoique supérieur) plus profonds qui ont le même faciès lithologique que les schistes et micaschistes affleurant dans tout l'ouest de la zone Sud armoricaine. Le métamorphisme dont l'évolution est synchrone de ces déformations intense varie du faciès schiste vert au faciès amphibolite avec anatexie et est caractérisé par une évolution précoce à très haute pression (faciès "schiste bleu" à glaucophane et faciès éclogite). On ne peut tracer de discontinuité d'évolution entre les terrains antérieurs au Cambrien et le Silurien, seule l'intensité des recristallisations varie.

Donc ici la structuration tectono métamorphique très intense est d'âge post Ludlow.

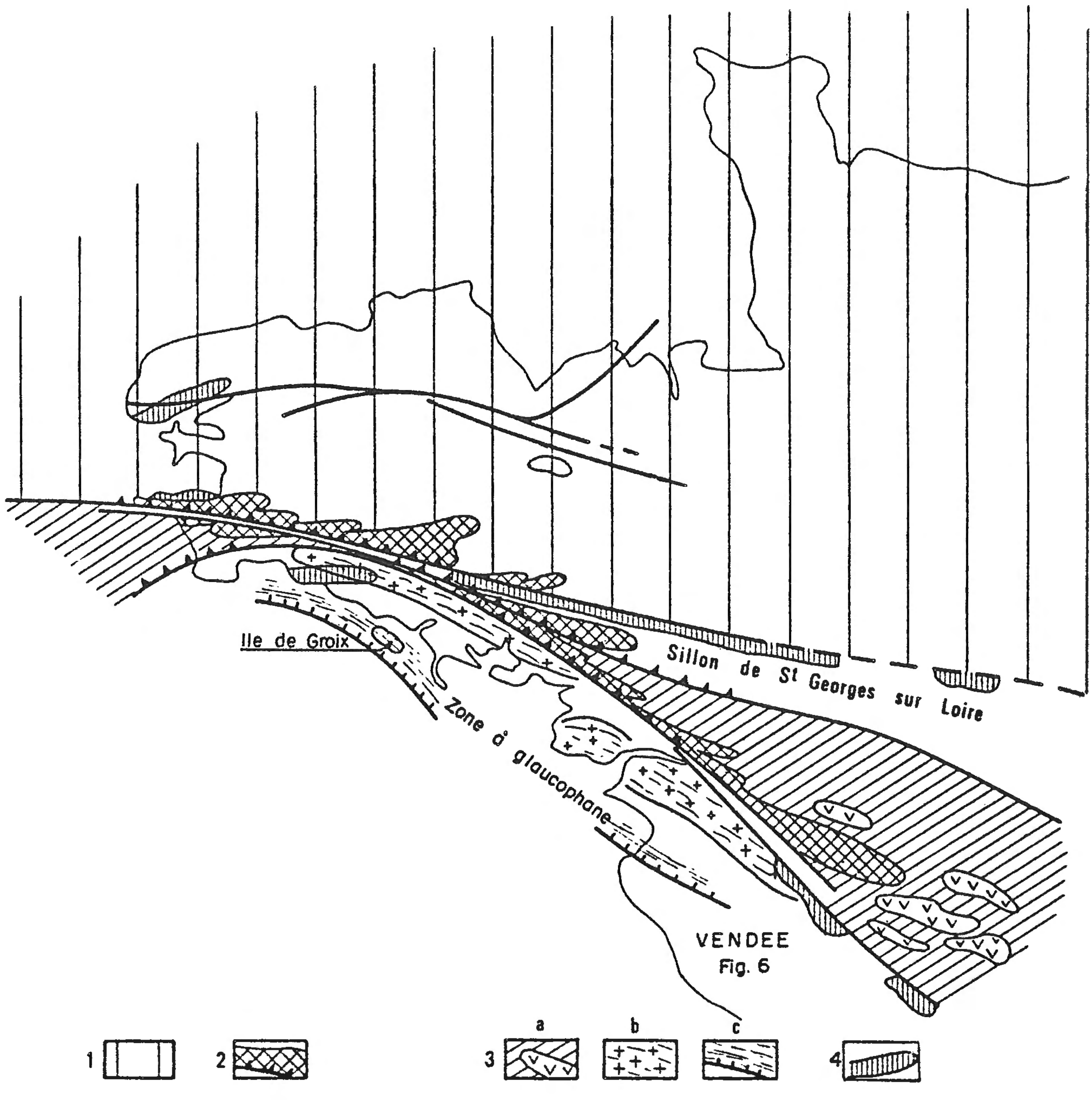
Un affleurement unique mais très symptomatique, montre à Ville Dé, un calcaire récifal Givétien non métamorphisé et simplement basculé par un plissement post Namurien, reposant sur des schistes épimétamorphiques cambriens affectés d'un plissement serré avec schistosité ardoisière, elle-même crénulée.

On a donc en Vendée un encadrement stratigraphique rigoureux pour une structuration orogénique majeure entre le Ludlow et le Givétien.

Par ailleurs, des granitoides d'anatexie cristallisés après la phase II des plis couchés EW (post-foliation), déversés vers le Sud, se trouvent tout au long de la zone Sud armoricaine; ils ont été datés par Ph. Vidal par isochrone Rb/Se de roche totale, à 365 Ma en Vendée et de 370 à 380 Ma au milieu de la zone (région de Morbihan).

La métamorphisme à glaucophane connu dans l'île de Groix, en Vendée et en quelques points par dragages en mer entre ces 2 secteurs, pourrait être aussi d'âge siluro-dévonien car il s'associe aux mêmes types de déformation que le métamorphisme de HP-HT affleurant plus au nord et les datations réalisées (Cogné et Peucat, 1977) suggèrent une homogénéisation isotopique Rb-Sr vers 420 Ma pour les roches totales (métavolcanites basiques et métashales) et la fin du métamorphisme à glaucophane vers 360 Ma. Il serait suivie à 320 Ma d'un réchauffement important (âge plateau unique par 39A/40A sur les glaucophanes) dont les nombreux massifs de leucogranite environnants, mis en place à cet âge, rendent facilement compte.

[1]Ces grands plis couchés replissent la foliation et sont donc considérés comme formés pendant une phase de déformation P_2. Les plis de la phase P_1, synfoliaux, sont plus rares et leur ampleur n'est pas connue.



Du Nord au Sud:

(1) Le "massif" armoricain proprement dit, partie intracratonique saxo-thuringienne de l'orogène hercynien: socle cadomien et couverture paléozoique épicontinentale.

(2) La Zone broyée sud-armoricaine, géosuture hercynienne chevauchante vers le Sud, dans le prolongement et au Sud du Sillon de Saint-Georges-sur-Loire, jalonnée par les leucogranites.

((3) L'Orogène ligérien, d'âge siluro-dévonien, constitué en marge méridionale du socle cadomien. On y reconnaît trois zones: (3a) dans les régions externes (Domaine ligérien): socle cadomien et magmatismes calco-alcalins anté- et pré-hercyniens; (3b) le long du bourrelet sédimentaire: métamorphisme de haute température et migmatisation (Domaine de l'Anticlinal de Cornouaille); (3c) dans la zone la plus interne: métamorphisme de haute pression - basse température (Ile de Groix - Vendée).

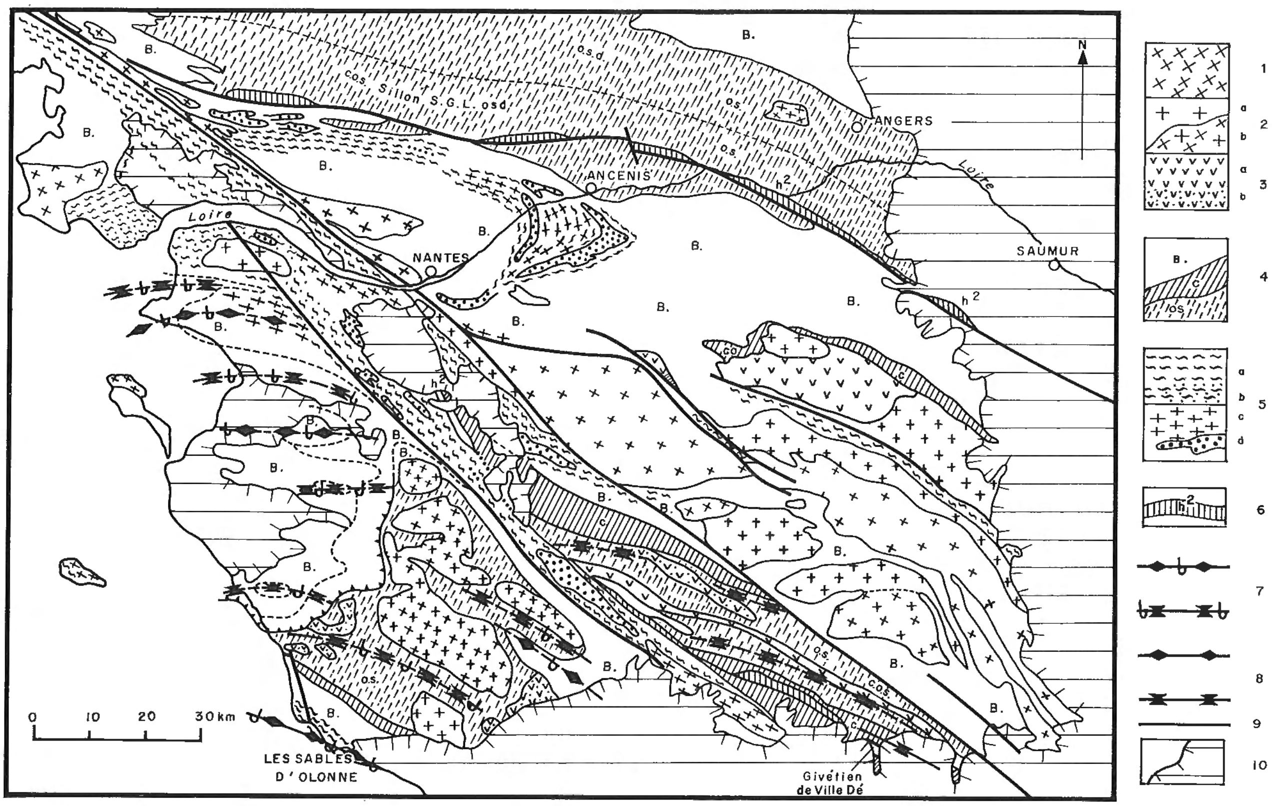
(4) Granites ordoviciens et siluriens.

Figure 21.5 La structure armoricaine

Ainsi les données actuellement recueillies, mais qui sont à renforcer, suggèrent une organisation structurale et métamorphique de cet orogène siluro-dévonien, voire strictement dévonien inférieur, apparentée au type "paired metamorphic belts" de Myiashiro, et plus particulièrement aux modèle impliquant une subduction au moins momentanée de croûte continentale (fig. 21.7).

Remarquons enfin la disposition, immédiatement au Nord de ce dispositif orogénique, d'un sillon ou d'une fosse à sédimentation détritique très épaisse, accompagnée d'un abondant volcanisme marin, qui a fonctionné pendant l'Ordovicien, le Silurien, au moins jusqu'au Ludlow. C'est le sillon de Saint Georges sur Loire qui pourrait être considéré comme une fosse de distension arrière d'un cordillère orogénique. Sa limite Nord est d'ailleurs jalonnée par une longue série d'intrusions granitiques d'âge Ordovicien à Silurien (450 à 520 Ma): Massif des landes de Lanvaux-Angers (fig. 21.5 et 21.7).

Ce sillon très intensément plissé au Carbonifère, peut-être déjà au Dévonien et cisaillé par la dislocation Sud Armoricaine devait représenter une barrière faunistique importante à l'Ordovicien, Silurien et Dévonien car les terrains de cet âge déposés au Sud (synclinal d'Ancenis, Vendée) ont des faunes stricte-



1. Granites varisques datés à 320 Ma, (généralement leucogranites à deux micas).

2a. Granites calco-alcalins non datés pouvant êtres dévonien supérieur. 2b: idem. de type alumineux leucocrate.

3a. Volcanites à la limite Ordovicien-Cambrien. 3b: idem. au Wenlock.

4. Formations épimétamorphiques d'âge briovérien supérieur (B) ou paléozoique: c = cambrien moyen, o = ordovicien, s = silurien, d = dévonien inférieur, hl = faménien à viséen (faciès culm).

5. Formations très métamorphiques d'âge probablement briovérien: a = faciès amphibolite, b = zones anatectique, c = orthogneiss datés à 450 Ma et à 550 Ma; d = amphibolites et éclogites.

6. Bassins de Namurien et Westphalien pincés le long des branches de la zone de cisaillement sud armoricaine.

7. Axes des mégaplis anticlinaux et synclinaux couchés de la phase II anté givetienne en Vendée occidentale.

8. Idem pour les plis droits ou simplement déversés au Sud de la Vendée centrale.

9. Failles: le grand cisaillement sud armoricain traverse la carte en diagonale du NW ou SE.

10. Transgression mésozoique ou tertiaire.

S.G.L.: sillon de Saint Georges sur Loire

S : Saumur A: Angers AN: Ancenis N: Nantes.

Figure 21.6. Structuration d'âge dévonien en Vendée. (A. Autran sur les indications de M. Ters).

ment apparentées à celles de Bohème centrale (zone Moldanubienne) qui ne se retrouvent pas dans la zone Centre et Nord armoricaine.

Le Massif armoricain montre donc la bordure Nord de la zone orogénique d'âge Calédonien tardif qui pourrait être l'équivalent des phases "erian" ou plutôt "acadian orogeny" des Appalaches du Nord.

Vers l'Est une centaine de kilomètres de dépôts mésozoiques cache la liaison avec le Massif Central. La limite Nord (sillon de Saint Georges sur Loire) peut se suivre en géophysique, (gravimétrie et magnétisme) jusqu'à la grande anomalie magnétique du Bassin de Paris (Weber, 1973).

C'est dans le Sud du Massif Central que se trouve la limite sud de cette même zone orogénique dont la largeur d'affleurement atteint environ 300 km.

Quant aux relations avec le cycle hercynien, la zone Sud armoricaine montre la continuité des 2 cycles:

- L'analyse précise de la succession des déformations et de l'évolution des conditions métamorphiques par rapport à celles-ci conduit en effet à considérer que la zone Sud armoricaine, du moins la partie Nord affectée par le métamorphisme de H.T., est restée dans des conditions thermiques intenses depuis le Dévonien (anatexie autochtone datée vers 370 Ma) jusqu'au Carbonifère, où se placerait la 3è phase de déformation caractérisée par un plan axial raide où déversé vers le Sud, dont l'encadrement d'âge est défini à présent de façon peu précise entre 360 et 290 Ma.

Cette évolution métamorphique, apparemment continue du Dévonien au Carbonifère, est aussi enregistrée dans l'importance de la "phase Bretonne" à la limite Dévonien/Viséen dans le Bassin de Chateaulin et le Léon (NW. de la zone centrale armoricaine) où 2 phases de déformations intenses post Strunien, avec métamorphisme faible à moyen, précédent le dépôt du Viséen Supérieur et du Namurien, eux-mêmes affectés par le plissement anté-Westphalien D.

Dans toute cette région on ne peut donc pas séparer clairement les cycles calédoniens et varisques surtout dans les domaines profonds de l'orogène.

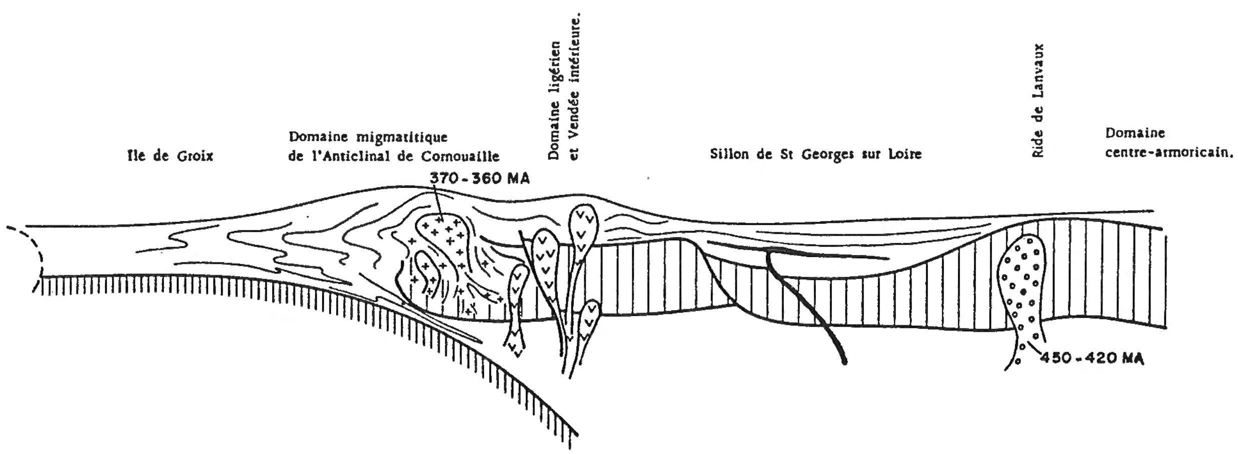
LE CYCLE CALEDONIEN DANS LE MASSIF CENTRAL

INTRODUCTION

Le socle du Massif Central (fig. 21.8 à 21.10) est constitué de terrains métamorphiqes dont l'âge stratigraphiqe est inconnu presque partout. Seule la région Sud (massif de la Montagne-Noire) montre une série Paléozoique assez complète du Cambrien inférieur au Viséen. Elle permet d'assurer que dans cette zone il n'y a eu que des déformations calédoniennes très modérées, l'essentiel de la tectonique et du métamorphisme s'étant développés après le dépôt du Viséen. Il semble y exister une lacune (ou une sédimentation très condensée) au niveau du Cambrien supérieur; on y constate aussi la transgression du Silurien supérieur ou du Dévonien inférieur sur différents termes du Cambrien au Caradoc, mais contrairement à la région Ardennaise la série infra-dévonienne n'était pas plissée avant (Arthaud, 1970; Hamet et Allegre, 1976).

Au Nord de la Montagne Noire on connait 3 types de dépôts de Paléozoique occupant des surfaces très réduites (fig. 20.10):

- à la bordure Ouest: une série mince où l'Ordovicien et le Silurien supérieur sont caractérisés est impliquée comme en Vendée dans les mêmes déformations que les terrains azoiques métamorphisés sur lesquels elle repose en discordance cartographique.



D'âge Siluro-Dévonien, cet orogène sud-amoricain présente des caractères cordillérains: du Sud vers le Nord:

- une zone en subduction responsable du métamorphisme de haute pression - basse température de l'Ile Groix (420 - 415 Ma).

- un bourrelet sédimentaire en marge du socle cadomien avec métamorphisme de haute température et migmatisation (diapirisme infrastructural et granites anatectiques du Domaine de l'Anticlinal de Cornouaille (370 - 360 Ma).

- les traces d'un arc magmatique calco-alcalin en Vendée intérieure.

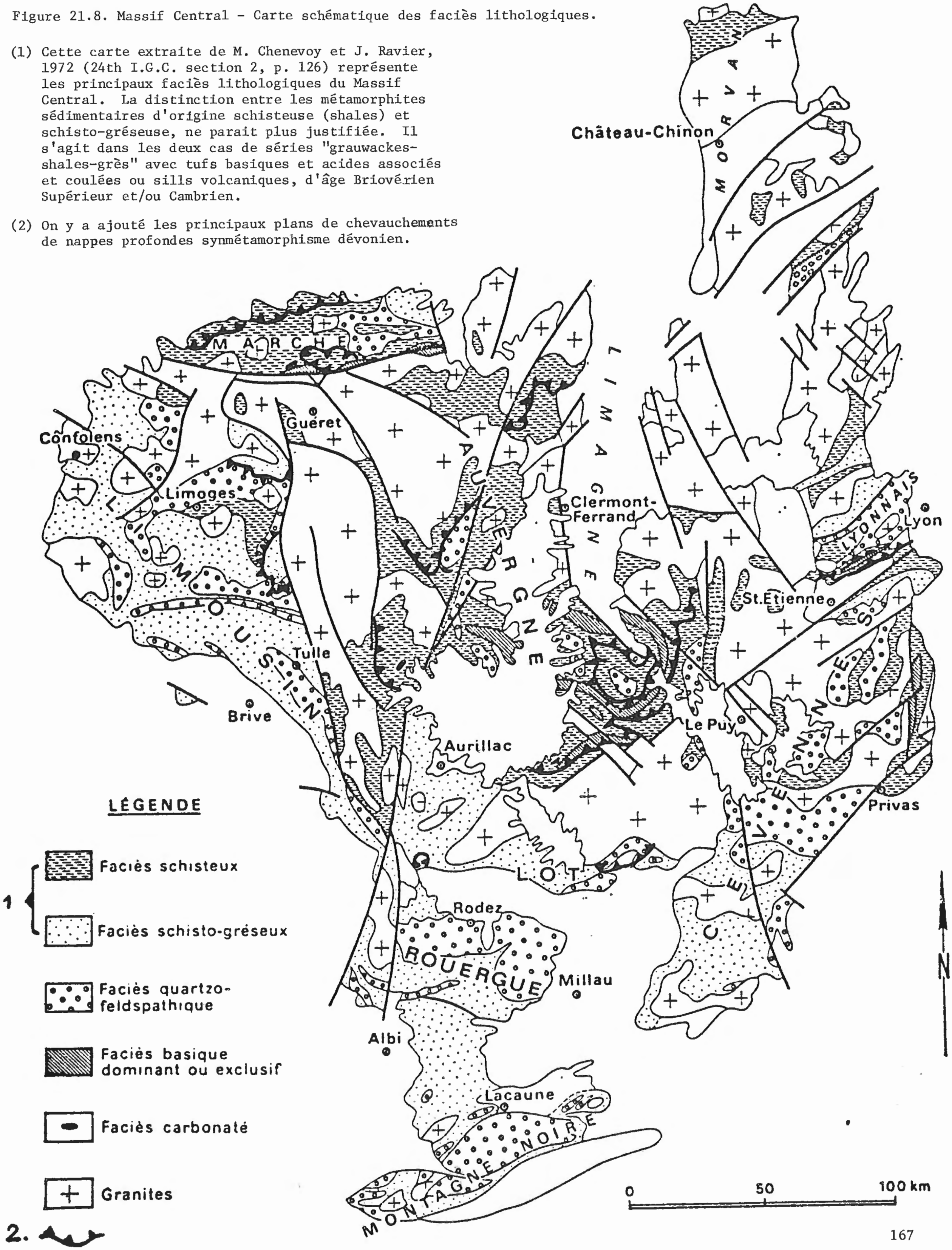
- un sillon marginal externe, lieu d'intrusions volcaniques basiques lors des périodes de distension et bordé d'intrusions alcalines (Sillon de Saint-George-sur-Loire et Ride de Lanvaux).

Figure 21.7. Coupe schématique de l'orogène Sud-Armoricain avant les serrages et cisaillements varisques.

Figure 21.8. Massif Central - Carte schématique des faciès lithologiques.

(1) Cette carte extraite de M. Chenevoy et J. Ravier, 1972 (24th I.G.C. section 2, p. 126) représente les principaux faciès lithologiques du Massif Central. La distinction entre les métamorphites sédimentaires d'origine schisteuse (shales) et schisto-gréseuse, ne parait plus justifiée. Il s'agit dans les deux cas de séries "grauwackes-shales-grès" avec tufs basiques et acides associés et coulées ou sills volcaniques, d'âge Briovérien Supérieur et/ou Cambrien.

(2) On y a ajouté les principaux plans de chevauchements de nappes profondes synmétamorphisme dévonien.



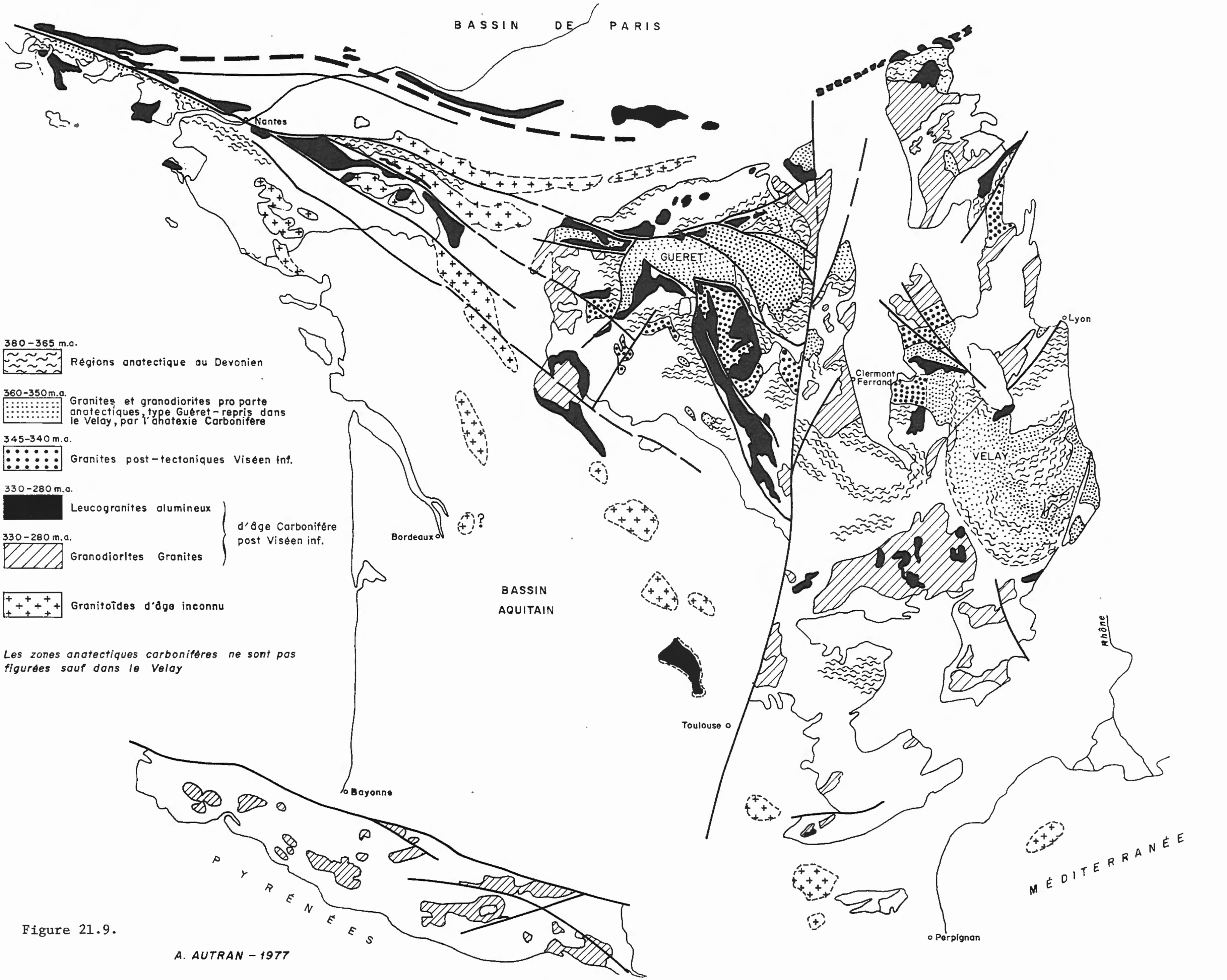

BASSIN DE PARIS
Nantes
GUERET
Clermont Ferrand
Lyon
VELAY
Bordeaux
BASSIN AQUITAIN
Toulouse
Bayonne
Perpignan
Rhône
MÉDITERRANÉE
PYRÉNÉES
380-365 m.a.
Régions anatectique au Devonien
360-350 m.a.
Granites et granodiorites pro parte anatectiques, type Guéret-repris dans le Velay, par l'anatexie Carbonifère
345-340 m.a.
Granites post-tectoniques Viséen inf.
330-280 m.a.
Leucogranites alumineux
330-280 m.a.
Granodiorites Granites
d'âge Carbonifère post Viséen inf.
Granitoïdes d'âge inconnu
Les zones anatectiques carbonifères ne sont pas figurées sauf dans le Velay
A. AUTRAN - 1977

Figure 21.9.

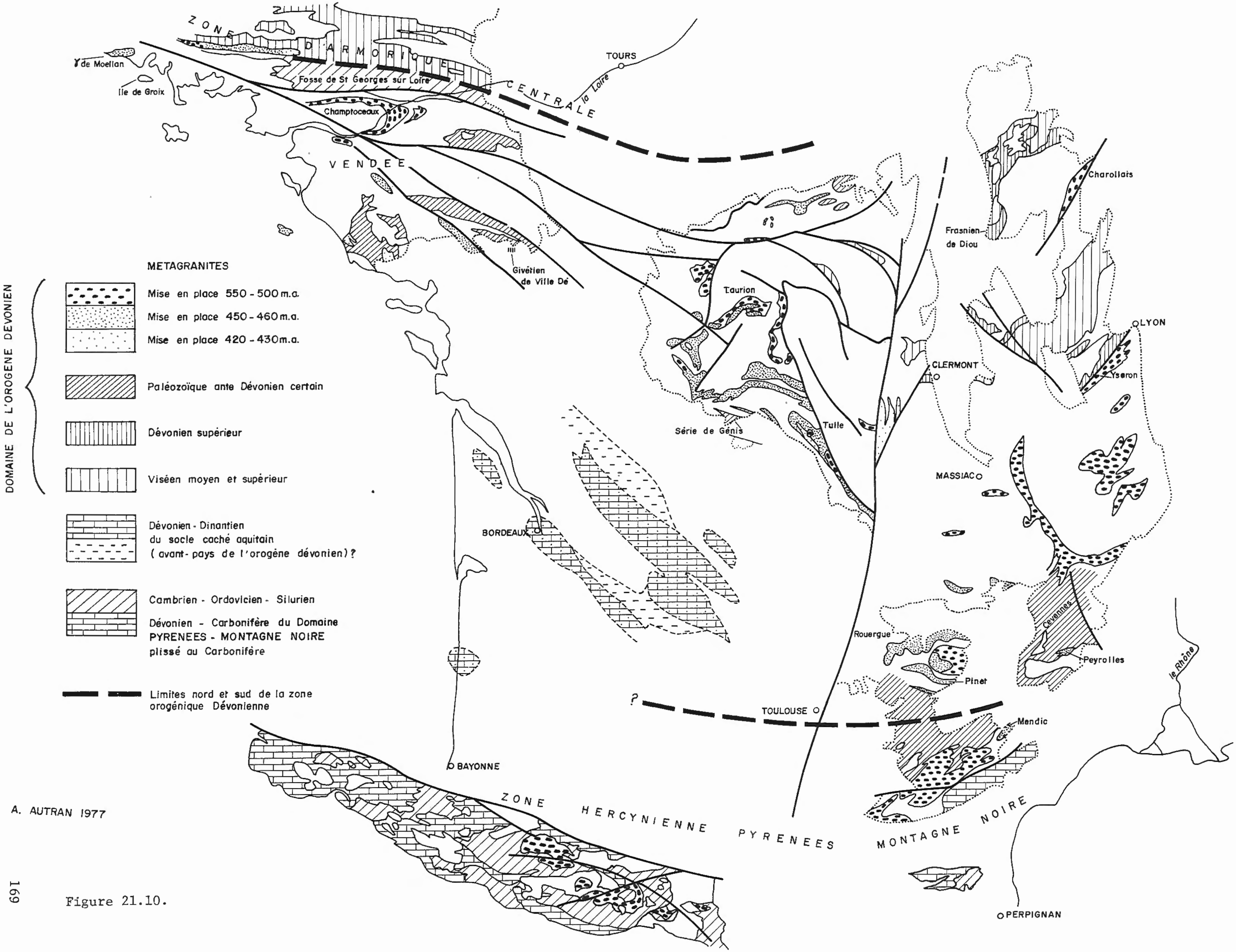

ZONE D'ARMORIQUE
CENTRALE
TOURS
la Loire
de Moellan
Ile de Groix
Fosse de St Georges sur Loire
Champtoceaux
VENDEE
Givétien de Ville Dé
Taurion
Série de Génis
Tulle
CLERMONT
MASSIAC
Frasnien de Diou
Charollais
LYON
Yseron
BORDEAUX
BAYONNE
TOULOUSE
Rouergue
Pinet
Mendic
Cevennes
Peyrolles
le Rhône
PERPIGNAN
ZONE HERCYNIENNE PYRENEES MONTAGNE NOIRE
METAGRANITES
Mise en place 550 - 500 m.a.
Mise en place 450 - 460 m.a.
Mise en place 420 - 430 m.a.
Paléozoïque ante Dévonien certain
Dévonien supérieur
Viséen moyen et supérieur
DOMAINE DE L'OROGENE DEVONIEN
Dévonien - Dinantien du socle caché aquitain (avant-pays de l'orogène dévonien)?
Cambrien - Ordovicien - Silurien
Dévonien - Carbonifère du Domaine PYRENEES - MONTAGNE NOIRE plissé au Carbonifère
Limites nord et sud de la zone orogénique Dévonienne
A. AUTRAN 1977

Figure 21.10.

- au N.E. une série marine d'âge Givétien-Frasnien-Faménien avec des volcanites à spilite kératophyre associées, repose sur le socle métamorphisé érodé. Cette série est généralement peu déformée sauf près de Lyon où elle est affectée d'un plissement très intense avec métamorphisme modéré, avant la transgression du Viséen moyen (phase bretonne de l'orogenèse varisque).

- le Viséen moyen et supérieur recouvre d'un épandage clastique molassique (Culm) puis d'un manteau d'ignimbrites rhyodacitiques tout le N. et N.E. du massif où ces dépôts sont préservés dans des gouttières synclinales fréquemment bordées de failles de cisaillement tardihercyniennes, de même que les bassins lacustres de Stéphanien à charbon.

Tous ces dépôts du Dévonien supérieur au Stéphanien sont en discordance structurale très nette sur le socle métamorphiques érodé auparavant par endroit, jusqu'au niveau des roches anatectiques (Lyonnais).

On trouvera le détail des séries métamorphiques du socle et l'historique des interprétations structurales dans la description de Chenevoy (1974)[1]. Nous présentons sur les figures 21.9 à 21.10 des faits qui nous paraissent significatifs de l'évolution du massif. Le fond géographique des figures 21.9 et 21.10 a été restitué par coulissage des accidents stéphaniens, dans l'état approximatif au Carbonifère).

Les principales études régionales et les synthèses successives publiées sur ce massif ont déjà envisagé une structuration tectono-métamorphique calédonienne pour une grande partie en basant cette interprétation, essentiellement sur deux types d'arguments:

1) La transgression du socle métamorphique central et Nord oriental par du Dévonien supérieur (Givétien-Frasnien) ou du Dinantien.

2) L'accumulation d'âges isotopiques par le laboratoire de géochronologie de Clermont-Ferrand (M. Roques) dans l'intervalle 550-360 Ma.

(travaux de Roques (1971); Chenevoy (1974); Cantagrel (1973); Collomb (1969).

Depuis 1970, 4 équipes ont recommencé les recherches géologiques et une cartographie précise du socle N.O. du Massif Central: équipe du Service géologique national (Autran, Chantraine, Boissonnas); équipes des Universités d'Orléans (Grolier, Guillot); de Limoges (Floc'h, Santallier) et de Lyon (Chenevoy, Piboule et leurs étudiants), enfin des recherches très poussées en géochronologie K/A et Rb/Sr étaient menées simultanément en partie en coopération, par Cantagrel, Duthou et Bernard-Griffiths, du centre de géochronologie de Clermont-Ferrand.

On dispose à présent de suffisamment de données nouvelles en stratigraphie, géochimie, pétrographie, évolution métamorphique, analyse structurale, datations isotopiques et cartographie pour apporter la preuve que la structuration tectonique et métamorphique majeure du N.O. du Massif Central s'est réalisée au Dévonien.

Après avoir montré la cohérence de l'ensemble des données obtenues par les diverses méthodes d'études des zones profondes dans cette partie du massif, nous exposerons plus brièvement les arguments qui permettent de retrouver ailleurs des signes de la même évolution et la limite méridionale de cette zone orogénique du Dévonien.

EVOLUTION TECTONO-METAMORPHIQUE DU N.O. DU MASSIF (Autran et Guillot 1974, 1975) (figure 21.11).

Tectonique: On a reconnu 4 à 5 phases de déformation superposées qui affectent toutes le Silurien daté.
Aucune phase plus ancienne que le Silurien n'a put être caractérisée. Les phases P1 et P2 accompagnées de métamorphisme régional sont les plus intenses. Elles se sont produites entre le sommet du Silurien (~ 400 Ma) et 362 Ma, âge obtenu par Rb/Sr sur des granites de fusion anatectique post P2. Par ailleurs les amphiboles donnent par K/A un âge de fermeture à 350 Ma. P1 à donné partout des plis couchés avec foliation ou schistosité de plan axial. C'est la phase la plus déformante. Une seule mégastructure P1 parait indiquer un déversement vers le Sud. La phase P2 plisse la foliation en direction ~ EW, en plis ouverts dans la région Sud devenant serrés et déversés vers le Nord dans l'ensemble du Limousin. Les phases 3 et 4 accompagnent la mise en place des massifs de granite d'âge Namuro-westphalien (320-300 Ma). Ce sont des plis droits post-métamorphisme, dans un niveau structural élevé, éventuellement accompagnés d'une métamorphisme de contact ou hydrothermal. Leurs directions mal réglées (~ 50 et ~ 150 à NS) interfèrent avec les grandes structures antérieures P1 et P2 qui étaient généralement en direction EW à NW-SE, cela donne les arcs, dômes et synclinoriaux pincés caractéristiques de la carte du Limousin.
Une cinquième phase P2 correspond à des antiformes et synformes, à plan axial vertical, contemporaines de la fin du métamorphisme au Dévonien Supérieur - Viséen.

Le métamorphisme régional: il montre de spectaculaires variations de ses conditions plysiques dans l'espace et le temps - Il est moins intense et se termine avant P2 dans la zone Sud, où déjà lors de P1, les plis n'étaient pas couchés. Ailleurs il débute partout, sauf peut être dans l'extrême N.O., par des conditions de HP-HT acquises de façon très précoce pendant les mouvements P1 (éclogitisation, localement faciès [disthène-orthose] (saxonien) et granulites de haute pression), puis partout au cours de P2 et après P2 la pression diminue et dans certaines mégastructures P2 les isogrades ante P2 sont entrainés dans le plis P2 et dessinent de vastes secteurs à polarité métamorphique renversée par la phase 2 sur lesquels s'établit une recristallisation statique hétérozonale post P2. La remontée du bati en surface est bien datée par le K/A sur les amphiboles des amphibolites vers 360-350 Ma (Cantragrel, 1973).

[1] Les données lithostratigraphiqes sur les régions métamorphiques couvrant l'ensemble du Massif Central sont encore très contradictoires. Elles sont évoquées dans cette note, mais feront l'objet d'une mise au point ultérieure du groupe de travail français PIGC no. 27.

Les granites: On connait par étude pétrologique et géochimique et on a daté par Rb/Sr (isochrones de R.T.) Duthou (1977) et Bernard-Griffiths (1976) trois générations de métagranites orthogneissifiés par P1: (figure 21.10)

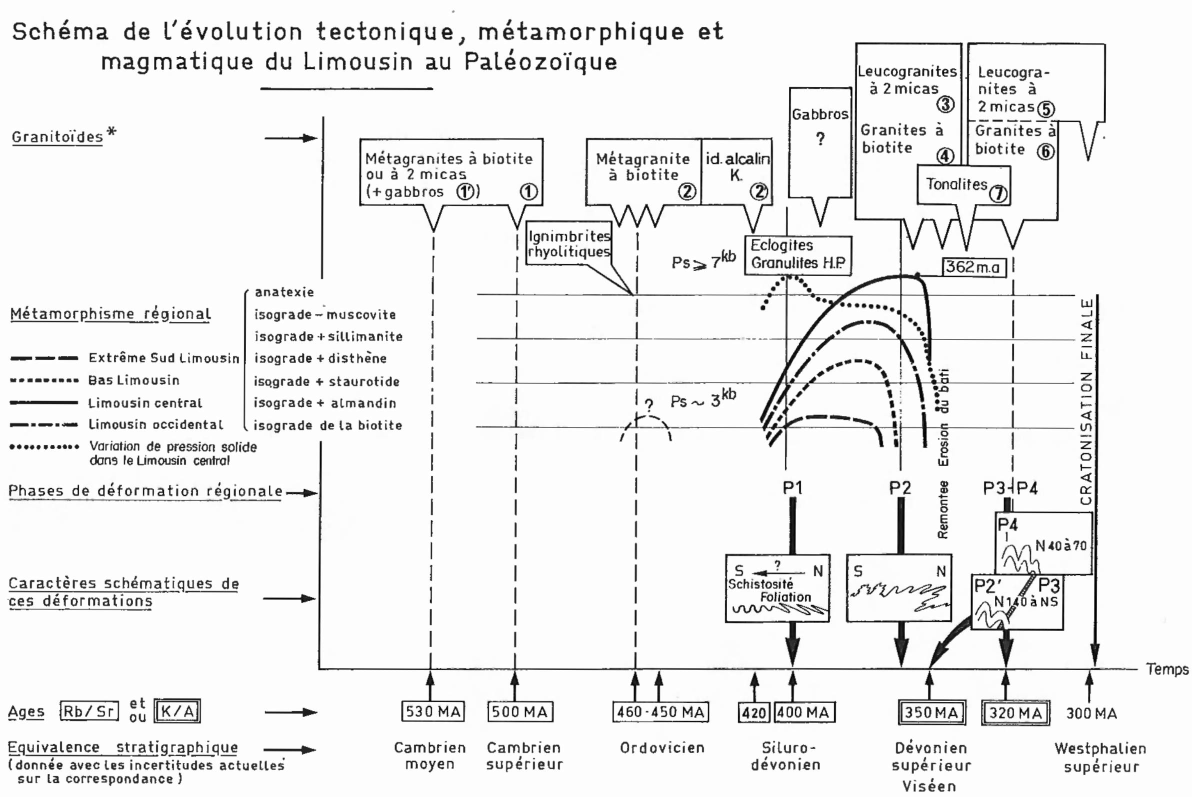
- granites cambriens datés à 520 et 500 Ma
- granites ordoviciens datés à 450 Ma
- granites siluriens datés à 420 Ma

Les granites syn à post P2, mais antérieurs au Viséen moyen sont très nombreux (figure 21.9) depuis des granites profonds liés aux zones anatectiques, jusqu'à des granites assez superficiels. Leurs âges par isochrone Rb/Sr vont de ⩾ 360 Ma à 340 Ma. Ces granites ont une origine essentiellement anatectique. Il est très possible que leur énorme volume soit le résultat d'une fusion intense du socle, par suite de sa remontée quasi adiabatique tardi-orogénique, après son enfouissement à plus de 30 km de profondeur lors des phases précoces à H.P. de l'orogenèse. Ensuite comme dans le centre et le Nord du Massif Central on a des granites viséens vers 335 Ma, peut-être contemporains des épanchements d'ignimbrites si caractéristiques du Viséen supérieur et surtout l'abondant cortège des granites à 320-300 Ma comprenant des tonalites, des granodiorites et des monzogranites d'une part, des massifs de leucogranites per-alumineux allochtones d'autre part.

Au niveau d'érosion accessible c'est ce plutonisme granitique carbonifère et les phénomènes thermiques qui lui sont liés, qui sont les attributs significatifs de l'orogenèse hercynienne par rapport à celle qui l'a précédée au Dévonien et qui est responsable de la structuration tectono-métamorphique essentielle du socle.

INDICES D'UNE EXTENSION DE STRUCTURATION DEVONIENNE A LA MAJEUR PARTIE DU MASSIF-CENTRAL (fig. 21.9 et 21.10)

On retrouve les caractères du métamorphisme et des déformations associées qui viennent d'être datés en Limousin dans toute la zone qui avait été appelée "noyau arverno - vosgien" ou zone "Lemovico-arverne" (Roques, 1971). Contrairement aux interprétations de Chenevoy (1974) il ne nous est jamais apparu indiscutable qu'un métamorphisme plus ancien de forte intensité, dont l'âge pourrait être cadomien, ait précédé le métamorphisme Dévonien. Ce point demande des études approfondies pour distinguer un métamorphisme plurifacial mono-orogénique (tel celui qui est mis en évidence en Limousin), d'une superposition de 2 métamorphismes polyorogéniques. De telles études sont en cours dans le Rouergue, le centre et l'Est du massif, de même que les études structurales précises nécessaires pour relier de proche les régions.



* *Granitoïdes* : ① *Orthogneiss œillés de l'arc du Taurion, de Tulle, de St. Germain les Belles - Orthogneiss à grain grossier de l'arc du Taurion et de Nexon* ①' *Gabbro du Monteil.* ② *Granite du Saut du Saumon orthogneissifié - Orthogneiss leptyniques.* ③ *Leucogranite de La Brame (syncinématique de P2) - Leucogranites type Chateauponsac.* ④ *Granites à biotite (Vaulry) ou à 2 micas - Granites de Guéret.* ⑤ *Leucogranites en massifs différenciés: St. Sylvestre - St. Mathieu - Marche massifs du Confolentais; Chirac - massif de Piégut.* ⑥ *Granite d'Auriat, des Courrières,* ⑦ *Diorites quartziques syncinématiques de P2 et P3.*

Figure 21.11. (in Autran-Guillot (1974) completée)

Dans le Rouergue, le Centre et l'Est du Massif, on dispose d'un encadrement des évènements tectono-métamorphiques, comparable à celui du Limousin:

- métagranites orthogneissifiés datés par isochrone Rb/Sr de 520-500 Ma et de 420-460 Ma. (fig. 21.10).
- anatexie datée par 4 isochrones de granites d'anatexie homogénéisés vers: 380-400 Ma en 4 localités éloignées. (fig. 21.9).
- constance de la granitisation tardi orogénique de type Guéret (granitoides alumineux subautochtone différentiés depuis des granodiorites alumineuses jusqu'à des leucogranites potassiques) qui prend en écharpe tout le nord et le centre du Massif jusqu'au Rhône. Ces granitoides sont constamment disposés au dessus des domaines anatectiques qui apparaissent dans des dômes ou aires anticlinales. Ils sont datés, encore en peu de points, de la limite Dévonien-Viséen (360 Ma.) et transgressés par le Viséen moyen.

Au Sud du Rouergue et dans les Cévennes s'étend une épaisse série de micaschistes et schistes épimétamorphiques d'âge inconnu, dont les déformations ne sont pas encore clairement situées, d'une part par rapport aux trois à quatre phases de déformation, toutes carbonifères post Viséen qui affectent la zone de la Montagne-Noire au Sud et d'autre part aux phases dévoniennes exprimées dans le Rouergue au Nord.

C'est dans cette région que devait passer la limite Sud de la zone orogénique calédonienne d'âge Dévonien. Les levés gravimétriques individualisent très bien cette zone et suggèrent de placer une limite majeure crustale allant du Sud du Rouergue au Sud des Cévennes (figs. 21.1, 21.9).

MASSIFS CRISTALLINS EXTERNES DES ALPES ET MAURES

Les travaux de Carme (1970-1975) ont montré que la structuration majeure de l'ensemble Aiguilles Rouges-Belledonne-Haut Dauphiné avait une ressemblance complète dans son évolution, avec celle du Massif Central oriental et qu'elle était aussi antérieure au Viséen. Aucune datation stratigraphique ou isotopique indiscutable n'existe encore ici. On peut seulement montrer que la structuration majeure est antérieure à des dépôts dont l'âge est présumé Viséen par analogie de faciès avec le Viséen des Vosges et du Massif Central.

De toute façon dans la mesure ou l'âge Dévonien de la structuration majeure aura été confirmé dans l'Est du Massif Central, il en sera de même dans les Massifs cristallins externes. Ils constituent des jalons typiques de la "zone Moldanubienne" qui en Bohême et Bavière est actuellement réinterprétée comme une zone de matériel Protérozoique qui a acquis sa structuration tectono-métamorphique au Calédonien (Fuchs, 1976).

Dans les Maures (Provence) il semble bien aussi que les déformations et le métamorphisme Barrowien typiques soient d'âge Dévonien à Carbonifère, car le Silurien daté est pris dans les plissements et les orthogneiss ne peuvent avoir un âge > à 560 Ma. Les datations par Rb/Sr des granitoides tardi métamorphes ne sont pas encore concluantes. Seuls des granites post tectoniques sont bien datés du Carbonifère. (Voir article de P. Tempier, cet tome, no 22, sur la région Maures, Corse, Sardaigne pour plus de détails.)

PYRENEES

Dans les Pyrenees on peut affirmer qu'il n'y a pas d'orogenèse calédonienne, mais il y a des évènements épirogéniques nets entre le Caradocien et le Dévonien inférieur:

- lacune localement de l'Ordovicien Supérieur;
- volcanisme acide et basique sur des rides étroites émergées au Caradocien-Ashgillien.

La série stratigraphique de l'Ordovicien inférieur et du Cambrien est trop mal datée pour savoir s'il y existe des lacunes.

CONCLUSIONS GENERALES

Récapitulons les arguments fondamentaux qui prouvent l'existence d'évènements tectono-métamorphiques majeurs au Dévonien inférieur à moyen:

a) Série stratigraphique préorogénique = Dernier étage représenté: passage Silurien/Dévonien en Vendée et Limousin.

b) Premiers dépôts post orogéniques:
   - Givétien récifal en Vendée;
   - Frasnien (et peut être Givétien) puis Famènien au N.E. du Massif Central;
   - Viséen moyen dans tout le Massif Central Nord.

c) Arguments de géochronologie isotopique:
   - Derniers granites préorogéniques
     - 418 Ma métagranite d'Aubazines (Limousin) et de Tauves (Auvergne)
     - 430 Ma métagranite du Pinet (Rouergue)
     - 450 Ma métagranite du Moellan (Bretagne)
   - Anatexie tardi phase 2 datée par isochrone de Rb/Sr de néogranites anatectiques
     - 380 Ma en Forez-Velay
     - 362 Ma en Limousin
     - 390 Ma en Haut-Allier (Centre Massif Central)
     - 375 Ma en Vendée
     - 370 Ma dans le Morbihan
   - Refroidissement post-métamorphique daté à ~ 350 Ma ou entre 350 et 300 Ma par K-Ar sur amphiboles
   - Premiers granites supracrustaux, post phase 2
     - 360 à 340 Ma - granites du Guéret, Vaulry - Aureil dans le Massif Central.

A l'intérieur de cet encadrement entre Silurien et Dévonien Supérieur, les phénomènes géologiques observés sont surtout en rapport avec une évolution de zones profondes orogéniques. Partout ils se caractérisent par une phase précoce de métamorphisme à très forte pression suivie d'une diminution de la pression solide qui accompagne de grandes structures couchées pouvant renverser la disposition précoce des isogrades métamorphiques. L'état actuel des connaissances sur la structure ne permet pas encore de la décrire dans son ensemble.

Il apparait déjà une cohérence d'ensemble dans son évolution polyphasée antécarbonifère. Ainsi de grandes nappes-écailles de socle, de mise en place synmétamorphe, paraissent bien caractériser toute la moitié Nord du Massif Central (Grolier, 1971; Demay 1948; Burg et Matte, 1977). Leur hétérochronisme apparait probable (nappes syn P_1, nappes P_2') mais leur signification reste la même: c'est un resserrement en direction NS à NE-SW d'une large zone de croûte continentale étirée en distension au Paléozoique inférieur, avec peut être quelques étroites bandes franchement océanisées.

Il est très probable que le métamorphisme précoce de type HP se soit installé lors d'épisodes de subduction sialique qui ont pris naissance dans cette zone de croûte amincie. L'absence apparente de volcanites ou magmatites typiquement alcoalcalines, du Cambrien au Viséen, sauf dans les Vosges du Nord, suggère que ces amorces de subduction sont probablement restées de peu d'ampleur.

Les quelques remarques qui suivent sur la nature et l'âge du matériel sedimentaire et magmatique, déformé et métamorphisé au cours du Dévonien, constituent un début d'argumentation du schéma orogénique qui vient d'être évoqué. Ce sont des sujets à approfondir ou même à élucider dans l'avenir.

Le socle créé au Cadomien (orogenèse de la fin du Protérozoique supérieur) que l'on connait au Nord et au Sud de la zone orogénique dévonienne, a très probablement été l'objet d'une importante distension dès le Cambrien, séparant les domaines à Cambrien néritique du Nord et du Sud de la France. En effet dans la zone intermédiaire, lieu de cette distension crustale, une épaisse série de grauwackes et volcanites basiques et acides s'est déposée. Les premiers résultats de géochimie (éléments majeurs "inertes", traces, terres rares) suggèrent qu'il s'agirait bien d'un magmatisme tholeitique à caractères "abyssal" ou de zones de distension (Piboule, 1977; Guillot, Tegyey, et al., 1977; Leyreloup et al., 1977)

Ce groupe lithologique particulièrement significatif est généralement très métamorphique, aussi est-il appelé "groupe leptyno-amphibolite" et il fut attribué jusqu'à présent au Briovérien inférieur ou supérieur.

Son équivalent très probable en Europe varisque centrale est à rechercher dans les "gneiss de Gföhl" ou le "varieted group" du Moldanubien. Il semble bien qu'il soit en fait du même âge dans tout le Massif Central et les premiers résultats de géochronologie au Pb, sur les zircons des amphibolites et métagabbros, indiquent plutôt un âge de mise en place de ces roches au Cambrien et non au Protérozoique supérieur. L'analyse lithostratigraphique réalisée en Limousin et Cévennes conduit à ce même âge.

On constate, toujours dans la même région et parallèlement à la future zone orogénique, un magmatisme important à l'Ordovicien et au Silurien, avec des termes superficiels marins ou aériens et des termes plutoniques. On ne connait pas de déformations ni de métamorphismes contemporains. Il doit correspondre lui aussi en partie à un magmatisme de distension sialique (tholeites "abyssales" identifiées en Vendée et Limousin, granitoides Trondjehmitiques). Mais on ne connait pas encore assez ses caractéristiques géochimiques. Le sillon de Saint Georges sur Loire, relativement préservé des transformations métamorphiques, permet d'observer effectivement l'association de ce magmatisme avec une sédimentation de zone très subsidente.

Les roches basiques et ultrabasiques sont très peu abondantes et aucune n'évoque des ophiolites. Les plus gros massifs ont un volume de l'ordre du km^3. Leur composition chimique a été précisée dans quelques cas (Coffrant et Piboule, 1975; Piboule, 1977) et les apparente aux magmas tholeitique océaniques. Si leur âge par géochronologie aux zircons confirmait leur formation au Paléozoique, ils pourraient représenter des reliques métamorphisés de zones d'océanisation commençante, de même que le volcanisme basaltique mélé aux grauwackes qui était évoqué ci-dessus. Cette océanisation est probablement restée modérée en largeur, car aucune des roches sédimentogènes associées n'évoque des sédiments océaniques. Ce sont plutôt des dépôts de marges.

Enfin, l'absence ou la très faible proportion de croûte continentale d'un âge précambrien ancien, dans toute cette zone orogénique est probablement la cause des faibles rapports isotopiques initiaux du Sr des nombreux granitoides d'âge Cambrien à Carbonifère, même parmi eux, ceux qui, par leurs caractères hyperalumineux leucocrate, dérivent certainement d'une fusion anatectique crustale (Duthou, 1977; Vidal, 1976).

Un dernier point sera évoqué, c'est la relation entre la phase orogénique dévonienne très intense qui semble avoir été un orogène fondamental, dans le sens où il a créé une croûte sialique nouvelle aux dépends d'une zone de distension, et les phases orogéniques du Carbonifère, classiquement classées dans l'orogenèse hercynienne. Ces relations ne sont pas l'objet de cet article.

Constatons seulement qu'il semble y avoir eu dans la bordure Sud de la zone centrale armoricaine et dans certaines parties du Massif Central (Lyonnais), un enchaînement rapide des phases de déformation dévoniennes et carbonifères. C'est dans ces régions que la phase bretonne (entre le Dévonien supérieur et le Viséen inférieur ou moyen) succède ainsi presque immédiatement à la phase dévonienne. Comme nous le signalions pour le centre de la zone Sud armoricaine certaines parties profondes de la zone orogénique dévonienne sont probablement restées des zones métamorphiques actives, (tout en changeant de conditions de métamorphisme) entre le Dévonien et le Viséen. On ne sait pas encore délimiter de telles zones.

On voit donc qu'en France les évènements orogéniques du cycle calédonien sont d'âge tardifs dans ce cycle:

- non décelés au Cambrien (à moins que le magmatisme datés vers 520-500 Ma soit bien de cette époque et qu'on puisse lui associer des phénomènes tectoniques et métamorphiques);
- inexistants pendant l'Ordovicien inférieur, sauf probablement en Corse et en Sardaigne;
- très modérés en Ardennes entre l'Ordovicien supérieur et le Silurien supérieur, ailleurs ils ne sont marqués à la même période que par des lacunes ou de larges discordances cartographiques;
- majeurs et avec tous les attributs d'une orogenèse au Dévonien.

On est donc très tenté de les intégrer dans un mégacycle orogénique calédono-varisque, analogue au cycle appalachien.

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LES EVENEMENTS CALEDONIENS DANS LES MASSIFS VARISQUES DU SUD-EST DE LA FRANCE, CORSE ET SARDAIGNE

OROGÈNE CALÉDONIEN
IUGS UNESCO
Projet PICG n°27

Claude Tempier

Laboratoire Associé au CNRS "Etudes géologiques ouest-africaines"
et Laboratoire de Géologie dynamique,
Faculté des Sciences St. Jérôme, 13397 Marseille Cedex 4,
France.

INTRODUCTION

Les massifs varisques, formés de terrains paléozoïques sédimentaires et de formations crystallophylliennes, qui se situent sur le pourtour septentrional du bassin méditerranéen occidental, sont moins connus que les autres massifs français. Les études récentes (structurales et géochronologiques) y sont rares et ne concernent que des secteurs limités. Cette note, fondée sur les travaux cités en référence, ainsi que sur les études en cours dans le massif des Maures et en Sardaigne, est destinée essentiellement à poser les les problèmes et à dégager les indices d'événements tectono-métamorphiques pouvant être rapportés au cycle calédonien.

A) MASSIFS DES MAURES ET DU TANNERON (FIGURE 22.1)

Les travaux récents portant sur ces massifs, situés entre Toulon et Cannes, ne concernent que des secteurs limités et ne permettent pas d'établir une synthèse; classiquement (Gueirard, 1957) ils sont formés par une succession continue de terrains métamorphiques, épizonaux à l'Ouest, catazonaux á l'Est, que l'on peut découper en quatre ensembles.

1°) Une ensemble de puissantes formations métasédimentaires, surtout grésoschisteuses, forme la partie occidentale des Maures, ainsi que la presqu'île de Giens, les îles de Porquerolles et Port Cros, le Cap Sicié et les petits lambeaux des environs de Toulon. A la partie supérieure (chapelle du Fenouillet au Nord de Hyères) un minuscule affleurement de shales siliceux noirs a été daté du Silurien inférieur (Schoeller, 1938; Gueirard *et al.*, 1970). Dans ce secteur, de même qu'au cap Sicié, à l'Ouest de Toulon, des formations lithologiquement différenciées (quartzites, schistes noirs, grès feldspathiques, rares lentilles de calcaires à crinoïdes) représentent probablement l'Ordovician supérieur. On notera l'absence de formations carbonatées pouvant être rapprochées des niveaux calcaires cambriens. On ne connaît pas non plus dans les Maures de termes attribuables au Dévonien qui est le plus souvent calcaire dans le pourtour méditerranéen occidental.

2°) Au-dessous de ce premier groupe dont la base est constituée par des amphibolites à lits de calcite (Collobrières) se situent des formations gneissiques (gneiss de Bormes). Des études en cours, quoique locales, ont montré la complexité structurale de ces gneiss et souligné l'importance des faciès "ortho" souvent blastomylonitiques dérivant d'anciens granites. Des indices de reliques catazonales ont été observés, dont l'affleurement de Barral au Nord de Bromes (Gueirard, 1976). Les études structurales et géochronologiques de H. Maluski (1968, 1971, 1972) révèlent l'existence de granites antérieurs au métamorphisme varisque (300 - 270 Ma) et mis en place à une date encore mal précisée, située entre 600 et 400 Ma.

3°) A l'Est des gneiss de Bormes se développe un ensemble comprenant des micaschistes à grenat, staurotide et plus rarement disthène, des gneiss fins (origine "para" probable) et des amphibolites avec localement des reliques catazonales telles que des boudins d'éclogites et des gabbros coronitiques (Lasnier, 1970), plus ou moins rétromorphosées. Les faciès blastomylonitiques sont probablement très fréquents. Les conglomérats et shales du Plan-de-la-Tour, d'âge stéphanien, non métamorphiques mais plissés suivant une direction méridienne, reposent en discordance majeure sur ces formations qui sont recoupées à l'Est par une zone mylonitique (faille de Grimaud), postérieure aux plissements et aux métamorphismes.

4°) Au-delà de ce grand accident tardif se développe un ensemble gneissique étudié dans les collines de Ste-Maxime (Le Marrec, 1976) et dans le Tanneron occidental (Orsini, 1968). Cet ensemble, où il serait prématuré de tenter des corrélations parmi les formations qui y ont été définies, comportent des orthogneiss conservant la trace d'une histoire complexe (Les Issambres), des métavulcanites acides (Les Adrets au Nord de Fréjus) et des lentilles d'éclogites localement à disthène, plus ou moins complètement rétromorphosées en amphibolites à grenats (La Nartelle, Le Cavalières, Forêt Royale). Les faciès blastomylonitiques sont fréquents dans ces formations qui sont, de plus, recoupées par des zones mylonitiques tardives orientées Nord-Sud, analogues à celle de Grimaud. Les conglomérats stéphaniens non métamorphiques du Reyran reposent en discordance sur les gneiss.

La migmatisation, qui s'est effectuée en deux épisodes, devient localement importante, surtout dans la presqu'île de St-Tropez. Il existe aussi quelques massifs intrusifs: diorite orientées syntectonique de Prignonet (Tanneron occidental), granite post-tectonique du Plan-De-La-Tour (Maures) et du Rouet (Tanneron).

Les mesures géochronologiques effectuées par plusieurs auteurs (Maluski, 1972; Roubault *et al.*,* 1970) par la méthode Rb/Sr permettent d'attribuer un âge varisque (320-340 Ma) au granite du Plan-De-La-Tour. L'âge des gneiss du Tanneron serait plus ancien: une isochrone sur roches totales de plusieurs faciès donne 410 ± 25 Ma* (Dévonien inférieur - Silurien), âge interprété par Roubault *et al.* (1970) comme étant celui du métamorphisme.

Dans les massifs des Maures et du Tanneron, les études structurales en cours ou publiées (Arthaud et Matte, 1966; Maluski, 1968; Le Marrec, 1976) ont mis

* âges corrigés avec la constante $\lambda^{87}_{Rb} = 1,39\ 10^{-11} \cdot an^{-1}$

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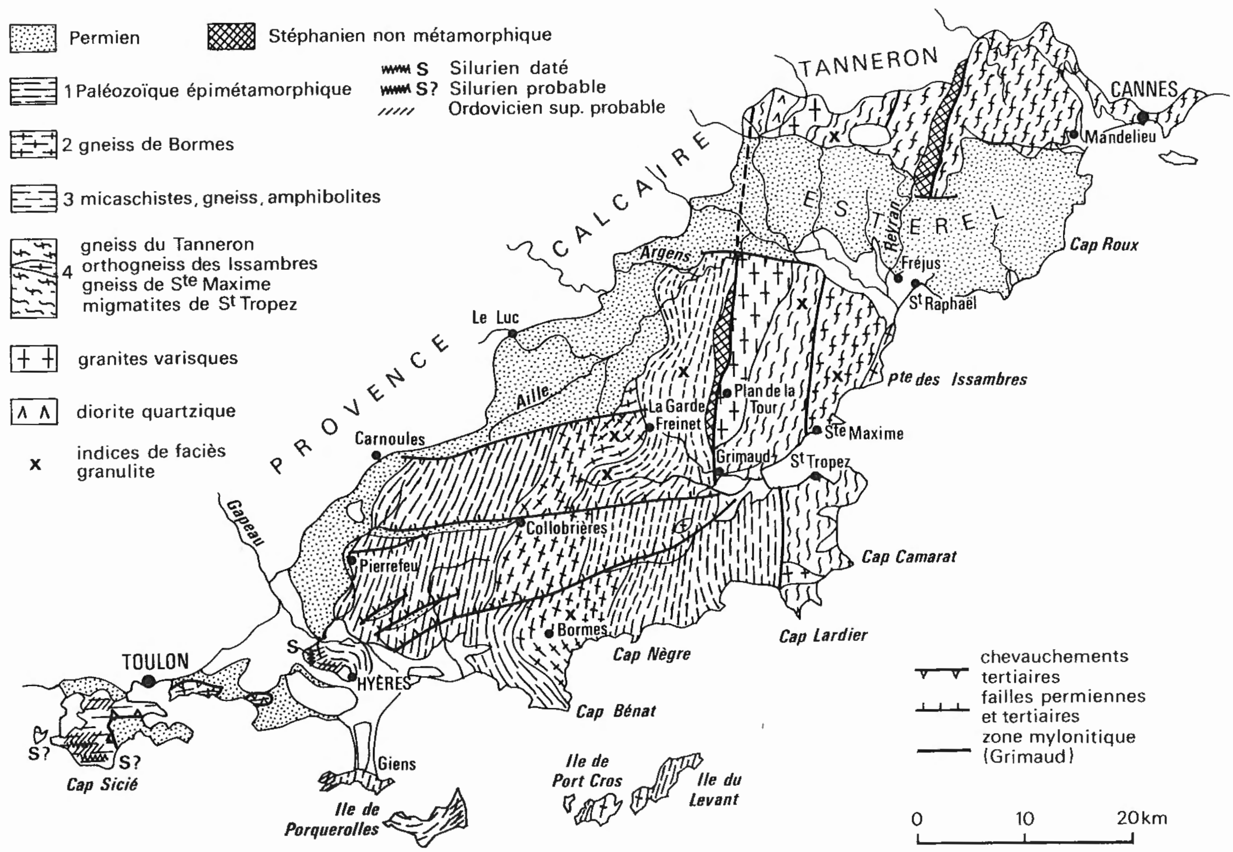

Figure 22.1. Schéma géologique des massifs des Maures et du Tanneron.

en évidence quatre phases de déformation dont les deux premières, isoclinales, sont liées à la schistosité régionale et au métamorphisme. Ce dernier, plurifacial ou peut-être polyphasé, semble débuter par un épisode à haute pression (qui demanderait de nouvelles investigations) suivi de la phase principale de type intermédiaire liée à la schistosité regionale. Il se termine localement par un épisode de basse pression et haute température qui pourrait être lié, à l'Est, à la migmatisation (Maquil *et al.*, 1976). Les reliques catazonales n'existent pas dans le groupe 1 (métasédiments paléozoïques). L'abondance des faciès blastomylonitiques indique que la tectonique des premières phases est probablement de style tangentiel.

En conclusion, dans l'ensemble Maures - Tanneron, déformations et métamorphisme majeurs sont post-siluriens et anté-stéphaniens. Les quelques données géochronologiques disponibles suggèrent qu'après des phénomènes magmatiques cambro-ordoviciens (mise en place de granites), la phase principale de métamorphisme, de type intermédiaire, localement suivie d'épisodes à basse pression, serait d'âge dévonien inférieur. : ceci s'appuie sur la présence de gneiss offrant des âges de 410-380 Ma. Or dans le Bas Limousin (Massif Central français) où l'on observe des faciès comparables à ceux des Maures, ainsi que deux phases de déformations synschisteuses suivies de plis post-schisteux et une évolution pétrogénétique identique, tectonique et métamorphisme majeurs sont d'âge calédonien tardif : phase acadienne (Autran et Guillot, 1975; Bernard-Griffiths *et al.*, 1977). Cependant des études de détail accompagnées de mesures d'âges radiométriques sont absolument nécessaires pour vérifier cette hypothèse.

B) MASSIF DE L'ARGENTERA - MERCANTOUR (FIGURE 22.2)

L'Argentera - Mercantour, le plus méridional des massifs cristallins externes des Alpes occidentales, est formé de gneiss catazonaux le plus souvent migmatisés (Faure-Muret, 1955) et probablement affectés par deux orogenèses dont l'une serait anté-varisque. Les faciès rappellent ceux des Maures orientales (groupe 4). Bogdanof et Prunac (1976) ont décrit quatre phases de déformation mais aucune mesure géochronologique ne semble avoir été publiée.

C) CORSE (FIGURE 22.2)

Les formations anté-stéphaniennes ne constituent en Corse que de petits lambeaux isolés dans les plutons granitiques varisques. Il s'agit surtout de gneiss (Belgodère) à lentilles basiques rappelant certains faciès de l'Argentera ou du Tanneron; ou de faciès particuliers (roches brunes) mylonitiques ? transformés par métamorphisme de contact. Toutefois, à l'Argentella (Sud de Calvi, Corse du Nord-Ouest), il existe des terrains paléozoïques, à peine schistosés et non métamorphiques: calcaires et grauwackes ("Culm") dévono-dinantiens de la Tour Maraghiu et shales siluriens du Monte-Martinu, décrits et datés par Baudelot *et al.* (1976). Les shales reposent sur des conglomérats qui sont légèrement discordants sur des siltstones argileux attribués par ces auteurs à l'Ordovicien. Ces siltstones reposent en discordance majeure sur une formation quartzo-feldspathique métamorphique ayant subi plusieurs phases de déformations intenses. Cet affleurement, malheureusement très peu étendu, permet cependant de pressentir l'existence d'une importante phase tectono-métamorphique anté-silurienne ou plutôt anté-Ordovicien supérieur, donc calédonienne ancienne à précambrienne, qui pourrait être à l'origine des gneiss du socle corse.

D) SARDAIGNE (FIGURE 22.2)

Les terrains paléozoïques affleurent largement en Sardaigne dans quatre secteurs principaux séparés par des granites varisques ou par des fossés à remplissage sédimentaire ou volcanique cénozoïque.

1°) Au Sud-Ouest, dans l'Iglesiente et le Sulcis, les formations paléozoïques plissées sont peu ou pas métamorphiques. Des conglomérats à galets de calcaires cambriens, des grès, schistes et pélites rouges souvent puissants, supportant un niveau fossilifère, sont attribués à l'Ordovicien supérieur. Ils reposent en discordance angulaire pouvant atteindre 80° sur un Cambrien puissant et bien caractérisé. Les plissements anciens qui ont affecté ce Cambrien, classiquement rapportés à la phase sarde (limite Cambrien Ordovicien) seraient en fait seulement connus comme anté-ordovicien supérieur. Le Silurien est schisteux et calcaire; il est surmonté (contact complexe) par des grès azoïques attribués au Dévonien.

2°) Au Sud-Est (Gerrei, Quirra), les formations paléozoïques épimétamorphiques, affectées par au moins deux phases de plissement dont la première d'axe E-W isoclinale, est liée à la schistosité régionale, se développent largement entre les massifs granitiques varisques du Sarrabus au Sud et de la Barbagia au Nord. Des travaux en cours dans ces terrains assez monotones (Naud et Tempier, 1977) mettent en évidence un groupe de base complexe sur lequel repose en discordance (Majeure?) l'Ordovicien supérieur débutant par des arkoses grossières et comportant associés à un horizon schisto-calcaire fossilifère, des schistes et siltstones, des grès et des formations volcaniques. Le Silurien, qui débute par des bancs très siliceux noirs, est formé par des schistes et calcaires à orthocères. L'existence du Dévonien calcaire est très probable.

Le groupe de base, non daté, est formé de schistes et quartzites verdâtres, marbres, conglomérats et "porphyroïdes". Ces derniers comportent des termes intrusifs à grands cristaux de feldspath pouvant correspondre à d'anciens granites ultérieurement déformés. L'importance et les caractères de cette tectonique anté-ordovicien supérieur, comparable du point de vue âge avec celle décrite dans l'Iglesiente, sont encore mal connues. Il est toutefois très probable que les terrains situés sous la discordance sont affectés par une phase de déformation plus ancienne, intense, absente dans les arkoses de l'Ordovicien supérieure et les formations sus-jacentes.

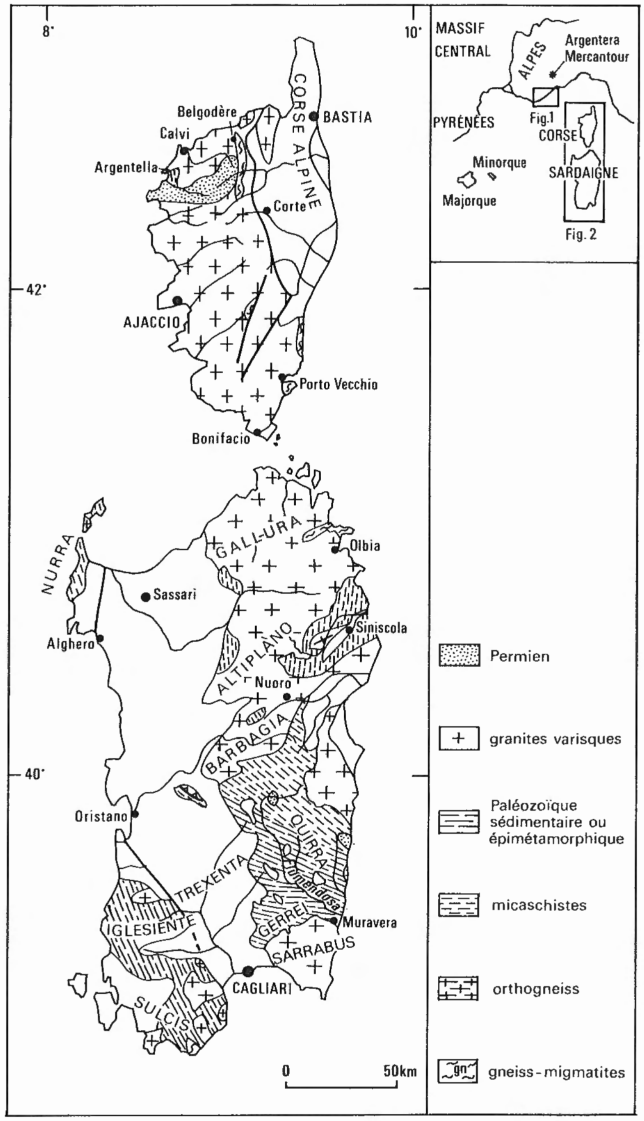

Figure 22.2. Schéma géologique de la Corse et de la Sardaigne.

3°) Au Nord-Est (Altiplano, Gallura) se développent des terrains non datés, affectés par un métamorphisme mésozonal de type intermédiaire, à staurotide et disthène, avec apparition de migmatites vers le Nord. Dans cet ensemble essentiellement schisteux existent localement (Ouest de Siniscola) des orthogneiss granodioritiques sur lesquels une isochrone Rb/Sr a donné un âge* de 470 ± 30 Ma avec rapport initial 0,709 (Di Simplicio *et al.*, 1975). On peut donc penser qu'un massif granodioritique mis en place à l'Ordovicien inférieur aurait ensuite été transformé en orthogneiss au cours du métamorphisme. Ce granite pourrait être rapproché de ceux qui, au Sud, dans le Gerrei, sont à l'origine des "porphyroïdes" à grands cristaux de feldspath. Les âges K/Ar sur des amphiboles de la série métamorphique (300 - 290 Ma), interprétés comme l'âge du métamorphisme par Di Simplicio *et al.* (1975) et

* âges corrigés avec la constante $\lambda^{87}_{Rb} = 1{,}39\ 10^{-11} \cdot an^{-1}$

légèrement plus récents que celui trouvé pour les batholites varisques, peuvent correspondre au refroidissement de ces derniers.

4°) Au Nord-Ouest, dans la Nurra, existent des terrains métamorphiques (schistes à chloritoïdes, quartzites, micaschistes) présentant des analogies avec les formations des Maures occidentales (groupe 1).

Les études structurales portant sur les terrains paléozoïques sardes sont rares. Elles ne concernent que des secteurs limités comme l'Iglesiente (Arthaud, 1970). Antérieurement aux phases principales, de déformation et de metamorphisme, l'existence d'une tectonique anté-ordovicien supérieur (sarde?, taconique précoce?) est sûre dans le Sud de l'île. Elle est accompagnée d'un magmatisme acide intrusif et effusif, ce dernier se poursuivant au cours de l'Ordovicien supérieur. Dans le Nord-Est, en Gallura, l'intrusion de granodiorite (ayant donné les orthogneiss) peut correspondre à cet épisode mais on ne connaît pas pour le moment, de discordance. Les déformations majeures, post-siluriennes et anté-stéphaniennes, comme dans les Maures, sont polyphasées (Matte et Arthaud, 1975): dans le Nord de l'île, la première phase, d'axes orientés Nord-Sud, serait associée au métamorphisme de type intermédiaire à staurotide - disthène avec apparition, vers le Nord, de sillimanite et migmatisation; une deuxième phase affecterait surtout le Sud, peu déformé précédemment où la présence du Dévonien plissé est probable; elle serait associée à un métamorphisme épizonal. Les âges K/Ar varisques (Di Simplicio *et al.*, 1975) ne datent que la fin du deuxième métamorphisme ou même le refroidissement des grands plutons granitiques.

CONCLUSION

Dans les massifs considérés (Maures, Tanneron, Argentera - Mercantour, Corse occidentale, socle sarde) généralement considérés comme formés lors de l'orogenèse varisque, l'existence d'événements calédoniens est probable et même localement démontrée.

En Sardaigne méridionale, une phase calédonienne précoce, antérieure à l'Ordovicien supérieur, est certaine. Dans l'Iglesiente, au Sud-Ouest, elle est caractérisée par des plis de niveau structural supérieur et au Sud-Est dans le Gerrei par des déformations plus intenses. Elle s'accompagne de magmatisme acide, intrusif et effusif.

Ailleurs, il y a partout des indices de phases anté-varisques et d'événements attribuables au cycle calédonien, mais dans l'état actuel des connaissances, on ne peut formuler que des hypothèses de travail qui devront être étayées par les études structurales en cours et par de nombreuses mesures d'âges radiométriques.

La phase antérieure à l'Ordovicien supérieur affecte aussi probablement les régions plus métamorphiques du Nord de la Sardaigne telle la Gallura où existent également d'anciennes roches intrusives acides mises en place à l'Ordovicien inférieur.

Une phase orogénique anté-silurienne (ordovicienne ou plus ancienne) pourait être à l'origine des gneiss du socle de Corse et peut-être aussi du massif de l'Argentera-Mercantour; cet ensemble devant alors être rapproché de la zone moldanubienne. Mais cette hypothèse, fondée sur l'âge et la situation structurale du seul affleurement paléozoïque de l'Argentella (dont il faut souligner les analogies avec les schistes de Steige dans les Vosges) a encore besoin d'être vérifiée.

Dans le massif des Maures, le Tanneron et peut-être dans le Nord de la Sardaigne, il semble que ce soit une phase acadienne (intra-dévonienne) qui domine avec une histoire structurale et pétrogénétique offrant de grandes analogies avec celle du Limousin dans le Massif Central français. Cette analogie d'évolution suggère que la zone orogénique Sud Bretagne - Limousin, structurée au cours du Dévonien (phase acadienne), pourrait se prolonger vers le Sud-Est dans les Maures et éventuellement le Nord de la Sardaigne. Par contre, dans le Sud de cette île, ce sont les phases varisques, affectant le Silurien et le Dévonien, qui semblent majeures bien que reprenant du matériel structuré avant l'Ordovicien supérieur.

Il apparaît ainsi que presque partout dans la région considérée, les phénomènes varisques sont précédés par des phases plus anciennes pouvant être très intenses. Ils affectent de ce fait du matériel pouvant être déjà structuré et métamorphisé et aboutissent alors à la formation de plis post-schisteux, de nouvelles migmatites, de zones mylonitiques ainsi qu'à la mise en place de grands plutons granitiques post-tectoniques.

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OROGÈNE CALÉDONIEN
Projet PICG nº 27

23. L'EVOLUTION CALEDONO-VARISQUE DES HESPERIDES

A.L. Guy Tamain

Laboratoire de Géologie Structurale et Appliquée
504 Faculté des Sciences
91405 Orsay Cedex, France

I. INTRODUCTION

Les Hespérides (= Massif hespérique, = Meseta ibérique) constituent la partie méridionale de l'arc ibéro-armoricain, à l'extrémité occidentale des chaînes cadomienne et varisque d'Europe dont elles sont, avec leurs 200.000 km^2 de superficie et au niveau d'érosion actuel, l'un des éléments les plus vastes et les plus continus.

Elles ont vue en effet, après une longue période de préparation protovarisque (= "calédonienne"), le segment varisque se mouler, en le reprenant parfois très énergiquement et en se surimposant étroitement à lui par endroits, sur le segment cadomien.

Leur histoire s'étend pratiquemment sur un milliard d'années, au moins.

Elles correspondent aujourd'hui à la moitié occidentale de l'Espagne et à la plus grande partie du Portugal (fig. 23.1).

Toutefois, l'état de notre connaissance des Hespérides varie encore très sensiblement d'une zone à l'autre, certaines régions - généralement, les mieux minéralisées - ayant attiré très tôt le prospecteur et le géologue, d'autres étant restées comme à l'écart.

II. GEOLOGIE REGIONALE

C'est à partir de l'architecture varisque que l'on subdivise les Hespérides en six zones paléogéographiques et litho-structurales (fig. 23.2), pouvant être d'ailleurs regroupées en trois grands domaines séparés par deux linéaments mégatectoniques.

1. LE DOMAINE SEPTENTRIONAL

Limité au Sud par le grand accident du Tage, il comprend, du NE au SW, les trois zones: cantabrique, léono - ouest-asturienne, et galaïco-castillane (Lotze, 1945).

1.1. La Zone cantabrique

Elle est située dans la concavité de l'arc ibéro-armoricain et présente les caractères suivants (Julivert *et al.*, 1974).

A. Stratigraphie

-Paléozoïque inférieur, très inégalement développé et fréquemment incomplet:

-Cambrien et Ordovicien *p.p.* (à volcanites: Cabo Peñas), Ordovicien moyen et supérieur souvent absents;

-Silurien manquant localement (Ponga);

-Dévonien bien exprimé à l'Ouest:

-de faciès asturo-léonais (sédimentation principalement calcaire, dans mer chaude peu profonde, à faune benthonique) au SW et W;

-de faciès palencien (en mer plus profonde, avec faune pélagique) au SE.

-Paléozoïque supérieur:

-Dévonien supérieur de faciès asturo-léonais et palencien (à Goniatites);

-transgression famennienne généralisée;

-notamment au Sud du bloc cantabrique: Tournaisien et Viséen (à lydiennes) extrêmement condensés, Namurien épais (à flysch calcaire);

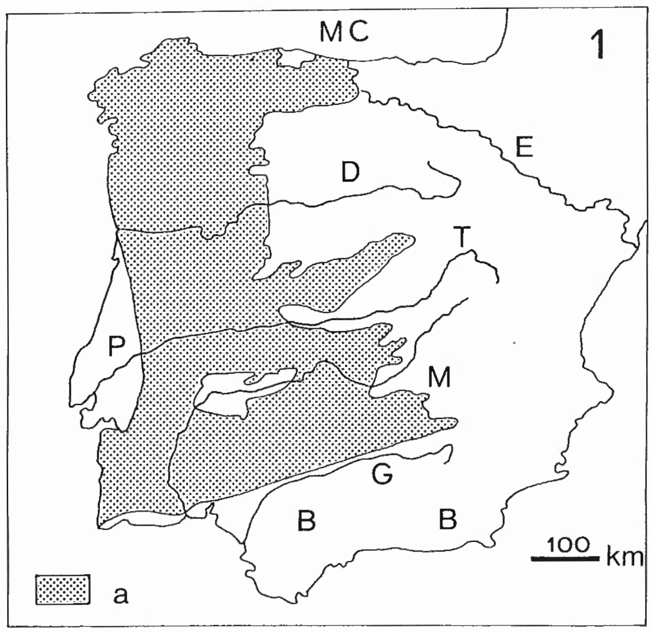
-Bassin Houiller Central: Westphalien très bien exprimé (B, C et D inférieur-moyen, paralique, molassique, puissant de 5000 m);

-NE du Léon et N de Palencia: Westphalien D - Cantabrien - Stéphanien A, marin, molassique, discordant et scellant la Nappe d'Esla déjà replissée;

-au SE: Westphalien B, molassique et discordant;

-Stéphanien B-C continental, discordant, et relativement étendu;

-Autunien, localisé - gréseux avec tufs volcaniques, ou bien à schistes bitumineux.



a : Protérozoique et Paléozoique. B : domaine alpin du Bétique. M : Manche. MC : Mer Cantabrique. P : Fossé du Portugal. D : Duero. E : Ebre. G : Guadalquivir. T : Tage.

Figure 23.1. LES HESPERIDES

PICG Projet 27, Contribution française n^0 13.
Caledonian-Appalachian Orogen of the North Atlantic Region; Geol. Surv. Canada, Paper 78-10 (1978).

B. Tectogenèse

-Tectogenèse varisque polyphasée (de Sitter, 1965; Julivert, 1971a et b; Julivert et Truyols, 1973):

-première phase, dite palencienne (= de Curavacas, connue dans le SE à la limite du Westphalien A et B ou bien intra - Westphalien B), donnant un style tectonique superficiel d'écailles et de nappes de charriage (d'amplitude hectométrique à décakilométrique) replissées par la suite, avec formation d'un premier système arqué de plis (à plan axial vertical);

-phase léonienne (intra - Westphalien D) génératrice d'un deuxième système rayonnant de plis (à plan axial également vertical) croisant le précédent;

-phase asturienne (entre le Stéphanien A et B), voyant le serrage des structures plissées antérieures;

-phase saalienne.

C. Métamorphisme

-Métamorphisme régional varisque: absent ou, exceptionnellement, présent (et très discret).

D. Plutonisme

-Plutonisme tardi-varisque, postérieur à la grande tectonique, très peu développé:

-en stocks granitiques calco-alcalins;

-ultimes porphyres acides, fini- à post-stéphaniens.

1.2. La Zone léono - ouest-asturienne

Elle comprend la zone initialement définie par Lotze (1945) ainsi que le Dôme de Lugo (en Galice orientale) et l'Anticlinal de l' "ollo de sapo" à sa limite occidentale.

Elle apparaît ainsi caractérisée (Matte, 1968; Capdevila, 1969; Marcos, 1971 et 1973; Julivert *et al.*, 1974):

A. Stratigraphie

-Protéroïque supérieur *p.p.*: "formation porphyroïde de l' "ollo de sapo", "série de Villalba" et "série du Narcea".

-Paléozoïque inférieur: à séries cambriennes et ordoviciennes en continuité, détritiques et très puissantes; Silurien schisteux à Graptolites et volcanites; Wenlockien discordant, en Galice orientale; Dévonien inférieur réduit.

-Paléozoïque supérieur: Carbonifère pré-stéphanien très peu développé, et Stéphanien B-C discordant (dans l'Ouest des Asturies).

B. Tectogenèse

-Tectogenèse cadomienne restant à définir;

-Tectogenèse varisque polyphasée:

-première phase, majeure, génératrice de grands plis déversés à couchés vers l'Est (Mondoñedo), et accompagnée de métamorphisme;

-deuxième phase, dans l'ensemble post-métamorphique, cisaillant et faisant chevaucher vers l'Est les structures plissées précédentes (Marcos, 1971);

-troisième phase responsable de plis à plan axial subvertical (reprenant les structures antérieurs, avec effet de rétrocharriage), post-métamorphique, anté - Stéphanien B-C et probablement méso-westphalienne (vers la limite du Westphalien B et C);

-phases ultérieures, mineures.

C. Métamorphisme

-Métamorphisme régional varisque de basse pression, croissant d'Est (faciès épizonal) en Ouest (faciès méso- et catazonaux) (Capdevilla, 1967), syn- à post-phase 1 et vraisemblablement antérieur à 312 Ma (*cf. infra*).

D. Plutonisme

-Plutonisme varisque polyphasé, croissant d'Est en Ouest (Capdevila et Vialette, 1965 et 1970; Capdevila et Floor, 1970):

-vieux granites et granodiorites calco-alcalins à biotite, post-phase 1 et anté-phase 3, tardi- à post-métamorphique - Santa Eulalia de Pena (318 ± 22 Ma) et Puebla de Parga (312 ± 10 Ma);

-leucogranites alcalins à 2 micas, syn-phase 3 et post-métamorphiques - Friol (304 ± 10 Ma);

-granites alcalins à 2 micas, post-phase 3, en intrusions discordantes - Forgoselo (303 ± 6 Ma);

-ultimes granodiorites, en massifs circonscrits - Castroverde (276 ± 10 Ma).

1.3. La Zone galaïco - castillane

Elle a jadis été interprétée (Staub, 1926) comme constituant le coeur "archéen" des Hespérides, autour duquel se seraient successivement moulées les ceintures "calédonienne" puis varisque.

Ses caractères principaux sont les suivants (Parga Pondal, 1960; Oen Ing Soen, 1970; den Tex et Floor, 1971):

A. Stratigraphie

-Protérozoïque moyen:

-la datation radiométrique d'une éclogite (900 ± 30 Ma), associée à certain faciès granulite de la Galice moyenne (Vogel, 1970, in den Tex et Floor, 1971) témoigne en faveur de l'existence d'un authentique matériel métamorphique pentévrien (fig. 23.3 B) comparable à certaines roches amphiboliques de même âge du Nord du Massif armoricain (Leutwein, 1968) ou bien, éventuellement rajeuni, corrélable avec certaines métabasites du Moldanubien bohémien (Svoboda, *et al.*, 1966; Zoubek, 1972) (voir aussi: Cogné, 1959 et 1974; Roach *et al.*, 1972; Leutwein *et al.*, 1973; pour le Pentévrien armoricain).

-Protérozoïque supérieur: avec notamment,

-un groupe comprenant trois complexes, l'un magmatique précoce de type ophiolitique (hyperbasites à Cr-Pt et basites de Vinhais - Bragança, Morais, Lalin, Santiago de Compostela, Cabo Ortegal), le second à amas sulfurés (Santiago de Compostela, Cabo Ortegal), le dernier à niveaux siliceux et phtanites (avec occurrences de Mn-Fe: Sobralhal) (fig. 23.3 C, D, E);

-un groupe sus-jacent, schisto-grauwackeux (fig. 23.3 F).

-Paléozoïque inférieur: Ordovicien inférieur transgressif et discordant, Silurien (à Graptolites), et Dévonien inférieur au SW;

-Paléozoïque supérieur très localisé: Westphalien D (basin de Porto), et Stéphanien supérieur.

B. Tectogenèse

-Tectogenèse pentévrienne:

-restant à définir dans le NW;

-donnée (Llopis Llado et Sanchez de la Torre, 1962, 1963 et 1965) comme alpinotype (à plis couchés NE-SW, donnant un style de méso-nappes) et dont les traces seraient conservées dans les migmatites de Tolède.

-Tectogenèse cadomienne polyphasée: à plis subméridiens en Galice occidentale, et à structure E-W en Galice moyenne.

-Tectogenèse varisque polyphasée:

-première phase, majeure, responsable des virgations de Galice et de Gredos - Guardarrama, avec formation de plis couchés de grande amplitude remobilisant le socle cristallin précambrien (en Galice moyenne), et d'anticlinaux et synclinaux étroits (à plan axial subvertical) à Ordovicien et Silurien pincés (NW du Portugal);

-phases fini- à post-stéphaniennes, mineures.

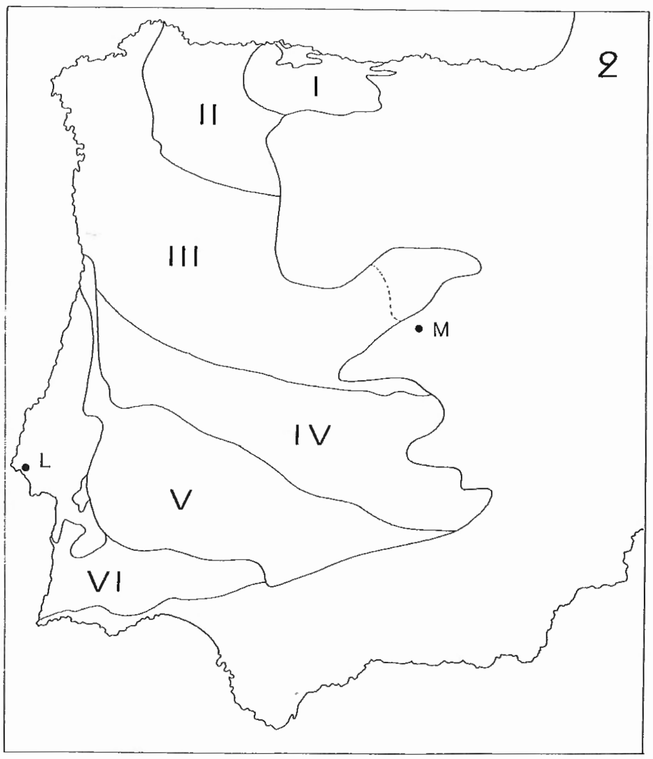
C. Métamorphisme

-Métamorphisme régional pentévrien: restant à définir;

-Métamorphisme cadomien: plurifacial (épi-, méso- et catazonal), variable selon la région considérée: de haute pression du type barrovien atteignant le faciès granulite (p.ex. en Galice moyenne) (den Tex et Floor, 1971);

-blastomylonitisation importante (den Tex et Floor, 1971);

-Métamorphisme varisque: de basse pression du type Abukuma ou intermédiaire de basse pression, voire du type barrovien dans les niveaux profonds, et accompagnant la phase plicative majeure.



I : Zone cantabrique. II : Z. léono - ouest-asturienne. III : Z. galaico-castillane. IV : Z. luso-orétane. V : Ossa Morena. VI : Z. sud-portugaise.

Figure 23.2. LES GRANDES ZONES LITHOSTRUCTURALES VARISQUES DES HESPERIDES

D. Magmatisme et Plutonisme

-Magmatisme initial ophiolitique du Protérozoïque supérieur;

-Plutonisme et migmatisation cadomiens encore mal connus;

-granites gneissiques et très tectonisés de Galice, dont l'âge radiométrique (349 ± 10 Ma) (Priem *et al.*, 1970) en fait une manifestation reussienne tardive à posthume;

-Plutonisme varisque plurifacial et polyphasé (Capdevila et Vialette, 1965, 1970; Priem *et al.*, 1967 et 1970; Mendès, 1968; Floor *et al.*, 1970; Oen Ing Soen, 1970; Capdevila *et al.*, 1973):

-granites palingénétiques, syn- à tardi-tectoniques, syn-métamorphiques, concordants, alcalins à 2 micas, datés de 298-297 ± 12 Ma, mis en place dans des domaines métamorphiques de basse pression du type Abukuma, très réduits dans les Sierras Centrales, et bien développés dans le NW hespérique;

-granodiorites et granites tardi- à post-tectoniques, discordants, calco-alcalins à biotite, très bien exprimés dans les Sierras Centrales (où ils sont post-tectoniques), ils comprennent dans le NW: une première série, syn-métamorphique et tardi-tectonique; une deuxième série, post-métamorphique et post-tectonique - Castro Daire (282 ± 7 Ma).

2. LE DOMAINE MERIDIONAL

Il comprend, du NNE au SSW, la Zone luso-orétane (=luso-alcudienne) et l'Ossa Morena que sépare le Linéament de Cordoue.

2.1. La Zone luso-oretane

Elle s'étend - depuis la vallée moyenne du Tage jusqu'au batholite de Los Pedroches - en une suite d'étroits domaines litho-structuraux, allongés WNW-ESE à NW-SE, particulièrement bien définis dans le SE hespérique (Ovtracht et Tamain, 1970a; Tamain et Ovtracht, 1971; Tamain, 1972 et 1975):

A. Stratigraphie

-Alcudien (=Protéroïque supérieur, = Précambrien le plus récent) (Tamain, 1970, 1971, 1972, 1973 et 1977; Ovtracht et Tamain, 1970a et b; Crespo Lara et Tamain, 1971; Tamain et Ovtracht, 1971a et b) moyen p.p. et supérieur (à reliques microscopiques d'un vieux socle métamorphique anté-alcudien (fig. 23.3 E, F).

-Paléozoïque inférieur: Cambrien inférieur (au NNE: dans le bras de mer tolédan), Ordovicien transgressif complet (2200 m dans les Hautes Sierras méridionales) (Tamain, 1967, 1971 et 1972; Tamain *et al.*, 1969; Tamain et Ovtracht, 1971a) et en général directement discordant sur le Protérozoïque supérieur (Mallada, 1896; Bouyx, 1969 et 1970), Silurien transgressif et discordant (Tamain, 1964; Arbey et Tamain, 1971), et Dévonien inférieur.

-Paléozoïque supérieur:

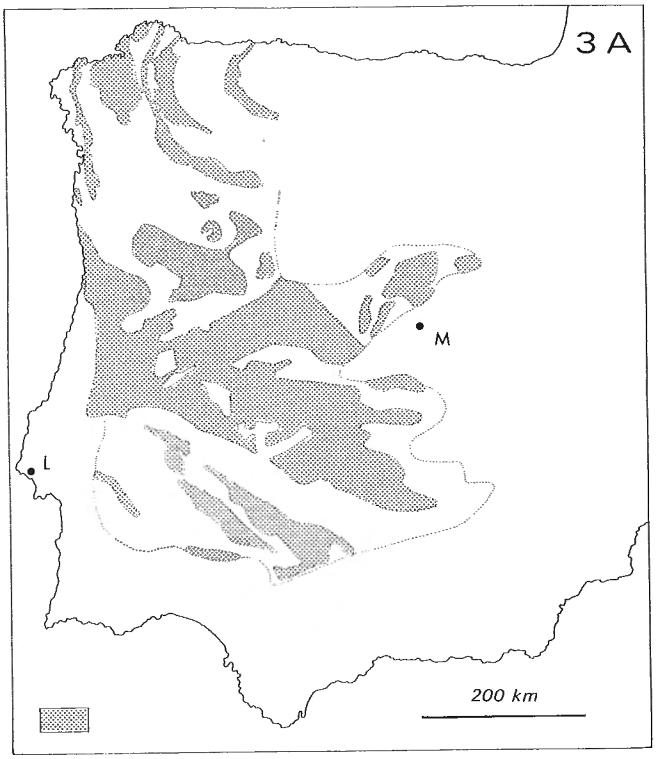
-Frasnien, Famennien (à Strunien) et Dinantien très puissant (à flysch de faciès Culm, au Viséen *cf.* supérieur) au SE;

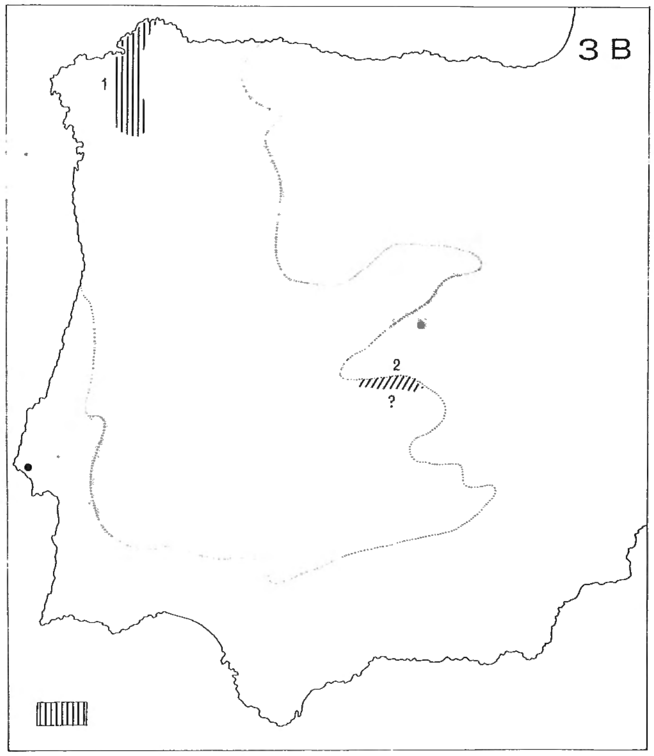
Stéphanien B-C et Autunien, avec volcanisme acide aérien (bassin de Puertollano) (Wagner et Utting, 1967).

B. Tectogenèse

-Tectogenèse cadomienne:

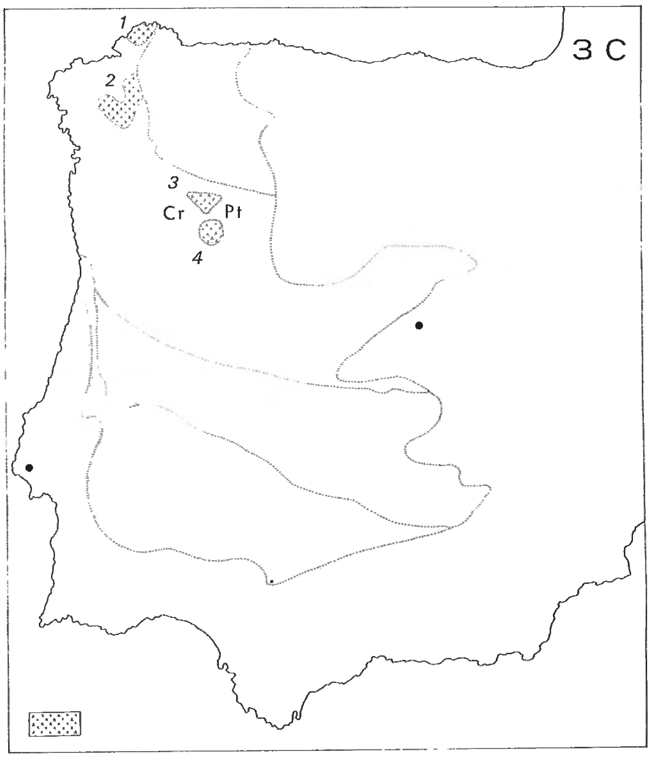
-phase majeure génératrice de plis isoclinaux NNW-SSE (à plan axial subvertical);





A. -- LES AFFLEUREMENTS.
L : Lisbonne ; -- M : Madrid.

B. -- LES AIRES A PROTEROZOÏQUE MOYEN (VRAI, EVENTUEL, OU HYPOTHETIQUE).
1 : Galice moyenne ; -- 2 : Tolède.



C. - G. -- LA MOBILISATION CADOMIENNE.

1. Etape préliminaire à précoce.

C. -- Magmatisme initial ophiolitique (à chrome-platine) de l'Alcudien moyen.
1 : Cabo Ortegal ; -- 2 : Lalin et périphérie du Bassin d'Ordenes ; -- 3 : Bragança - Vinhais ; -- 4 : Morais.

D. -- Amas sulfurés (Cu-Fe) de l'Alcudien moyen.
1 : Cabo Ortegal ; -- 2 : Bassin d'Ordenes ; -- 3 : Cerro Muriano.
SC : Santiago de Compostela ; -- C : Cordoue.

E. -- Niveaux siliceux et phtanites de l'Alcudien moyen.
Mn : occurrences manganésifères.
B : Bragança.

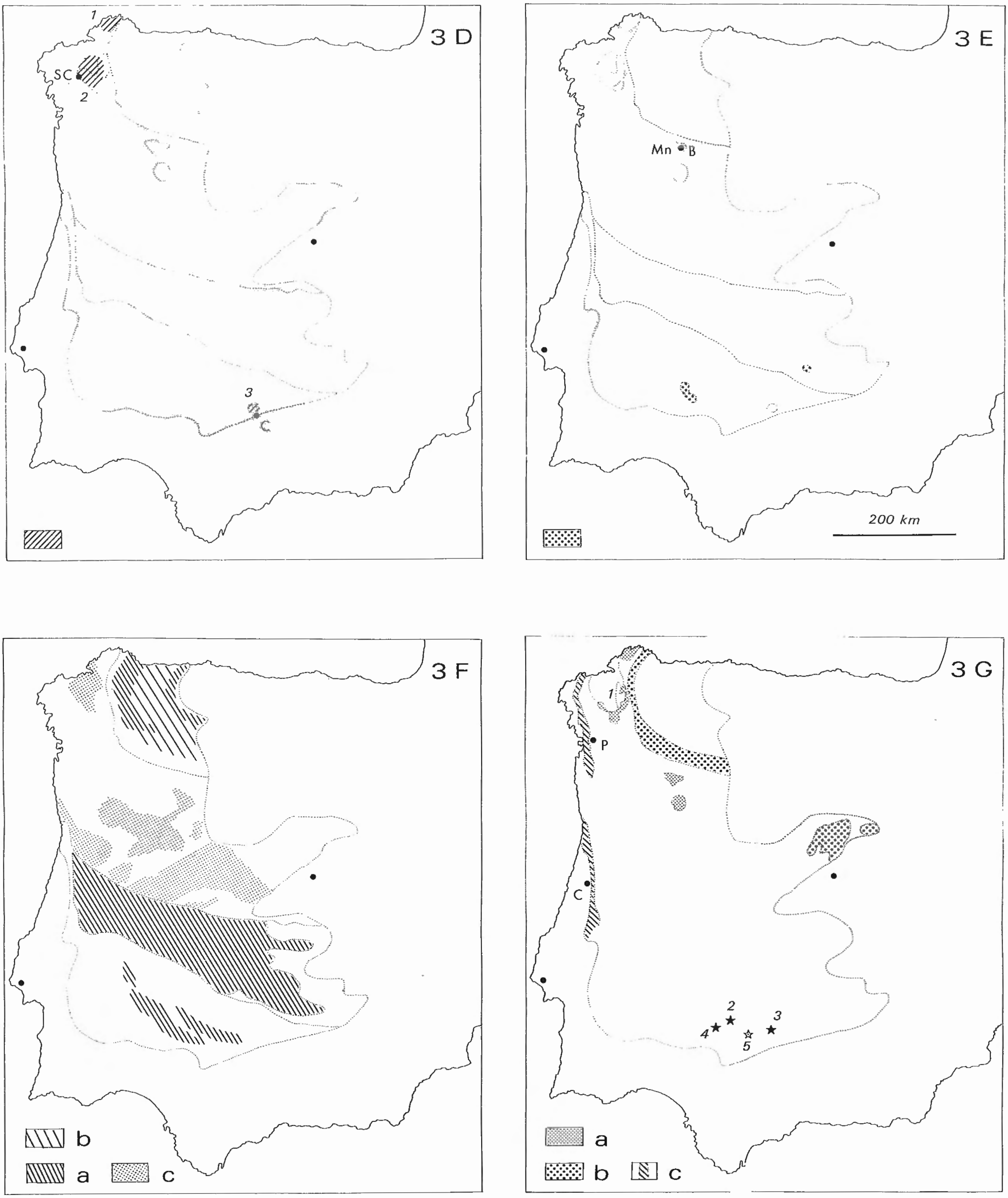
2. Etape moyenne.

F. -- Aires à flysch schisto-gréso-grauwackeux de l'Alcudien supérieur.
Peu ou pas métamorphique, a : affleurant ; -- b : supposé existant en profondeur ;
Métamorphique, c : de la Zone galaïco-castillane.

3. Fin de l'étape moyenne et étape tardive.

G. -- Quelques métamorphites et granitoïdes cadomiens (ou considérés comme tels) remarquables.
a : métamorphisme de faciès méso- à catazonal des Internides de Galice - Tras-os-Montes (affectant notamment le cortège ophiolitique, et comprenant aussi les orthogneiss [1]) ; -- b : porphyroïdes de Galice orientale et du Guadarrama oriental, du faciès "ollo de sapo" ; -- c : linéaments et couloirs mylonitiques à blastomylonitiques de Pontevedra (P) et de Coimbra (C), à métamorphites cadomiennes reprises au Calédonien et à l'Hercynien.
1 : orthogneiss de la partie orientale du Bassin d'Ordenes ; -- 2 : orthogneiss de Granja - Torrehermosa ; -- 3 : granitoïde du N de Villaharta (?) ; -- 4 : granite de Valverde de Lierena - Ahillones ; -- 5 : granite de las Minas "Gloria" (?).

Figures 23.3 A-C LE PROTEROZOÏQUE DES HESPERIDES (partie)



Figures 23.3 D-G   LE PROTEROZOÏQUE DES HESPERIDES   (fin)

-déformations ultérieures, restant à préciser;

-phases épirogéniques proto-varisques: sarde, taconique, et reussienne;

-absence de toute phase plicative bretonne *s. str.* (c-à-d. à la limite du Dévonien - Carbonifère).

-Tectogenèse varisque polyphasée:

-phase majeure post-viséenne (à post - Namurien basal), responsable des grandes structures plissées WNW-ESE (à NW-SE) déversées au SSW, puis cisaillées par un écaillage d'amplitude hectométrique, avec interférence possible d'un plissement mineur ENE-WSW;

-phases plicatives ultérieures, mineures - NNE-SSW et NNW-SSE; phase tangentielle avec chevauchements du SSW au NNE, d'amplitude kilométrique, et anté-Stéphanien B;

-rôle important de la fracturation profonde et de la tectonique de blocs.

C. Métamorphisme

-Métamorphisme cadomien: anchizonal;

-Métamorphisme varisque:

-selon les régions, inexistant, anchi- et épizonal: de basse pression, accompagnant la phase majeure de plissement.

D. Plutonisme

-Plutonisme granodioritique et granitique calco-alcalin tardi-varisque:

-très réduit dans le SE hespérique - stocks de la Garlitos et de Fontanosas (daté de 302-301 ± 7 Ma) et cortèges de filons basiques d'Almaden (305-301-298 ± 15 Ma) (Leutwein *et al.*, 1970) et de filons rhyolitiques de Chillon (302 ± 15 Ma) (Bellon et Rides, 1977);

-bien développé en haute Estrémadure - massifs de Montanchez (313 % 10 Ma) (Penha et Arribas, 1974), Miajadas, Plasenzuela, Trujillo, Caceres, Ceclavin - Zarza la Mayor, et stock de Logrosan (325 ± 16 Ma) (Bellon et Rides, 1977).

E. Minéralisation

-Minéralisations filoniennes mésothermales plombo-argentifères d'El Centenillo et de "Diogenes" (270 ± 30 Ma) (Tamain, 1968, et 1972).

2.2. Le Linéament de Cordoue

Cette bande linéamentaire, large de 60 km, est bien définie sur 300 km (du Guadalquivir à Badajoz) tout au long desquels elle est parallèle aux mégastructures varisques (Tamain, 1972 et 1975; Blachère *et al.*, 1977).

Apparu à l'aube du Cambrien et ayant joué en barrière paléogéographique au Paléozoïque inférieur, ce linéament a vu ainsi, pendant la mobilisation varisque, se mettre successivement en place:

-la ligne ophiolitique du Varas - Guadalbarbo (Crousilles *et al.*, 1976) - avec hyperbasites (à Cr) et basites (348 ± 18 et 334 ± 16 Ma) (Bellon et Rides, 1977);

-le complexe basique de Los Ojuelos (Delgado Quesada et Fontboté, 1970; Delgado Quesada, 1971), supposé précoce;

-le sillon à flysch de Los Pedroches;

-les hauts-fonds viséo-serpoukhoviens d'Adamuz, Villaharta;

-le couloir blastomylonitique de Badajoz - Cordoue (Bladier et Laurent, 1974);

-le bassin houiller à Westphalien B supérieur - C inférieur de Peñarroya -Belmez (précédé, d'ailleurs, par une première sédimentation houillère au Namurien) (Ortuño, 1970);

-le batholite granitique polyfacial tardif de Los Pedroches, avec ses faciès granodioritiques localement uranifères (305 ± 10 Ma et 291 ± 15 Ma), ses cortèges de filons rhyolitiques (315 ± 16 Ma) et rhyodacitiques 297 ± 15 Ma), et qui se continue par le massif d'Albuquerque - Valencia de Alcantara (287-281 ± 10 Ma) - Nisa (301-290 ± 9 Ma) (Mendès, 1968; Penha et Arribas, 1974; Bellon et Rides, 1977);

-une authentique ligne Plomb-Cuivre (au sens de Laznicka et Wilson, 1972), juste sur le bord nord-du dit batholite.

On y trouve également de très étroites bandes de métamorphites précambriennes: l'une d'elles est constituée de gneiss fins envahis, au Nord de Villaharta (N de Cordoue), par un granitoïde daté de 452 ± 22 Ma (Bellon et Rides, 1977).

Au NW de Badajoz, déjà au Portugal, le Linéament de Cordoue est moins bien connu.

En fait aussi partie, selon toute vraisemblance, le massif granitique alcalin de Portalegre, qui - avec un premier âge absolu de 358 ± 44 Ma (Mendès, 1968) puis un second (moyen) de 466 ± 12 Ma (Priem *et al.*, 1970), et malgré une cataclase ultérieure - pourrait être considéré, dans une approche directe et sous réserve, comme une manifestation soit tardi-reussienne, soit taconique.

2.3. L'Ossa Morena

Comme la Zone luso-orétane, elle est découpée en étroits domaines allongés selon la direction varisque majeure. Elle voit la réactivation, pendant la mobilisation varisque, d'un socle précambrien métamorphique. Ses caractères sont les suivants (Bard, 1969; Delgado Quesada et Fontboté, 1970; Delgado Quesada, 1971; Julivert *et al.*, 1974):

A. Stratigraphie

-Protérozoïque moyen, anté-alcudien: ?

-Protérozoïque supérieur (Alcudien):

-inférieur: restant à définir, mais peut-être (?) représenté en partie par les "quartzites de la Albarrana", blancs, très purs, en bancs puissants, de plus en plus feldspathiques vers le haut, et recouverts successivement par des paragneiss, des amphibolites, et de l'Alcudien supérieur ("schistes d'Azuaga");

-moyen: notamment à basites et petits amas sulfurés du Cerro Muriano (N de Cordoue), phtanites de basse Estrémadure, (fig. 23.3, D et E);

-supérieur: schisto-grauwackeux (Azuaga) (fig. 23.3, F);

-posthume: avec magmatisme andésitique (Cordoue) comparable à celui du "Précambrien III" du Maroc (Capdevila *et al.*, 1971), p.ex. de la "Série de Ouarzazate".

-Paléozoïque inférieur: Cambrien inférieur et moyen *p.p.* bien développé (avec basites, spilites et dolérites, dans le S et SW), Ordovicien réduit ou manquant, Silurien irrégulièrement exprimé (puissant au SW; à volcanites et lydiennes dans l'Alentejo), et Dévonien inférieur épicontinental, schisteux (Robardet, 1976).

-Paléozoïque supérieur:

-Dinantien: houiller continental de Val-de-Infierno (Wagner, 1976) au Tournaisien supérieur (à Viséen inférieur ?); Tournaisien - Viséen inférieur, Viséen supérieur (flysch schisto-grauwackeux), Viséen supérieur à terminal - Namurien (calcaires et houiller) (Armengot et Martinez Diaz, 1972; Crousilles *et al.*, 1976; Ortuño, 1970) du NW de Cordoue; flysch schisto-grauwackeux du Viséen inférieur (?) d'Estremoz - Terena - Barrancos;

-Westphalien B supérieur - C inférieur, paralique et discordant, du NW de Cordoue (Ortuño, 1970); Westphalien B cf. inférieur de Villanueva del Rio y Minas (Ortuño, 1970); Westphalien D de Santa Susana; Westphalien D - Stéphanien A et B des petits bassins du Nord de Cazalla de la Sierra (Mingarro Martin, 1962);

-Stéphano-Autunien du bassin de Guadalcanal: avec Stéphanien C - D (à Autunien) à la Cantera de Ladrillos, et Autunien (inférieur, et passage au supérieur) au Charco de la Sal (Broutin, 1974; Doubinger et Broutin, 1976); Stéphanien supérieur - Autunien du bassin d'El Viar.

B. Tectogenèse

-Tectogenèse cadomienne: avec une phase majeure génératrice des structures NNW-SSE;

-absence apparente de toute phase orogénique bretonne *s. str.*, ou post-emsienne et anté-viséenne (Schermerhorn, 1971).

-Tectogenèse varisque polyphasée:

-phase majeure post-Namurien basal et anté-Westphalien B (du NW de Cordoue), responsable des plis WNW/NW-ESE/SE;

-reprises ultérieures, l'une post-Westphalien C inférieur et anté-Westphalien D, une autre post-Stéphanien A-B et anté-Stéphanien supérieur (=Stéphano-Autunien).

C. Métamorphisme

-Métamorphisme régional cadomien plurifacial (épi-, méso, voire catazonal, (variables dans l'espace et encore mal connu;

-Métamorphisme varisque de basse pression (Aracena - Lora del Rio) à intermédiaire de basse pression (Azuaga - Villaviciosa de Cordoba) (Bard, 1969); avec blastomylonitisation syn-métamorphique à la limite des domaines, unités et blocs.

D. Magmatisme

-Magmatismes variés (Bard, 1969 et 1971; Bard et Fabriès, 1970; Barros e Carvalhosa, 1970; Fabriès, 1963; Mendès, 1968):

-plutonisme tardi-cadomien, granitique - stock de Valverde de Llerena - Ahillones (Delgado Quesada et Fontboté, 1970; Delgado Quesada, 1971);

-complexes granitoïdiques varisques, syn-tectoniques - gneiss de Montemoro-Novo (323 Ma);

-granites et granodiorites, postérieurs à la phase majeure, d'âge méso-westphalien - diorite quartzique d'Evora (304 ± 4 Ma), granite calco-alcalin à biotite de Pedrogâo (308 ± 4 Ma);

-ensemble post-tectonique, tardif - granite alcalin à biotite de Monforte - Santa Eulalia, d'âge moyen stéphanien (303 et 281 ± 12 Ma).

3. LE DOMAINE SUD-PORTUGAIS

Séparé de l'Ossa Morena par le grand accident de Beja - Aracena, ce domaine, qui comprend aussi toute la partie espagnole de la Ceinture pyriteuse, présente les caractères suivants (Schermerhorn, 1971):

A. Stratigraphie

-Paléozoïque supérieur, très bien exprimé:

-à la lisière NW: Westphalien D discordant (bassin de Santa Susana);

-dans la Ceinture pyriteuse: Famennien schisto-gréso-quartzitique, Tournaisien - Viséen inférieur volcano-siliceux (à basites, spilites et albitophyres, kératophyres et tufs, amas sulfurés, jaspes et lydiennes, occurrences de Mn-Fe), Viséen supérieur (à puissant flysch de faciès Culm: Mertola);

-dans le SW: Dinantien et Namurien schisteux (à Goniatites), Namuro-Westphalien (à flysch de plus en plus tardif vers le SW - Aljustrel).

B. Tectonique et Tectogenèse

-Tectonique de blocs et fracturation profonde à la lisière nord: dès le Cambrien (sur la marge même de la géosuture de Beja - Aracena);

-absence de toute phase orogénique bretonne *s.str.*, ou post-emsienne et anté-viséenne;

-Tectogenèse varisque polyphasée:

-phase majeure post-namurienne (à post-Westphalien A ? dans le SW) et anté-Westphalien D (au NW), responsable des grandes structures E-W à WNW-ESE, déversées au SSW/SW et dessinant une légère virgation (à concavité au NE), puis cisaillées, écaillées et chevauchantes du NNE au SSW (à l'amplitude croissante de Ficalho à Carrapateira, où la translation est d'ordre décakilométrique) (Feio et Ribeiro, 1971);

-reprises ultérieures, mineures, phase NW-SE post-Westphalien D (Santa Susana) replissant les structures antérieures de façon homoaxiale, puis phase NNW-SSE (Feio et Ribeiro, 1971).

C. Métamorphisme

-Métamorphisme régional varisque:

-absent, anchi- à épizonal, dans la Ceinture pyriteuse, il est polyphasé et atteint son degré maximal (sous-faciès épizonal des schistes verts) dès la phase majeure de plissement (Lécolle, 1977).

D. Magmatisme et Plutonisme

-Magmatisme initial, basique puis acide, spécifique de l'Étape préliminaire à précoce de la mobilisation varisque.

-Plutonisme tardi-varisque, postérieur à la grande tectonique, au chimisme varié, avec des faciès subvolcaniques, et localisé à la bordure NE du domaine.

III. STRUCTURATION DES HESPERIDES ET MOBILISATION PROTO-VARISQUE (= CALEDONIENNE).

Pour avoir une meilleur idée de cette mobilisation proto-varisque, il n'est pas inutile de la replacer dans le cadre plus vaste de la structuration progressive des Hespérides.

1. INTRODUCTION

Les géologues du XIX° siècle distinguèrent deux grands ensembles structuraux: un socle "archéen", et une couverture paléozoïque plissée à l'Hercynien.

Par contre, Stille (1927), Lotze (1945) et leurs disciples considérèrent les Hespérides comme un bâti essentiellement et fondamentalement varisque, issu d'une très longue préparation quasi ininterrompue depuis l' "Algonkien": étonnante singularité par rapport au restant de la vieille Europe, qui transparut d'ailleurs dans la Carte Tectonique de l'Europe au 1/2.500.000° (Moscou, 1962).

Ce n'est que récemment - à partir d'études locales ou régionales détaillées et de levés cartographiques très précis, puis d'essais d'intégration des Hespérides aux chaînes cadomienne et varisque de l'Europe moyenne - qu'a pu être mise en évidence l'individualité de Cadomides hespériques sur lesquelles et à partir desquelles s'est développée ultérieurement la partie SW du segment varisque européen.

2. LE "VIEUX" SOCLE PRECAMBRIEN

Bien que le nombre, l'importance et la distribution des noyaux pentévriens ou, d'une façon plus générale, anté-alcudiens ne nous soient pas encore connus, on peut néanmoins distinguer aujourd'hui, dans les Hespérides, un voire deux domaines à Anté-Alcudien (fig. 23.3, B): - la Galice, à coeur pentévrien; - et l'éventuel noyau migmatique de Tolède.

Rappelons aussi que, dans l'Alcudien supérieur schisto-grauwackeux de la Zone luso-orétane, ont été observés au microscope (Bouyx, 1969 et 1970) d'exceptionnels microgalets de micaschiste, de très rares fragments de micropegmatite, et quelques esquilles de quartz filonien, vestiges qui témoignent bien de la présence, au loin, d'un vieux socle métamorphisé et granitisé, alors exondé (Tamain et Ovtracht, 1971a et b).

3. LA MOBILISATION CADOMIENNE

Elle s'est développée tout au long de l'Alcudien (=Protérozoïque supérieur hespérique), dans un vaste territoire qui comprendra plus tard les zones varisques léono - ouest-asturienne, galaïco-castillane, luso-orétane et ossa-marianique.

Notre connaissance, à son sujet, reste éminemment perfectible. En voici, néanmoins, les grandes lignes (fig. 23.3 et 23.4).

3.1. L'Etape préliminaire a précoce

On ne sait encore rien du début de cette mobilisation, ni des prémices d'organisation des futurs domaines litho-structuraux.

La première phase identifiable se manifeste à l'aube de l'Alcudien moyen en Galice et dans le Nord du Portugal (Tras-os-Montes), puis en Ossa Morena.

S'individualise en effet, dans le NW hespérique, un domaine de type eugéosynclinal, avec apparition d'un magmatisme initial ophiolitique (à péridotites, dunites, saxonites) accompagné d'une première métallogenèse à chrome - platine (Ann. 1): Bragança - Vinhais, etc. (fig. 23.3, C) et se poursuivant par un volcanisme bimodal basique puis acide, accompagné d'une métallogenèse spécifique, sous la forme d'amas sulfurés du type cupro-pyriteux: Cabo Ortegal (Ann. 2), bassin d'Ordenes (Ann. 3), etc. (fig. 23.3, D).

Presque au même moment, un magmatisme basique se développe évoluant en un volcanisme *cf.* bimodal (ortho-amphibolites et complexe leptyno-amphibolitique) (Deloche *et al.*, 1977), puis de petits amas sulfurés se constituent (Tamain, 1977) au Cerro Muriano, en Ossa Morena orientale (fig. 23.3, D).

En suite, la série volcano-sédimentaire évolue et voit le dépôt de niveaux siliceux (phtanites): dans la Zone luso-orétane, en Ossa Morena, tandis que se développe localement une troisième métallogenèse, à manganèse ou fer-manganèse cette fois-ci: Sobralhal (Edrosa, près de Bragança), etc. (fig. 23.2, E).

Par contre, on ne connaît pas encore la délimitation exacte du domaine de type miogéosynclinal.

3.2. L'Etape moyenne

Elle voit l'installation et le développement, pendant l'Alcudien supérieur, d'une sédimentation terrigène caractérisée par le dépôt d'un flysch schisto-gréso-grauwackeux qui s'étend, en outre, à la future Zone léono - ouest-asturienne (fig. 23.3, F). Ce flysch admet, dans sa partie supérieure, des conglomérats intraformationnels (Ann. 4) et des passées carbonatées également lenticulaires (Ann. 5).

Elle se termine par la phase majeure de l'orogenèse cadomienne, au cours de laquelle la totalité du territoire (fig. 23.3, A) est plissée (Ann. 6), mais dont seuls les domaines les plus internes (Galice - Tras-os-Montes - Sierra Centrales castillanes, et Ossa Morena *p.p.*) sont métamorphisés et plus ou moins granitisés (fig. 23.3, G).

3.3. L'Etape tardive

Ses caractéristiques restent encore mal définies, mis à part certain plutonisme acide tardi- à post-tectonique (fig. 23.3, G) (Ann. 7).

3.4. Alcudien, Algonkien, Briovérien

Les tableaux 23.1, 23.2, et 23.3 donnent l'équivalence des différents termes constituant respectivement chacun de ces trois systèmes.

De leur confrontation, il ressort une identité quasi terme à terme et, si l'on veut tenir compte du schéma classique de l'évolution des zones mobiles (selon U.A. Bilibine, ou P.M. Tatarinov et coll., p.ex.), on a en l'Algonkien bohémien une référence idéale, surtout si l'on considère que la base, ou la partie inférieure, de l'Alcudien ne nous est pas encore connue ou n'a pas encore été différenciée.

3.5. Conclusion

Au Protérozoïque supérieur, une grande partie des hespérides est ainsi intervenue activement dans la structuration de la chaîne cadomienne d'Europe, avec, notamment: la remobilisation de vieux noyaux pentévriens ou anté-alcudiens; l'individualisation, dès le début de l'Alcudien moyen, d'un domaine eugéosynclinal au cortège ophiolitique bien exprimé; et, tous les caractères d'une évolution "classique" complète de zone mobile en plateforme.

Il y a donc eu une continuité certaine de cette chaîne cadomienne - et des structures l'ayant préfigurée - depuis, au moins, les confins orientaux du Massif bohémien jusqu'au Guadalquivir actuel, et ce n'est qu'intégrée à l'Europe que sa partie hespérique doit être désormais envisagée, bien qu'il ne soit pas, non plus, sans intérêt d'en rechercher l'éventuel devenir nord-africain.

TABLEAU 23.1.

L'ALCUDIEN HESPERIQUE

		OSSA MORENA ORIENTALE Cerro Muriano	SIERRA MORENA CENTRALE Alcudia	NORD-OUEST HESPERIQUE
ALCUDIEN	SUPERIEUR	Schistes d'Azuaga	Série conglomératique : (conglomérats (± aurifères) à galets (de phtanite ((Calcaires de Cabezarrubias-Hinojosas (lenticulaires et interstratifiés	
			Série schisto-gréso-grauwackeuse : flysch schistes, grès, grauwackes	
	MOYEN		Série schisto-siliceuse : à phtanites ?	Schistes de Sobralhal : à Mn-Fe
		Série des paragneiss à amas sulfurés (pyrrhotite, pyrite,...)		Série des amas sulfurés (chalco-pyrite, pyrite,...) d'Ortegal - Compostela
		Série à basites : à ortho-amphibolites, leptynites - - - - - - - ? - - - - - - - - - -		Série à basites et hyperbasites (à Cr-Pt) de Vinhais - Bragança ?
	INFERIEUR	Série de base : ((Quartzites de la Albarrana ((?		?
SOCLE METAMORPHIQUE ANTE-ALCUDIEN				

TABLEAU 23.2.

LE BRIOVERIEN DU MASSIF ARMORICAIN (FRANCE)

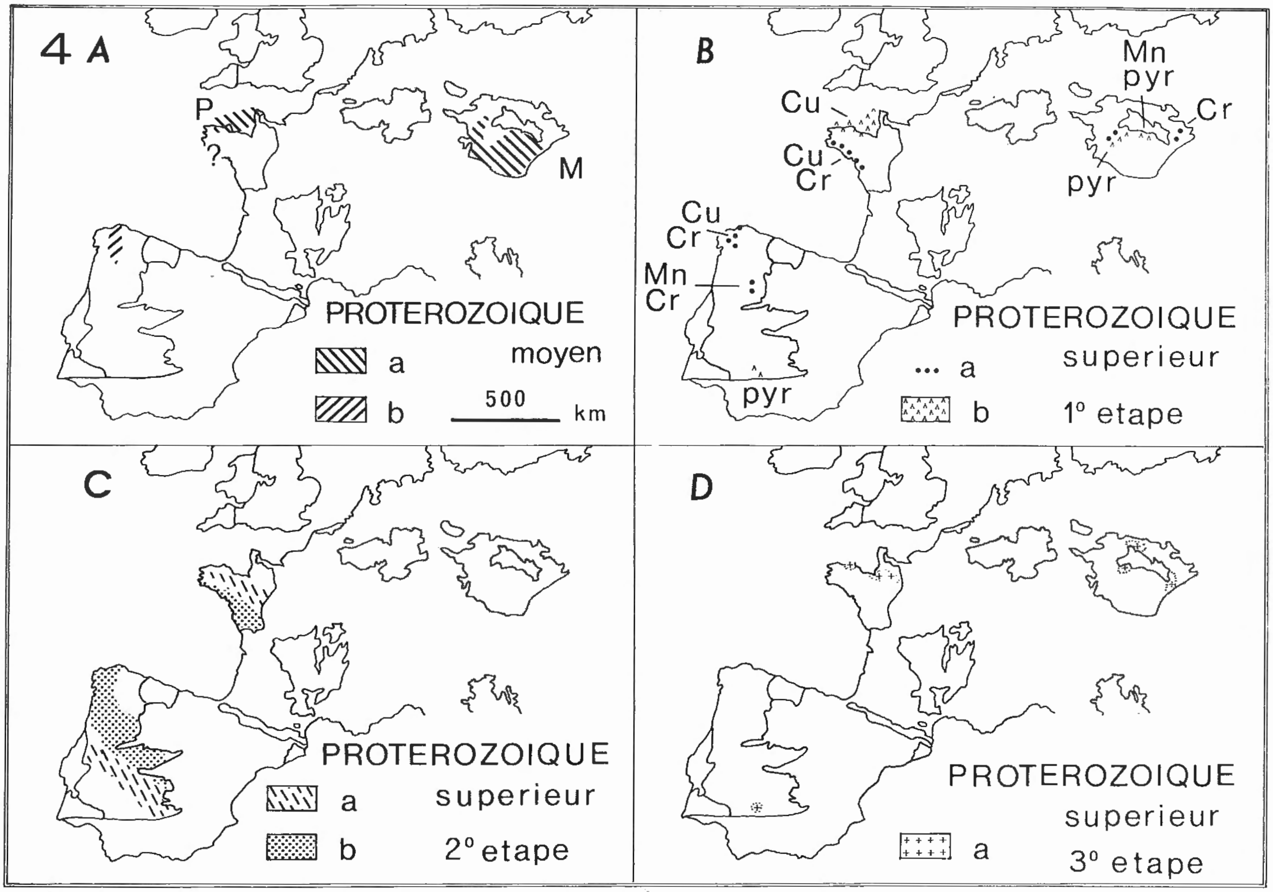
		BRETAGNE MERIDIONALE ET VENDEE	BRETAGNE MOYENNE	BRETAGNE SEPTENTRIONALE ET NORMANDIE
		Cambrien (moyen)	Ordovicien	Cambrien inférieur
^^ 1 ^^^	^^^^	^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^^ 1 ^^^	^^^^^^^^^^^^^^^^^^^^^^^^^^^^^ 1 ^^^	^^ 1 ^^^^
BRIOVERIEN	SUPERIEUR	Schistes, grès, grauwackes, conglomérats, et <u>Cipolin de Beaupréau</u>	<u>Dalles de Néant</u> : flysch schisto-gréseux, et <u>Calcaire de Saint-Thurial</u>	<u>Etage de la Laize</u> : flysch schisto - grauwackeux, à grès calcaires
		<u>Poudingue de Saint-Fulgent</u>, = <u>Poudingue de Loiré</u> : brèches et conglomérats (à galets de phtanite)	<u>Poudingue de Gourin</u> : conglomérats ± aurifères (à galets de phtanite)	<u>Etage de Granville</u> : tillites et schistes varvés, conglomérats (à galets de phtanite)
				ou ^^^^^^^^^^^^^^^ 2 ^^^^
	MOYEN	Quartzophyllades	Schistes et phyllades	<u>Etage de Villiers - Fossard</u> = <u>Schistes et phyllades de Saint - Lô</u> : à grès
		Schistes et phtanites, ou : micaschistes graphiteux	<u>Schistes de Lamballe</u> : à phtanites	<u>Etage de la Lande des Vardes</u> = <u>Schistes de Lamballe</u> : à phtanites et schistes graphiteux
	INFERIEUR	Schistes verts, amphibolites, micaschistes albitiques (= métaspilites), métadolérites, métagabbros, serpentinites	?	<u>Amphibolites de Lanvollon</u> : à petits amas sulfurés, pillow-lavas
		Migmatites		PENTEVRIEN ^^^^^^^ 3 : 1100 - 900 MA. ^^^^^^^ 4 : 2000 - 1900 MA. ^^^^^^^ 5 : 2700 - 2550 MA.

1 : orogenèse cadomienne ; -- 2 : hypothétique orogenèse constantienne ; -- 3 - 4 - 5 : orogenèses respectivement pentévrienne, lihouienne et icartienne.

TABLEAU 23.3.

L'ALGONKIEN DU MASSIF BOHEMIEN

		T H U R I N G E : R. D. A. ANTICLINAL DE SCHWARZBURG	T C H E C O S L O V A Q U I E BASSIN BARRANDIEN	ZELEZNE HORY
		Cambrien	Cambrien moyen	Cambrien moyen
^^^ 1 ^^^		^^^^^^^^^^^^^^^^^^^^	^^^^^^^^^^^^^^^^^^^^	^^^^^^^^^^^^^^^^^^^^ 1 ^^^
A L G O N K I E N	RECENT	COUCHES DE KATZHÜTTE SUPERIEURES : schistes, grauwackes, conglomérats	GROUPE POST - SPILITIQUE : flysch (Conglomérat de Dobris : à tillites (et varves (schistes, grauwackes, conglomérats	SUB - CAMBRIEN : flysch (Conglomérat de Litosice : à tillites (schistes, grauwackes, conglomérats ou ^^^^^^^^^^^^^^^ 2 ^^^
	M O Y E N	COUCHES D' ALTENFELD : schistes et phyllades (schistes siliceux (et phtanites (schistes alunifères (diabases, métadiabases	GROUPE SPILITIQUE : (Formation Zbiroh : schistes, tufs (et phtanites (Formation Kamenec : schistes, psam- (mites, (à phtanites, (à schistes pyriteux, tufs, spilites (Formation Hromnice : schistes et (grauwackes, + schistes alunifères, (amas sulfurés, spilites	(Formation Post-minéralisation : (schistes siliceux, tufs, phtanites (Formation minéralisée : schistes gra- (phiteux, gréseux, grauwackeux, (à occurrences de Mn-Fe, (à amas sulfurés (Formation Pré-minéralisation : schis- (tes, grauwackes, calcaires, conglomé- (rats, + tufs, pillow-lavas, spilites, (diabases et gabbros
	ANCIEN	COUCHES DE KATZHÜTTE INFERIEURES : schistes, grauwackes, quartzites	GROUPE PRE - SPILITIQUE : pélites, phyl- lades, schistes, arkoses, grauwackes	? socle métamorphique
		vieux socle métamorphique	MOLDANUBIEN	1 : Orogenèse cadomienne ; -- 2 : Orogenè- se zéleznohorienne (= de l'Eisengebirge).



A. PROTEROZOÏQUE MOYEN. a: de référence (M: Moldanubien; P: Pentévrien); b: hespérique. (Est aussi posée la question -- compte tenu de l'âge radiométrique de 900 Ma. trouvé en Galice, de la symétrie des ensembles hespérique et armorician de part et d'autre de la Mer cantabrique, de certaines analogies avec le Massif bohémien, ainsi que de nouvelles hypothèses de recherche (Brillanceau et Nicolas, 1971) -- de l'existence du socle pentévrien dans le domaine vendéo - sud-breton).

B. PROTEROZOÏQUE SUPERIEUR: Etape préliminaire à précoce. a: magmatisme initial ophiolotique (hyperbasites et basites); b: *id.* (basites); Cr: chrome; pyr. et Cu: amas sulfurés du type pyriteux et du type cupro-pyriteux; Mn: Manganèse (+ Fer).

C. PROTEROZOÏQUE SUPERIEUR: Etape moyenne. a: flysch schisto-gréso-grauwackeux, peu ou pas métamorphisé ultérieurement; b: flysch fortement métamorphisé au Cadomien, puis à l'Hercynien. (N'est pas pris en considération le flysch, de même nature, des domaines normano - nord-breton, d'Ossa Morena et de la Zone léono - ouest-asturienne).

D. PROTEROZOÏQUE SUPERIEUR: Etape tardive. a: plutonisme granitique tardi-cadomien.

Figures 23.4 A-D. CORRELATIONS DU PROTEROZOÏQUE D'EUROPE CENTRALE (MASSIF BOHEMIEN) ET OCCIDENTALE (ARC IBERO-ARMORICAIN).

4. LA PREPARATION PROTO-VARISQUE, OU CALEDONIENNE

Elle commence avec les premières incursions, dès l'aube du Cambrien, des mers paléozoïques sur la pénéplaine cadomienne.

Elle se poursuit avec l'individualisation progressive de grandes aires, qui vont évoluer d'une façon plus ou moins autonome:

-au Nord, l'ensemble des futures zones cantabrique et léono - ouest-asturienne, où la sédimentation est continue pratiquement du Cambrien à la fin de l'Ordovicien;

- au centre, la Galice occidentale et la Zone luso-orétane, où, d'une façon générale, la préparation proto-varisque comporte trois cycles sédimentaires: cambrien *p.p.*, ordovicien et siluro-éodévonien, séparés par les crises épirogéniques sarde et taconique;

- au Sud, la moitié méridionale de l'Ossa Morena, où l'Ordovicien se présente sous des faciès particuliers.

Elle se termine: dans l'ensemble centro-méridional, au Dévonien moyen, avec la phase reussienne (de caractère également épirogénique); dans l'ensemble septentrional, à la limite Frasnien - Famennien.

Jusqu'à ces tout derniers temps, elle était unanimement considérée comme la première étape de la mobilisation varisque. Elle correspond, en réalité, à l'évolution d'une aire de plateforme soumise aux contrecoups de la mobilisation calédonienne qui, elle, se développe de la Scandinavie à la Mauritanie, sur la marge occidentale des Hespérides.

4.1. L'Etape cambrienne

Trois aires sont envahies par la mer cambrienne, transgressive: le bassin cantabrique, au Nord; le bassin tolédan, ou tolédano-salmantin, au centre; et, le bassin andalou, au Sud.

A. Le bassin cantabrique

Il apparaît dès le début du Cambrien - base des "Couches de Barrios" détritiques, à *Lunolenus lotzei* Sdzuy - sur l'emplacement des futures zones varisques léono - ouest-asturienne et cantabrique (franges W et S) (Llopis Llado, 1962; Lotze, 1970) et fonctionne apparemment pendant la plus grande partie du Cambrien, alimenté par une sédimentation - dans ses grandes lignes - d'abord détritique, puis carbonatée, et de nouveau détritique.

L'existence du Cambrien supérieur est attestée en Asturie par la présence de l'ichnite *Cruziana semiplicata* Salter (Lotze, 1970; Marcos, 1973) dans des faciès détritiques gréso-quartzitiques passant au sommet, en apparente continuité, à l'Ordovicien inférieur de mêmes caractéristiques lithofaciales et à *Cruziana rugosa* d'Orbigny, *Cr. furcifera* d'Orb.

Il se poursuit au SE dans la Sierra de la Demanda et la Chaîne celtibérique, hors des Hespérides. Élément de la sous-province hispano-sarde, il offre une certaine parenté faunique avec la Montagne Noire (S. France) pendant le Cambrien inférieur et une partie du Cambrien moyen (H. et G. Termier, 1964).

Il a une forme dissymétrique marquée et présente un maximum de subsidence (avec près de 6000 m de sédiments) dans l'axe du rio Navia, ainsi qu'une pente très redressée presque à l'aplomb de l'actuel Anticlinorium du Narcea (à Protérozoïque supérieur), à la limite des deux futures zones varisques: cette limite est donc tectoniquement active tout au long du Cambrien et voit le lent enfoncement des Asturies occidentales au pied de la Zone cantabrique formant alors un relief sous-marin. Autre élément tectonique important à cette époque: la "Ligne du Léon", déjà active (de Sitter, 1965) et qui relaie en quelque sorte, avec sa direction E-W, l'accident précédent.

B. Le bassin tolédan, ou tolédano-salmantin

Apparu également au Cambrien inférieur, il voit une sédimentation détritique à carbonatée - avec, selon des niveaux, Algues, Archéocyathidés (Martin Escorza et Perejon, 1972), Méduses, Brachiopodes, Lamellibranches et Trilobites: *Eodiscus (Serrodiscus) speciosus* Ford (Aparico Yagüe et Gil Cid, 1972) du Cambrien inférieur, ou encore: *Realaspis strenoides* Sdzuy, *Pseudolenus weggeni* Sd., *Ps. glaber* Sd., *Kingaspis cf. velatus* Sd., d'un Cambrien inférieur élevé (Lotze, 1970; Gil Cid *et al.*, 1976) - se développer selon deux bras de mer orientés E-W (Martin Escorza, 1976).

Le premier de ceux-ci est le plus méridional, et aussi le mieux défini; son axe passerait juste au Nord de Tolède. Le second a son axe passant par Salamanque; il présente un Cambrien inférieur assez bas, à *Pararedlichia sp.* (Garcia de Figuerola et Martinez Garcia, 1972).

C. Le bassin andalou

C'est dans sa partie orientale, à Cordoue, que s'installe la sédimentation marine au tout début du Cambrien.

Et durant le Cambrien basal puis inférieur, vont alterner les épisodes détritiques et les épisodes carbonatés, dans des eaux peu profondes et agitées, parfois en ambiance paralique (avec décharges de matériel alluvial ou deltaïque) (Liñan Guijarro et Dabrio, 1974).

Y prolifèrent, selon les faciès et les horizons, Algues, Archéocyathidés (Zamareño et Debrenne, 1977), Méduses (*Anthoichnites cabanasi* Melendez: Cabanas et Melendez y Melendez, 1960), et Trilobites: *Anadoxides richterorum* Sdzuy, *Lunolenus lotzei* Sd., *Dolerolenus formosus* Sd., *Dolerolenus sp.* (Linan Guijarro, 1972 et 1974).

Dans sa partie centrale et au Portugal, il est caractérisé par une sédimentation détrituque à carbonatée.

Les faciès détritiques, d'eaux peu profondes, ont livré des faunes néritiques à Brachiopodes, Lamellibranches, Ptéropodes et Trilobites:

p.ex. *Delgadella souzai caudata* Delgado, *Hicksia elvensis* Delg., *Callavia choffati*, de Vila Boim;

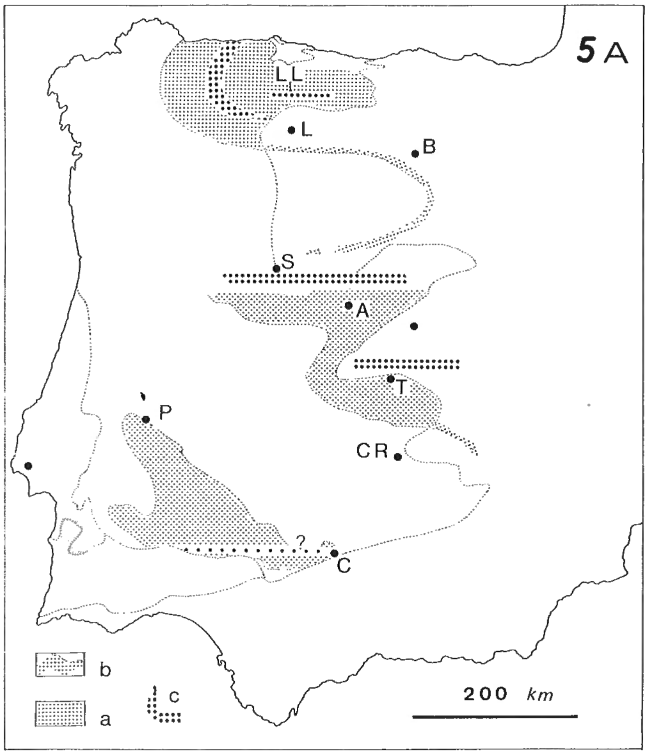
Delgadella souzai caudata Delg., *Protolenus schneideri* Richter, *Gigantopygus cf. bondoni* Hupé, etc. du niveau inférieur de Cala;

Eodiscus (Serrodiscus) serratus Richter, *E. (Serrodiscus) cf. speciosus* Ford, *Strenuaeva cf. vigilans* Matthew, etc. du niveau supérieur de Cala;

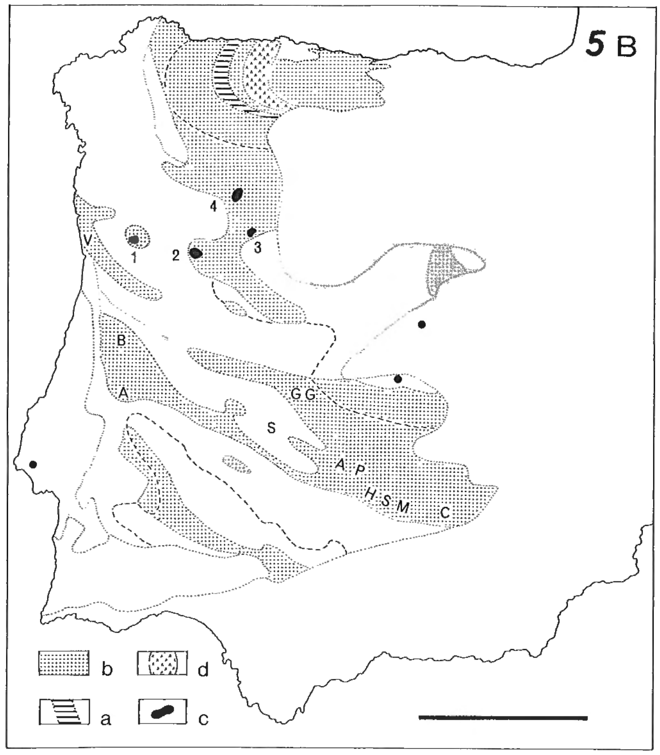
Saukianda andalusiae Richter, *Ellipsostrenua alanisiana* Sdzuy, *Perrector perrectus* Richter, *Alanisia guillermoi* Richter, etc. d'Alanis.

Les faunes de Cala et de Vila Boim sont identiques à celles qui, au Maroc, sont caractérisées par l'association *Micmacca - Serrodiscus*.

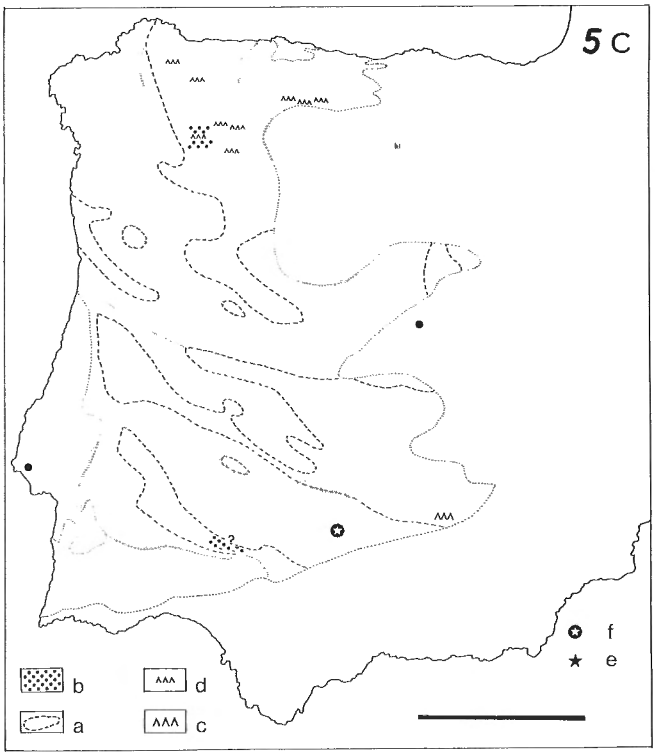
Par contre, la faune plus récente d'Alanis, qui comprend à la fois des *Redlichiacea* et des *Ellipsocephalacea*, montre que le bassin andalou, à cette époque-là, correspond - comme le Maroc et la Sardaigne - à une aire de transition entre la province à *Olenellus* nord-atlantique et la province à *Redlichia* "téthysienne" (Henningsmoen, 1957).



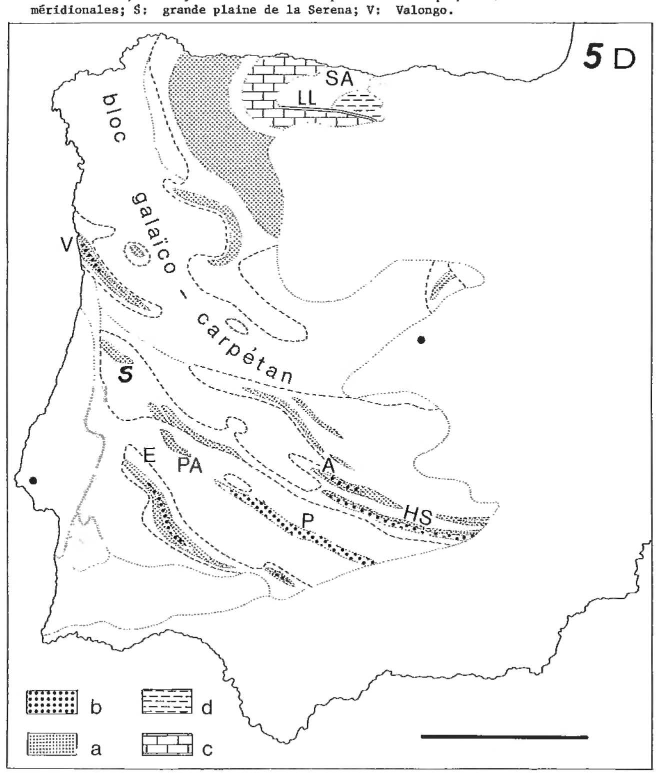
A. AU CAMBRIEN. a: aires à Cambrien; b: front de la mer cambrienne, à son maximum d'extension; c: axes des bassins à Cambrien; LL: "Ligne du Léon"; ?: hypothétique axe d'Aracena - Cordoue.
A: Avila; B: Burgos; C: Cordoue; CR: Ciudad Real; L: Léon; P: Portalegre; S: Salamanque; T: Tolède.



B. A L'ORDOVICIEN. a: emplacement de l'ancien sillon cambrien de Navia; b: aires à Ordovicien (+ a, c, d); c: principales occurrences de fer oolithique méso-ordovicien: 1, Serra de Marão; 2, Moncorvo; 3, Miranda do Douro; 4, Guadramil; d: aire à volcanisme basique du Cabo Peñas.
A: Amêndoa; AP: synclinal d'Almaden - Puertollano; B: Serra de Buçaco; C: El Centenillo; GG: synclinaux de Guadalupe - Guadarranque; HSM: Hautes Sierras méridionales; S: grande plaine de la Serena; V: Valongo.



C. PENDANT LA PHASE TACONIQUE, A LA LIMITE ORDOVICIEN - SILURIEN. a: aires à Ordovicien; b: éventuels dépôts glacio-marins; c: surface de ravinement glaciaire (El Centenillo); d: éventuelles surfaces de ravinement glaciaire; e: magmatisme anorogénique (granites, granites hyperalcalins, syénites); f: granitoïde du N de Villaharta, daté de 452 ± 22 Ma.



D. AU SILURIEN ET AU DEVONIEN INFERIEUR. a: aires à Silurien; b: aires à Dévonien inférieur des Hespérides centro-méridionales; c: bassin léono-asturien; d: bassin palencien.
S.A.: seuil asturien, limité au Sud par la "Ligne du Léon" (LL).
A: Almaden; E: Evora; HS: Hautes Sierras méridionales; PA: Portalegre - Albuquerque; P: Peñarroya; S: Serra de Buçaco; V: Valongo.

Figures 23.5 A-D LA PREPARATION PROTO-VARISQUE OU CALEDONIENNE

D. Remarques

En Sierra Morena, il existe deux niveaux stratigraphiquement différents de biohermes à Archéocyathidés (Zamarreño et Debrenne, 1977): celui de Las Ermitas (Cordoue), qui appartient à la moitié supérieure de l'Atdabanien; celui de Zafra, sous-jacent à la faune à *Saukianda* et qui appartiendrait au début du Lénien.

Ceci semblerait se retrouver dans le bras de mer tolédan.

Enfin, il reste à définir en termes de pétrologie moderne le volcanisme cambrien.

C'est le cas du complexe volcano-sédimentaire, à rhyolites, pyroclastites et cinérites récemment mis en évidence au-dessus du Cambrien inférieur et en-dessous de l'Ordovicien des Monts de Tolède (Martin Escorza, 1976).

C'est aussi le cas du volcanisme du bassin andalou, au chimisme moyennement acide à basique (spilitique), et qui est actuellement à l'étude.

E. Conclusion

La transgression cambrienne se propage sur la plateforme cadomienne à la faveur d'aires en voie de lent affaissement et que semble contrôler une tectonique cassante de direction E-W qui se manifeste notamment: au Nord, avec la "Ligne du Léon" à laquelle serait conjugé, dans les Asturies occidentales, un accident de direction apparemment subméridienne à NNE-SSW; dans le centre, avec les axes mêmes de bassins tolédan et salmantin; en Ossa Morena, éventuellement avec l'alignement d'Aracena - Cordoue.

Si l'on considère que la tectogenèse cadomienne a bien été contrôlée par une direction principale NNW-SSE et sa conjuguée E-W, on aurait alors un bel exemple de réactivation de celle-ci au Cambrien.

D'autre part, c'est seulement dans la partie septentrionale des Hespérides qu'est connu le Cambrien supérieur (daté par des ichnites en Asturie) alors qu'il faut aller jusque dans la Sierra de la Demanda pour le trouver daté par une faune trilobitique.

Or, c'est pendant la plus grande partie du Cambrien moyen et la totalité du Cambrien supérieur que les Hespérides méridionales, de Tolède au Guadalquivir, sont affectées par l'épirogenèse sarde.

4.2. L'Etape ordovicienne

Elle revêt une importance toute particulière (fig. 23.5, B).

Tout d'abord, elle voit la première transgression marine paléozoïque généralisée se développer sur l'ensemble des Hespérides, c'est-à-dire principalement dans les zones cantabrique, léono - ouest-asturienne, luso-orétane et en partie ossa-marianique, alors que la Galice moyenne et la région transmontane *p.p.* restent apparemment émergées.

Cette transgression est bien plus étendue que celle du Cambrien inférieur.

Néanmoins, en Ossa Morena, son existence dans la frange NE est beaucoup plus discrète qu'on ne l'a cru, tandis que son extension dans la partie S et SW est notablement réduite par rapport à l'aire cambrienne et que les caractères, qu'elle y présente, sont nettement différents de ceux des autres zones.

Peut-être voit-elle aussi se préciser les premiers grands traits paléogéographiques et structuraux qui, ultérieurement, vont guider et contrôler la mobilisation varisque?

Elle se termine enfin sur des évènements d'importance planétaire.

A. Le cycle ordovicien

D'une façon très générale, son évolution est assez uniforme et homogène sur l'ensemble des zones cantabrique, léono - ouest-asturienne, galaïco-castillane *p.p.* et luso-orétane:

- l'Ordovicien inférieur est notamment caractérisé par une sédimentation détritique spécifique des aires eurafricaines - celle des grès et quartzites "armoricains";

- l'Ordovicien moyen voit l'expansion des mers épicontinentales et le dépôt des fameux "schistes à Calymènes";

- l'Ordovicien supérieur, détritique (avec, localement, un épisode carbonaté éoashgillien), est le témoin d'une lente régression, au déroulement irrégulier et oscillant, qu'accompagne une activité volcanique (Ann. 8) parfois intense, le long des fractures profondes d'un système ou d'un réseau restant à définir.

Chacune de ces quatre régions présente, certes, ses particularismes régionaux.

Les deux premières forment une grande aire, dans la partie léono - ouest-asturienne de laquelle le sillon, nettement dissymétrique et largement ouvert à l'Est, apparu au Cambrien inférieur, évolue: sont flanc oriental (région de Luarca) voit l'accumulation de puissants dépôts détritiques à l'Ordovicien inférieur, ce qui aboutit à un rétrécissement du sillon dont le nouveau flanc oriental (région du Navia, à l'emplacement de l'ancien sillon à subsidence cambrienne optimale) devient à sont tour le lieu de dépôt privilégié des schistes méso-ordoviciens; puis, le sillon lui-même est comblé par des turbidites, et avorte à la fin de l'Ordovicien (Julivert *et al.*, 1972; Marcos, 1973).

En Galice et dans le Nord du Portugal, des occurrences de fer oolithique méso-ordoviciennes jalonnent le pied des terres émergées (fig. 23.5, B).

La Zone luso-orétane présente un Ordovicien complet et bien fossilifère, discordant ici sur l'Alcudien et là sur le Cambrien (quand celui-ci existe): série-type d'El Centenillo (2000-2200 m) (Tamain, 1967, 1971 et 1972; Tamain *et al.*, 1969; Tamain *et al.*, 1971), série d'Almaden - Puertollano, séries tolédano -nord-estrémègnes, dans le Sud-Est hespérique; série-type de la Serra de Buçaco (1360 m/min.) (Teixeira, 1955; Henry *et al.*, 1974) dont le conglomérat basal à grands éléments témoigne bien de la proximité des terres émergées, série d'Amêndoa (800 m), au Portugal.

La partie méridionale de l'Ossa Morena est caractérisée par un Ordovicien qui ne rapelle en rien celui des régions précédentes (Robardet, 1976):

- absence des grès "armoricains" et des "schistes à Calymènes";

- Ordovicien supérieur avec pôle carbonaté, dont la faune trilobitique très particulière a des affinités avec des populations de Pologne et du Kazakhstan.

B. Corrélations

Depuis la seconde moitié du siècle dernier, il était usuel de comparer l'Ordovicien ibérique et armoricain, soit globalement, soit formation à formation, membre à membre (p. ex. Henry *et al.*, 1968; Chauvel *et al.*, 1969; Carré *et al.*, 1970; Henry *et al.*, 1971) (Ann. 9).

Récemment, des essais de corrélations ont envisagé la comparaison de séries entières et en ont montré l'identité rigoureuse, parfois terme à terme (Tamain *et al.*, 1969; Tamain et Ovtracht, 1971a et b; Tamain, 1971 et 1972; Henry *et al.*, 1974): principalement entre la Zone luso-orétane (séries-types d'El Centenillo et de Buçaco) et le Synclinorium médian (série-type de Crozon - Cap de la Chèvre), et aussi entre la frange SW de la Galice (Valongo) (Ann. 10) et les Synclinaux du Sud de Rennes (Martigné-Ferchaud) (Ann. 9).

Par contre, l'Ordovicien de l'Ossa Morena méridionale semble bien ne pas avoir d'équivalent dans le domaine normano - nord-breton.

C. Paléogéographie

On est ainsi amené à reconnaître l'appartenance de la Zone luso-orétane et du Synclinorium médian, de la Galice sud-occidentale et du Synclinorium de Martigné-Ferchaud, etc. à autant d'unités paléogéographiques et structurales de l'Arc ibéro-armoricain, à l'Ordovicien.

Reste toutefois posé le problème du prolongement de l'Ossa Morena méridionale dans la moitié nord dudit arc.

D'autre part, l'examen de la carte de répartition des affleurements ordoviciens (fig. 23.5, b) suggère quelques remarques.

En Ossa Morena, l'aire ordovicienne méridionale s'est installée presque dans l'axe de l'aire cambrienne.

La structuration varisque de la Zone luso-orétane, dans sa moitié orientale, révèle une disposition en bandes alternativement anticlinales et synclinales qui, si l'on tient compte d'une variation latérale progressive des bio-lithofaciès et des puissances du SSW au NNE, pourrait être héritée d'une configuration paléogéographique ordovicienne avivée par le jeu de fractures plus ou moins profondes (dont certaines auraient pu guider la moitiée d'émissions volcaniques). Dans le même ordre d'idées, d'autres arguments suggéreraient une variation latérale comparable de l'ESE à l'WNW, la grande plaine de la Serena (Estrémadure) et son prolongement WNW ayant pu alors constituer comme un seuil.

4.3. La glaciation et l'épirogenèse taconiques

A. Le glaciaire continental d'El Centenillo

A El Centenillo, en Sierra Morena orientale, le Silurien transgressif (Tamain, 1964) sous son faciès typique de schistes fins noirs, à Monograptidés du Llandovérien moyen-supérieur, repose sur l'Ordovicien selon une surface de ravinement glaciaire, moutonnée et bosselée, localement cannelée, striée et montrant des figures de broutage ("crescentic gouges", "chatter marks", "crescentic fractures") qui indiquent un sens général du mouvement des glaces du Nord vers le Sud (Arbey et Tamain, 1971).

Cet épisode continental, témoin de la présence d'une masse glacée ayant vigoureusement buriné son substratum rocheux, est postérieur à l'Ashgillien inférieur et antérieur à la zone graptolitique 20 de Elles - Wood.

B. Les comparaisons et corrélations

La signification et l'importance de cette découverte se voient encore soulignées par l'existence, à des niveaux stratigraphiques identiques ou proches et dans des régions plus ou moins éloignées: d'autres surfaces de ravinement glaciaire (Ann. 11); de dépôts glaciaires, ou glacio-marins (Ann. 12); de discordances (de ravinement) vers la limite Ordovicien - Silurien (Ann. 13); de la transgressivité du Silurien.

On peut formuler un certain nombre d'hypothèses, de plusieurs ordres et intéressant des formations et des anomalies observées dans une grande partie des Hespérides.

Dans la Zone cantabrique (bassins supérieurs des rios Bernesga, Torio, Esla: N. Léon), les schistes ampéliteux et plus ou moins riches en pyrite du Llandovérien moyen-supérieur reposent sur un Ordovicien quartzitique assez inférieur selon une surface de "pseudo-transgression" (Comte, 1959), assimilable aux surfaces de ravinement d'El Centenillo.

Dans la Zone léono - ouest-asturienne (cluse du Sil, Anticlinal de Vegadeo - Fonsagrada, Mondoñedo, Anticlinal de Teleño, Sierra de Caurel), les schistes noirs à nodules, ampélites et Monograptidés siluriens reposent sur différents termes de la série ordovicienne, voire sur le Cambrien (Matte, 1968). Dans la Sierra de Caurel en particulier (Matte, 1968; Guillou, 1969), la transgressivité du Silurien est soulignée par une authentique surface de ravinement traçable sur quelque 100 km.

Dans la partie occidentale de la Zone luso-orétane, le Silurien à Monograptidés reposerait en discordance soit sur les "Quartzites culminants", soit sur les "Schistes culminants" (sous-jacents), d'âge valentien, de la Sierra de Buçaco (Delgado, 1908; Teixeira, 1955).

Enfin, dans la frange SW de la Galice, dans le Synclinal de Valongo, les "grauwackes de la Sierra de Murta" reposent en discordance directement sur les "schistes à *Uralichas ribeiroi*" du Llandeilien.

Dans la Zone léono - ouest-asturienne, toujours dans la Sierra de Caurel, les "schistes intermédiaires", sous-jacents au grès valentiens et sus-jacents aux calcaires ashgilliens, contiennent des passées siliceuses, gréseuses et conglomératiques (à galets calcaires). Dans la Sierra de la Peña Redonda, à peu près au même niveau, des schistes mauves et des pélites jaunes emballent des lentilles conglomératiques à micro-conglomératiques (à grains de quartz subanguleux à arrondis, galets quartzeux) (Guillou, 1969).

En Sierra Morena occidentale, des conglomérats seraient associés à des quartzites blancs, des schistes noirs et des argiles vertes, en-dessous des ampélites siluriennes (Bard, 1969).

Enfin, à l'Est des Hespérides, dans les Chaînes ibériques, les "schistes d'Orea" contiennent des galets de quartzite et de calcaire (ashgillien) et seraient comparables aux niveaux glacio-marins à galets et blocaux des "Lederschiefer" de Thuringe (Greiling, 1967).

C. Conclusion

Ces hypothèses permettent d'avoir une idée préliminaire sur la répartition des différents témoins, connus ou supposés, de cette glaciation fini-ordovicienne dans les Hespérides et alentour (fig. 23.5, D; fig. 23.7).

En raison des grandes analogies lithofaciales existant entre l'Ordovicien supérieur - final du Sud-Est hespérique et celui de la Meseta marocaine (Anticlinorium occidental, Rehamna orientaux, et Jbilete orientales), il n'est pas impossible que des occurrences argilo-pélitiques micro-conglomératiques soient un jour trouvées dans la Zone luso-orétane.

La question se pose aussi de savoir si les "Quartzites supérieurs" d'El Centenillo, qui terminent le cycle ordovicien est-marianique, ne seraient pas, dans leur partie inférieure, d'anciens sables englacés?

Il y a peut-être, en effet, deux épisodes glaciaires: l'un, datant de l'Ashgillien supérieur, et voyant se déposer des sédiments glacio-marins; l'autre, plus récent, voyant des glaces continentales buriner, sculpter et polir leur substratum rocheux.

Quoi qu'il en soit, ce dernier épisode témoigne bien de la réalité de l'épirogenèse taconique dans les Hespérides, avec émersion de terres tant au Nord qu'au Sud-Est (Tamain, 1964 et 1972).

Et c'est précisément sur ces terres que s'étendra la mer silurienne transgressive.

4.4. Le magmatisme taconique

D'après les seules observations de terrain, on ne peut jusqu'à présent attribuer un âge taconique indiscutable à aucune métamorphite ni plutonite. On ne dispose pour ce faire que de datations d'âges absolus.

Un premier groupe de mesures radiométriques (Ann. 14) a été réalisé sur des roches de la Galice occidentale et moyenne.

Une datation préliminaire, pourtant sur 2 gneiss granitiques envahissant des métasédiments précambriens de la région de Vigo, a fourni un âge de 500-475 Ma (Priem *et al.*, 1966).

Une nouvelle série de datations systématiques porta ensuite sur 9 gneiss granitiques du graben blastomylonitique (Malpica - Noya - Vigo), sur 6 du complexe orthogneissique précambrien de la bordure orientale du bassin d'Ordenes, et sur 2 autres du complexe de Lalin: l'âge moyen, d'ailleurs assez homogène, fut de 429 ± 8 Ma (Priem *et al.*, 1970).

Un deuxième groupe de mesures (Ann. 14) a été effectué sur un ensemble de roches plutoniques acides gneissifiées de la région d'Elvas - Portalegre (Haut Alentejo, Portugal).

Cette série porta sur une demi-douzaine de granites alcalin et hyperalcalin à aegyrine, et de syénites à biotite et à néphéline - sodalite; l'âge moyen fut de 466 ± 12 Ma (Priem *et al.*, 1970) (Ann. 15).

Doit-on considérer que tous ces gneiss granitiques de Galice, dont certains semblent bien appartenir à des complexes précambriens, sont d'âge taconique?

D'autre part, l'existence même du magmatisme hyperalcalin anorogénique du Haut Alentejo ne trahirait-elle pas l'existence d'une aire continentale proche et de fractures profondes? (Ann. 16).

Enfin, dans le cas du granitoïde du Nord de Villaharta (Cordoue) qui scelle des métamorphites cadomiennes et qui est daté de 452 ± 22 Ma (Bellon et Rides, 1977), doit-on le considérer comme une manifestation effectivement taconique et anorogénique? Ou bien, plutôt, n'y voir que la réactivation d'un magma plus ancien (cadomien) et dont l'âge aurait été rajeuni lors des processus épirogéniques taconiques, en une zone tectoniquement sensible du Sud-Est hespérique?

4.5 L'Etape siluro-éodévonienne

A. *L'évolution paléogéographique*

La transgression de la mer silurienne à Graptolites, aux eaux calmes, sans courant, peu aérées et en voie de réchauffement, progresse lentement sur les reliefs ordoviciens antérieurement sculptés par les glaces.

Elle débute au Llandovérien inférieur en Ossa Morena sud-orientale (Robardet, 1976), au Llandovérien inférieur - moyen dans la Zone léono - ouest-asturienne (au moins, en partie) (Matte, 1968), au Llandovérien moyen-supérieur dans les zones luso-orétane (Haberfelner, 1931; Tamain, 1972) et cantabrique *p.p.*(Comte, 1959) (fig. 23.5, D).

Puis, se développe un épisode volcano-sédimentaire (à cinérites, lydiennes et schistes siliceux) assez généralisé: en Ossa Morena méridionale, dans la Zone luso-orétane, en Galice (Valongo, Serra de Marâo, NE. Trans-os-Montes).

Pendant la deuxième moitié du Silurien, la sédimentation schisteuse fine est progressivement remplacée par des dépôts de plus en plus gréseux, avec développement localisé d'occurrences carbonatées.

La limite Silurien - Dévonien est caractérisée, à l'échelle des Hespérides, par la continuité de la sédimentation marine.

Le début du Dévonien voit la division des Hespérides en deux grands ensembles (Llopis Llado *et al.*, 1967): l'un septentrional, regroupant les zones léono -ouest-asturienne et cantabrique, et se poursuivant jusqu'aux Pyrénées et la Catalogne; l'autre centro-méridional, et s'étendant de la frange SW de la Galice jusqu'au Guadalquivir actuel (fig. 23.5, D).

Le premier ensemble est caractérisé par une sédimentation de faciès bohémien (= hercynien), se développant sur une aire affectée seulement par quelques pulsations tardi-calédoniennes épirogéniques (Ann. 17) (Brouwer, 1967b).

Il comprend (Brouwer, 1967a):

-le bassin léono-asturien, où alternent les dépôts clastiques et carbonatés d'une mer peu profonde, chaude et bien oxygénée, à la riche faune benthonique;

-le "Seuil asturien" (fig. 23.5, D: SA), aire de plateforme relativement stable, caractérisée soit par le non-dépôt de la quasi-totalité du Dévonien, soit par une sédimentation réduite suivie d'une dénudation sous-marine; limité au Sud par la "Ligne de Léon" (E-W) (fig. 23.5, D: LL);

-le bassin palencien, où alternent les dépôts carbonatés fins et schisteux d'une mer fermée, aux eaux tranquilles et à la faune rare, peu diversifiée et principalement pélagique; limité au Sud par la "Ride de Santibañez" (= de Riaño - Cervera de Pisuerga: WNW-ESE).

Le second ensemble est caractérisé par une sédimentation terrigène clastique, de faciès rhénan, intéressant des domaines extrêmement étroits fréquemment situés dans l'axe des domaines siluriens antérieurs: synclinaux de Valongo, Almaden, Hautes Sierras méridionales (Guadalmez), Portalegre - Albuquerque, Nord de Peñarroya - Pueblo Nuevo, Barrancos (Ann.18), (Llopis Llado *et al.*, 1967; Teixeira et Thadeu, 1967). Des particularismes régionaux apparaîtront sans doute, quand le Dévonien en sera connu plus en détail.

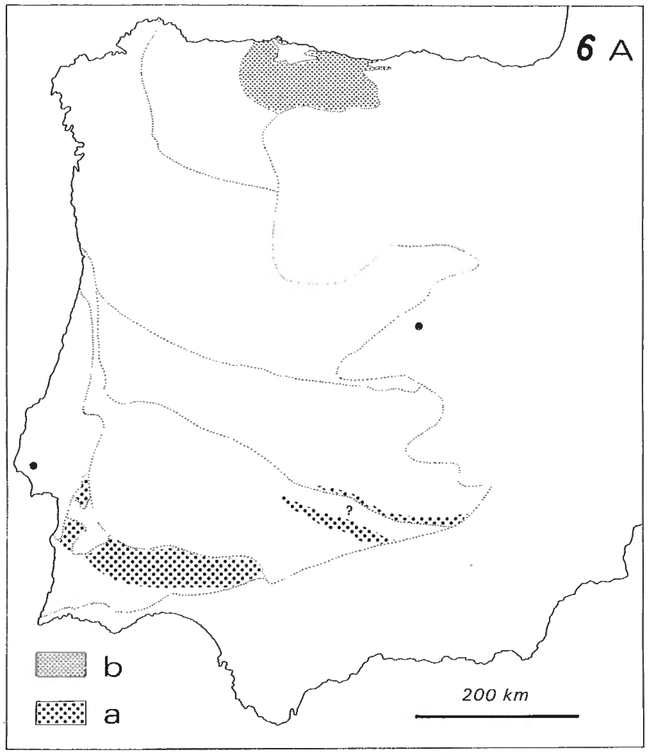
B. *L'évolution structurale*

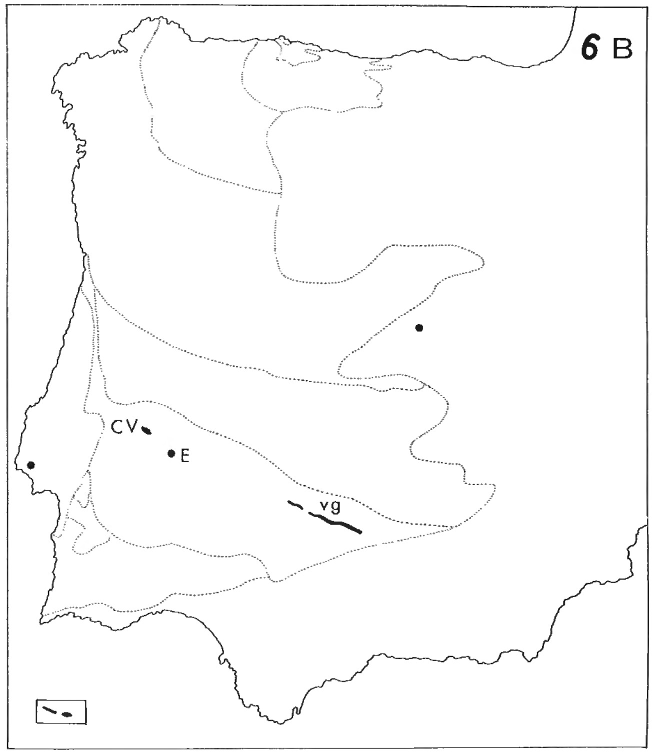
L'étape siluro-éodévonienne voit ainsi se développer un premier mouvement transgressif généralisé pendant tout le Llandovérien, puis, après inversion, un second mouvement généralisé, régressif.

Ce dernier mouvement s'achève par une phase épirogénique importante: d'âge reussien (pendant le Dévonien moyen, en totalité ou en partie) (Ann.19), dans l'ensemble centro-méridional; et d'âge néodévonien (vers la limite Frasnien - Famennien) dans l'ensemble septentrional.

Ces deux grandes aires présentent donc, dès le début du Dévonien, une évolution différente de part et d'autre du bloc galaïco-carpétan (fig.. 23.5, D): une aire relativement stable au Nord, avec déjà un certain retard dans la mobilisation qui se prépare par rapport à une aire déjà plus mobile au Sud, où les bassins de sédimentation sont de plus en plus réduits depuis l'Ordovicien - coincés qu'ils semblent être entre les terres émergées: massif d'Evora (fig. 23.5, D), etc. - et où sont déjà inscrites les grandes lignes directionnelles de la structuration varisque.

Enfin, la fracturation profonde, de direction E-W, est, comme déjà au Cambrien, bien représentée dans l'aire septentrionale, avec la toujours active "Ligne du Léon".





A. -- A LA FIN DU DEVONIEN.

a : aires à Famennien supérieur - terminal, du Sud ; -- b : aire à Famennien supérieur transgressif, du Nord (z. cantabrique).

B. - C. -- AU TOURNAISIEN.

B : Les hyperbasites.

vg : du Varas - Guadalbarbo ; -- CV : de Cabezo de Vide.
E : Elvas.

C : Les basites.

vg : du Varas - Guadalbarbo ; -- CM : de Campo Maior ; -- CV : de Cabezo de Vide.
B : Badajoz ; -- E : Elvas.

D. -- AU TOURNAISIEN.

a : amas sulfurés de la Ceinture pyriteuse (CP) du Sud-Ouest hespérique ; -- b : flore continentale du bassin houiller de Val-de-Infierno (V).

E. -- AU TOURNAISIEN - VISEEN INFERIEUR.

a : aires de sédimentation volcano-siliceuse à siliceuse ; -- Mn : occurrences de Mn à Mn-Fe ; -- b : calcaires.

BC : bassin cantabrique ; -- CP : Ceinture pyriteuse ; -- SMO : Sierra Morena orientale.

F. -- AU VISEEN SUPERIEUR.

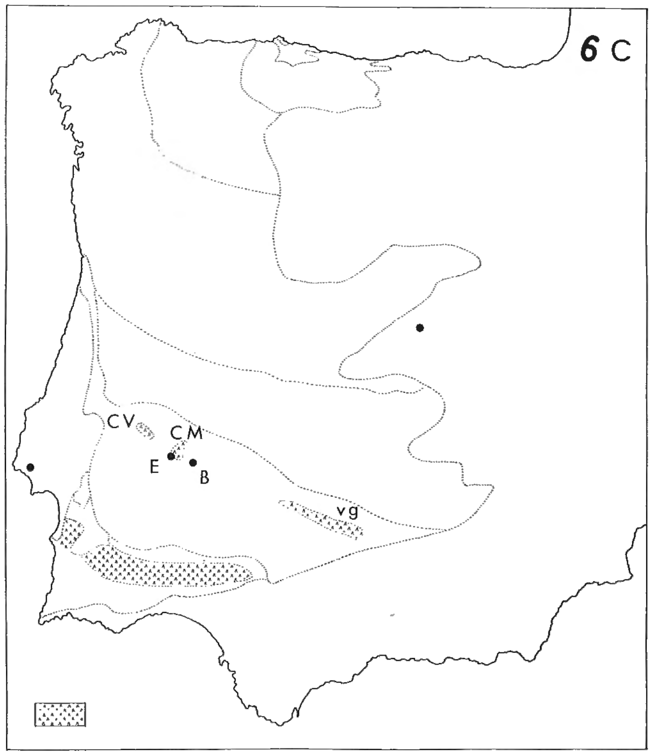
a : aires à sédimentation flysch (schisto-gréso-grauwackeux), de type Culm ; b : marge à sédimentation flysch (a) tardive ; -- c : aires à sédimentation carbonatée.

OMO : Ossa Morena orientale ; -- EA : bassin est-asturien ; -- NL : bassin nord-léonais.

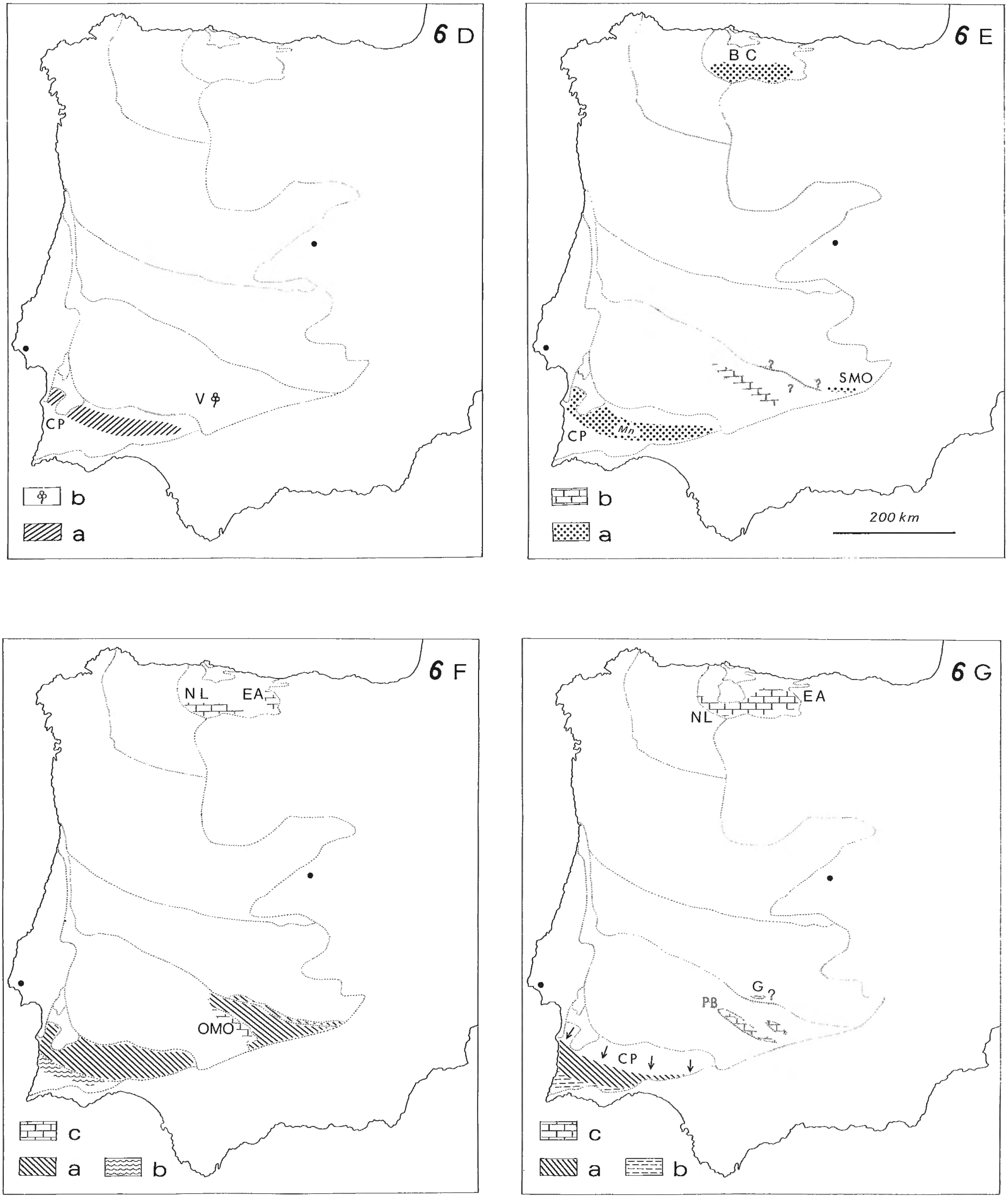
G. -- AU NAMURIEN.

a : aire à sédimentation flysch, de type Culm ; -- b : marge à sédimentation flysch (a) tardive ; -- c : aires à sédimentation carbonatée.
Les flèches indiquent le sens de la mobilisation dans le SW.

CP : Ceinture pyriteuse ; -- PB : bassin de Peñarroya - Belmez (à calcaires du Namurien basal) ; -- G : Guadalmez ; -- EA : domaine est-asturien ; -- NL : domaine nord-léonais.



Figures 23.6 A-C LA MOBILISATION VARISQUE (partie)



Figures 23.6 D-G LA MOBILISATION VARISQUE (fin)

TABLEAU 24.4.

LE DEVONIEN SUPERIEUR ET LE CARBONIFERE INFERIEUR DU SUD DES HESPERIDES

Echelle stratigraphique :		CEINTURE PYRITEUSE (SW HESPERIQUE)	SIERRA MORENA ORIENTALE (SE HESPERIQUE)	
NAMURIEN inférieur		"<u>CULM</u>"	"<u>COMPLEXE DE FACIES FLYSCH DE TYPE CULM</u>"	F O R M A T I O N C A M P A N A
VISEEN supérieur		à rares conglomérats	à niveaux de plantes à conglomérats intraformationnels lenticulaires (à galets de lydienne)	
VISEEN inférieur		"<u>COMPLEXE VOLCANO-SILICEUX</u>" à calcaires lenticulaires de Sotiel (à Conodontes) à jaspes, radiolarites, et occurrences de Mn-Fe (mine "El Teniente") à tufs, kératophyres	"<u>SCHISTES A NIVEAUX SILICEUX</u>" à radiolarites à témoins microscopiques d'une lointaine activité volcanique explosive, acide	
TOURNAISIEN		à amas sulfurés à basites, spilites, pillow-lavas		
STRUNIEN	Tn1a	"<u>GROUPE DES PHYLLITES-QUARTZITES</u>"	"<u>QUARTZITES CAMPANA</u>"	"<u>Schistes Aquisgrana</u>"
FAMENNIEN	Fa2		"<u>SCHISTES SATINES INFERIEURS</u>"	
FAMENNIEN	Fa1	?		?
FRASNIEN		?	Complexe à dominante gréseuse	

C. *Les manifestations magmatiques éventuellement tardives à posthumes*

Des mesures radiométriques effectuées sur des granites gneissiques à deux micas, très déformés, de la Galice occidentale (Lage, Carril et Cabo Silleiro) ont donné des âges absolus de 369, 352, 343 et 330 Ma, soit un âge moyen de 349 ± 10 Ma (Priem *et al.*, 1970): ce qui, dans une interprétation littérale, correspondrait - pour la première citée au moins - à une manifestation tardi-reussienne (Ann. 20) plutôt que varisque précoce.

5. LA MOBILISATION VARISQUE

Elle se développe d'une façon très différente selon que l'on se trouve au Nord ou au Sud du bloc galaïco-carpétan.

5.1. Dans les domaines méridional et sud-portugais

A. *L'Etape préliminaire à précoce (fig. 23.6, A-E)*

Elle s'étend du Frasnien au Viséen inférieur inclus, et se manifeste essentiellement dans la partie sud de la Zone luso-orétane, la frange nord de l'Ossa Morena, la Ceinture pyriteuse et le SW hespérique.

Au Dévonien supérieur *(fig. 23.6, A)*

Une première sédimentation, détritique et parfois assez grossière, s'installe dans le Sud-Est (Sierra Morena extrême-orientale à orientale), au Frasnien.

Elle évolue pendant le Famennien, toujours dans le SE et aussi dans la Ceinture pyriteuse.

Elle passe localement, au Famennien supérieur - terminal, à une sédimentation calme, peu oxygénée et apparemment d'une relative profondeur de dépôt, ainsi que le révèle l'apparition de faciès lutitiques à Ostracodes, Lamellibranches à test fin et Orthocères à l'Est de Los Pedroches (Charpentier *et al.*, 1976); puis, elle se charge en nouveaux apports détritiques grossiers.

Nous avons là, avec l'enfoncement progressif du fond sous-marin, l'apparition de dépressions éphémères et une instabilité tectonique croissante, les premiers témoins de l'organisation de la partie hespérique de la Zone mobile varisque.

Au Tournaisien et Viséen inférieur (fig. 23.6, B-E)

Le premier évènement notable est la mise en place du "complexe ophiolitique du Varas - Guadalbarbo" (Crousilles *et al.*, 1976) selon un train de fractures profondes de direction générale déjà varisque, à l'intérieur du "Linéament de Cordoue", juste à la limite des zones luso-orétane et ossa-marianique. On le retrouve au Portugal, à Cabezo de Vide (fig. 23.6, B et C).

Il s'agit d'une association ophiolitique incomplète, à harzburgites complètement serpentinisées (à lizardite et orthoserpentine à 6 feuillets, kotschubéite et kammérérite, et inclusions de chromite, magnétite, pentlandite), gabbros rodingitisés et gabbro noritique pegmatoïdique (Ann. 21) chevelu filonien doléritique, et cortège à spilites, pillow-lavas.

Des basites de ce dernier cortège ont été datées de: 348 ± 18 Ma et 334 ± 16 Ma (Bellon et Rides, 1977).

Dans la Ceinture pyriteuse, au même moment, se met en place un magmatisme bimodal, basique (à spilites, pillow-lavas) (fig. 23.6, C) à acide (à kératophyres et quartz-kératophyres, tufs) qui donne lieu à un complexe volcano-siliceux particulièrement bien développé: à amas sulfurés (fig. 23.6, D), puis occurrences de Mn ou Mn-Fe associées à des jaspes, radiolarites et lydiennes (fig. 23.6, E), au Viséen inférieur.

Un épisode carbonaté se développe momentanément: calcaire, à fragments de Radiolaires, de Campillo de Llerena et du N de Villanueva del Rey (Ortuño, 1970), à l'emplacement du futur bassin houiller de Peñarroya - Belmez, au Tournaisien et/ou Viséen inférieur (fig. 23.6, E); calcaire de Sotiel, à Conodontes du Viséen inférieur de la Ceinture pyriteuse.

En outre, on sait de façon indiscutable - avec la flore houillère du petit bassin de Val-de-Infierno (fig. 23.6, D: V), datée du Tournaisien supérieur (plutôt que du Viséen inférieur) (Wagner, 1976) - que la frange NE de l'Ossa Morena était, au moins partiellement, émergée à cette époque.

B. *L'Etape moyenne (fig. 23.6, F et G)*

Elle s'étend du Viséen supérieur au Namurien basal inclus, dans le Sud-Est, et perdure jusqu'à un Namurien assez élevé (voire au Westphalien inférieur ?) dans le Sud-Ouest.

Elle voit l'installation, dans ces régions, d'une sédimentation schisto -gréso - grauwackeuse: flysch, de type Culm.

Une épisode carbonaté se développe momentanément et très localement, dans le Sud-Est, au Viséen supérieur puis terminal et au Namurien basal (fig. 23.6, G), et révèle une surrection des bordures des sillons à flysch, ainsi rétrécis.

Elle se termine par la phase majeure du plissement varisque, qui se produit donc après le Namurien basal - inférieur et avant le Westphalien B (Villanueva del Rio y Minas, Peñarroya - Belmez), dans la plus grande partie de la moitié méridionale des Hespérides.

5.2. Dans le domaine septentrional

Il s'agit essentiellement de la Zone cantabrique (fig. 23.6).

S'y développe par endroits, à la suite de la transgression généralisée du Famennien supérieur, une première sédimentation gréso-calcaire famenno-strunienne à struno-éotournaisienne.

Une lacune plus ou moins importante traduit, ensuite, la réalité de mouvements précoces, notamment dans les confins orientaux et selon la "Ligne du Léon".

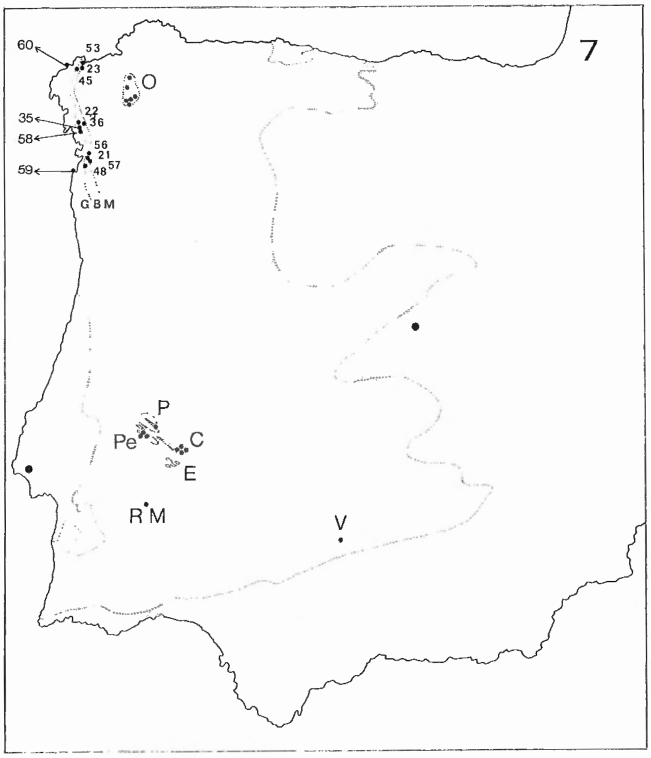
La sédimentation reprend, ici au Tournaisien supérieur et d'abord schisteuse, là seulement au Viséen avec une nette dominante carbonatée et se poursuivant pendant le Namurien (en totalité, ou quasi) voire au-delà (jusqu'au Westphalien A: San Emiliano): cette sédimentation, interprétée localement comme un flysch calcaire, s'oppose ainsi à celle du Sud hespérique, schisto-gréso-grauwackeuse en aire tectoniquement très instable.

Ce cycle sédimentaire se termine avec la phase palencienne, première grande phase plicative varisque connue ici: antérieure au Westphalien B inférieur dans le Nord et Nord-Est du Léon, elle est rapportée au Westphalien B supérieur au Nord de Palencia.

5.3. Dans le bloc galaïco - carpetan

Dans ce bloc, qui correspond à la Zone galaïco-castillane et à la plus grande partie de la Zone léono - ouest-asturienne, on ne connaît aujourd'hui aucun dépôt méso- à néodévonien, dinantien et namurien; seuls y sont connus quelques dépôts de Westphalien D et de Stéphanien C.

Par contre, c'est dans cette grande aire géanticlinale sur socle précambrien cristallin que le métamorphisme, la migmatisation et le plutonisme varisques seront les plus intenses.



C : Cevadais. E : Elvas. O : bassin d'Ordenes. P : Portalegre. Pe : Pedroso. RM : Reguendos de Monsaraz. V : Villaharta.

21, 22, 23, 36, 45, 48, 53, 56 et 57 : échantillonage de H.N.A. Priem *et al.*, pour le Graben blastomylonitique (GBM) de Galice.

35, 58, 59 et 60 : granites gneissiques de Lage, Carril et Cabo Silleiro, d'âge reussien tardif à posthume, voire varisque.

Figure 23.7. AGES RADIOMETRIQUES "TACONIQUES" ET "TARDI-REUSSIENS"

IV. CONCLUSION

Les Hespérides ont été le siège, à plus de 300 Ma d'intervalle, de deux mobilisations importantes: l'une cadomienne et l'autre varisque, dont les évolutions sont fort comparables.

La mobilisation varisque, sur les caractères tectorogéniques de laquelle nous n'insisterons pas ici, a été précédée d'une longue préparation qui, dans le segment mauritano-scandinave, a abouti à l'orogenèse calédonienne.

Ici, cette préparation proto-varisque a avorté bien qu'elle portât déjà inscrites certaines grandes caractéristiques structurales varisques. D'autre part, le premier cycle la constituant voit sa paléogéographie et sa sédimentation partiellement guidées par le jeu - ou mieux, le rejeu - d'une fracturation E-W de carctère profond, apparemment héritée de la structuration cadomienne. On ne connaît encore rien des implications morphologiques et structurales des mouvements épirogéniques sardes. Par contre, les bassins de sédimentation ordoviciens et siluro-éodévoniens sont, eux, déjà orientés - semble-t-il - selon ce qui sera la direction varisque majeure.

Les temps taconiques sont marqués à la fois par une nouvelle crise épirogénique et une glaciation, peut-être polyphasée, bien connue par ailleurs dans le Sud et tout le Nord-Ouest africain et dont on commence à trouver des témoins de plus en plus nombreux en Europe.

En outre, la fracturation profonde joue un rôle prépondérant dans ces diverses structurations: cambrienne (avec les axes E-W des bassins de Tolède, de Salamanque); cambrienne et ordovicienne (avec les axes arqués du Navia, puis du Narcea qui, lui, deviendra la limite des zones cantabrique et léono - ouest-asturienne, là précisément où s'enracineront les nappes cantabriques varisques); etc. La permanence de cette fracturation est bien illustrée par la "Ligne du Léon", extraordinairement mobile pendant le Paléozoïque.

En conclusion, cette préparation proto-varisque, ou calédonienne, constitue l'originalité non seulement des Hespérides mais aussi du restant des Variscides ouest-européennes (Massif armoricain).

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VI. ANNOTATIONS

(1): chromitites contenant jusqu'à 12 g Pt/t. (Cotelo Neiva, 1948).

(2): chalcopyrite (dominante) et pyrite dans des amphibolites (à anthophyllite et grenat): gîte de Cerdido Moeche.

(3): chalcopyrite - pyrite - pyrrhotite dans des micaschistes (à amphibole et grenat) associés à des amphibolites correspondant à un volcanisme basaltique (tholéitique et hyperalumineux) d'origine cf. circum-océanique: gîtes de Fuente Rosas (Chabod *et al.*, 1976), Arinteiro, Fornas, près de Santiago de Compostela.

(4): cf. tillitiques (Lotze, 1956 et 1961; Bouyx, 1969 et 1970; Arbey *et al.*, 1975), remaniant des galets notamment de phtanite; par endroits, aurifères (jusqu'à 16 g Au/t. de préconcentré) dans la Zone luso-orétane (Crespo Lara, 1972), comme le "Poudingue de Gourin" faiblement aurifère à Roudouallec (domaine normano - nord-breton, France) (Chauris et Guiges, 1969).

(5): calcaires dolomitiques, généralement ferrugineux, quelquefois bréchiques et souvent oolithiques: SW de l'Ossa Morena, Zone luso-orétane (Cabezarrubias, Hinojosas de Calatrava, vallée du rio Tirteafuera, et en haute Estémadure: vallée de l'Ibor, environs de Guadalupe).

(6): les grandes et moyennes structures s'allongeant NNW-SSE dans la moitié orientale de la Zone luso-orétane, apparemment aussi de l'Ossa Morena, et selon une direction sub-méridienne dans le Nord-Ouest hespérique.

(7): Dans le Massif armoricain (France), Cogné (1974) considère que le stade molassique terminant l'évolution cadomienne est représenté par les dépôts du Cambrien inférieur et moyen *p.p.*

(8): volcanisme moyennement acide de la Sierra Morena orientale (SE hespérique), basique (à diabases et spilites) de la Sierra de Buçaco (E de Coimbra, Portugal), et basique du Cabo Peñas (bordure W de la Zone cantabrique), à l'Ashgillien inférieur.

(9): ...comparaisons parfois aussi, d'ordre purement lithofacial: telles les occurrences de fer oolithique de Galice (fig. 23.5, B) et d'Anjou (W France), de l'Ordovicien respectivement moyen et inférieur.

(10): dont le prolongement SE idéal passerait par Tolède, ou légèrement plus au Nord.

(11): -surface de ravinement tourmentée, constituée par des collines de psammite et des falaises de grès, avec des profondes vallées entaillées dans le substratum localement cannelé,... dans le Zemmour (Mauritanie) (Sougy, 1956 et 1969; Sougy *et al.*, 1963);

-paléovallées (10 km x 2 km x 400 m) en forme d'auge, à verrou glaciaire et à flancs très redressés, aux surfaces striées et polies, avec alignements de roches moutonnées et striées, figures d'arrachement en croissant,... sur tout le pourtour du Hoggar (Sahara) (Gabriel *et al.*, 1965; Debyser *et al.*, 1965; Rognon *et al.*, 1968);

-planchers glaciaires à stries et figures d'arrachement en croissant, orientées N-S à NNW-SSE,... de l'Anti-Atlas (S Maroc), où l'épisode glaciaire est daté avec précision: postérieur à un certain Ashgillien supérieur, et antérieur à un certain Llandovérien inférieur (Destombes, 1968a et b); etc.

(12): a. - "Série de Garat-el-Hamoueid": grès-quartzites à grains ronds mats, galets (polyédriques, à arêtes émoussées, angles rentrants et faces concaves) striés et impressionnés (à cannelures, cupules et cassures parfois bipolaires), et blocs erratiques,... du Zemmour (Sougy, 1956; Sougy *et al.*, 1963); - grès tendres non stratifiés, micro-conglomératiques, et "grès moutarde conglomératiques" mal classés, à blocs de grès tordus et "plissés", à galets (avec traces en coup de gouge),... de Mejeria (Dia *et al.*, 1969), dans le Taganet; - grès argileux jaunes, à argiles micro-conglomératiques et à conglomérats (avec galets éolisés, striés, et à faces d'usure),... d'Aoucert (Bronner *et al.*, 1969);

-conglomérats à blocs, avec slumpings et varves,... de la Gara Sayada (Gevin, 1966 et 1968), tillite à blocs allogènes énormes (jusqu'à 150 m^3), varves,... de la Gara Assaba (Gevin *et al.*, 1968)... dans l'Est du bassin de Taoudeni;

-conglomérats polygéniques, grès plus ou moins micro-conglomératiques, brèches à blocs remaniés, avec slumpings, stratifications anarchiques, argile rouge à gravillons quartzeux,... et dépôts de déglaciation, varves,... du Mouydir (Chanut *et al.*, 1958; Borocco *et al.*, 1959; Rognon *et al.*, 1968), argiles jaunes à grains de quartz,... du bassin de Fort-Polignac (Corriger *et al.*, 1963), argiles à gros grains de quartz arrondis, blocs anguleux,... du secteur de Fort-Gardel (Borocco *et al.*, 1959),... dans le Tassili N'Ajjer; - grès fins et grès argileux à gros grains de quartz épars de l'Aïr (Joulia, 1959);

-argiles finement gréseuses à micro-conglomératiques de Ben-Zireg, du Teniet Ghenia,... dans les confins algéro-marocains méridionaux (Massa *et al.*, 1965);

-"grès du Ksar d'Ougarta", argileux et micro-conglomératiques, "argiles jaunes d'El Kseib" à graviers épars, et série quartzitique à conglomératique sous-jacente,... de l'Ougarta (NW Sahara): correspondant en partie à des sédiments glaciaires, proglaciaires et périglaciaires ayant subi des déformations glacio-tectoniques, avec gradins, slumpings, fentes en coin,... (Arbey, 1968 et 1970);

-"grès du 2° Bani" à niveaux détritiques grossiers, hétérogranulaires, conglomératiques à galets souvent polyédriques exotiques, et à grès argileux micro-conglomératiques,... de l'Anti-Atlas (S Maroc) (Destombe, 1968a et b);

-argiles gréseuses et micro-conglomératiques, à grains de quartz, surmontant des schistes ashgilliens et surmontées par des quartzites (sous-jacents à un Llandovérien non basal schisteux), du NW du Jbel Mouchchène (Anticlinorium de Khouribga - Oulmès, Meseta marocaine septentrionale); - argiles micro-conglomératiques et quartzites sous-jacents aux ampélites à Monograptidés d'El Mesrane (Rehamna orientaux); - argiles micro- et macro-conglomératiques, et quartzites bleutés (à bleu noir) riches en pyrite de Piton + 725 m, Koudiat Bou Marhara (Jbilete orientales);

-sédiments d'origine glaciaire transportés en partie par icebergs, en partie par glissements sous-marins,... au Fezzan; etc.

b. - pélites à fragments polyédriques (et galets à facettes striées, à cannelures, à cupules,...), situées entre des schistes cf. ashgilliens et des schistes gréseux à fucoïdes cf. valentiens,... du Synclinal d'Urville (Normandie, France) (Dangeard et Doré, 1971), et passant à des argilites feuilletées noires, micro-conglomératiques, à Saint-Germain-sur-Ay (Cotentin); - "Lederschiefer" ashgilliens: schistes argileux, avec conglomérats, pélites à blocaux et galets (parfois énormes) striés,... avec slumpings, glissements sous-aquatiques et traces d'un transport par icebergs... du flanc SE de l'Anticlinal de Schwarzburg (Thuringe, R.D.A.) (Katzung, 1961); - série glacio-marine, au matériau non classé, à petits galets (anguleux et arrondis) de quartz,... de la Nouvelle Zemble (U.R.S.S.) (Miloradowic, 1935); - etc.

(13): discordance (de ravinement glaciaire) dans l'Ougarta, le Mouydir, dans la région d'Ouallène (NW Hoggar) (Dourthe, 1959), au SW de Fort-Flatters (Adrar Tan Elak) (Claracq *et al.*, 1958), dans la région d'In Azaoua (Aïr, SSE Hoggar) (Joulia, 1959a et b), dans l'Est du bassin de Taoudeni (W Sahara), dans l'Anti-Atlas (S Maroc), dans le Zemmour, l'Adrar maure et le Taganet, en Mauritanie,... dans la région de Kayes (Soudan) (Dars *et al.*, 1957),... au Fezzan, etc.

(14): Rb/Sr sur roche totale; Rb/Sr et K/Ar sur micas.

(15): Une première analyse (Rb/Sr) (Mendès, 1968) du granite alcalin gneissique, tectonisé et cataclasé, de Portalegre avait donné un âge de: 358 ± 44 Ma.

(16): jalonnant habituellement, dans de tels contextes, la limite aire continentale - domaine marin.

(17): à l'origine de décharges clastiques (faciès bohémien "impur").

(18): à Psilophytales, Protolycopodiales,... Coralliaires, Crinoïdes, (Teixeira et Thadeu, 1967).

(19): comme c'est le cas dans le SE hespérique (Puschmann, 1967), par exemple.

(20): Rappelons l'âge (Rb/Sr) de la diorite quartzique - dans un cortège de granodiorites à biotite ou à biotite-hornblende - de Reguengos de Monsaraz, dans le massif d'Evora (fig. 23.7: RM): 359 ± 76 Ma (Mendès, 1968).

(21): toutes roches volumétriquement très peu exprimées.

24. BREVE DESCRIPTION DU SEGMENT CALEDONO-HERCYNIEN DU MAROC

André Michard

Institut de Géologie
Université Louis Pasteur
67084 Strasbourg Cedex
France

AVERTISSEMENT

La courte synthèse qui suit a été préparée sur la base du mémoire que le Service Géologique du Maroc vient de publier (Michard, 1976); son texte reprend à peu près celui d'une communication à un colloque, en cours de publication (in Michard et Sougy, 1975). Notre but est simplement, et sant prétendre faire oeuvre nouvelle, de fournir au lecteur du fascicule spécial préparé par le Projet "Caledonian Orogen" une introduction pratique à la géologie du Primaire marocain, largement documentée dans les publications du Service géologique de ce pays.

I-INTRODUCTION: L'EXTENSION DE LA CHAÎNE ET DE SES AFFLEUREMENTS

LES HERCYNIDES EN AFRIQUE DU NORD

Au nord de la plate-forme saharienne, le sous-sol nord-africain porte les marques superposées des orogenèses hercynienne et alpine (fig. 24.1). Cette dernière, qui a érigé les Atlas et les chaînes rifo-telliennes, est la mieux connue. Le bâti hercynien, ou mieux: calédono-hercynien, se laisse moins bien décrire parce qu'il n'affleure que discontinuement et parce qu'il a été disloqué par les mouvements tardi- et post-hercyniens.

Les affleurements nord-africains de terrains primaires sont essentiellement concentrés au Maroc et à ses confins algériens. Vers l'est, les seuls massifs anciens visibles sont les noyaux kabyles, inclus dans l'orogène rifo-tellien; le reste du socle est caché sous la couverture secondaire et tertiaire. On a cependant des raisons de penser que les affleurements occidentaux fournissent une "coupe" à peu près représentative de l'ensemble des Hercynides nord-africaines.

LES AFFLEUREMENTS CALEDONO-HERCYNIENS DU MAROC

On peut les grouper en trois ensembles, suivant les domaines structuraux d'âge alpin dont ils relèvent (fig. 24.1 et Saadi, 1976).

1. Dans le Sud algéro-marocain, plate-forme stable depuis le Trias, les terrains anciens constituent de vastes unités structurales très continues. Ce sont d'abord les chaînes hercyniennes de l'Anti-Atlas et de l'Ougarta, où le socle précambrien affleure en larges "boutonnières" (fig. 24.2). Outre ces chaînes, ce sont des bassins (synéclises) à couches généralement sub-tabulaires: bassins de Tindouf et Reggane à l'ouest, de Béchar (le plus déformé) et Timimoun à l'est, qui font passage à la dalle saharienne, à socle précambrien rigide.
2. Au nord de la ligne sud-atlasique (limite sud approximative des mouvements germanotypes secondaires et tertiaires), le socle ancien n'affleure qu'en massifs discontinus. On trouvera les noms des principaux sur la figure 24.2. Les plus vastes appartiennent à la Meseta marocaine, peu disloquée depuis le Permien: ils constituent les pièces maîtresses pour l'étude des Hercynides maghrebines. Les autres sont inclus dans les chaînes atlasiques: ils sont moins étendus et plus ou moins disloqués par des failles post-hercyniennes.
3. Enfin, tout au nord du Maghreb, des affleurements de socle primaire et probablement précambrien s'égrènent de Ceuta à Annaba. Ce sont les nappes paléozoïques et cristallophylliennes du Rif (Ghomarides et Sebtides) et les noyaux kabyles, ressortissant des zones internes rifo-telliennes. En outre, des témoins du socle des zones externes sont également connus en nappes (Senhadja).

PROBLÈME DES DEPLACEMENTS POST-HERCYNIENS

Des déplacements post-hercyniens d'ampleur considérable et inconnue ont évidemment affecté les derniers éléments cités: les raccorder aux ensembles plus méridionaux est actuellement impossible. De fait, les raccords semblent plus plausibles avec les zones bétiques internes et l'ensemble aurait initialement constitué un bloc continental unique, le bloc d'Alboran des auteurs (Andrieux *et al.*, 1971; Didon *et al.*, 1973).

Les massifs du domaine atlaso-mésétien ont été, certes, moins bousculés depuis la fin de la tectogenèse hercynienne. Ils appartiennent, avec le Sud anti-atlasique, à ce que l'on peut appeler le "Maroc africain", relativement stable par rapport au "Maroc méditerranéen" durant le cycle alpin. Il ne faudrait pas croire pour autant qu'ils sont restés immobiles les uns par rapport aux autres, ni chacun par rapport à la plate-forme saharienne. C'est ainsi que le contraste structural entre l'Anti-Atlas occidental et les massifs anciens situés juste au nord (bloc occidental du massif haut-atlasique, Jbilete, etc.), ainsi que d'autres considérations (Voir III) conduisant à localiser un décrochement tardi-hercynien d'au moins 100 km de rejet dextre au niveau de l'accident du Tizi n'Test (Mattauer *et al.*, 1972). Cet accident a nécessairement un prolongement ou des équivalents orientaux et isole le bâti Anti-Atlas - Ougarta des segments orogéniques situés plus au nord. De plus, des arguments structuraux (caractères tectoniques des Atlas) et paléomagnétiques (divergence entre les paléo-déclinaisons triasiques atlaso-mésétiennes et méridionales) conduisent à envisager des mouvements de rotation et translation durant le Secondaire et le Tertiaire au nord de la ligne sud-atlasique (Michard *et al.*, 1975).

PICG Projet 27, Contribution marocaine nº 1.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

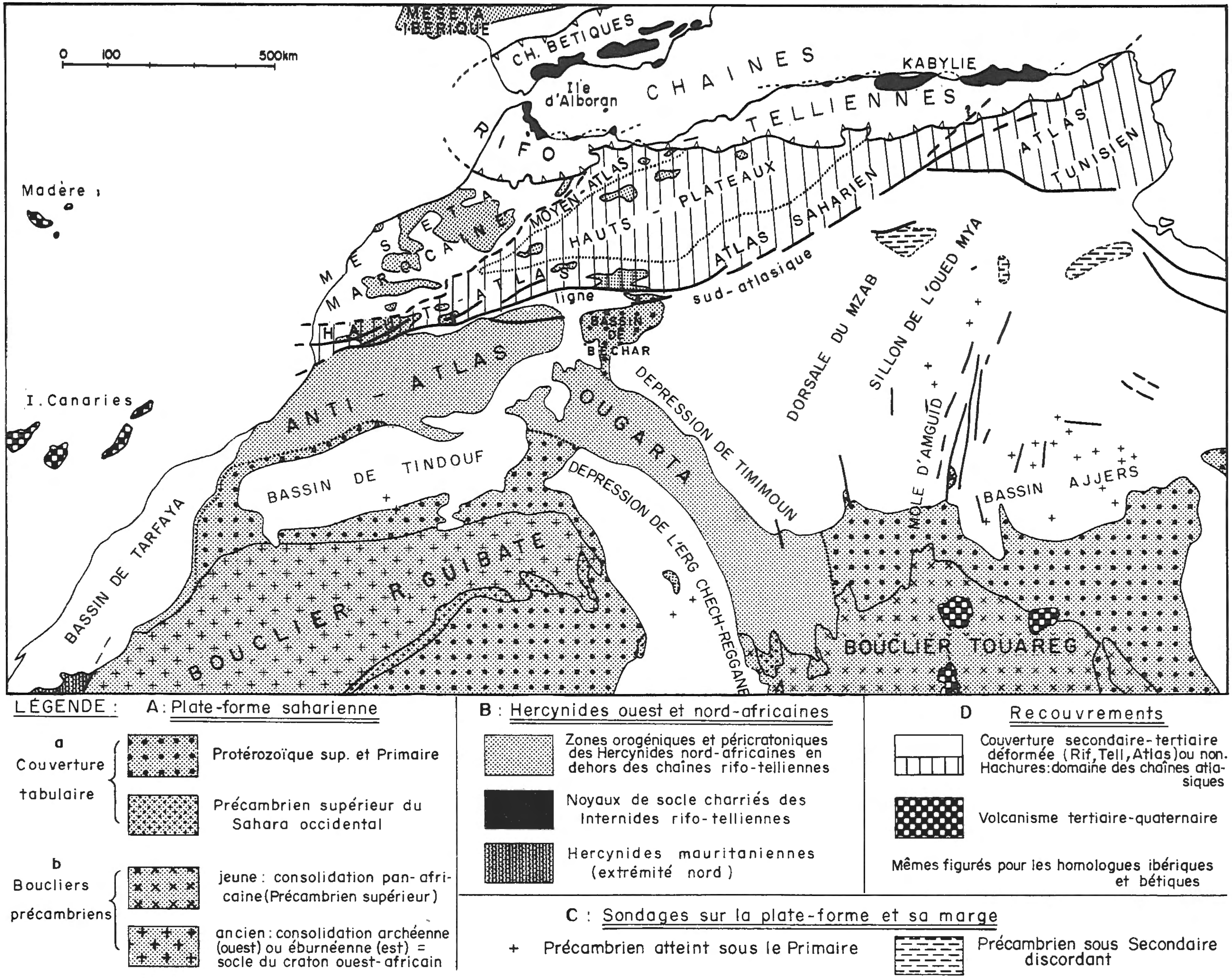


Figure 24.1. Extension des affleurements anté-triasiques en Afrique du Nord et dans le nord-ouest du Sahara; de la plate-forme continentale aux zones orogéniques hercyniennes et rifo-atlasiques.

Tableau 24.1

LES PRINCIPALES FORMATIONS PRECAMBRIENNES ET INFRACAMBRIENNES DISTINGUEES AU MAROC

Domaines / Périodes	ANTI-ATLAS		ATLAS	MESETA	RIF
Base du Cambrien inférieur (« Géorgien ») daté paléontologiquement	*Secteur W :* Calcaires supérieurs de l'Anti-Atlas { Calc. à *Archaeocyathus* / Dolomies	*E :* *Id.* mince, ou lacune, ou faciès détritiques	Schistes pyroclastiques et Calcaires à *Archaeocyathus* Série schisto-calcaire à Trilobites (Ounein).	Calcaires et dolomies (El-Jadida, J. Irhoud des Jbilet, Rehamna occid., Pays Zaïan)	*Nappes cristallophylliennes des Sebtides,* mal datées : micaschistes, amphibolites et gneiss du Filali (anté-Llanvirn : Calédonien ? Eocambrien ?)
Infracambrien supérieur (1) ou (Adoud. sup.)	Série lie de vin	Grès de Tikirt, ou lacune	? Schistes lie de vin,		
« *Adoudounien* » (Adoud. inf.) (Eocambrien ?)	Calcaires (dolomies) inférieurs de l'Anti-Atlas. Intercal. pyroclastiques Volcan andés. d'Alougoum Dolomies de Tamjout Série de base	Lacune ou f. détritiques	dolomies, intercalations volcano-détritiques, andésitiques (Agoundis) dolomies (Tamlelt)		
Infracambrien inférieur (Eocambrien ? Précambrien terminal ?)	*Précambrien III :* Série de Ouarzazate, couches rouges de Tiouine; niveau à *Collenia* d'Amane-n'Tourhat, etc. (série volcano-détritique à ignimbrites, andésites, granites annulaires sub-volcaniques...)		Précambrien III du « promontoire de l'Ouzellarh » (Bou-Ourhioul, O. Rdat...), du pays de Skoura, de Tamlelt et Talzaza (Bou-Kaïs)	Rhyolites d'El-Jadida Couches volcano-détritiques du Pays Zaïan (J. Hadid, Bou-Acila, Goaïda)	
Précambrien supérieur (et moyen ?)	*Précambrien II-III :* Séries de Tidiline, d'Anezi (volcano-détritiques) — *Précambrien II s.s.* (1 et 2 ?) — Granite de Tafraoute (Pr. II-III ?) — Série des quartzites (Lkest) et des calcaires à Oncholites (Tachdamt) — Ophiolites de Bou-Azzer—El-Graara; diorites quartziques.		Granites, Schistes et quartzites du Massif ancien du Haut Atlas (Bloc E : Ouzellarh, J Toubkal, O. Ourika), du Pays de Skoura, de Tamlelt (Aïn-Chaïr)	en pays Zaïan ? (voir aussi formation des granites hercyniens par palingenèse du Précambrien)	gneiss (kinzigites) et ultrabasites (péridotites, pyroxénolites) des Beni-Bouchra (Précambrien ?)
Précambrien ancien (et Archéen ?)	*Précambrien I :* Micaschistes et gneiss des Zenaga, du Tizi-n'Taratine; granites d'Azguémerzi et de Tazenakht. Micaschistes, amphibolites et granite de Tasserhirt, schistes du Kerdous, granite de Tazeroualt		Gneiss migmatitiques du Massif ancien haut-atlasique (Ouzellarh)		

N.B. : Selon A. Boudda & G. Choubert (1972), la base du Cambrien est située en (1)

Les nombreuses études sur les Hercynides marocaines réalisées depuis les travaux fondamentaux de H. Termier (1936) sont consignées pour la plupart dans les Notes et Mémoires du Service Géologique du Maroc. On en trouvera un exposé dans Michaud (1976). Les lignes qui suivent sont présentées comme un bref essai de synthèse des données disponibles, accompagné de quelques hypothèses de travail. L'essentiel portera sur le "Maroc africain".

II - STRATIGRAPHIE: LE MATERIEL, LES PHASES CALEDONIENNES

On se limitera ici à quelques indications sur le matériel constituant la chaîne primaire du Maroc, en complément aux tableaux 24.1, 24.2 et 24.3.

LE SOCLE PRECAMBRIEN

C'est dans le Sud marocain qu'il est le plus développé à l'affleurement. De nombreuses études récentes lui sont consacrées (Choubert et Faure-Muret, 1970; Choubert *et al.*, 1974; Hassenforder et Jeannette, 1974; Benziane, 1974; Charlot, 1976; Jeannette et Tisserant, 1976; Yazidi, 1976).

Le matériel le plus ancien, et en général aussi le plus métamorphique, paraît bien procéder d'une croûte continentale archéenne, vieille de quelque 3 milliards d'années, totalement reprise (ou quasi) dans des événements tectoniques et métamorphiques plus "jeunes", c'est-à-dire relevant seulement d'un Précambrien ancien: 2 milliards d'années environ. On interprète ainsi le "Précambrien I" et l'ex- "Précambrien 0" comme des fragments du Craton de l'Ouest africain, qui affleure largement plus au Sud dans le Bouclier R'guibate (= Reguibat).

Là, ainsi que sous le bassin de Tindouf qui s'étale à son bord nord, cette croûte ancienne est restée stable, rigide, jusqu'à nos jours. Par contre, dans l'Anti-Atlas et notamment dans ses parties septentrionales et centrales, cette croûte a été engagée dans une "zone mobile" dont l'orogenèse (mono- ou polyphasée) s'est essentiellement déroulée, semble-t-il, à la fin du Précambrien supérieur et au tout début du Cambrien. C'est l'interprétation du "Précambrien II" s.s. et du "Précambrien II-III" (Anti-Atlasides) comme segments de la chaîne pan-africaine, édifiée vers 650-550 Ma et contemporaine (voisine aussi) des chaînes avalonienne, briovérienne, etc.

Les molasses rouges discordantes et le volcanisme subsequent acide constituent le "Pr. III" ou Infracambrien inférieur, mais des faciès analogues existent déjà dans le "Pr. II-III" plissé.

Cette période du Précambrien terminal-Eocambrien est enfin caractérisée par un climat très froid et humide. Des glaciations répétées ont été bien identifiées sinon au Maroc, du moins plus au Sud, dans les séries sahariennes et voltaïques de cet âge, ainsi que, chose remarquable, dans le Précambrien supérieur de Terre-Neuve.

Tableau 24.2

LES PRINCIPALES FORMATIONS DU PALEOZOIQUE ANCIEN DISTINGUEES AU MAROC

Périodes / Domaines		ANTI-ATLAS	ATLAS	MESETA	RIF
Silurien	Ludlow Wenlock Llandovery	Schistes ampélitiques à Graptolithes des Plaines du Dra, à nodules et bancs calcaires et grès Lacune à la base	Cf. Anti-Atlas	(mêmes coupes que l'Ordovicien en général) vers la base : Schistes de Mokattam	*Nappes paléozoïques = Ghomarides :* griottes à Orthocères, phtanites, schistes du Llandovery sup.
ORDOVICIEN	Ashgill	Grès du Deuxième Bani : grès grossiers et argiles microconglomératiques	Quartzites grossiers, grès rubanés et argiles microconglo. (Adrar n'Dgout, J. Bou-Dahar, Tazzeka, Mts d'Oujda)	Cf. Anti-Atlas et Atlas : Jbilet, Rehamna (Kharrou), Smaala (Chaaf), Méséta côtière	? ? Schistes et greywackes, Conglom. à galets étirés
		Schistes du Ktaoua supérieurs à lentilles carbonatées		? ? Schistes et grès flyschoïdes d'El-Krad (nappes de Ziar) et du Goulibet (Rehamna)	
	Caradoc sup.	Grès et quartzites de Rouïd-Aïssa	Schistes et grès (H. Atlas central) Quartzites à Scolithes (Tamlelt)	Grès et quartz. d'Allahia sup. et du Kharrou; de Feddan-Tabba; de Sidi-Saïd et du Tirmah-Beddouz. Sch. et grès des Ouled Fennane, d'Asfar (p.p. ?) de Khénifra (?) - Schistes «OBK»...	
	Caradoc inf. moyen	Schistes (et grès) du Ktaoua inférieurs	Schistes psammitiques *p.p.* (Oujda)		
	Llandeilo	Grès et quartzites du Premier Bani (avec lumachelles; fer oolithique à la base)	Grès et quartzites (coupes : cf. Ordov. inf.)	...d'Ouljet—Bou-Khemis. Grès et qz. d'Imfout (sup.), d'Allahia inf.; schistes en dalles des Smaala; grès des Ouled-Bahloul.	
	Llanvirn	Schistes du Tachilla (Schistes des Feijas externes)	Schistes et grès verts et rouges à pistes (Ida-ou-Zal, Adrar n'Dgout, Skoura, Foum-Zabel, Mougueur, Sebbab-Kébir...)	Grauwackes d'El-Harcha; argiles gréseuses de Demja; psamm'tes de Sidi-Khriali	
	Arenig	Grès et quartzites du Zini (à l'Ouest) Sch. du Fezouata sup.		Schistes et grès micacés bioturbés (Jbilet, Rehamna, Imfout); sch. du Tergou; de Chabet-el-Oukaref; Sch. et qz. des Zaïan (partie sup. ?)	*Nappes cristallophylliennes = Sebtides :* anté-Llanvirn (voir tableau précédent)
	Tremadoc	Sch. du Fezouata inf. Grès ou lacune basale			
CAMBRIEN	Lacune du Cambrien supérieur ? Cambrien moyen (Acadien moyen et inférieur)	Lumachelles à Brachiopodes (localement) Grès à *Conocoryphe* et Lingules = du Tabanit Schistes à *Paradoxides* = des Feijas internes, avec greywackes pyroclastiques	cf. Anti-Atlas : Massif ancien du Haut Atlas occid. (blocs E et W) Pays de Skoura, Tamlelt...)	Grès et qz. d'El-Hank, du Zguit, des Zaïan (partie inf. ?) Psammites à arenicoles Sch. à *Paradoxides*, sch. de Bouznika, de Ouardane. Sch. et greywackes inf. Rehamna Séricitosch. et Qz. des Sehoul ?	
	Cambrien inférieur (« Géorgien ») daté	Brèche à *Micmacca*, schistes à Protolenidés Grès terminaux, tufacés, à Tigillites (étage d'Asrhir) Série schisteuse (ét. d'Issafene) et calcaires « scoriacés » Série schisto-calcaire (ét. d'Amouslek) passant vers le Sud-Ouest au faciès « Cal. sup. » Calcaires sup. à *Archaeocyathus*	Haut Atlas occid. (Massif du J. Tichka) : Complexe volcano-sédimentaire andésitique à bancs de calc. à *Archaeocyathus;* calc. « scoriacés » des Ida-ou-Zal. (voir tableau précédent)	? ? trachyandésites de l'O. Rhebar Schistes troués des Rehamna occidentaux (Lalla Mouchaa) Série du Bou Gader (*p.p.* ?) Calc., calcschistes et dolomies (Adoudounien ?)	

Dans le Maroc atlasique et mésétien s'étendait probablement, à l'aube du Paléozoïque, un croûte continentale de nature analogue à celle de l'Anti-Atlas. Les seuls affleurements de ce socle sont ceux du Haut Atlas occidental (Ouzellarh, etc.) et oriental (Tamlelt) et ceux, très exigus, de Méséta (El-Jadida, J.Hadid..), mais les indices indirects sont nombreux.

Vers le futur domaine rifain, ce même socle se prolongeait probablement sous une partie de l'actuel sillon externe. Mais au-delà? On ne retrouve de Précambrien probable que dans les nappes septentrionales des Sebtides (et dans leurs homologues des Cordillères bétiques). L'importance des péridotites et le métamorphisme catazonal lui confèrent un caractère particulier: peut-être s'agit-il de fragments du manteau, déformés lors d'une orogenèse du Précambrien supérieur, mais l'idée de diapirs ultrabasiques tertiaires est aussi défendue.

LE MATERIEL PROPRE

Le matériel propre des Hercynides moracaines est une série très puissante: 6000 à 8000 m en moyenne, 10 000 m dans le sud-ouest. C'est qu'elle englobe le plus souvent tout le Primaire, de l'Infracambrien au Carbonifère supérieur: aucune discordance généralisée n'interrompt cette accumulation jusqu'au Viséen supérieur (domaine atlaso-mésétien, socle rifain interne) et au Houiller (ensemble du Maroc), et les domaines sont restés pour l'essentiel en dehors de l'aire calédonienne.

On se fera une idée des séries "calédono-hercyniennes" d'après quelques-unes des figures suivantes et les tableaux 24.1, 24.2 et 24.3. On notera un "air de famille" prononcé entre les séries atlaso-mésétiennes et les séries anti-atlasiques. Les unes et les autres montrent des variations de faciès d'ouest en est, qui sont analogues au nord et au sud. Citons notamment la réduction du Cambrien inférieur calcaire vers l'est et le maintien des faciès marins au

Tableau 24.3

LES PRINCIPALES FORMATIONS DU PALEOZOIQUE RECENT DISTINGUEES AU MAROC

Périodes	Domaines	ANTI-ATLAS	ATLAS	MESETA	RIF
Permien	Permo-Trias				*Ghomarides et Sebtides sup. (Federico) :* conglomérats polychromes, sch. et grès rouges ou noir de fumée
Carbonifère supérieur	Stéphano-Autunien	Série argilo-gréseuse rouge (Merkala = Betana sup.; Béchar)	Grès Conglom. rouges de l'Ourika, de la Tessaout, du Tazekka, série limnique (des Ida-ou-Zal)	Couches de Bou-Achouch, conglo. de Khénifra et des Chougrane	
	Westphalien Namurien sup.	Houiller paralique du Bassin de Kenadza-Béchar Série de la Betana inf. (Bassin de Tindouf) avec les Grès fluviatiles du J. Berouina	Houiller paralique du Bassin de Jerada Grès à plantes	Conglomérats de Mechra-ben-Abbou et de Sidi-Kassem sem ? ?	*Nappes paléozoïques = Ghomarides :*
Carbonifère inférieur	Namurien inf. Viséen sup.	Calcaires de l'Ouarkziz et de la Zousfana Flysch de Ben-Zireg, à olistostromes	Flysch andésitique (Oujda) conglom., flysch à brèche intraform. de l'Atlas de Marrakech Sch de Debdou et du Mekkam (?)	Sch. à roches vertes du Sarhlef et des Rehamna. Grès à plantes d'Ennta. Flysch du Fourhal, de Kassem-Rahal. Flysch à olistostromes de Karrouba, de Ziar-Mrirt Calc. et conglomérats de base	Flysch, conglom. à galets de roches métam.; calc. récifaux transgressifs (Beni-Hozmar, Akaïli)
	Viséen inf. moyen	Sch. et grès de la Betaïna		Conglom. du Jbel Bakach Flysch du Khorifla (p. sup. ?)	
	Tournaisien	Grès du		Sch. à goniatites, flysch	« Culm. » à Calizas ala-
	Strunien	Tazout		Grès et qz. de Skhirat. Flysch du Khatouat (?)	
Dévonien supérieur	Famennien	Argiles et grès du Dra et du Zemoul; flysch du Maïdère; calc. pélagiques et couches à Arthrodires du Tafilalt		« Flyschs » du bassin de Si. Bettache à conglomérats ± chaotiques (Biar-Setla, Tiflet) et arkoses; des Ouled-Abbou; de Moulay-Hassane. Lumachelles et brèches pélagiques de Touchchent; Grauwackes transgr. de Mechra-ben-Abbou	beadas (calc. contournés) Sch. et phtanites (Beni-Hozmar)
	Frasnien	Schistes et calc. en miches du Dra; calc. pélagiques et griottes du Tafilalt		Argilo-pélites des Ouled-Abbou; marbres de Skhirat; griottes de Moulay-Hassane	
Dévonien moyen	Givétien	Faciès construits du Maïdère	Calc. à Tentaculites des Aït-Mdioulal	Marbres récifaux de l'O. Ykem, de Tiliouine, du Bou-Khemis (D. Aït-Abdallah) Grès, et calc. à flore terrestre (Bou-Nebdou). Calc. zoogènes (Ouled-Abbou, Mechra-ben-Abbou) Conglomérats rouges et calcaires dolomitiques des Rehamna (Skhirat -	Calc. récifaux (Al-Hoceïma, Talembot)
	Eifélien	Schistes marnes et calcaires gréseux Calcaires à Goniatites	Calcaires de Talmakent		
Dévonien inférieur	Emsien Praguien	W : (Dra) Trilogies des Rich (grès/argiles/calc.) El-Ansar, Mersakhsaï, Assa E : (Tafilalt) : Récifs d'Hamar Laghdad; calc. crinoïdiques ou calc. pélagiques	Schistes et Calcaires	Es-Slimane) discordants et déformés (Kef-el-Mouneb) Calc. et Dol. de l'O. Akrech, de Safsaf, d'Aïn-Targa. Grès calc. de Moulay-Hassane Calc. du Bou-Regreg, du Grou, de Ghtira, Rechoua, Tiflet.	Sch. et calc. à Tentaculites (Koudiat-Tizian, Bokkoya)
	Lochkovien	(avec grès vers l'W) Argilo - sch. et calc. Volcanisme	Conglom. rouges de Talmakent	Sch. et cal. de Hossei; calc. de Mograne; Sch. du Zemmour (Rabat), de Touchchent. Sch. et calc. des Ouled Abbou, de Mechra-ben-Abbou.	

Westphalien dans le nord-est. Les lignes isopiques ne sont pas connues précisément partout; elles paraissent discontinues entre le nord et le sud des Atlas; néanmoins, la constatation précédente doit conduire à minimiser les décrochements le long de la ligne sud-atlasique.

Les dépôts indiquent généralement des mers épicontinentales de profondeur faible à moyenne, où dominent les apports terrigènes méridionaux; font exception la mer cambrienne, où les apports pyroclastiques sont importants, surtout au nord-ouest; et les mers dévono-dinantiennes, à bathymétrie et subsidence contrastées, où apparaissent des rides à dépôts pélagiques condensés, des bassins à flyschs, des talus faillés à brèches chaotiques (voir III et IV).

Les faunes indiquent que ces mers, transgressives au sud sur le craton saharien, dépendaient au nord de la Téthys (faunes ardennaises et bohémiennes, notamment); elles présentent aussi des correspondances avec les faunes "atlantiques" de l'est des Appalaches; les faunes américaines pénètrent soudain au Maroc à la fin du Dévonien inférieur.

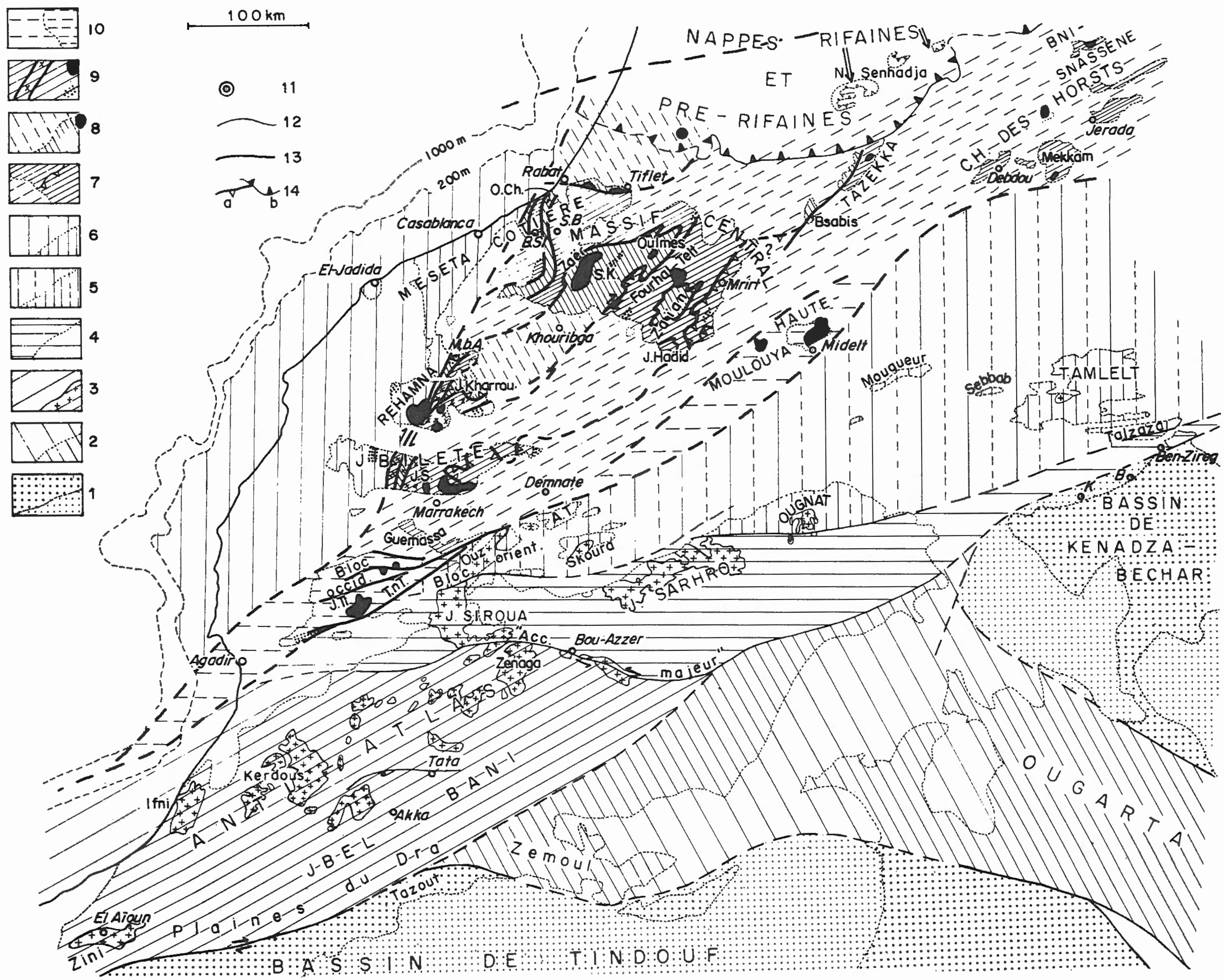

100 km
1
2
3
4
5
6
7
8
9
10
11
12
13
14
a
b
NAPPES RIFAINES
ET
PRE-RIFAINES
N. Senhadja
BNI-SNASSENE
CH. DES HORSTS
Jerada
Mekkam
Debdou
TAZEKKA
Bsabis
Rabat
Tiflet
O.Ch.
Casablanca
El-Jadida
MESETA
COTIERE
S.B
B.S.
MASSIF CENTRAL
Oulmes
Fourhal
Telt
Mrirt
Khouribga
J.Hadid
HAUTE MOULOUYA
Midelt
Mougueur
Sebbab
TAMLELT
Talzaza
Ben-Zireg
M.b.A.
REHAMNA
J.Kharrou
J.BEBLET
J.S.
Demnate
Marrakech
Guemassa
Bloc occid.
Bloc orient.
Ouz
T.n.T.
J.T.
Skoura
J. SARHRO
OUGNAT
BASSIN DE KENADZA-BECHAR
J. SIROUA
Acc.
Bou-Azzer
majeur
Zenaga
Agadir
ANTI-ATLAS
Kerdous
Tata
Ifni
Akka
BANI
JBEL
Zemoul
OUGARTA
Dra
Tazout
Plaines du
El Aioun
Zini
BASSIN DE TINDOUF

Figure 24.2.

Sur la stratigraphie des divers systèmes paléozoïques, on consultera notamment Destombes (1971), Hollard (1967, 1974), Boudda *et al.* (1974), Bensaïd (1974), Kelling *et al* (1975), Cavaroc *et al.* (1976), Willefert (1963 et en prép.).

LES PHASES CALEDONIENNES

Avant les "phases" hercyniennes proprement dites (voir IV), des terres émergées se sont édifiées sur la marge ouest du Maroc, à l'Ordovicien supérieur et au Dévonien inférieur, à la faveur des mouvements "taconiques" et "acadiens".

Dans le Nord-Ouest de l'Anti-Atlas, le Llanvirn repose en discordance sur l'Acadien. Les dépôts post-glaciaires de l'Ashgill sont discordants sur le sud-ouest de l'Anti-Atlas. Une lacune plus ou moins longue existe vers la base de la série silurienne en Meseta (Willefert, 1963; Destombes et Jeannette, 1966).

Mais les manifestations orogéniques les plus nettes sont celles de l'orogénie acadienne. Hollard *et al.* (1976) notent "qu'il s'agit:

- dans le Sud marocain (domaine de l'Anti-Atlas), de divers indices d'instabilité de l'écorce: transgression puis régression du Lochkovien, avec émersions locales et avec volcanisme basaltique dans l'Est; transgression au Praguien, puis apports terrigènes particulièrement épais dans l'Ouest (Rich) jusqu'à l'Eifelien inclus; discordances intra-givétiennes dans l'Est (Maïdere); apports sableux venant du Nord dans le Frasnien inférieur-moyen oriental.
- sur la frange occidentale des domaines de l'Atlas et de la Meseta, de la présence de conglomérats souvent très grossiers et de grès rouges à plantes, localement discordants jusque sur le Cambrien et datés du Lochkovien (Talmakent) ou du Dévonien moyen (Rehamna). Ces dépôts "molassiques" sont le plus souvent suivis de récifs. l'Eifelien est souvent transgressif.
- dans l'anticlinorium Rabat-Tiflet, tout au nord-ouest de la Meseta et au-delà d'un décrochement E-W dextre, d'un granite daté de 415 Ma, intrusif dans des schistes et quartzites (Cambro-Ordovicien ?) et remanié dans le Dévonien moyen transgressif.

Divers âges analogues pourraient lui correspondre dans le Sud marocain, également comparables à l'"événement thermique" postérieur à 450 Ma. décrit au Hoggar occidental.

La paléogéographie du Dévonien inférieur et moyen marocain est dirigée par cette orogénie acadienne, qui dût ériger une chaîne à la bordure occidentale du pays. Elle doit être distinguée des phases hercyniennes qui, dès le Dévonien supérieur puis au Dinantien, établissent une paléogéographie nettement différente (rotation des lignes isopiques dans le Sud marocain)."

L'intérêt de ces observations pour les corrélations avec les Appalaches canadiennes est évident.

Notons qu'une orogenèse anté-Llanvirn est décrite dans le socle rifo-kabyle; il pourrait s'agir d'une phase calédonienne précoce (?), mais un âge précambrien supérieur est plausible; elle succèderait à une orogenèse d'un âge précambrien plus ancien (Kornprobst, 1971; Bossière *et al.*, 1972). Au-dessus de ce socle métamorphique poly-orogénique se présente une série primaire qui n'est pas sans évoquer celles du nord-est atlasique (voir IV).

LA COUVERTURE

Mis à part des couches permiennes localisées (tableau 24.3), elle débute généralement (Meseta, Atlas, Rif externe) par une série rouge gréso-argileuse et salifère du Trias supérieur - Infralias, où des basaltes abondants indiquent l'effet des distensions atlasiques. Cependant, la mer ne gagnera le Sud marocain qu'au Crétacé.

Figure 24.2. Carte structurale schématique des Hercynides du "Maroc africain". - En dehors des zones d'affleurements, en trames serrées et délimitées par un pointillé, les prolongations des unités structurales sont plus ou moins hypothétiques suivant les secteurs.

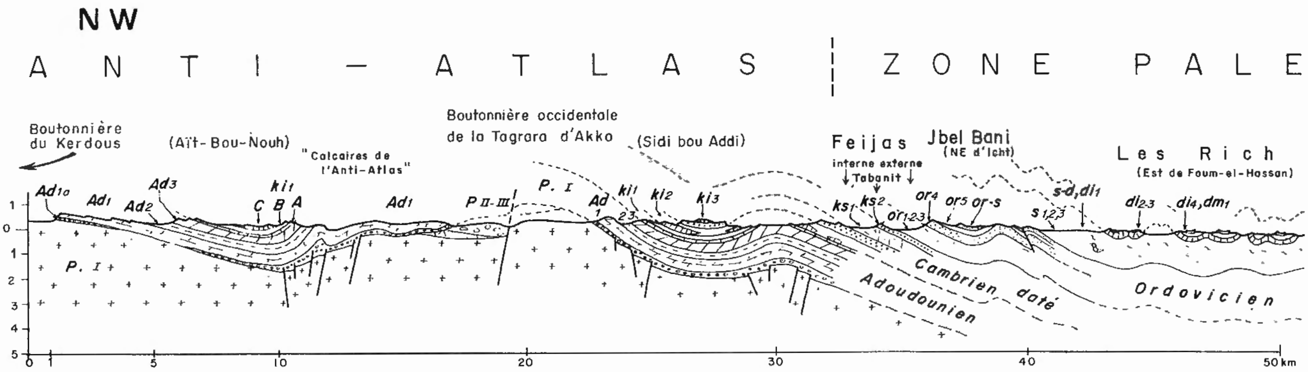
Dalle saharienne: 1. Bassins tabulaires (ouest) à sub-tabulaires (est).

Marge péricratonique non déformée après le Trias: 2. Chaîne d'Ougarta et zone de jonction avec 3 et 4. -3. Anti-Atlas *s. str*; croix: "boutonnières" de Précambrien ancien et moyen (?) - supérieur plissé - 4. Massifs anti-atlasiques nord-orientaux, sauf Ougnat et Talzaza (rattachés à 5 ?)

Hercynides atlaso-mésétiennes: 5. Marge sud = massifs haut-atlasiques à l'est de Marrackech et frange anti-atlasique NE (caractères péricratoniques). - 6. Marge ouest = môle côtier de Meseta occidentale. - 7 à 9. Unités orogéniques *s. str.*: 7. Zones synclinoriales dévono-dinantiennes, supportant des nappes en Meseta orientale; 8. Zones anticlinoriales et zone de Rabat-Tiflet; 9. X: Couloir de métamorphisme de la marge ouest; noir: granite; points: molasses continentales post-orogéniques. -10. Blocs métamorphiques de la marge sud. -11. Sondage ayant atteint un socle granitisé.

Limites de zones structurales: 12. dans la marge péricratonique (décrochements *p.p.*); -13. dans les Hercynides atlaso-mésétiennes (décrochements *p.p.*); -14. chevauchements: a: hercynien; b: rifain.

Abréviations: A.T. = Aït-Tamelilt; B. = Bechar; B.Sl. = Ben-Slimane; (Bloc occ. et or. = Blocs occidental et oriental du Haut-Atlas de Marrakech; J.S. = Jbel Sarhlef; J. Ti. = Jbel Tichka; K = Kenadza; M.b.A. = Mechraben-Abbou; O. Ch. = Oued Cherat; Ouz. = Ouzellarh; S.B. = Sidi-Bettache; S.K. = Sidi-Kassem; TnT. = Tizi n'Test.



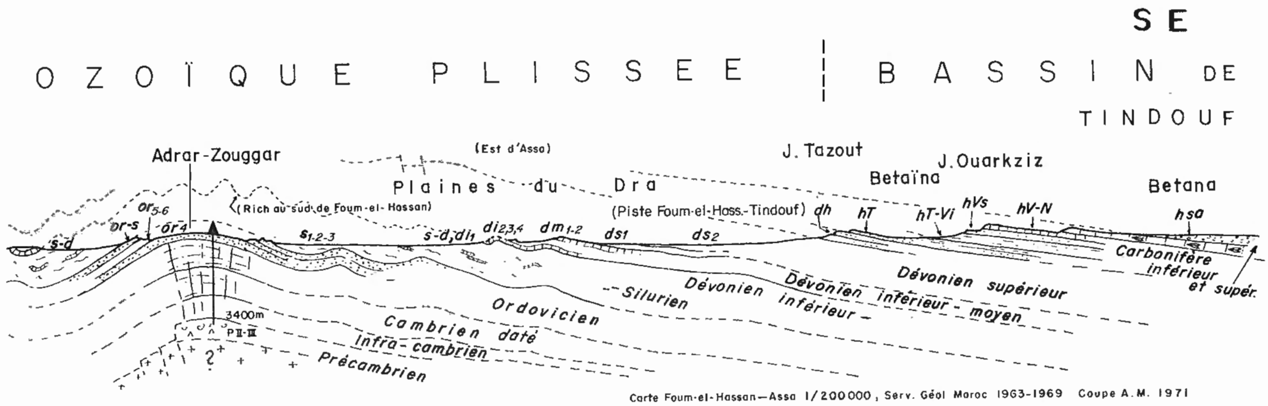
PI: Archéen (?) ou Précambrien ancien, granito-gneissique, à roches vertes, non subdivisié.

PII-III: Précambrien supérieur = série d'Anezi à rhyolites et andésites puis pélites, grès et conglomérats.

Ad: Adoudounien (Infracambrien supérieur)
Adla: Série de base de l'Adoudounien, conglomératique.
Adl: Adoudounien inférieur: "Calcaires (dolomies) inférieurs de l'Anti-Atlas".
Ad2: Adoudounien moyen: "Série lie-de-vin".
Ad3: "Calcaires (dolomies) supérieurs" *pro parte*.
ki: Cambrien inférieur = "Géorgien", daté paléontologiquement.
kilA: Calcaires à *Archaeocyatha* dolomies et schistes.
B: "Série schisto-calcaire" à Trilobites, avec prédominance gréso-pélitique.
C: Idem avec prédominance de calcaires à *Archaeocyatha*.
ki2: Biohermes à *Archaeocyatha* et "Calcaires scoriacés", puis "Schistes supérieurs".
ki3: "Grès terminaux" Grès pyroclastiques puis grès grossiers à Scolites (Tigillites).
Le sommet du Cambrien inférieur est un peu au-dessus, à la base de ksl (niveau d'Ouriken n'Ourmast).
ks: Cambrien moyen (partie moyenne)
ksl: Schistes à *Paradoxides*, greywackeux, verts.
ks2: Grès à *Conocoryphe* et Lingules.
or: Ordovicien
orl: Grès blanc Trémadoc?
or2: Schistes de Fezouata et Grès du Zini à intercalations argileuses, Trémadoc supérieur? - Arenig inférieur et moyen.

Figure 24.3 (p.p.) Coupe schématique du domaine anti-atlasique occidental, extrait de Michard (1976), d'après la carte géologique au 1/200 000 Foum-el-Hassane-Akka.

Figure 24.3 (p.p.)



or3: Schistes du Tachilla, Llanvirn.
or4: Grès et quartzites du 1er Bani, Llandeilo.
or5: Formations du Ktaoua inférieur et de Rouid-Aïssa, schisto-gréseuses, Caradoc.
or6: Ktaoua supérieur, Ashgill (*pro parte*).
or-S: Grès terminaux du 2è Bani, Ashgill (*pro parte*).
s: Silurien
s1: Argilo-pélites à Graptolites: Llandovery, Tarannon, Wenlock *p.p.*
s2,3: *Idem* avec, vers le haut, lits calcaires à nautiloïdes orthocônes, *Cardiola*, crinoides.Ludlow,*p.p.*
s-d: Argilo-pélites et grès à pistes, couches de passage Ludlow-Dévonien inférieur.
di: Dévonien inférieur
di1: Schistes et calcaires gréseux, ferrugineux, Gédinnien-Siegenien *p.p.* (Lochkovien).
di2,3: 1er et 2ème Rich (Assa et Mersakhsaï), chacun fait d'une trilogie calcaire - argiles pélitiques - grès et lumachelles; Siegenien supérieur (?) - Emsien inférieur (Praguien).
di4: 3ème Rich (El-Ansar), Emsien supérieur.
dm: Dévonien moyen
dm1,2: Schistes, marnes, calcaires gréseux ou non: Eifelian puis Givétien.
ds: Dévonien supérieur
ds1: Schistes et calcaires du Frasnien.
ds2: Argiles à intercalations calcaires ou gréseuses: Famennien.
d-h: Grès inférieurs du Tazout, Strunien.
h: Carbonifère
hT: Grès supérieurs du Tazout et schistes de base de la Betaïna, Tournaisien.
hT-Vi: Schistes et grès de la Betaïna: sommet du Tournaisien et Viséen inférieur.
hVs: Ouarkziz gréseux, Viséen supérieur *p.p.*
hV-N: Ouarkziz calcaire, suite du Viséen supérieur et Namurien inférieur; dans la partie sommitale, couches marno-gypseuses.
hsa: Base de la Série de la Betaïna: grès continentaux du Namurien supérieur. La série se poursuit par des molasses continentales du Westphalien Stéphanien.

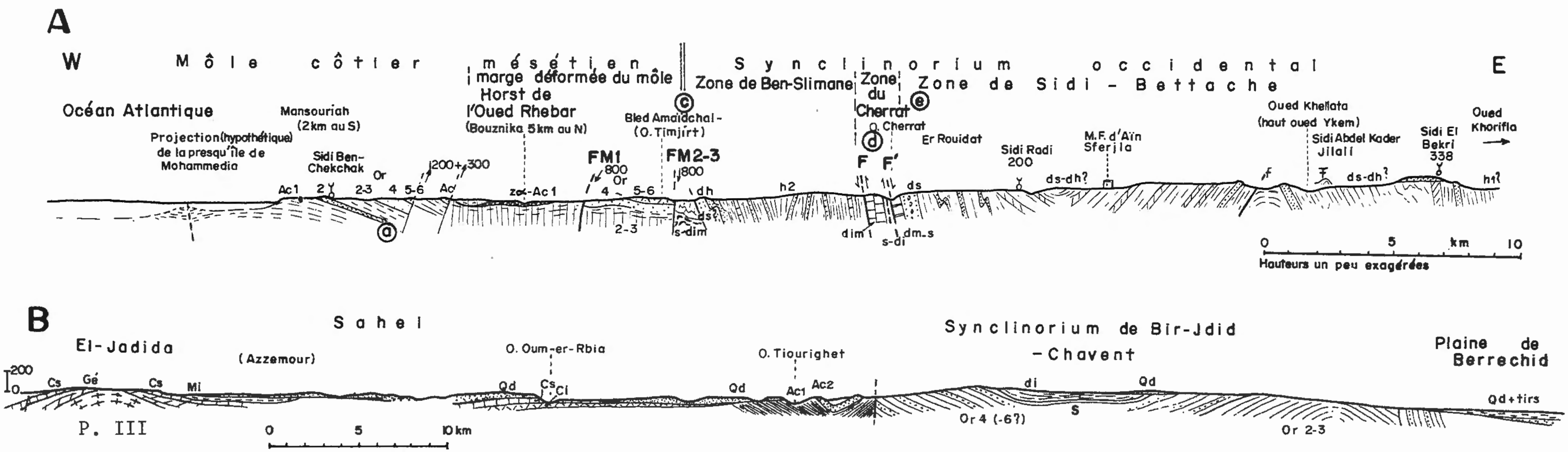


Figure 24.4

Deux coupes E-W dans la Meseta côtière:

A: Transversale nord (à environ 30 km au sud de Rabat) d'après J. Destombes et A. Jeannette (1966) et Lecointre (1926) (Sidi-Bettache).

B: Transversale sud (El-Jadida), d'après M. Gigout (1951).

P.III: "Précambrien III" Infracambrien inférieur (?) rhyolitique; Gé: "Géorgien" dolomitique; Ac: Acadien moyen, 1: "Schistes à *Paradoxides*" et schistes de Bouznika; 2: Quartzites d'El-Hank; zα: trachy-andésite; Or: Ordovicien, 2-3: Arenig inférieur-moyen, 4: Llanvirn-Llandeilo, 5-6: Caradoc; s: Silurien - d: Dévonien, im: inférieur moyen, s: supérieur - dh: Strunien quartzitique; dh-ds?: Strunien (?) flyschoïde. - hl: Tournaisien (limites?). - h2: Viséen supérieur, molassique, puis flyscheux. Ts: Trias supérieur (et moyen?) argilo-salière; Ci: Crétacé inférieur; Cs: Crétacé supérieur; Mi: Miocène marin; Qd: Quaternaire ancien et Moghrébien dunaire.
FM: Failles marginales du môle côtier; F et F': Failles-limites de la zone du Cherrat; f: Faille avec filon de microdiorite quartzifère (marge W de l'anticlinal du Khellata).

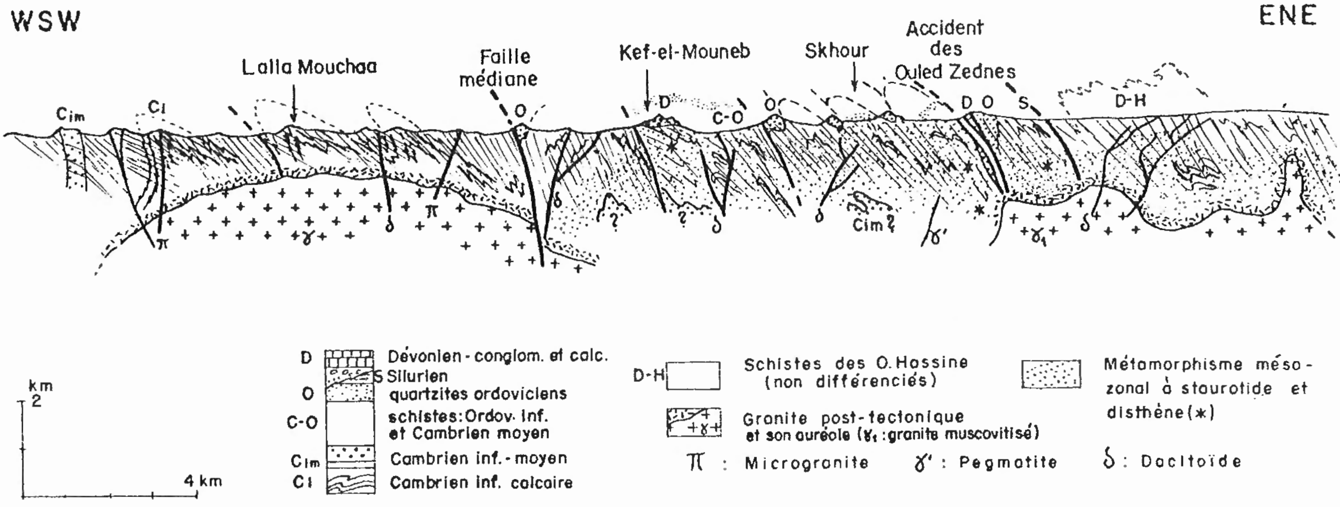


Figure 24.5. Coupe interprétative dans les Rehamna centraux, d'après Michard (1967 à 1969), Piqué (1972), Hoepffner (1974), Jenny (1974). N.B.: Les schistes des Ouled-Hassine (D-H), formation flyschoïde avec roches vertes (amphibolites) et kératophyres (porphyroïdes) dans sa partie supérieure, est attribuée au Dévono-Dinantien.

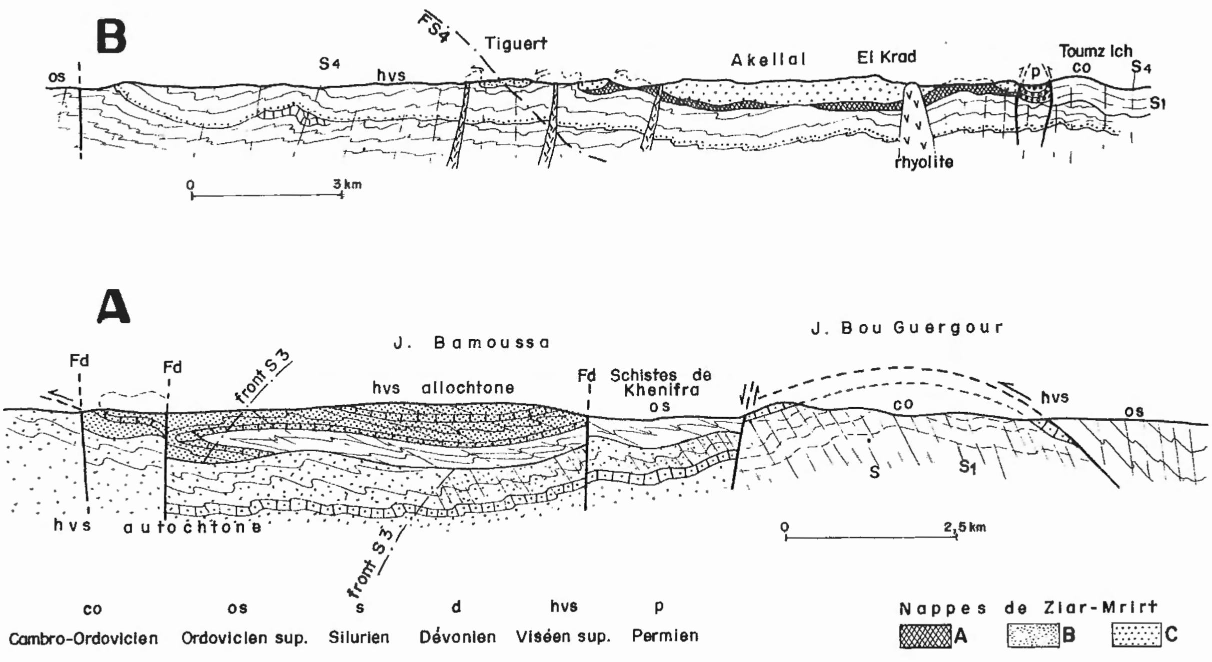


Figure 24.6 Nappes hercyniennes du sud-est du Massif Central.

A: Coupe composite dans les structures méridionales: la nappe pli couché-cisaillé de Khenifra et l'autochtone du J.Bou Guergour (équivalent du J.Hadid), d'après Allary (1972).

B: Coupe générale dans le nord du secteur: les nappes de Ziar-Mrirt et le bassin molassique permien de Khenifra, d'après Ribeyrolles (1972).

III TECTONIQUE: LES UNITES STRUCTURALES DU "MAROC AFRICAIN"

LE SUD ALGERO-MAROCAIN, UNE MARGE PERICRATONIQUE

Au sud de la ligne sud-atlasique, la tectonique hercynienne reste modérée et de caractère germanotype; la couverture épaisse, avec localement (ouest) une schistosité sub-verticale dans les niveaux inférieur, s'adapte à un socle découpé en blocs plus ou moins déplacés.

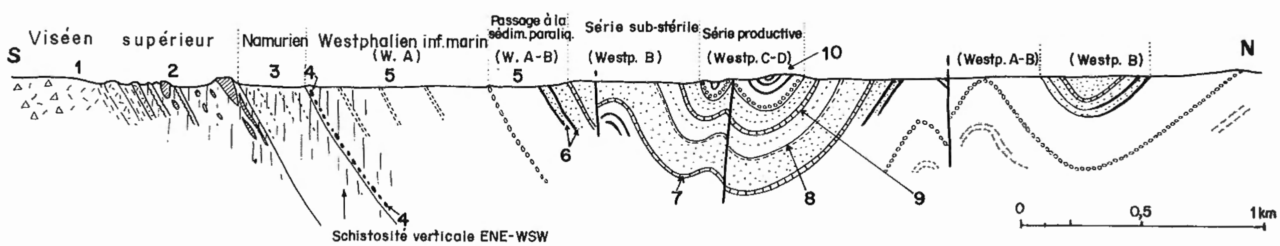
On peut distinguer trois segments déformés, agencés en une sorte d'arc (fig. 24.2).

1. *A l'Ouest, l'Anti-Atlas s.str.* présente des plis de direction SW-NE "atlantique", au-dessus du socle précambrien découpé en blocs. Les plis, de style concentrique à concentrique-aplati, sont disharmoniques d'un niveau à l'autre, les principaux niveaux de disharmonie correspondant au Cambrien moyen, au Siluro-Lochkovien et au Dévonien supérieur (fig. 24.3). Au plissement est associé un épimétamorphisme dont Chennaux *et al.* (1970) ont montré l'extension dans le nord du Sahara. La schistosité, sub-verticale, est particulièrement marqués à l'ouest (au moins jusqu'au Silurien). Tout au sud-ouest, on observe des plis déversés et chevauchants vers l'ESE (J. Khannfra, au sud du Zini: Destombes et Hollard, *in litt.*). On connaît peu de plis dans le Carbonifère (Tazoût à l'est d'Assa), mais on peut supposer que les couches de cet âge ne sont conservés qu'en dehors de la chaîne anticlinoriale. La limite entre l'anticlinorium et le bassin de Tindouf est progressive à l'est et tout au sud-ouest, dans les chaînons qui s'étirent en direction des Mauritanides. Elle est brusque et oblique aux structures entres ces secteurs, où l'on peut envisager un décrochement dextre est-ouest dans le Bas Dra (fig. 24.2). Cette hypothèse, formulée aussi par Fabre (1971), n'exclut pas celle de plis en partie néodévoniens (voir III,3). Dans le sondage de l'Adrar-Zougar (situation: Fig. 24.3), l'épimétamorphisme de la fraction argileuse a été daté de la phase saalienne (269±14 Ma) par Esquevin et Menendez (1974), mais une première phase de cristallisation des micas serait intervenue des 400 Ma Cette "phase ardennaise" n'est cependant pas enregistrée par une discordance dans la sédimentation (cf. fig. 24.3), contrairement aux phases du Dévonien supérieur.

2. *A l'est, l'Ougarta* est orienté à environ 120° de l'anticlinorium précédent. Son style est analogue, avec des plis majeurs larges, voire coffrés, souvent croisés, parfois des plis mineurs plus serrés, une schistosité vers la base, pouvant passer dans le socle. Ce plissement découlerait du coincement de l'aulacogène ougartien entre le craton ouest-africain et le prolongement nord du bouclier touareg (Fabre, 1969; Donzeau, 1974), ceci à une époque relativement tardive, permienne à triasique (plis du bassin de Kenadza-Béchar qui seraient localement accompagnés de chevauchements, selon Fabre, comm. orale).

3. *Entre les deux branches de "l'arc" Anti-Atlas - Ougarta, une région centrale* montre l'interférence des directions qui les caractérisent. Au sud du môle précambrien du J.Sarhro, s'étend une région de dômes et cuvettes peu accentués. Une zone de tectonique complexe la parcourt, "l'accident majeur de l'Anti-Atlas", qui aurait fonctionné dès le Précambrien (Choubert, 1963), puis a rejoué verticalement au Famennien (Hollard) et au Néogène. Son fonctionnement en décrochement sénestre durant l'orogenèse hercynienne est admis par Leblanc (1972) dans la région de Bou-Azzer, où il provoque des plis et chevauchements dans la couverture. Au sud-est du J. Sarhro (Maïdere, Tafilalt), les déformations précoces d'âge frasnien et famennien sont bien marquées dans la série sédimentaire; variations de faciès et discordances angulaires traduisent des mouvements verticaux et des plis amples (Hollard 1967, 1974); une inversion du relief se produit alors au nord de la plate-forme saharienne, le bassin de Tindouf se creusent, alors que l'Anti-Atlas se soulève.

D'une manière générale, on peut admettre avec Donzeau (1974) que la structure de l'"arc" Anti-Atlas - Ougarta, avec plis SW-NE et NW-SE, résulte de l'adaptation de la couverture au mouvement de blocs de socle, plutôt qu'à des phases superposées d'orientation différente. Mais l'inventaire de ces blocs et la description systématique de leurs coulissements reste à faire.



1: Tufs et brèches andésitiques, cinérites; 2: Flysch argilo-calcaire et pyroclastique (pélites à Goniatites; turbidites bioclastiques plus ou moins pyroclastiques; blocks exotiques calcaires ou dolomitiques. 3: Flysch pélitique à Goniatites; 4: 1er conglomérat à galets de phtanites siluriens; 5: Grès plus ou moins pyroclastiques et argilopélites à Goniatites; 6: 1ères veines de houille et niveaux à Fougères; 7: Niveau marin (calcaire crinoïdique + Goniatites); 8: *id.* (calcaire à *Productus*); 9: id. (à Goniatites); 10: Assise de Jerada: cinq veines d'anthracite dans stérile paralique (quelques niveaux à Goniatites, pélites à Plantes, à faune limnique, etc.).

Figure 24.7 Coupe dans un massif du nord-est atlasique: Le Carbonifère de Jerada d'après Horon (1952) et Owodenko (1952), avec quelques additions.

LE SOCLE ATLASO-MESETIEN, DOMAINE OROGENIQUE HETEROGENE

Les échantillons calédono-hercyniens offerts par les massifs anciens au nord de la ligne sud-atlasique sont très hétérogènes (fig. 24.2). Certains correspondent à des segments profondément déformés, d'autres à des zones faiblement plissées. Parmi celles-ci, les segments ou môles marginaux, que l'on décrira d'abord.

1. Les marges du domaine orogénique

a) *Marge sud* - Il s'agit des divers massifs haut-atlasiques, exception faite du "bloc occidental" de l'Atlas de Marrakech. Celui-ci mis à part, les autres se présentent comme des dépendances de l'Anti-Atlas, une frange accidentée mais de structure comparable.

Dans le "bloc oriental" de l'Atlas de Marrakech, le Précambrien de l'Ouzellarh semble prolonger celui du Siroua et sa couverture est analogue. Cependant, comme plus au nord, une discordance existe à la base du Viséen supérieur, représenté par un flysch. On remarque que vers le sud, cette discordance ne dépasse que légèrement la ligne sud-atlasique (marge NE de l'Anti-Atlas oriental).

Ces caractères se retrouvent dans l'Atlas de Demnate. L'ouest de la "boutonnière" de Skoura est sub-tabulaire; vers l'est s'observent des plis E-W; dans celles de Aït-Tmelilt plus au nord, deux plissements post-viséens se superposent (Laville *et al.*, 1973).

Beaucoup plus à l'est, des couches cambro-ordovico-siluriennes constituent les petits massifs de Mougueur, où elles montrent un léger métamorphisme séricìteux (Caïa, 1969), et du Sebbab - Bou Dahar, où elles sont fossilifères (Agard et Du Dresnay, 1965). Ces mêmes couches montrent leur soubassement infracambrien et précambrien dans le Tamlelt, môle qui semble avoir joué le rôle de haut-fond mobile durant le Viséen supérieur (Pareyn, 1961; Du Dresnay, 1965; Pareyn *et al.*, 1975).

b) *Marge ouest* - Le deuxième ensemble de massifs où les déformations hercyniennes sont restées modérées se présente dans une position plus inattendue que le premier: il s'agit du môle côtier mésétien (Michard, 1969) affleurant en Meseta côtière et dans les parties ouest des Rehamna et des Jbilete, donc entièrement séparé de l'Anti-Atlas.

Sa structure ne laisse pas, pourtant, de ressembler à celle de certains secteurs anti-atlasiques (Bani, Dra), comme permet d'en juger la figure 24.4 Sur sa bordure orientale, la seule observable, on voit apparaître progressivement une schistosité quasi verticale pendant que s'accentue le plissement subméridien (Michard, 1967), puis un faisceau de failles bordières N 20° fait passer aux zones orogéniques de Meseta moyenne. Etudiées dans les Rhamna, ces failles hercyniennes se présentent comme des décrochements dextres importants, s'accompagnant de déversements et de chevauchements ver l'Ouest, (Michard, 1969; Hoepffner *et al.*, 1975b).

Ce bloc était probablement individualisé dès le Dévonien inférieur: des conglomérats rouges discordants éo- et méso-dévoniens sont connus sur sa bordure sud-orientale (Talmakent dans le Haut-Atlas; Jbilete occidentales; Rehamna) indices de mouvements "calédoniens" et "acadiens". La question de ses prolongements occidentaux mériterait d'être étudiée.

2. Les zones orogéniques proprement dites

Les zones de tectogenèse et de pétrogenèse profonde hercyniennes actives s'étendent dans le "triangle" ménagé entre les môles précédents (fig. 24.2). Les directions tectoniques y sont, en moyenne, orientées SW-NE, mais dessinent souvent des virgations entre la direction sub-méridienne et la direction E-W: virgation de Rabat et des Zaër, virgation des Rehamna et Jbilete orientaux. Les lignes isopiques, dont on ne connaît que l'allure générale, sont également orientées en moyenne SW-NE, au moins depuis le Dévonien: les unités structurales paraissent bien le fruit d'une longue évolution calédono-hercynienne, même si l'essentiel de la structuration est dévono-dinantienne.

a) *Le centre du dispositif* - L'inventaire de ces unités sera fait en partant de la plus centrale et plus large bande transversale d'affleurements, celle du Massif Central et de la Meseta côtière. D'ouest en est on distingue:

- la zone synclinoriale ouest, à flyschs dévono-dinantiens, comportant un fossé occidental (Ben Slimane) où le Viséen supérieur est transgressif et un fossé oriental (Sidi Bettache) où le Dinantien paraît complet; entre les deux, le faisceau siluro dévonien du Cherrat (fig. 24.4). Au nord, cette zone est limitée par les accidents E-W de Rabat-Tiflet (virgation de Rabat), qui ramènet des schistes probablement cambro-ordoviciens, avec un granite calédonien et des conglomérats dévoniens. Vers le sud, on peut l reconnaître dans le nord des Rehamna (Mechra-ben-Abbou), puis elle se pince entre le môle côtier et l'uni é anticlinoriale suivante (fig. 24.2 et 24.5).

- l'unité anticlinoriale ouest Khouribga-Oulmès, à armature cambro-ordovicienne ne qui paraît s'ennoyer régulièrement, au nord, sous les flyschs néoviséens discordants, à l'ouest sous ceux, concordants (?), du Dévonien supérieur - Dinantien. A l'est, une zone d'accidents, avec chevauchement des flyschs orientaux sur l'anticlinorium et décrochements dextres (Cailleux, 1974). Au sud-ouest enfin, cette zone paraît se retrouver dans les Rehamna orientaux et centraux (massif du Kharrou) où elle se biseaute à son tour contre les accidents de bordure du môle côtier, dans une zone de cisaillement et de métamorphisme intenses (staurotide et disthène syn- à tarditectoniques (Michard, 1968; Hoepffner *et al.*, 1975a). Des mouvements tardifs déterminent l'accentuation des virgations déjà ébauchées et des renversements et chevauchements locaux vers le sud-ouest.

- la zone synclinoriale axiale (Fourhal-Telt), à flyschs viséo-namuriens épais, qui semble pouvoir être prolongée dans les Rehamna méridionaux (micaschistes à amphibolites: fig. 24.5) et les Jbilete centrales (schistes du Sarhlef). On ne préjuge pas ici de l'autochtonie de ces flyschs, ni au nord (Cailleux, 1974) ni au sud (Hoepffner *et al.*, 1975a; Huvelin, 1975).

- l'anticlinorium oriental (Cambro-Ordovicien des Zaïn); il est entièrement enveloppé de flyschs néoviséens discordants et paraît une simple culmination du socle de la zone des flyschs orientaux. Les déformations anté-Viséen supérieur sont particulièrement intenses dans ce secteur (fig. 24.6b); elles peuvent dater du Famennien ou du Dinantien.

- les nappes orientales, qui chevauchent les flyschs précédents; elles semblent provenir de l'E, du SE ou du SSE. Dans les Jbilets orientales, elles sont interprétées comme des nappes de glissement sous-marin précédées d'un wildflysch (Huvelin, 1967); leur matériel va de l'Ordovicien au Dévonien. Des nappes analogues, à l'est du Massif Central, sont celles de Ziar-Mrirt (fig. 24.6b), cependant que celle de Khenifra (Ordovicien et Viséen supérieur) paraît résulter du cisaillement d'un pli couché (fig. 24.6a) (Allary *et al.*, 1972; Allary, 1972; Ribeyrolles, 1972), mais ces descriptions sont contestées par Mullin *et al.* (1976).

b) *Le nord-est* Il semble que l'on puisse retrouver la prolongation approximative des deux dernières zones citées dans le massif du Tazekka et dans ceux du Nord-Est atlasique (Debdou, chaîne des Horsts, Bni-Snassene); l'accident de Bsabis-Tazekka (Morin, 1973) y introduit une division majeure (Hoepffner, trav. en cours). Une particularité de ces zones très orientales est la persistance de la sédimentation marine durant le Westphalien (Houiller paralique de Jerada, fig. 24.7). Au contraire, dans les secteurs occidentaux, le Westphalien est inconnu, sauf localement sous un faciès de poudingues continentaux discordants (Sidi Kassem), peu distincts de ceux du Stéphano-Autunien (Khenifra, etc.). On remarque donc que, dans le domaine atlaso-mésétien comme dans celui de l'Anti-Atlas Ougarta, les plissements post-viséens sont plus précoces à l'ouest qu'à l'est. Des grès à brachiopodes du Westphalien inférieur probable jalonneraient, dans le nord-est du Massif Central (W d'Ito), la limite entre les deux secteurs (Hollard, *in litt.*).

c) *La limite sud* - Il reste à considérer la limite sud de ce domaine orogénique hercynien, vers l'Anti-Atlas.

On ne peut l'observer que dans le Haut-Atlas de Marrakech. Là, un "bloc occidental" fait de Cambro-Ordovicien profondément déformé (Schaer, 1964) et métamorphisé (Termier, 1971): Jbel Tichka, faciès amphibolite à grenat et staurotide affronte, par un faisceau de décrochements, en partie tardi- et post-hercyniens (voir I), un bloc oriental de type anti-atlasique. De deux choses l'une:

- ou l'on considère que le massif métamorphique haut-atlasique occidental représente la prolongation sud de la zone métamorphique des Rehamna et Jbilete centraux,
- ou bien on le considère comme un "copeau" du domaine anti-atlasique, séparé, et de celui-ci, et des zones orogéniques mésétiennes, par des décrochements orientés WSW-ENE (N 70°).

L'existence de flyschs viséens, au nord du bloc ancien atlasique (zone de Guemassa) est en faveur de cette deuxième hypothèse. On serait alors fondé à voir dans le massif de Haute-Moulouya (Midelt, Aouli-Mibladen), où Vauchez (1976) vient de montrer l'existence de déversements au SE, l'équivalent du "bloc occidental" haut-atlasique: ses schistes métamorphiques (à amphibolites) et ses granites ont fourni des âges isotopiques hercyniens (Tisserant, 1968). Dans cette hypothèse, les culminations métamorphiques du Haut-Atlas et de Haute Moulouya seraient à la lisière des zones peur déformées péricratoniques, en relation avec des accidents tectoniques majeurs (décrochements), comme la culmination des Rehamna à la lisière du môle côtier. La tectonique de ces massifs mérite de nouvelles descriptions, en particulier celle du massif haut-atlasique où sont signalés des décrochements (Schaer, 1964) et des chevauchements (*id.* et De Koning, 1957) hercyniens.

IV CONCLUSIONS: CARACTERES GENERAUX DES HERCYNIDES MAROCAINES

LES STRUCTURES DANS LE MAROC "AFRICAIN"

Leur style est celui de structures édifiées à partir d'une épaisse série sédimentaire reposant sur un socle de nature continentale, faiblement remobilisée.

La présence d'un tel socle précambrien polyorogénique est directement observable à la marge sud de la chaîne (domaine péricratonique de l'Anti-Atlas). Dans le domaine orogénique lui-même, la présence de ce socle se déduit:

- de la sédimentologie du Cambro-Ordovicien (de type anti-atlasique);
- de la rareté des roches vertes dinantiennes anté-tectoniques (seulement représentées par des sills ophitiques, des filons doléritiques et quelques coulées de basaltes en coussins localisées: Rehamna et Jbilete centraux, Meseta côtière au SE de Rabat),
- de l'importance des granites, à mise en place en partie syntectonique, en partie post-tectonique (intrusions supracrustales jusque dans le Dinantien, à partir des anticlinoriums cambriens),
- du style tectonique, marqué par un découpage en blocs et lanières séparés par de longs couloirs de décrochements, sièges privilégiés des phénomènes métamorphiques; au-dessus des blocs de socle, les niveaux supérieurs de la couverture, notamment les flyschs dinantiens, sont susceptibles de se "décoller" et de fournir des nappes de charriage; l'affrontement des blocs peut s'accompagner de déversements et écaillages plus ou moins importants;
- enfin de l'existence d'enclaves de gneiss granitoïdes remontées par les "microdiorites" post-tectoniques des Jbilete (Huvelin, 1975).

Les unités orogéniques atlaso-mésétiennes sont faiblement obliques à la marge péricratonique saharienne, tendant à se disposer "en échelon" le long de sa limite nord (fig. 24.2). Vers l'ouest, elles sont prises en écharpe par un "môle côtier" peu déformé. Dans l'hypothèse de rotations et translations post-triasiques exposée par Michard *et al.* (1975), la disposition des unités à la fin du Primaire serai repoussée d'environ 100 km vers l'WNW.

Les structures de ces unités orogéniques montrent en général un déversement vers le nord-ouest (plis, nappes), de sorte que si l'on voulait rechercher la polarité tectonique, on la trouverait curieusement dirigée de "l'avant-pays" vers les "zones internes": en fait cette nomenclature de chaîne géosynclinale ne convient pas ici. Par contre, tout au sud-ouest de l'Anti-Atlas, on remarque des structures déversées à l'est-sud-est: elles pourraient représenter déjà l'avant-pays des Mauritanides (voir III).

L'AGE DES DEFORMATIONS

Les Hercynides du "Maroc africain" se caractérisent par une évolution très longue, englobant les "cycles" calédonien et hercynien *s. str.* (voir II).

L'orogenèse hercynienne proprement dite présente une succession de "phases" du Dévonien supérieur au Trias, inégalement marquées suivant les régions, et qui sont sans doute des enregistrements discontinus d'une évolution continue:

- *"phase bretonne" précoce* (cf. "phase reussienne"), déterminent la discordance du Famennien supérieur jusque sur l'Ordovicien (Tafilalt) et l'apparition de flyschs chaotiques. Les mouvements sont déjà importants au Givétien et surtout au Frasnien (enchaînement avec les phases "acadiennes").
- *"phase bretonne" tardive*, déterminant des dépôts grossiers transgressifs du Viséen inférieur et, semble-t-il, une discordance du Viséen supérieur sur le Dévonien supérieur dans le nord du Maroc central.

Il est difficile de préciser, en général, l'âge des mouvements anté-viséen supérieur, qui ont édifié de véritables chaînes plissées à l'ouest et surtout dans l'est (fig. 24.6).

- *"phase sudète" précoce*, marquée par des discordances internes dans les flyschs néo-viséens et surtout par le glissement des nappes orientales, nappes "humides" intra-viséen supérieur ou nappes ou plis cisaillés, sub-contemporaines (fig. 24.6). Le Viséen supérieur est également l'époque où se manifeste un discret magmatisme basique dans les flyschs occidentaux et un important volcanisme andésitique tout au nord-est (cf. fig. 24.7), où le Viséen supérieur est discordant sur des schistes attribués au Viséen inférieur ou même à la base du Viséen supérieur.

- *"phase asturienne"*, manifeste dès le Namurien (qui est régressif dans l'ouest et le centre), paroxysmale durant le Namuro-Westphalien dans les mêmes secteurs (plissement des flyschs viséo-namuriens, métamorphisme), gagnant durant le Westphalo-Stéphanien les régions orientales tandis que les granites intrusifs se mettent en place à l'ouest. Dans les mêmes zones occidentales s'accumulent localement des conglomérats rouges discordants, parfois volcanisés (trachy-andésites).

-*"phase saalienne"* déformant ces molasses continentales westphalo-autuniennes, avant le Trias. L'érosion de la chaîne hercynienne, entamés dès le Houiller, est parachevée avant le Trias supérieur.

LES PROLONGEMENTS EN AFRIQUE DU NORD

Les premiers à envisager sont les prolongements vers les zones internes rifo-kabyles (voir I).

En fait, dans ces éléments de socle disjoints et charriés, on ne retrouve plus les grandes structures anté-triasiques. Par contre, les caractères de l'évolution paléozoique de ces zones internes sont assez analogues à ceux des zones atlaso-mésétiennes les plus orientales. On remarque en effet (Kornprobst, 1974, Bourrouilh *et al.*, 1976):

la réduction apparente du Cambrien; l'opposition entre faciès pélagiques (calcaires) et détritique (flyschs) au Dévonien, et l'existence de lacunes, révélant une activité orogénique acadienne;

l'importance de la "phase bretonne" (flyschs néodévoniens, Viséo-Namurien discordant);

l'âge probablement tardif des phases finales (Permo-Trias discordant).

La chaîne hercynienne devait aussi se prolonger longuement ver l'est de l'Afrique du Nord, ses unités tendant à se mouler sur la plate-forme saharienne (ce qui n'exclut pas des relais en échelon). Les données de sondage s'accordent avec l'idée d'une limite sud des zones orogéniques grossièrement parallèle à la ligne sud-atlasique. On note aussi le maintien prolongé de la Tethys permo-carbonifère (voir en particulier le Permien à fusulines du J.Tebaga, dans le Sud tunisien: Domergeu et al., 1952); la migration progressive (?) de l'orogenèse fini-paléozoïque, déjà sensible entre l'ouest et l'est du Maroc, semble donc se vérifier au-delà, ce qui offrirait un premier élément de corrélation avec la Meseta ibérique (cf. Bard *et al.*, 1971).

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25. LES MAURITANIDES, AFRIQUE OCCIDENTALE - ESSAI DE SYNTHESE

J.-P. Lécorché et J. Sougy

Laboratoire de Géologie dynamique
Faculté des Sciences St. Jérôme, 13397 Marseille Cedex 4, France

INTRODUCTION

La chaîne des Mauritanides (Sougy, 1962b) s'étend sur 1800 km depuis le Sud-marocain au Nord jusqu'à la Sierra Leone au Sud. Elle est déversée à l'Est sur le craton ouest-africain soit découvert (dorsale réguibat au Nord, dorsale de Léo au Sud), soit revêtu de sa couverture, protérozoïque supérieur à dévonienne (bassin de Tindouf au Nord, bassin de Taoudeni au centre et au Sud, bassin bové au SW). Elle s'ennoie à l'Ouest sous le bassin côtier sénégalo-mauritanien, jurassique à quaternaire. Entre ses limites orientales et occidentales, la partie visible de la chaîne ne dépasse pas 200 km de large, même au Sud où elle se dédouble en deux branches (fig. 25.1).

GEOLOGIE REGIONALE

Du Nord au Sud, la chaîne se répartit longitudinalement en cinq tronçons dont les rapports ne sont pas toujours faciles à établir.

1. Le tronçon nord, au N de la dorsale réguibat, comprend les bordures plissées du bassin de Tindouf (Zemmour: Sougy, 1964) et la région des nappes du Sahara occidental ex-espagnol (Sougy, 1962a; Sougy et Bronner, 1969). Il est orienté NNE-SSW. Peut-être les chevauchements de même orientation de l'extrême-Ouest de l'Anti-Atlas (Mazéas et Pouit, 1968) en font-ils également partie.

2. Le tronçon centre nord, orienté NW-SE, entre la dorsale réguibat et le complexe dunaire de l'Aouker, est le domaine de la série d'Akjoujt" (Blanchot, 1955), où ont été mis en évidence les charriages (Tessier *et al.*, 1961) et dont la moitié orientale est étudiée par Lecorché (thèse en cours).

3. Le tronçon central, sub-méridien, s'étend avec une belle continuité de L'Aouker au fleuve Sénégal. C'est "l'arc Bakel-Moudjéria" étudié par Lille (1968), Chiron (1973) et Dia (thèse en cours).

4. Le tronçon sud est caractérisé par la bifurcation de la chaîne entre le fleuve Sénégal et le bassin bové. Une branche se place dans le prolongement sud du tronçon central N-S, tandis que l'autre marque une nette inflexion vers l'WSW, en direction de l'océan. L'essentiel de ce tronçon, situé au Sénégal oriental, a été étudié par Bassot (1966). Ses relations avec le tronçon central font actuellement l'objet de travaux de A. Le Page.

5. Le tronçon extrême-sud correspond à la réapparition de la branche sud-méridienne au S du bassin bové, en Sierra Leone (Allen, 1965, 1968, 1969). Les auteurs du présent article n'en ont, pour le moment, qu'une connaissance bibliographique.

D'Est en Ouest, on peut distinguer transversalement quatre grandes régions:

1, L'AVANT-PAYS;
2, LA ZONE EXTERNE;
3, LA ZONE DES FENÊTRES, SITUEE EN POSITION SENSIBLEMENT MEDIANE;
4, LA ZONE INTERNE.

L'AVANT-PAYS

L'avant-pays est constitué, du Nord au Sud, par:

- l'Anti-Atlas occidental,
- l'extrémité occidentale du bassin de Tindouf et le Zemmour noir,
- la bordure occidentale de la dorsale Réguibat, revêtue localement au voisinage de la chaîne d'une mince couverture paléozoïque ne débutant qu'à l'Ordovicien supérieur et pouvant aller jusqu'au Dévonien,
- les massifs tabulaires occidentaux du bassin de Taoudeni: Adrar, Taganet, Assaba, Tambaoura,
- le bassin bové, par rapport à la branche occidentale de la chaîne,
- la bordure ouest de la dorsale de Léo en Sierra Leone, par rapport à sa branche orientale.

Le socle plissé, métamorphisé, largement migmatisé et granitisé, est définitivement cratonisé depuis 1600 Ma[1], âge des granitisations ultimes (Vachette *et al.*, 1975). Ses éléments stratigraphiques les plus anciens (Amsaga, Rhallaman) ont été datés de près de 3000 Ma (Vachette et Bronner, 1975), d'autres assimilés au Birrimien, sont antérieurs à 2000 Ma. Ce socle supporte en discordance fondamentale des formations sédimentaires gréso-schisto-calcaire. Les plus anciennes, qui sont datées d'environ 1000 Ma (Bassot *et al.*, 1963), constituent la base de la série 1 de Trompette (1973), restée extraordinairement subhorizontale, sauf en bordure de la chaîne. Il faut aller loin à l'Est pour retrouver cette série impliquée dans les plissements pan-africain du Hoggar occidental.

La série 1, essentiellement détritique dans l'Affolé (Aymé *et al.*, 1962; Bense, 1964) et la Tambaoura (Bense, 1964; Bassot, 1966), est caractérisée, au Nord, par des épisodes dolomitiques à Stromatolites, rares dans le Zemmour (série d'El Thlethyate, Sougy, 1964), nombreux en Adrar (Monod, 1952; Trompette, 1973) et dans le Hank (Marchand *et al.*, 1972) et, au Sud, par une base calcaire à Stromatolites dans le secteur de Segou-Madinakouta (Bassot, 1966). Une chrono-stratigraphie, fondée sur ces Stromatolites, a été tentée (Bertrand-Sarfati, 1972; Bertrand-Sarfati et Trompette, 1976) et s'est avérée comparable à celle du Riphéen supérieur de l'Oural.

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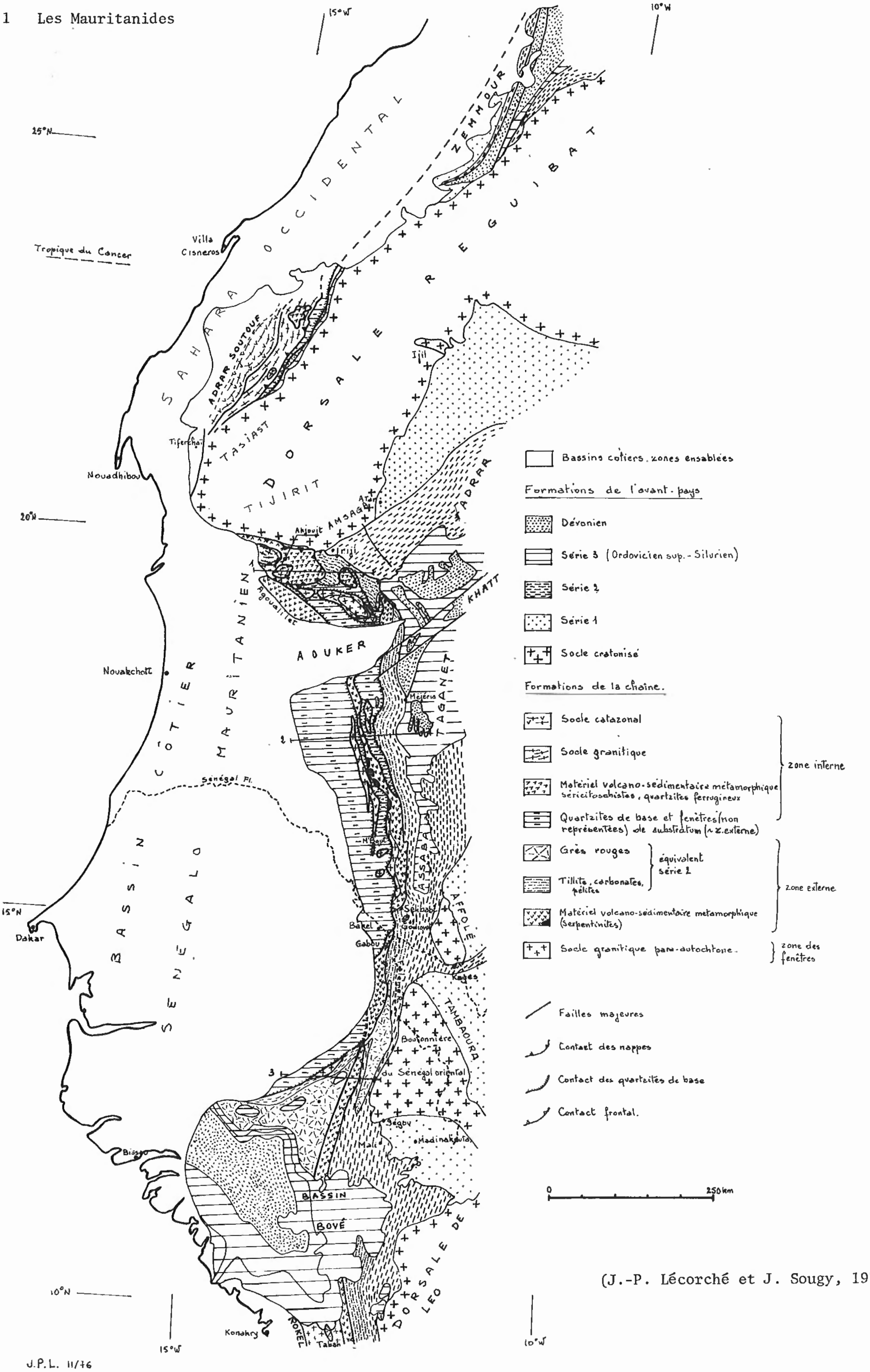
[1] Tous les âges sont donnés avec $\lambda^{87}Rb$ 1,47, 80^{-11}. y^{-1} (roches totales ou minéraux)

[2] Âge recalculé pour $\lambda^{87}Rb$ 1,47. $10^{-11}y^{-1}$

PICG Projet 27, Contribution française n° 4.
Caledonian-Appalachian Orogen of the North Atlantic Region;
Geol. Surv. Canada, Paper 78-13 (1978).

Figure 25.1 Les Mauritanides



(J.-P. Lécorché et J. Sougy, 1976)

La série 2, presque totalement azoïque, débute par une tillite (Zimmermann, 1960), observée en plusieurs points sur plancher glaciaire strié, témoin de la glaciation "éocambrienne". Cette tillite constitue généralement le terme inférieur d'une "triade" (Trompette, 1973) comprenant de bas en haut: formations à caractères glaciaires, calcaire à barytine et silexites, souvent un peu phosphatées. La discordance de la série 2 sur la série 1 n'est pas encore datée avec précision. Plus ancienne, en Adrar, que 580 Ma[2] (Clauer, 1975), âge mesuré au-dessus des formations glaciaires, on la situe généralement autour de 650 Ma. Vers le Sud de la chaîne, des formations volcaniques ou volcano-sédimentaires acides et basiques sont associées aux niveaux de base ou les précèdent.

Au-dessus de cette triade, la sédimentation, d'abord constituée de shales verts, passe progressivement à des formations rouges, shales et grès fins, de nature continentale. Ensuite viennent des faciès gréseux marins transgressifs, soulignés par des niveaux à Scolites. Dans l'Adrar, ils contiennent des Brachiopodes inarticulés attribués par Legrand (1969) à la limite Cambrien-Ordovicien; c'est un premier repère paléontologique. Les séries rouges se situent donc vers la fin du Précambrien ou pendant le Cambrien.

La série 3 (Ordovicien supérieur et Silurien) débute-elle aussi par des formations tillitiques gréso-argileuses qui soulignent une discordance de ravinement (Sougy, 1955, 1964) d'origine glaciaire (Sougy et Lécorché, 1963) bien décrite dans le Sahara algérien (Beuf *et al.*, 1972) et occidental (Deynoux *et al.*, 1972; Trompette , 1973), donc certainement due à une calotte glaciaire de grande extension. Un discordance de ravinement assimilée au même épisode glaciaire a été datée de l'Ashgillien supérieur dans le Sud-marocain (Destombes, 1968a et b).

C'est cet âge fini-ordovicien (environ 440 Ma) qui est généralement admis pour les autres régions, en l'absence de repères chronostratigraphiques. Mais on ne peut, par exemple, critiquer J.-P. Bassot, au Sénégal oriental, de situer ses "grès blancs" à la base du Silurien en raison des âges radiométriques fini-ordovicien obtenus pour leur substratum. Cette coupure se complique d'une discordance angulaire discrète dans le Taganet (Dia *et al.*, 1969), plus nette vers le Sud, dans l'Assaba (Le Page, 1976) et surtout au Sénégal oriental et en Sierra Leone (Allen, 1965). Cet "Ordovicien supérieur" à matériel détritique complexe comporte des surfaces de ravinement internes. Il est surmonté par les sédiments gréseux, argileux et calcaires, fossilifères, du Silurien et du Dévonien, ce dernier connu jusqu'au Frasnien compris. Des sédiments carbonifères, marins puis continentaux sont également présents à l'intérieur des deux bassins de Tindouf et de Taoudeni mais ne sont pas connus au contact ou à l'intérieur de la chaîne. Il est donc impossible de préciser la position chronologique de ces dépôts par rapport à l'orogenèse.

Du point de vue structural, le bassin de Tindouf est orienté E-W; la dorsale réguibat est allongée NE-SW, de même que le vaste pli de fond (bassin dévonien) du bassin de Taoudeni; le bassin bové est orienté NNW-SSE. Ces différents ensembles sédimentaires tabulaires sont interceptés à leur actuelle bordure occidentale par des plis de couverture dont l'orientation est sensiblement parallèle à celle de la chaîne:

- NNE-SSW dans le Zemmour (bassin de Tindouf) et le Sahara occidental (NW de la dorsale réguibat), et peut-être faut-t-il y lier les plis écaillés, de même direction, de l'extrême-Ouest de l'Anti-Atlas,

- NW-SE subméridienne de l'Adrar à l'Assaba (bassin de Taoudeni),

- au S du fleuve Sénégal, NE-SW à l'E de la branche est, qui passe sous le bassin bové, ainsi qu'en bordure occidentale de celui-ci, que semble bien intercepter la branche sud-ouest.

Cette zone plissée de l'avant-pays est de largeur variable (1 à 30 km). Les plis qui, à l'Ouest, sont parfois chevauchants vers l'Est (chaîne de Dhlou au Zemmour) s'amortissent à l'Est. Ils affectent généralement des terrains appartenant aux séries 2, 3 et dévonienne, au moins jusqu'au Frasnien inclus. Le Dévonien plissé qui n'était connu qu'au Zemmour, au Sahara occidental ex-espagnol, en bordure de l'Adrar et dans le Taganet, a récemment été trouvé au S de l'Assaba, à Godiovol, au N du fleuve Sénégal (Crévola *et al.*, 1974).

LA ZONE EXTERNE

La zone externe ne s'individualise nettement qu'au S de la dorsale réguibat et forme alors une bande quasi continue jusqu'au S de la chaîne. Elle est occupée par des terrains d'origines diverses: sédimentaires, volcano-sédimentaires ou volcaniques, acides et basiques, voire ultra-basiques. Ces terrains sont plissés et plus ou moins métamorphisés; des faciès de la mésozone sont connus au S de l'Aouker, dans les formations volcano-sédimentaires, alors que les formations sédimentaires ne dépassent pas un stage épizonal léger qui a d'ailleurs permis d'y reconnaître des niveaux repères comme celui de la mixtite "éocambrienne". Du Dévonien fossilifère plissé y a également été observé au S de l'Aouker (Dia, 1974) en position normale sur des grès ordovico-siluriens probables. Le matériel sédimentaire comporte donc de la couverture Paléozoïque.

Au N de l'Aouker, la zone externe repose en contact anormal peu penté sur la bordure occidentale plissée et morphologiquement en relief de l'avant-pays (Lécorché et Sougy, 1969). Au S de l'Aouker, le contact est mal visible. Pour certains géologues la couverture, même métamorphique, appartient à l'avant-pays (Chiron, 1973; Lille, 1968; Bassot, 1966) et repose en discordance sur le matériel volcano-sédimentaire qui caractérise une "zone axiale" plissée et métamorphisée précambrienne (Chiron, 1973), témoin d'un sillon intracratonique autochtone précambrien. Il nous semble plutôt que cette "zone axiale" et la partie métamorphique de la couverture ont une origine allochtone par rapport aux formations non métamorphiques de l'avant-pays qu'ils côtoient.

LA ZONE DES FENETRES

En position sensiblement médiane dans la partie visible de la chaîne, apparaît, depuis le N de l'Aouker jusqu'au fleuve Sénégal, un chapelet d'affleurements en fenêtres, souvent allongés suivant une direction sub-méridienne, de matériel cristallin et crystallophyllien comparable, au Nord, au socle de la dorsale réguibat (Amsaga, Tasiast), et au Sud, au socle birrimien (2 000 Ma), comme il en existe par exemple, à proximité, dans la région de Kayes.

Certaines de ces fenêtres conservent parfois des témoins de leur couverture sédimentaire sous forme d'un tégument gréseux plissé, à peine décollé, et légèrement métamorphique (fenêtre de Bou Naga au N de l'Aouker). Bien que cette association confère *à priori* à ces socles un caractère autochtone, des arguments souvent d'ordre géométrique amènent à envisager des déplacements différentiels de blocs vers l'Est.

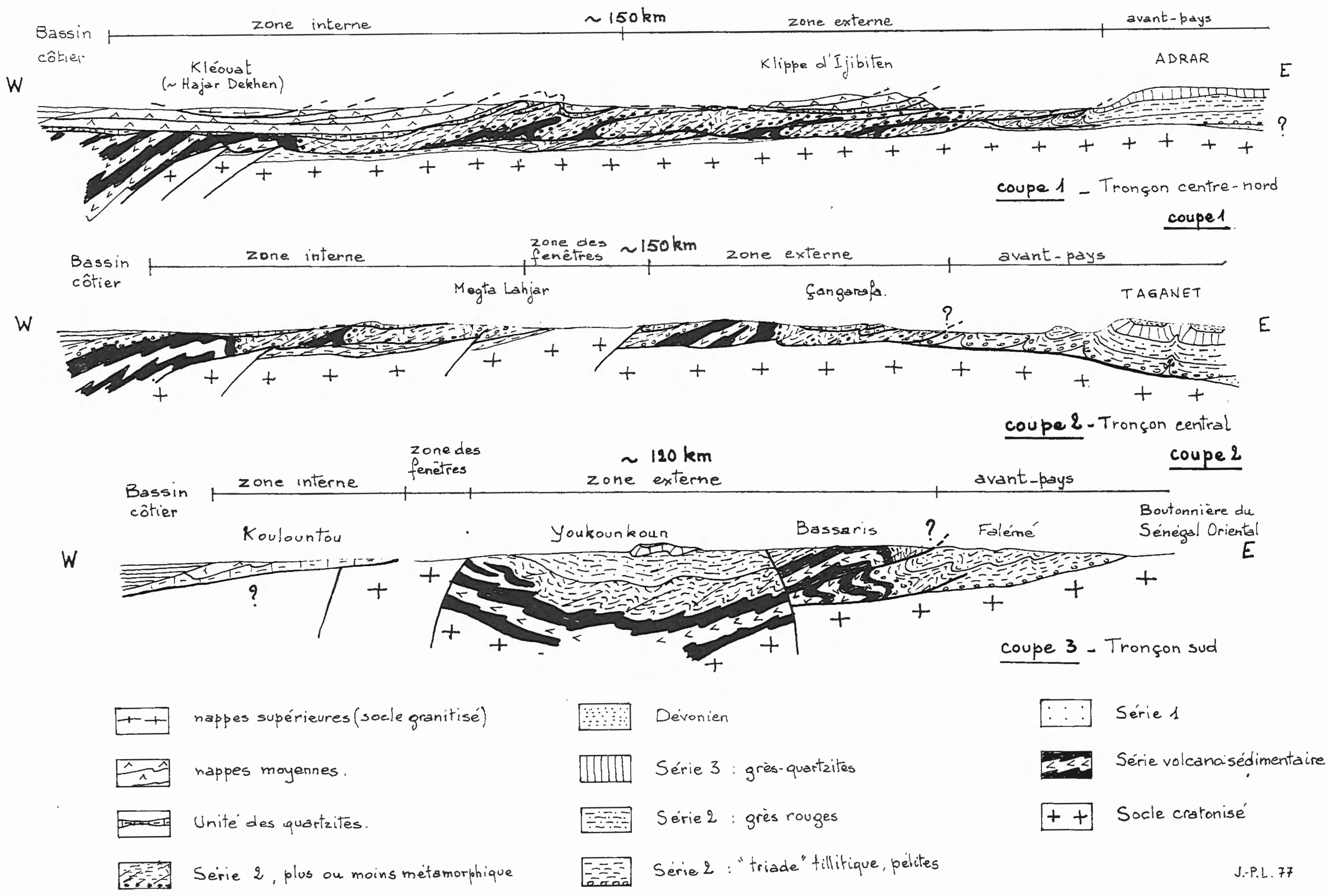


Figure 25.2 Coupes interprétatives des tronçons centre-nord, central et sud des Mauritanides.

Il s'agirait donc plus de la remontée de panneaux du substratum, revêtus ou non de leur couverture, que de véritables "culminations".

Dans d'autres cas, notamment à la latitude de Méjéria, l'écrasement du matériel suggère un écaillage et des déplacements plus importants du socle. Les deux fenêtres situées au S de M'Bout, dont seule l'enveloppe présente un très fort écrasement, constituent un cas intermédiaire.

LA ZONE INTERNE

C'est essentiellement la zone des grandes nappes. On y distingue le plus souvent très schématiquement:

- une unité de base ou nappe inférieure formée uniquement de quartzites divers grossiers à très fins, souvent mylonitiques, mais parfois peu déformés et conservant des structures sédimentaires (Lécorché, 1969). Ces quartzites sont, dans l'ensemble, peu métamorphiques et fortement étirés vers l'Est.

- des nappes moyennes de matériel épizonal à mésozonal rétromorphosé, plissé, comprenant des quartzites à séricite et séricitoschistes, des carbonates et des quartzites ferrugineux à hématite ou à magnétite généralement rubanés, des prasinites et amphibolites chloritisées, des chloritoschistes, des calc-chloritoschites, des épidotites et des diorites quartziques.

- des nappes supérieures constituées de lames de socle cratonisé de nature très comparable à celui de l'Amsaga dans la dorsale réguibat (Giraudon et Sougy, 1963), mais très mylonitique. Ces unités de socle sont coiffées par des nappes de gabbro à hypersthène et charnockites (Sougy et Bronner, 1969), uniquement visibles au Sahara occidental et qui constituent le terme le plus élevé actuellement connu dans l'édifice.

Les quartzites de l'unité de base ont une extension quasi générale. A l'E d'Akjoujt, ils reposent en troncature basale sur les terrains sous-jacents (Lécorché, 1973). Cette disposition qui n'est pas partout admise, est notamment reconnue au S de l'Aouker par Dia (1977). Ces quartzites forment une véritable semelle, parfois replissée, pour les nappes qui les recouvrent. On en retrouve des témoins reposant aussi bien directement sur les fenêtres de socle que sur la zone externe, parfois jusqu'au contact de l'avant-pays plissé.

Les nappes moyennes ne sont actuellement reconnues que dans la partie nord de la chaîne, dans la région d'Akjoujt essentiellement et aussi au Sahara occidental ex-espagnol où, en l'absence de zone externe, elles reposent, par l'intermédiaire de la nappe de quartzite, directement sur le Dévonien ou le Siluro-Ordovicien plissés de l'avant-pays (W d'Aouker). Leur témoin le plus oriental se situe à l'E d'Akjoujt, conservé dans une petite klippe synclinale (massif d'Ijibiten) (fig. 25.2, coupe 1).

Les nappes supérieures ont la même répartition mais une extension encore plus réduite du fait de l'érosion et de la couverture mésozoïque à l'Ouest. Elles ont tout de même un développement probablement assez important dans l'ex-Sahara espagnol.

Signalons enfin au SW du tronçon d'Akjoujt une bande très redressée de formations volcano-sédimentaires, d'orientation NW-SE, comprenant quartzites à épidote et épidotites, carbonates, calcschistes, roches vertes, plagioclasites. C'est la série d'Agoualilet de Marcelin (1967). Ces terrains ne semblent pas affectés par la tectonique tangentielle, malgré les analogies de faciès qu'ils présentent tant avec les formations volcano-sédimentaires de la zone externe qu'avec le matériel de certaines nappes plus internes. Mais le problème n'est pas résolu, l'hypothèse d'un matériel de nappe redressé ne pouvant être écartée.

HISTOIRE DE LA CHAÎNE

La difficulté majeure pour aborder l'histoire de cette région réside dans l'absence de repères chronologiques précis, qu'il s'agisse de l'avant-pays ou des formations de la chaîne proprement dite. On est donc généralement réduit à des estimations.

AGE DES MATERIELS DE LA CHAÎNE

Dans la zone externe, une "triade" qui débute par des formations très comparables aux formations glaciaires de la base de la série 2 du bassin de Taoudeni, permet de "corréler", au moins en partie, la série plissée avec cette série 2 et d'assimiler à la série 1, notamment dans le tronçon central, les grès sous-jacents. On reste néanmoins dans l'indétermination au point de vue âge, puisque la série 2 se serait déposée dans un vaste intervalle de temps entre environ 650 et 440 Ma, période qui couvre l'Eocambrien, le Cambrien et l'Ordovicien. L'indétermination est encore plus grande en ce qui concerne le "substratum" volcano-sédimentaire de cette série, puisqu'on ignore s'il s'agit d'un terme de la série 2 plus ancien que la tillite, d'un équivalent latéral de la série 1 ou d'un ensemble plus ancien. La fourchette s'établit dans ce cas entre 1600 Ma, date de stabilisation du craton ouest-africain et 650 Ma, date approximative attribuée à la "tillite".

Le matériel des fenêtres est constitué du craton sous-jacent, donc offre la même fourchette d'âge que celui-ci. On ignore cependant l'âge exact des intrusions alcalines qui l'affectent au N de l'Aouker. Or on connaît des trachytes très analogues dans le volcano-sédimentaire de la zone externe.

La zone interne fournit deux types de repères:

- à sa base, l'unité des quartzites dont on peut, à titre d'hypothèse, faire un équivalent des grès de base de la série 3 (Lécorché, en cours) ce qui impliquerait un âge ashgillien.

- à son sommet, des lames de matériel cratonisé, plus vieux que 1600, ou même 2700 Ma.

Entre les deux, les nappes moyennes, dont la patrie reste inconnue, ont fait l'objet d'hypothèse diverses: équivalents latéraux et autochtones du Cambro-Ordovicien du bassin de Taoudeni (Bassot et Delpy, 1960), équivalents du volcano-sédimentaire "enraciné" d'Agoualilet *pro-parte* et de séries reliques du craton ouest-africain (Marcelin, 1967), par exemple. On conçoit qu'il sera difficile de lever de telles indéterminations. Aussi, pour l'instant, est-il plus raisonnable de dire que l'âge du matériel de ces nappes est probablement situé dans une fourchette 1600-350 Ma, puisqu'elles ne sont pas granitisées.

Un programme de datations radiométriques est envisagé depuis plusieurs années. Il attend pour le moment une détente du climat politique pour être mis en oeuvre.

AGE DES PLISSEMENTS

L'âge fini ou post-dévonien de la tectonique qui a définitivement structuré la chaîne des Mauritanides est suggéré par un certain nombre d'arguments:

- présence de Frasnien plissé dans la zone de bordure de l'avant-pays, au contact de la chaîne et parallèlement à elle;

- présence de Dévonien plissé dans la zone externe;

- âge radiométriques Rb/Sr mesurés sur des biotites des nappes supérieures de la région d'Akjoujt - 327 Ma, 243 Ma (M. Bonhomme, 1962); sur des muscovites de quartzites et de micaschistes au S du fleuve Sénégal - 357 et 358 Ma (Bassot *et al.*, 1963), ainsi que sur des biotites - 209 Ma. Mais il s'agit plutôt d'âges de refroidissement qui peuvent varier d'une région à l'autre, que d'âges de recristallisation de ces minéraux.

Toutefois, en allant du Nord au Sud, il apparaît dès le Taganet une tectonique antérieure soulignée, dans l'avant-pays, par la discordance angulaire au moins locale des formations glaciaires fini-ordoviciennes sur les séries cambro-ordoviciennes sous-jacentes légèrement plissées. Mais cette tectonique est nettement mieux marquée dans la chaîne, dont la branche sub-méridienne plissée passe sous les formations tabulaires fini-ordoviciennes, siluriennes et dévoniennes du bassin bové, elles même cependant plissées au contact de la branche occidentale. Des âges radiométriques Rb/Sr (Bassot et al., 1963) de 430, 435, 440 Ma ayant été mesurés dans ces formations, il semble donc logique d'en déduire l'existence d'une phase calédonienne, située plus précisément vers la fin de l'Ordovicien, mais avant la glaciation fini-ordovicienne dont les dépôts apparaissent discordants. Ce serait ainsi une phase taconique. Mais la mise en évidence, dans le prolongement de cette branche méridienne "taconique", du tronçon des Rokelides (Allen, 1965, 1968, 1969) daté de 550 Ma pose néanmoins le problème de la superposition de cette tectonique calédonienne sur une tectonique plus ancienne, pan-africaine.

Ainsi, le matériel de la chaîne a subi des phases tectoniques superposées dont seules les dernières peuvent être rapportées aux charriages fini ou post-dévoniens. Une partie de ce matériel peut donc avoir été précédemment impliqué dans des tectoniques pan-africaine (tronçon central, Chiron, 1973) ou calédonienne (tronçon d'Akjoujt, Marcelin, 1964).

Un certain nombre d'événements ont été enregistrés soit directement par la chaîne, soit indirectement par son avant-pays. On peut tenter de reconstituer leur succession, mais en sachant que toute datation précise paraît pour le moment prématurée.

L'événement le plus ancien semble être lié à la présence en différents points de la chaîne de matériel volcanique ou volcano-sédimentaire acide, basique et ultra-basique formant le substratum des formations glaciaires éocambriennes. Pour Chiron (1973), au Sud de l'Aouker, ce matériel correspondrait aux vestiges d'une chaîne intracratonique précambrienne autochtone. Mais on peut aussi penser que ce matériel, polyphasé, est allochtone, la chaîne précambrienne terminale s'étant plissée plus à l'Ouest, probablement sous la bordure actuelle du bassin côtier. Cette bordure est marquée en gravimétrie (Blot *et al.*, 1962; Crenn et Rechenmann, 1965) par un axe lourd encore inexpliqué. Un témoin marginal pourrait en être, au N de l'Aouker, la série d'Agoualilet, probablement en fenêtre, dans la zone interne.

Le matériel de cette chaîne contient des reliques de matériel affecté par une première phase métamorphique synschisteuse. Une deuxième phase synchisteuse liée à des plis déversés vers le Nord-Est, à schistosité de flux plan-axial, affecte l'ensemble.

La première phase, (Φ1) pourrait être précambrienne comme le propose J.-C. Chiron, mais la deuxième, (Φ2) qui intéresse du matériel encore mal daté mais probablement éocambrien et cambrien, serait peut-être contemporaine de la tectonique pan-africaine de Kennedy (1964) et affecterait vers le Nord. Le prolongement, ensuite déplacé vers l'Est, de la chaîne des Rokelides d'Allen (1969).

Le plissement des termes inférieurs marins de la série 2 et de leur substratum volcano-sédimentaire déjà plissé et métamorphisé, aurait donc débuté lors du deuxième épisode métamorphique et aurait été suivi d'une période de démantèlement dont témoignent, d'après Allen (1969), le grès de Taban, au S du bassin bové et, pour Bassot (1966), les grès rouges du Sénégal oriental. Ces grès sont observés dans les Rokelides en discordance angulaire sur la formation de Marampa, en bordure de l'Assaba (Le Page, 1976) sur les termes inférieurs de la série 2 et au Sénégal oriental en discordance cartographique sur ces mêmes termes (Bassot, 1966). Des formations rouges détritiques (cambriennes?) appartenant à la série 2 du bassin de Taoudeni, existent dans l'avant-pays, mais leur caractère beaucoup plus évolué et leur provenance très différente (SE) semblent exclure qu'elles puissent être liées à cet orogène.

Au Sud (Sierra Leone, Guinée et Sénégal oriental), ces grès rouges sont, à leur tour plissés avec leur substratum avant le dépôt des grès blancs fini-ordoviciens, donc avant la nouvelle glaciation qui a marqué cette période. Cette phase coïncide avec un troisième phase de plissement, Φ3, accompagnée, au moins au SE du tronçon d'Akjoujt, de charriages vers le NE.

Après ces deux dernières phases, Φ2 et Φ3, qui seraient donc à rapporter à un cycle pan-africain - calédonien, la glaciation fini-ordovicienne, générale dans l'Ouest africain, s'est étendue à la zone de la chaîne, au moins dans sa partie sud. Là, les "grès blancs" qui lui correspondent n'ont pratiquement plus été affectés sauf au Nord-Ouest du bassin bové où ils ont été quelque peu plissés. Il sont signalés plus au Nord, au Sud de l'Aouker (Dia *et al.*, 1974), associés localement à du Dévonien, plissés et, semble-t-il, en contact chevauchant sur les formations de la zone externe ou de l'avant-pays plissé (Dia, 1977). Au N de l'Aouker un ensemble de quartzites épimétamorphiques et fortement mylonitisés sont situés à la base des nappes internes et en discordance tectonique sub-horizontale sur toutes les formations antérieures (Lécorché, 1973). On est alors tenté d'étendre à ces deux régions, le schéma du Sud et à envisager l'extension de la glaciation et de ses dépôts au domaine précédemment plissé. Dans cette hypothèse et compte tenu du métamorphisme qui les affecte, ces dépôts clastiques ont dû s'étendre fort loin vers l'Ouest et être éventuellement recouverts par les transgressions silurienne et dévonienne.

Les nappes de la zone interne marquent un nouvel épisode, nouveau par le style et probablement aussi par l'origine. Il semble en effet logique de rechercher leur patrie au-delà des régions qu'elles recouvrent, ce qui implique l'existence d'un domaine de sédimentation encore plus occidental, probablement en bordure de l'actuel continent. Le matériel, d'âge inconnu, très comparable à celui du substratum de la série 2 rencontré dans les unités précédentes peut-être lui même en partie repris a subi un premier plissement de type isoclinal accompagné d'un métamorphisme synschisteux (Marcelin, 1964), puis a été plissé et charrié en nappes du deuxième genre à troncatures basales sur un avant-pays ordovico-dévonien, probablement aplani, qu'il a laminé, étiré, parfois plissé, remodelant et entraînant les structures antérieures sous-jacentes jusqu'à chevaucher leur couverture dévonienne. L'extension du phénomène serait alors mesurable au degré de mylonitisation et de recristallisation de la semelle quartzitique, seule conservée dans la plupart des régions. Ceci implique que, de la dorsale réguibat à l'Assaba, ces nappes ont sensiblement atteint l'actuelle limite des plateaux primaires, alors qu'à partir du fleuve Sénégal, où leur front s'infléchit nettement vers l'Ouest, elles laissent à découvert les zones antérieurement structurées.

CONCLUSION

La chaîne des Mauritanides n'est étudiée que depuis une quinzaine d'années et par un petit nombre de géologues qui s'attachent en même temps à l'étude du craton voisin et de sa couverture. On ne s'étonnera donc pas si les interprêtations,encore bien hésitantes, évoluent d'un essai de synthèse au suivant. Celle que nous proposons ici repose encore sur beaucoup d'hypothèses insuffisamment étayées, mais permet d'envisager une articulation entre les différents tronçons d'une chaîne dont on postule du même coup l'unité. Schématiquement et sous toutes réserves concernant l'âge exact des plissements, il semble que l'on puisse considérer cette bande plissée comme issue de la superposition de plusieurs événements tectonique successifs: pan-africain, calédonien et hercynien précoce.

Les premiers marquent l'évolution d'une zone mobile qui se serait individualisée lors du cycle pan-africain directement à l'W du domaine actuellement visible occupé par la chaîne, suivant l'axe des anomalies positives indiquées par la gravimétrie. Le dernier événement serait dû à l'apparition encore plus à l'Ouest, peut-être en bordure de l'actuel continent, d'une nouvelle zone mobile dont l'évolution aurait abouti à la mise en place des nappes post-frasniennes. Dans les deux cas, l'absence dans les formations étudiées, malgré l'ampleur des charriages, de matériel océanique caractérisé, incite à penser que ces zones mobiles se seraient individualisées dans le craton. Par contre, l'absence de granitisations contemporaines des événements tectoniques pourrait souligner le caractère marginal de la bande actuellement visible de la chaîne. Enfin la question de l'origine du socle ancien des nappes les plus élevées de l'édifice, pose, comme pour les Calédonides de Scandivanie, le problème de la présence d'un socle à l'Ouest. Mais l'existence d'un tel socle serait ici compatible avec les hypothèses ci-dessus formulées.

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