David L. Bruton (ed.)

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aspects of the Ordovician System

British Series	Atlantic Grapt. zones	Baltic Series	N.American Series	N.Atlantic Conodont zones
ASHGILL	*persculptus*	HARJU	CINCINNATIAN	– – – – – – – –
	extraordinarius			*Amorphognathus ordovicicus*
	anceps			
	complanatus			
CARADOC	*linearis*			*Amorphognathus superbus*
	clingani	VIRU	CHAMPLAINIAN	
	multidens			*Amorphognathus tvaerensis*
LLANDEILO	*gracilis*			*Pygnodus anserinus*
LLANVIRN	*teretiusculus*			*Pygodus serrus*
	murchisoni			*E.suecicus*
	'bifidus'	OELAND		*E.variabilis*
ARENIG	*hirundo*		CANADIAN	*M.parva* *B.navis* *B.triangularis*
	extensus			*Oepikodus evae* *P.elegans* *P.proteus*
	?			
TREMADOC	?			*P.deltifer* *C.angulatus*
	Dictyonema spp.			?

Correlation of the British, Baltic and North American series with the Atlantic graptolite and conodont zones. Modified from Bergström 1977 and Jaanusson 1979.

Aspects of the
Ordovician
System

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Edited by: DAVID L. BRUTON

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Editorial

The IV International Symposium on the Ordovician System was held in Norway in August 1982 and followed similar meetings in France (1971), Great Britain (1974) and the U.S.A. (1977). The Symposium was sponsored and partly financed by the International Union of Geological Sciences (IUGS) and numerous oil companies, but it was left to the "Friends of the Ordovician" to organise. This group has no constitution, and membership is open to all with interest in "things Ordovician". The friends, via Reuben J. Ross Jr. and Stig M. Bergström approached me in 1977, suggested the date and venue of the fourth symposium, and gave me a free hand in organising the programme and meetings. From the start I came to rely heavily on the support of Valdar Jaanusson and much later especially Henry Williams. Valdar Jaanusson and I decided that part of the programme should be devoted to the presentation of those aspects of the Ordovician System such as climate, volcanicity, global transgressions and regressions etc., often neglected yet so important in their bearing on the more commonly studied stratigraphy and faunas. The former aspects have been given scant attention, so much so that some of the lecturers invited to present individual themes had to start preparation from scratch often using only a few published results.

This book contains twenty of the invited thematic lectures presented on two days of the Symposium and it is hoped that their content and style of presentation will be of general interest to the specialist and non-specialist alike.

I take this opportunity of thanking all authors for agreeing to lecture and for meeting the December '82 dead-line set for submission of manuscripts. These have since been assessed by a number of referees, some of whom wish to remain anonymous. Nevertheless I wish to thank Henry Williams, Alan W. Owen and David A. T. Harper for their assistance. Hans Arne Nakrem kindly re-drafted some of the original text-figures and Elisabeth Sundholm re-typed sections of manuscript where necessary.

All views expressed in the papers represent the opinion of the authors as of 1 June 1983, after which date no amendments or additions have been made to manuscripts.

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What is so special about the Ordovician?

By VALDAR JAANUSSON

Some of the most distinctive features of the Ordovician Period appear to have been (1) an extreme thalassocratic distribution of the land and sea, with associated extensive development of epicontinental carbonate sediments, (2) profound changes in the composition of skeleton-bearing faunas during the Period because of the appearance or diversification of numerous major taxonomic groups, and (3) a pronounced biogeographical differentiation. The Ordovician south pole had an ice cap whereas in the northern polar area no ice cap was developed. The important change in the biogeographical pattern from the Ordovician to the Silurian might have been associated with the disappearance of the southern ice cap.

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The purpose of this paper is to outline those general aspects that make the study of Ordovician fossils and rocks both challenging and rewarding. Emphasis is placed on features which may be said to be particularly distinctive to the Peroid as a whole.

The Ordovician Period was pronouncedly thalassocratic, in the sense that epicontinental seas had a greater extent than in almost any other period and land areas were relatively small, being represented on most cratons by archipelagos rather than continents. The cratonic land areas were mostly of low relief, the rivers had gentle gradients and low energy of transport. The resulting low supply of terrigenous material to the epicontinental seas facilitated widespread deposition of carbonate sediments, mostly on a sea floor that was almost perfectly flat. It is difficult to find modern counterparts for these conditions, and this in turn renders it difficult to understand some of the depositional processes.

Movements along several plate boundaries gave rise to mountain belts, many as island arcs. Erosion of both these uplifted areas and cratonic land areas adjacent to the Ordovician south pole appears to have been the principal producer of terrigenous sediment. The widespread volcanic activity associated with the plate movements was more intense than during almost any other geological period (see, e.g., Stillman in this volume).

The composition of the skeleton-bearing benthic fauna and flora changed markedly during Ordovician times. In this respect, the Ordovician is one of the most interesting periods in the history of marine faunas. A number of major taxonomic groups appeared or became common and diverse for the first time. The list of such groups is long, but the most important are stromatoporoids, various corals, pelecypods, cephalopods, cystoids, crinoids, echinoids, ostracodes with a calcitic exoskeleton, and bryozoans. Representatives of these groups are either absent or very rare in the Cambrian. Among planktic groups graptoloids appeared and euconodonts became common and diverse.

The successive appearance of new major groups affected the composition of skeleton-bearing benthic associations, so that to a large extent the quantitatively dominant organisms in the later part of the Ordovician belong to taxonomic groups that had either not evolved or were very rare in the earliest part of the Period. The change in the composition of the skeleton-bearing associations is even more accentuated if one takes into account the successive appearance of new taxa at superfamily or order level within major groups that were already common in the Cambrian. Examples are the enteletacean, plectambonitacean, strophom-

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enacean, rhynchonellacean, and atrypoid articulate brachiopods, and trinucleid trilobites. Soon after their appearance these became common to dominant constituents in many assemblages. In the Lower Ordovician, organic reefs are rare, mainly because the principal skeleton-bearing groups which formed the frame in later Ordovician non-algal reefs had not yet acquired the ability, or lacked the selection-induced necessity, to secrete skeletons.

Concomitantly with the successive appearance of the new groups, a biotic organisation of the benthic epifauna developed that characterises many of the subsequent Palaeozoic epifaunas but differs in several respects from the organisation of modern counterparts. The presence of many forms with an ambitopic (attached in juvenile stages, recumbent as adults) or recumbent mode of life make the distinction between soft-bottom and hard-bottom epifaunas less sharp than on the modern sea floor (Jaanusson *in* Jaanusson *et al.* 1979: 272). In modern benthic faunas, sedentary organisms lying free on the substrate are rare, and for this reason a community itself does not produce patches of hard-bottom to the same extent as in the Palaeozoic.

The Ordovician Period was also characterized by an extreme biogeographical differentiation of marine faunas (for a recent summary, see Jaanusson 1979), more extensive than during most other periods. This affected both benthic and planktic faunas, but the degree of differentiation in the planktic graptoloids and conodonts is far weaker than in the benthic faunas. In modern seas the primary factor regulating the distribution of faunal provinces is temperature (see Ekman 1953). According to palaeomagnetic data, the Ordovician north pole was situated somewhere in the present northwestern Pacific ocean, apparently far away from any continental plates (for recent, somewhat conflicting reconstructions of Ordovician geography see Scotese *et al.* 1979; Smith 1981; Smith *et al.* 1981; Spjeldnæs 1981). This implies that there was no northern ice cap, because water in the polar region had free exchange with the vast thermal reservoir of the oceans. This in turn suggests that climatic zones of the Ordovician northern hemisphere were probably poorly defined. On the other hand, the Ordovician south pole was most probably situated on a continent, possibly in northwestern or western Africa. An ice cap was present, and southern climatic zones were well defined. An important event at the end of the Ordovician Period was an extensive glaciation in areas close to the south pole (Beuf *et al.* 1971; Allen 1975), which obviously had important eustatic effects. One of the main reasons for the strong Ordovician biogeographical differentiation appears to have been the pronounced climatic zonation in the southern hemisphere, coupled with restriction of communication between various epicontinental seas.

The pronounced biogeographical differentiation in the Ordovician contrasts sharply with the almost cosmopolitan distribution pattern of the succeeding Llandovery faunas. In fact, the change in the degree of biogeographical differentation at the Ordovician-Silurian boundary appears to be one of the most remarkable events in the history of marine faunas. The reason for this change is not clear. The late Ordovician glaciation itself has been suggested as playing a decisive role in the change of biogeographical pattern (Sheehan 1973, 1975), but the far more extensive Pleistocene glaciations do not appear to have produced effects of a comparable magnitude on marine faunas. A strong reduction or disappearance of the southern polar ice cap at that time would have produced a sufficient effect, particularly if the location of the pole had shifted from a continent to an ocean. However, the latter possibility is not supported at present by reconstructions of Silurian geography. A model for Silurian climate without polar ice caps has also been suggested by Spjeldnæs (1981).

We are still far from understanding Ordovician biogeographical relationships, partly because it is difficult to distinguish between what in the known spatial distribution of faunas is due to ecological factors and what is due to geographical factors. Another reason is that we still lack reliable palaeomagnetic data on the approximative latitudinal position of many Ordovician plates, and in several cases it is not clear what constituted an individual plate.

The pronounced biogeographical differentiation of faunas is also one of the main reasons why it is difficult to correlate and classify Ordovician deposits, not only on an intercontinental scale but even between adjacent regions. With regard to stratigraphical classification, the Ordovician is perhaps the most confused of all periods. The problems of correlation have not been made easier by the fairly recent observations (Jaanusson 1976) that changes in benthic and planktic faunas are often not contemporaneous. Changes in both graptoloid and conodont faunas occur at levels at which no conspicuous change can be detected in local benthic faunas and, likewise, major changes in benthic faunas take place at levels where no corresponding change can be proved in graptolite faunas. The factors that influenced the change in planktic faunas were mostly different from those that affected local benthic faunas.

Many problems remain as a basis for future research on Ordovician rocks and fossils. Some pertinent fields in which our knowledge is particularly unsatisfactory, such as the paucity of reliable palaeomagnetic data, are mentioned above. A solid taxonomic framework is a necessary foundation for stratigraphical, ecological and even palaeogeographical conclusions, and much taxonomic work remains to be done. Large faunas await description or revision, and I would like to emphasise the importance of more extensive studies of calcareous algae. My particular wish for future research is that sedimentologists and palaeontologists should work more closely together and individual workers should combine lithological and faunal studies as a mean of presenting more coordinated interpretations of Ordovician patterns and processes.

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The Ordovician Time Scale – New Refinements

By REUBEN J. ROSS JR. and CHARLES W. NAESER

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Introduction

Seven fission-track dates and one K/Ar date constitute great progress in isotopic dating of the Ordovician Period since the Ordovician Symposium at Birmingham in 1974. Speaking at Birmingham Professor Frank J. Fitch warned that the few Ordovician dates then available would be in need of constant revision because of inadequate geological interpretation and analytical methods. Fitch *et al.* (1976) charged that there are two approaches to the construction of a time scale. The first is the "direct, strictly scientific approach of stratigraphical geology" which requires that the stratigraphic position of the rocks to be dated be fully documented; ideal are "datable sediments, lavas, tuffs, and bentonites intercalated within richly fossiliferous strata". The derived isotopic data should be "precise and unambiguous". At that time no dates met these requirements and without them there could be no "viable Ordovician time-scale".

The second approach is indirect, subjective, and geologically intuitive. All available age data, both good and bad, are weighed against "total geological background". "Although a very poor alternative" this was the only source of any Ordovician dates in 1975, and the method is still very much in evidence, particularly in interpreting emplacement of ophiolite complexes.

Although the paper by Fitch *et al.* (1976) seemed pessimistic when presented, it clearly divided Ordovician isotopic dates into good and not-so-good categories, for which the score was nil to too many. The paper served as an inspiration for a project that was undertaken shortly thereafter in Wales, Shropshire, and southern Scotland (Ross *et al.* 1982). It may also have inspired the critical dating of the Upper Tremadoc Rhobell Volcanics in north Wales by Kokelaar, Fitch, and Hooker (1982).

Dating Stratotypes and other sections

In the autumn of 1976 fourteen British and North American colleagues took 35 samples of bentonites and tuffs interbedded in the classical sections of each of the Series of the Ordovician and Silurian Systems, except the Tremadoc. The gratifying results of 11 samples were reported in abstract form two years later (Ross *et al.*, 1978) and a complete report including all supporting stratigraphic and chemical evidence has recently been published (Ross *et al.*, 1982), greatly enhanced by the contributions of our British colleagues. The entire effort would have failed had not the British contingent unearthed from countless geologic reports the whereabouts of a great many Ordovician and Silurian bentonites and tuffs, many more than those from which we were able to collect, and had they not guided us to them. Also essential was the financial assistance of the National Geographic Society.

The useable dates are included in Figure 1 and Table A (Ross *et al.*, 1982, Table 1. Kokelar *et al.*, 1982). They do not include a date for the base of the Tremadoc or any way to accurately date the Caradoc-Ashgill boundary. They do include significant dates for the Series of the Ordovician and Silurian, all stratigraphically defined to conform to the first category of Fitch et al. (1976).

The Rhobell Volcanic Group, exposed near Dolgellau in northern Wales, is considered to be uppermost Tremadoc on biostratigraphic

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grounds and has been dated by the K/Ar method at 508 ± 11 m.y. by Kokelaar *et al.* (1982). This date supports the fission-track dating of the Llyfnant Flags as a minimum age (Figure 1), but not the assignment of 482–484 m.y. dating to the Arenig-Tremadoc boundary by Gale et al. (1980). This Tremadoc date provides an important extension of the Ordovician and with the new Cambrian dates reported below requires that the base of the Ordovician System should be about 515 m.y. in age. The boundary between the Arenig and Tremadoc should probably be dated at 500–505 m.y.

North America problem dates

Other fission-track dates that do not seem to have been reset have produced puzzles for which we have yet to find the solutions. For a score of years the age of the Caradoc was based on isotopic dating of the *Chasmops* Limestone of Sweden – 450 m.y. (Byström-Asklund *et al.* 1961) and the Carters Limestone of Alabama. A fission track date on the latter is 454 ± 8 m.y. Two other Rocklandian formations in the Mississippi Valley averaged with the correlative Carters are dated by fission tracks at 453 ± 3 m.y. (Ross *et al.*, 1982; 146). By contrast the Actonian of the Onny River is dated at 466 ± 16 m.y. and the Longvillian near Bala 464 ± 13 m.y. Further, a sample collected by S. M. Bergström from the Dalby Limestone of Sweden (zone of *D. multidens* = Mid Caradoc?) is dated at 470 ± 10 m.y. by fission tracks (Figure 1). According to strict statistical practice these dates may not be significantly different, but there is a consistent suggestion that the North American Rocklandian dates are about 10 m.y. younger than the supposedly correlative Caradoc. In this same interval the graptolite zonation in the United States is the source of much disagreement (Ross, 1982; 2; Finney, 1982; 20–21). The shelly fauna of the type Caradoc is virtually unknown in North America. Is it possible that the isotopic dates are calling attention to a biostratigraphic miscorrelation? An attempt to obtain more evidence on this question by collecting two bentonites from the Rocklandian Tyrone Limestone of Kentucky produced fission track dates of 447 ± 15 m.y. and 462 ± 16 m.y., averaging 455 ± 15 m.y. and merely reinforced the problem, not its solution.

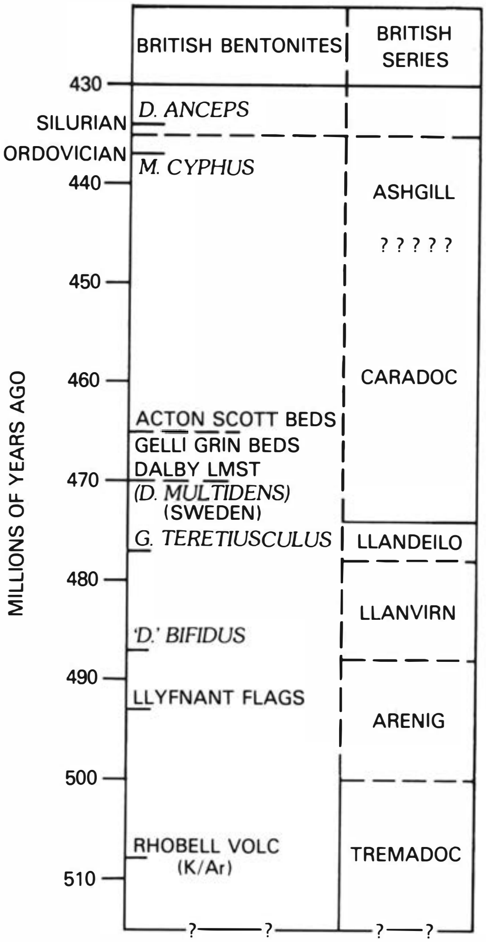

Figure 1 – British Ordovician fission-track dates and deduced ages of Series. Also date from Swedish Dalby Limestone (Zone of Diplograptus multidens*) and from Upper Tremadoc Rhobell Volcanics.*

Chinese Ashgillian and North American Cambrian dates

During the visit to China of five members of the Subcommission on Ordovician Stratigraphy in 1978, Ross collected a sample from the lower Wufeng Shale with the assistance of Dr. Wang Xiaofeng and Dr. Ma Kuogan of the Geological Institute of Yichang. The collection was made on the north side of the Yangtze River and the west side of the Huanling anticline. The fission track date of this sample is 447 ± 10 m.y. If these beds are correctly correlated with the ear-

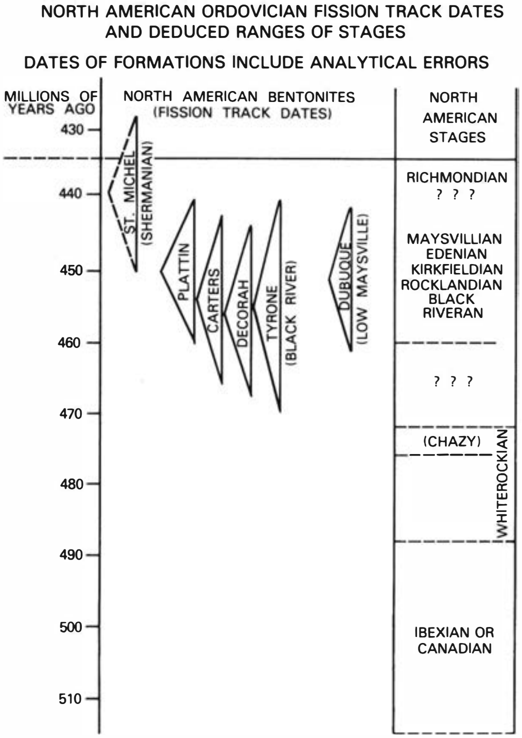

Figure 2 – North American Ordovician fission-track dates and one possible deduction of ages of Stages in the United States.

ly Ashgill, this date suggests that the Rocklandian is equivalent to the youngest Caradoc.

Two new fission track dates, both Middle Cambrian, are important in putting limits on the base of the Ordovician. Within the past two years Naeser has collected and dated a bentonite from the base of the Peach Spring Member of the Muav Limestone at 535 ± 12 m.y. (*Bathyuriscus – Bolaspidella*) and a bentonite from the *Glossopleura* zone of the low Bright Angel Shale at 563 ± 12 m.y. Both collections came from within the Grand Canyon of the Colorado River. As far as we know these are the only stratigraphically controlled Cambrian dates currently available.

Controversy

The publication of the fission track dating of the Series of the Ordovician and Silurian has been greeted almost immediately by controversy. Two examples, one involving yet another dating of the Ballantrae ophiolite complex of southwestern Scotland (Bluck *et al.*, 1980), and the other involving the dating of the Stockdale Rhyolite of northwestern England (Gale *et al.*, 1980), have prompted their authors to reject the fission track dates.

One of the most recent examples of the intuitive approach to dating is the work of Bluck *et al.* (1980), who have used isotopic dating to derive a revised history of the Ballantrae igneous complex. However, they have also noted the presence of a variety of clasts of metamorphic rocks interbedded in a black shale. Lithologic similarity led them to assume that the clasts were derived from the complex and that the shales were of the same age as shales 300 km distant in Ireland. In the Irish shales, not in the Scottish, graptolites of the Arenig zone of *Didymograptus nitidus* have been reported (Dewey *et al.*, 1970; 39–40). The metamorphic aureole of the complex was dated isotopically as 478 m.y. Therefore it was concluded that the date of the middle Arenig is 480 m.y. Here we do not question the isotopic date or that it is important in interpreting some important geologic event, probably in Llanvirn time. Using it to establish the time scale against which one is attempting to date the same sequence of geologic events is surely circular reasoning. Until 1978 this was the pragmatic approach to building an Ordovician time scale.

The Ashgill Stockdale Rhyolite has been dated at 421 ± 3 m.y. by the whole rock method of isotope analysis (Gale *et al.* 1980), resulting in their placement of the Ordovician-Silurian boundary at 418 m.y. Naeser (*in* Ross *et al.*, 1982; 147–149) has commented on this date. Wyborn and others (1982) have derived an isotopic date from the late Silurian early Ludlow Laidlaw Volcanics of southeastern Australia of 421 ± 3 m.y. The Laidlaw is composed of ignimbrite flows with minor airfall tuffs and has highly fossiliferous beds, containing conodonts as well as shelly fossils, both above and below. The fauna is correlated without question with the earliest Ludlow of Great Britain.

Very precise analysis has produced the following dates: K/Ar = 418.5 ± 1.9 m.y: Rb/Sr = 421.1 ± 1.3 m.y.. Rb/Sr = 424.5 ± 7.8 m.y. (whole rock).

Wyborn and coauthors consider 421 m.y. as "a precise estimate for the age of the earliest Ludlow". They comment on the identity of the date for the Stockdale Rhyolite (Gale *et al.*, 1979) and conclude that their very impressive data invalidate the Stockdale date as Ashgillian.

Summaries of previous time-scales

McKerrow *et al.* (1980) have presented the best modern summary of the status of the early Paleozoic time scale. We note with some satisfaction that they have made good use of Ordovician and Silurian dates published by Ross *et al.* (1978, now available with supporting data, 1982). Their graphic presentation includes a traditional distrust of and skepticism about isotopic dates. The horizontal axis, which we might consider a close approach to objectivity, is devoted to isotopic dates in millions of years. The other axis – the subjective axis – shows the relative lengths of stages and periods as estimated by the authors, based on rates of evolution for fossils and on rates of deposition of sedimentary rocks. Because the thicknesses of sedimentary rocks provide an almost instinctive measure of relative time when two similar sequences are compared, the subjective ordinate axis has been evolving since geologists first attempted to gauge the age of the earth. An early and surprisingly successful assembler of such data was Barrell (1917), who applied esti-

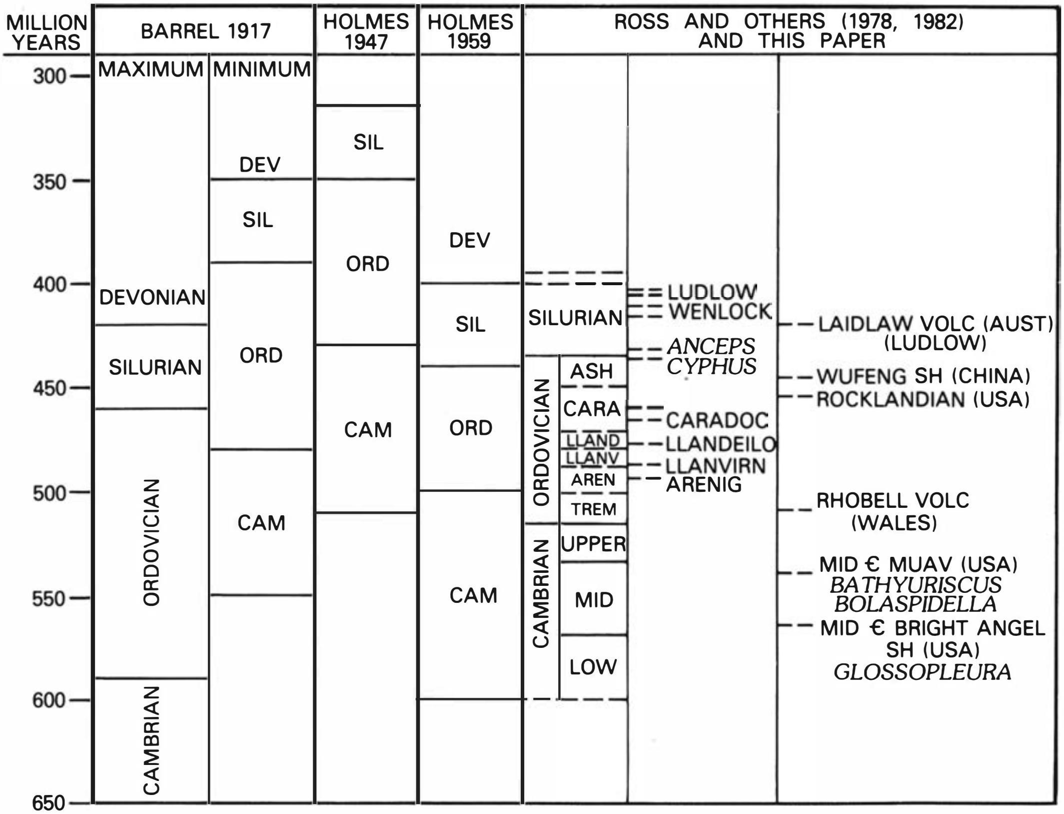

Table A – Comparison of dates for the early Paleozoic of Barrell (1917), of Holmes (1947, 1959), of Ross et al. *(1978 & 1982), and of this paper. Under Ross* et al., *left side shows deduced relative lengths of Series; two right columns show fission-track dates. Date of Laidlaw Volcanics tends to compress possible length of Silurian. Date of Ashgillian Wufeng Shale limits youngest Caradoc. Rocklandian dates correlate with youngest Caradoc. Date of Rhobell Volcanics lengthens duration of Ordovician at expense of Cambrian.*

mated minimum and maximum rates of sedimention to arrive at the estimated duration of each Period. Holmes (1947, see particularly Table III; 1959, 205–206, fig. 2) based his geologic time scale in a similar manner, taking stock of Barrell's efforts, but adding such isotopic constraints as were available. There has been some improvement over Holmes' results, as we have attempted to show in Table A.

McKerrow *et al.* (1980, Fig. 1, no. 10) have shown that our fission track date on the Acton Scott beds (466 ± 16 m.y.) does not conform to their straight line plot of all dates. They conclude (their Table 1) that the isotopic date should be younger, assuming that (1) they have correctly equated the Actonian with the zone of *D. clingani* and (2) they have proportioned that zone correctly within the Ordovician. To us the possibility is attractive that neither of these assumptions is correct, that the Actonian may be as old as the zone of *D. multidens* (= zones of *C. peltifer* + *C. wilsoni*), and that no part of the Caradoc is younger than 450 m.y. The last of these may be supported by our date on the Ashgillian Wufeng Shale (447 ± 12 m.y.)

The combination of the new Late Tremadocian date and the dates for the Middle Cambrian suggests that the age of the Cambrian-Ordovician boundary should be close to 515 m.y. Therefore the duration of the Ordovician Period seems to have approximated 80 m.y.

The future

Compston (1979) in a paper particularly important to biostratigraphers calls attention to a growing divergence between isotopic geochemists and stratigraphers. Compston charges biostratigraphers with the responsibility of selling the "scientific importance of numerical time scale work. They need to convince more isotope geologists of the need to direct their interests and research funds to stratigraphic applications; they should also engage them as early as possible in the project rather than at the end of one and as part time consultants".

Compston raises a question concerning importance of isotopic dating in the choice of stratotypes of internationally important boundaries. He might very well ask whether any of the proposed stratotypes for either boundary of the Ordovician System can be dated isotopically and whether that possibility should be given as much weight as the presence of any particular fossil phylum.

There may be easily dateable units, such as bentonite beds, in stratigraphic sections which are either ignored or unrecognized. It is the responsibility of the stratigrapher to call these beds to the attention of the isotope geologist.

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$^{40}Ar/^{39}Ar$ Age Spectrum Dating of Biotite from Middle Ordovician Bentonites Eastern North America

By MICHAEL J. KUNK and JOHN F. SUTTER

Middle Ordovician bentonites in sedimentary sequences of eastern North America are potentially useful time markers. Previous attemts to determine their radiometric age have yielded discordant results for various reasons. Biotites from thermally undisturbed bentonites from the interval of North American Mid-continent Conodont Faunas 7 and 8, corresponding to an age of Blackriveran to early Kirkfieldian, have been analyzed by $^{40}Ar/^{39}Ar$ age spectrum techniques. Plateau ages of biotites from bentonites in the Stones River Formation (as used by Drahovzal & Neathery in 1971) in Alabama and the Carters Limestone of central Tennessee range from 453 to 456 Ma, providing a mean age for the above-mentioned biostratigraphic interval of about 454 Ma. Additional $^{40}Ar/^{39}Ar$ dates from the Stones River Formation in Alabama, the Hermitage Formation of central Tennessee, and the Lexington and Tyrone Limestones of central and northern Kentucky, although analytically less reliable, provide a mean age of about 455 Ma. The age spectrum data show a direct correlation between apparent K/Ca ratio and apparent age as a function of argon extraction temperature. Biotites that have individual temperature steps with apparent K/Ca ratios above about 50 tend to define age plateaus, whereas temperature steps with apparent K/Ca ratios below about 50 tend to yield discordant ages. Comparison of plateau ages with their corresponding total gas ages suggest that conventional K/Ar dating of these biotites would generally yield only minimum age estimates for these bentonites.

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Introduction

The primary obstacle to the production of a detailed radiometric time scale for the Ordovician and, in fact, the whole of the Palaeozoic, has been the lack of suitable datable material. Ideal materials (Fitch *et al.*, 1977) include datable sediments, lavas, tuffs, and bentonites interbedded with fossiliferous strata. Of these ideal materials, bentonites have a widespread occurrence in Middle Ordovician rocks in eastern North America.

Since the late 1950's, several attempts, using various techniques, have been made to date these bentonites isotopically. Faul & Thomas (1959) and Faul (1960) reported an average, conventional K-Ar age of 419 ± 5 Ma (old constants) for biotite from a bentonite near the top of the Stones River Formation (of Drahovzal & Neathery, 1971) of Alabama. Additionally, Faul (1960) reported Rb-Sr ages of bentonitic biotites from the Carters and Eggleston Limestones of Tennessee that range from 437 ± 50 to 466 ± 50 Ma (old constants). Edwards *et al.* (1959) and Adams *et al.* (1960) reported U/Pb ages ranging from 438 ± 10 to 453 ± 10 Ma (old constants) for bentonitic zircons from the Carters Limestone and Bays Formation of Tennessee. Ghosh (1972) reported conventional K/Ar ages for bentonitic biotites and a sanidine, from the Carters Limestone (equivalent to the upper part of the Stones River Formation), and from the Little Oak Limestone in Alabama that range from 424 to 453 Ma (old constants, no analytical errors given). Ross *et al.* (1981) have reported fission track ages for 3 zircon samples from bentonite beds in the Tyrone Limestone of Kentucky of 447 ± 15, 462 ± 15, and 438 ± 15 Ma. Additionally, Ross *et al.* (1981, 1982) reported

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zircon fission track ages of 454 ± 12, 456 ± 12, and 450 ± 10 Ma for bentonites from the Carters Limestone of Tennessee and the Decorah Formation and the Plattin Limestone of eastern Missouri, respectively.

The stratigraphic position from which the samples of the above mentioned analysis were collected has been limited to a narrow biostratigraphic interval by Fetzer (1973) on the basis of his own work as well as conodont and bentonite data in the literature. This interval ranges from North American Midcontinent conodont Fauna 7 to 8. On this basis it is obvious that all of the above mentioned analyses cannot be correct, even with corrections for new decay constants (Steiger & Jäger 1977), which would result in a maximum variation of any quoted age of less than 2%.

Isotopic ages from other portions of the Ordovician in North America are rather few and do not have adequate biostratigraphic and/or geologic control to allow for their use as points on a geological time scale. At best they can be used as age maxima or minima. A good example of these is the data given by Dallmeyer & Williams (1975) average $^{40}Ar/^{39}Ar$ incremental release age of 460 ± 5 Ma for the Bay of Islands ophiolite metamorphic aureole in Newfoundland, Canada. The biostratigraphic age of the emplacement of this complex has only been confined to the interval of the late Arenig to late Llandeilo due to complex structural relationships. This level of biostratigraphic control is inappropriate for constructing a geologic time scale. In addition, the age is a cooling age of hornblendes in the metamorphic aureole and, thus must be viewed only as a minimum age.

The age of the Ordovician-Silurian boundary, however, has been defined by $^{40}Ar/^{39}Ar$ total fusion ages of two hornblende samples from sedimentary breccias in the Lower Silurian *Monograptus cyphus* Zone of Esquibel Island, Alaska (Lanphere *et al.*, 1977). Although these samples are from the Lower Silurian, Lanphere *et al.* (1977) were able to suggest an age estimate for the boundary of 433–435 Ma, on the basis of their data and of a sedimentation model.

A variety of methods have been used in efforts to isotopically date Middle Ordovician bentonites. Unfortunately, as shown above, the various isotopic techniques have, in general, produced disordant results. While a complete discussion of the causes for this discordancy is beyond the scope of the present paper, a few brief comments on some basic assumptions used in most isotopic dating techniques would be of use to the reader in evaluating existing and future data.

Most of the isotopic dating techniques in use today rely on the decay of a radioactive parent nuclide to produce a stable daughter nuclide. For these dating techniques to reflect the age of the material being dated, several assumptions must be fulfilled:

1) The decay constants of the parent nuclide must be accurately known.
2) It must be possible to meansure, precisely, the isotopic composition and/or concentration of both the daughter and parent nuclides.
3) The sample being dated must have neither gained nor lost (except by radioactive decay) any parent nuclides since its formation.
4) The sample must have neither gained nor lost (except by decay of the parent nuclide) any stable daughter nuclides since its formation.

The decay constants (assumption 1) in use today for most isotopic dating techniques are those recommended by the Subcommission on Geochronology (Steiger & Jäger 1977). These constants are known to a precision of 1–2% and are internally consistent (Jäger & Hunziker 1979). The only commonly used isotopic dating technique for which various decay constants are used is the Fission Track method (Jäger & Hunziker 1979).

The ability to measure precisely the isotopic composition of both daughter and parent nuclides (assumption 2) has improved markedly in the past twenty years. In most techniques, it is now possible to measure isotopic compositions to a precision of 0.1% or less (1σ) and concentrations to a precision of about 1% or less. The only commonly used technique in which the analytical precision is worse than this, is the Fission Track dating method. The concentrations of daughter and parent nuclides are not measured directly in this method. Due to the method of measurement and also due to the inhomogeneous distribution of uranium in some minerals, the analytical precision of this technique cannot be better than about 3–4% (for a detailed explanation of Fission Track systematics, procedures, and statistics the reader is directed to Naeser 1976, and Johnson *et al.* 1979). While this level of analytical precision is inappropriate for use in the construction of a time scale, Fission Track data can be useful as corroborating evidence for samples dated by other techniques.

The closure (assumptions 3 and 4) of the isotopic system with respect to both daughter and parent nuclides is the most common problem encountered when trying to determine the age of a rock or mineral sample. Most, if not all, isotopic systems of Palaeozoic age have been open, to one degree or another, for some period of time since their formation. In the conventional K-Ar, Rb-Sr and Fission Track (track annealing in sphene and zircon) techniques there is no direct way of measuring the degree of system closure but the interpretation of the results is, at present, difficult to make for discordant data. The $^{40}Ar/^{39}Ar$ age spectrum technique, however does allow for the measure of system closure and, in many cases, if the disturbance has not been too severe, it is possible to accurately date disturbed systems. For a detailed explanation of the $^{40}Ar/^{39}Ar$ age spectrumtechnique, the reader is referred to Dalrymple & Lanphere (1971, 1974) and Dalrymple *et al.* (1981).

It is our conclusion from this brief examination of these basic assumptions that, currently, the most useful isotopic technique for high precision time scale studies in the Paleozoic is the $^{40}Ar/^{39}Ar$ age spectrum technique. It is important, however, that $^{40}Ar/^{39}Ar$ age spectrum results be confirmed by means of other isotopic dating techniques to ensure geological accuracy.

In the present study, the $^{40}Ar/^{39}Ar$ age spectrum technique has been employed to date biotites from Middle Ordovician bentonites of eastern North America. The samples analyzed were interbedded with fossiliferous carbonates located in the Middle Ordovician of Alabama, Tennessee, and Kentucky (Fig. 1) and are believed to represent the interval of North American Midcontinent Conodont Faunas 7 and 8 (Fetzer 1973). This interval corresponds to an age of Blackriveran to early Kirkfieldian (Sweet & Bergström 1976). Conodont Alteration Index values from conodonts in the associated carbonates range from < 1.5 to < 3.0 indicating paleotemperature maximums of < 90° to <200°C (Epstein *et al.* 1977). These temperatures are well below the argon closure temperature of biotite (> 250°C, Harrison & McDougall 1981). It is our opinion that this biostratigraphic interval (Faunas 7 to 8) is probably less than 2–3 million years in duration and, in terms of the resolution of isotopic dating techniques, can be considered a "Horizon".

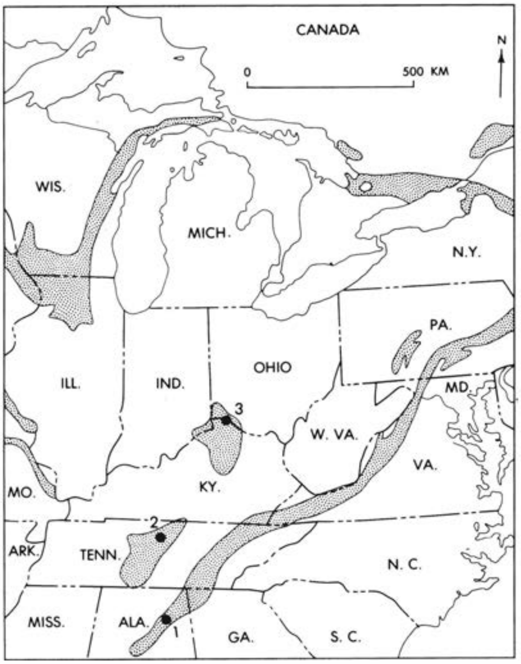

Figure 1 – Middle Ordovician outcrops in eastern North America. Sample sites are indicated by numbered black dots. Samples 31D and 30B are from Location No. 1, 40 and 29 are from Location No. 2 and 1A, CM-10 and 5A are from location No. 3 (modified from Fetzer, 1973).

Analytical Techniques

All samples were processed using standard magnetic separator, heavy liquid, and paper shaking techniques to produce biotite separates of > 99% purity. The samples were then irradiated in the central thimble facility of the U.S. Geological Survey TRIGA reactor (Dalrymple *et al.* 1981) along with aliguants of monitor mineral MMhb-1. The geometry of the irradiation was arranged to ensure that all of the samples and monitors received the same neutron fluence. The purpose of this geometrical arrangement was to provide a constant J value so that the analytical results could be more critically compared. The samples were then analyzed and their $^{40}Ar/^{39}Ar$ ages calculated using:

$$t_u = \frac{1}{\lambda} \ln \left[\left({}^{40}Ar_R / {}^{39}Ar_K \right) J + 1 \right]$$

$$J = \frac{e^{\lambda t_m} - 1}{\left({}^{40}Ar_R / {}^{39}Ar_K \right)_m}$$

where, $^{40}Ar_R$ = Radiogenic ^{40}Ar

$^{39}Ar_K$ = Potassium derived ^{39}Ar

λ = Total decay constant for $^{40}K = 5.543 \times 10^{-10}/yr$

and, t = Age of unknown Sample (u) and Monitor (m)

t_m for MMhb-1 was taken to be 519.4 Ma (Dalrymple *et al.* 1981). This monitor mineral has been described in detail by Alexander *et al.* (1978).

The analytical precision reported for the individual temperature step analyses in this study is based on the estimated error in the $^{40}Ar_R/^{39}Ar_K$ ratio. As J was a constant for all samples analyzed in this study, this (the $^{40}Ar_R/^{39}Ar_K$ ratio) is the only possible source of analytical error for internal comparisons. Errors reported in Table 1 are at the 67% confidence level ($\sim 1\sigma$) but have been compared using the Critical Value Test reported by Dalrymple & Lanphere (1969: 120) at the 95% confidence level ($\sim 2\sigma$).

During preliminary analyses, it was noted that the biotites discussed in this study contained small, but significant, amounts of ^{37}Ar. ^{37}Ar is produced during neutron irradiation by the reaction $^{40}Ca\ (n, \alpha)^{37}$ and, in addition to its use as a correction factor in calcium bearing samples, can be used, together with $^{39}Ar_K$, in the study of K/Ca ratios (for example, Harrison & McDougall 1981). In this study, a direct relationship between apparent K/Ca and apparent age as a function of argon release temperature was noted and it was possible to define for these specific samples a critical value for this ratio of ~ 50. Those having temperature steps with apparent K/Ca > 50 tended to form plateau age spectra as defined by Fleck *et al.* (1977). Samples with apparent K/Ca < 50, however, tended to yield spurious age results. The cause for this variance has not yet been determined. It may, however, be due to small amounts of inseparable, intimately intergrown chlorite or some other alteration product of the biotites.

The relationship of apparent K/Ca versus apparent age permitted the formulation of criteria for evaluation of the data reported in this study:

1) Is the age spectrum concordant or discordant?
2) If the age spectrum is discordant, does it define an age plateau (Fleck *et al.* 1977)?
3) If an age plateau is defined, are the apparent K/Ca ratios of at least two temperature steps above the critical value of 50?
4) If at least two temperature steps on the plateau have apparent K/Ca ratios above 50, do they together represent more than 50% of the total $^{39}Ar_K$ in the sample?

In all samples analyzed, the age spectra are discordant and, thus, fail criterion 1. The reader should note at this point that the failure of *all* samples analyzed to pass criterion 1 means that it is very likely all conventional K-Ar ages for these materials are spurious and of questionable value. Some samples, however, meet criteria 2–4, and, in our opinion have a high probability of being accurate indicators of geologic age. These samples will be discussed in the remainder of this report.

In a few cases, we were not able to separate enough biotite from our bentonite samples for a 7–9 step age spectrum experiment. In these cases, a 3–4 step age spectrum (3–4 step fusion) experiment was conducted. As it was not possible to apply all of the above criteria to these experiments, their reliability is somewhat lower than that of the more detailed experiments. In addition, 3–4 steps fusion experiments were conducted for all samples analyzed by the more detailed age spectrum experiments as a test of analytical reproducibility.

Results

I–59 Roadcut, Gadsden, Alabama

The roadcut is located on U.S. Interstate Highway 59, 19.3 km north of Gadsden, Alabama, and west of the trace of the Helena thrust fault. Six bentonites are present here in the upper Stones River Formation of Drahovzal and Neathery (1971) (Fig. 2) and two of them yielded biotite both of which meet $^{40}Ar/^{39}Ar$ age spectrum criteria 2–4. The biostratigraphic assignment of this bentonite complex, although somewhat uncertain due to poor conodont preservation, is believed to represent the interval of Fauna 7 and 8 (Fetzer 1973) of Blackriveran to Kirkfieldian age (Sweet & Bergström 1976, Ross *et al.* 1982). Sample 31D yields a plateau age of 454 ± 1.9 Ma (Fig. 3, Table 1) and a preferred 4-step fusion age (Table 1) of 454.2 ± 3.0 Ma. Sample 30B (Fig. 3, Table 1) yields a plateau age of 455.8 ± 1.4 Ma and a preferred 3-step fusion age (Table 1) of 453.3 ± 3.0 Ma. As all of these ages agree within the limits of analytical uncertainty and, because the age spectra passed criteria 2–4 these results are considered to be of the highest quality (Class I).

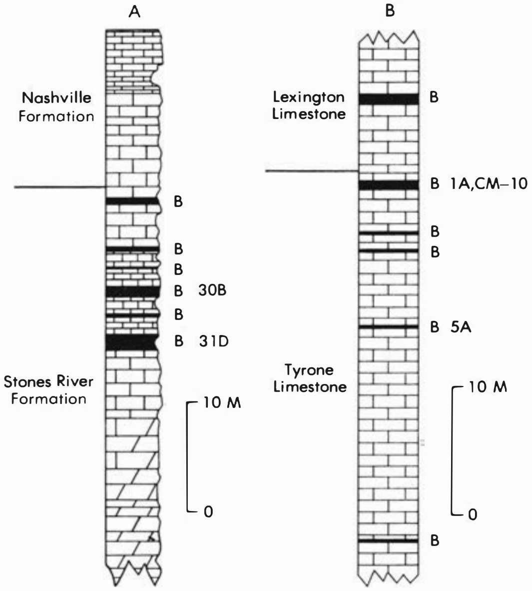

Figure 2 – Partial columnar section of (A) I–59 roadcut near Gadsden, Alabama, includes portions of the Stones River and Nashville Formations of Drehovzal & Neathery, 1971, and (B) outcrop in mine at Carntown, Kentucky. B = bentonite horizon.

Carthage, Tennessee

The outcrop sampled is located in a new roadcut on Tennessee State Route 53, 4.4 km south of Cartage, Tennessee. One sample (our no. 40) from a 41 cm thick bentonite was collected and donated to the authors by A. Soderberg of the Tennessee Valley Authority. The bentonite is located, stratigraphically, 7.6 m below the contact of the Middle Ordovician Carters and Hermitage Limestones at the contact of the upper and lower members of the Carters Limestone and its equivalent to T-3 of Wilson (1949). The Carters Limestone, at this locality, contains late representatives of conodont Fauna 7 and early representatives of Fauna 8 (Fetzer 1973); they are probably of late Blackriveran to Rocklandian age (Sweet & Bergström 1976, Ross *et al.* 1982).

Sample 40 (Fig. 3, Table 1) yields a plateau age of 452.8 ± 0.5 Ma and a preferred 3-step fusion age (Table 1) of 454.2 ± 3.0 Ma. For the same reasons as described for the data from Gadsden, these data are also considered to be of highest quality (Class I).

Watertown, Tennessee

The bentonite locality sampled (our sample no. 29) is on the farm of J. Fite near Watertown, Tennessee. Due to the poor outcrop at this locality, it was impossible to determine the precise stratigraphic position. However, this shaley bentonite is in the lower portion of the laminated argillaceous member of the Hermitage Limestone and has been identified as Wilson's (1949) T-5 by R. A. Miller of the Tennessee Division of Geology. Though slightly higher stratigraphically, the biostratigraphic assignment of sample 29 is the same as that of sample 40; late Conodont Fauna 7 to early Fauna 8 (Fetzer 1973) of late Blackriveran to Rocklandian

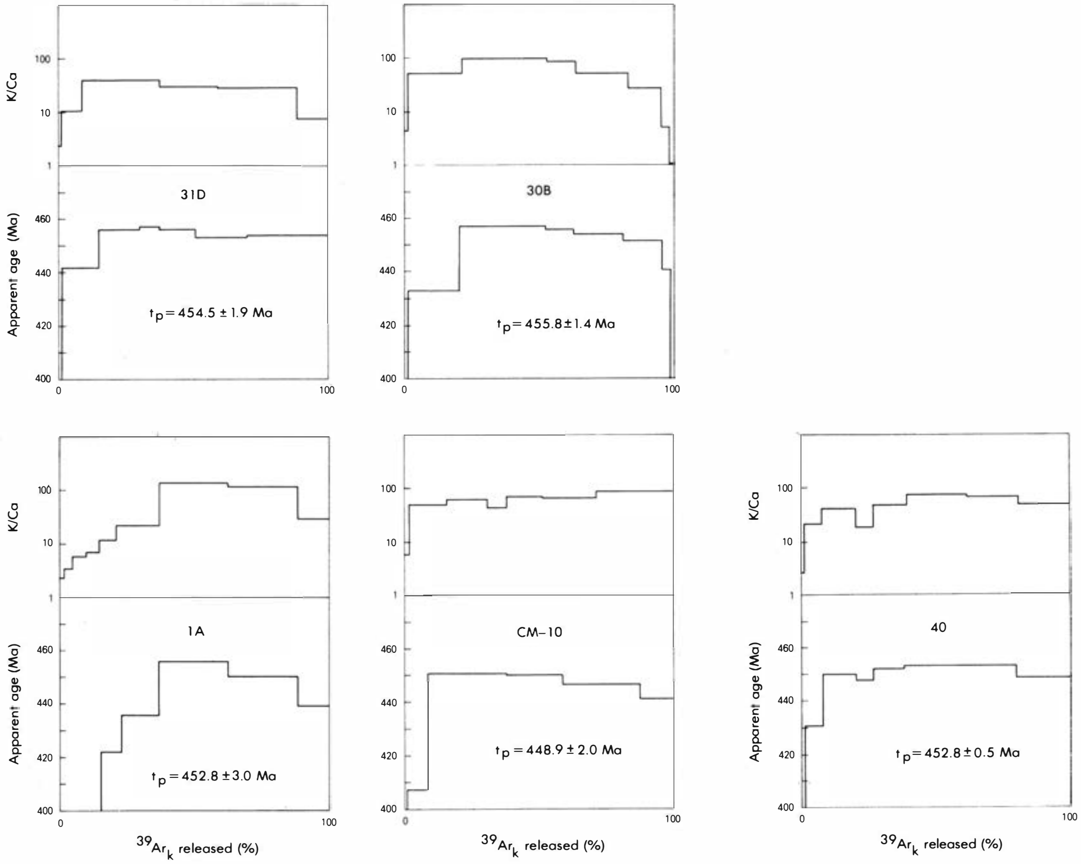

Figure 3 – Age spectra and K/Ca plots for biotites from Middle Ordovician bentonites of eastern North America.

age (Sweet & Bergström 1976, Ross *et al.* 1982).

Sample 29 was analyzed only by the 3-step fusion technique due to low biotite recovery and has a preferred age of 453.8 ± 3.6 Ma (Table 1). Even though the data are not of the highest analytical quality, they do agree, within the limits of analytical precision with other results from this horizon and thus are probably geologically meaningful (Class II data).

Carntown, Kentucky

This sample is from the shaft of an underground mine near Carntown, Kentucky. Six bentonites have been found here in the Lexington and Tyrone Limestones (Fig. 2) and of these six bentonites, two, from the Tyrone, will be discussed. The rocks in which these two bentonites occur have been placed in the interval of latest conodont Fauna 7 (?) (Fetzer 1973) of late Blackriveran (Sweet & Bergström 1976) to Rocklandian age (Ross *et al.* 1982).

Sample 1A yields a plateau age (Fig. 3, Table 1) of 452.8 ± 3.0 Ma and a preferred 3-step fusion age of 457.7 ± 3.0 Ma. Although the age spectrum meets criteria 2–4, the 3-step fusion experiment shows that this sample is not reproducible in terms of total gas age and apparent K/Ca. In addition, a separate sample, CM-10, was collected from this bentonite at the same locality by D. Stith of the Ohio Geologi-

Table 1 – Analytical data for samples discussed in this study.

Temp °C	40-Ar / 39-Ar	37-Ar[1] / 39-Ar ($x10^{-3}$)	36-Ar / 39-Ar ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	$39\text{-}Ar_K$[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
	31 D Biotite Age Spectrum I-59 Roadcut, Gadsden, Alabama J = .008898 Sample Wt = .0971 g							
350	36.67	81.6	4.497	1.4	63.8	0.16	6.0	344.1 ± 15.0
600	32.62	9.8	0.594	13.6	94.6	1.70	50.1	441.7 ± 1.8
800	33.53	7.7	0.521	15.0	95.4	1.84	63.1	455.8 ± 1.8
900	34.37	11.2	0.770	7.6	93.4	0.93	43.8	457.2 ± 1.9
1000	33.57	6.9	0.531	13.2	95.3	1.62	70.7	454.9 ± 1.8
1050	33.21	7.4	0.491	20.1	95.6	2.47	66.1	452.9 ± 1.8
1150	33.95	5.6	0.730	28.6	93.6	3.51	86.9	453.4 ± 1.8
Fuse	165.65	263.3	48.44	0.2	13.6	0.03	1.9	332.4 ± 224.9
						Total-gas Age		= 451.0
						Weight Average Plateau Age		= 454.5 ± 1.9
	31 D Biotote 4-Step I-59 Roadcut, Gadsden, Alabama J = 0.008898 Sample Wt = .0372 g							
550	46.30	40.1	6.309	4.7	59.7	0.26	12.2	400.5 ± 6.8
650	36.38	27.1	2.257	4.5	81.7	0.26	18.1	427.0 ± 5.5
1100	33.00	4.5	38.36	70.5	96.5	3.96	109.	454.2 ± 3.0
Fuse	35.27	12.1	1.257	20.3	89.3	1.14	40.5	449.7 ± 3.1
						Total Gas Age		= 449.6
						Preferred Age		= 454.2 ± 3.0
	30 B Biotite Age Spectrum I-59 Roadcut, Gadsden, Alabama J = .008898 Sample Wt = 0.0965 g							
350	37.60	113.3	4.530	1.1	64.4	0.16	4.3	355.3 ± 15.4
650	31.64	9.2	0.492	19.3	95.4	2.67	53.1	433.0 ± 1.7
850	32.81	4.9	0.248	31.7	97.7	2.37	92.2	456.9 ± 1.8
950	33.50	5.2	0.516	10.8	95.4	1.49	93.4	455.7 ± 1.8
1000	33.94	9.3	0.708	19.4	92.8	2.67	52.8	454.0 ± 1.8
1050	33.71	18.0	0.701	13.5	93.8	1.86	27.3	451.4 ± 1.8
1100	38.77	98.8	2.703	3.3	79.4	0.45	5.0	440.6 ± 5.3
Fuse	106.34	468.1	29.32	0.8	18.6	0.11	1.1	294.5 ± 44.8
						Total Gas Age		= 447.9
						Weight Average Plateau Age		= 455.8 ± 1.4
	30 B Biotite 3-Step I-59 Roadcut Gadsden, Alabama J = .008898 Sample Wt = 0.0332 g							
700	31.85	20.4	0.509	24.3	95.3	87.1	24.0	435.0 ± 2.9
1100	33.85	22.3	0.696	73.2	93.9	2.62	22.0	453.3 ± 3.0
Fuse	69.39	194.5	18.05	2.4	23.1	0.09	2.5	243.2 ± 31.0
						Total Gas Age		= 444.0
						Preferred Age		= 453.3 ± 3.0

40 B Biotite Age Spectrum Carthage, Tennessee
J = .008898 Sample Wt = .0834

Temp °C	40-Ar / 39-Ar	37-Ar[1] / 39-Ar ($x10^{-3}$)	36-Ar / 39-Ar ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	$39\text{-}Ar_K$[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
350	50.28	185.1	12.82	0.9	24.6	0.11	2.7	190.5 ± 23.1
600	35.45	23.0	1.843	6.7	84.6	0.85	21.3	430.7 ± 2.1
800	33.99	11.5	0.832	12.6	92.8	1.60	42.6	450.0 ± 1.9
900	35.22	25.8	1.310	6.7	89.0	0.85	19.0	447.7 ± 2.0
1000	34.10	9.9	0.813	12.5	92.9	1.59	49.6	452.2 ± 1.9
1050	33.06	6.3	0.442	22.6	96.0	2.87	73.3	452.9 ± 1.8
1150	33.12	6.9	0.461	19.2	95.9	2.43	70.8	452.9 ± 1.8
Fuse	33.3	10.0	0.655	18.9	94.2	2.39	49.0	448.4 ± 1.8

Total Gas Age = 447.6
Weight Average Plateau Age = 452.8 ±0.5

40 Biotite 3-Step Carthage, Tennessee
J = .008898 Sample Wt = 0.0243 g

Temp °C	40-Ar / 39-Ar	37-Ar[1] / 39-Ar ($x10^{-3}$)	36-Ar / 39-Ar ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	$39\text{-}Ar_K$[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
700	34.25	49.7	1.656	13.0	85.7	0.50	9.9	422.4 ± 3.5
1150	33.03	6.05	0.397	73.1	73.1	2.84	81.0	454.2 ± 3.0
Fuse	36.61	19.7	1.972	13.9	13.9	0.54	24.9	440.6 ± 3.7

Total Gas Age = 448.2
Preferred Age = 454.2 ±3.0

29 Biotite 3-Step Watertown, Tennessee
J = .008898 Sample Wt = 0.0063 g

Temp °C	40-Ar / 39-Ar	37-Ar[1] / 39-Ar ($x10^{-3}$)	36-Ar / 39-Ar ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	$39\text{-}Ar_K$[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
650	34.95	107.8	2.964	16.5	74.9	0.14	4.6	381.4 ± 14.4
1100	35.12	24.75	1.116	70.5	80.6	0.60	19.8	453.8 ± 3.6
Fuse	50.50	80.79	6.780	13.1	60.3	0.11	6.1	436.6 ± 19.8

Total Gas Age = 439.8
Preferred Age = 453.8 ±3.6

1A Biotite Age Spectrum Carntown, Kentucky
J = .008898 Sample Wt = 0.0920 g

Temp °C	40-Ar / 39-Ar	37-Ar[1] / 39-Ar ($x10^{-3}$)	36-Ar / 39-Ar ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	$39\text{-}Ar_K$[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
350	24.29	206.9	5.939	1.3	27.8	0.18	2.4	106.3 ± 11.6
450	19.49	142.4	1.947	2.9	70.5	0.41	3.4	210.2 ± 3.1
600	29.74	84.02	1.330	5.1	86.8	0.72	5.8	376.4 ± 2.4
700	30.66	70.19	1.179	5.8	88.6	0.82	7.0	394.2 ± 1.7
800	31.74	41.02	0.820	7.7	92.4	1.09	11.9	422.0 ± 1.8
900	32.99	21.90	0.883	14.2	92.1	2.00	22.4	435.5 ± 1.8
1000	33.13	3.59	0.389	25.4	96.5	3.57	137.	455.7 ± 1.8
1050	34.10	4.27	0.870	26.0	92.4	3.66	115.	450.0 ± 1.8
Fuse	34.29	16.95	1.227	11.6	89.4	1.64	28.9	439.1 ± 1.9

Total Gas Age = 428.4
Weight Average Plateau Age = 452.8 ± 3.0

Temp °C	$\frac{\text{40-Ar}}{\text{39-Ar}}$	$\frac{\text{37-Ar}^1}{\text{39-Ar}}$ ($x10^{-3}$)	$\frac{\text{36-Ar}}{\text{39-Ar}}$ ($x10^{-2}$)	39-Ar % of total	40-Ar % Radiogenic	39-Ar_K[2] (moles $x10^{-12}$)	Apparent[3] K/Ca (mole/mole)	Apparent[4] Age (Ma)
	1A Biotite 3-Step Carntown, Kentucky J = .008898 Sample Wt = 0.0440 g							
700	31.01	25.31	0.595	14.7	94.3	0.96	19.4	421.0 ± 2.8
1100	32.47	3.95	0.114	73.3	98.9	4.80	127.	457.7 ± 3.0
Fuse	33.30	16.32	0.818	12.0	92.7	0.79	30.0	441.9 ± 3.0
								±
						Total Gas Age		= 450.4
						Preferred Age		= 457.7 ± 3.0
	CM-10 Biotite Age Spectrum Carntown, Kentucky J = .008898 Sample Wt = 0.0888 g							
350	43.04	208.6	12.04	0.6	17.4	0.08	2.4	117.2 ± 33.4
650	30.89	45.45	9.469	7.9	90.9	1.07	10.8	406.1 ± 1.8
850	32.66	11.86	3.740	29.2	96.6	3.92	41.3	450.4 ± 1.8
950	32.90	15.84	4.594	20.9	95.5	2.80	30.9	450.1 ± 1.8
1100	32.55	16.67	4.397	28.8	96.0	3.87	29.4	446.5 ± 1.8
Fuse	33.64	62.42	9.589	12.6	91.6	1.69	7.9	440.9 ± 1.9
						Total Gas Age		= 442.8
					Weight Average Plateau Age			= 448.9 ± 2.0
	CM-10 Biotite 3-step Carntown, Kentucky J = .008898 Sample Wt = 0.0399 g							
700	30.57	48.72	1.153	14.4	88.8	0.69	10.1	394.0 ± 3.2
1100	32.27	9.36	0.165	68.2	98.5	3.24	52.4	453.3 ± 3.0
Fuse	34.02	72.25	1.475	17.4	87.2	0.83	6.8	426.4 ± 3.0
						Total Gas Age		= 440.2
						Preferred Age		= 453.3 ± 3.0
	5A Biotite 4-Step Carntown, Kentucky J = .008898 Sample Wt = 0.0489 g							
550	25.03	114.0	2.483	6.3	70.7	0.41	4.30	266.2 ± 4.1
650	29.92	87.03	1.350	4.1	86.7	0.27	5.6	377.9 ± 4.9
1100	32.69	7.93	0.188	78.4	98.3	5.10	61.8	457.7 ± 3.0
Fuse	33.70	19.24	1.002	11.1	91.2	0.73	25.5	440.1 ± 3.5
						Total Gas Age		= 440.9
						Preferred Age		= 457.7 ± 3.0

[1] ^{37}Ar corrected values were determined using a decay constant of 8.25×10^{-4} disintegrations/hour for ^{37}Ar.

[2] $^{39}Ar_K$ concentrations were calculated using the measured sensitivity of the mass spectrometer and thus have a precision of about 5%.

[3] Apparent K/Ca ratios were calculated using the equation given in Fleck, Sutter and Elliot (1977).

[4] The isotopic composition of argon was measured with a VG-Isotopes MM 1200 B mass spectrometer at the U.S. Geological Survey in Reston, Va. Samples were irradiated in the Central Thimble facility of the U.S. Geological Survey TRIGA reactor in Denver, Co. and $(^{36}Ar/^{37}Ar)_{Ca}$, $(^{39}Ar/^{37}Ar)_{Ca}$ and $(^{40}Ar/^{39}Ar)_K$ ratios used were those reported by Dalrymple *et al.* (1981). The monitor mineral used in this study was MMhb-1, which has been described by Alexander, Michelson & Lanphere (1978).

Table 2 – Summary of highest quality (Class I) and supporting (Class II) age data.

		Results of Age Data	
Sample	Locality	Age (Ma)	Preferred Age for Unit
		Class I	
31D (1)	Gadsden, Ala	454.5 ± 1.9	454.4
31D (2)	Gadsden, Ala	454.2 ± 3.0	
30B (1)	Gadsden, Ala	455.8 ± 1.4	454.6
30B (2)	Gadsden, Ala	453.3 ± 3.0	
40 (1)	Carthage, Tenn	452.8 ± 0.5	453.5
40 (2)	Carthage, Tenn	454.2 ± 3.0	
		Mean age 454.1 ± 2.1[3] (3.1)[4] Ma	
		Class II	
29 (1)	Watertown, Tenn	453.8 ± 3.6	453.8
1A (1)	Carntown, Ky	452.8 ± 3.0	454.6
1A (2)	Carntown, Ky	457.7 ± 3.0	
CM-10 (2)	Carntown, Ky	453.3 ± 3.0	
5A (2)	Carntown, Ky	457.7 ± 3.0	457.7
		Mean age 455.1 ± 4.9 Ma[3]	

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(1) Age Spectrum
(2) 3–4 Step Fusion
(3) Standard Error of the Mean, 2 σ
(4) Probable Error (includes uncertainty in age of monitor MMhb-1)

cal Survey. CM-10 yields a plateau age of 448.9 ± 2.0 Ma (Fig. 3, Table 1) and a preferred 3-step fusion age of 453.3 ± 3.0 Ma (Table 1). The plateau age spectrum of CM-10 fails criteria 1, 3, and 4. The 3-step fusion results, however, agree with the results of the age spectrum experiment of sample 1A. The conclusion to be drawn from this comparison is that this bentonite is not homogeneous in terms of the chemical composition of its biotite as reflected by apparent K/Ca ratios. Even though this sample is not reproducible in terms of apparent K/Ca ratios, the age results of both analyses of 1A and the 3-step fusion results of CM-10 are in excellent agreement with the results from Gadsden, Alabama, and Carthage, Tennessee. Due to this agreement, we believe that these data, while not of the highest quality have a good probability of being geologically meaningful and that they can be used as supportive data (Class II).

Sample 5A, also from the Tyrone Limestone, was analyzed only by the 4-step fusion technique due to the small amount of biotite in our sample. It yields a preferred age (Table 1) of 457.7 ± 3.0 Ma. Even though these data could not be properly tested by criteria 1–4, the K/Ca ratio of the preferred age fraction and the agreement in age with other samples analyzed from this "Horizon" leads us to believe that it is probably geologically meaningful (Class II).

Discussion

The age data presented in this report are of two qualities (Class I and II), defined on the basis of criteria 1–4. Class I data are of higher analytical quality than Class II and have both a low analytical error and a low probability of geologic error due to their reproducibility over a wide geographic area. Class II data are of somewhat lower analytical quality but, due to their close agreement in age with Class I data, also have a low probability of geologic error. Both Class I and Class II data are reviewed in Table 2. The estimated analytical error placed on the mean of both Class I and Class II data are at the 95% confidence level ($\sim 2\sigma$). The probable error includes a 0.5% uncertainty in J which is based on the uncertainty in the age of the monitor mineral MMhb-1.

The best age, from our data, for the interval of North American Midcontinent Conodont Faunas 7 and 8 of Blackriveran to early Kirkfieldian age is, thus, 454.1 ± 2.1 Ma. The Class II data support this conclusion and yield an age for this "Horizon" of 455.1 ± 4.9 Ma. A recent study by Ross et al. (1982) of zircon from bentonites in the Carters Limestone (equivalent in the upper part of the Stones River Formation) of Alabama and the Plattin and Decorah Formations of eastern Missouri lend support to these results. Using the fission track dating technique, they were able to establish ages of 450 ± 8, 454 ± 10, and 456 ± 11 Ma, respectively, and suggested a mean of 453 ± 3 Ma for this stratigraphic interval.

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Graphic correlation of upper Middle and Upper Ordovician rocks, North American Midcontinent Province, U.S.A.

By WALTER C. SWEET

Ranges of conodont species are used to effect a graphic correlation of upper Middle and Upper Ordovician rocks in the North American Midcontinent Province. A Composite Standard Section (CSS) synthesized from the resulting network of correlated sections provides the total stratigraphic ranges of all species for which there is information in the system, stated in terms of the 477m sequence of upper Middle and Upper Ordovician rocks in the Cincinnati Region of Kentucky, Ohio and Indiana. The CSS is divisible into 80 vertically continuous 6m units, representing approximately equal time intervals, and also into a succession of chronozones with boundaries defined as the levels at which certain conodont species first appear in the CSS.

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Stratigraphic interpretation of Mohawkian and Cincinnatian rocks in the North American Midcontinent has been carried out within a chronostratigraphic framework divided into eight stages (cf. Ross *et al.* 1982), which are based on diachronous lithic units that vary in thickness from 14 to more than 80 m, can nowhere be shown to form a chronologic continuum, and formed in sedimentary environments with diverse but disparate organic assemblages. Distribution of most fossil groups has not been determined in stratotypes of Mohawkian and Cincinnatian stages, hence chronostratigraphic resolution below the stadial level is not possible.

In this report, the distritubion of conodonts in 61 stratigraphic sections in 18 sectors of the North American Midcontinent Province (Fig. 1) is used graphically to effect a high-resolution chronostratigraphic framework for the Mohawkian and Cincinnatian Series (Fig. 2) that is conceptually absolute. This framework includes data from stratotypes of seven of the eight stages in the traditional scale (Fig. 2), hence permits chronostratigraphic evaluation of those units and may enable their continued use.

Graphic correlation

The graphic-correlation method (Shaw 1964; Miller 1977; Sweet 1979 b) has not been much used because it requires data, in feet or meters above an arbitrary base, on the ranges of fossil species in all the stratigraphic sections to be compared. Data like those in Figs. 4, 5, and 7 are suitable. Range-data sets from pairs of section are compared graphically by plotting the range-bases and range-tops of species common to the two sets as points on graphs like those in Figs. 6 and 8–13. If the array of points is rectilinear, the equation of a straight line (LOC of Shaw 1964) fitted to it expresses the relationship between the sections that yielded the compared data sets. Commonly, the LOC is drawn through the lowest of the plotted range-bases and the highest of the plotted range-tops. Points at the common bases or tops of the sections compared may be ignored, for they may represent range-limits below or above the bases or tops of those sections. Points widely separated from the axis of the array represent range-limits that are not well established in one or both of the sections being compared and may also be ignored.

The LOC equation is used to translate range-data from the section plotted on the Y axis

In Bruton, D. L. (ed.), 1984. *Aspects of the Ordovician System*. 23–35. Palaeontological Contributions from the University of Oslo, No. 295, Universitetsforlaget.

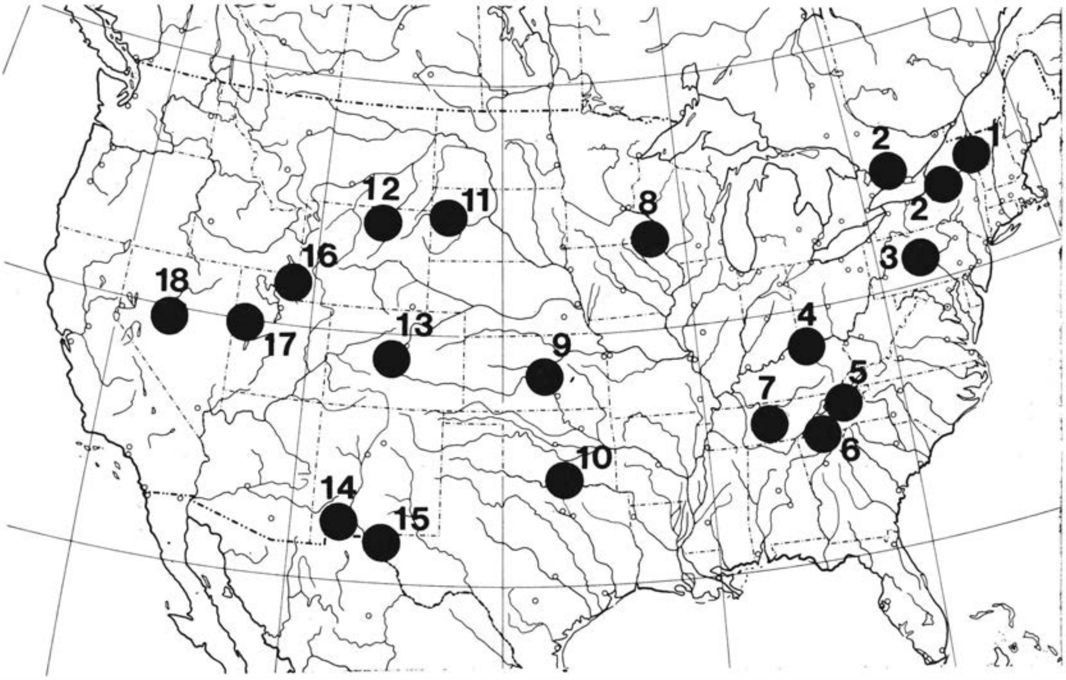

Fig. 1 – General location of sections or groups of sections that are now part of the graphic network described in this report. 1, Chazy and Crown Point, New York; 2, Black River and Trenton Group localities, southern Ontario and New York; 3, central Pennsylvania Salona-Coburn localities; 4, Cincinnati Region, Kentucky, Ohio, Indiana; 5, northeast Tennessee localities of Carnes (1975); 6, Friendsville, Tennessee sections; 7, Nashville Basin, Tennessee; 8, southeast Minnesota; 9, subsurface localities, Kansas; 10, Arbuckle and Hunton anticline localities, south-central Oklahoma; 11, northern Black Hills, South Dakota; 12, northern Wyoming localities of Sweet (1979 b); 13, central Colorado localities of Sweet (1979 b); 14, southwest New Mexico locality of Sweet (1979 b); 15, west Texas locality of Sweet (1979 b); 16, northeast Utah locality of Sweet (1979 b); 17, Ibex District, western Utah; 18, Antelope and Monitor Ranges, Nevada.

into terms of the one plotted on the X axis, which is initially a thick, well-controlled section chosen as a Standard Reference Section (SRS), but subsequently becomes a Composite Standard Section (CSS). From the two data sets, now stated in terms of the SRS, the lowest range-base value and the highest range-top value for each species is chosen and these are assembled into a CSS, which has the vertical dimension and extent of the SRS, but is synthetic in that it now includes range data from both initially compared sections.

Subsequent steps in compiling range data from additional sections are like the initial ones, save that the CSS is plotted on the X axis and its range values are modified as each new section is added. After all available sections have been compiled, component sections are serially recorrelated with the ultimate CSS, from which values controlled by the recorrelated section have been removed. Recorrelation is continued through as many rounds as may be necessary to achieve a stable network. In the one described in this report, stability was reached at the end of the fifth recorrelation round.

Compilation of a Mohawkian-Cincinnatian CSS

A stable CSS for the Mohawkian and Cincinnatian Series of the North American Midcontinent has been compiled from range data on more than 100 conodont species in 61 stratigraphic sections in the 18 areas designated in Fig. 1. The well-controlled 477m section of Middle and Upper Ordovician rocks in the Cin-

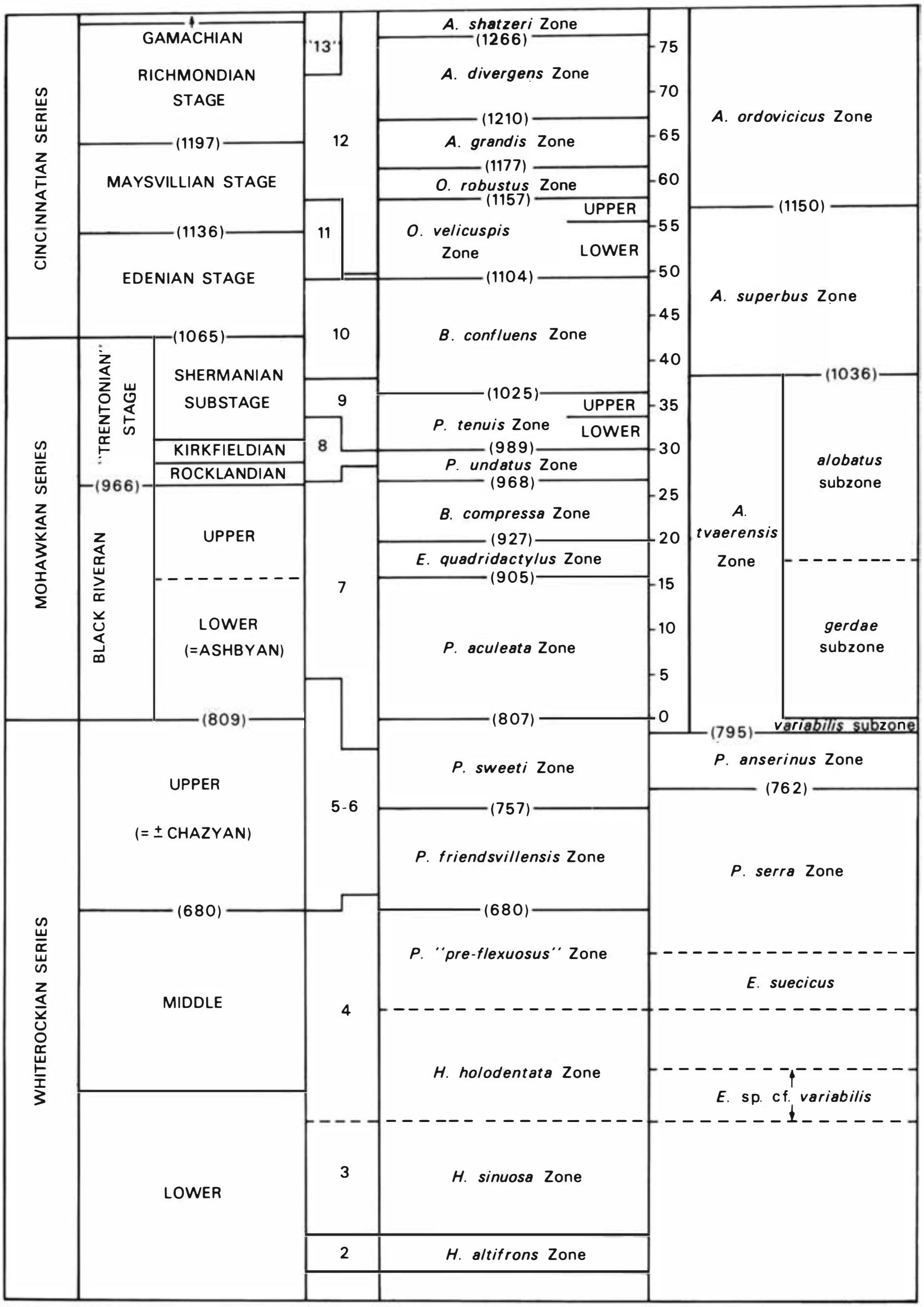

Fig. 2 – Middle and Upper Ordovician chronostratigraphic units (left two columns); conodont faunal units of Sweet, et al. (1971) (third column from left); conodont-based chronozones (fourth column); post-Whiterockian Standard Time Units (fifth column); and North Atlantic conodont zones of Bergström (1971) (right column). Vertical dimension and extent of all units determined from graphic correlation. Pre-Mohawkian conodont chronozones are defined in this report. CSS and SRS values of important stratigraphic boundaries given in parentheses.

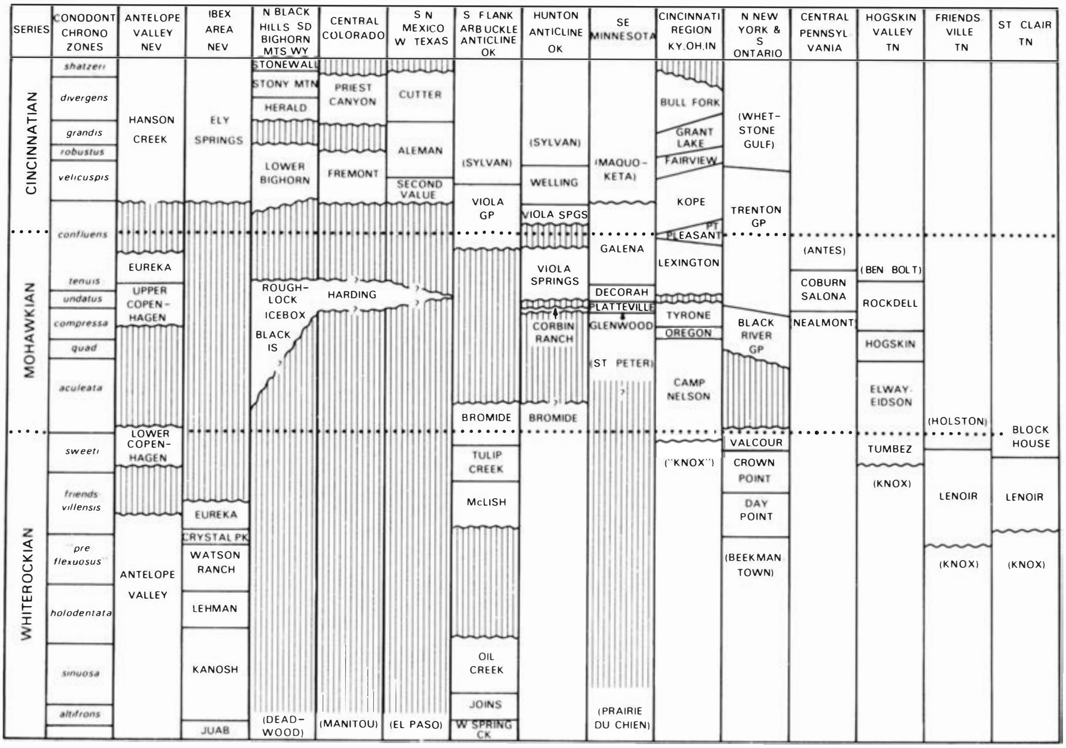

Fig. 3 – Correlation of lithostratigraphic units of selected localities in the United States, as determined by graphic correlation based on conodonts. Names in parentheses are those of under- or overlying units for which data usable in graphic correlation are not currently available.

cinnati Region of Kentucky, Ohio and Indiana was chosen as SRS. Pre-Mohawkian rocks have also been considered, primarily to ensure that the Mohawkian-Cincinnatian network could ultimately be extended to include older Ordovician strata, but also to be certain that there would not be overlap or gaps between the results of separate exercises. Results of graphic correlation at the end of the fifth recorrelation round are summarized in Figs. 6 and 8–13, and in the Appendix, which includes the names and fifth-round CSS ranges of all species used in correlation.

The upper, post-Whiterockian part of Fig. 2 summarizes the extent and relations of stratotypes of the chronostratigraphic units assembled by graphic correlation and indicates their relationship to pre-Mohawkian units. The vertical scale of Fig. 2 is that of the CSS; hence the extent of the units shown is proportional to their chronologic extent. Fig. 3 shows the lithic units recognized in the sections considered, at the vertical scale and the position indicated by graphic correlation.

Resolution. – Widths of the arrays (or W) used to establish the LOC's that relate sections of the graphic network to the SRS may be used empirically to set limits of maximum error. In the network described here, W has a maximum value of 6 m; hence a division of the SRS 6 m thick is the thinnest one that can be recognized with confidence in all component sections of the network. Thus the 477-m SRS (and the CSS based on it) may be divided into 79.5 6m units, the equivalents of which may be located in all sections of the network through use of the appropriate equations. Shaw (1964) argues persuasively that each of these SRS (and CSS) divisions, termed "Standard Time Units" (STU), represents the same length of time and

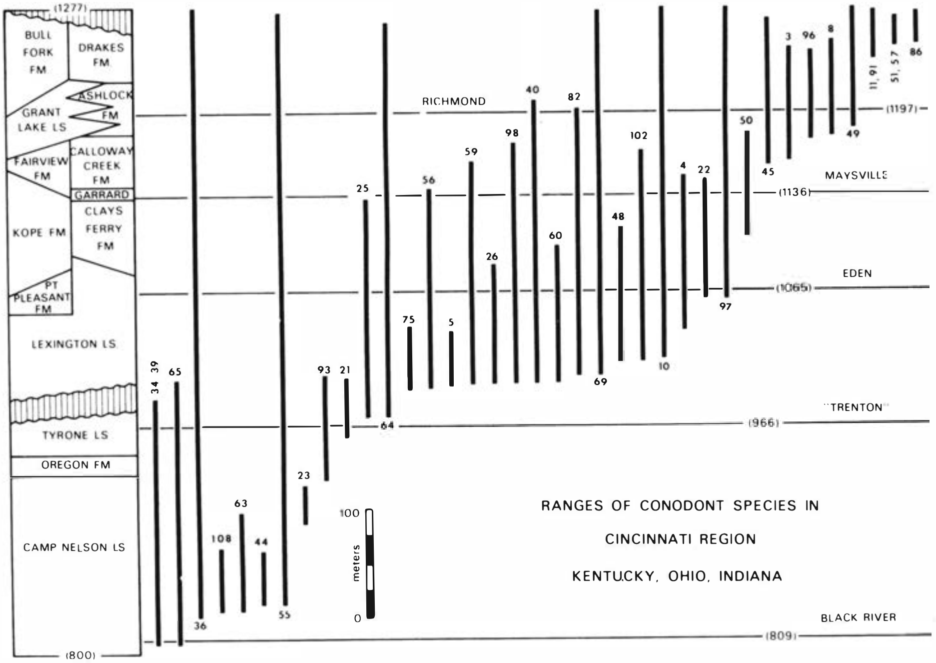

Fig. 4 – Ranges of conodont species in the Middle and Upper Ordovician of the Cincinnati Region, Kentucky, Ohio and Indiana. Data from Sweet (1979 a) and Votaw (1971). Names of species listed by number in Appendix.

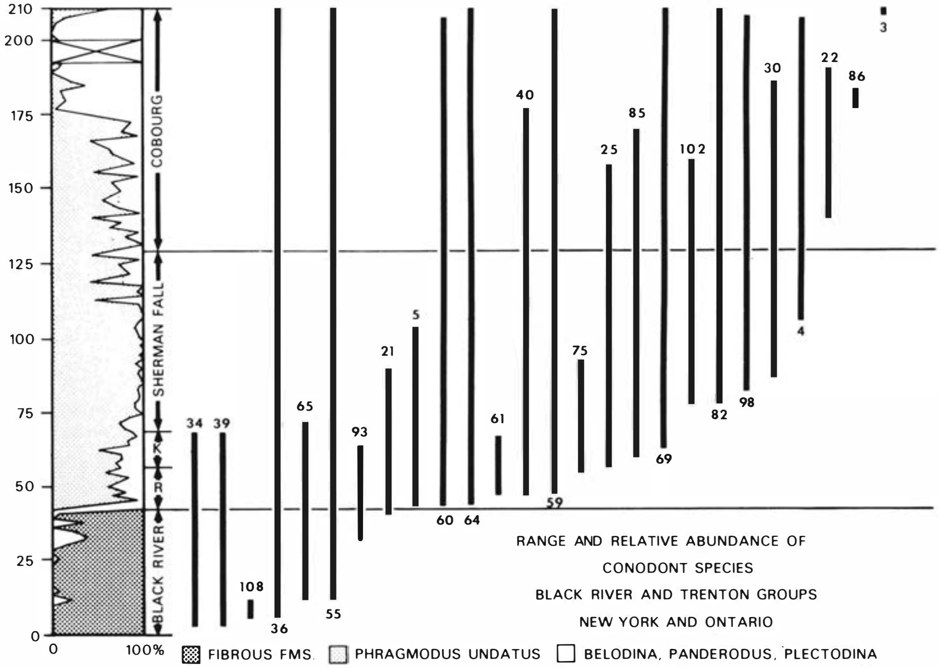

Fig. 5 – Ranges of Middle and Upper Ordovician conodont species in New York and southern Ontario. Relative-abundance log at left charts contribution of major faunal components. Data largely from Schopf (1966) and Votaw (1971); names of species listed by number in Appendix.

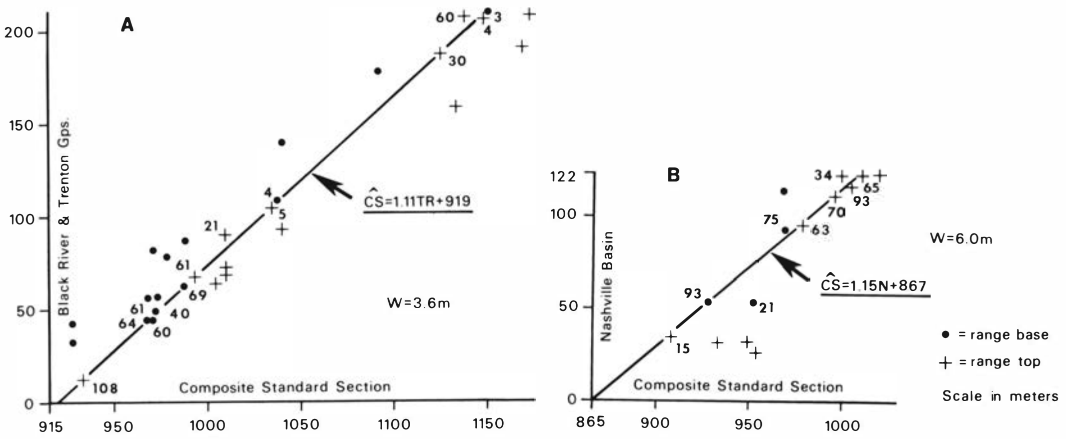

Fig. 6 – A, Graphic correlation of Black River and Trenton groups of New York and southern Ontario with Composite Standard Section. B, Graphic correlation of upper Murfreesboro, Pierce, Ridley, Lebanon and Carters formations of the Nashville Basin, Tennessee, with the Composite Standard Section. Data from Votaw (1971), sections 70VJ, VK, VL and VM. Names of species listed by number in Appendix.

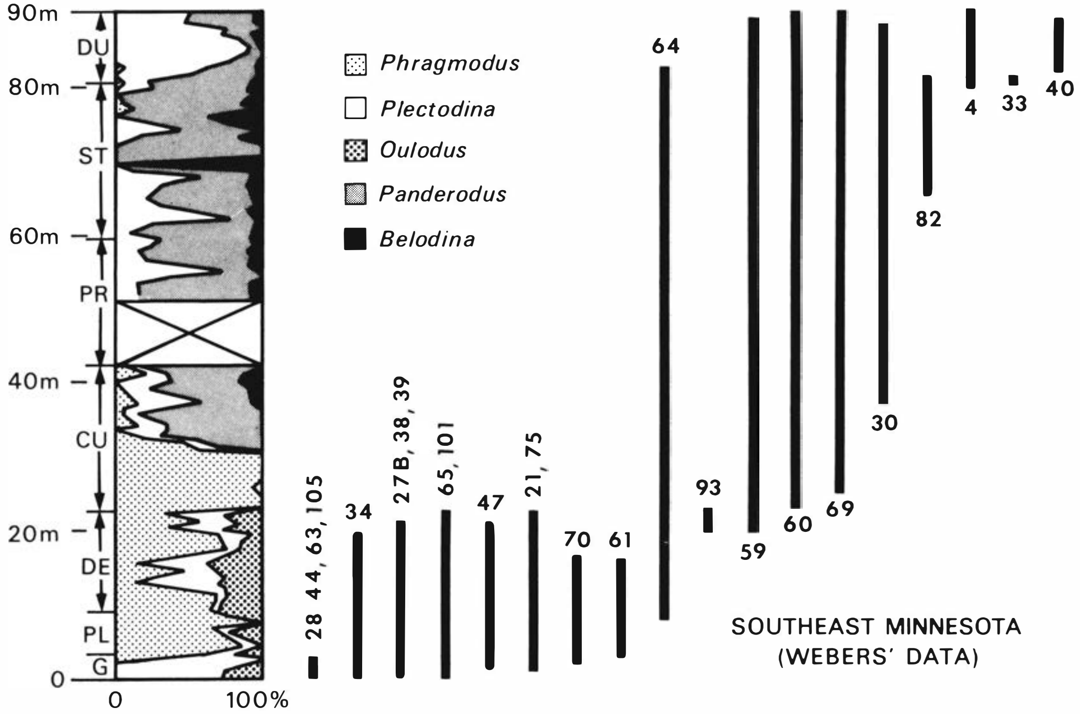

Fig. 7 – Ranges and relative abundances of conodont species in Glenwood (G), Platteville (PL) and Decorah (DE) formations and in Cummingville (CU), Prosser (PR), Stewartville (ST) and Dubuque (DU) members of Galena Formation in southeast Minnesota. Names of species listed by number in Appendix. Data from Webers (1966) and Votaw (1971, section 70VH). Species 33 is not known from southeast Minnesota, but occurs in uppermost Stewartville near Kendallville, Iowa.

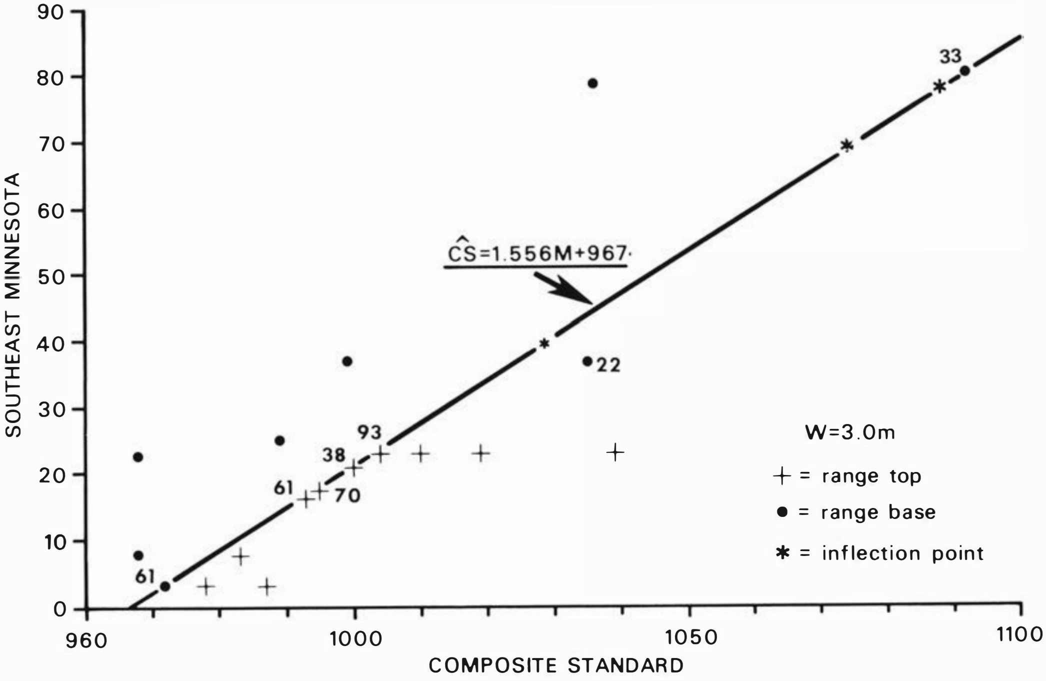

Fig. 8 – Graphic correlation of Glenwood, Platteville, Decorah and Galena formations of southeast Minnesota with Composite Standard Section. B33 plots on, but has not been used to determine LOC; its position, however, appears to confirm the fit made. Names of species listed by number in Appendix.

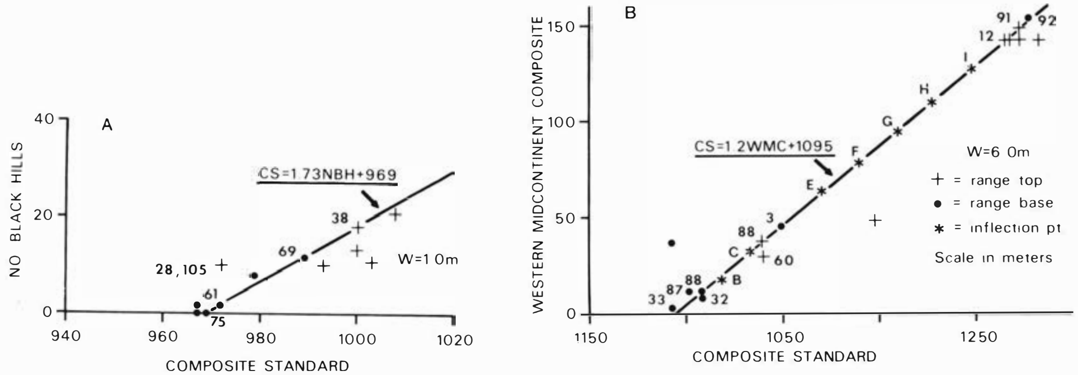

Fig. 9 – A, Graphic correlation of Icebox and Roughlock members of Winnipeg Formation of northern Black Hills, South Dakota, with the Composite Standard Section. Data from Sweet (1982). B, Graphic correlation of Western-Midcontinent CSS with Composite Standard Section; Western-Midcontinent CSS on Y axis summarizes range-data from 10 localities in South Dakota, Wyoming, Utah, Colorado, Kansas, New Mexico and Texas (Sweet, 1979 b). Names of species listed by number in Appendix.

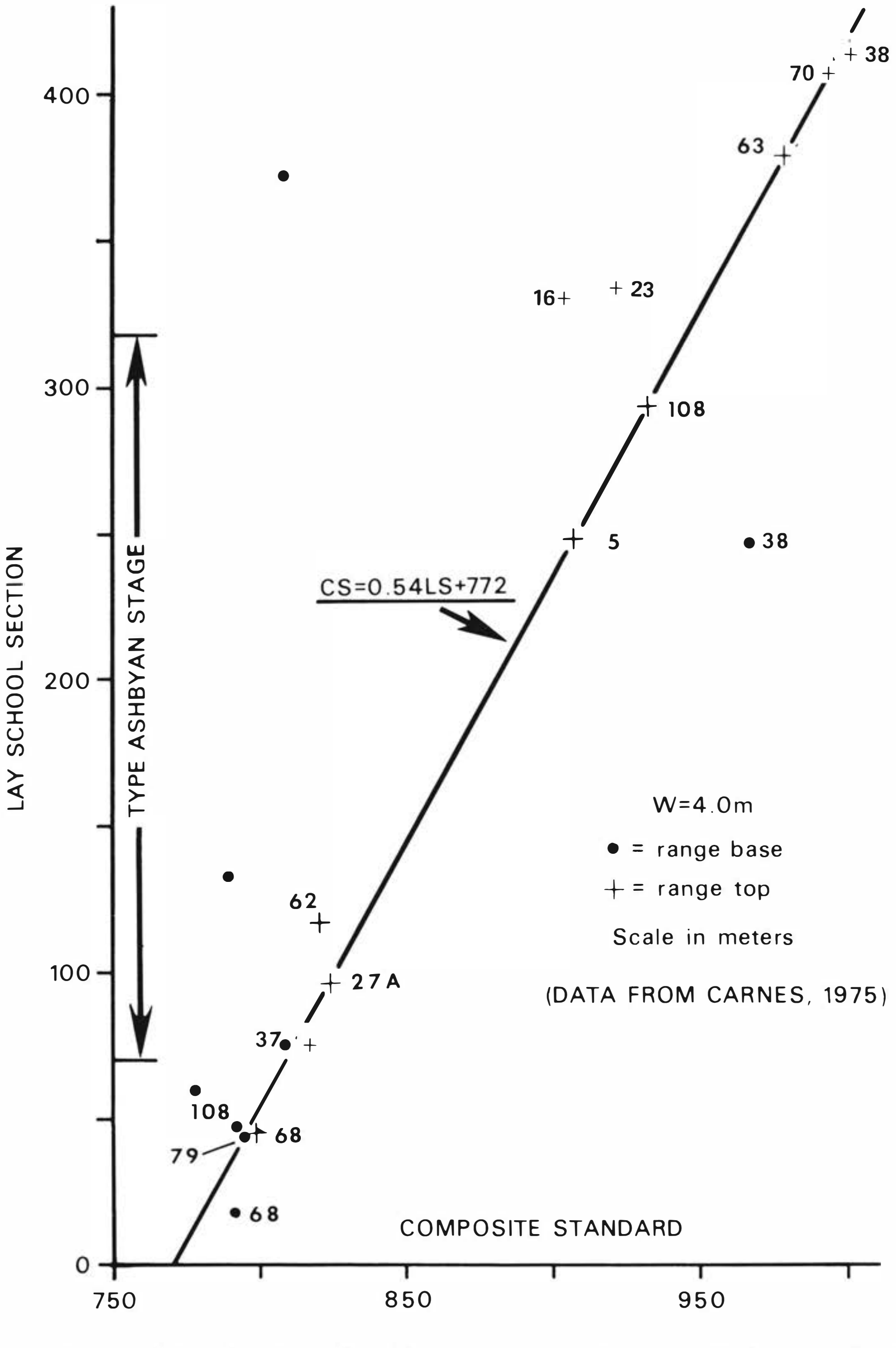

Fig. 10 – Graphic correlation of section at Lay School, northeastern Tennessee, with Composite Standard Section. Data from Carnes (1975). Elway-Eidson and Hogskin formations, 70–318 m above base of section, are stratotype of Ashbyan Stage of Cooper (1956). Bergström (in Ross et al. 1982) proposes that base of Elway-Eidson in this section be base of redefined Mohawkian Series. Names of species listed by number in Appendix.

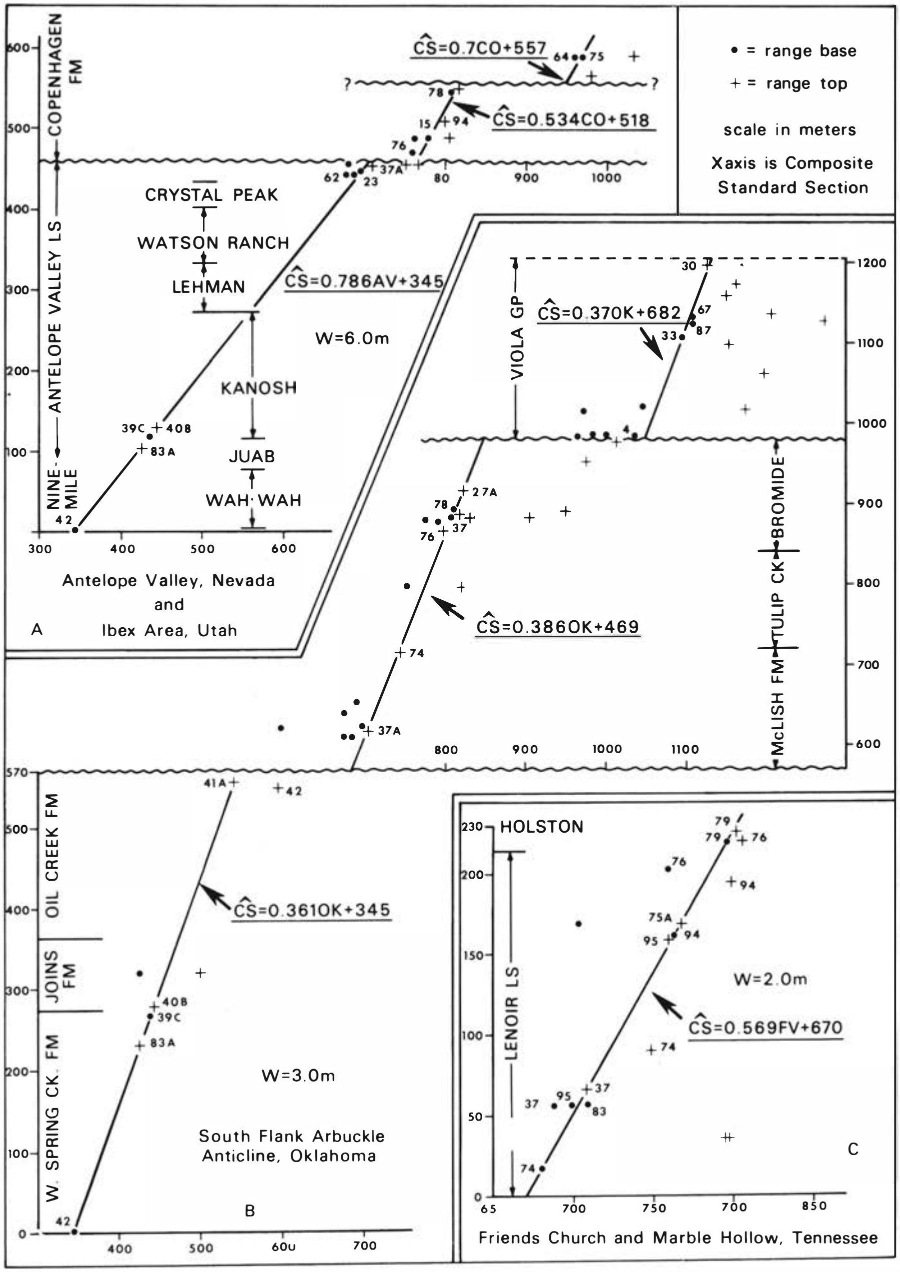

Fig. 11 – A, Graphic correlation with Composite Standard Section of Antelope Valley, Nevada – Ibex, Utah, composite section. Data largely from Harris et al. (1979) and Ethington & Clark (1982). Regional SRS is Antelope Valley-Copenhagen Limestone section described by Ross (1970) and Harris et al. (1979). B, Graphic correlation with Composite Standard Section of the upper West Spring Greek-Viola Group sequence along Interstate Highway 35, Arbuckle Mountains, Carter County, south-central Oklahoma. Data largely from undescribed Ohio State University collections, parts of which are now under study by Mr. Russell Dresbach. C, Graphic correlation with Composite Standard Section of the Friends Church-Marble Hollow section of Friendsville, Tennessee. Data from Bergström (1973) and Bergström & Carnes (1976). Lenoir strata in lower 90 m of this section are the stratotype of Cooper's (1956) Marmorian Stage. Names of species listed by number in Appendix.

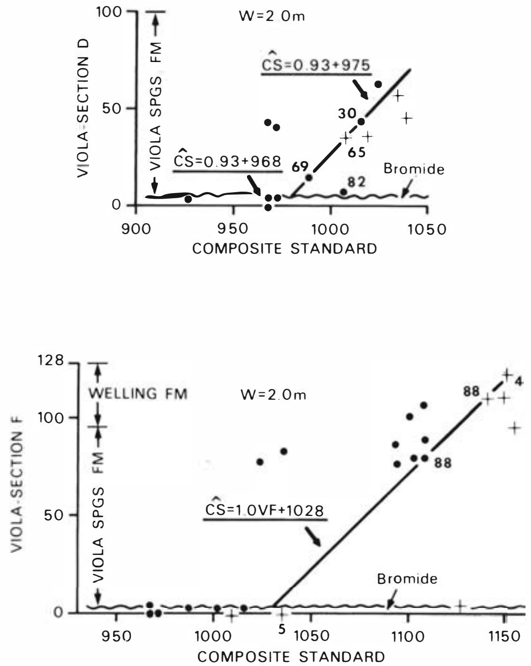

Fig. 12 – Graphic correlation of Viola Group at Alberstadt's (1973) localities D (above) and F (below) in the Arbuckle Mountains, south-central Oklahoma. Data from Oberg (1966) and Amsden & Sweet (1983). Names of species listed by number in Appendix.

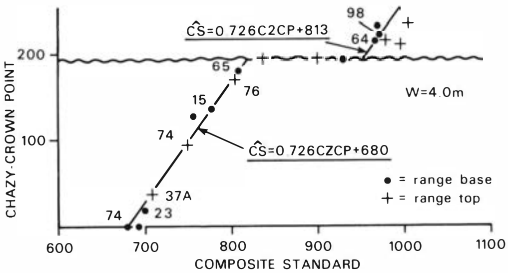

Fig. 13 – Graphic correlation of Chazy Group (below unconformity) and Black River Group (above unconformity) of Lake Champlain Valley, New York. Data for Chazy from Raring (1972) and Roscoe (1973); Black River data from Roscoe (1973). Raring's qualitative statements of range assembled on Oxley & Kay's (1959) measured section of Chazy Group and compared graphically after network based on other sections had reached stability. Names of species listed by number in Appendix.

that the scale built from them is absolute.

Of the 79.5 STU's in the SRS, 78 represent post-Whiterockian time. Ordovician rocks at the top of the Bighorn succession in northern Wyoming (Fig. 3) are younger than any in the Cincinnati Region, however, and extend the SRS upward by 11 m, or nearly two STU's. Thus, a major result of the graphic-correlation exercise summarized here is division of a CSS for the post-Whiterockian Ordovician of the North American Midcontinent into 80 STU's, each closely similar or identical to all the others in temporal extent and each recognizable with confidence in all component sections of the graphic network. If the post-Whiterock Ordovician was about 37 million years long (Ross *et al.* 1982), then each STU represents 462,500 years.

The post-Whiterockian Ordovician of North America is divided into eight stages (Ross *et al.* 1982) and is embraced by just three of Bergström's (1971) North Atlantic conodont zones, the *A. tvaerensis, A. superbus* and *A. ordovicicus* zones. Thus the capacity to resolve 80 divisions within this same stratigraphic interval increases resolution by 10 to as much as 27 times.

Chronozones

In the central column of Fig. 2, the post-Whiterockian Ordovician is divided into 11 named zones. Six others are indicated provisionally for the Whiterockian primarily to set the younger ones in chronostratigraphic context. All these zones are subdivisions of the CSS derived from graphic compilation of conodont range data at the localities shown in Fig. 1. They are vertically contiguous groups of STU's, the boundaries of which are the actual or projected levels in the SRS at which the name-giving conodont species first appear. These zones are thus chronozones rather than biostratigraphic zones. Their chronologic equivalents may be recognized with confidence only in sections that are parts of the graphically correlated network, or in sections that can be added to the network by the same procedures used to establish it. Informally, however, these units may be used with perhaps greater precision of meaning than the numbered conodont faunal intervals of Sweet *et al.* (1971), which were never intended to be biostratigraphic or chronostratigraphic units but have been treated as such by numerous authors.

The post-Whiterockian conodont chronozones named in Fig. 2 have boundaries in the CSS and SRS indicated in that figure. Lists of conodont species characteristic of those chronozones may be compiled from the CSS in the Appendix.

Conclusions

The procedures and results summarized in this severely abbreviated report are sufficient to indicate that a chronostratigraphic network that resolves at a level 10 to 27 times higher than any previously proposed, can be constructed from existing data on the stratigraphic ranges of conodonts. The stable framework may be divided into units of various sorts, to suit different stratigraphic purposes and, because it is conceptually absolute, the framework may also be useful for constructing detailed paleogeographic maps or for considering biologic or sedimentologic problems in which rate is an important consideration. However, it should also be noted that each new section compiled will necessitate reconsideration of the entire network, and that chronostratigraphic divisions of the framework are recognizable only in sections that are components of the network or can be added to it by use of the same procedures employed in assembling it.

Appendix

The Composite Standard Section

Species No.	*Species Name*	*Range in CSS*
3	*Amorphognathus ordovicicus*	1150–1269
4	*A. superbus*	1036–1151
5	*A. tvaerensis*	968–1035
6	*Aphelognathus divergens*	1210–1264
7	*A. floweri*	1153–1255
8	*A. grandis*	1177–1248
11	*A. pyramidalis*	1234–1270
12	*A. shatzeri*	1266–1288
15	*Appalachignathus delicatulus*	778– 907
16	*Belodella nevadensis*	603– 949
21	*Belodina compressa*	927–1019
22	*B. confluens*	1025–1169
23	*B. monitorensis*	698– 953
25	*Bryantodina abrupta*	973–1133
26	*B. staufferi*	997–1085
27A	*B.* n. sp. cf. *B. typicalis*	795– 824
27B	*B. typicalis*	967–1000
28	*Chirognathus duodactylus*	967– 987

30	*Coelocerodontus trigonius*	1015–1126
32	*Culumbodina occidentalis*	1104–1154
33	*C. penna*	1092–1167
34	*Curtognathus expansus*	807–1008
36	*Drepanoistodus suberectus*	680–1288
37	*Eoplacognathus elongatus*	811– 822
37A	*E. foliaceus-reclinatus* transition	687– 708
37C	*E. suecicus*	613–651
38	*Erismodus quadridactylus*	905–1000
39	*E. radicans*	807–1008
39C	*Histiodella altifrons*	437– 500
39D	*H. holodentata*	537– 690
39E	*H. sinuosa*	461– 699
40	*Icriodella superba*	972–1205
40B	*Jumudontus gananda*	264–445
41	*Leptochirognathus* sp.	532– 797
41A	*"Microzarkodina" marathonensis*	193– 546
42	*Oistodus multicorrugatus*	345– 598
44	*Oneotodus ovatus*	797–972
45	*Oulodus robustus*	1157–1283
46	*O. rohneri*	1217–1285
47	*O. serratus*	969–1000
48	*O. oregonia*	1014–1111
49	*O. ulrichi*	1102–1288
50	*O. velicuspis*	1104–1184
51	*Panderodus angularis*	1243–1274
55	*P. gracilis*	680–1286
56	*P. panderi*	793–1286
57	*P. staufferi*	1236–1282
59	*Dapsilodus mutatus*	795–1278
60	*Periodon grandis*	968–1149
61	*Phragmodus cognitus*	972– 993
61A	*P.* n.sp. *("pre-flexuosus")*	614–692
62	*P. flexuosus*	680– 835
63	*P. inflexus*	789– 978
64	*P. undatus*	968–1282
65	*Plectodina aculeata*	807–1008
67	*P. florida*	1102–1271
68	*P. joachimensis*	781– 798
69	*P. tenuis*	989–1272
70	*P.* n.sp.	680– 995
74	*Polyplacognathus friendsvillensis*	680–749
75	*P. ramosus*	969–1039
75A	*P. rutriformis*	703– 766
76	*P. sweeti*	757– 805
78	*Prioniodus gerdae*	809– 814
79	*P. variabilis*	795– 799
82	*Protopanderodus liripipus*	980–1198
83	*P. varicostatus*	702–822
83A	*Protoprioniodus aranda*	354– 427
85	*Pseudobelodina dispansa*	986–1280
86	*P. inclinata*	1092–1268
87	*P. kirki*	1101–1257
88	*P. ? obtusa*	1108–1141
91	*P. vulgaris vulgaris*	1108–1273
92	*P. vulgaris ultima*	1278–1286
93	*P.* n. sp.	927–1004
94	*Pygodus anserinus*	762– 799
95	*P. serra*	702– 761
96	*Rhipidognathus rowlandensis*	1176–1240
97	*R. symmetricus*	1058–1278
98	*Rhodesognathus elegans*	971–1174
101	*Scyphiodus primus*	967–1003
102	*Staufferella falcata*	890–1167
105	*Stereoconus gracilis*	967–986
108	*Triangulodus* n.sp.	793– 932

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Global earlier Ordovician transgressions and regressions and their biological implications

By RICHARD A. FORTEY

Global marine transgressions and regressions serve to define the original Series into which the earlier part of the Ordovician System was divided. The biological effects of these cycles are variously, but simultaneously expressed on what were independent continental blocks at the time. The faunal changes which occur at Series boundaries are as much a product of environmental shift as of evolutionary novelty. Scarcity of recoverable deep water facies from tectonic causes and partly from lack of searching) during regressive phases has meant that "ancestral" faunas have been generally overlooked, but they can be found in the correct sites in areas peripheral to former continents. Thus some of the "Llanvirn faunas" (transgressive) of Ordovician Gondwanaland can be identified in Arenig off-shelf occurrences in peripheral sites, and the "Middle Ordovician" North American fauna has a progenitor in what is believed were earlier rocks in Spitsbergen. Conversely, times of regression exposed offshore islands, and many (but not all) of these island faunas correspond with regressive intervals on the platform. Faunal interchange in relatively uniform deep water faunas may have proceeded in advance of major changes in provinciality, which are manifest when these faunas move shelfwards during transgression. These ideas are discussed in relation to trilobite and graptolite biofacies during the Tremadoc to Llandeilo. The eustatic changes could have been caused by fluctuations of a Gondwanan Ice Sheet.

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In recent years a number of authors have drawn attention to the broad geological effects of transgressions and regressions during the Ordovician (Vail *et al.* 1977; Leggett 1978; Leggett *et al.* 1981). Major sea level changes of this type have been implicated in faunal changes which occurred during the Ordovician (Shaw & Fortey 1977; Jaanusson & Bergström 1980; Ludvigsen 1982). In this paper I attempt a brief overview of the biological effects of transgressions and regressions during the earlier part of the Ordovician. In a review of this length it is not possible to give detailed documentation of all the assertions; I have concentrated on a number of critical instances which may be used in support of the general picture.

Recognition of global as opposed to local events

Regressive-transgressive events can operate on various scales; those which are addressed here are believed to have been major eustatic events independent of local tectonic circumstances. Especially in mobile sites at active continental margins during the Ordovician, there may be complex regional transgressions or local unconformities which may seem locally more significant than the more general eustatic events, and which may operate to enhance or oppose such events. Criteria which may be used to distinguish eustatic events of more than local significance are judged to be particularly:

1. That simultaneous regressions or transgressions occur on what were separate continental blocks (i.e. belonging to separate lithospheric plates), thus minimising regional influence.
2. That the events will be regionally correlatable in a consistent way. A transgressive event, for example, may be expected to in-

In Bruton, D. L. (ed.), 1984. *Aspects of the Ordovician System.* 37–50. Palaeontological Contributions from the University of Oslo, No. 295, Universitetsforlaget.

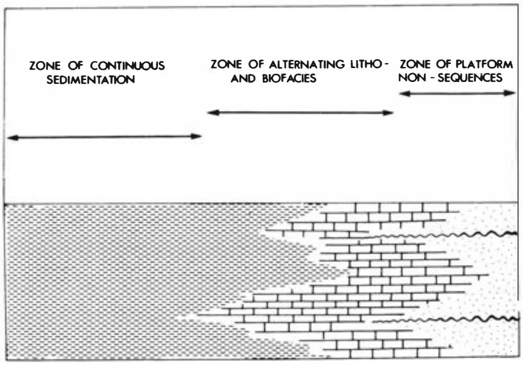

Fig. 1 – The simplest type of facies distribution patterns connected with transgressive and regressive cycles. Platform sandstones to right grading through limestones to deeper water shales. During regressive phase unconformities should coincide with exterior migration of limestone facies.

troduce sediments onto cratonic areas for the first time since the last transgresssion, and this should be directly correlatable with events such as facies changes in peripheral cratonic sites where sedimentation is otherwise continuous. (Fig. 1.)

3. Only truly off-shelf sites, which are generally developed in graptolite facies in the Ordovician, are relatively immune from the effects of transgressive-regressive cycles. These are to be taken as the standard when assessing what might be "missing" in platform regions.

Both correlation between open-ocean facies and shelf faunas, and between widely separated cratonic areas – especially if they were at different palaeolatitudes – pose particular stratigraphic problems, a product of the more-or-less patchy distribution of the biostratigraphic indices. It is important to avoid circular arguments (for example, making an *a priori* assumption that two approximately coeval regressive

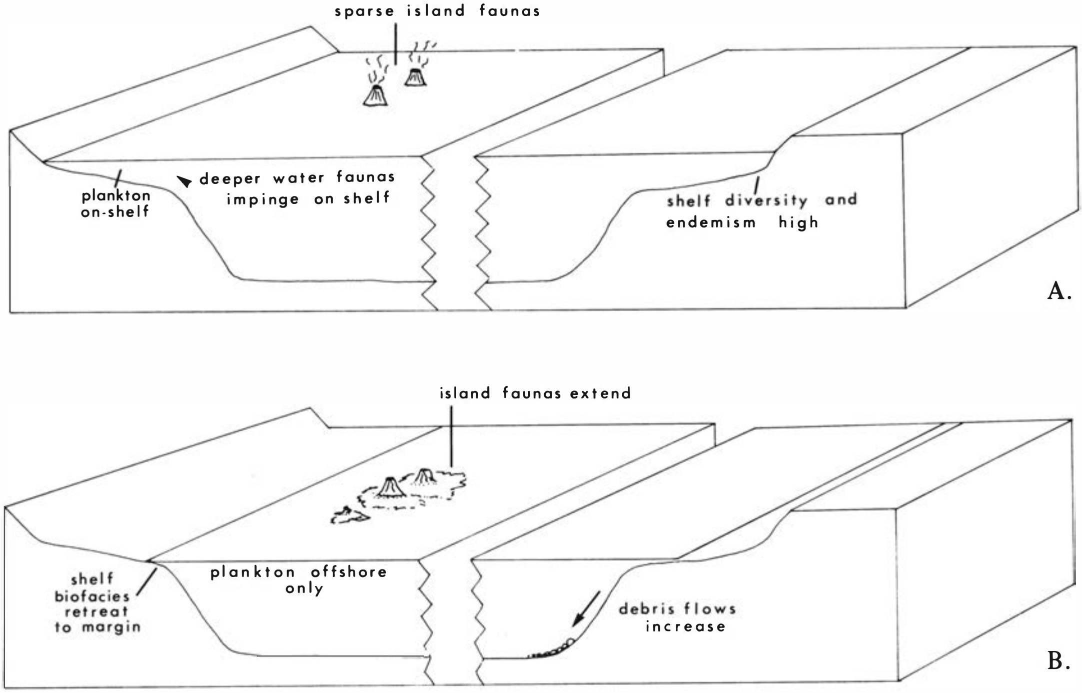

Fig. 2 – Cartoon representing some of the effects predicted during times of transgression (A) and regression (B) on opposing continental blocks.

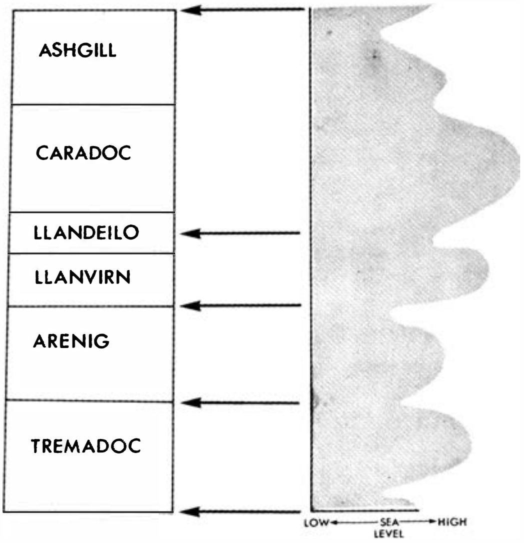

Fig. 3 – Times of proposed maximum regression (arrowed) with hypothetical sea-level curve to right.

cycles are exactly contemporary just *because* they are regressive) by treating faunal evidence on its own merits, and separately. The correlations used here are, naturally, my own, and it would be surprising if there were not areas of disagreement with other specialists. I have chosen to treat the Tremadoc to Llanvirn in detail because it is best known to me; there have been numerous recent studies on the regression associated with the late Ordovician glaciation (Sheehan 1973, 1975; Brenchley & Cocks, 1982) and further comment here would be nugatory.

Biological effects of transgressive-regressive cycles

As a basis for considering the effects of transgressive-regressive cycles on marine organisms, assumptions are made concerning the distribution of biofacies in a profile running from the interior of cratonic areas to the open ocean (Fig. 2). This is developed from the trilobite-based biofacies/community-type analyses of Fortey (1975, 1980), Ludvigsen (1975) and Cocks & Fortey (1982). Inner shelf-to-slope transects are interrupted in tropical latitudes by a carbonate mound facies which does not have a counterpart at higher latitudes. Volcanic islands are introduced on the assumption that continental edges were active margins in many places.

A. Transgression

During a relatively transgressive phase the following biological effects are predicted: (Fig. 2)

1. Flooding of cratonic areas will produce widespread shallow epieric seas. Spatial heterogeneity (Eldredge 1974) and the "species area" effect (Ludvigsen 1982) will play a part in inducing high speciation rates in epicontinental areas. Cratonic areas separated – especially by latitude – from their neighbours will generate endemic taxa.
2. As transgression proceeds, and especially in more exterior sites, previously extra-cratonic biofacies will be brought on to the shelf. Such changes will be more or less diachronous according to the rapidity of transgression (below). In some cases this may entail displacement shelf-wards of faunas usually living below the thermocline, and may produce extinctions of shelf forms.
3. Because deep water faunas are more independent of continental boundaries, times of transgression may *appear* as times of provincial breakdown.
4. In tropical areas mound faunas will migrate on-shelf.
5. Island faunas will be generally rarer as offshore 'highs' are immersed. This will apply only in the broadest sense, because active vulcanity will overtake transgressive events in some circumstances.

B. Regression

During a relatively regressive phase the following biological effects are proposed:

1. Interior cratonic sites will either have stratigraphic gaps, or, seawards, super- or infratidal deposits poor in fossils. These might be dolomites in the tropical regions or Grès Armoricain facies (Dean 1976) at high latitudes.
2. On the cratonic interior, across such a regressive phase, faunas will appear to change

with a taxonomic jump, which will not be so apparent in off-shelf faunas.

3. Retreat of biofacies seawards will mean that the "ancestors" of the faunas which are found on-shelf during transgressive phases must be sought in peripheral sites. Because such sites are narrow, and often involved in subsequent tectonism, or over-ridden by nappes, they will be uncommon.
4. Conversely, regressive phases will tend to *increase* the incidence of island faunas (or faunas fringing microcontinents) by a greater extension of surrounding productive shelves, and by bringing formerly submerged volcanic islands to shallow sub-littoral depths. These faunas will thus tend to belong in the cratonic "gaps"; exceptions are noted above.
5. Mound faunas in the tropics will be near the edge of shelf areas. Regressive phases will coincide with times when allochthonous debris-slides from such sites were at a maximum, as in the Cow Head Group, western Newfoundland (James *et al.* 1979).
6. Both mounds and islands may provide sites of retreat for shelf forms, and, if new taxa also originate there, may appear to anticipate the younger faunas of ensuing transgressive phases.

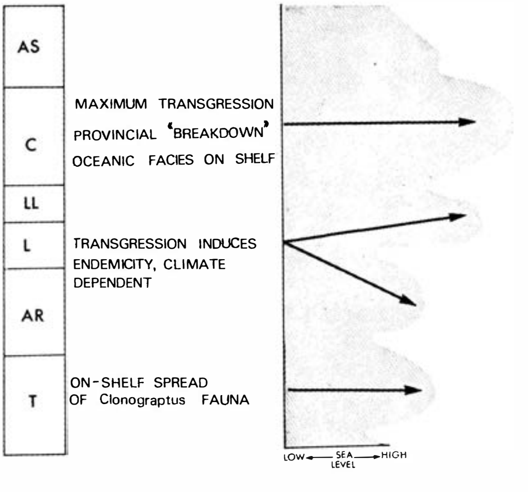

Fig. 4 – Some of the important events in Ordovician faunal dynamics related to the sea-level curve.

Times of regression and transgression

Using the criteria for recognition of worldwide transgressive-regressive events listed above, the following times are suggested for the climaxes of cycles (Fig. 3).

Basal Tremadoc

This is a time of regression; its world-wide character was suggested by Miller (1978), and reiterated on different evidence by Leggett *et al.* (1981). In this paper, evidence of its effects can be deduced from what happens around several independent continental areas at the time.

(a) On the Gondwanan margin of Iapetus, in the Anglo-Welsh area, which embraces a craton-to margin profile, sedimentation across the Cambro-Ordovician boundary is complete only in the peripheral area in North Wales recently documented by Rushton (1982). Elsewhere in North Wales the *Acerocare* Zone is missing, or represented in a phosphate horizon. At the edge of the craton, in Shropshire, there is a considerable non-sequence, with a shallowing-upward Cambrian sequence (glauconitic shales) doubtfully as young as *Peltura* Zone, overlain by Tremadoc beds with *Dictyonema flabelliforme,* but excluding the oldest subspecies of this species group.

(b) On the North American craton the interval embraced by the *Corbinia apopsis* Subzone of Upper Cambrian age, the *Missisquoia* Zone, and the early *Symphysurina* Zone (Ordovician) marks the maximum regression and the beginnings of the ensuing transgression (Miller 1978; Ludvigsen 1982; Fortey *et al.* 1982). It has recently been shown that the base of the Tremadoc can be correlated with an horizon, either in the upper part of the *Missisquoia* Zone or more probably, early in the *Symphysurina* Zone. The regression in North America is therefore coincident with the regressive event in the type Tremadoc area.

(c) In platformal Australia (Queensland), the Upper Cambrian terminates with a typical shallowing-upwards sequence of regressive sandstones, overlying fossiliferous limestones. As Miller (1978) pointed out, the

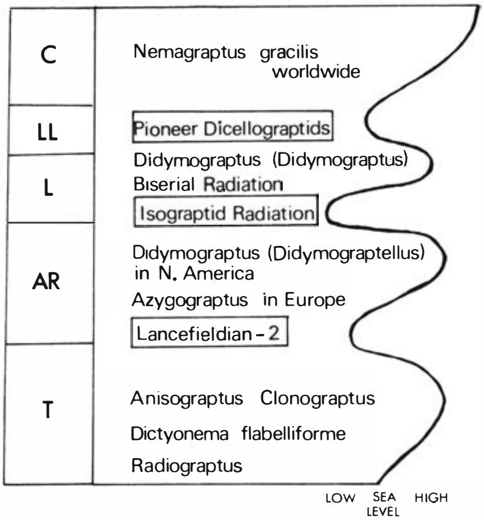

Fig. 5 – Taxonomic and geographical events in graptolite history (see text) related to sea level curve for Tremadoc to Caradoc. Times of regressive maxima shown in boxes.

following Tremadoc transgression, which reintroduces limestone deposition of the Ninmaroo Formation, correlates closely with the "Cambrian-Ordovician" boundary in North America. Miller also notes that the same "event" is recorded in platformal NE Siberia.

(d) In Scandinavia the regression at the level of the *Acerocare* Zone is well-known (Martinsson 1974; Rushton 1982). The Zone is present in the peripheral facies of the Oslo Region (Bruton *et al.* 1982) and Bornholm, for example, but is absent over much of central Sweden and eastwards on to the Russian Platform.
There is thus evidence for a regression at this time in four widely separated (and presumably tectonically independent) regions.

The ensuing transgression brought Tremadoc graptolite faunas more and more on-craton (Fig. 5), with progressively younger *"Dictyonema" flabelliforme* subspecies achieving wider dispersal, and culminating in the anisograptid fauna which alone penetrates into peripheral platform deposits in North America (Erdtmann & Comeau 1980). Endemic speciation of on-shelf trilobite faunas paralleled the transgressive phase, laying the foundations of the bathyurid faunas of North America, endemic asaphids in Baltoscandia, and such "Gondwanan" elements as Dikelokephalinidae, Orometopidae and Taihungshandiidae.

Basal Arenig

The possibility of a similar regressive-transtressive event near the base of the Arenig was noted by Fortey (1979); however, correlation between different faunal provinces is especially difficult at this stratigraphic level.

(a) Again, the type area presents a profile running from relatively inshore facies in Shropshire to peripheral cratonic facies in West and North Wales, and open oceanic in the Lake District. In western and northern Wales (Lynas 1973), the top of the Tremadoc is often a series of regressive sandstones, while the base of the Arenig is everywhere a coarsely arenaceous unit which is probably diachronous eastwards. It is difficult to assess exactly how much of the Tremadoc/Arenig interval is missing, or represented by these shallow-water deposits. Certainly there is as yet no evidence of the Zone of *Tetragraptus approximatus* in the type area, a Zone which is widespread in truly oceanic facies (condition 3 above) and frequently adopted as the arbitrary base for the Series in continuous graptolitic sequences. It seems likely that the equivalents of the Lancefieldian and possibly the early Bendigonian (Australian oceanic standard) are present in regressive facies in most of England and Wales. As might be expected, in Shropshire the gap is even larger; work in progress by myself and R. M. Owens suggests that here the "basal" transgressive sandstone is middle Arenig.

(b) On the North American platform shallow water carbonate deposition is predominant through this interval. In western Newfoundland, the St. George Group is generally poor in shelly fossils, but in the Catoche Formation a rich trilobite fauna appears relatively suddenly, including a number of oceanic trilobites and rare graptolites, a response to a sudden "deepening event" (Fortey 1979), which we would associate with the early Arenig transgression. This may be of *T. approximatus* Zone age or slightly younger and zone H in terms of the sheily North American zones (Ross 1951). Beneath this interval there is some evidence (W. D. Boyce pers. comm.) of missing trilobite zones in the early upper Canadian. Elsewhere along the eastern seaboard of North America where sparse trilobite faunas appear in the Beekmantown (Midcontinent) facies, they also seem to be of Catoche type. In the Basin Ranges of Utah and Nevada, the regressive interval may be represented in the thick intraformational conglomerates at the base of the Fillmore Limestone (Hintze 1973) – here in basinal facies, and overlain by earliest Arenig graptolites (Braithwaite 1976). Intraformational conglomerates are widespread in the Zone F – Zone G interval in the Garden City Formation (Ross 1951). Even in the Marathon region, Texas, a presumably off-shelf graptolitic sequence, the regressive interval may be represented by the Monument Spring Dolomite Member, which Berry (1960) comments "lies completely within Zone 3" (*T. approximatus*). A trilobite from Zone 4, presumably at the early stage of the transgression, is identical to one from the Catoche Formation in Newfoundland mentioned above. Even allowing for the correlation problems involved, it does seem reasonable to assume a regressive event at the Tremadoc-Arenig boundary over the North American platform.

(c) In central Australia (western Queensland) there is a stratigraphic gap between the Kelly Creek and Nora formations, with an intraformational conglomerate at the base. Unpublished work by myself and J. Shergold suggests that the transgression does not reach the continental interior here until probably Middle Arenig times. In a more peripheral site in the Canning Basin (Legg 1976), so-called Fauna 2 of probably earliest Arenig age is transgressive over ?Precambrian. The evidence is still incomplete on how much late Tremadoc is "missing" in platform Australia, but does support the notion of the early Arenig transgression. The graptolitic facies of Victoria is, of course, unaffected by the regression, and may be the most complete anywhere across this interval.

(d) Over the Scandinavian platform the diachronous early Arenig transgression has been documented in detail by Tjernvik (1956) and details need not be repeated here. Even progressing from South to North on the island of Öland (Tjernvik 1956, fig. 19) the earlier Tremadoc zones of *Apatokephalus serratus* to *Plesiomegalaspis planilimbata* are cut out – that of *A. serratus* first. This is presumably the time at which the regression was at a maximum. In the more off-shelf facies of the Oslo Region Norway the event is marked by the sudden appearance of the correlative *Ceratopyge* Limestone within the otherwise deeper-water sequence.

The evidence for a world-wide regression again seems persuasive. Since the Ceratopyge Limestone is overlain by *T. approximatus*-bearing beds, the time at which the regression was at a maximum may have been Lancefieldian Stage 2 in terms of the complete oceanic sequence of Victoria, Australia (see condition 3 above). This may well account for the extreme scarcity of graptolite faunas of this age. In any case, the transgression probably commenced within the range zone of *T. approximatus*, but may not have reached interior cratonic sites until much later.

During the subsequent Arenig transgression graptolites again appeared in cratonic sequences. In former low latitudes endemic speciation produced a variety of pendent didymograptids, which were absent from cratonic facies at former high latitudes at that time; here, the genera which apparently penetrated into epicratonic deposits were *Azygograptus* and *Corymbograptus*, the former entirely absent from the "Pacific" province. These genera presumably constitute the epiplanktonic graptolite fauna, and as such were strongly under latitudinal control for their distribution, but able to penetrate into relatively shallow-water sediments (Fig. 6). The more oceanic graptolitic facies (here termed the isograptid biofacies) is characterised by a richer fauna, including *Isograptus* and *Pseudisograptus*, slender, many-branched sigmagraptines such as *Sigmagraptus* and *Laxograptus*, and probably *Pseudotrigonograptus*. Unlike the epiplanktonic species these *were* capable of crossing latitudinal barriers but are only found in the most exterior sites. Among trilobite faunas, the Arenig transgression accom-

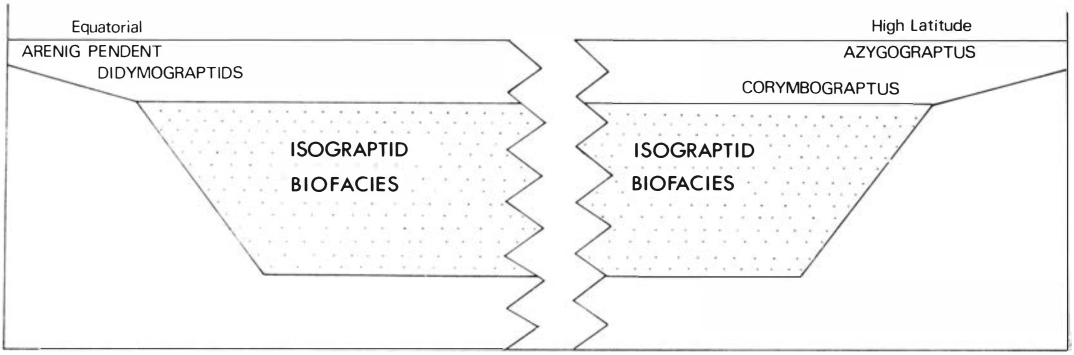

Fig. 6 – Biogeography of isograptid biofacies between North America and Gondwanaland in the early Ordovician. Some epiplanktic forms with restricted distribution are indicated.

panied what was probably the climax of bathyurid evolution in North America and endemic asaphid evolution in Sweden (Tjernvik 1956), and is associated with the appearance of bizarre endemics in the Nora Formation, central Australia.

Arenig-Llanvirn boundary

I believe that this was another time at which an important regressive-transgressive event occurred, although it does not figure on the sea-level curve of Leggett *et al.* (1981). The time of maximum regression is considered to be in the upper Arenig, equivalent to the *Didymograptus hirundo* Zone of the British standard and equating with the Castlemainian stages 2–3 and (possibly) Yapeenian Stage of Australia (Cooper & Fortey 1982). This affects both shelf and graptolitic facies worldwide, and local causes seem improbable.

(a) I have reviewed the evidence for a stratigraphic gap in shelly facies over platform North America below the Middle Ordovician (Fortey 1980). Briefly, there is a missing series of shelly faunas between the Canadian and Whiterock in sections spanning this interval. There may be regressive dolomites, with a low diversity mid-continent conodont fauna. The gap varies from place to place; it appears to be at the least in the Basin Ranges of the western United States. Shelly faunas of this age are present in only a few shelf-edge or off-shelf localities: in the Cow Head Group, western Newfoundland, Glenogle shales, British Columbia (Norford & Ross 1978), and, most prolifically, in the Valhallfonna Formation northern Spitsbergen. The interval has been termed the Valhallan Stage, and is the youngest subdivision of the Canadian Series. The ensuing Whiterock to Chazyan transgression is diachronous, and corresponds to the Llanvirn of Europe.

(b) In southern Wales the upper part of the Arenig Series is developed in a relatively deep-water graptolitic facies of black mudstones with huge-eyed bathypelagic and blind benthic trilobites. Below the Arenig-Llanvirn boundary there is a distinct facies change: light coloured shales appear with a different fauna of normal-eyed trilobites (*Ectillaenus, Barrandia, Ormathops,* abudant *Placoparia*) known from the Tankerville Flags, Shropshire. The Llanvirn boundary is marked by a dramatic influx of pendent didymograptids, which are absent in the Arenig beds here. The succession deepens upwards again into a thick, graptolitic Llanvirn turbidite-shale sequence, or, westwards, into euxinic black shales. The interruption of the deeper water sedimentation at the Arenig-Llanvirn boundary is remarkable, and it is tempting to associate the appearance of the pendent didymograptids with the onset of the ensuing transgression. What happens in North Wales is not fully known, but preliminary work suggests that the upper Arenig is absent in the type area.

(c) On the Australian platform the Nora Formation (western Queensland) is capped by a regressive sandstone facies which passes up into the Carlo sandstone. The succeeding Mithaka Formation is probably transgressive, and may be Llanvirn in age. In the more marginal Canning basin, Fortey & Cooper (1982) summarise evidence that the regression serves to remove the latest Arenig graptolite faunas, and that the Llanvirn again introduces graptolitic shales. Elsewhere on platform Gondwanaland, the Llanvirn marks a general transgression over often shallow-water Arenig facies. This applies over the Armorican-Iberian region, North Africa, and even over platform Saudi Arabia, where the *Neseuretus* trilobite fauna, and pendent didymograptids, appear in the Hanadir Shale abouve the Saq Sandstone (Fortey & Morris 1982). Unfortunately, the faunal control is insufficient to say how much of the Arenig Series may, or may not be missing beneath the Llanvirn transgression.

(d) Tjernvik (1972) and Tjernvik & Johansson (1980) have presented a detailed discussion of the correlation of the beds near the Arenig-Llandvirn boundary in Sweden. The correlation problems at this level are evidently highly complicated and contentious. It may be that the introduction of the concept of a regression and subsequent transgression at this level will help to solve some of these problems. Several remarks made by Tjernvik & Johansson (1980) are suggestive of a regression: for example, they observe that in Scania (presumably relatively marginal) there are beds "which may perhaps occupy a hiatus between the Zone of *Megistaspis limbata* and that of *Asaphus expansus*", and an environmental control at this horizon may be responsible for regional absence of the Lepidurus Limestone they claim for much of Sweden. In more marginal facies the situation may be clearer. The appearance of the "Orthoceras" Limestone and correlative units in the Oslo Region between the Lower and Upper Didymograptus Shale is exactly where it would be expected. I am indebted to Dr S. Stouge for pointing out to me that the Komstad Limestone of Bornholm and adjacent Scania occupies a comparable stratigraphic position, and Dr Stouge has recognised a succession of conodont "Iso-communities" recording the shallowing sequence.

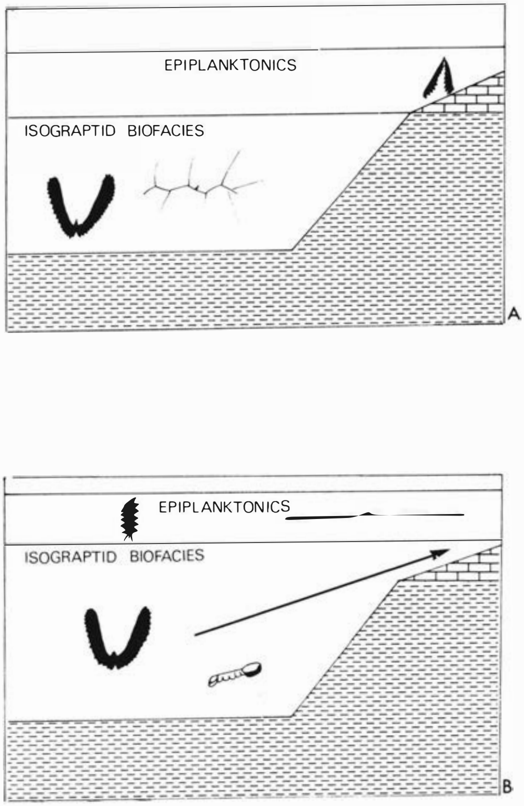

Fig. 7 – Shelfward encroachment of the isograptid biofacies in the Llanvirn transgression (B) brings this fauna onto the platform for the first time in North America. Compare this with the Arenig (A).

The evidence seems to be very good for a simultaneous regressive-transgressive event the world over – and a eustatic cause is again probable.

The Llanvirn transgression brought a flood of pendent didymograptids of the subgenus *Didymograptus (Didymograptus)* into cratonic successions around Ordovician Gondwanaland – Shropshire, England; Bohemia; France; Spain; Saudi-Arabia. The same transgression caused an onstep of the isograptid biofacies on to the edge of the North American craton (Finney & Skevington 1979). It is perhaps not surprising that the wide dispersal of "oceanic" graptoloids at this time permits relatively sound correlation in graptolitic facies (Fig. 7). The same transgression accompanied the endemic radiation of dalmanitacean trilobites in Eastern Gondwanaland, the Chazy "reefs" in North America, and later megalaspid evolution in Scandinavia.

Llandeilo and Caradoc

I do not propose to examine the evidence for the later Ordovician in detail, but for the sake of completeness it is noted that the Llandeilo and Caradoc also appear to be defined by eustatic events. It has been recognised for some time that the Llandeilo represents a relatively short interval compared with the other standard divisions of the Ordovician; in the type area the Llandeilo limestone with its characteristic fauna is the expression of a regression and subsequent transgression (Wilcox & Lockley 1981) similar to that which terminated the Arenig. Comparable facies changes occur elsewhere in Britain – for example in Shropshire. The evidence for its world-wide extent is perhaps less convincing than for the examples discussed above, but it might be noted that in platformal successions on Gondwanaland (e.g. Saudi-Arabia, SE China) Llanvirn graptolitic deposits are succeeded by regressive sandstones, which in turn underlie fossiliferous Caradoc rocks. It may also be significant that a series of extra-cratonic island deposits (see condition 5, p. 40) of supposed Llandeilo age, occur across the mobile belt in Newfoundland. The transgressive nature of the Caradoc is well known, and needs no elaboration here. It was probably of greater extent than any that preceded it, judging from the widespread introduction of facies of oceanic aspect over equatorial areas which had previously only had platform carbonate deposition. This is shown, for example, by the relatively wide geographical spread of the trilobite family Cyclopygidae during the upper compared with the early Ordovician (Fig. 8). Cyclopygidae are invariably associated with exterior facies, and may well have had bathypelagic habits. Only during the Caradoc did they penetrate over former cratonic areas and for the first time, into North America. When rare exterior sites can be recognised through a long interval of time as they have been in Kazakhstan, their associated cyclopygid faunas show very little morphological change (Apollonov 1975, 1976).

In summary, all the major divisions of the earlier part of the Ordovician are represented by regressive-transgressive events at their

Fig. 8 – Distribution of cyclopygid trilobites (left) in the early Ordovician and (right) the later Ordovician. Based on various sources. Map after Whittington & Hughes 1972.

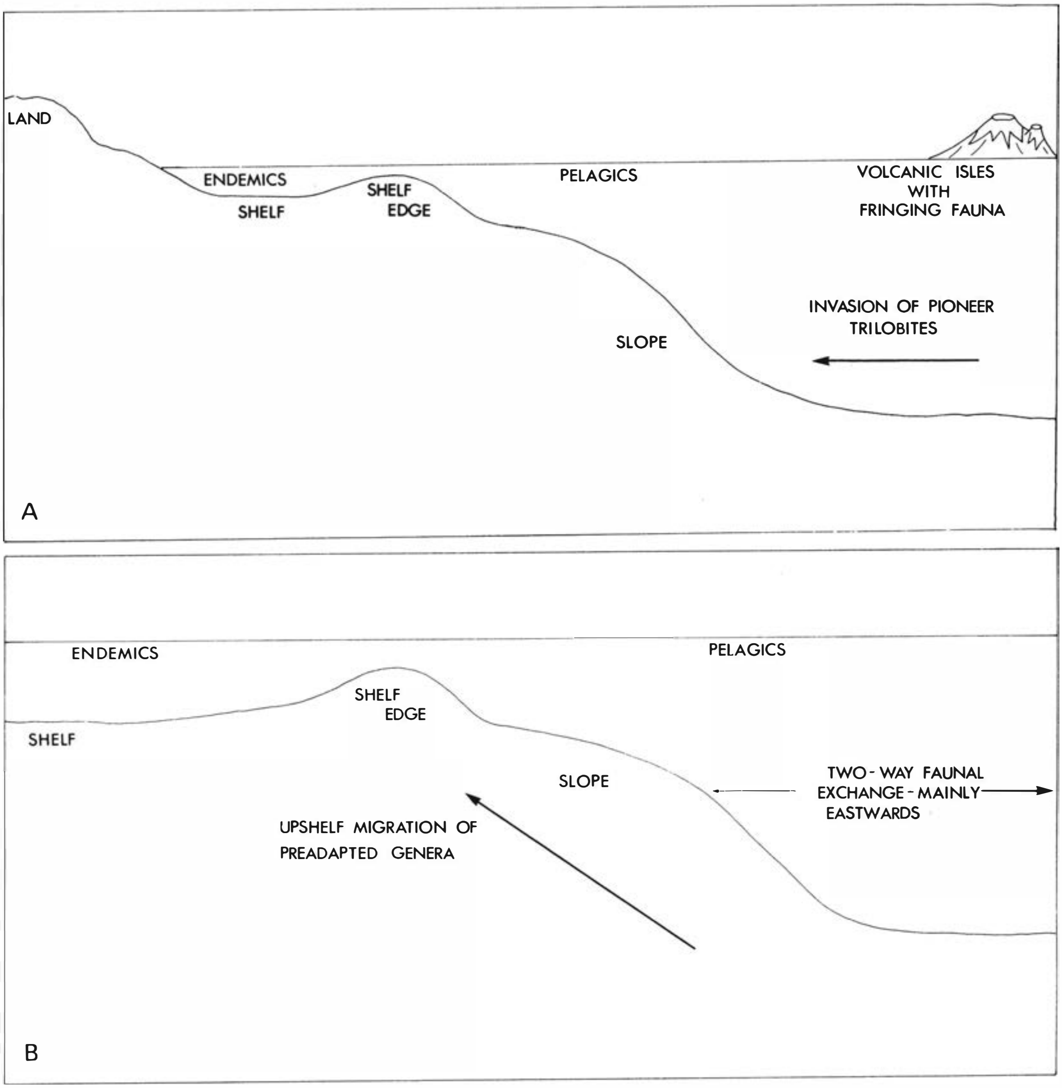

Fig. 9 – Contribution of a transgression to the Caradoc 'provincial breakdown' (A). Pre-Caradoc faunal exchange via islands or deep water facies, becomes manifest over a wider area (B) as transgression proceeds.

boundaries. The Llandeilo may be unique in that it probably consists of the regressive and only part of the transgressive event *alone,* which possibly accounts for the difficulties in its precise definition. This coincidence is not surprising, because the shifts in biofacies, lithofacies, and local uncomformities at these times introduce precisely the kind of "natural" divisions which were astutely recognised by geologists unravelling the complexities of Ordovician correlation. Ironically, it is precisely the naturalness of the divisions that makes the recognition of their boundaries – not least the Cambrian-Ordovician boundary – so fraught with difficulties.

Provincial breakdown (Fig. 9)

Much is made of the breakdown of the faunal "provinces" during the Caradoc, which is usually attributed to tectonic causes such as the impending closure of Iapetus. In the present context it is worth asking how much of this breakdown may be attributable to the Caradoc transgression itself. We have already seen how the extracratonic biofacies overstep the North American continent at this time, and are widely distributed elsewhere (Whittington 1963; Whittington & Hughes 1972, 1974). Could the replacement of the earlier trilobite "provinces" by a unified "Remopleuridid Province" (except for a relict *Selenopeltis* Province surrounding the pole) simply be greater spread of an outer shelf or upper slope biofacies? It does seem likely that the trinucleids, for example, were a group with an origin around Gondwanaland, and that they spread into outer shelf or slope benthic faunas in the Arenig, whence they may have been free to disperse to similar sites around other continental blocks. They appear in abundance in North America as the Caradoc transgression proceeds, but they *are* known from earlier marginal occurrences (Shaw & Fortey 1977). Other elements of the "unified" fauna (e.g. calymenids, raphiophorids) may have crossed Iapetus early on in the same fashion. Conversely, *Remopleurides* and dimeropygids probably had North American origins, occurring in the faunas of the Tremadoc transgression. By the Llanvirn, however, they can be found in exterior facies, whence they were free to disperse in the opposite direction to the trinucleids.

It would, however, be an over-simplification to attribute too much to the effects of the transgression alone. It would indeed be responsible for the wider spread of more cosmopolitan biofacies, and in a statistical sense for a greater appearance of faunal uniformity. North American endemic shelf trilobite faunas retreated during the Caradoc, but survived with the last bathyurids at least until the early Caradoc in the Decorah of the Upper Mississippi Valley. On the other hand the equator-wards movement of the Baltic-Welsh continent would have brought shelf environments into latitudes where compatible environments existed on both sides of Iapetus; at this stage migration could have been more or less direct. Thus, the establishment of carbonate "mound" faunas in the Ashgill of Sweden, Kazakhstan, Salair, Ireland, and northern England includes many genera with an ultimately North American pedigree (e.g. *Isbergia, Toernquistia, Heliomera, glaphurids*) together with other of Gondwanan origins (*Prionocheilus, Tretaspis* etc.). This fauna is genuinely an amalgamation of biogeographically separate earlier genera, and the mixing that occurs cannot be explained by transgression alone; the relative convergence of climatic belts during oceanic closure is reasonably invoked in this case. Note that the widespread occurrence of "mound" faunas is in accordance with the prediction (5, above) that they will occupy shelf positions in times of transgression.

Island faunas

If regressive phases expose islands (volcanic, or microcontinents), or increase the length of shorelines about them, we might expect a concentration of records of island faunas at about the times of regression. Since vulcanicity can happen at any time this will not be an invariable rule; it will better apply to "dead" volcanoes and microcontinents. One example of the latter has very recently come to light in the Scottish Caledonides. Curry et al. (1982) report an early Arenig fauna from the Highland Border rocks, a shallow-water fauna of platform North American bathyurid biofacies which may have related to an island beneath the present Midland Valley. Previous evidence of this island consisted of a derived boulder of very early Canadian age (Rushton & Tripp 1979). It is coincidence that both these occurrences correspond with, or immediately postdate, our regressive climaxes at the first two Series boundaries of the Ordovician? Similarly, it is striking how many of the "Celtic Province" island faunas lie near the Arenig-Llanvirn boundary, and hence close to, or immediately after, the climax of the late Arenig regression. The most recently described of these is that from the Otta Conglomerate, south Central Norway (Bruton & Harper 1981); this was suggested as of "late Arenig-early Llanvirn age" on the basis of brachiopods, and early Llanvirn age on the trilobites. Faunas of similar age have been described from elsewhere in the Caledonides (Neuman & Bruton 1974; Bruton & Bockelie 1980) and in the Appalachians (Neuman 1972). Island faunas approximating to the Llandeilo regression are known from several sites in the mobile belt of Newfoundland. The "Celtic Province" is a

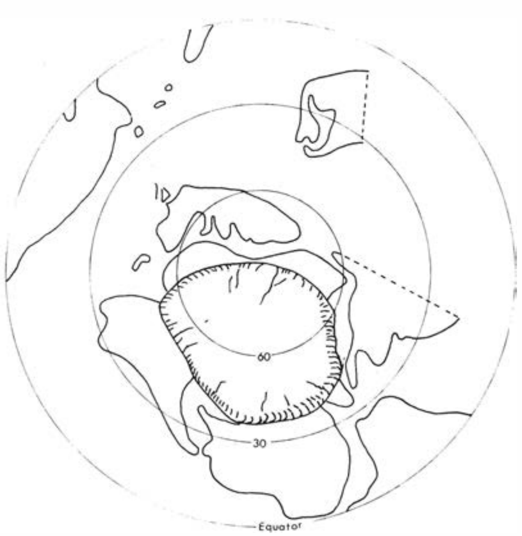

Fig. 10 – Possible extent of early Ordovician ice sheet. Its limits are taken inside the distribution of the inshore Neseuretus *facies over Gondwanaland (from Fortey & Morris 1982).*

rather loose term to describe opportunistic biotas that fringe such islands: they often seem to be derived from shallow-water sites, but I can see no reason in principle why they should not be derived from several depth associations. It is not surprising to find that they are a taxonomic/provincial mixture in terms of platform faunas. The islands may, however, be important havens during regressive phases, and if their isolation stimulated allopatric speciation, they may be implicated in the evolution of forms which aquire importance in the ensuing transgressions. They may account for the "precocious" appearance of some of the brachiopod genera (Neuman 1972) in these island faunas.

Causes of regressive-transgressive cycles

The major causes of eustatic cycles (Hallam 1981) include the effects of retreating and advancing polar ice sheets, or fluctuations in the rate of sea-floor spreading (Hays & Pitman 1973). Because there were large continental masses in the south polar region in the earlier Ordovician, the conditions were appropriate for the establishment of major ice sheets long before the well-known Ashgill glaciation. If it is assumed that the ice sheet occupied an area in the Arenig-Llanvirn, and polewards of the *Neseuretus* biofacies (Fortey & Morris 1982), which is the most inshore, we would have a possible ice sheet approximately twice the area of that in the Antarctic today (Fig. 10). It has been estimated that melting of the Antarctic icecap would induce a sea-level rise of about 50 metres; if a major advance-retreat cycle occurred in the earlier Ordovician it could have produced a transgression of even greater magnitude. Given the virtually peneplaned topography of the epicontinental areas at this time this could be sufficient to account for the biofacies and lithofacies shifts outlined in this paper, without invoking tectonic causes. However, the Caradoc transgression appears to be of greater magnitude, with the displacement of truly oceanic biofacies over the shelf edge in some places. Since the same period has been identified as one where subduction (and presumably concomitant ocean floor spreading) was particularly active, it seems possible that tectonic and glacial causes were operating together at this time. It is interesting to observe that the reconstructions of Scotese *et al.* (1979) show Gondwanaland having drifted off the pole at this period.

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Early Ordovician eustatic events in Canada

By CHRISTOPHER R. BARNES

Four areas are discussed for which the Lower Ordovician and lower Middle Ordovician stratigraphy, sedimentology and paleontology are well documented: southern Rocky Mountains, Mackenzie Mountains, Arctic Canada, western Newfoundland. For each area a curve is developed representing major regional facies shifts through this time interval. The four areas are widely separated around the outer passive continental margin and slope of the ancient Canadan (Laurentian) craton. In comparing these areas, similar changes in the curves suggest that these are the result of eustatic changes in sea level. A generalized curve of transgressions and regressions is developed for the craton with transgressions in the early and late Tremadoc, mid to late Arenig and late Whiterock; regressive phases occur in the early Arenig and early Whiterock with a brief regressive pulse in the late Arenig. Paleogeographic maps for the Canadian craton during the Tremadoc, Arenig and Whiterock stages illustrate the major facies belts and the changing patterns of epeiric seas on the craton.

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Several attempts have been made in recent years to document eustatic sea level changes from Lower Paleozoic successions (e. g. Benedict & Walker 1978; McKerrow 1979; Johnson & Campbell 1980; Johnson *et al.* 1981; Leggett *et al.* 1981; Lenz 1982). Most of these studies have used a combination of stratigraphical, sedimentological, and paleontological data to establish local apparent changes in sea level. Such apparent changes can, however, be induced by a variety of factors including eustacy, progradation, epeirogeny, tectonism which may operate individually or in combination. Space does not permit a review of this problem (see e.g. Pitman 1978). For the discrimination of unequivocal eustatic events the widespread (preferably global) occurrence of specific sea level changes must be established. It is the purpose of this paper to demonstrate that such widespread eustatic events can be recognized in Lower Ordovician strata in Canada, across distances of several thousand kilometers. Such events will need to be further tested by comparative analysis of data from outer cratonic areas (e. g. Australian and Siberian platforms).

As Lenz (1982) has noted, there is usually a larger data base available for cratonic platform facies than for abyssal facies in terms of stratigraphical, sedimentological and paleontological information. Further, the less deformed platformal sequences allow greater precision in applying the various criteria available to determine eustatic change (Benedict & Walker 1982). This paper will, therefore, only consider the platformal and slope facies of the Lower Ordovician in Canada. There are only a few major regions in Canada where most of this stratigraphic interval is well exposed, and has been studied in terms of its stratigraphy, sedimentology and paleontology to a degree that allows interpretation of eustatic changes. These areas are confined to the outer margins of the craton because Early Ordovician seas did not transgress across the interior region of the Canadian Shield as shown by paleogeographic reconstructions presented below. The areas selected for detailed review and for the interpretation of eustatic change are the southern Rocky Mountains, Mackenzie Montains, Arctic Canada and western Newfoundland. All were located on or adjacent to passive continental margins during the Lower Ordovician; only in western Newfoundland was the margin influenced by an adjacent subduction zone and in this area it did not significantly affect the margin until Middle Ordovician time (mar-

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gin collapse and ophiolite obduction). Discussion of the biostratigraphy and stratigraphical correlation of most of these areas, together with detailed references, has been presented by Barnes *et al.* (1976; 1981). Together, these areas allow an analysis of eustatic events experienced by the northern half of Laurentia (ancient North American craton) during the Early Ordovician. The object is to identify first order events and to filter out local second or third order eustatic events; these events are interpreted, and then compared, through a series of figures showing the stratigraphy and interpreted depositional environments for each region (Figs. 1–4).

Southern Rocky Mountains

The Lower Ordovician formations for western and eastern parts of the southern Rocky Mountain Fold Belt are shown in Figure 1, together with a general indication of the horizons with accurate biostratigraphic control based on conodonts (c), graptolites (g) and shelly fossils (s). The two areas represent stable platform environments with the western sections (Kicking Horse River, North White River areas) being located near the ancieng platform margin. The sections in the Main Ranges of the Rocky Mountains are the best documented, with the Survey Peak, Outram, Tipperary and Skoki formations being approximately 515, 440, 175, and 185 m in thickness, respectively. These formations are predominantly, carbonate with some shale; the Tipperary is a quartzite (Aitken *et al.* 1972).

The Survey Peak Formation overlies with sharp lithologic change the massive stromatolitic limestones and dolomites of the upward shallowing Mistaya Formation (Trempealeauan). The four informal members of the Survey Peak, in ascending order, are the basal silty, putty shale, middle and upper massive members. Aitken & Norford (1967) and Aitken 1966, 1978) considered the formation to represent a single "Grand Cycle" with the lower two members comprising a predominantly clastic, inner (?) detrital facies and the upper two members representing an upward shallowing, prograding, middle carbonate facies. Both the carbonates and clastics display a variety of shallow water sedimentary structures and a rather sparse fauna of trilobites, brachiopods, gastropods, sponges and conodonts, with thrombolites common in the upper member.

The Outram Formation is predominantly limestone, with calcareous quartzose siltstone and brown shale. The limestones are of variable lithology, but thick bedded, clotted-nodular limestone and thin beds of trilobite-brachiopod-pelmatozoan-gastropod grainstone are common; chert nodules occur throughout. The grainstones become widespread toward the top of the formation. There are no sedimentary structures indicative of intertidal environments. The formation grades westwards into the graptolitic Glenogle Shale.

Locally developed on top of the Outram Formation, or within the Glenogle Shale, the Tipperary Quartzite is thick bedded, unfossiliferous and in places dolomitic. It pinches out depositionally to the north and west.

The Skoki Formation is composed of dolomitized limestone typically pelmatozoan grainstone in the lower part passing upward into packstone and wackestone and near the top into oncolitic packstone with abundant gastropods (*Maclurites, Palliseria*). An interdigitating, diachronous contact with the underlying Outram Formation has been established. The formation spans the Lower-Middle Ordovician boundary.

Figure 1 includes an interpretative curve for the changing depositional environments within this sequence expressed in terms of intertidal, shallow subtidal and deep subtidal environments. These changes in environment may be interpreted as being produced by eustatic change. The intertidal stromatolitic facies of the upper Mistaya is replaced abruptly by open circulation conditions, but still relatively shallow, of the lower Survey Peak clastic members. The carbonate facies of the upper Survey Peak represents a shallowing event climaxing at the top with the massive thrombolitic limestone. The Outram Formation is clearly a deep subtidal facies and the change at the lower formational boundary is relatively abrupt. In the Main Ranges area, the influx of the Tipperary Quartzite marks a brief return to nearshore or intertidal conditions. The overlying Skoki Formation represents a complex of shallow subtidal environments with a relative shallowing up-

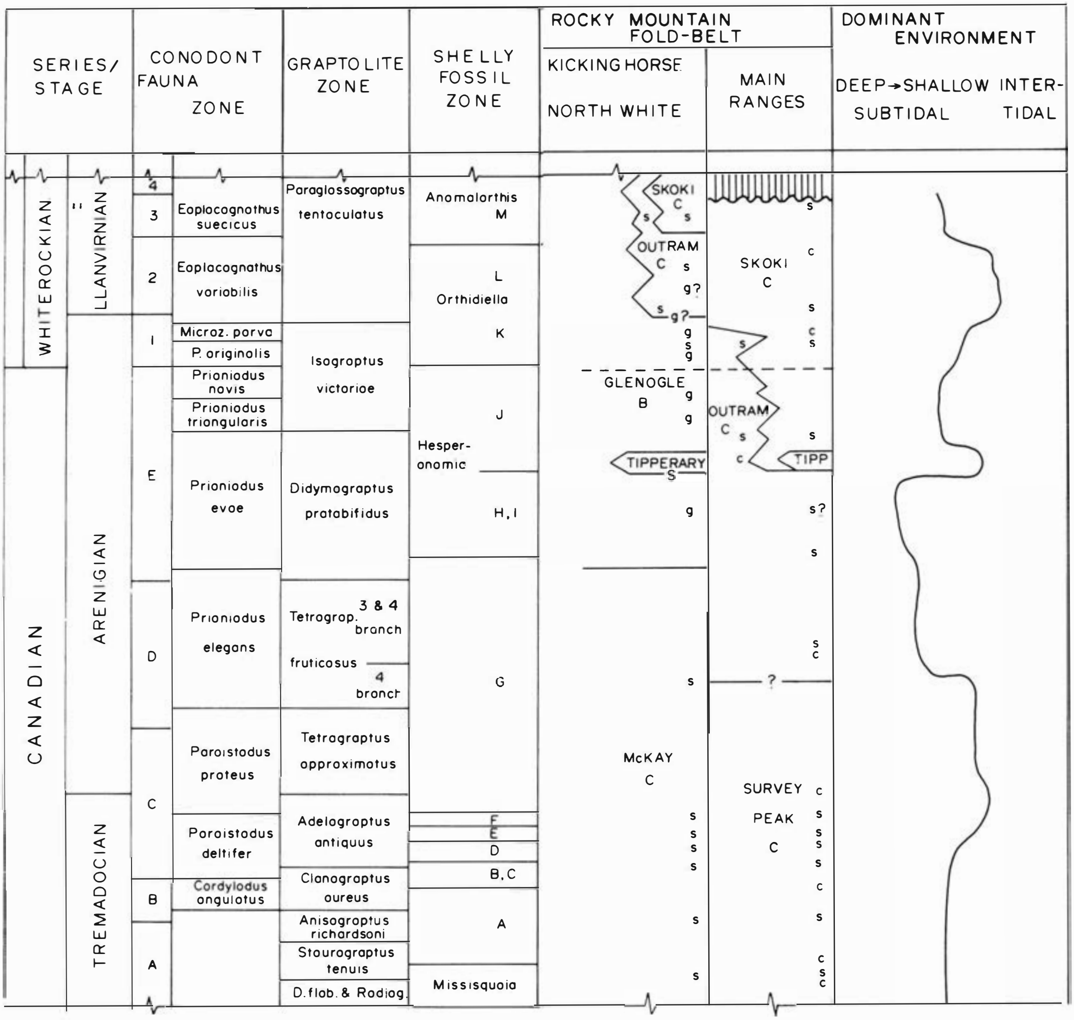

Fig. 1 – Stratigraphy, biostratigraphical control levels, and curve showing major changes in depositional environment for southern Rocky Mountains (c = conodont, g = graptolite, s = shelly, fossil, as control levels). Chronostratigraphy from Barnes et al. 1981.

wards marked by the oncolitic and gastropod limestones.

Mackenzie Mountains

In the northwestern part of Canada, Lower Ordovician strata are well exposed in the Selwyn and Mackenzie Fold Belts (Fig. 1). The central Mackenzie Mountains preserve a carbonate platform facies and part of a transitional slope facies. Further west, basinal shales are preserved in the Selwyn Basin and Misty Creek Embayent. The stratigraphy and paleontology of these areas has been detailed by Gabrielse *et al.* 1973; Ludvigsen 1975; 1978; 1979; 1982; Copeland 1977; Tipnis *et al.* 1978; Gordey 1980; Landing *et al.* 1980; and Cecile 1982.

The Broken Skull Formation consists of a thick sequence of dolostones and limestones, commonly sandy. It is correlative in part with the dolostones of the Franklin Mountain Formation (760 m thick) of the Mackenzie Moun-

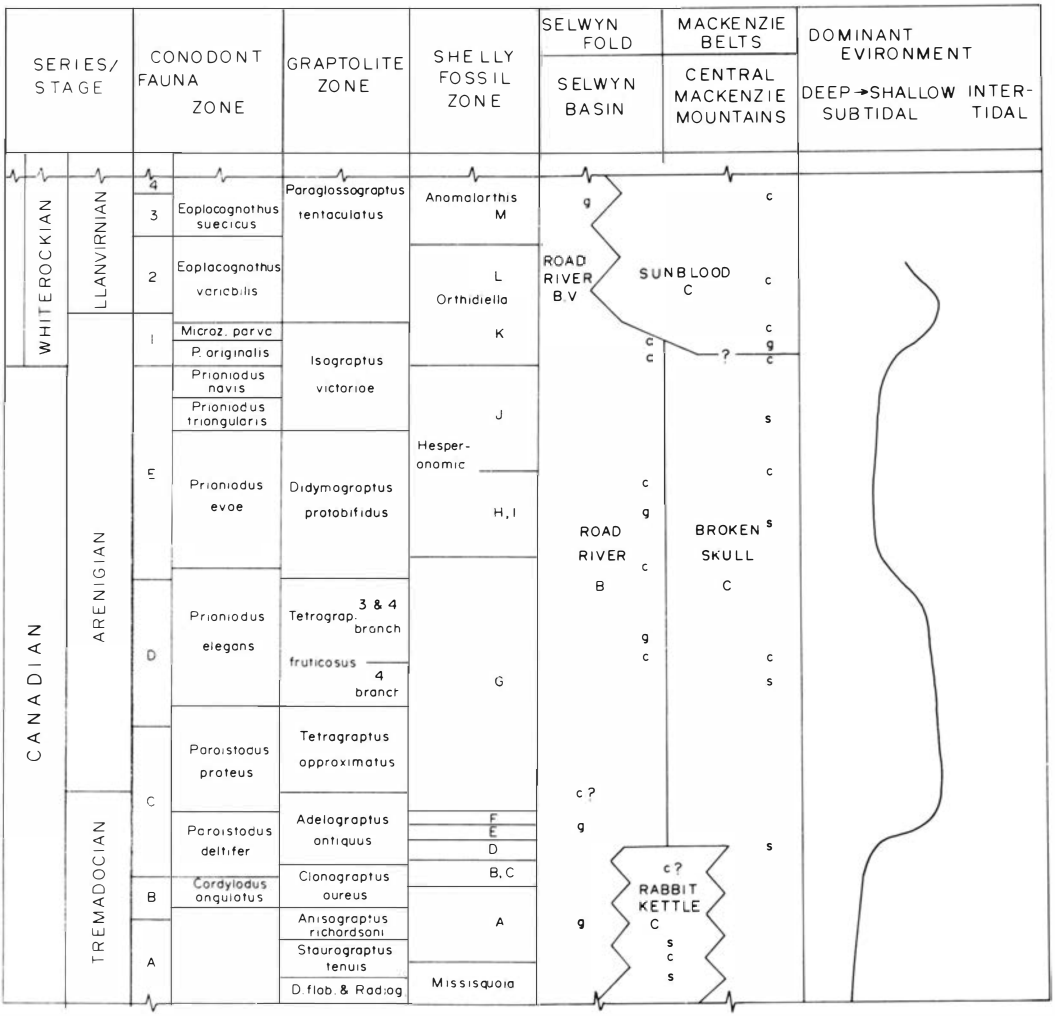

Fig. 2 – Stratigraphy, biostratigraphical control levels, and curve showing major changes in depositional environment for Mackenzie Mountains.

tains and Franklin Mountains to the east (Norford & Macqueen 1975). Shelly fossils allow correlation with the Ross-Hintze zones of the Great Basin sequence in Utah-Nevada but more detailed study is needed to better document minor facies changes as reflected by the cyclic, rhythmic, and cherty units.

The Broken Skull Formation passes westwards into the transitional slope facies of the Rabbitkettle Formation. Both formations were initiated in the Late Cambrian. The Rabbitkettle Formation (up to 750 m thick) consists of thin bedded silty limestone with shaly partings. Ludvigsen (1982) has shown subtle but important petrographic differences within the formation with some of the limestones being black laminated lime mudstones whereas others below are burrowed lime wackestones. He attributed the change to a deepening phase at the base of the *Corbinia apopsis* Subzone of the *Saukia* Zone coincident to the base of the "Hystricurid" Biomere and a Grand Cycle. The Rabbitkettle Formation extends into the Late Tremadoc in some areas (e.g. western District of Mackenzie; Ludvigsen 1982) and then is overlain by a black barren dolostone member of

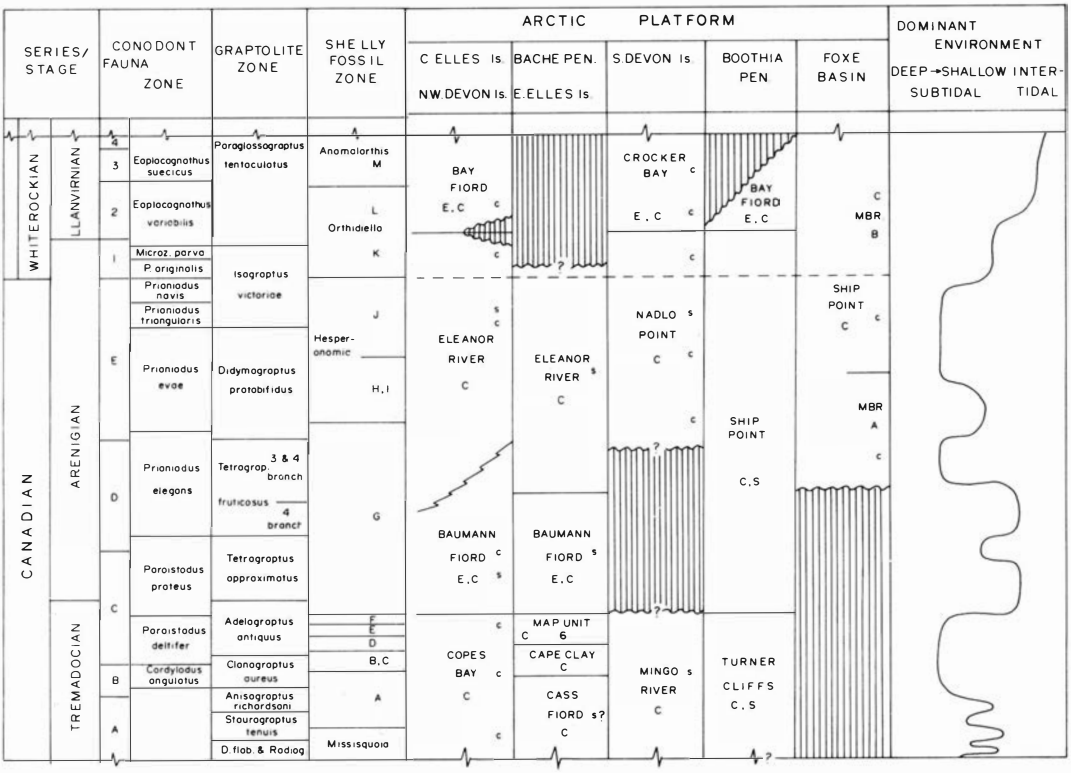

Fig. 3 – Stratigraphy, biostratigraphical control levels, and curve showing major changes in depositional environment for Arctic Canada.

the Road River Formation probably reflecting a brief relative shallowing. In other areas (e.g. southwest Mackenzie Arch; Cecile 1982), the Rabbitkettle Formation persists to approximately the base of the Whiterock before being overstepped by the basal Sunblood Formation as a result of another relative shallowing event.

The Sunblood Formation (up to 1400 thick) consists of dark grey dolostone, alternating dark and light grey dolostone overlain by limestone, locally with interbedded sandstone, and finally by grey limestone and dolostone, locally bioclastic (stratigraphy revised by Ludvigsen 1975). The formation is characterized by vivid weathering colours due to a high silt content. Ludvigsen (1975) advocated a shallow sublittoral, occasionally littoral environment for the Sunblood.

In terms of depositional environments, a generalized curve for the Mackenzie Mountains area is shown on Figure 2. It must be emphasized that this reflects changes in the areas of carbonate platform edge to transitional slope facies and is based largely in Ludvigsen's recent studies. He has demonstrated, in applying the Grand Cycle concept to the Rabbitkettle Formation, that a marked relative deepening in the slope facies occurs at the base of the *Corbinia apopsis* Subzone (late Trempealeauan). The black dolostone member, basal Road River Formation, and the nature of the middle Broken Skull Formation indicates an overall shallowing phase in both slope and platform facies. The more diverse faunas of the upper Broken Skull Formation and the development of more anoxic conditions in the laterally equivalent part of the Road River Formation suggests a relative deepening. The basal Sunblood Formation and its overstepping relationship to the slope facies indicates a fairly abrupt shallowing phase in early Whiterockian time. These changes are reflected in shifts in trilobite and conodont biofacies (Ludvigsen 1975, 1978; Tipnis *et al.* 1978; Landing *et al.* 1980).

Arctic Canada

In Arctic Canada, Lower Ordovician rocks are widely exposed in many areas of the Arctic Islands and northern mainland, together comprising the Arctic Platform (Fig. 3). Formational names change across this vast territory but the basic lithological successions are remarkably similar. Details of the Lower Ordovician stratigraphy and paleontology provides a basis for review comments have been published by several authors including Kerr 1968; Mossop 1979; Barnes 1974; Morrow & Kerr 1977; Mayr 1978; Miall & Kerr 1980.

The lower and late Middle Ordovician succession, where complete, typically consists of an alternation of two major facies: shallow subtidal carbonate, and evaporite. The Copes Bay and Eleanor River formations represent the former and the Baumann Fiord and Bay Fiord formations represent the latter.

The Copes Bay Formation (commonly 300–500 m in thickness) has a lower and upper part that consists of thick to massive bedded, mottled, limestone with a variety of sedimentary structures indicating a shallow subtidal environment. In many localities, an interval within the formation consists of thin bedded dolostones with desiccation cracks indicative of regional shallowing.

The overlying Bay Fiord Formation (300–350 m thick) is siminar in lithology to the evaporitic Baumann Fiord and both are recessive in outcrop. Limestones and shales predominate toward the top of the formation. As with the Baumann Fiord Formation, the unit is locally thin or absent and marked by an hiatus.

The Lower Ordovician succession in the Arctic Platform thus represents an oscillation of shallow subtidal and evaporitic intertidal-supratidal environments. The facies persist for significant periods of time. A curve representing these shifts through time is included in Figure 3; the oscillations at the base of the chart are diagrammatic, representing the stromatolitic cycles in the lower Copes Bay Formation.

Western Newfoundland

This area contains excellent exposures of Lower Ordovician strata of the carbonate platform facies (St. George Group) and also in the continental slope facies of the Cow Head Group preserved in adjacent allochthonous sheets. These two facies belts have long been studied in detail and much recent work has also been completed or is in progress. The review below is based on many of these detailed studies including Kindle & Whittington 1958; 1959; Whittington 1968; Hubert *et al.* 1977; Knight 1977 a, b, 1978; Fåhraeus & Nowlan 1978; Fortey 1979; Fortey & Skevington 1980; Stouge 1980, 1981, in press; Fortey *et al.* 1982; Stouge & Godfrey 1982.

The St. George Group (about 500 m thick) consists of four formations, in ascending order the Watts Bight, Boat Harbour, Catoche and Port au Choix formations. It is overlain, commonly disconformably, by the Table Head Group. The St. George Group consists of dolostones and limestones, commonly stromatolitic or bioturbated, and with chert at some horizons. There are a few brecciated horizons produced during karst formation following regressive phases. The pattern of shifts between exposure, intertidal and shallow subtidal facies has been detailed by Stouge (in press, Fig. 6) who has interpreted these changes in terms of major and minor eustatic events and the major ones are incorporated into the curve shown on Figure 4. Important lead-zinc deposits (e.g. Daniels Harbour) are associated with these paleokarst horizons.

The Cow Head Group (Ordovician part being about 160 m thick) is a sequence of thin bedded ribbon limestone and graptolitic shale that are interbedded with carbonate conglomerate and breccia with clasts or blocks up to many tens of metres in diameter. The breccias were derived through down-slope slumps or flows carrying carbonate blocks from the shelf margin (James 1981). The thin bedded limestones and graptolitic shales represent slow deposition on the continental slope whereas the thick breccia units represent sudden brief influxes of carbonate debris. There are major faunal differences between the two carbonate facies. It has been proposed (Stouge, in press;

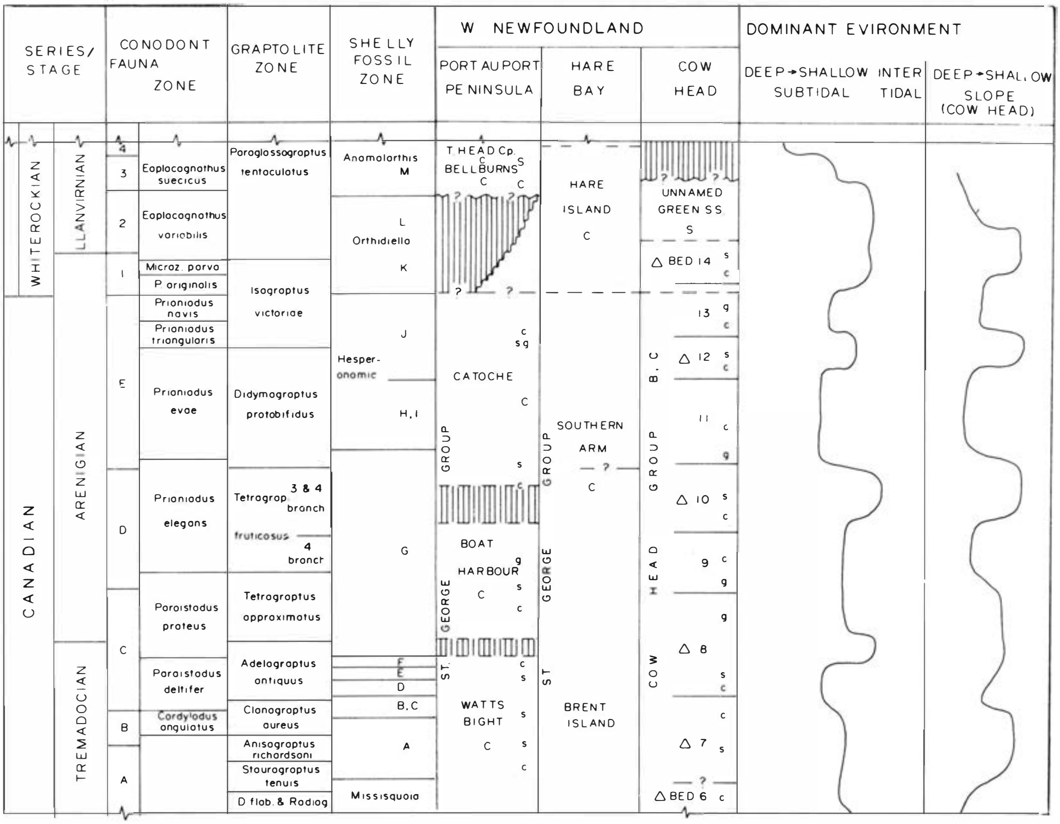

Fig. 4 – Stratigraphy, biostratigraphical control levels, and curves showing major changes in depositional environment for western Newfoundland. Open triangles show location of main megabreccias within numbered units (Beds) of Cow Head Group.

N. P. James and R. K. Stevens, pers. comm. 1981) that the megabreccias were generated during regressive phases when the carbonate platform margin was exposed and brecciated with the formation of karst surfaces. In Figure 4, the formal beds numbered within the Cow Head Group that contain the large megabreccias are shown with an open triangle. Several, but not all, seem to correlate well with the hiatuses demonstrated by stratigraphic studies and conodont biostratigraphy (e.g. Stouge, in press) in the St. George Group strata of the carbonate platform facies. Work in progress to try to refine further these events and correlations.

The carbonate platform suffered a major collapse in early Middle Ordovician time with the shallow water carbonates of the lower Table Head Group passing up into graptolitic shales in the upper Table Head. These shales are overlain by a flysch sequence and by obducted ophiolites. The continental margin is interpreted as being drawn down into an easterly dipping subduction zone. Some of the late Early Ordovician hiatuses may have resulted from a temporarily upwarped margin prior to collapse. However, as shown below, they do correlate with other apparent eustatic events elsewhere in Canada.

Summary

In four areas of Canada, the Lower Ordovician and lower Middle Ordovician stratigraphic record is remarkably complete and has been well documented in terms of its stratigraphy, sedimentology and paleontology: the southern

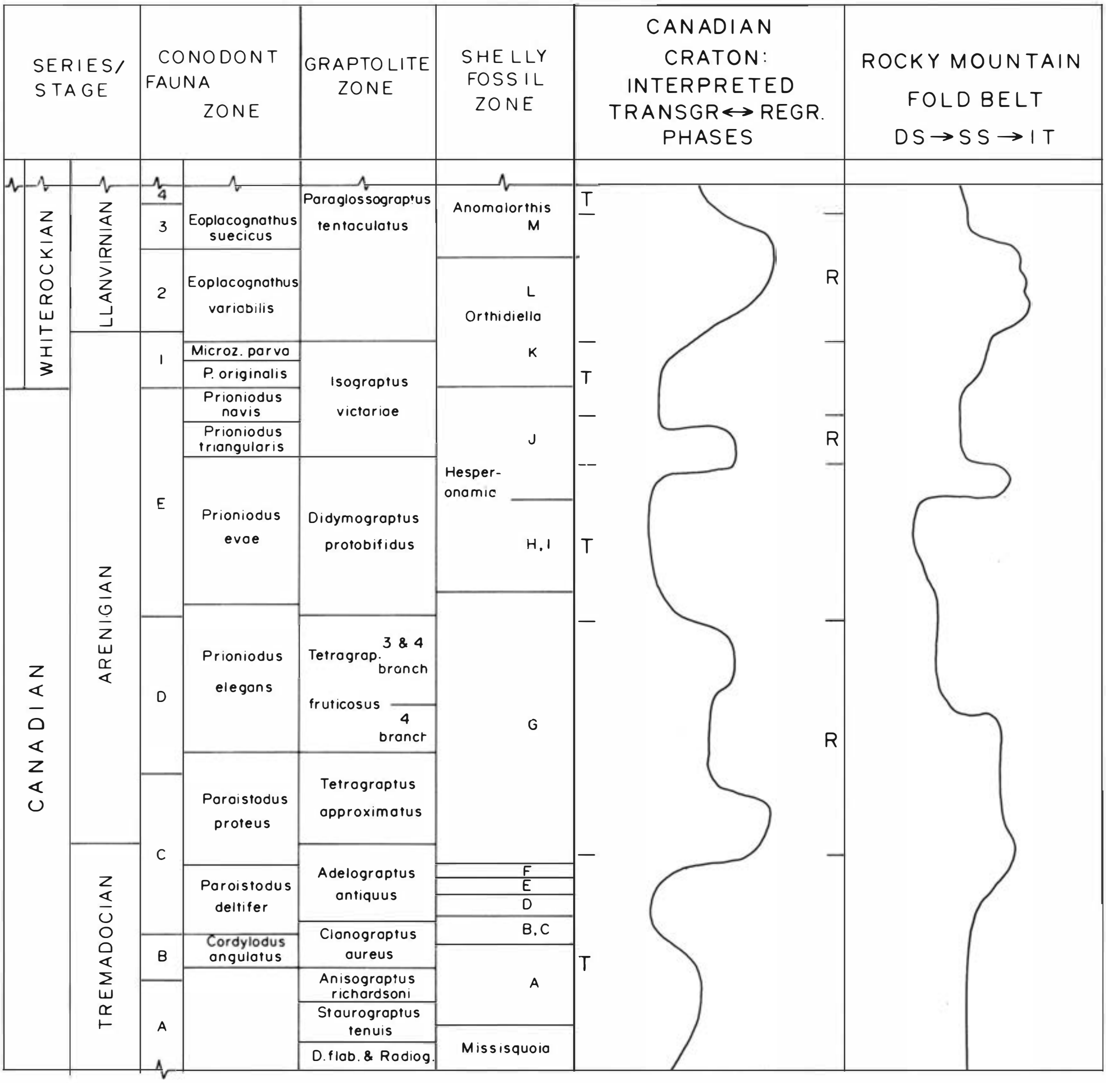

Fig. 5 – Interpreted transgressive–regressive phases for the Canadian craton during the Early and early Middle Ordovician, constructed by comparison of curves developed from southern Rocky Mountains, Mackenzie Mountains, Arctic Canada, and western Newfoundland.

Rocky Mountains, the Mackenzie Mountains, Arctic Canada, and western Newfoundland. Published data, not fully repeated herein because of space constraints, allows a curve to be developed for each area that plots the changing depositional environments. First order facies shifts are recognized and correlated against the chronostratigraphy and zonations adopted by Barnes, Norford and Skevington (1981).

In any particular region such major regional facies changes may be a result of various factors such as eustacy, progradation, epeirogeny, tectonism, or any combination of such factors. In this review paper, it is suggested that if major facies changes occur at the same time in the four widely separated areas around the margin of the ancient Laurentian craton then these are likely caused by eustatic sea level changes. This will need to be further checked against similar data from other cratons.

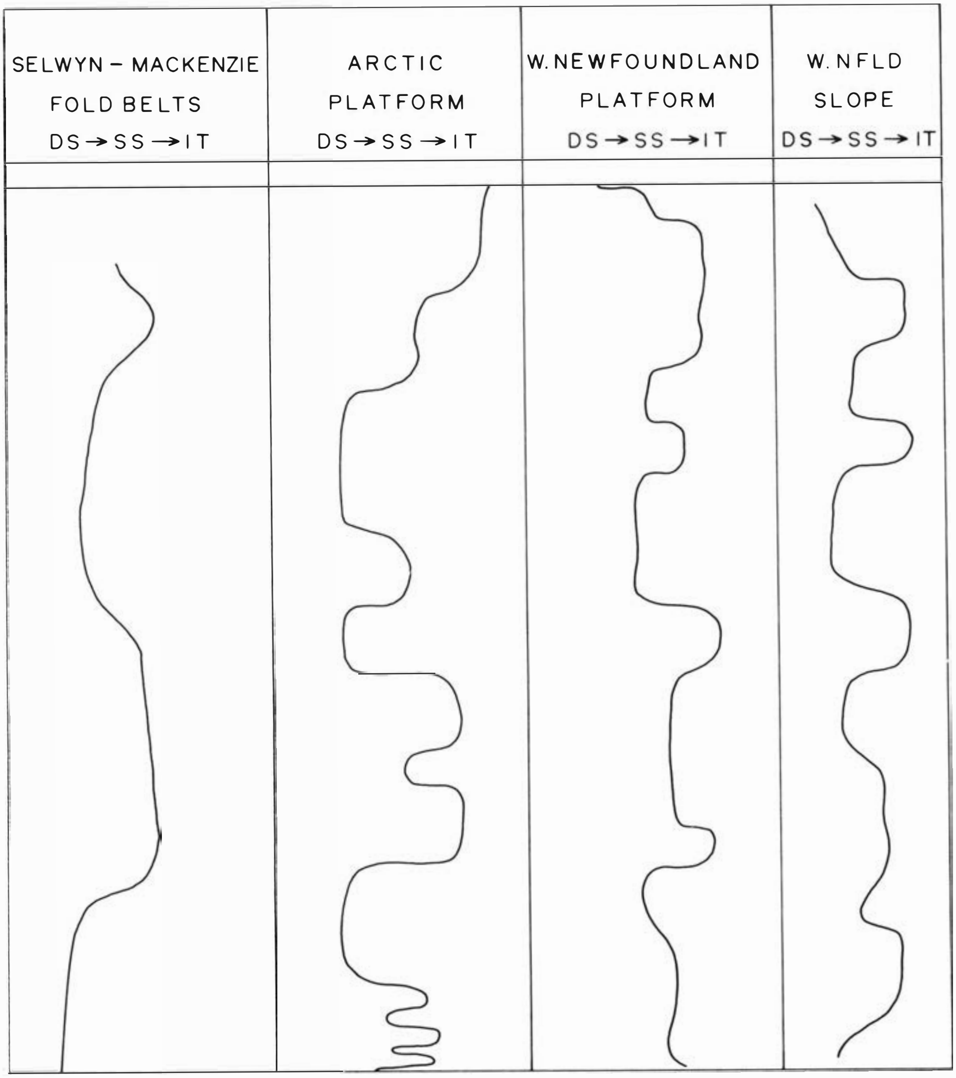

Figure 5 provides a comparison of the curves developed in the four areas. From these, a subjective generalized curve is derived which identifies major transgressive-regressive eustatic changes that affected the Canadian craton during the Early and early Middle Ordovician. Transgressive phases are recognized for the early and late Tremadoc, the mid to late Arenig and the late Whiterock. Major regressive phases occur during the early Arenig and the early Whiterock with a brief regressive phase in the late Arenig. These changes are portrayed as paleogeographic maps for the Tremadoc, Arenig, and Whiterock stages for the Canadian craton (Figs. 6–8), although such divisions cannot express the several eustatic changes that occur within each stage or series and must be viewed as generalized reconstructions. Although developed independently these results compare closely with the conclusions reached by

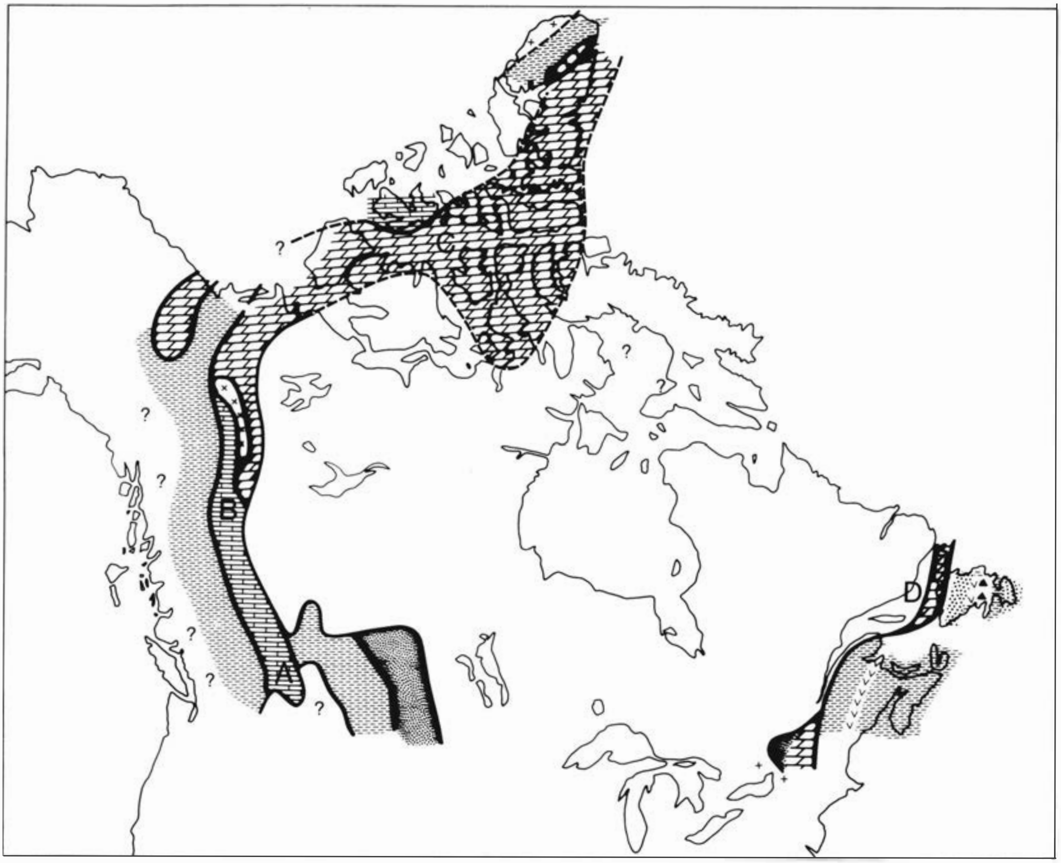

Fig. 6 – Paleogeographic recounstruction for the Canadian craton during the Tremadoc. General areas of southern Rocky Mountains, Mackenzie Mountains, Arctic Canada, and Western Newfoundland marked by letters, A, B, C and D, respectively.

Fortey's analysis (this volume) of Early Ordovician eustatic events from orther areas.

It is stressed that while full documentation could not be included here, it seems evident that eustatic events can be recognized for the Early and early Middle Ordovician in Canada. Similar unpublished data show that more detailed curves can be generated for the Middle and Late Ordovician. Although imperfect, it would be extremely valuable to see such curves generated for many other areas of the world in order that global coverage be attained to properly test the hypothesis that the curves do truly reflect eustatic events.

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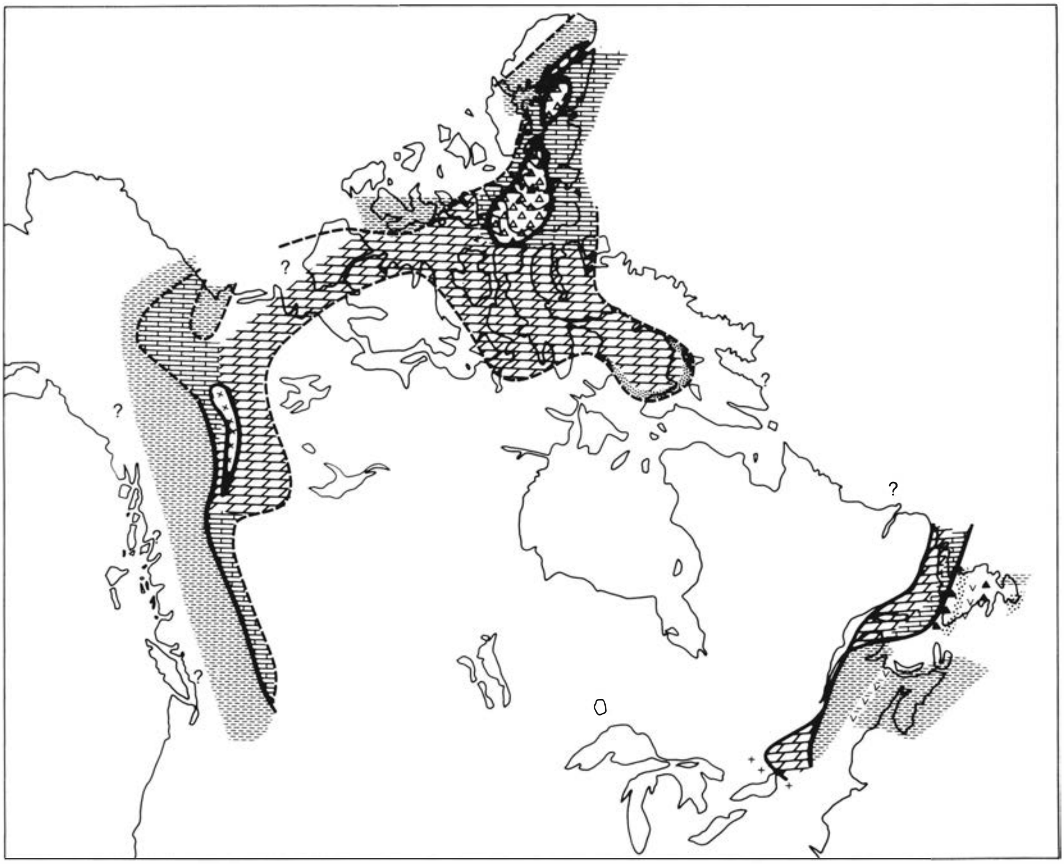

Fig. 7 – Palaeogeographic reconstruction for the Canadian craton during the early to middle Arenig. Open triangles indicate evaporite facies, crosses indicate local land areas.

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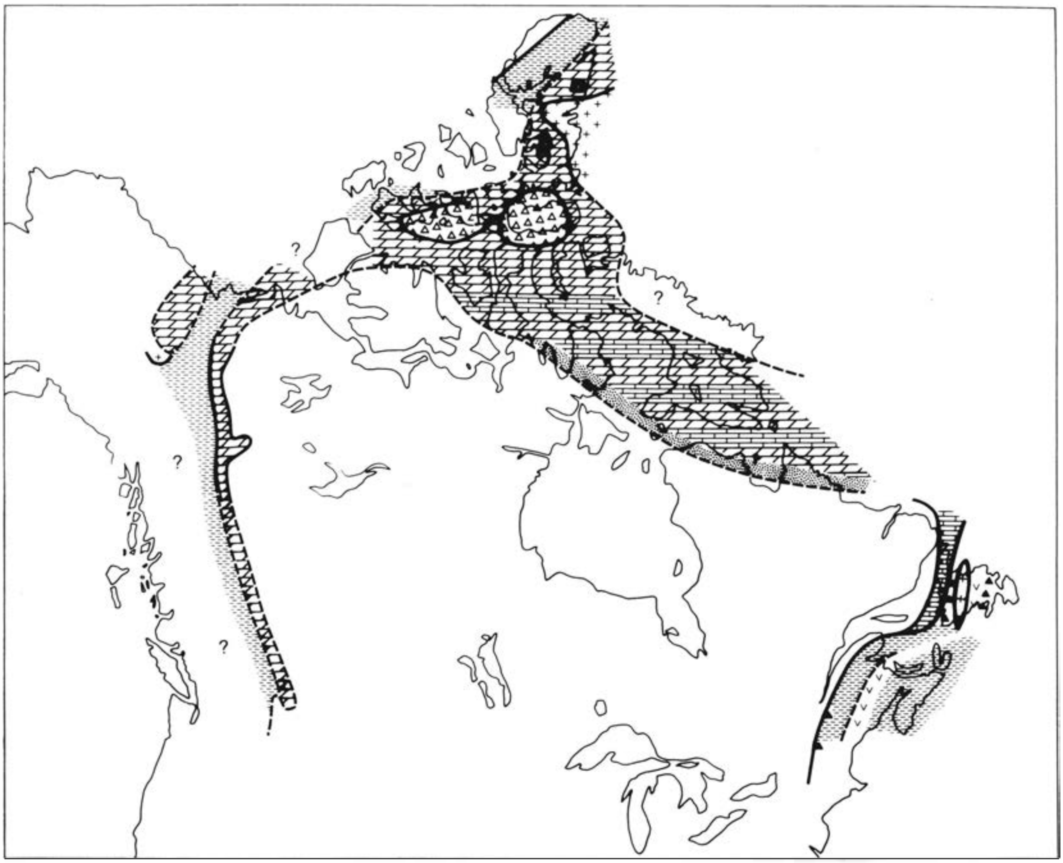

Fig. 8 – Paleogeographic reconstruction for the Canadian craton during Whiterock. Open triangles indicate evaporited facies.

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Late Ordovician environmental changes and their effect on faunas

By PATRICK J. BRENCHLEY and the late GEOFFREY NEWALL

It is estimated that the late Ordovician glaciation extended to 40° latitude from the southern pole, but that there was no glaciation in the northern hemisphere. There was a related < 100 m lowering of sea-level. The lowering of sea-level produced profound changes in palaeogeography, notably (1) disconformities in shelf areas with widespread karst surfaces in carbonate environments and channels on terrigenoclastic shelves, (2) deep erosion at shelf margins to produce marked disconformities, (3) a variety of syn-sedimentary deformation structures on shelves and slopes, (4) mass flow deposits and fans at the base of the slopes. The ensuing transgression flooded platform areas to produce a mantle of argillaceous sediment, often carbonaceous, which extended down the slopes and across the basin floors. These environmental changes, recorded from outgrop geology, confirm and expand previous models of transgressive/regressive sequences.

Contemporaneous with these late Ordovician environmental changes was a striking extinction of both benthonic and planktonic faunas. However, major taxonomic groups were differently affected and the timing of the extinction was also variable. The extinction of trilobites and graptolites occurred close to the Rawtheyan-Hirnantian boundary and preceeded the main drop in sea-level. Evidence suggests that the initial wave of extinction was related to global cooling while reduction in habitable area may have been important in later extinctions.

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Although there have been several glaciations during the history of the earth, the glaciation at the end of the Ordovician is of particular interest because the glacial maximum was pronounced but relatively brief and the resultant facies changes can often be precisely identified in the stratigraphical record. There is widespread evidence of continental glacial deposits on those continents which formerly composed Gondwanaland, and those in Saharan Africa, have been particularly well described by Beuf *et al.* (1971). The related glacio-eustatic sea-level changes caused environmental changes which can be recognised on separate continents and plates (Berry & Boucot 1973). Coincident in time and almost certainly causually related was an extinction of a substantial part of the late Ordovician biota (Berry & Boucot 1973; Sheehan 1973).

Because it is possible to identify the environmental changes related to the Gondwana glaciation in many stratigraphic sequences in a wide variety of environments, late Ordovician rocks are used as evidence in this paper to test both models of environmental change, resulting from eustatic sea-level change (Vail *et al.* 1977), and hypotheses concerning the causes of extinction. Furthermore, since the environmental changes caused by the rise and fall in sea-level were synchronous around the globe, and as they occurred close to the Ordovician-Silurian boundary, an understanding of these changes might influence decisions as to which is the best position to place the boundary between to two systems.

The Late Ordovician Glaciation

Glacial Deposits

Continental glacial deposits of approximately late Ordovician age are known from widely se-

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parated localities on the Gondwana continental plate. They are known from several sites in and around Saharan Africa, from South Africa, South America (see Spjeldnæs 1981 and references) and have been reported from Saudi Arabia (McClure 1978). Well preserved striated pavements provide convincing evidence of land-based ice and the variety of glacial geomorphological features, glacial and fluvio-glacial sediments and ice deformation structures are consistent with the former presence of a major ice-cap based on a generally flat continental shield area (Beuf *et al.* 1971; Allen 1975; Spjeldnæs 1981).

Tilloids, typically composed of mudstones containing dispersed angular or sub-angular carbonate clasts, together sometimes with well rounded quartz grains, are found at many localities in northern France and Iberia. Clasts with well preserved ice scratches have been described by Doré & Le Gall (1972) from the "Tillite de Fueguerolles" in Normandy which convincingly demonstrate the glacial origin of these tilloid beds. The similarity of texture and stratigraphic position of tilloids elsewhere in western Europe suggests that they have the same origin, though ice scratches have not been identified. However, the presence of "drop-stones" which have deformed delicate laminae in the Schistes du Cosquer in Brittany (Hamoumi 1981) also attests to the former presence of floating ice in the region. Tilloids of probably glacial origin are now known from localities in Normandy (e.g. Dangeard & Doré 1971; Robardet 1973), and Brittany (Hamoumi *et al.* 1980) in France; Celtiberia (Carls 1975), Montes de Toledo (Robardet 1982), and Sierra Morena (personal observation) in Spain; Valongo in northern Portugal (Romano & Diggens 1973–1974), and many localities in central Portugal (personal observation). The Ledershiefer in the German Democratic Republic (Greiling 1967) may also be of glacial origin.

Although the tilloids at most of these localities are rather similar in appearance, many of the vertical sequences differ in detail because there are variable amounts of intercalated laminated mudstones and bedded sandstones, commonly showing soft sediment deformation. Not only are the vertical sequences varied, but lateral facies changes can occur over a few kilometres. This variability was used by Greiling 1967) as evidence against a glacio-marine origin for the tilloids. However, there is no reason to expect uniformity of lithology under ice floating across a relatively shallow platform, particularly towards the distal margin of ice influence. This is confirmed by samples of Recent bottom sediments around the Antartic which show considerable heterogeneity (Anderson 1972; Anderson *et al.* 1977).

On the Gondwana continent the distribution of continental glacial deposits of late Ordovician age shows that land-based ice extended outwards across at least 40 degress of latitude from the south pole and that floating ice may have extended a further 10 degrees (Fig. 1). This is comparable to the spreads of ice at glacial maxima during the Pleistocene (Flint 1971). However, whereas the Pleistocene glaciation was bi-polar, the configuration of the continents in the late Ordovician (Smith *et al.* 1973; Ziegler *et al.* 1977) makes it unlikely that there was a north polar ice-cap during the Ordovician because open oceanic circulation apparently prevailed there.

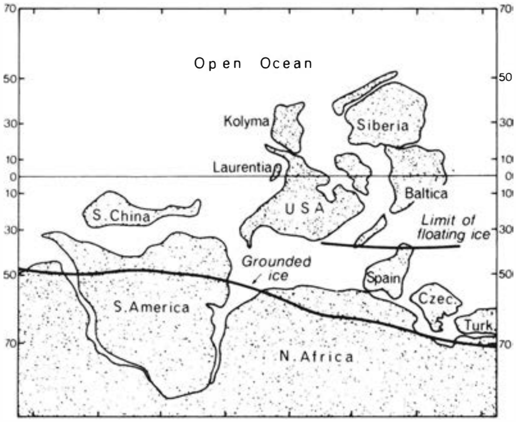

Figure 1 – Estimated limits of the late Ordovician ice-cap and of floating ice. The distribution of continents is based on the early Silurian palaeogeographical reconstruction by Ziegler et al. (1977).

Timing of the glaciation

Continental glacial deposits are notoriously difficult to date stratigraphically because they generally lack fossils. This is true both of Pleistocene deposits and more ancient tills where estimates of age are usually made by using well dated fossiliferous beds above and below. The late Ordovician glacial deposits are best dated in Morocco where an upper Ashgill, *Hirnantia* fauna is intercalated with them (Destombes 1968). In South Africa the main tillites lie above an upper Ordovician fauna and are overlain by sediments which have not been dated (Cocks *et al.* 1970). Elsewhere, all that is known is that the beds above the tillites are generally Silurian and the beds below are of rather variable Ordovician age.

An alternative approach to the dating of glacial periods is to monitor sea-level changes. The growth and decay of an ice-sheet requires that water be withdrawn and subsequently returned to the oceans and that these changes are reflected in glacio-eustatic sea-level changes and in the isotopic composition of sea water (e.g. Shackleton 1977).

In order to establish that eustatic sea-level changes have occurred it must be demonstrated that there were synchronous sea-level changes on several unconnected plates. It is more difficult to establish whether the changes were glacio-eustatic or were tectono-eustatic, i.e. the result of changes in the configuration of ocean basins. However, it does appear likely that there was a substantial difference in the rates of sea-level change associated with glacio-eustatic and tectono-eustatic events (Pitman 1978). Evidence from the Pleistocene indicates that whilst the growth of an ice-cap may take several million years, the cyclic changes of climate within the major glacial period last tens of thousand of years and cause large glacio-eustatic sea-level changes over the same time scale (Shackleton 1977). The decay of an ice-cap and the related rise in sea-level are apparently particularly rapid. In contrast tectono-eustatic changes usually operate over millions of years. In favourable circumstances it should be possible to discriminate between transgressive deposits formed by rapid glacio-eustatic sea-level and those deposits which were formed during the slower tectono-eustatic changes.

Throughout Ordovician times Gondwanaland apparently occupied a polar position so there existed the potential for a land based, polar icecap. The strong climatic zoning demonstrated by Spjeldnæs (1961, 1981) supports the existence of cold polar climates. Within the Ordovician there were several Ordovician transgressions at least one of which appears to have been eustatic (McKerrow 1979), but the evidence for eustatic regression is not strong, except at the end of the Ashgill. It is therefore uncertain whether there was any substantial growth or decay of ice-caps before the Hirnantian.

The late Ordovician sedimentary changes are, in contrast to those earlier in the Ordovician, particularly distinctive because in many sections there is an abrupt change from deep to shallow water deposits, and a subsequent abrupt reverse change to deeper water facies, reflecting a rapid rise in sea-level. These particular changes are widely recognised (Berry & Boucot 1973) on plates which were separate in Ordovician times, and fulfill the criteria for glacially controlled sea-level changes. Furthermore, these events correlate very well with the Moroccan evidence for the glaciation being late Ashgill.

Using this evidence of sea-level changes, it is possible to identify quite precisely the time at which the ice caps grew. The first evidence of regional shallowing occurs at the base of the Hirnantian, where in the type Ashgill sequence in England, for example, the Cautley Mudstones with a fauna generally dominated by trilobites give way upwards to mudstones with a sparse brachiopod fauna (Ingham 1966). A similar change occurs at the same stratigraphic level in the Oslo region of Norway (Brenchley & Cocks 1982). However, in both these places shelf mudstones persist through the lower part of the Hirnantian and the first influx of sandstones marking a strong regressive phase is higher in the sequence. It therefore appears that the drop in sea-level which drained many of the continental shelves did not reach completion until sometime well within the Hirnantian Stage (see bathymetric curve, Fig. 7). The reverse change from shallow to deeper water deposits is a sharp one in most sections and is at the top of the Hirnantian. It apparently occurred either within the *G. persculptus* Zone (lowest Zone of the Silurian) or at the base of the zone, but the exact position is still a matter for debate (Ingham & Williams 1982). If the late Ordovician glaciation spanned only the Hirnantian Stage and the glacial maximum occupied only a part of the stage it is likely to have lasted for less than 2 m.y. (Brenchley & Newall 1980).

Oscillations of sea-level within the glacial period

Pleistocene oceanic sequences preserve a record of $0^{16}/0^{18}$ isotopic composition in planktonic and benthonic foraminifera which shows that oceanic temperatures have changes through cycles of approximately 20,000, 40,000 and 100,000 years duration (Hays *et al.* 1976). The 20,000 and 40,000 year cycles are similar to those predicted by Milankovich (1938) on astronomical grounds and are believed to be related to changes in the orientation and obliquity of the earth's axis. The 100,000 year cycle appears to be related to the eccentricity of the earth's orbit, but why it should have a dominant influence is obscure because its likely effect on insolation appears to have been too small to have been a major climatic influence.

Similar climatic cycles might be expected to have taken place during earlier glacial periods, so we have looked for evidence, both in the field and from existing literature, for oscillations in sea-level within the span of the late Ordovician ice age. The evidence is unfortunately scattered and equivocal.

In the Hirnantian of the Oslo Region there are locally developed successions showing three tidal channel sequences stacked vertically one above the other (Brenchley & Newall 1980). These could reflect oscillation in sea-level, but equally well they might have been formed by three episodes of tidal channel migrations across the area in conditions of constant subsidence. Possibly more significant are the karstic surfaces, one within and one at the top of a Hirnantian carbonate sequence, in the Oslo Region (Hanken 1974). Further suggestive evidence comes from Iberia where, in Celtiberia an Ashgill limestone formation (the Urbana Limestone) has a karstic top reflecting emergence (Carls 1975) and is succeeded by glaciomarine tilloids (Orea Shales) indicating an episode of marine transgression. Quartzites above suggest a later regressive phase, whilst a further trangression is indicated by dark graptolitic shales of Silurian age. A rather similar sequence is developed in clastic rocks in the Dornes area of Central Portugal where shales pass upwards into quartzites in a regressive sequence. These are succeeded by tilloids, more quartzites and finally by shales of Silurian age (Cooper 1980). Although the stratigraphic control is poor at these horizons the sequences might reflect an early Hirnantian regression followed by a rise in sea level associated with floating ice, then a further period of emergence and finally the drowning of the region at the start of the Silurian.

In conclusion, although at least one oscillation in sea-level is tentatively recognised, there is no clearly preserved record of climatic cycles comparable with those of the Pleistocene. However, the Pleistocene record on the continental shelves is very incomplete, and without the foraminiferal and $0^{16}/o^{18}$ record in oceanic cores, the long and complex climatic history of the Pleistocene would not have been recognised. It seems unlikely that there was a long succession of late Ordovician glacials and interglacials, but the matter has not yet been proved.

Late Ordovician Environmental Changes

The evidence for world-wide development of regressive facies or disconformities and their relationship to glacio-eustatic sea-level changes has been reviewed bu Berry & Boucot (1973). The following account describes the environmental changes in stable platform to deep basin areas of Europe which in Ordovician times were situated in peri-Arctic to tropical latitudes.

Rawtheyan high standard of sea-level

While sea-level stood at a relatively high level during the Ashgill the continental shelves and platforms were covered by sea and blanketed by rather uniform terrigenous muds or carbonates (Fig. 2). The basin also received mainly muddy sediment while turbidite deposition was limited to tectonically active regions.

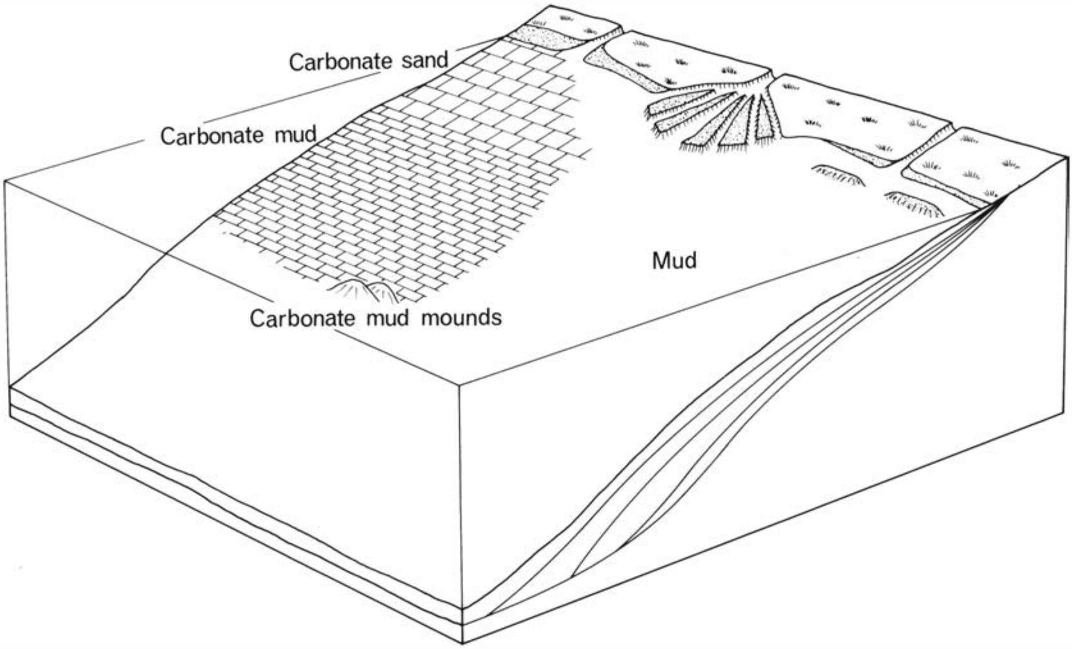

Figure 2 – Reconstruction of generalised facies distribution during Rawtheyan high stand of sea-level.

Hirnantian low stand of sea-level

The glacio-eustatic lowering of sea-level during the Hirnantian produced radical environmental changes both on shelves and in basins, which left a clear imprint on the stratigraphic record.

1) On clastic shelves in areas where sand was generally scarce, there was little or no aggradation of sediment during the period when sea-level dropped. Shales of Rawtheyan age were commonly cut by channels which subsequently filled with silts and sands, so that the sequences show Hirnantian sandstones lying on Rawtheyan mudstones with a sharp erosional contact (Brenchley & Newall 1980). Many of the sandstones filling the channels contain an *Hirnantia* fauna and may exhibit large-scale cross-stratification, but are more commonly massive, so that deposition within the channels appears to have been generally rapid. The exact timing of the filling is uncertain, but they might have been filled either during the initial regressive phase, during an intra-ice-age rising in sea-level, or during the rise in sea-level at the end of the glaciation.

2) On clastic shelves where there was a sufficient supply of sand to maintain continuous deposition during the early Hirnantian regression, there was shoreline progradation (Fig. 3), forming regressive, upward coarsening sequences (e.g. Oslo Region, Norway; Brenchley & Newall 1980; and Central Portugal, Cooper 1980). The shoreface sandstones associated with these sequences commonly have ball-and-pillow structures and other syn-sedimentary deformation structures suggesting that the rate of sedimentation of at least some of the beds was particularly rapid (i.e. centimetres per day).

3) In deep shelf areas there was continuous deposition of mainly argillaceous sediments from the Rawtheyan through the Hirnantian and these areas were never emergent. However, the beginning of a fall in sea-level at the Rawtheyan/Hirnantian boundary is indicated by a change from mainly trilobite-dominated faunas in the Rawtheyan to the brachiopod-dominated faunas of the Hirnantian (e.g. at Cautley in Northern England; Ingham 1966; near Bala in North Wales, Bassett *et al.* 1966).

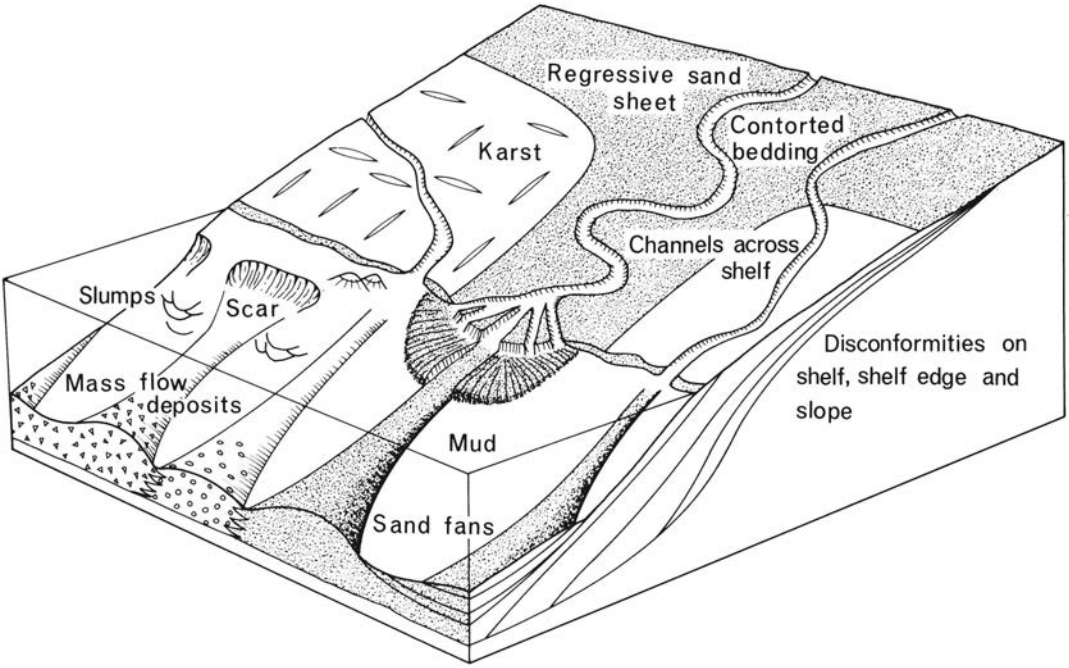

Figure 3 – Reconstruction of generalised facies distribution during Hirnantian low stand of sea-level.

Later in the Hirnantian there were times when channels were formed in these areas and filled with sediment derived from near the shoreline. The fill of these channels, which cut the deeper part of the shelf, is very variable and includes ooids at Bala (Bassett *et al.* 1966), and a variety of ill-sorted breccias suggesting rapid deposition (Cautley) and possible mass flows such as the breccias in the Coniston Limestone in Northern England (the latter were described as fault breccias by Mitchell 1956) and breccias at Portrane, Ireland (Lamont 1941).

4) Carbonate shelves were generally emergent during the Hirnantian and have karst surfaces (Fig. 3). In some areas there is relief of several metres on the erosional surface (e.g. Celtiberia, Spain; Carls 1975) but elsewhere the karst surface may only have a relief of tens of centimetres and the upper surface of the limestone is planar for tens or hundreds of metres (e.g. Cystoid Limestone, central Spain; Hafenrichter 1979). The carbonate mud mounds of central Sweden became exposed during the Hirnantian and have well developed microkarst surface on the crown and upper flanks (Brenchley & Newall 1980).

5) The shelf edge, never an extensive area, is not easily recognised in ancient rocks. However there is some evidence to suggest that the shelf margin might have been deeply notched during the Hirnantian low stand of sea-level (Fig. 3). At two places in Wales close to the shelf-slope break there is evidence of deep, late Rawtheyan or early Hirnantian erosion. At Bala, nearly 500 m of Moelfryn Mudstones (Rawtheyan age) could be missing locally (Bassett *et al.* 1966) and at Garth on the Towy anticline nearly 100 m of Rawtheyan mudstones are missing (Williams & Wright 1981). The stratigraphical relationships have been interpreted as unconformities of tectonic origin but an alternative explanation is that the erosion was caused by channelling of the shelf margin comparable to the notching of the shelf margin by canyon formation during the Pleistocene. The erosion surface at Bala is covered by a thin veneer of Hirnantian sandstones, whereas Garth there is a lenticular sandstone fill as

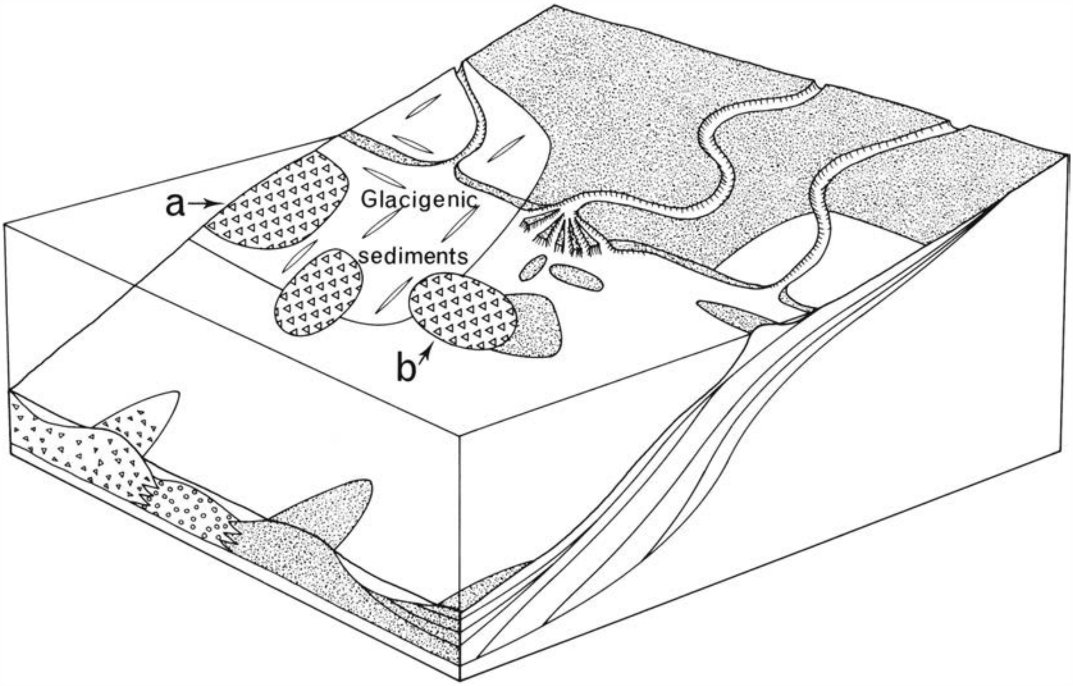

Figure 4 – Reconstruction of generalised facies distribution during minor rise of sea-level in late Hirnantian times, with deposition of glacio-marine tilloids. "a" indicates localities with tilloids over limestone with karst, "b" shows tilloids on a regressive sandy sequence.

much as 50 m thick.

6) On slope environments the Hirnantian is marked by a phase of sediment instability and the formation of large slump sheets (for example in the Towyn-Abergynolwyn District in Wales; James 1971). In stratigraphic sequences the monotonous mid-Ashgill shales are succeeded by shales and turbidites with slumps on a variety of scales.

7) In basin areas the Hirnantian was marked by an influx of gravity flow sediments, ranging from mass flow deposits to turbidites (Fig. 3). Major deep sea fans have been recognised in Wales at Plynlymmon (James 1972), Llangranog and Corris (James & James 1969). A base of slope fan, developed in much shallower water and containing bioclastic and shelf carbonates, was formed near Conway, North Wales (James & James 1969). The widespread evidence of sediment instability and accumulation of thick gravity flow deposits, suggests that sedimentation rates on the basin slopes and floor were unusually high during the Hirnantian. Comparably high rates of sedimentation, an order of magnitude higher than normal, are known from Pleistocene deep sea sediments (Davies *et al.* 1977). The late Ordovician glacio-eustatic sea level changes are reflected in basin areas such as the Welsh Basin by stratigraphic sequences with Ashgill mudstone overlain by gravity-flow deposits. The exact age of the change in sedimentation cannot always be determined but it is close to the Rawtheyan/ Hirnantian boundary.

Hirnantian glacio-marine deposition

Many localities in Iberia have sequences with tilloids which usually succeed either a regressive sandy sequence or lie on a karstic limestone surface. The presence of karst with only a shallow erosion surface at several localities suggest that the climate may have been cold and dry and chemical weathering was consequent-

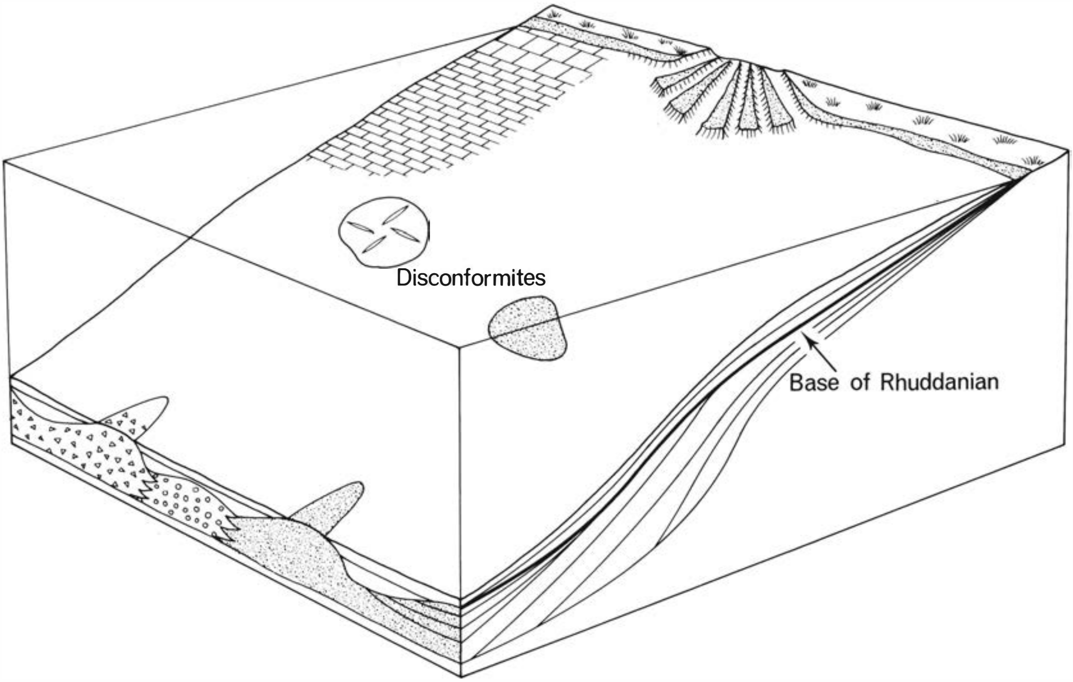

Figure 5 – Reconstruction of generalised facies distribution during Rhuddanian high stand of sea-level.

ly restricted. The presence of limestone clasts (of Ashgill age) in the tilloids indicates that ice was at least locally grounded on a limestone surface and that, although some mechanical erosion occurred, the ice cover may have prevented deep chemical erosion. Figure 4 shows a generalised reconstruction of the kind of palaeogeography which would have produced the Hirnantian sequences containing tilloids.

The Ordovician/Silurian rise in sea-level

In most areas in both shelf and basin successions there is an abrupt upward transition from the heterogeneous Hirnantian facies to the uniform, dark grey and generally carbonaceous shales of the Silurian (Fig. 5). In shelf regions the transgressive sea did not substantially rework and remould the underlying clastics (e.g. Oslo Region, Brenchley & Newall 1980), and the rise in sea level was apparently as fast as the Recent post-glacial rise (i.e. it spanned only a few thousand years). Only in the channels, which were incised across the shelves could there have been substantial deposition of sand during the post Ordovician rise in sea level, and even some of these were incompletely filled. In deep shelf, slope and basin successions, dark grey graptolitic shales usually lie with a sharp contact on the Hirnantian grey mudstones and turbidites.

The oldest graptolitic shales contain a fauna of the *G. persculptus* Zone, but at many localities, both in shelf and the basin sequences several zones are missing at this level and the lowest Silurian faunas may be as young as Wenlock age. Such disconformities in deep marine areas are becoming well known from D.S.D.P. cores, and apparently reflect a balance between rates of sedimentation and erosion by oceanic currents (Kennett 1982: 93). Experience of Tertiary to Recent sequences has shown that, although there were vigorous bottom currents and associated erosion during the Pleistocene, the high sedimentation rates nevertheless produced a substantial net accumulation. The periods for which the Tertiary record is most incomplete (Moore *et al.* 1978) are earlier in the Tertiary when although bottom currents were probably more sluggish, sedimentation rates were even more reduced. The early Silurian disconformity might therefore reflect a rise in sea-level which flooded the shelves and reduced sedimentation rates to a minimum.

The environmental changes outlined above are believed to have been the result of sea-level changes of 50–100 m (Brenchley & Newall 1980). The exposure of shelves during lowstand of sea-level with resultant formation of channels and disconformities, the development of submarine fans in deep water, and rapid coastal onlap during the major episode of sea-level rise are all predicted by the models of Vail *et al.* (1977) based on a knowledge of Pleistocene to Recent changes and evidence from seismic profiles.

The models we have outlined apply to a "usual" range of sedimentation rates and vertical tectonic movements and would be modified by a) unusually high sedimentation rates, or b) very active tectonics. High sedimentation rates would have particularly affected sequences developed during the earliest Silurian transgression. In extreme cases, instead of there having been coastal retreat at this time, unusually high rates of sedimentation could have caused coastal progradation even in the face of rapid sea-level rise (Vail *et al.* 1977; Heward 1981: 233). In such situations the Ordovician/Silurian boundary could lie within an unbroken sandstone sequence, as for example in the Queenston delta of the Eastern North American (Dennison 1976).

Active vertical tectonics with the same sense of movement as the sea-level changes could have effectively masked the glacio-eustatic effects. If, however, the sea-level changes were as rapid as we have suggested, tectonic movements of a comparable rate are likely to have been uncommon.

Because glacio-eustatic sea-level changes are synchronous, and because they leave a distinctive mark on the stratigraphic record, they have considerable chronostratigraphic significance, as in the final choice of level at which the Ordovician/Silurian boundary should be drawn. The widespread development of a disconformity at the top of the Hirnantian, both in shelf and basin areas suggests that if the boundary is placed either at the base of the *persculptus* zone or the base of succeeding *acuminatus* zone it will lie within a depositional hiatus in many sections throughout the world.

The late Ordovician extinction

An extinction episode of late Ordovician age was one of several such events recognised by Newell (1967) in his review of the Phanerozoic fossil record. More recently Raup & Sepkoski (1982) have identified a late Ordovician peak of family extinction, which is significantly above background extinction levels (Raup & Sepkoski 1982, Fig. 1) and is one of four mass extinctions which they identify in the fossil record.

The late Ordovician extinction peak is, however, almost certainly a composite peak composed of a late Caradoc to early Ashgill reduction in species diversity and a separate late Ashgill episode of extinction. The first wave of extinction resulted from plate movements which reduced the width of the Iapetus Ocean and allowed interchange of benthic faunas. Separate trilobite faunal provinces on either side of the Iapetus Ocean had lost their distinctive character before the middle Caradoc (Whittington & Hughes 1972) and brachiopod provinces had essentially merged by early Ashgill times (Williams 1976). The unification of these separate provinces into one single province almost certainly accounts for the reduced late Caradoc early Ashgill diversity in some groups. The second wave of extinction is stratigraphically quite distinct from the earlier one, and was confined to the late Rawtheyan and Hirnantian stages.

The magnitude of the extinction was considerable but different groups were affected to a different degree. Most data on changes in taxonomic abundance are plotted for series and not stages, so the following account initially refers to extinctions within the Ashgill Series. Later, data will be used to demonstrate specifically Hirnantian extinctions.

There is little evidence of a late Ordovician extinction of orders, but several phyla show evidence of reduction in the number of families and genera present (Fig. 6). Thirty eight families of trilobites in the Ashgill were reduced to fourteen in the early Silurian (Jaanusson 1979)

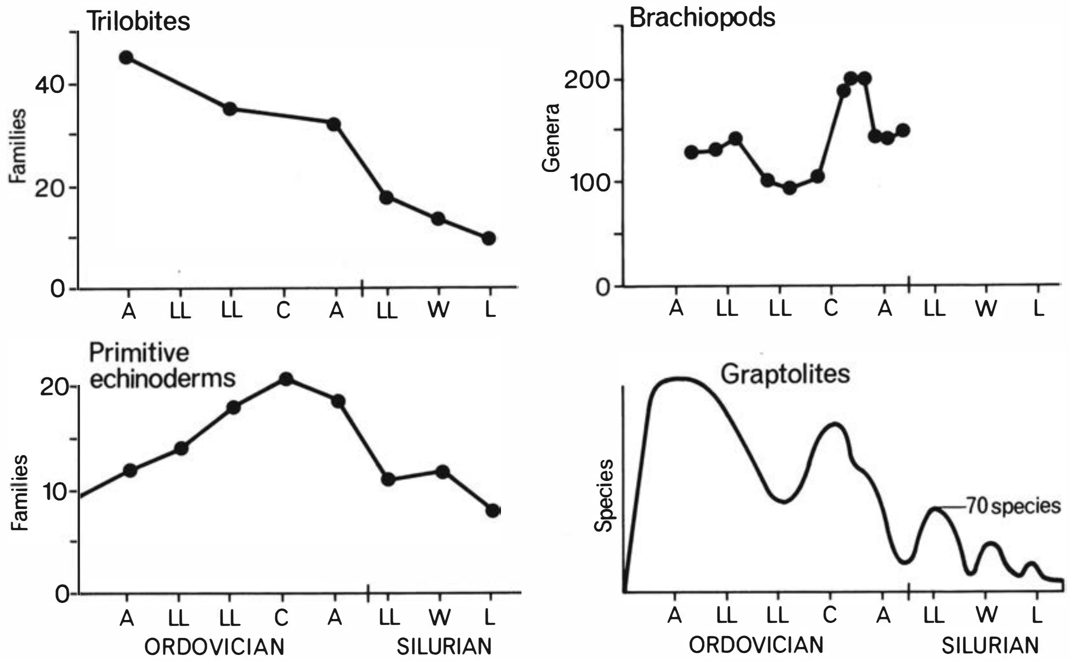

Figure 6 – Changes in taxonomic abundance through the Ordovician and Lower Silurian of (a) trilobites (data from Harrington 1959), (b) brachiopods (after Fig. 151, Williams 1965), (c) Cystoidea, Cyclocystoidea and Edrioasteroidea, unpublished data C. R. C. Paul) and (d) graptolites (after Fig. 3, Koren & Rickards 1979).

and cystoids too show a substantial drop in family diversity (Paul 1980). Generic and species diversity was substantially reduced in groups such as the brachiopoda (Williams 1965) and graptolites (Koren & Rickards 1979). Other elements of the plankton such as acritarchs (C. Downie, pers. comm.) and conodonts (cf. Orchard 1980) were apparently reduced in specific diversity in Hirnantian rocks though this stratigraphic interval still remains to be studied in detail. The diverse tabulate and heliolitid coral faunas of the late Ordovician, which are represented by about 70 genera were drastically reduced by the loss of about 50 genera prior to the Silurian (Kaljo & Klaaman 1973).

From the above discussion it is clear that the late Ordovician extinction affected most elements of the biota, including the sessile, filter feeding, shelled benthos, the vagile benthos, the phytoplankton and zooplankton.

The following data suggest that the time of extinction was slightly different for different groups, and ranged from late Rawtheyan or early Hirnantian times to late Hirnantian times. At least 10 of the trilobite families that became extinct, did so in the Rawtheyan and especially towards the end of the stage. At the generic level they show there was a high percentage of survival from stage to stage through the Ashgill until the end of the Rawtheyan, when only about 15% survived into the Hirnantian (Fig. 7). These figures are drawn from European localities which would have ranged from mid-latitudes to the southern tropics in late Ordovician times, but the data appear to be valid for all the climatic zones.

The timing of graptolite extinction relative to that of benthonic groups is difficult to determine because the late Ordovician graptolite zones are not precisely correlated with the Ash-

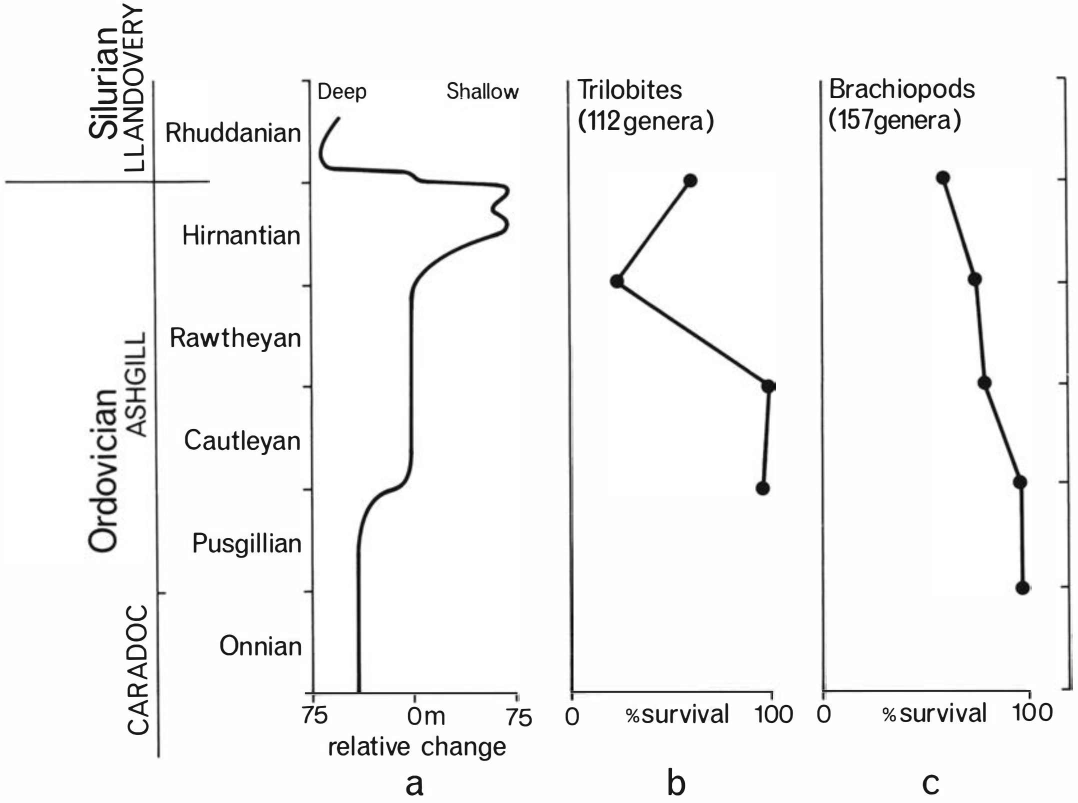

Figure 7 (a) Estimated relative changes in sea-level in the Upper Ordovician (Brenchley & Newall 1980), (b) Percentage survival of genera from stage to stage in the Upper Ordovician/Lower Silurian e.g. 93% of Pusgillian tilobite genera from selected localities in Europe, survived into the Cautleyan, (c) Percentage survival of brachiopod genera. The graphs are based on genera found in sequences with good stratigraphic control in Britain, Scandinavia and Poland.

(The data are drawn from twenty two references, including Bassett et al 1966; Bergström 1968; Brenchley & Cocks 1982; Dean 1959, 1971, 1974, 1977; Hiller 1981; Ingham 1966; McNamara 1979; Owen 1981; Price 1980; 1981; Temple 1965; Williams & Wright 1981, and Wright 1963, 1964). Data concerning total ranges of the genera are mainly from the Treatise of Invertebrate Palaeontology.

gillian stage boundaries based on shelly facies. The trough in graptolite diversity occurred in the *extraordinarius* Zone and though it may have extended into the *persculptus* Zone (Koren & Nikitin 1982) it certainly preceded the end of the Hirnantian and could thus have coincided with trilobite extinction at the Rawtheyan/Hirnantian boundary, but this has yet to be proved. The conodonts and acritarchs appear to have been reduced in numbers at the Rawtheyan/Hirnantian boundary within Europe but elsewhere assemblages persisted in more tropical regions and on Anticosti Island (Achab & Duffield 1982) are found nearly up to the Ordovician/Silurian boundary. There is however, a short length of section at the boundary which lacks chitinozoa and a significant change in the composition of the acritarch floras occurs here. Skevington (1974) has suggested that the low level of provinciality amongst graptolites in late Ordovician times reflects a reduction in the number of habitable climatic belts to a single tropical zone. This interpretation could be reasonably applied to the other planktonic groups.

In contrast to the trilobites, the brachiopods do not appear to have been nearly so severely affected at the Rawtheyan/Hirnantian boundary (Fig. 7), and the persistence of many genera from the Rawtheyan into the Hirnantian is reflected in some Hirnantian brachiopod faunas of considerable diversity (Bergström 1968; Williams & Wright 1981; Brenchley & Cocks 1982). There was, however, some reduction of diversity before the Hirnantian (Lespérance 1974) and a further reduction before the basal stage of the Silurian, implying a late Hirnantian extinction.

Two principle hypotheses have been proposed to account for the late Ordovician extinctions. One is a lowering in water temperature related to the Gondwana glaciation, and the second is exposure of the continental shelves following the glacio-eustatic fall in sea-level (Sheehan 1973, 1975, 1979; Jaanusson 1979).

The significance of the stratigraphic dating of the extinctions outlined above is the demonstration that the first wave of extinction commenced when sea-level started to fall but preceded the main rop in sea-level. This fall is unlikely to have caused extinction by reducing the area of the continental shelves. On the other hand, the second wave of extincttions which reduced the variety of shelly sessile benthos could have been related to lowered sea-level and the decrease of habitable areas.

The cause of the first wave of extinctions remains obscure, particularly because substantial changes of temperature during the Pleistocene glaciation did not produce such waves. Biotas were apparently able to move with the shifting climatic belts (Berger & Berger 1981; Ford 1982) and, given time, some appear to have been able to adapt to a temperature regime outside their accustomed range (Ford 1982: 29). It could be that the crucial effect was not temperature *per se* but the cooling of surface waters outwards from the polar region which contracted the plankton belts to such a degree that habitable areas was severely reduced, and extinction resulted. Moreover, the very extensive epicontinental seas of the mid-Ashgill may have had niche-specific, highly adapted faunas, which were ill-equipped to withstand rapid environmental changes. This was in marked contrast to the conditions in the Pleistocene. Furthermore it is likely that the rate at which the Ordovician ice cap developed was considerably faster than the Pleistocene. There appears to have been a gradual cooling of climate from the early Tertiary and ice probably began to develop in the Antarctic at least as early as the Miocene (Kennett 1982: 730). The build up of the polar ice caps was progressive, though step-like, over as much as 20 million years. The evidence for the Ordovician though less precise, suggests that, although polar climates may have been present throughout the Ordovician, there was a particularly rapid onset of glaciation at the beginning of Hirnantian times implying a rapid decrease in marine temperatures and a sharp contraction of the climatic belts in a matter of 1 million years. Temperature, and related climatic changes certainly have a profound affect on the distribution of fauna in the short term (Ford 1982) and if the decline in temperature was sufficiently rapid this might have been fatal to many species.

It appears likely that no single factor caused extinctions (cf. Jaanusson 1979), but that extinctions were caused by a complex combination of circumstances. Nevertheless we believed it is possible to isolate some of the major causes of extinction, and that contraction of the climatic belts in early Hirnantian times as envisaged by Skevington (1974) and Sheehan (1979) was initially a significant factor while contraction of habitable area in later Hirnantian times was a later contributory cause. It is also possible that there were even further extinctions, as Jaanusson (1979) has suggested, when the early Silurian rise in sea level flooded many shelf areas to considerable depths and there was an accumulation of black euxinic muds inimicable to a bottom living shelly benthos.

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The Ordovician climate based on the study of carbonate rocks

By MAURITS LINDSTRÖM

The climate evidence contained in Ordovician carbonate rocks is extremely difficult to decipher because of large gaps in our knowledge of Recent carbonates and ignorance of things like the ecological demands of many contributing Ordovician biota, geophysical parameters for the Ordovician, and depth of deposition of important Ordovician carbonate rocks. Areas, for which a subequatorial position is indicated by palaeomagnetic evidence, resembled Recent subequatorial areas by containing carbonates with algal structures and oolites. Unless sea-water temperatures were considerably warmer in the Ordovician than in the post-Palaeozoic, cold waters, (including relatively deep water), might have been characterized by trilobites as major sediment contributors. A relatively high carbonate concentration in sea-water is indicated by persistent carbonate mud deposition and cementation at high latitudes. Eustatic events are reflected in carbonate successions: if these events are of glacial origin, their existence suggests that Ordovician climate zonation might have resembled the present one. Under this condition, and assuming that approximately $-5.5^\circ/oo$ $^{18}0$ (PDB) corresponds to $\pm$ $0^\circ C$ sea-water temperature for the early Ordovician, $^{18}0$ data can probably be used for palaeotemperature determinations. If the basic interpretations involved are correct or nearly so, mean annual temperatures of marine surface waters at 60°S may have been near + 8°C, which differs very little from present values.

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Climate evidence in Recent marine carbonates is a difficult theme, and one finds little enthusiasm when discussing it with marine scientists. Our state of ignorance regarding the Ordovician makes it almost certain that the accuracy of interpretations is inversely proportional to the assurance with which they are made. The ground that can be covered is well treated in two papers by Spjeldnæs (1961, 1981) and the reader is referred to these for references and discussions on Ordovician palaeoclimatology. In the following I shall deal mainly with climatic interpretations of carbonate rocks.

Any deposit about which the original latitudinal position is unknown is more likely to have been deposited within 30° of the equator than in any other zone because slightly more than half of the earth's surface is between those latitudes. The surface ratio between the zones: equator to 30°, 30° –60°, and 60° to poles is roughly 4 : 3 : 1. Arctic and subarctic sediments should not be particularly widely distributed. Add to this that most limestones form in subtropical to subequatorial climates and inevitably, the first guess regarding any carbonate succession would be that it formed in relatively warm water.

It is difficult to discuss the effects of climate unless the approximate depth of deposition is known, as is the case with the large areas of Ordovician sedimentary carbonates in North America. However, there is conflict of evidence and opinion as regards the depths of deposition of the important carbonate successions of Baltoscandia. Otherwise a comparison between these two provinces of carbonate sedimentation could be expected to be rewarding.

Evidence of cold climate is restricted mostly to clastic sequences (Spjeldnæs 1981). Theoretically, in the case of carbonate successions

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such evidence might consist of dropstones, glacially fractured sand grains, ice-push-structures, or tillites or structures of glacial origin occurring in the same narrow palaeogeographic province. I do not know of any Ordovician limestone with dropstones, but such beds are known from the Permian of Tasmania (Rao 1981a) and they might occur in other Systems. Lindström (1972) reported the occurrence of glacially fractured sand grains in a Lower Ordovician limestone in Sweden, but they have not been found in the large number of other beds sampled. However, in view of palaeolatitude determinations by Noltimier & Bergström (1976), it is not unlikely that Sweden, positioned at 60°S, was occasionally invaded by drift ice that carried sand populations with glacial markings.

In arctic to cool-temperate littoral and shallow-water deposits, one might expect to find occasional ice-push structures and depressions caused by stranded ice blocks but I do not know of any such structures from Ordovician limestones. Outside the continuous ice margin, grounding ice may normally deform the bottom sediment to depths of 30–75 m (Dell 1972; Reimnitz *et al.* 1972).

Some generalizations apply to climate included processes at all ages, but generally speaking climate and its effects depend on so many variables that climatic reconstructions are uncertain even for the immediate geologic past. Conclusions drawn from comparisons between ancient and modern sediments must be supported by other evidence, such as palaeomagnetic data. Energy transfer across the latitudes depends on whether or not polar ice caps are present. Spjeldnæs (1981) doubts the existence of polar continental ice in the earlier parts of the Ordovician, while other authors (Fortey, this volume; Barnes, this volume) consider the possibility that continental glaciation was responsible for worldwide lowering of sea-level in the Ordovician prior to the Caradoc. The actualistic aspect of Ordovician climate zonation will depend on whether such glaciation did or did not exist. Furthermore, it is not altogether certain that the diameter of our planet was as great in the Ordovician as it is now (Carey 1976; Glikson 1980). If it were considerably smaller, then this circumstance must have influenced the width and stability of climate belts. Last but not least, vascular land plants influence the flow of sediment to modern oceans by promoting weathering and reducing the amount of mechanical erosion; they were of negligible importance in the Ordovician (Gray *et al.* 1982). This environmental difference greatly reduces the usefulness of actualistic comparisons (Spjeldnæs 1981), although comparison with Recent conditions can be illuminating.

Actualistic model

1) Temperature

In the zone between 0° and 30° latitude, one can expect normal mean annual ocean surface temperatures of 20–27°C; between 60° and the poles corresponding temperature in Recent oceans is 0°C and lower (Moore 1972). The surface temperature gradient is greatest, 0–20°C, between 30° and 60° latitude, and the mean annual variations is also greatest. It reaches a mean range of about 14°C at 40° latitude and is 2°C and lower in subequatorial and subpolar areas. The tidal range is also greatest in the 30°–60° latitude belt. This belt shows the greatest physical gradients and can be expected to have the largest rates of transfer of physical energy in its shallow waters. Possibly, the ocean was appreciably warmer in the Ordovician than at present (Schopf 1981). This must if anything have steepened the energy gradients in areas towards the poles.

Temperatures in the open ocean are generally not above 10°C at depths greater than 500 m; as a rule one can reckon with stable temperatures of + 4°C or less at this and greater depths. In landlocked oceans and very extensive shelf areas, even the deepest parts can have stable temperatures approximately equal to mean annual surface temperature. Deep water facies should contain evidence of environmental stability in any climate. Moisture is another important climate factor. There is a net water loss from the ocean between 10° and 40° latitude, with a maximum about 22° (Starr & White 1955). These are the dry areas, with surface waters of high salinity and desert or semidesert conditions over much of the land. There is a net flux of water vapour polewards from 40°, so there will be a tendency for surface waters to freshen in this direction. Again, the most rapid variation is in the 30°–60° latitude belt. The equator receives a certain net influx of water vapour.

2) Clay minerals

To some extent the above conditions are reflected by the clay mineralogy. Kaolinite is produced mainly in the warm, humid equatorial ocean areas (Rateev *et al.* 1979; Kolla *et al.* 1976). Near-polar areas have relatively much chlorite. A large amount of kaolinite in the non-carbonate residue of a limestone might indicate sedimentation in the warm zone. However, considerable amounts of kaolinite derived from tropical weathering during much older sedimentary cycles can be found even in Arctic seas (Bjørlykke & Elverhøi 1975). A further problem is that in the absence of vascular plants, weathering conditions in the Ordovician might not have favoured the formation of kaolinite. However, Spjeldnæs (1979) reports kaolinite as an important consituent of the Middle Ordovician Harding Sandstone, thought to have been deposited near the palaeoequator in western North America. This is precisely where one would expect it to be according to an actualistic model. On the other hand, Lower Ordovician limestones of southern Sweden (deposited at approximately 60°S according to Noltimier & Bergström 1976) also contain appreciable quantities of kaolinite (Lindström & Vortisch 1983).

Sediments from the present day arid tropical zones may be lacking in kaolinite. To take an Ordovician example, the type Cincinnatian, a largely carbonatic succession deposited at a latitude that would have been arid by comparison with Recent climate zonation, contains little if any kaolinite (Booth & Osborne 1971). Because kaolinite can be derived from much older deposits, and because it can be transported a long way, for instance by winds, the absence of this clay mineral in some cases might be more revealing than the presence of it.

Comparison with the distribution of carbonate in present oceanic sediments (Lisitzin 1971), suggests that carbonates in the temperate and cold zones should be first and foremost relatively sparse, and, furthermore, impure. The carbonate content of Recent marine sediments decreases greatly towards the continents, with the exception of some arid areas, like Australia. It decreases rapidly towards the subpolar and polar zones. Whitman & Davies (1979) give an actualistic model for a north-south oriented ocean with narrow shelf areas. According to this model shallow waters are likely to be dominated by terrigenous mud, except at arid continents. By comparison with present conditions it would be unlikely that any major, epicontinental carbonate succession formed outside a belt 30° north and south of the equator, yet early Ordovician carbonate sedimentation took place persistently over large areas of Baltoscandia, which was about 60°S (based on consistent palaeomagnetic data Noltimier & Bergström 1976). This observation is clearly at odds with the actualistic model.

Carbonates

The present ocean is undersaturated with respect to carbonate and carbonate deposited in cold water shows evidence of early solution on grain boundaries (Alexandersson 1976). I do not know anly clear evidence of this kind from the Ordovician. Discontinuity surfaces in Ordovician limestone have been described by Jaanusson (1961) and they have beeen referred to as corrosion surfaces, implying unknown amounts of solution. The surfaces referred to in such terms (Fig. 1) undoubtedly show evidence of loss of material, but it is seldom evident that the material was cemented before removal, or that the removal was effected by solution. The morphology of many discontinuity surfaces in the Lower to early Middle Ordovician of Baltoscandia suggest that the surfaces represent the upper boundaries of cemented, and therefore erosion-resistant, portions of carbonate beds. The overlying, non-cemented portions might have been removed by currents (Lindström 1979).

The persistence of carbonate sedimentation and preserved cementation, in the presumably quite cool early Ordovician water of Baltoscandia (Jaanusson 1979), suggests that the ocean here was not as undersaturated with carbonate as it is now. Because abundant carbonate sedimentation went on in very extensive shelf seas, large quantities of carbonate cations must have been carried regularly from land areas to the sea. In other words, sufficient land area must have been exposed to constant weathering.

According to Lees (1975) modern shelf carbonate sediments are made up mainly of associations of skeletal and non-skeletal components. While there might be no harm in attempting to compare Ordovician carbonates with these associations, the great differences for instance between Recent and Ordovician skeleton-producing biota make it necessary to regard the results of such attempts with much scepticism. The non-skeletal group of associations might be the least controversial object of comparison. Non-skeletal pellet associations ("bahamites") require high salinity and temperature and are subtropical to equatorial in distribution. They have been reported from the late Ordovician of Baltoscandia (Jaanusson 1973), the Lower Ordovician of Argentina (Serpagli 1974), and the Ordovician of North America (Cloud & Barnes 1957; Read 1980). A further association, with ooliths and grapestone aggregates, requires salinities of at least about 36°/oo in the present ocean: within the tropical and arid zones. Such facies have been reported by Mazzollo & Friedman (1975) from the Lower Ordovician of North America. Lees (1975) identified three skeletal component associations, the foramol, chlorozoan and chloralgal.

The foramol association contains benthic foraminifera, bivalves, barnacles, bryozoa, and calcareous algae as typical, but not omnipresent, components. This association is widespread and not very diagnostic. Since neither benthic foraminifera nor barnacles occur in the Ordovician and bivalves are rare in many Ordovician limestones (their place being largely occupied by the brachiopods), it might be hazardous to identify any particular Ordovician skeletal association with the foramol.

The chlorozoan association is characterized by contributions from corals and calcareous green algae. This association occurs within 30° north and south of the equator and requires temperatures not below + 15°C and warmest annual temperatures of at least + 25°C. If stromatoporoids are accepted as members of the equivalent association in the Ordovician (Webby 1980), then this can be identified throughout much of the North American shelf, whilst in Baltoscandia it first appears in the Middle Ordovician at a time when this geological province might have drifted into the warm climate belt.

The chloralgal association contains calcareous green algae but lacks coral because of extreme salinity. This is the algal-mat association that can be identified in much of the North American Ordovician shelf areas and in Baltoscandian Middle and Upper Ordovician.

Lees (1975) did not classify calcareous muds, though he indicated that they are sparse outside the warm zones and occurrences outside the tropics are impure and sporadic. Nevertheless, the Lower Ordovician of Baltoscandia, though deposited probably at about 60°, characteristically contains relatively pure carbonate mudstones.

Rao (1981a) described a bryozoan-rich limestone with glacial dropstones from the Lower Permian of Tasmania. Early cementation of these beds was probably connected with upwelling and the same appears to be the case with modern cold-water cementation of bryozoan-rich calcarenites (Rao 1981b). Somewhat similar instances might have occurred in the Upper Ashgill of Spain (samples from San Benito, courtesy E. Serpagli) and Brittany (Calcaire de Rosan, samples provided by Y. Plusquellec; see Hamoumi 1981). These beds contain biocalcarenites rich in bryozoans and echinoderms and are roughly coeval with tillites that occur in the same geological provinces.

Since the study of the Ellenburger Group by Cloud & Barnes (1957) it has been well established that during the Ordovician the North American craton was warm and shallow, although the depth of deposition of other areas is often difficult to judge. Biological evidence is of little help in determining depth, because it is difficult to differentiate between the affect of cold climate, and coolness owing to deep water (Taylor & Forester 1979). Furthermore, most of the known groups might have been relatively well represented even at great depths. This generalization even applies to ahermatypic corals (Sartori 1980). However, certain groups tend to form a greater proportion of shallow-water than of deep-water communities. In Recent oceans corals obviously belong to the shallow-water environment as do the majority of bryozoa (Hyman 1959; Dahl *et al.* 1976; Carey 1981). Organic productivity (but not necessarily diversity) is much greater in relatively warm and shallow sees than in cold (and deep) water (Clarke 1962; Sokolova 1972; McGowan 1977).

Autochthonous algae and desiccation cracks are the most reliable criteria of very shallow water although identification of the algae must be accurate. Desiccation cracks are notoriously difficult to identify as such, even where the internal structure is known. The best arguments for the widely accepted hypothesis that the Lower Ordovician rocks in Baltoscandia were deposited in shallow water (Jaanusson 1982), are based on the identification of shrinkage cracks (Jaanusson 1973), and stromatolites (Larsson 1973), although the definitive details

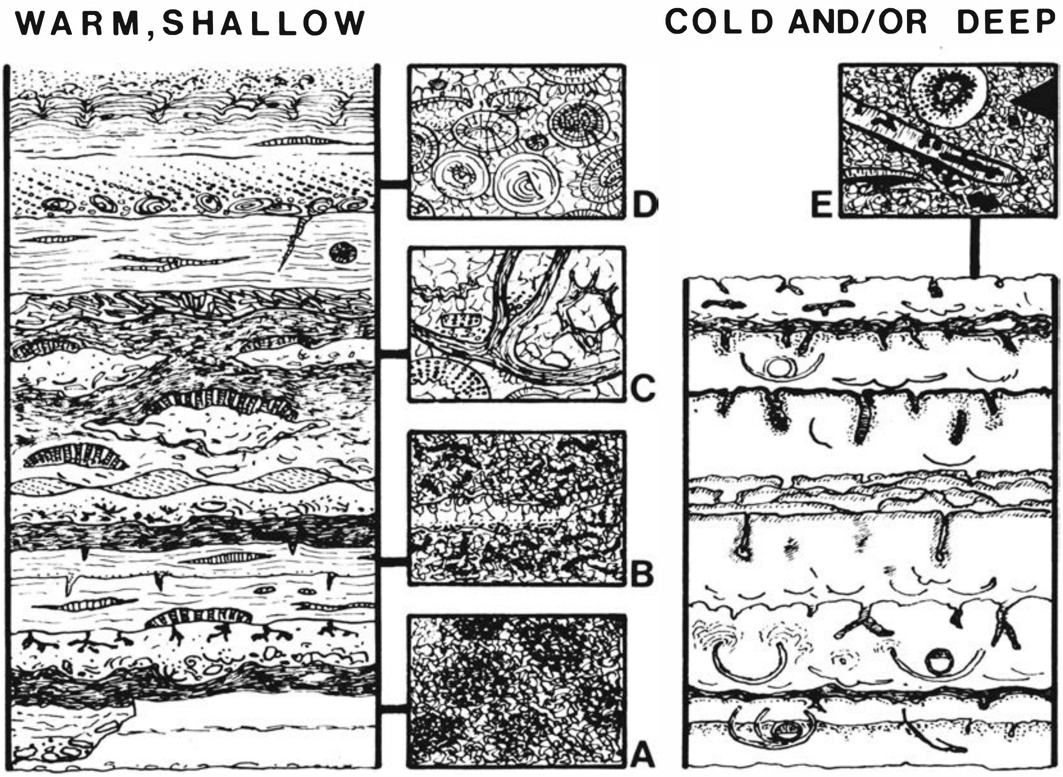

Fig. 1 – Comparison between features of carbonate deposits formed in warm, shallow seas and those formed in cold and/or deep seas. The warm section (left) may represent about 2 m thickness; the cold, deep section (right) corresponds to about 0.5 m. The left section shows, in ascending order, erosional channel with lag deposit; storm grading with load marks; bored hardground with attached skeletal epifauna; laminate with birdseyes, sheetcracks, and desiccitation cracks, ripple bedding, marly bed with concretions on which epifauna has grown, edgewise mud-chip conglomerate, non-ferrugineous onkoids and oolite, the latter current-bedded; algal stromatolites; sandstone with skeletal carbonate components. Boxes show corresponding thin sections of (A) micrite with peloids ("bahamite"), (B) micrite with "ghosts" of algal threads, and sheet-crack, (C) packstone with skeletal fragments of major benthic organisms such as brachiopods, echinoderms, and bryozoa, (D) calcareous oolite. The right section shows several hardgrounds and complexes of hardgrounds that are mineralized to different degrees and in different ways. They lack skeletal epifauna but have borings and burrows. Several beds are graded. Thin trilobite and cephalopod fragments may project above the hardgrounds. Evidence of mechanical reworking within the sedimentary environment is rare. Laminated crusts can occur sporadically and are partly of diagenetic origin. Thin section (E) shows calcilutite with trilobite fragment bored by sponges (?), echinoderm fragment with Fe mineralization at core, and pyrite crystal. Right section based mainly on Baltoscandian examples.

are missing. It is still not proven that the Lower and early Middle Ordovician limestones of Baltoscandia were formed in shallow water, and the lack of structures caused by waves, currents, and ground ice make this appear improbable, particularly in view of the assumed latitudinal position (60°S) where the transfer of physical energy should have been vigorous.

Fig. 1 shows the comparison between North American Ordovician carbonates formed in warm, shallow water with those from Baltoscandia (Lower–early Middle Ordovician) formed in water that probably was cool and also deeper than modern epicontinental seas. The warm, shallow water facies realm is characterized by pelletal mudstones, cyptalgal laminites, calcareous ooliths, onkoids, birdseye structures and sheet-cracks, desiccation cracks, erosion channels, edgewise conglomerates, current lamination, a greatly diversified megafauna, hardgrounds with conspicuous, sessile epifauna, and great lateral and vertical variation in fauna and facies. The cool water bedded limestone succession lacks these features and contains numerous discontinuity surfaces, often with phosphatic, glauconitic, or ferruginous crusts. A characteristic feature are trilobite fragments with minute boring patterns made possibly by sponges (this feature also occurs in similar limestones from the Ashgill of the Carnic Alps; samples provided by E. Serpagli). Several features indicate a very tranquil sea-bed environment (Lindström 1963, 1979) with a laterally and vertically stable facies and fauna.

Palaeotemperature

Palaeotemperature determinations for the Ordovician based on ^{18}O measurements have been discredited because it has been a longstanding assumption that the aberrant values obtained for much of the Palaeozoic were due to fresh water causing late-diagenetic equilibration (Hoefs 1980). However, recent results indicate that one must reckon with a lower ^{18}O-content of sea-water for Palaeozoic limestones than for younger limestones (Walls *et al.* 1979). It would be very difficult to interpret the numerous consistent data obtained from the Swedish Upper Cambrian (M. Dworatzek, pers. comm. 1982), the Lower to early Middle Ordovician (Friedrichsen & Lindström, in prep.) and uppermost Ordovician (Jux & Manze 1979), unless they record the original composition of these sequences. Upper Cambrian limestones yield about −8‰ ^{18}O, Lower Ordovician limestones −5.5 to −8‰ ^{18}O, and uppermost Ordovician limestones mostly −3 to −6‰ ^{18}O, all by PDB standard. The span of variation of mean annual sea-water temperature calculated from the variation of ^{18}O data for the early Ordovician (−5.5 to −8‰) is about 8°C. Since it appears improbable than mean annual sea-water temperature at 60°S was much above +8°C or below ±0°C, it is suggested that this was indeed the temperature range.

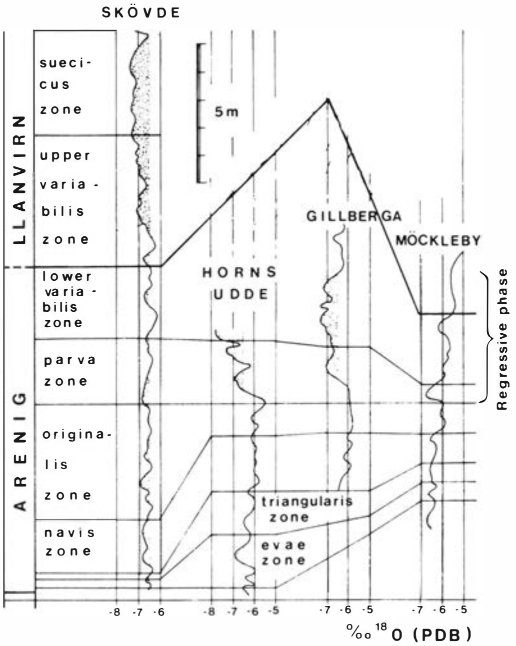

Fig. 2 – ^{18}O variation in four limestone sections in Sweden, with conodont zonation. Stippled zones have values lower than −6.5‰. The curves are based on means for each pair of successive samples. Data from Friedrichsen & Lindström (in prep.).

Fig. 2 shows the stratigraphic variation of ^{18}O in four Ordovician limestone sections in south Sweden. Granted that the curves reflect water temperatures, they indicate less cold water for Skövde than for the other sections during much of the time involved. The cause of this difference could be that Skövde faced towards the relatively warm Iapetus Ocean, whereas the other three sections faced towards the South Pole. However, during a late Arenig to early Llanvirn regressive phase (corresponding to the Whiterock regression of North America), the sedimentary environment of Skövde became somewhat cooler whereas that of the other sections became warmer. There might be a complex explanation for this reversal of temperature polarity. Faunal content and lithofacies suggest that the Skövde succession was deposited in deeper water than the other sections (though Möckleby might also have been relatively deep). To judge from the extent and amplitude of the regression (as suggested for instance by Lindström & Vortisch 1983) a major, continental glaciation was its most likely cause. The resulting reduction of deep-water temperature affected the Skövde section that remained relatively deep. The warming at Horns Udde and Gillberga could be a combined effect of shallowing and drift of Baltoscandia towards latitudes with warmer surface waters.

These interpretations bring us back to the all-important question of depth. If the whole of the Baltoscandian limestone succession was deposited at very shallow depth, then the interpretation must be drastically modified.

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Ordovician reefs and climate: a review

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Ordovician reefs have a relatively restricted distribution within carbonate depositional belts of the world. Most may be inferred to have formed in warm waters of the tropical-subtropical belts, like the associated bahamitic sediments. Early Ordovician reefs have a restricted distribution and mainly comprise undifferentiated algal- or sponge-dominated mounds of small size. The largest have a rigid organic framework in their upper parts and are true patch reefs. Middle Ordovician reefs have a much broader geographical distribution, a wider range of morphological differentiation (on-shelf patch reefs, shelf-edge reef complexes and on-shelf and down-slope carbonate mud mounds), and some attain barrier-reef dimensions. A much greater variety of frame-building organisms are represented including stromatoporoids, corals and bryozoans. Late Ordovician reefs are similarly widely distributed but not as large. They include restricted marine algal-dominated "pinnacle" reefs associated with evaporites, normal marine stromatoporoid-coral patch reefs and carbonate mud mounds. Of the two phases of warming suggested by reef occurrences in Baltoscandia, the first in the late Caradoc probably results from a short-lived maximum expansion of the tropical-subtropical belts with accompanying reduction in the latitudinal temperature gradient. The second in the late Ashgill coincides with the period of major glaciation. An associated increased latitudinal gradient is suggested by representatives of the cool (possible deeper) *Hirnantia* fauna occurring in relatively close proximity to reefs both in North Europe and North America. Reefs continued to grow at the height of the glaciation but apparently in much narrower tropical-subtropical belts.

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A wide variety of reef and reef-like structures have been reported from Ordovician carbonate successions of North America, Europe, Asia and Australia (Fig. 1). Best documented are those in North America and Baltoscandia. The range of morphologies includes carbonate mud mounds, patch reefs and reef complexes. These structures are mainly of small size and usually lack the sort of clear-cut differentiation into fore-reef, reef-core and back-reef facies seen in some Middle Palaeozoic, and most modern, reefs. However, the Ordovician reefs still show a remarkable degree of variation in form and organic composition.

A most dramatic evolutionary development of the reef ecosystem occurred during the Ordovician, with successional changes from algal-sponge dominated assemblages in the Early Ordovician to bryozoan dominated and stromatoporoid-coral dominated associations in the Middle Ordovician. Both these latter associations apparently co-existed for a time but then the stromatoporoid-coral assemblages seem to have taken over as the main frame-builders in the late Middle and Late Ordovician (Newell, 1971; Heckel, 1974; Copper, 1974).

Secondly, in the organically more complex patch reefs (or bioherms), Walker & Alberstadt (1975) have demonstrated the presence of vertical ecological changes from stabilization to colonization, then to diversification and finally to domination stages as the structure grew upward into the surf zone. James (1979) argues that carbonate mounds in the sense that they only exhibit stabilization and colonization stages may be viewed as "half-reefs", and presumably they did not grow into the zone of turbulence. He classified the reefs of the Early Ordovician as entirely carbonate mud mounds, while those of the Middle-Late Ordovician included

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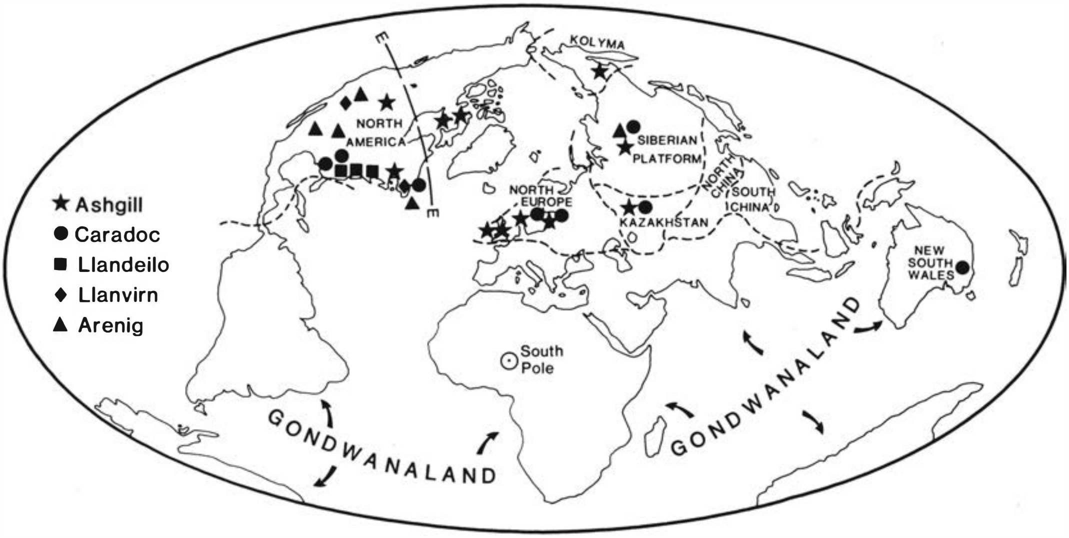

Fig. 1. Map showing worldwide distribution of Ordovician reefs. Note also probable position of Ordovician south pole in Africa and equator (E–E) for North America.

both mud mounds and patch reefs.

The term "reef" is here used in a broad sense to include all sorts of carbonate buildups with topographic relief, irrespective of their internal composition. This includes various types of carbonate mounds, ranging from the "half-reefs" of James (1979) to the stromatactis-bearing mounds of Jaanusson (1979a; 1982). Some lack a preserved "rigid organic framework" perhaps because they were formed by non-calcareous mud-trapping organisms such as some algae and sea grass, or their small areas of framework were destroyed by physical or biological processes prior to burial (Longman 1981). Others, like the stromatactis-bearing mounds, appear to lack any trace of organic control, and may have been formed by a process of early lithification converting loose sediments into hard rigid structures as in modern lithoherms (Neuman *et al.* 1977). But since only a few stromatactis-bearing Ordovician occurrences of this latter type, for example in the probable late Arenig of Nevada (Ross *et al.* 1975; 1982) and in the late Caradoc Kullsberg and Ashgill Boda mounds of Sweden (Jaanusson 1979 a; 1982), have so far been differentiated from carbonate mounds of more strictly organic origin, and they usually occur at times and in regions where organically constructed buildups are also well developed, there seems little justification for excluding them from this survey of Ordovician reefs.

For convenience I have used a tripartite division of the Ordovician including Tremadoc and Arenig in the early Ordovician, Llanvirn, Llandeilo and Caradoc (to the top of *Dicranograptus clingani* Zone) as Middle Ordovician, and the topmost Caradoc and the Ashgill as Late Ordovician.

Early Ordovician reefs

The world-wide distribution is confined to the North America and to Siberia (Figs. 1–2). In the western United States reefs are recorded as having formed in shallow, normal marine waters of gently shelving platforms to the north and south of the Transcontinental Arch (Ross 1976; 1977). To the north, the oldest in the House Limestone of western Utah (Rigby 1966) are of Tremadoc age. They are predominantly algal with sponges only a minor consistent. Others from Arenig horizons (Fillmore and Wahwah Formations) have sponges and the receptaculitid *Calathium* as dominant consistuents (Rigby 1966; 1971; Church 1974). There are also large stromatactis-bearing carbonate mud mounds of probable late Arenig age in Nevada (Ross 1972; Ross *et al.* 1975; 1982). The largest, though up to 76 m high and about 300 m across, have no organic frame.

South of the Transcontinental Arch, Toomey & Ham (1967), Toomey (1970) and Toomey & Nitecki (1979) have described late Canadian (early-mid Arenig) mounds in the McKelligton Canyon Formation of Western Texas and the Kindblade Formation of southern Oklahoma. They are typically small domical structures, but much larger mounds also occur, up to 20 m high and 87 m across. The largest has stabilization and colonization stages with cystoids, sponges (*Archaeoscyphia*), *Calathium*, brachiopods and trilobites; a diversification stage including digitate stromatolites, *Girvanella, Nuia, Archaeoscyphia, Calathium*, cystoids, the problematical stromatoporoid-like *Pulchrilamina* and various shelly fossils; and a domination stage with frame-building *Pulchrilamina*. Despite the predominant lime mud content (60–75% by volume) these larger mounds are strictly patch reefs since their upper parts include a rigid organic framework of *Pulchrilamina* allowing them to grow up into the surf zone.

Early Ordovician algal-sponge buildups have also been recorded from western Canada (Rigby 1965) and Newfoundland (Stevens & James 1976), and algal buildups from the Moiero River section, Siberian Platform (Miagkova *et al.* 1977). The Newfoundland structures have been documented recently by Pratt & James (1982) as algal (thrombolite)-dominated buildups, with associated corals (*Lichenaria*) in the earliest Ordovician (Tremadoc), and rare sponges and *Pul-*

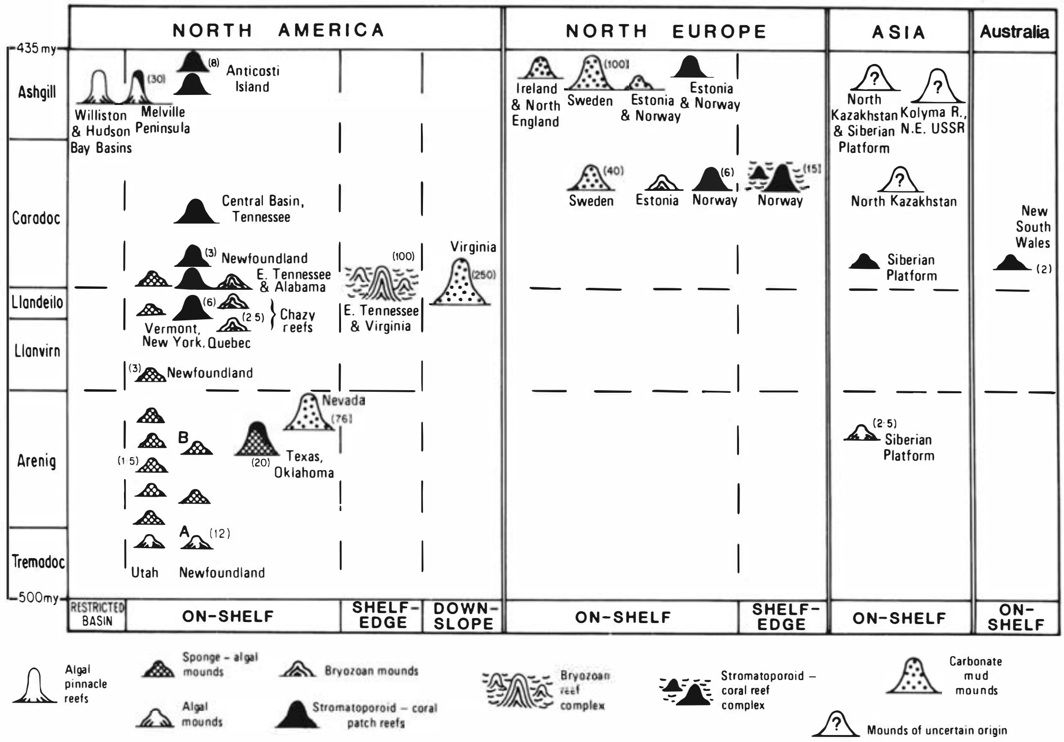

*Fig. 2 – Diagrammatic representation showing regional and tectono-environmental distribution of main types of reefs through Ordovician time. Note that the numbers in brackets refer to the maximum heights of reefs in metres. Also note different representation of Early Ordovician Texan and Oklahoman sponge reefs which have stromatoporoid-like upper parts, and Late Ordovician algal pinnacle reefs from Melville Peninsula which exhibit abundant corals in upper parts. In Newfoundland earliest corals (*Lichenaria*) appear in algal mounds (A), and earliest stromatoporoid-like organisms (*Pulchrilamina*) in sponge-algal mounds (B).*

chrilamina in the succeeding Arenig, occurrences (Fig. 2).

Middle Ordovician reefs

These have a much wider distribution in North America, Baltoscandia, Siberia, Kazakhstan and New South Wales. They also exhibit morphological differentiation into a number of types (on-shelf and down-slope carbonate mud mounds, on-shelf patch reefs and shelf-edge reef complexes).

The record in the pre-Chazyan (early Llanvirn) is limited to occurrences of small, 3 m high algal-sponge patch reefs in the Table Point Formation of western Newfoundland (Klappa *et al.* 1980, Klappa & James 1980).

In the Chazyan (late Llanvirn-Llandeilo), however, there was a dramatic appearance of more complex reef communities (Pitcher 1964; 1971; Newell 1971; Copper 1974; Heckel 1974). For the first time in the geological record a number of different groups of massive skeletal organisms began to contribute significantly as frame builders and binders in the growth of reefs. In the lower part of the Chazy Group (Day Point Formation) of New York and Vermont, the reefs are relatively small, apparently lacking any vertical zonation and dominated by bryozoans. Kapp (1975) has argued that these reefs formed in slightly deeper water (at wave base or below) than those of the succeeding Crown Point Formation. The Crown Point reefs are much larger, and some exhibit vertical growth stages – stabilization at the base with pelmatozoan debris, colonization by stromatoporoids and bryozoans, diversification by various groups including corals, stromatoporoids, bryozoans, algae and sponges, and domination at the top by frame-building stromatoporoids (Alberstadt *et al.* 1974). The reefs contain a lime mud matrix (up to 60% by volume) and those dominantly composed of stromatoporoids grew up into shallow waters above wave base. Bryozoan mounds have also been recorded from the Chazy Group of Quebec (Hofmann 1963; Kobluk 1981).

While the Chazy Group reefs are taken to represent on-shelf accumulations, the much larger, and slightly younger (latest Llandeilo – eraly Caradoc) Holston mass of east Tennessee (southern Appalachians) is regarded by Walker & Ferrigno (1973) as having formed in a shelf-edge location. It is composed of interconnected reef core and flank deposits, and in being up to 100 m high and at least 40 km long assumes barrier-reef proportions. The reef tract is overwhelmingly dominated by bryozoans, with up to 19 different, predominantly encrusting types in the core, and mainly ramose types in the flanks. The absence of algae has suggested to Walker (1977) that it formed at moderate depth, and recently this has been confirmed Ruppel & Walker 1982; Stock & Benson 1982). Thus a deeper water origin may help to explain the absence of stromatoporoids and account for the difficulties of Alberstadt *et al.* (1974) in defining the vertical growth stages of the Holston mass.

To the north, in Virginia, there are even larger carbonate masses (Read 1982), up to 250 m high and 60 km in length. Apparently they formed during a period of continuing rise in sea level (or subsidence). The main types of accumulation include shelf edge (shallow ramp) and down-slope buildups, both with associated core and flank deposits. Usually the core of the shelf-edge buildups exhibit more encrusting bryozoans, algae and sponges. The core of the down-slope buildups has more ramose bryozoans. Algae are present in the down-slope mounds suggesting that they grew up into the photic zone, although they are surrounded by deeper-water deposits. Stromatactis is common especially in the down-slope buildups.

Other small patch reefs are recorded from the Blackriveran (early Caradoc) Lourdes Limestone of western Newfoundland (Fåhraeus 1973; Bergström *et al.* 1974; Copeland & Bolton 1977; James 1979). The reefs are dominantly built by the coral *Labyrinthites.* In the Rocklandian (middle Caradoc) Carters Limestone stromatoporoid-coral patch reefs were developed in shallow on-shelf sites of the Central Basin, Tennessee (Walker & Alberstadt 1975). The reefs are most similar to Chazy reefs in having frame-building stromatoporoids and corals and exhibiting a fourfold vertical growth succession.

In Baltoscandia, Middle Ordovician reefs have been recorded from Norway, Sweden and Estonia. Harland (1981) has described relatively small on-shelf coral-stromatoporoid patch reefs and a larger shelf-edge reef complex in late

Caradoc successions of the Oslo region. The shelf-edge reef complex is a composite of a number of separate reefs, the largest being 15 m high and 50 m wide. This larger reef mass exhibits a crude lateral zonation with massive, close-packed domical stromatoporoids and talus on the "offshore" side more loosely packed laminar to domical stromatoporoids in the middle, and more diverse assemblages of corals, algae and ramose bryozoans on the "in-shore" side. Mounds with an organic fame also occur in the Vasalemma Limestone (Oandu stage) of north-west Estonia (Röömusoks 1980; Jaanusson 1979 a). Both Norwegian and Estonian buildups have associated bahamitic sediments (Jaanusson 1973).

More-or-less contemporaneous stromatactis-bearing carbonate mud mounds occur in the Kullsberg Limestone of Sweden (Thorslund & Jaanusson 1960; Jaanusson 1982). Reef cores are 40–50 m high and 300–350 m wide, and the shelly faunas and algae occur in nests or lens-shaped pockets. The mounds which lack an organic frame seem to have formed in on-shelf sites.

Nestor (1977) and Miagkova *et al.* (1977) noted occurrence of a small reef with stromatoporoids and algae in the Krivoluk Stage (early-middle Caradoc) of the Moiero River, Siberian Platform. Also much more extensive developments of reefs occur in the middle-late Caradoc of northern Kazakhstan (Vinogradov 1968; Nikitin 1973).

Australian occurrences are restricted to the small coral-stromatoporoid patch reefs of the Fossil Hill Limestone, central New South Wales (Webby & Packham 1982). The reefs, of probable early Caradoc age, are dominantly composed of the coral *Tetradium.*

Late Ordovician reefs

North American successions of Edenian and Maysvillian (latest Caradoc – early Ashgill) age seem to be devoid of reefal structures. Carbonates (Red River Formation and equivalents) are well developed in the western Mid-continent but no reefs have been reported.

In the succeeding Richmondian (middle Ashgill), however, there is a remarkable development of pinnacle reefs on the Melville Peninsula of Arctic Canada. Each reef rises up to 30 m above the surrounding interreefal deposits and is 0.8 km or more in diameter. Reef cores are described by B. V. Sanford (1977) as primarily composed of an "algal and stromatolitic limestone framework". The reefs are believed to have been linked by an open seaway to the north-east, explaining the presence of corals and other organisms. But to the south, on the margins of the Hudson Bay Basin, only reefs formed in more restricted conditions (B. V. Sanford 1977). Evaporites formed in the centre of the Hudson Bay Basin. Late Ordovician "algal buildups" shown by Ross (1976) at the margins of the Williston Basin in Montana and North Dakota are similarly flanked by basin-centre evaporites.

On Anticosti Island coral-stromatoporoid dominated patch reefs occur in late Richmondian-Gamachian horizons of the Vaureal and Ellis Bay Formations (Petryk, 1981 a). The Vaureal buildups in the upper part of Member 4 are not well documented. In the succeeding Ellis Bay Formation however one isolated reef occurs in Member 4 (Copeland & Bolton 1975; Bolton 1981), and a more significant development is recorded in what was originally the basal part of Bolton's (1972) Member 6, now Petryk's (1981 a) Member 7. The reefs of Member 7 are up to 8 m high and 100 m across and contain a variety of frame-builders and encrusters but especially corals and stromatoporoids (Copeland & Bolton 1975; Bolton 1981).

Reef development also occurred in the latest Ordovician of northern Europe. Small patch reefs dominated by corals and stromatoporoids occur in the 5b-limestones of the Oslo region (Hanken & Owen 1982), and in the Porkuni stage of Estonia (Jaanusson 1979 a). The sediments associated with these "organic reefs" are of bahamitic type (Jaanusson 1972). In the Swedish Boda Limestone there are much larger stromatactis-bearing carbonate mud mounds, 100–140 m high and up to 1 km across (Jaanusson 1982). They appear to lack any sort of organic frame. Similar reef-like bodies of limestone are represented in the Keisley Limestone of northern England (Ingham & Wright 1972), in the Chair of Kildare Limestone of Ireland (Wright 1968), the Pirgu stage of Estonia (Männil 1966)and the 5a-limestones of Norway (Brenchley & Newall 1980). Elongate, mound-like, massive limestone bodies with a fauna of bryozoans and pelmatozoans have also been reported by Hafenrichter (1980) from the early-middle Ashgill of Spain, but may represent erosional remnants of more continuously bedded sequences rather than true reefs.

In the Asian part of the Soviet Union, Late Ordovician "reef massifs" have been recorded from the Kolyma River, and small single reefs, from various localities in northern Kazakhstan and the Siberian Platform (Vinogradov, 1968).

Climatic significance

Spjeldnæs (1961) first recognized the marked temperature fluctuations of the Ordovician Period, and predicted polar ice caps prior to Late Ordovician glacial deposits being found in the Sahara (Beuf *et al.* 1971). He recognized major phases of cooling, the first in the early Llanvirn and the second in the early Ashgill (Fig. 3A). The polar ice cap in North Africa is now thought by Spjeldnæs (1981) to have existed continuously from Arenig to latest Ordovician time, expanding greatly in one or more episodes, even into low latitudes, at the end of the Ordovician. The effect of its continued existence was to markedly increase the climatic zonation (i.e., latitudinal temperature gradient) so that Ordovician climate (with the possible exception of the Tremadoc which was climatically similar to the Late Cambrian) was more like that of the present than that of the Cambrian or Silurian. With such a marked latitudinal gradient, the extremes of climate, i.e. at the pole and the equator, could be easily and reliably identified. Accordingly, the "bioherm/bahamite, warm water facies" could be regarded as a good starting point for regional analysis of the equatorial zone (Spjeldnæs 1976). However identification is not so easy because of the relatively limited number of preserved Ordovician reefs, and their tendency to exhibit a very wide range of morphological and evolutionary diversification, of apparently differing climatic significance.

Secondly there are no direct means of establishing the geographic limits of Ordovician reef growth, other than by comparing the distribution with palaeomagnetically-determined latitudes. The likely limits of coral-stromatoporoid reef growth relative to the pole within a single lithospheric plate can be obtained in younger (Middle Palaeozoic) successions. Middle Devonian reefs in north-west Africa (Dumestre & Illing 1967) are about 55° of latitude away from the position of the south pole, as determined palaeomagnetically and by the distribution of elements of the "cool water" Malvinokaffric Province (Oliver 1976; Heckel & Witzke 1979; Boucot & Gray 1979) in southern Africa. This suggests that these particular reefs flourished about 35° south of the equator. In their reconstructions of the Devonian world Heckel & Witzke (1976) have shown the reef and stromatoporoid distribution restricted to the warm climatic belt between 35° N and 40° S.

Of the various types of buildups stromatoporoid-coral reefs of the Middle-Late Ordovician may similarly have been useful indicators of deposition in warm climatic tropical-subtropical belts, especially where they are associated with bahamitic sediments. Counterparts in the Early Ordovician may have been the sponge-dominated reefs. Algal buildups seem also to suggest warm conditions of low latitudes particularly where associated with evaporites. Middle-Late Ordovician bryozoan reefal complexes appear to have had a wider climate range. The Holston and Day Point reefs of the Appalachians apparently grew in relatively deeper waters (Kapp 1975; Walker 1977; Ruppel & Walker 1982; Stock & Benson 1982), but in similar low latitudes to the stromatoporoid-coral buildups. In contrast the Spanish bryozoan mounds may have formed in higher latitudes but do not seem to be true reefs. There is the record of a near-polar bryozoan-dominated biostrome in the late Caradoc of the Anti-Atlas mountains in Morocco (Destombes 1971).

The climatic significance of on-shelf and down-slope carbonate and mounds is less certain. Apparently, the late Caradoc Kullsberg mounds of Sweden are not associated with "warm water" bahamitic sediments (Jaanusson 1973), and the late Ashgill mound complex in Ireland has intercalated mudstones towards its top containing the "cool water" *Hirnantia* fauna (Wright 1968). The carbonate mounds may have had a slightly wider latitudinal spread than the stromatoporoid (or stromatoporoid-like) dominated buildups, but they mainly occur at times and in regions where such reefs are also well developed (Fig. 2).

Apart from the buildups in Newfoundland (Pratt & James 1982) and Utah (Rigby 1966) there is very little evidence of reef development in the Tremadoc, even though algae, sponges and earliest corals (*Lichenaria*) were available to contribute to buildups. Possibly this reflects a period of relatively less favourable (possibly cooler) climatic conditions in the equatorial zone. However by Arenig times conditions in the equatorial zone had become far more favourable to reef growth with the rise of frame-builders like *Pulchrilamina* and the development of ecologically zoned patch reefs of comparatively large size. This would seem to be consistent with the suggestion by Spjeldnæs (1961) of a temperature rise through the Arenig (Fig. 3A).

The only reefs recorded during a period of

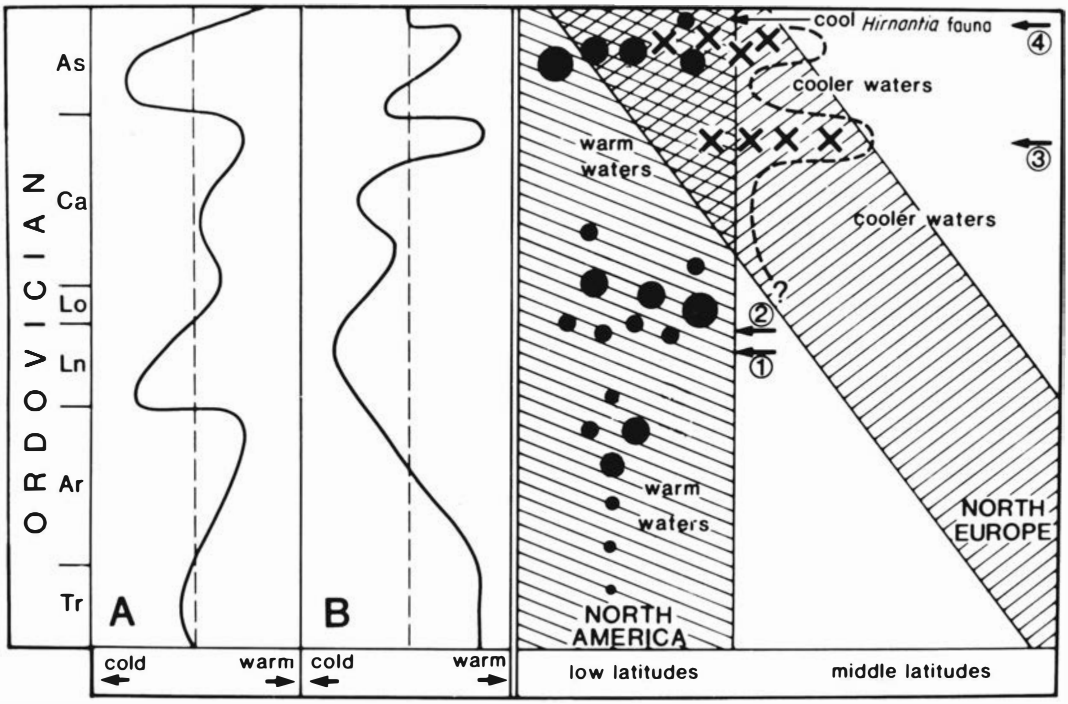

Fig. 3. Diagram showing Ordovician climatic curves for world (A) and North Europe (B) after Spjeldnæs (1961; 1978), the temporal distribution of reefs in North America (circles) and North Europe (crosses), inferred latitudinal positions of North American and North European plates with movement of North European plate into low latitudes (see Fig. 4), and the superimposed warming and cooling event, based in part on the occurrences of reefs. Circled numbers depict periods in the Ordovician when the following events occurred: 1–2, advent of main frame-bulding assemblages (1, bryozoans; 2, stromatoporoids and corals); 3, warming and reduced climatic zonation; and 4, intense polar glaciation and increased climatic zonation.

major early Llanvirn regression and cooling (Spjeldnæs 1961; see Fig. 3) are the small patch reefs in Newfoundland. These occurrences all seem to have formed in a position near the palaeomagnetically inferred equator (Ross 1976).

Significantly, the complex Middle Ordovician reef communities of bryozoans, corals and stromatoporoids made their appearances in the warming phase immediately following the major early Llanvirn cooling event (Spjeldnæs 1961). From this time onwards a wide variety of frame-building organisms particularly stromatoporoids and corals were available to contribute to the growth of reefs.

Some of the palaeomagnetically-based Ordovician world map reconstructions show northern Europe (including Baltoscandia) in low latitudes (Smith *et al.* 1973; Zonenshayn & Gorodnitzkiy 1977; Morel & Irving 1978; Kanasewich *et al.* 1978), while others place it in the middle-high latitudes (Noltimier & Bergström 1976; Bergström 1977; Ziegler *et al.* 1977; 1979; Scotese *et al.* 1979). The development of Baltoscandian late Caradoc reefs, with the accompanying bahamitic sedimentation and temporary invasion of new faunal elements in the Oslo region and Estonia is explained by Jaanusson (1973; 1979b) as due to a short-lived marked increase in temperature. If the North European plate had occupied a middle-high latitude position in the Middle Ordovician (see Llandeilo-earliest Caradoc reconstructions of Ziegler *et al.* 1977; 1979; and Scotese *et al.* 1979) then an abnormally high rise in temperature and concomittant expansion of the tropical-subtropical belts would have been required. The more likely alternative would involve the North European plate having already moved from high to middle-low latitudes by late Caradoc times (Figs. 3–4). This move also coincided with the narrowing of the Iapetus Ocean (Spjeldnæs 1978; Webby 1980). Northward movement of the North European plate towards the equator may be assumed to have been at a relatively uniform rate, with the temperate to warm (subtropical-tropical) to temperate climatic changes indicated by Jaanusson (1973) for the middle Caradoc-middle Ashgill interval superimposed on it (Fig. 3). The warming event, if it took place while the North European plate was still in middle-low latitudes, records a period of maximum expansion of the tropical–subtropical belts, and reduced climatic zonation (Fig. 3).

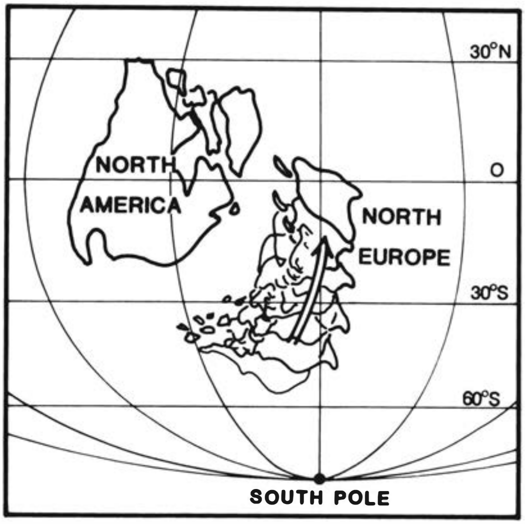

Fig. 4. Map reconstruction showing inferred northward drift of North European plate and closing of the Iapetus Ocean during Ordovician time.

The lack of reefs in contemporaneous North American successions is less easily explained, though it may be in part due to glacioeustatic rise in sea level which accompanied the warming. Certainly there was a period of widespread flooding and deposition of graptolitic shales in the eastern Midcontinent (J. T. Sanford 1978; Copper 1978).

By middle–late Ashgill time the North European plate have moved into low latitudes, and reef growth was again flourishing but this time while there was an increasingly intense continental glaciation developing in North Africa (Beuf *et al.* 1971; Spjeldnæs 1981). There was also contemporaneous reef formation in North America including major algal buildups on the flanks of evaporite basins in the continental interior. This seems to have been a period of heightened climatic zonation for while "warm-water" bahamitic sediments were associated with the Oslo and Estonian coral-stromatoporoid patch reefs and the large Swedish Boda carbonate mounds, these latter were forming contemporaneously with a slightly deeper water facies containing the "cool" *Hirnantia* fauna

(Sheehan 1979). The *Hirnantia* fauna is seemingly even more closely associated with the carbonate mounds in Ireland and northern England (Wright 1968; Ingham & Wright 1972).

Brenchley & Newall (1980) have argued that at the end of the Ordovician a dramatic lowering of sea level occurred and this seems to be related to the major expansion of the polar ice caps. They have interpreted both the Stage 5b patch reefs of the Oslo region and the Boda mounds as developing a karst topography after becoming emergent during the major regressional phase. There was substantial erosion of previously deposited facies in the Oslo region (Størmer 1967; Brenchley & Newall 1980).

No such erosional break is reported from the latest Ordovician–earliest Silurian succession on Anticosti Island. Petryk (1981 b) has interpreted glacioeustatic changes of sea level as responsible for the regressional/transgressional cycles of deposition in the upper Vaureal and Ellis Bay Formations, with the reefal horizons being inferred to have formed in the shallower, regressive phases. The extensive development of reefs in the Member 7 bioherms of the Ellis Bay Formation formed during the last and perhaps the most significant regressional phase marking the end of the Ordovician Period. The regressional phase and period of formation of the Member 7 reefs probably coincides with the maximum of glaciation and expansion of polar ice caps. By analogy with glacial phases of the Quaternary where there is evidence of "intensified oceanic circulation during times of expanded continental ice sheets", there may have been a higher carbonate productivity in the oceans of the equatorial zone (Bloom 1974). Although there may have been increased climatic zonation with a narrowing in width of the tropical–subtropical belts during the later Ordovician glacial phases, it is unlikely that there would have been any significant reduction in the degree to which reef growth flourished in well circulated shallow seas of the near equatorial zone. This narrowing of tropical–subtropical belts is supported by palaeomagnetic evidence of Anticosti Island lying within 20° of the late Ordovician equator (Petryk 1981 b), and the presence of a possible cool-water *Hirnantia* fauna in beds 6 m below the regressive deposits of Member 7 (Cocks & Copper 1981).

It may be concluded that during the Ordovician Period the tropical and subtropical belts alternately expanded and contracted (Newell 1971) with the world-wide oscillations of climate, as first recognized by Spjeldnæs (1961). At times of reduced climatic zonations, as for instance in the late Caradoc warming, reef building appears to have been flourishing in middle–low latitudes, possibly to at least 40° from the equator. In contrast at times of increased climatic zonation, as for example during the period of most intense glaciation at the end of the Ordovician, reef formation seems to have been limited to within a relatively few degrees, perhaps to within 20° of the equator.

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Distribution of the *Hirnantia* fauna and its meaning

By RONG JIA-YU

The geographic and stratigraphic distribution of the late Ashgill *Hirnantia* brachiopod fauna is here recorded and discussed based mainly on Chinese data. The *Hirnantia* fauna from the Kuanyinchiao Beds and their equivalents in southern China is approximately contemporaneous with the late Ordovician graptolite zones of *Diceratograptus mirus, Paraorthograptus uniformis* and *Diplograptus bohemicus.* The *Hirnantia* faunal assemblage became extinct before the *Glyptograptus persculptus* Zone, although the fauna is undoubtedly diachronous. Ecologic communities of the *Hirnantia* fauna are discussed. The present data reveal that the distribution of the *Hirnantia* fauna might not be restricted to one hemisphere and its development was not necessarily controlled by a cold water environment. It probably extended into the temperate and even subtropical zones during the latest Ordovician. The ecologic distribution of the main elements of the *Hirnantia* fauna were much broader than previously thought.

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The *Hirnantia* fauna is a geographically widespread and distinctive brachiopod fauna that occurs near the boundary between the Ordovician and Silurian systems. It seems to have appeared abundantly at the beginning of the Hirnantian (i.e. Late Ashgill), spread rapidly, and become extinct as a whole fauna beneath the *Glyptograptus persculptus* Zone. Temple (1965) introduced the term *Hirnantia* fauna for the brachiopod assemblage in the *Dalmanitina* Beds of the Holy Cross Mountains of S. Poland, the Ashgill Shales of N. England, and the Hirnant Beds of N. Wales. The common taxa in this fauna are *Hirnantia sagittifera* (M'Coy), *Dalmanella testudinaria* (Dalman), *Kinnella kielanae* (Temple), *Paromalomena polonica* (Temple), *Eostropheodonta hirnantensis* (M'Coy) and *Plectothyrella crassicosta* (Dalman) (= *P. platystrophoides* Temple). The unusually distinctive *Hirnantia* fauna is commonly associated with the trilobites *Dalmanitina (Mucronaspis), Brongniartella* and *Leonaspis* (Kielan 1960). Since 1965 the *Hirnantia* fauna has been recognized from the following areas: Bohemia (Marek & Havlíček 1967; Havlicek 1977), Västergötland, Sweden (Bergström 1968), Kildare, Ireland (Wright 1968), Keisley, England (Temple 1968), Maine, U.S.A. (Neuman 1968), Anti-Atlas, Morocco (Havlicek 1971; Destombes 1976), Carnic Alps, Austria (Schönlaub 1971; Jaeger *et al.* 1975), W. Libya (Havlicek & Massa 1973), Quebec, Canada (Lesperance 1974; Lesperance & Sheehan 1976, 1981), Southwest Wales (Cocks & Price 1975), Kazakhstan (Nikitin 1976), Southern China (Rong 1979), Kolyma (Oradovskaya & Sobolevskaya 1979), Anticosti Island, Canada (Cocks & Copper 1981), the Garth area, Wales (Williams & Wright 1981), and the Girvan District, Scotland (Lamont 1935; Harper 1981) (Fig. 1). The Burmese brachiopod assemblage described by Reed (1915), from the Northern Shan States, should also be considered as a representative of the *Hirnantia* fauna in southeastern Asia. Amsden (1971, 1974) compared the brachipod fauna from the Noix Limestone (lower part of the Edgewood Group), with the *Hirnantia* fauna and noted that these two faunas had some genera in common and were probably of the same age. However, in general these two faunas have a different composition. There are many genera in the older Edgewood fauna, such as *Stegerhynchus, Brevilamnulella, Dolerorthis, Eospirigerina, Leptoskelidion, Thebesia, Orthostrophella, "Homoeospira", Dictyonella* and others, which are un-

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known in the typical *Hirnantia* fauna, and conversely there are a number of genera in the *Hirnantia* fauna, for example, *Draborthis, Kinnella, Paromalomena, Leptaenopoma, Eostropheodonta, Plectothyrella, Hindella* and others, which are not present in the Edgewood fauna (Amsden 1974: 28). The older Edgewood fauna, therefore, seems significantly different from the *Hirnantia* fauna at both the generic and specific level.

Hirnantia fauna in China

Recent discovery of the widespread occurrence of the *Hirnantia* fauna in China (Rong 1979) has considerably expanded the geographical distribution of this fauna. During the last two decades, large collections of the *Hirnantia* fauna, usually associated with the trilobites *Dalmanitina, Platycoryphe* and *Leonaspis,* have been made from a great number of places in China (Fig. 2). In northern China no representatives of the fauna are as yet known.

The *Hirnantia* fauna is well developed in the upper Yangtze Basin, including north-eastern Yunnan, Sichuan, southern Shaanxi, northern Guizhou, and western Hubei. Lu (1959) named the strata bearing the fauna in these regions the Kuanyinchiao beds, which are intercalated between the Wufeng and Lungmachi Formations. They include shelly and graptolitic facies, and consist mainly of marls, mudstones, and argillaceous limestones, usually 0.2–1 m thick. Zhang & Sheng (1958) reported the following taxa from these beds in Tongzi, northern Guizhou: *Hebertella* aff. *occidentalis* (Hall), *Rafinesquina* cf. *alternata* (Emmonds), *Platystrophia lynx* (Eichwald) and *Orthis* sp. The specimens collected by Zhang & Sheng from these beds at the same localities have been reidentified by the author as follows: *Hirnantia sagittifera* (M'Coy), *Aphanomena* cf. *ultrix* Marek & Havliček, *Plectothyrella crassicosta* (Dalman) and *Dalmanella testudinaria* (Dalman), as well as

Fig. 1. Global distribution of the Hirnantia *fauna.*

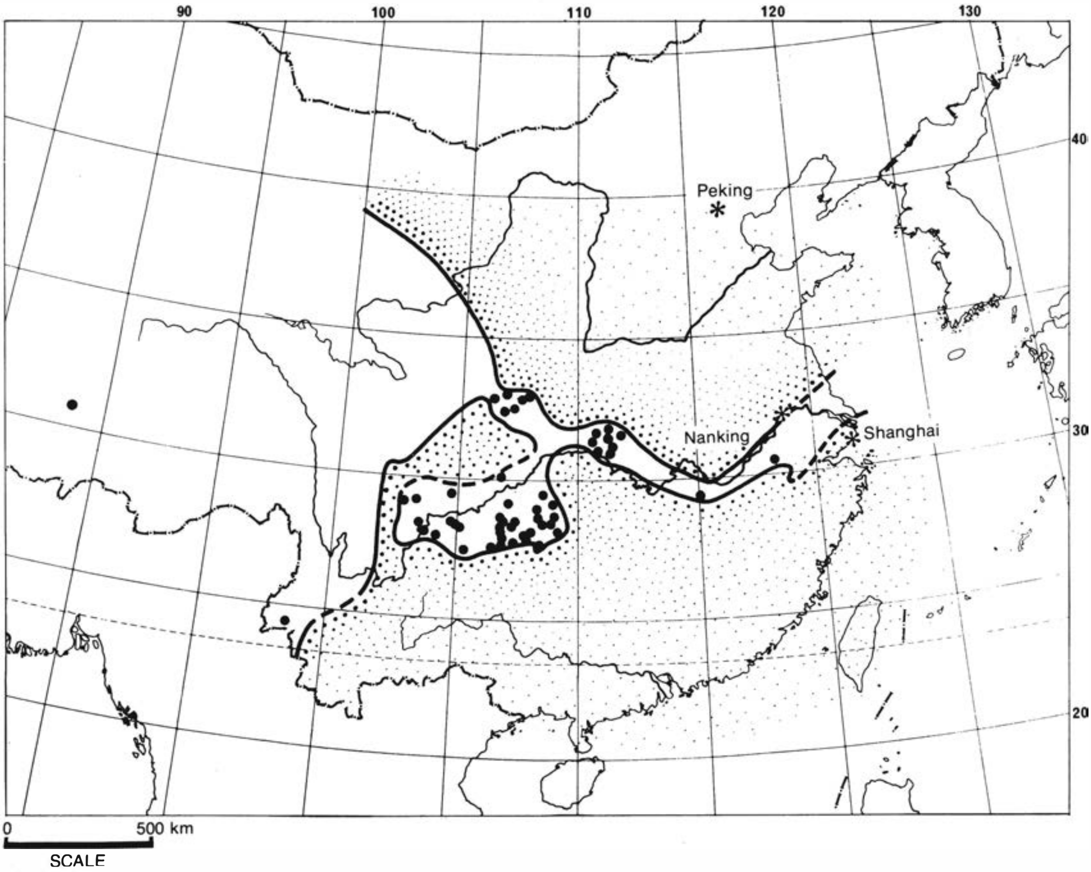


Fig. 2. Occurrences of the Hirnantia *fauna in China.*

Kinnella kielanae (Temple) and *Hindella crassa incipiens* (Williams). This is a typical (or strict) *Hirnantia* fauna which has been known from a great number of localities not only in northern Guizhou and southern Sichuan, but also in north-eastern Yunnan, northern Sichuan, southern Shaanxi and western Hubei (Rong 1979). A strongly diversified *Hirnantia* fauna has been found in Shuanghe, Changning in south-western Sichuan and Huanghuachang, Yichang in western Hubei by Mu and others during their field work in 1972 (Mu *et al.* 1978; Chang 1982). The Kuanyinchiao Beds yield the following taxa: *"Paracraniops"* sp. nov., *Philhedra* sp., *Philhedrella* sp., *Acanthocrania* sp. nov., *Comatopoma* sp., *Toxorthis* sp. nov., *Dalmanella testudinaria* (Dalman), *Horderleyella inexpectata* (Temple), *Trucizetina* sp., *Hirnantia sagittifera* (M'Coy), *Kinnella kielanae* (Temple), *Draborthis caelebs* Marek & Havlicek, *Cliftonia* cf. *oxoplecioides* Wright, *Aegiromena ultima* Marek & Havlicek, *Paromalomena polonica* (Temple), *Leptaenopoma trifidum* Marek & Havlicek, *Aphanomena ultrix* Marek & Havlicek, *Coolinia* sp., *Dorytreta* sp. nov., *Plectothyrella crassicosta* (Dalman), and *Hindella crassa incipiens* (Williams). The composition of this fauna is quite similar to those of Bohemia (Marek & Havlicek 1967), northern England (Temple 1968) and Västergötland (Bergström 1968). In the Upper Yangtze Basin, a highly diverse *Hirnantia* fauna has also been recorded from Qiaoting, Nanjiang in northern Sichuan, Wanjiawan and Tangya, Yichang in western Hubei and Sanlangpu, Xixiang in southern Shaanxi (Rong 1979).

In the Lower Yangtze Basin, in the provinces of Jiangxi and Anhui, the strata containing the *Hirnantia* fauna lie between the Kaochiapien and Wufeng formations. The fauna comprises only three brachipod taxa, viz. abundant *Paromalomena polonica* (Temple), fairly common *Aegiromena ultima* Marek & Havlicek, and very rare *Coolinia* ? sp. They are associated with *Dalmanitina* and *Platycoryphe*.

A *Hirnantia* fauna has most recently been found in western Yunnan and northern Xizang (northern Tibet). A good sample from the uppermost unit of the Ordovician rocks near the Mangjiu reservoir in Luxi, western Yunnan has been collected by Sun Dong-li and his colleagues during their field work in 1981. The unit yields a shelly fauna with *Hirnantia sagittifera* (M'Coy), *Cliftonia* sp., *Paromalomena polonica* (Temple), *Aphanomena* cf. *ultrix* Marek & Havlicek, *Coolinia* cf. *dalmani* Bergström, *Plectothyrella crassicosta* (Dalman) and *Hindella crassa incipiens* (Williams) associated with *Dalmanitina*. The first occurrence of the typical *Hirnantia* fauna in northern Xizang (northern Tibet) has been reported by Ni *et al.* (1981) from the upper part of the Xainza Formation in Xainza. It contains *Hirnantia, Kinnella, Cliftonia, Paromalomena, Aphanomena,* and *Hindella,* but the identifications are at present only at the generic level.

Stratigraphic distribution of the *Hirnantia* fauna

The time of both the appearance and extinction of the *Hirnantia* fauna in China is apparently not the same at all localities, although it is always within the Late Ashgill (= Hirnantian). Mu (1974) distinguished six graptolite zones in the Wufeng Formation, in ascending order, the zones of *Pleurograptus lui* (W_1), *Dicellograptus szechuanensis* (W_2), *Tangyagraptus typicus* (W_3), *Diceratograptus mirus* (W_4), *Paraorthograptus uniformis* (W_5), and *Diplograptus bohemicus* (W_6). The relationship of these zones to the base and top of the beds with the *Hirnantia* fauna is as follows (Fig. 3):

1) In Ganxi, Yanhe in north-eastern Guizhou, the Kuanyinchiao Beds with the *Hirnantia* fauna lie between the *D. bohemicus* (W_6) and *T. typicus* (W_3) zones (Mu *In* Lu & Mu 1980).
2) In Honghuayuan, Tongzi in northern Guizhou, the beds yielding the *Hirnantia* fauna occur beneath the *D. bohemicus* (W_6) Zone and above the *D. mirus* (W_4) Zone (Zhang *et al.* 1964).

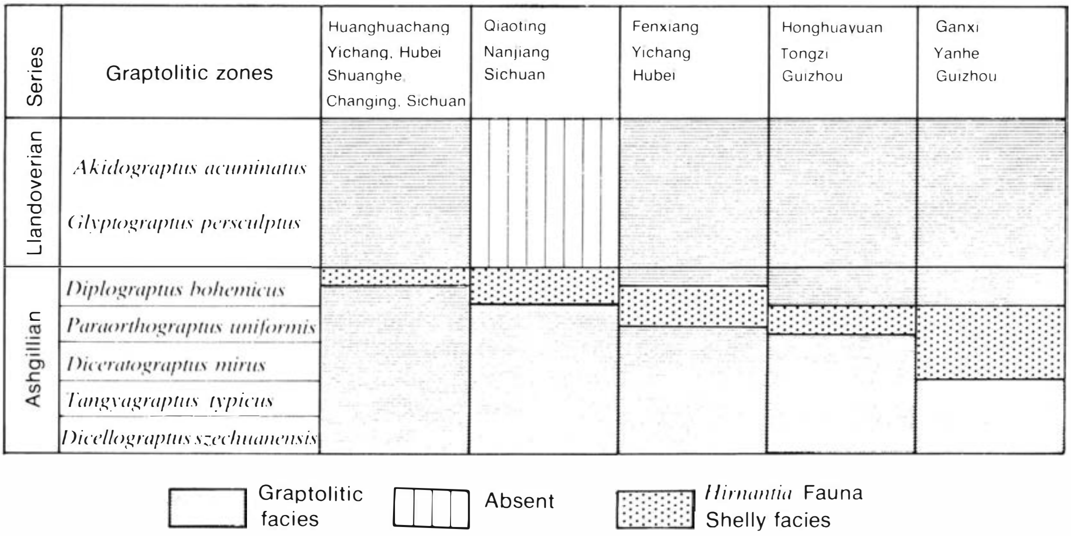
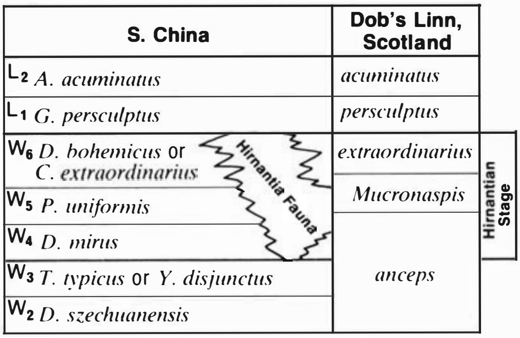


Fig. 3. Different levels of the Hirnantia *fauna in south-western China.*



S. China	Dob's Linn, Scotland
L_2 *A. acuminatus*	*acuminatus*
L_1 *G. persculptus*	*persculptus*
W_6 *D. bohemicus* or *C. extraordinarius*	*extraordinarius*
W_5 *P. uniformis*	*Mucronaspis*
W_4 *D. mirus*	*anceps*
W_3 *T. typicus* or *Y. disjunctus*	
W_2 *D. szechuanensis*	

Fig. 4. Correlation of the Hirnantia *fauna with the graptolitic zones across the Ordovician and Silurian boundary in southern China and Scotland.*

3) In Fenxiang, Yichang in western Hubei, the Kuanyinchiao Beds lie between the *D. bohemicus* (W_6) and *P. uniformis* (W_5) zones (Mu en-zhi, pers. comm.).
4) In both Huanghuachang, Yichang in western Hubei and Shuanghe, Changning in southwestern Sichuan, the beds containing the *Hirnantia-Dalmanitina* fauna occur between the *G. persculptus* (L_1) and *D. bohemicus* (W_6) zones (Mu *et al.* 1978).

Thus, the *Hirnantia* fauna in China is basically contemporaneous with the *D. mirus* (W_4), *P. uniformis* (W_5), and *D. bohemicus* (W_6) zones, although the *Hirnantia* fauna-bearing beds may also be within or above the *D. bohemicus* (W_6) Zone. It always occurs beneath the *Glyptograptus persculptus* (L_1) Zone which, in turn, is followed by the *"Akidograptus" acuminatus* (L_2) Zone.

It is also worth mentioning that in Shahe, Baosha (Western Yunnan) *Diplograptus bohemicus* (Marek) occurs in association with *Climacograptus extraordinarius* (Sobolevskaya) and *Diplograptus* cf. *orientalis* Mu *et al.*, in a grey black silty shale (Ni *et al.* 1982). This indicates that the *D. bohemicus* Zone may be correlated with the *C. extraordinarius* Zone. The latter zone has been reported from Scotland (Rickards 1979), Kazakhstan and Kolyma Basin (Koren' *et al.* 1979). Thus, the Ordovician and Silurian boundary could be placed at the level where the *Hirnantia* fauna disappears in the shelly facies and at the level between *G. persculptus* and *D. bohemicus* (or *C. extraordinarius*) zones in the graptolitic facies (Fig. 4).

Within the last decade or so, a mixture of Ordovician shelly fossils with so-called earliest Silurian graptolites have been reported from a few localities in various parts of the world (Rukavishnikova *et al.* 1968; Mikhajlova 1970; Nikitin 1972, 1976; Lespērance 1974; Jaeger *et al.* 1975; Koren' *et al.* 1979; Apollonov *et al.* 1980). Some workers have suggested that the Ordovician-Silurian boundary should be placed at the base of the *"Akidograptus" acuminatus* (L_2) Zone. However, *Glyptograptus ? persculptus* forma A or B (Koren' *et al. In* Apollonov *et al.* 1980), known in the association with the *Hirnantia-Dalmantina* fauna of the Chu-Ili Monuntains, Kazakhstan, is probably not *G. persculptus* (Salter) (Mu & Ni, in press). The *G. ? persculptus* Zone of Koren' *et al.* has been considered in China to correspond to the *D. bohemicus* (W_6) or *Cl. extraordinarius* Zone, rather than to the *G. persculptus* (L_1) Zone (Lin, pers. comm. 1981; Mu & Ni 1982). In the Portage River area, near Percē, Quebec, Lesperance (1974) reported the association of *Dalmanitina, Cryptolithus* and *Brongniartella* with the early Llandovery graptolite *Climacograptus rectangularis medius.* However, Rickards (*In* Cocks & Price 1975), on the other hand, considered the graptolite in question to be a Ordovician form and Ingham (1977) believes this shelly fauna to be older than the late Ashgill. It should also be noted that no *Hirnantia* fauna is present with these graptolites. In Feistritzgraben of the Karawank Alps, southern Austria, Jaeger *et al.* (1975) reported the occurrence of *G.* cf. *persculptus* in a black slate which has been correlated with the *Hirnantia* fauna-bearing beds in the Cellon section, Carnic Alps. However, the shelly fauna associated with *G.* cf. *persculptus* contains neither a *Dalmanitina* nor *Hirnantia* fauna. The identification of *G.* cf. *persculptus* is in doubt. The above occurrences of these taxa are insufficient to correlate the *G. persculptus* Zone with the *Hirnantia-Dalmanitina* fauna.

Ecologic communities within the *Hirnantia* fauna of northern Guizhou and southern Sichuan

When Cocks (1972; 627) discussed the ecology of the *Clarkeia* shelly fauna of S. America, he postulated that the low-diversity *Hirnan-*

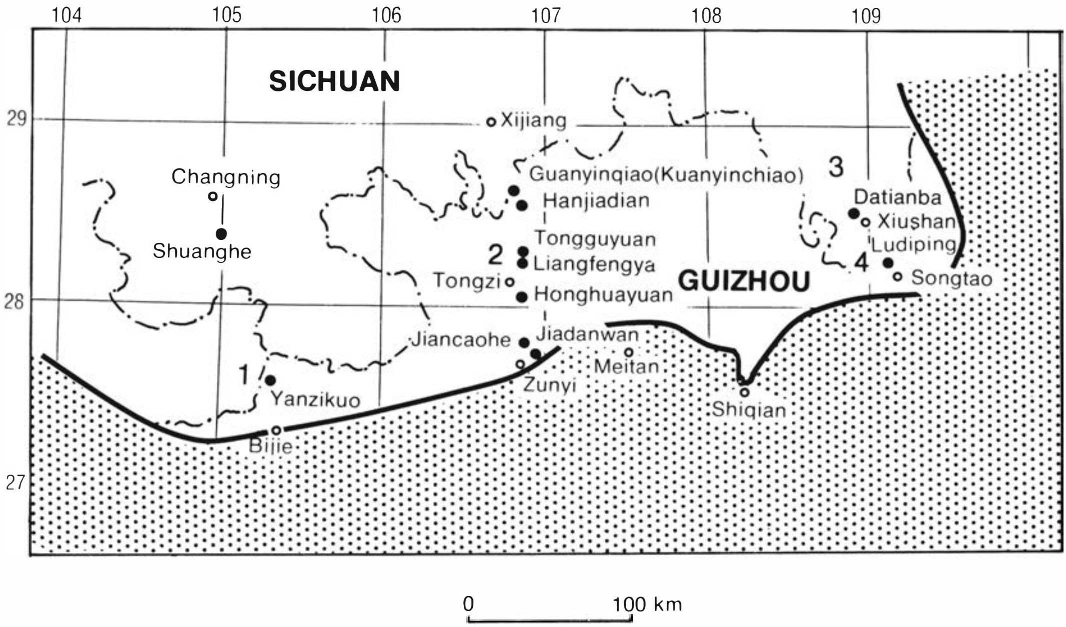


Fig. 5. Occurrences of the Hirnantia *fauna in southern Sichuan and northern Guizhou.*

tia fauna represents a community filling an ecologic niche comparable to the *Eocoelia* and *Crytothyrella* communities (i.e. Benthic Assemblage 2 of Boucot 1975) of the early Silurian. Lespérance & Sheehan (1976: 720) considered that the term *Hirnantia* Community should be used instead of the "*Hirnantia* fauna". The *Hirnantia* Community has been assigned to Benthic Assemblage 4 (possibly 5) by Sheehan (1979). The available data in southern China indicate, however, that the *Hirnantia* fauna may comprise more than one ecological community.

During the latest Ordovician, the southern margin of the Upper Yangtze Basin was located approximately in the Bijie-Zynyi-Songtao region of northern Guizhou. There are seven localities yielding the *Hirnantia* fauna from Zunyi to Xijiang (Fig. 5). The sandy shale Kuanyinchiao Beds at Jiadanwan, Donggongsi, Zunyi immediately north of the Mid-Guizhou Late Ordovician Land area, have yielded a rich brachiopod assemblage along with the trilobite *Dalmanitina.* The assemblage is characterized by the abundant occurrence of "*Whitfieldella*" n.sp., comprising 97% of the whole fauna, associated with a few specimens of *Hindella crassa incipiens* (Williams) and *Coolinia* sp. This is not a typical *Hirnantia* fauna, although *H. crassa incipiens* is one of the most common taxa in the *Hirnantia* fauna elsewhere. The assemblage may have lived in a very shallow water environment near the shore and probably is a representative of Benthic Assemblage 2. Northwards, the conditions are quite different. From Jiancaohe, Zunyi in northern Guizhou to Guanyinqiao, Xijiang in southern Sichuan, no "*Whitfieldella*" has been found from the mudstone or argillaceous limestone of the Kuanyinchiao Beds and *Hirnantia sagittifera, Dalmanella testudinaria, Kinnella kielanae, Paromalomena polonica, Aphanomena ultrix, Cliftonia psittacina, Plectothyrella crassicosta* as well as *H. crassa incipiens* are all common elements of the typical *Hirnantia* fauna (Fig. 6). It seems possible that these *Hirnantia*-bearing faunas may occupy a position equivalent to high Benthic Assemblage 3, although the relative frequencies of the genera vary. It is suggested that the sea floor bottom of the Upper Yangtze Basin in N. Guizhou appears to have been relatively even, and the depth of the basin there did not change appreciably.

It is also interesting to compare the composition of the brachiopod fauna along the line (see Fig. 5) from locality 1 (Yanzikou, Bijie in north-western Guizhou), through 2 (Liangfengya, Tongzi in northern Guizhou) to 3 (Datianba, Xiushan in south-eastern Sichuan) and 4 (Ludiping, Songtao in north-eastern Guizhou). The line runs approximately parallel to the northern coast of the Mid-Guizhou Late Ordovician Land. The brachiopod assemblages from the Kuanyinchiao Beds at these four localities represent a typical *Hirnantia* fauna and contain about 5–8 taxa (Fig. 7). It is *Dalmanella testudinaria* (Dalman) which is the most abundant taxon comprising about 66.7% (1), 34.3% (2), 38.95% (3), 48.7% (4) in the respective assemblages. The overall similarities in both diversity and composition between these assemblages might indicate that they have belonged to a single palaeoecologic community. All of them probably inhabited a normal shallow water environment and may occupy a high Benthic Assemblage 3 position. With respect to the dominance of *Dalmanella*, the *Hirnantia*

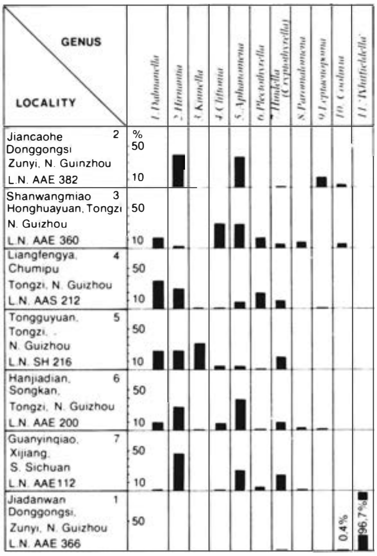


Fig. 6. Percentage of elements of the Hirnantia *fauna in different localities from Jiadanwan to Guanyinqiao in northern Guizhou and southern Sichuan.*

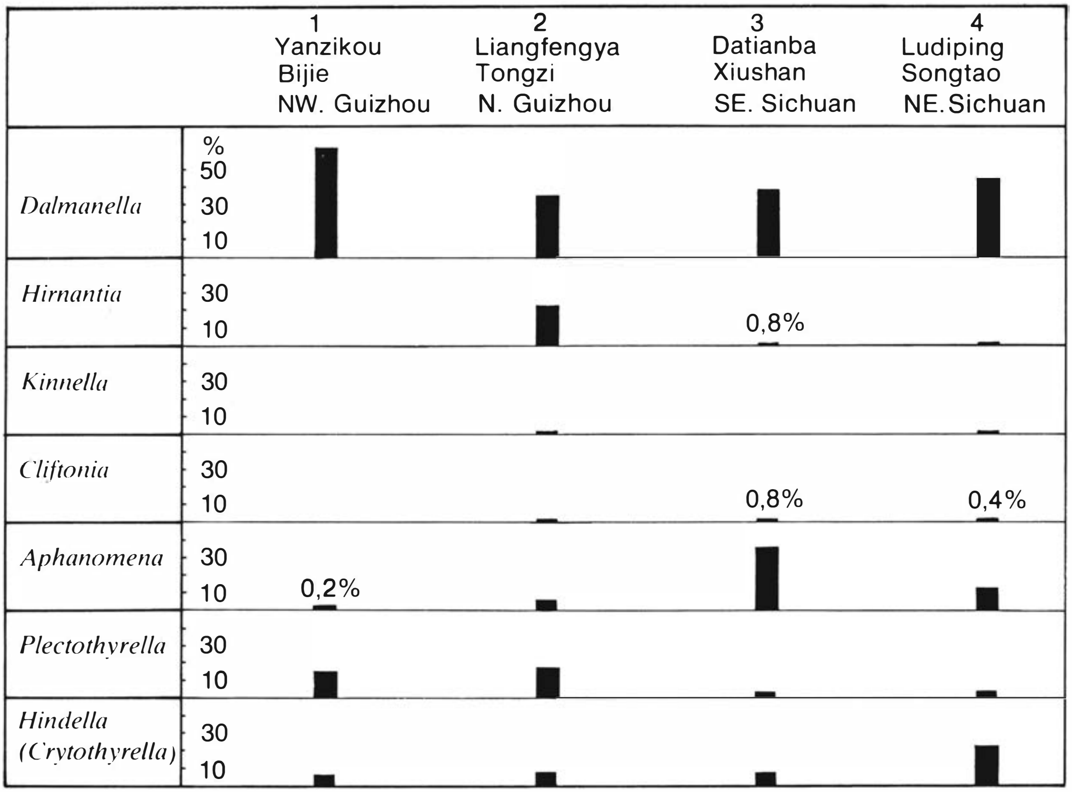


Fig. 7. Percentage of the components of the Hirnantia *fauna in Yanzikou, Liangfengya, Datianba and Ludiping.*

fauna mentioned above is comparable to the Caradoc *Dalmanella* Community of the southern Berwyns in North Wales (Pickerill & Brenchley 1979). The latter has been also assigned to Benthic Assemblage 3.

In a northerly direction towards Shuanghe, at the locality Changning in south-western Sichuan, situated in the central part of the basin (Fig. 5), there occurs a strongly diversified *Hirnantia* fauna which consists of 21 taxa associated with *Dalmanitina, Platycoryphe* and *Leonaspis.* Here the *Hirnantia* fauna includes not only the common representatives of the typical *Hirnantia* fauna (such as *Dalmanella, Hirnantia, Kinnella, Cliftonia, Paromalomena, Aphanomena, Plectothyrella* and *Hindella*), but also such genera as *Toxorthis, Comatopoma, Trucizetina, Horderleyella, Draborthis, Triplesia, Dorytreta, Sphenotreta, Philhedra, Philhedrella, Paracraniops,* and *Acanthocrania* which are very rare or absent in the typical *Hirnantia* fauna. Within this fauna there are four very abundant genera: *Hirnantia, Kinnella, Aphanomena* and *Paromalomena.* However, *Dalmanella, Plectothyrella* and *Hindella* are rare, usually making up only 1–4% of the whole fauna, or absent.

It appears that this highly diversified *Hirnantia* fauna may have lived in a deeper water environment than that of the strict *Hirnantia* fauna and may occupy a position about that of low Benthic Assemblage 3.

Evidently, the strict or typical low diversity *Hirnantia* fauna lived in more restricted conditions (compared to the highly diversified *Hirnantia* fauna) but also reached a wider distribution. It is possible that the Late Ordovician glaciation lowered sea levels (Sheehan 1973) and decreased broad shallow water environments during that time.

Climatic significance of the *Hirnantia* fauna

The occurrence of continental glaciation centered in North Africa during the Late Ordovician has been documented by Beuf *et al.* (1971), Bennacef *et al.* (1971), Fairbridge (1971), Berry & Boucot (1973) and others. Evidence of periglacial deposits has been obtained not only from Morocco, Mauritania, Mali, Chad, Ethiopia, Niger and Sierra-Leone, but also from South Africa, the Amazon Basin of Brazil, Argentina, Peru, Bolivia and southern Spain. Possible later Ordovician glacio-marine deposits also occur in the Armorican Massif (Normandy) of France (Dore & Legall 1972). Based on the discovery of a *Hirnantia* fauna in the Upper 2nd Bani Formation, a detrital deposit of glacial origin, in the central Anti-Atlas of Morocco (Havlicek 1971; Destombes 1971), it has been concluded that the maximum glaciation was reached at the end of the late Ashgill (Destombes 1976). Havlíček (1976) pointed out that the distinctive features of the *Hirnantia* fauna in the Kosov Beds were presumably a reflection of the very cool climate, whose influence can be traced all over the Mediterranean Province. Sheehan (1975) and Sheehan & Lespérance (1978, 1979) postulated that the development of glaciers was accompanied by the expansion of a cold water fauna which originated in the high latitudinal Mediterranean Province. Thus they considered that the *Hirnantia* fauna was composed largely of Mediterranean Province genera and was part of a relatively cold water province. As Cocks & Price (1975) suggested, however, the extent to which the *Hirnantia* fauna is a direct reflection of cold water condition is as yet uncertain.

Recently, the distribution of the *Hirnantia* fauna has been shown to be much wider than previously assumed. In addition to Morocco and Libya, where both the *Hirnantia* fauna and Late Ordovician glacial deposits occur, this brachiopod fauna is known not only in Europe and North America, but also in vast regions of Asia including most of southern China (from Xizang (Tibet) on the west to Anhui on the east), Burma, Kazakhstan and Kolyma. In these regions no late Ordovician deposits of glacial origin have been encountered so far. It is evident that most of these regions were far from the occurrences of the late Ashgill continental glaciation. On palaeomagnetic evidence the Ordovician south pole was in northwest Africa (McElhinny & Luck 1970; Whittington & Hughes 1972). Regions with the *Hirnantia* fauna in Asia, northern Europe and North America were situated much closer to the palaeoequator than to the pole during that time. Southern China, Kazakhstan and Kolyma may have been within the temperate zone, or even subtropical zone, according to palaeomagnetic data or other evidence (Morel & Irving 1978; Scotese *et al.* 1979; Seslavinsky 1979). It is also clear that the distribution of the *Hirnantia* fauna might not be restricted to one hemisphere. The fauna was apparently distributed far beyond the area of influence of the late Ordovician glaciation, into temperate and even subtropic climatic belts.

A much more diverse *Hirnantia* fauna than the typical one has been observed in Bohemia, Västergötland, Yunnan, Sichuan and Hubei. Here it is generally composed of 13 to 20 brachiopod taxa with a diversity greater than one would expect if it were a cold water fauna of the type described from the Malvinokaffric Realm in the Silurian and Devonian (Boucot *et al.* 1969; Boucot 1975). As yet no highly diversified *Hirnantia* fauna has been found in the cold water regions of North Africa and neighbouring areas.

A typical *Hirnantia* fauna comprising *Dalmanella testudinaria, Aphanomena* sp., *Coolinia* sp., *Plectothyrella crassicosta, Dorytreta* n. sp., and *Hindella crassa incipiens,* associated with *Dalmanitina,* has been found to occur in a 1.5 m thick argillaceous limestone in Yanzikou, Bijie in north-western Guizhou, south-western China (Rong 1979). It is associated with many solitary corals, such as *Brachyelasma, Streptelasma, Grewingkia, Borealasma, Crassilasma, Siphonoplasma, Pycnactics, Paramplexoides, Singkiangolasma, Lambeophyllum,* and *Kenophyllum* (Ho 1980). Rugose corals, in part the same genera, also occur in the Borenshult fauna in Östergötland, Sweden, where there are also some elements of the *Hirnantia* fauna. A similar rugose coral fauna with some associated tabulates, such as *Palaeofavosites, Schedohalysites* and *Propora,* is known from an argillaceous limestone of the Kuanyinchiao beds in Leijiatun, Shiqian in north-eastern Guizhou (Yang Sheng-wu, pers. comm. 1982). The abundance of corals of southern China seems to cast doubt on the view that this *Hirnantia* fauna lived exclusively in a cold water regime.

Of special interest is the occurrence of the *Hirnantia* fauna in bioclastic, argillaceous limestones and bituminous limestone in many localities in south-western China (Rong 1979). The intimate association of the *Hirnantia* fauna with the upper part of the Chair of Kildare reef limestone at Kildare, Ireland and with oolitic limestones in the Hirnant beds at Aber Hirnant, North Wales is also known (Wright 1968). The presence of oolite indicates that the limestone was deposited in shallow water at an estimated water temperature of at least 25 to 27°C (Jaanusson 1973). Moreover, Bahaman-type sediments (Beales 1958; Bathurst 1971; Jaanusson 1973), such as peloids and ooids, occur in the *Dalmantina* Beds in Västergötland (Stridsberg 1980) together with the highly diversified *Hirnantia* fauna (Bergström 1968). Modern bahamitic sediments have been described by Newell *et al.* (1960) and Purdy (1963). Therefore, the author considers it more likely that the *Hirnantia* fauna did not necessarily live only in cold water. It probably extended into the temperate and even subtropical zones during the latest Ordovician, although these zones would have been influenced indirectly by the glacial events, accompanied by a significant eustatic lowering of sea-level (Sheehan 1973) and a cooling of the oceans (Jaanusson 1979).

The above discussion seems to suggest that the ecologic tolerances of the main elements of the *Hirnantia* fauna, such as *Hirnantia, Kinnella, Paromalomena, Eostropheodonta* and *Plectothyrella,* were much broader than previously implied. Other members of the fauna, such as *Dalmanella, Aphanomena, Cliftonia, Leptaena* and *Leptaenopoma* may have expanded to the regions near the equator associated with *Dolerorthis, Brevilamnulella, Stegerhynchus, Eospirigerina* and *Dictyonella,* which are almost absent in the typical *Hirnantia* fauna. Interestingly, an approximately contemporaneous *Holorhynchus* brachiopod fauna is located immediately to the north of the known limits of the *Hirnantia* fauna in Eurasia. It is possible that the *Holorhynchus* fauna lived in a warmer water environment than the *Hirnantia* fauna, although in central Sweden, the two are mixed (Jaanusson, per. comm. 1978).

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The environmental distribution of associations belonging to the *Hirnantia* fauna – Evidence from North Wales and Norway

By PATRICK J. BRENCHLEY and BRIAN CULLEN

The differentation of the *Hirnantia* fauna into two or more associations has been possible in the Hirnantian of Wales and Norway. The *Hirnantia* fauna *s.l.* is shown to have occupied a rather wide bathymetric range on late Ordovician shelves. In North Wales, Rawtheyan shelly faunas are used as evidence indicating a palaeoslope dipping gently to the west in the vicinity of the Berwyn Hills; brachiopod dominated, plectambonitacean/dalmanellid faunas pass westward into sparse trilobite-dominated faunas. Slumped deposits indicate the shelf/basin transition was south-west of Bala. The *Hirnantia* fauna in the nearer shore environments of the east Berwyns is dominated by *Hindella* while the more offshore environments to the west have *Eostropheodonta* and *Hirnantia.* Palaeogeographical and sedimentological evidence from the Oslo–Asker district of Norway confirms the inner shelf position of *Hindella* dominated assemblages and the deeper shelf position of the *Hirnantia* association. A third, intermediate, *Dalmanella* association has also been recognised. The faunal associations which together comprise the *Hirnantia* fauna *s.l.* appear similar in nature to brachiopod-dominated associations of communities recognised at other horizons in the Lower Palaeozoic.

Nearly all the assemblages recorded in North Wales can be assigned to the *Hirnantia* fauna, but in the formerly more tropical area of Norway there were, in addition to the *Hirnantia* fauna, two offshore associations, the *Onniella* and *Holorhynchus* associations and two nearshore associations in carbonate facies, the *Brevilamnulella* and *Thebesia* associations. Sandy shoreface facies characteristically contain the trace fossil *Monocraterion.*

The distinctive features of the *Hirnantia* fauna are discussed, and it is concluded that the fauna was probably a cold water fauna.

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The *Hirnantia* fauna is a brachiopod-dominated fauna which is characteristically found in the uppermost part of the Ordovician (i.e. the Hirnantian Stage). The fauna has attracted special attention and a special designation because elements of the fauna are unusually widespread. Even when the continents are reassembled into a probable late Ordovician configuration, the *Hirnantia* fauna is still found at sites which were widely separated and ranged from circumpolar to near tropical positions, though it is more common in the former (Jaanusson 1979).

The core of the *Hirnantia* fauna is composed of several eurytopic genera, such as the brachiopods *Hirnantia, Eostropheodonta, Kinnella, Plectothyrella,* and the trilobites *Mucronaspis* and *Brongniartella,* which also have a wide geographical distribution. At any particular locality there may be, in addition, several endemic taxa (Lespérance 1974). The differing abundance of genera and species at different locations has been noted by Temple (1965), Bergström (1968), Wright (1968), Lespérance (1974), Cocks & Price (1975) and Harper (1981). A subdivision of the *Hirnantia* fauna into distinct associations has been made in China by Rong (1979), in Scotland by Harper (1981), and in Scandinavia by Brenchley & Cocks (1982). In some regions there are, in addition to the *Hirnantia* fauna, other distinct assemblages. For example, an association characterised by *Holorhynchus* is found at localities as separates as

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China (Rong 1979) and Scandinavia (Jaanusson 1979, Brenchley & Cocks 1982), while other associations confined mainly to carbonate facies are recorded by Amsden (1974) from the U.S.A. and Brenchley & Cocks (1982) from Scandinavia.

Although the variability of the *Hirnantia* fauna has been well described, there is still little information about how this variability is related to environmental parameters, or indeed about the environmental range of the fauna as a whole. Wright (1968) noted the presence of the fauna, in association with oolites at Bala, North Wales, and in association with reef limestones at Kildare, in Ireland, and concluded that the fauna lived in a shallow water environment. Further, the restricted diversity of some assemblages strongly suggested a lagoonal type of environment (Wright 1968). Reinvestigation of Hirnantian sediments in North Wales suggest that the oolites near Bala lie in channels and were probably allochthonous, while the late Ordovician bioherms, such as those at Kildare in Ireland, Keisley in northern England and Boda in Sweden are carbonate mud mounds and not organic framework reefs. The evidence that the *Hirnantia* fauna lived in a shallow water environment is therefore questionable. An alternative view is that of Sheehan (1979) who has suggested that the *Hirnantia* fauna in Västergötland, Sweden, can be assigned to benthic assemblage 4 (cf. Boucot 1975), implying a moderately deep shelf environment for the fauna.

The purpose of this paper is to assess the environmental range of the *Hirnantia* fauna in the Upper Ordovician of North Wales and southern Norway, and to show that the fauna in both areas can be sub-divided into associations which were depth related and similar in nature to palaeocommunities described from other levels in the Lower Palaeozoic. The distinctive features of the *Hirnantia* fauna are then discussed.

Faunal Associations in North Wales

The *Hirnantia* fauna is found at more than ten localities in the Hirnantian rocks of North Wales (Fig. 1) and faunal lists have been published for Bala (Bassett *et al.* 1966) and Glyn Ceiriog (Hiller 1981). The main elements of the fauna found at selected localities are shown in Table 1. Study of this shows that *Hirnantia, Eostropheodonta* and *Dalmanella* occur in most localities but *Eostropheodonta, Hirnantia, Bancroftina* and *Plectothyrella* are particularly common in the western localities, around Bala, whereas *Hindella* is the dominant brachiopod at the eastern localities, such as Meifod and Llanfyllin, where *Plectothyrella* is absent. We will refer to these western and eastern associations as the *Hirnantia* associations and *Hindella* associations respectively.

The *Hindella* association has, in addition to the core genera *Hindella, Dalmanella, Hirnantia* and *Eostropheodonta,* a large number of corals, bryozoa and molluscs, together with a relatively large number of other brachiopod genera. This association is therefore considerably more diverse than the *Hirnantia* association.

In the most easterly area, near Glyn Feiriog, faunal assemblages are found which are similar to the *Hindella* association in having large numbers of solitary corals and bryozoans although they lack *Hindella* itself. Instead, *Hesperorthis* and *Nicolella* are particularly common, and since these rather variable and undoubtedly nearshore assemblages cannot be assigned to the *Hindella* association they are termed orthid-coral assemblages.

Assemblages in the west Berwyns are geographically intermediate between the eastern and western regions and have some similarities with both the *Hindella* and *Hirnantia* associations. Although *Hindella* is common it does not dominate the fauna to the same extent as at Meifod or Llanfyllin, and the number of genera making up the brachiopod fauna increases to a maximum of 12 with poorly preserved ?*Platymena* being the dominant member of the fauna.

The *Hirnantia* association as seen in the type area for the Hirnantian Stage, Bala, consists almost wholly of the brachiopods *Hirnantia, Eostropheodonta, Bancroftina* and *Plectothyrella* (Temple 1965; Wright 1968; Cocks & Price 1975), although bivalves also occur in conjunction with the *Hirnantia* fauna at a few localities.

One of the most notable features of the *Hirnantia* association is the almost complete absence of corals and bryozoans which are so common in the more easterly faunas. The association has a low diversity even though the

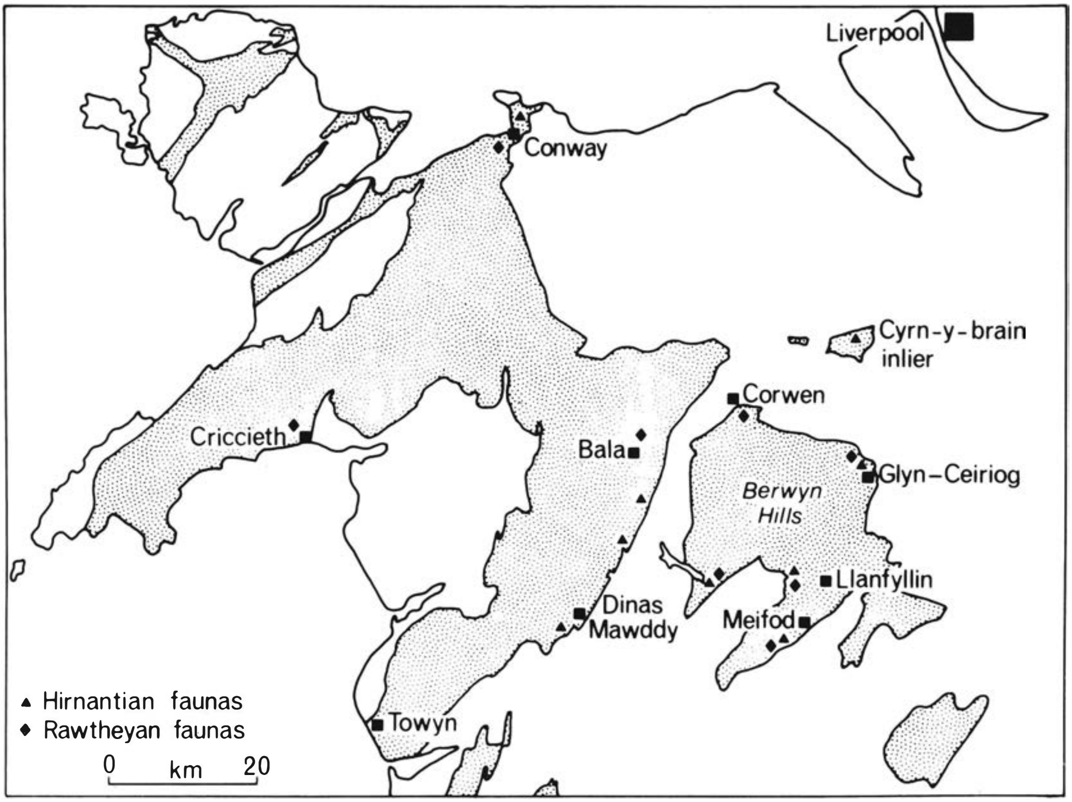

Fig. 1. Location of selected localities for Rawtheyan and Hirnantian faunas in North Wales. Outcrop of Ordovician rocks shaded.

number of specimens found at Bala is large.

Beyond the areas where the brachiopod-dominated *Hirnantia* fauna occurs there are sections where Hirnantian faunas are trilobite-dominated. Usually only the single genus, *Dalmanitina* is present although it is often represented by specimens of unusually large size. Lesperance (1974) refers to a *Mucronaspis (Dalmanitina)-Hirnantia* fauna as a combination of the trilobite and brachiopod-dominated *Hirnantia* faunas. The Welsh localitites have mainly yielded *Dalmanitina* on its own and it appears that in Wales the *Hirnantia* fauna (s.s.) and the *Dalmanitina* bearing assemblages are separate entities.

Palaeogeographic distribution of the *Hirnantia* faunas in North Wales

The Hirnantian faunas under discussion all occur at localities which lay within the Welsh basin. This was a fault-bounded trough in which a large volume of sediment accumulated, although during the Ordovician the faults did not always mark the topographic divide between shelf and basin. In mid-Caradoc times, for example, a shallow shelf extended fully across the northern part of the basin. Further south the transition from shelf to basin lay near Bala, some 40 km west of the Church Stretton Fault (Brenchley & Pickerill 1980) and the topographic basin occupied only the medial part of the overall fault-bounded trough.

The Ashgill palaeogeography was apparently broadly similar to that of the middle Caradoc as shown by the occurrence of varied brachiopod and trilobite shelf faunas across the northern part of North Wales as far west as Bala, whereas graptolitic or unfossiliferous mudstones occur beyond Bala (Fig. 2 & 3). It is possible to determine a more detailed palaeogeography for Rawtheyan times based on the

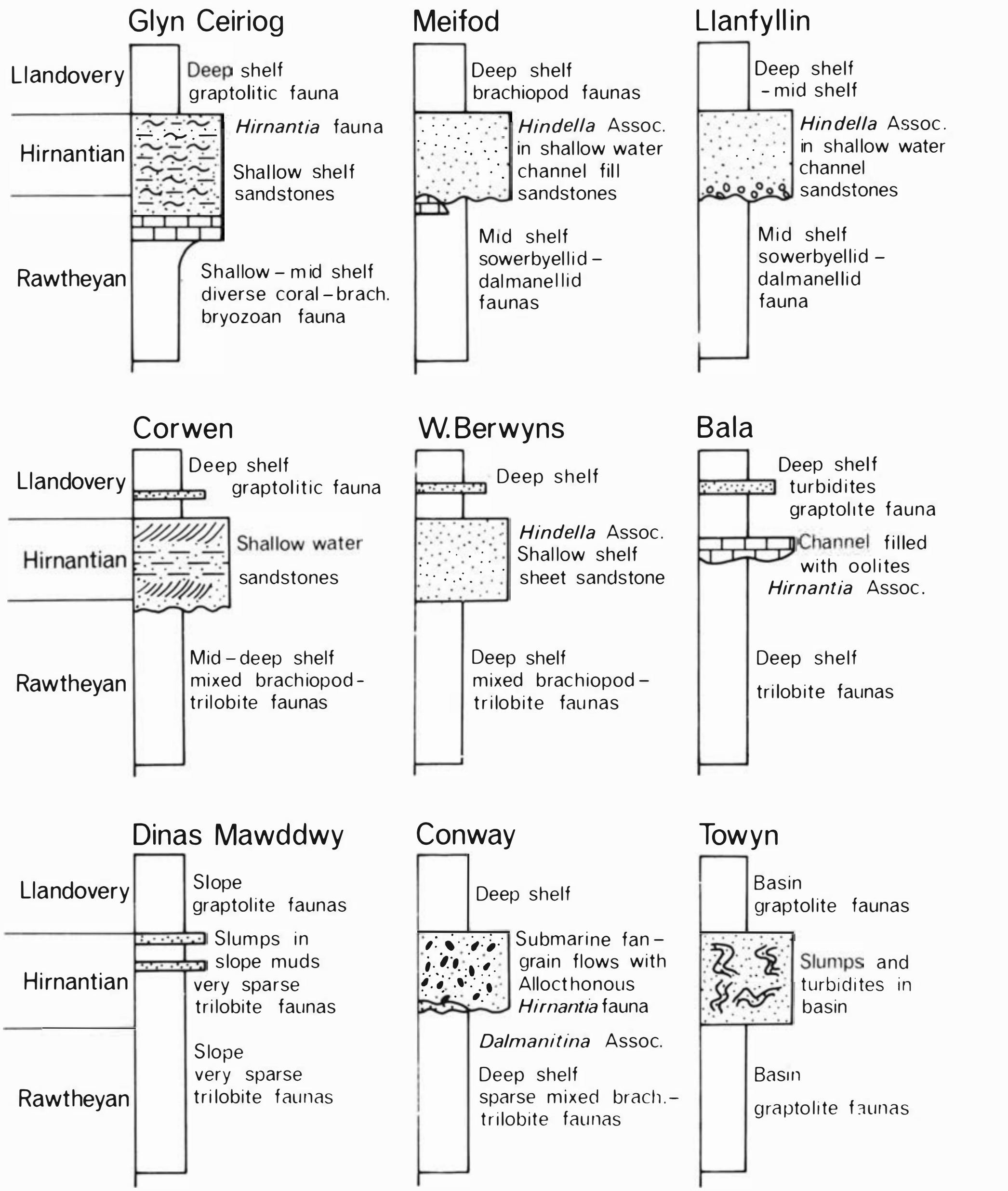

Fig. 2. Simplified stratigraphical columns and environmental interpretations for sections through the Upper Ordovician and lowermost Silurian in North Wales.

composition of the benthic faunas which can be divided into a number of broad faunal assemblages. The more easterly faunas, as illustrated by sections near Glyn Ceiriog (Fig. 2), consist of brachiopods, corals, trilobites, bryozoans and molluscs (Table 2) found in fine sandstones which were deposited in a shallow shelf environment. There is also some localised development of carbonates in this area and the substrate is generally coarser than in the more western areas. The most striking feature of these inshore assemblages is the high proportion of solitary corals and molluscs which only occur rarely in the more offshore assemblages. Table 2 shows that a number of eurytopic brachiopod genera, particularly *Sowerbyella* and *Dalmanella,* occur in most of the areas investigated in North Wales. However these two brachiopod genera are particularly dominant in the brachiopod assemblages at Meifod and Llanfyllin where they account for more than 50% of the total preserved biota. These assemblages are assigned to a *Sowerbylla-Dalmanella* association. Trilobites are present in the association at some localities but the number of trilobite genera is relatively small and the fauna is generally dominated by brachiopods.

Further west near Lake Vrynwy in the western Berwyns there is a more varied fauna which includes brachiopods, trilobites and abundant bryozoa. The high proportion of articulated brachiopod valves and associated unbroken bryozoans probably reflect quiet water conditions. It is suggested that the mixed brachiopod assemblages lived in the outer part of the midshelf. The faunas from areas further west are often very variable but may be placed into two main types, namely diverse trilobite-dominated faunas such as those at Llanystumdwy, near Criccieth (Price 1981) and in the Rhiwlas Limestone at Bala (Bassett *et al.* 1966), and sparse mixed brachiopod-trilobite faunas such as those at Conway. In the Bodeidda mudstones (Cautleyan-Rawtheyan) at Conway only a few genera of brachiopod are present (see Table 2) and these are generally small and thin-shelled. *Christiania* is the dominant brachopod and is locally abundant at Conway but is rare at all other localities. There are in addition several genera of trilobites notably *Tretaspis,* in greater abundance than in most other western localities. One striking feature of the sparse mixed fauna is the almost total lack of any fauna other than brachiopods and trilobites.

In the basinal areas such as Dinas Mawddwy only a very sparse fauna is found in the Rawtheyan sediments. The fauna is most usually graptolitic and trilobites are only rarely found.

It is believed that the faunal associations found in the Rawtheyan are similar in nature to the depth-related associations described from other horizons in the Ordovician. A *Dalmanella-Sowerbyella* association is generally regarded as being an inner shelf association (Pickerill & Brenchley 1979; Hurst 1979; Lockley 1983), while assemblages with genera such as *Eoplectodonta* and *Christiania* are regarded as indicating a deep shelf environment (Williams & Wright 1981; Brenchley & Cocks 1982). Trilobite-dominated assemblages which pass laterally into graptolitic faunas would be consistent with an environmental range from slope to basin.

The rather monotonous Rawtheyan mudstones which mantle large parts of both the shelf and basin were succeeded by a varied suite of Hirnantian sediments deposited during a period of low stand sea level (Fig. 2). The eastern areas are characterised by sheet sandstones and thin lenses of conglomerates which were deposited in channels eroded into the Rawtheyan mudstones.

The orthid-coral assemblages dominated by *Hesperorthis* or *Nicolella* occur in thin ripple-laminated fine sandstones of the Glyn Formation (Rawtheyan-Hirnantian) near Glyn Ceirog (Fig. 2). These assemblages may have been transported from further east and may therefore be representatives of faunas which lived inshore of the *Hindella* association. The overall character of the Glyn Formation is strongly suggestive of deposition within a wide channel developed near the shoreface.

The *Hindella* association also occurs within similar channel-fill sandstones and there was probably some transport of fossils, which makes it difficult to determine the exact environmental position of the fauna. However, the ratio of opposing valves for the brachiopods is very close to unity and no large net transport is indicated. The *Hindella* association probably inhabited a shallow shelf environment and may have lived in or close to the channels in which the association is commonly found. The *Hirnantia* association as seen at Bala (Fig. 2) is almost certainly autochthonous and occurs in mudstone in an outer shelf position although water depth was probably not great. The *Hirnantia* association also occurs in sandy facies and in oolitic limestones (Hirnant Limestone). The ooids of the Hirnant Limestone are almost certainly allochthonous but the fauna collected from the limestone and the surrounding mudstones is not significantly different and the fossils in the limestone do not show any evidence of being transported over a great distance, since they are mainly unbroken and opposing valves occur in similar numbers. The fauna

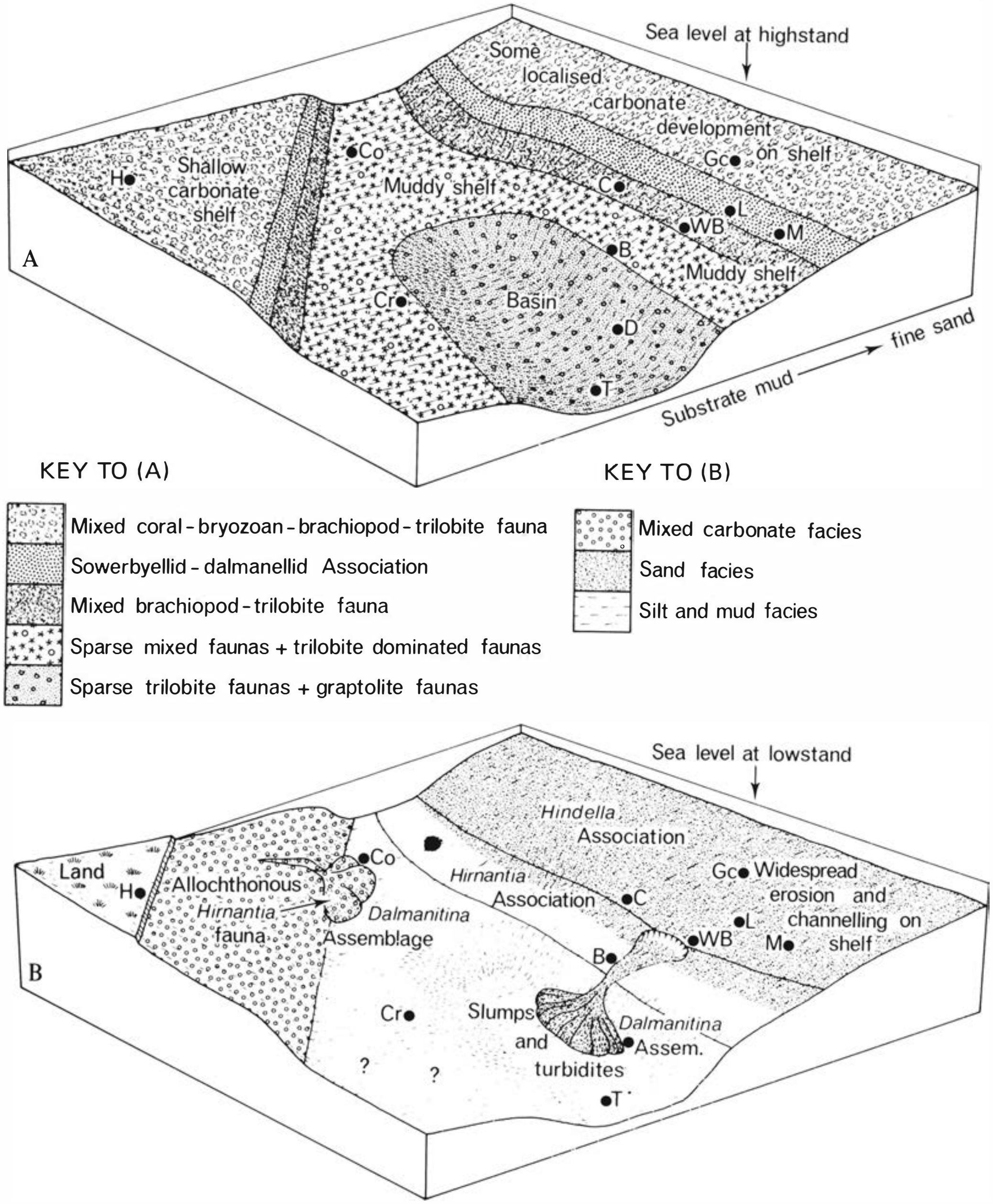

Fig. 3. Palaeogeographic reconstruction of the Northern part of the Welsh Basin during the Rawtheyan (A) and Hirnantian (B) stages, showing distribution of the faunal associations.
Locations: B = Bala, C = Corwen, Co = Conway, Cr = Criccieth, D = Dinas Mawddwy, Gc = Glyn Ceiriog, H = Holyhead, L = Llanfyllin, M = Meifod, T = Towyn, WB = Western Berwyns.

was probably derived from the adjacent muds and swept into the channel during the deposition of the ooids.

The *Dalmanitina* assemblage occurs at Conway 1.2 m below the base of the Conway Castle Grits (Fig. 2) which are a series of calcarenitic proximal turbidites containing an allochthonous *Hirnantia* fauna transported from the

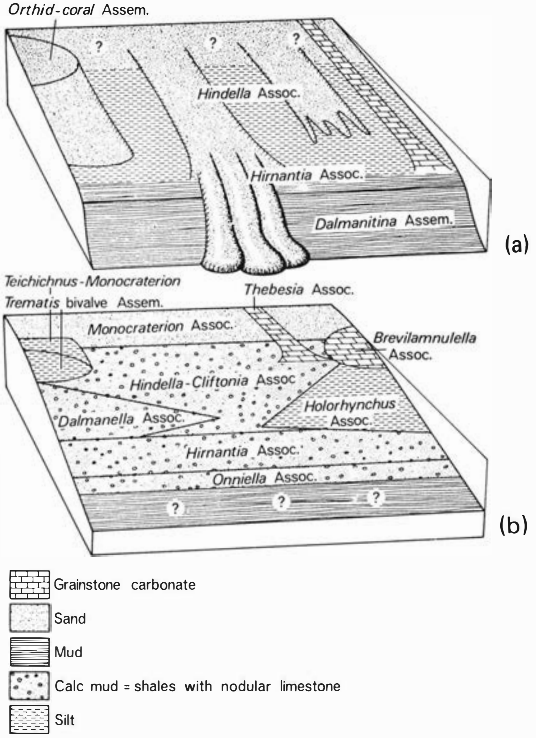

Fig. 4. Reconstructions of the environmental distribution of Hirnantian faunal associations in (a) North Wales and (b) Oslo–Asker district, Norway.

west. These turbidites accumulated in a submarine fan when sea-level fell in Hirnantian times. The *Dalmanitina* assemblage would have been the local autochthonous fauna occurring in mudstones of the deep shelf. The *Dalmanitina* assemblage also occurs in the Foel-y-Ddinas Mudstones (Hirnantian) at Dinas Mawddwy (Fig. 2) but the same formation when traced laterally contains the *Hirnantia* association 18 km to the north east at Bala. The *Dalmanitina* assemblage appears to have occupied a position further down the palaeoslope that the *Hirnantia* association. The environmental distribution of the *Hirnantia* fauna in North Wales is summarised in Fig. 3.

Associations of the Hirnantia fauna in the Oslo region

Three associations belonging to the *Hirnantia* fauna have been recognised at the very base of the Hirnantian in the Oslo–Asker district of southeastern Norway (Brenchley & Cocks 1982). They are the *Hirnantia* association *s.s.*, the *Dalmanella* association and the *Hindella-Cliftonia* association. Several genera are common to the three associations but they are present in very different proportions.

In addition to the three associations of the *Hirnantia* fauna there are four other associations present in Hirnantian rocks of the Oslo–Asker district. An *Onniella* association persisted from the Rawtheyan into rocks probably belonging to the lower part of the Hirnantian, a *Holorhynchus* association is locally present in the area in the middle part of the Hirnantian, and two associations, the *Thebesia* association and the *Brevilamnulella* association, are found in shallow marine facies as the top of the Hirnantian stage (Brenchley & Cocks 1982).

Palaeogeographical distribution of the *Hirnantia* fauna in the Oslo region

In Ordovician times an extensive, relatively shallow marine platform occupied the Baltic region and extended about 1500 km from western Norway to eastern Estonia. The western margin of the platform was raised to form a palaeoslope which dipped gently eastwards toward the central part of the platform. The Oslo–Asker district lay on this gentle palaeoslope which influenced the nature of the sediments. In the west of the district these were more richly fossiliferous and calcareous than those in the east (Kiær 1902; Størmer 1967; Brenchley & Newall 1980). A broadly similar palaeogeography persisted into early Hirnantian times when block faulting of the basement coupled with a probable glacio-eustatic lowering of sea level radically changed the distribution of facies.

The three associations of the *Hirnantia* fauna known from the Oslo–Asker district are all found in bioturbated shelf mudstones which lie in a regressive sequence below sandy shoreface sediments.

The *Hindella-Cliftonia* association is found in an up-slope position, in relatively calcareous mudstones (Fig. 4b), the *Hirnantia* association is found downslope in more argillaceous sediments to the east and the *Dalmanella* association occurs in an intermediate position (Fig. 4b). The presence of many articulated valves of *Cliftonia* and *Hindella,* and a well preserved life assemblage of the latter indicate that the *Hindella-Cliftonia* association probably lived below wavebase, while the other associations presumably lived in somewhat deeper parts of the shelf. Some elements of the *Hirnantia* fauna are found stratigraphically higher in the sequence where shelf mudstones pass upwards into shoreface sandstones. *Hirnantia, Eostropheodonta, Hindella* and *Plectothyrella* are found associated with *Trematis,* corals and bivalves in silty and sandy sediments reflecting an environment near the bottom of the shoreface (Brenchley & Cocks 1982). The shoreface, and associated sediments deposited in tidal channels, have trace fossils such as *Monocraterion,* but most lack body fossils, except for the local presence of the *Thebesia* and *Brevilamnulella* associations. Both of these associations are found in coquinas formed of locally transported valves deposited in channels (Fig. 4b).

The *Onniella* and *Holorhynchus* associations are both found in bioturbated shelf mudstones. The former is believed to have occupied a deep shelf environment while the latter may have lived in an inner shelf environment comparable to that occupied by *Pentamerus* in the succeeding Silurian (Brenchley & Cocks 1982).

In conclusion, the position of the *Hirnantian* fauna in a regressive sequence, below shoreface sediments, provides convincing evidence of a shelf environment for the fauna. Although some elements of the fauna probably lived close to the shoreface, the shallowest of the well-defined associations, the *Hindella-Cliftonia* association, typically occurred below wave base and the other two associations in positions slightly further down the palaeoslope. Co-existing with the *Hirnantia* fauna but living in other parts of the shelf were the *Onniella* and *Holorhynchus* associations. Sediments of the shoreface generally only supported a fauna recorded as trace fossils but two low diversity and locally restricted associations, the *Thebesia* and *Brevilamnulella* associations, colonised shallow maine, channel environments.

Discussion

The eight associations in the Hirnantian of the Oslo–Asker district are believed to have been eight distinct biofacies representing eight biotic communities. This degree of partitioning of the shelf biota into associations appears comparable with that found lower in the Ordovician where about six associations, excluding ichnofacies, are typical of the Caradoc (Lockley 1983). Similarly, about five brachiopod associations are typically found in Lower Silurian shelf sediments (Ziegler 1965). The presence of oolites and reefs in the Upper Ordovician of the Oslo region, Norway, supports the limited palaeomagnetic data in placing Baltoscandia in or near the southern tropics (cf. Ziegler *et al.* 1977). The variety of associations could therefore be a feature of the Ordovician tropics.

In North Wales, an area which was probably more temperate, there were no clearly defined associations in addition to the *Hirnantia* fauna, but the fauna itself, can be sub-divided into two associations. Other nearshore assemblages appear distinctive in being coral-bivalve rich and may represent a separate association. In deep shelf or slope environments assemblages generally lack brachiopods and are characterised by the trilobite *Dalmanitina.* There were therefore only two brachiopods associations in the Hirnantian which contrasts with the six in the Caradoc of the same region. Elsewhere in the world there is insufficient information on the environmental distribution of late Ordovician faunas to determine whether or not there were several co-existent biofacies in any particular area. However, the lack of evidence may in itself be significant and might suggest that the differentiation of the Hirnantian biota into associations was not pronounced outside the Ordovician tropics.

The associations of the *Hirnantia* fauna apparently occupied an environment which ranged from the inner to the deep shelf, and could be assigned to benthic assemblages three, four and possibly five according to the scheme proposed by Boucot (1975).

Several of the species of the *Hirnantia* fauna are eurytopic and are present in all the associations. Eurytopic species are not unique to the *Hirnantia* fauna but are found in earlier Ordovician associations. For example, *Howellites antiquior* and *Sowerbyella sericea* are found in at least three associations in the Caradoc of North Wales (Pickerill & Benchley 1979; Lockley 1980) and the inarticulate brachiopod, *Paracraniops doyleae* is found in as many as eight associations in the Caradoc of Shropshire (Hurst 1979). However, in both these areas there are no species, except *Paracraniops doyleae,* which have a range across most of the shelf, as is the case for several species of the *Hirnantia* fauna.

The characteristic genera and species of the *Hirnantia* fauna are not only eurytopic, as noted above, but are also cosmopolitan (Lespérance 1974; Rong 1979; Jaanusson 1979). Brachiopod faunas in general became progressively more cosmopolitan during the late Ordovician and though there were still local, endemic species in Hirnantian times, faunal provinces were no longer clearly differentiated (Jaanusson 1979). The existence of this cosmopolitan fauna at a time when there was a large polar ice cap, and therefore probably well developed climatic belts, is unexpected (Lespérance 1974). It does not apparently conform with the models of Valentine (1973), which predict that faunas should be particularly well differentiated when global climates are most strongly contrasted. One possible explanation of this apparent anomaly is that the *Hirnantia* fauna was a cold water fauna as suggested by Sheehan (1979) and that during the late Ordovician glaciation the temperate climatic belts contracted to such an extent that cool oceanic waters extended from the poles to areas which were formerly tropical (Skevington 1974, Sheehan 1979).

The evidence for the *Hirnantia* fauna being a cold water fauna is:

1. It characteristically occurs in areas which were previously occupied by faunas of the Mediterranean Province, and is composed of some genera which have Mediterranean origins (Jaanusson 1979; Sheehan 1979). The Mediterranean fauna probably occupied a temperate or circum-polar position (Spjeldnæs 1961, 1981; Havlíček & Vaněk1966).. However the evidence is not conclusive. There had been a progressive breakdown of provincial barriers from mid-Ordovician times onwards, probably related to changes in the geography of the continental plates, so that by the late Ashgill there was a general increase in the number of pandemic genera (Williams 1973). The *Hirnantia* fauna is found in areas well outside the former Mediterranean province and in addition to cosmopolitan genera of Mediterranean origin (e.g. *Hirnantia, Drabovia*) there are several new pandemic genera (e.g. *Plectothyrella, Eostropheodonta*) and other pandemic genera persisting from earlier times (e.g. *Dalmanella, Bancroftina* c.f. Havlíček 1976). The Hirnantian was thus a time when cosmopolitan genera flourished, some of which were of Mediterranean origin.

2. The *Hirnantia* fauna shows, in common with the Mediterranean fauna (except in the mid-Ashgill) a rather low diversity of brachiopod species, but a eurytopism of these species. It is likely that the cool water faunas in the Ordovician were generally poorly differentiated into associations and were geographically widespread. This is consistent with many colder water faunas at the present day which tend to have generalist species and fewer associations (Valentine 1973; 261, 365).

The evidence against the *Hirnantia* fauna being a cold water fauna is that it is found in a few sequences, such as those in the Oslo region, with characteristically tropical facies, including oolites. However, in these sequences the *Hirnantia* fauna is confined to clastic facies and is not found in the warm water carbonates, so the fauna could have colonised the shelf area during the colder phases of the Hirnantian.

Although the evidence is inconclusive we believe it favours a cold water environment for the *Hirnantia* fauna. It might therefore be significant that at the end of the Hirnantian, when the ice cap melted and a warmer regime was probably initiated, there was an extinction of most of the remaining remnants of the Mediterranean trilobite fauna and the early Silurian diversification of the brachiopods came not from Mediterranean stocks, but from the more tropical, Hiberno-Salairian faunas (Jaanusson 1979). It therefore appears that the cold water cosmopolitan *Hirnantian* fauna of the late Ordovician might have been replaced by cosmopolitan, but warm water faunas in the Silurian.

Table 1. Composition of the Hirnantian faunal associations based on assemblages at selected localities in North Wales, percentages refer to percentages of total fauna. Diversity is measured by the number of taxa in an assemblage. Locations refer to those marked on Figure 1.

	BALA	*W. BERWYNS*	*LLANFYLLIN*	*MEIFOD*	*GLYNCEIRIOG*
Brachiopods					
Hirnantia	21.1%	11.2%	13.8%	19.8%	5.9%
Eostropheodonta	32.1%	11.2%		8.9%	2.0%
Dalmanella	6.8%	5.2%	5.1%	5.5%	13.7%
Plectothyrella	5.0%	2.2%			
Hindella		9.0%	25.1%	20.3%	
Kinnella	3.6%	4.5%		0.5%	
Bancroftina	23.3%	0.7%			
Sowerbyella	0.4%	2.2%	3.1%		3.9%
Christiania			1.6%		
Diambonia			1.0%		
Leangella				0.5%	
Rhychotrema		0.7%	1.5%	2.0%	
Plaesiomys		0.7%			
Leptaena					2.0%
Hesperorthis					21.5%
Dolerorthis					7.8%
Sampo					2.0%
Platymena		20.8%			
Epitomyonia		0.7%			
Orthambonites					3.9%
rhynchonellids	5.2%				
Cnidarians					
streptelasmatids			34.4%	11.9%	15.6%
favositids			12.8%		2.0%
conulariids	0.7%				
Bryozoa					
Hallopora	6.1%	20.9%	abundant	27.7%	3.9%
trepostome	1.1%				
dendroid bryozoans				0.5%	3.9%
Molluscs					
bivalves		1.5%	0.5%		2.0%
gastropods			0.5%	0.5%	
Others		2.4%	0.5%	2.3%	10.2%
TOTAL	100.2%	99.1%	99.9%	99.9%	100.3%
Diversity	11	18	14	15	17
Sample size (number of individuals)	280	134	206	202	55

Table 2. Occurrence of the commoner elements of the Rawtheyan faunas at localities in North Wales shown in Figure 1.

	Glyn Ceirog	*Meifod*	*Llanfyllin*	*W. Berwyns*	*Corwen*	*Conway*
Brachiopods						
Dalmanella	9.3%	49.5%	11.4%	12.9%	10.9%	10.9%
Sowerbyella	0.8%	2.8%	59.6%	5.8%	31.3%	17.4%
Anisopleurella		7.3%	1.2%	2.3%		2.2%
Dolerorthis	2.3%		0.8%		3.1%	
Christiania		2.8%				37.0%
Saukrodictya	0.8%	1.8%	0.4%	1.2%		
Skenidioides	1.6%		0.4%	7.0%	0.8%	
Orthambonites		6.4%	3.1%	1.2%		
Glyptorthis			2.4%	10.5%	3.9%	
Schizophorella	2.3%					
Triplesia		2.8%			4.7%	
Rhynchotrema		2.8%			0.8%	
Nicolella	7.0%				0.8%	
Schizonema		3.7%				
Leangella			0.4%			8.7%
Sampo					3.1%	
Orbiculoidea		0.9%		7.0%	0.8%	2.2%
other brachiopods	3.9%	13.8%	2.8%	9.3%	6.4%	4.4%
Trilobites						
Gravicalymene	0.4%	0.9%		0.1%		
Encrinurus	0.4%	0.3%			0.2%	
Prionocheilus	0.1%					
Remopleurides		0.1%		0.1%		
Paracybeloides		0.1%				
Tretaspis				0.1%	0.3%	3.3%
Octillaenus				0.1%		
Stenopareia					0.4%	
Phillipsinella						0.2%
Trinodus						0.4%
Lonchodomas						0.4%
other trilobites	0.4%					2.1%
Bryozoa						
stick bryozoans	17.8%	3.2%	12.9%	22.2%	26.6%	
trepostomes	8.5%			15.2%	2.3%	2.2%
"dendroid"	2.3%				2.3%	
Corals						
streptelasmatids	25.6%	0.9%				
Molluscs						
Bivalves	4.7%		0.4%			
Gastropods	10.9%			1.2%		
Echinoderms						
Cystoids			1.4%	1.2%	0.8%	
crinoid collumnals	common	common	common	abundant	present	rare
conulariids				2.3%		2.2%
TOTAL	99.1%	100.1%	99.2%	99.7%	99.5%	100.1%
Diversity	20	19	19	20	21	17
Sample size (number of individuals)	130	110	255	90	140	46

Table 3. Occurrence of the commoner taxa in the three associations of the Hirnantia *fauna in the Oslo–Asker district, shown as a percentage of the total fauna, and a percentage of the brachiopod fauna (in brackets). Diversity is measured by the number of taxa in an assemblage. (Data showing the variation between localities is given in Brenchley & Cocks 1982.)*

	Hirnantia Association	*Dalmanella* Association	*Hindella/Cliftonia* Association
Hirnantia sagittifera	23% (41%)	< 1%	< 1%
Dalmanella testudinaria	15% (24%)	69% (81%)	4% (9%)
Eostropheodonta hirnantensis	6% (10%)	- -	2% (4%)
Hindella cassidea	2% (3%)	2% (2%)	25% (32%)
Cliftonia aff. psittacina	6% (3%)	2% (3%)	25% (32%)
	6% (13%)	2% (3%)	34% (49%)
Mucronaspis mucronata	4%	< 1%	< 1%
Mean total diversity	14.3	5.3	11.6
Mean brachiopod diversity	6.5	4.0	6.3
Mean sample size	97.5	30.8	63.7

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Ordovician benthic macrofaunal associations

By VALDAR JAANUSSON

A major problem in the quantitative analysis of Ordovician macrofaunal assemblages is not only to take into account groups in which individuals are no longer recognisable (most echinoderms, small bryozoans etc.), but also to integrate the data from large and small macrofaunas. In Baltoscandian carbonate rocks the quantitatively important constituents in the small macrofauna (adult dimensions c. 2–5 mm) are mostly gastropods and acrotretid brachiopods. The numerically dominant macrofaunal species belong to this size category. This suggests that because of difficulties in sampling, the importance of small macrofaunas has been underestimated in many cases.

In graptolitic shales, benthic large macrofaunas are very poorly represented. The small macrofauna is commonly dominated by small inarticulate brachiopods. It is customary to regard both these and associated small plectambonitacean brachiopods as epifauna attached to floating weeds, but all are more probably benthic. The rarity of large macrofaunal organisms in such shales may be due to unsuitable substrate, an oxygen-dependent zonation with large forms disappearing at higher oxygen levels than much of the small macrofauna, or the competitive disadvantage of large macroorganisms in areas of very low food supply.

Along the environmental gradient reflected in the transition from graptolitic shales to progressively coarser grained rocks, the successive benthic macrofaunal associations appear to have been more substrate-related than depth-dependent.

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The purpose of this paper is to summarise the available information on the composition of Ordovician benthic macrofaunal associations. Only level bottom associations are considered, not those of organic reefs or carbonate mounds. Hirnantian faunas are also excluded because they are discussed elsewhere in this book.

Studies of faunal associations require a quantitative approach. In Ordovician rocks such study may be said to have barely started, and therefore quantitative data are few and concentrated to certain areas. The data are insufficient for a synthesis, but some trends are discernible, and the main emphasis in this paper is on tracing such trends. A complicating factor at present is that there is some confusion as to how to obtain and evaluate the quantitative data and what to include and exclude. Thus many published data are difficult to compare because of different approaches in methodology.

For individual-based quantitative estimates of the composition of faunal assemblages a number of different sampling methods are available. The commonest are (1) bulk sampling in friable rock, (2) surface sampling, and (3) bulk sampling in hard rock (for a recent summary see Jaanusson 1979 b: 256–259). Almost all data available from Ordovician rocks are given in terms of individual-based frequencies, that is, groups in which individuals can no longer be recognised (most echinoderms, sponges if represented only by spicules, many small articulate brachiopods, ramose and fenestrate bryozoans, "chitinous" hydrozoans etc.) are seldom included, or the estimates hardly reflect their real importance. For example, studies on the composition of skeletal sand in Ordovician limestones (Martna 1955; Jaanusson 1972; Põlma 1972a, b; Lauritzen 1975; Gnoli & Serpagli 1980; Hints & Põlma 1981) show that echinoderms were important producers of ske-

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letal material, although they are rare or absent in data based on counting individuals. Echinoderm skeletons disintegrate easily into consistuent plates, which in turn are easily fragmented because each plate is porous and composed of a single crystal. That the relative frequency of echinoderms in Ordovician macrofaunal assemblages is almost always strongly underestimated can be shown in cases where the importance of various species can be evaluated in terms of sample-frequency (Hints 1979). Cystoids (e.g., Jaanusson *in* Bruton & Williams 1982: 38, Fig. 6) and crinoids were normally the commonest Ordovician echinoderm groups, but in the Upper Ordovician echinoids were also fairly common in places (Põlma *et al.* 1977). Conversely, the importance of articulate brachiopods is mostly much higher in individual-based data than in skeletal sand. Because of the tough, finely fibrous structure of the shell, the brachiopod skeleton does not break easily, and this contributes to the over-representation of articulate brachiopods in individual-based estimates. The original composition of skeleton-bearing faunal associations therefore differs widely from data on the frequencies of individuals obtained from the rock. The sample-frequency approach (Jaanusson 1979 b: 258–259) yields an estimate of the original composition of the skeleton-bearing fauna that is more in accordance with the composition of the skeletal sand. Because of the lack of other data, the discussion in this paper has to be based on individual-based quantitative data, but the above considerations should be borne in mind.

A serious problem is that the data published by various workers are not always comparable because of different tabulation conventions. This is particularly notable with regard to trilobites. In many papers only the dominant exoskeletal element of a species is considered and, in addition, the number of the element is divided by an arbitrary figure (up to ten) to account for ecdysis. For example, what for one worker would be a trilobite-dominated assemblage, would be dominated by another group for those who use a different tabulation convention for trilobites. Because of limitation of space the problem cannot be analysed here in detail. It is sufficient to mention that the effect of ecdysis is commonly vastly exaggerated (see also Cisne 1973), and that several assumptions on which recalculations are based are doubtful (such as that all specimens of the rarer exoskeletal element belong to individuals represented by the dominant element). The fact is that the primary quantitative data of many groups are distorted to varying degrees relative to the original number of individuals, but, on the other hand, recalculations of the data tend to increase the distortion of the original relative frequencies rather than vice versa. The original data presented in this paper, have not been recalculated.

The importance of various groups in macrofaunal assemblages can also be estimated in terms of volume (Walker 1972; Walker & Alberstadt 1975; Copper & Grawbarger 1978). This is a useful approach when large skeletons are involved (stromatoporoids and corals in organic reefs etc.) but it is difficult to apply where small fossils are involved. The quantitative information based on volume and number of individuals is not comparable.

Serious methodological problems also arise when widely different size groups are combined in the same quantitative analyses. Marine zoologists (see Mare 1942) distinguish between meiofaunas (c. 0.5–2 mm), small macrofaunas (c. 2–5 mm) and large macrofaunas (larger than about 5 mm). Because of the effect of the frequency dominance of the smallest forms (Jaanusson 1979 b: 260–262) and differences in effective sampling methods, it is very difficult to integrate the meiofauna and macrofauna in the same analysis. For example, in my opinion, attempts to consider ostracodes together with the macrofauna (Bretsky *et al.* 1977; Pickerill & Brenchley 1979; Lockley 1980 etc.) have not led to reliable results. In most rocks it is also difficult to integrate small and large macrofaunas in the same analysis. In many Lower and Middle Ordovician limestones of Sweden various authigenic minerals (mostly glauconite and chamosite) fill cavities within small shells. Insoluble residues from such limestones abound in small gastropods of the size range of the meiofauna and small macrofauna (Jaanusson & Mutvei 1953, Pl. 4; Jaanusson 1955, Pl. 10, 1960, Pl. 1, 5:1; Dzik 1978, Fig. 5). The existence of this gastropod fauna would hardly have been recognised with normal sampling. Studies of thin sections show that it is also

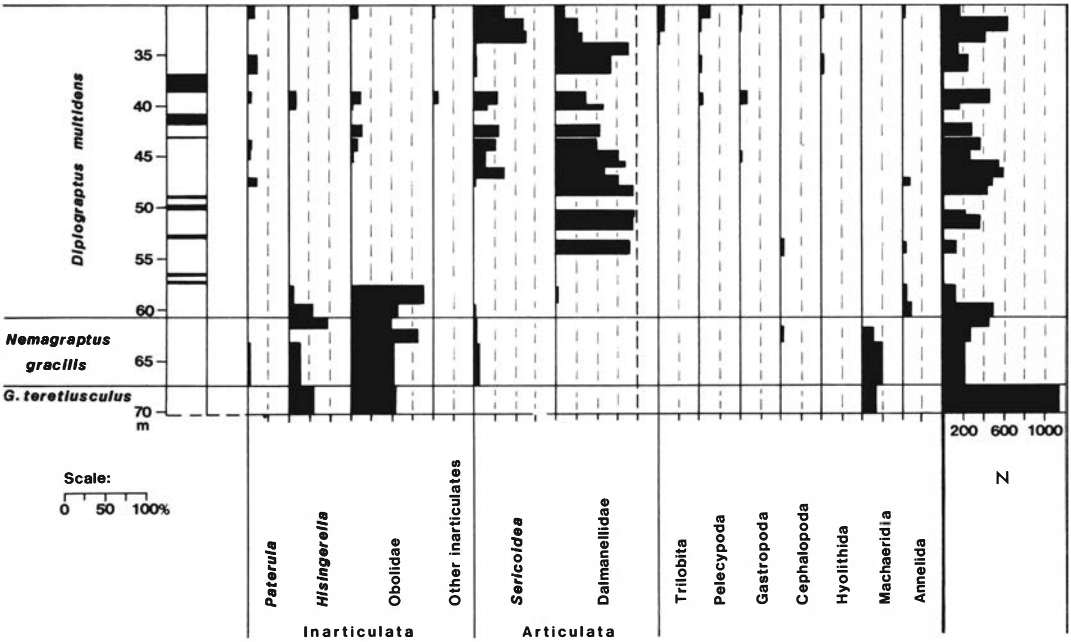

Fig. 1. Composition of the benthic microfauna in the lower part of the Koängen core, Scania, southern Sweden (S. Bergström & Nilsson 1974; Nilsson 1977), based on counts by Ragnar Nilsson (personal communication). In order to obtain an adequate sample size, data are pooled within various intervals of the core; note that this procedure causes a certain loss of resolution. Occurrences of less than 2% are not indicated. Black portions in the rock column refer to relatively thick bentonitic beds. Inarticulate brachiopods belong almost exclusively to the small macrofauna, while other groups represent the large macrofauna. Sericoidea *forms a borderline case with large adult valves somewhat exceeding a width of 5 mm (J. Bergström 1968, Fig. 6).*

present in many rocks without authigenic mineralisation. The gastropods include probable protoconchs (Dzik 1978) but possibly also small species which may have occupied a niche comparable to that of modern hydrobiids or small rissoids. Insoluble residues also invariably include numerous acrotretacean brachiopods in the size range of meiofauna and small macrofauna. The number of specimens of small inarticulate brachiopods vastly exceeds that of the total large macrofauna in many samples (for an example see Fig. 7). Finally, experience has shown that it is very difficult to obtain reliable quantitative data on the small macrofauna from a hard rock which can be compared with data on the large macrofauna in the same beds.

Thus, for several reasons meio- and macrofaunas should not normally be integrated in quantitative analyses of faunal assemblages, and in many cases it is also advisable to analyse small and large macrofaunas separately. In the following discussion small and large macrofaunas are integrated in the few cases where sampling has been adequate in this respect (e.g., Figs. 1–2,5), but because of the lack of reliable information mostly only the large macrofauna is considered.

The term *assemblage* is used here for faunal remains in a sample and *association* for a consistent co-occurrence of taxa. The existence on the sea floor of communities that form discrete ecological entities in which various species are dependent on each other because of strong biological interaction is doubtful. Most species are distributed individually in an overlapping manner and form continua along environmental gradients (Gray 1974).

A great number of various major taxonomic groups provide dominant species in Ordovician macrofaunal assemblages in terms of relative

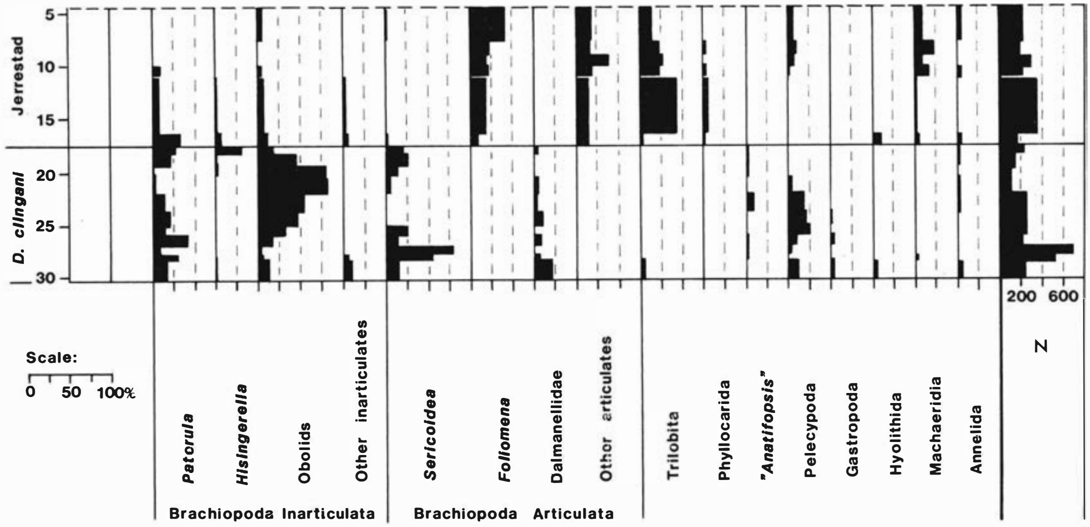

Fig. 2. Composition of the benthic macrofauna in the upper part of the Koängen core, Scania, southern Sweden (Nilsson 1977), based on counts by Ragnar Nilsson (personal communication). In this core there is a break between the Dicranograptus clingani *Zone and the Upper Ordovician Jerrestad Mudstone (*Dicellograptus complanatus *and* D. anceps *zones), comprising the* Pleurograptus linearis *Zone. None of the trilobite species in the Jerrestad Mudstone reaches as high as 10% of the total fauna. See also explanation of Fig. 1.*

frequency of individuals. Such groups are articulate and inarticulate brachiopods, trilobites, leperditiid ostracodes, bryozoans, pelecypods, gastropods, cephalopods, tentaculitoids (see Hurst 1979), hyoliths (see Jaanusson & Mutvei 1982, Fig. 5), cystoids, and in places probably also some other benthic group. In addition, there are assemblages without any obvious dominant species (equitaxial associations; see Appendix, samples 4–5; from personal observations, the large macrofauna in the Ashgill rocks (5a) in Ringerike, Norway, also appears to have a similar equitaxial composition). A benthic macrofauna without numerically dominant species is unusual (for recent discussion see Jaanusson 1979 b: 259–261), and in such cases the question arises as to whether the dominant species might have been soft-bodied organisms.

Examples are given below of various Ordovician associations, separately for each major taxonomic group that provides dominant species and with emphasis on successions of associations that were controlled by environmental gradients. The majority of the available quantitative data on the composition of the fauna is from the Middle Ordovician, and this is considered first in the following discussion.

Brachiopod associations

A common environmental gradient reflected in the lithological transition from graptolitic shale to coarser grained, shelly deposits, can be illustrated by data (Figs 1, 2) provided by Ragnar Nilsson from the Koängen boring in Scania, southern Sweden (Nilsson 1977). In graptolitic shales from *Glyptograptus teretiusculus, Nemagraptus gracilis* and basal *Diplograptus multidens* zones, large macrofaunas are very rare and machaeridians form the dominant component. The non-graptoloid fauna consists mainly of inarticulate brachiopods, mostly obolids, in the size class of small macrofauna. Such an abundance of small inarticulates is a fairly common phenomenon in graptolitic shales (Ruedemann 1934; Bulman 1964). It is customary to interpret the inarticulates as epifauna attached by their pedicles to floating weeds. However, this is hardly probable (see also Cisne 1973) for several reasons which cannot be fully discussed here. Lingulelline obolids, which form an important constituent of the small macrofauna in several portions of the core (particularly in the *G. teretiusculus* Zone), were obviously endobionts; large lingulellines have frequently been found in vertical life position, and they had a mode of life and function of the fleshy pedicle comparable with those of modern lingulids. Obolellines and paterulids most probably also belonged to the endofauna. In my view, the problem in this context is not the relative abundance of small inarticulates but the rarity of associated large benthic macrofaunal organisms. Several explanations are possible. (1) The very fine grained terrigenous mud was an unsuitable substrate for large skeleton-bearing organisms. (2) There may have been an oxygen-dependent zonation of the fauna, with the large skeleton-bearing macrofauna disappearing at higher oxygen levels than much of the small macrofauna. (3) The large macrofauna may have had a competitive disadvantage over the small macrofauna in areas of very low food supply.

The articulate brachiopods normally found in Middle Ordovician graptolitic shales are small aegiromenines, such as *Sericoidea* and *Chonetoidea,* in the size class of small macrofauna but with large adults of several species somewhat exceeding 5 mm in width. It has been claimed that these forms also lived as epiplanktic epifaunas (Havlíček & Vaněk 1966: 39–40; Havlíček 1967; J. Bergström 1968) or were attached to algal fronds which sank to the bottom (Sheehan 1976). However, aegiromenines obviously lacked a pedicle, and it is difficult to understand how their shells were attached (see also Hurst 1979: 225). It is probable that these forms, like many other plectambonitaceans, had a recumbent mode of life, lying free on the sediment surface.

With increasing coarseness of the substrate dalmanellids become a common or even dominant component of the large macrofauna (Figs. 1–2). A pair of species, consisting of a recumbent aegiromenine and a pedunculate dalmanellid, forms an important macrofaunal consistuent in many Middle Ordovician shales and mudstones (see also Hurst 1979 and Lockley 1980).

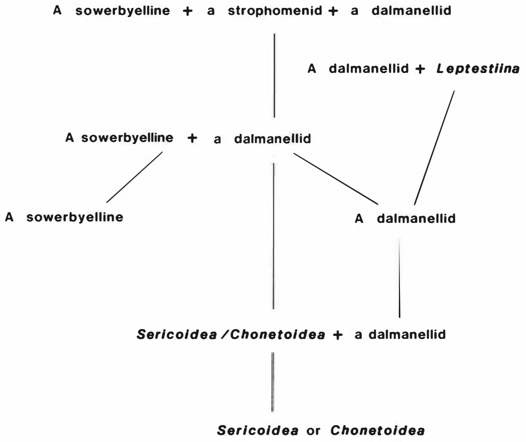

Fig. 3. Diagrammatic presentation of a common Middle Ordovician succession of articulate brachiopod associations along the gradient from graptolitic shales to coarser grained rocks.

In somewhat coarser sediments sowerbyellines (*Viruella, Sowerbyella* or *Thaeorodonta,* rarely *Eoplectodonta*) appear to replace aegiromenines as dominant recumbent macrofaunal forms. Williams *et al.* (1981: 685) suggested that sowerbyellines were mobile, but this appears to be very unlikely. The earliest known assemblages in which a sowerbyelline forms the dominant constituent of the large macrofauna occur in the Fairfach Group (upper *Didymograptus murchisoni* Zone) of Wales (Williams *et al.* 1981). In these beds *Viruella antiqua* (Jones) occurs in association with a common dalmanellid, and such an association of a recumbent sowerbyelline and a pedunculate enteletacean species as dominant elements in skeleton-bearing large macrofaunas is widely distributed in the Middle and Upper Ordovician. It has been described from the British Isles (Hurst 1979; Pickerill & Brenchley 1979; Lockley 1980; Williams *et al.* 1981) and the U.S.A. (Fox 1962; 1968; Bayer 1967; Bretsky 1970a; Titus & Cameron 1976; Cisne & Rabe 1978; Rabe & Cisne 1980). A similar association characterises parts of the upper Middle Ordovician Saue beds of the Keila Stage in north-western Estonia (Appendix, sample 2). The sowerbyelline and dalmanellacean species also occur separately in almost unitaxial assemblages, crowding bedding planes or thin beds (see Appendix, sample 3). At a further stage, a relatively large, common or even dominant strophomenacean species, normally with a recumbent mode of life and thus belonging to

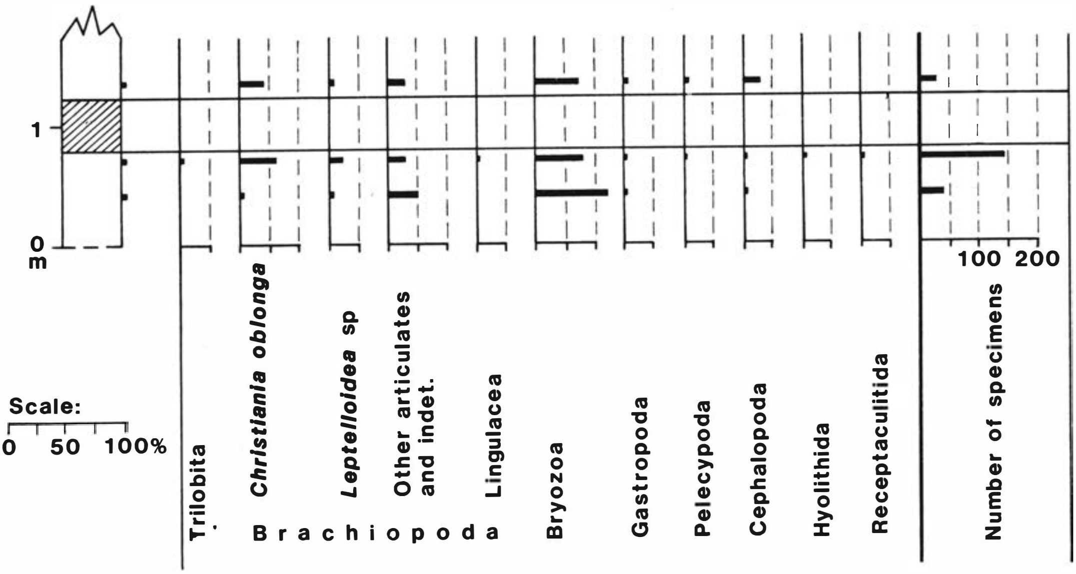

Fig. 4. Composition of the large macrofauna in the Kallaste Beds of the Lasnamägi Stage (upper Didymograptus murchisoni *Zone), Kadaka quarry (Rõõmusoks 1970:52), north Estonia; surface sampling by V. Jaanusson in 1940. The portion in the rock column with diagonal shading refers to the index dolomite bed. The composition of the large macrofauna in contemporaneous limestones (lower Folkeslunda Limestone) of the central Baltoscandian confacies belt is shown in Fig. 7 (see also Jaanusson & Mutvei 1982:7).*

the soft bottom epifauna, accompanies a sowerbyelline and an enteletacean species (e.g., Cisne & Rabe 1978; Pickerill & Brenchley 1979; Hurst 1979; Rabe & Cisne 1980). In another, fairly distinctive association the sowerbyelline species is replaced by a species of *Leptestiina,* a recumbent form of about the same size, which occurs in association with a common to dominant dalmanellid species (Hurst 1979; a similar association characterises the Moldå Topoformation of Sweden).

The succession of Middle Ordovician brachiopod associations along the environmental gradient beginning with graptolitic shales is summarised diagrammatically in Fig. 3. There are associations with other dominant brachiopods, but their relationship to the environmental gradient under consideration is not clear. The small strophomenacean *Foliomena,* of about the same shape and size as small sowerbyellines, is the dominant brachiopod in certain Upper Ordovician mudstones (Fig. 2; Sheehan & Lesperance 1978, Table 1; Harper 1979, 1980), in a habitat which appears to be roughly comparable to that of *Sericoidea* or *Chonetoidea.*

In some Middle Ordovician environments the fairly small and globose strophomenacean *Christiania,* a probable ambitopic form, provides the dominant macrofaunal species (Fig. 4; *C. subquadrata* at the base of the Arline Formation in Tennessee etc.).

Several Middle and Upper Ordovician associations in which an articulate brachiopod species is dominant have a biogeographically restricted occurrence. This is particularly evident in some very shallow water assemblages of the North American Midcontinent Province, where rhynchonellaceans such as *Rostricellula* (Walker & Alberstadt 1975; Shaw 1980; the genus is known to be dominant in several other areas), *Ancistrorhyncha* (e.g., Corbin Ranch Formation of the Arbuckle Mountains, Oklahoma) and *Orthorhynchula* (Bretsky 1969, 1970a), and the atrypid *Zygospira* (Bretsky 1969, 1970a; Copper & Grawberger 1978) are very common in places.

The Lower Ordovician brachiopod associations along the gradient beginning with graptolitic shales are poorly known. Fjelldal's (1966)

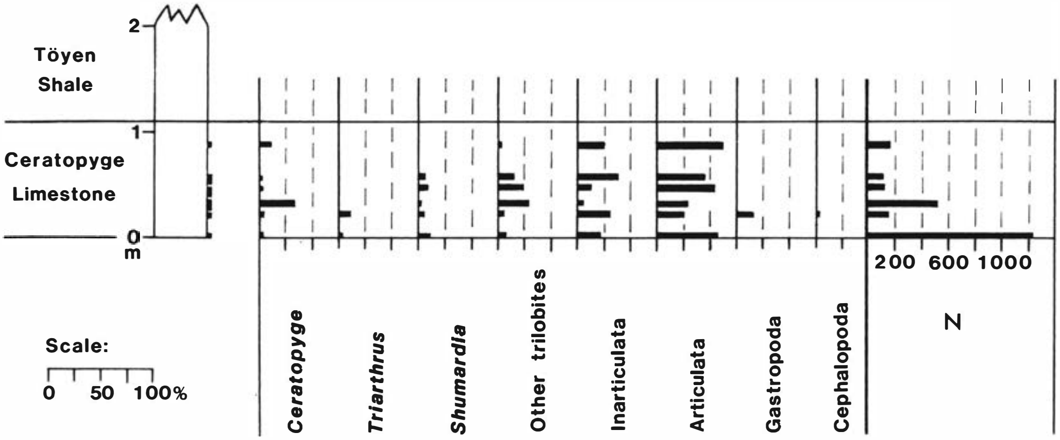

Fig. 5. Composition of the macrofauna in the uppermost Tremadoc Ceratopyge Limestone at Bjerkåsholmen, Oslo Region (after Fjelldal 1966). Bulk sampling. The relative frequencies are shown only for those trilobite genera that form 10% or more of the total fauna in at least one sample.

data from the uppermost Tremadoc Ceratopyge Limestone of the Oslo region, Norway (Fig. 5) show a dominance of small orthid brachiopods among which *Nanorthis? christianiae* (Kjerulf) is the dominant species. Such small species, belonging to the small macrofauna or somewhat larger, are mostly homoeomorphs of later dalmanellids and provide dominant species in many Lower Ordovician assemblages. Examples are *Nanorthis? suecica* Tjernvik in the *Armata* Limestone (*Tetragraptus phyllograptoides* Zone) of Sweden and *Ranorthis parvula* Rubel in the Mäeküla beds (*Phyllograptus densus* Zone) of northern Estonia. The earliest known assemblage in which enteletaceans provide the dominant species is in the Vääna Limestone (Volkhov Stage, middle *D. hirundo* Zone) of northern Estonia (Appendix, sample 1), and from then on enteletaceans form an important constituent in many assemblages.

Trilobite associations

A diagrammatic presentation of a Middle Ordovician succession of trilobite associations along the environmental gradient beginning with graptolitic shales, is given in Fig. 6. The olenid *Triarthrus* tends to be a common to dominant component at the transition from graptolitic shales to graptolitiferous mudstones (Cisne 1973; Fortey 1975: 347; Cisne & Rabe 1978; Rabe & Cisne 1980); similar *Triarthus*-dominated assemblages occur, for example, in the upper part of the Örå Shale, *Dicranograptus clingani* Zone, of Jämtland, northern Sweden). This is a low-diversity, Middle Ordovician equivalent of the Upper Cambrian-Arenig olenid-dominated associations which have a much higher taxonomic diversity of olenids in

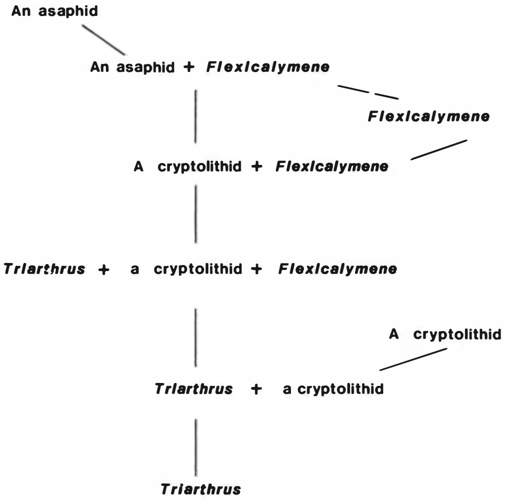

Fig. 6. Diagrammatic presentation of a common Middle Ordovician succession of trilobite associations along the gradient from graptolitic shales to coarser grained rocks.

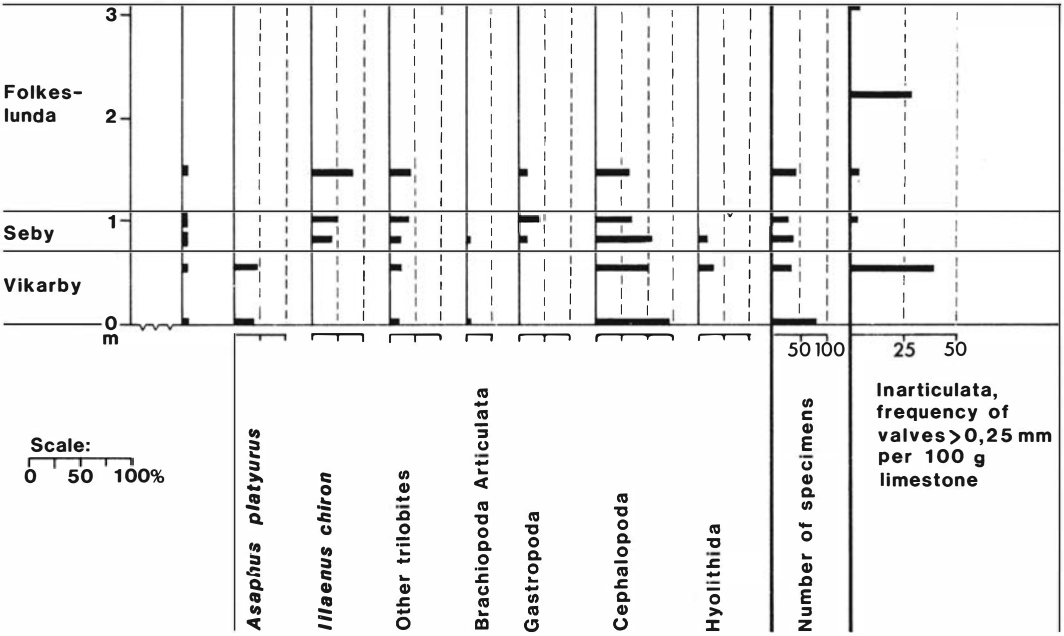

*Fig. 7. Composition of the large macrofauna in the Vikarby, Seby and Folkeslunda Limestones (*Didymograptus murchisoni *Zone) at Vikarbyn (Jaanusson & Mutvei 1953), Siljan district, central Sweden. Bulk sampling by V. Jaanusson and H. Mutvei in 1947. The column on the right shows the frequency of inarticulate brachiopods, mainly acrotretaceans, in the insoluble residue from five limestone samples treated with acetic acid.*

a roughly similar habitat. In sediment reflecting increased water energy, *Triarthrus* commonly occurs in association with a subdominant to dominant *Flexicalymene* or a trinucleid species (Cisne 1973). At a somewhat later stage *Flexicalymene* occurs either alone (Cisne & Rabe 1978; Williams *et al.* 1981) or together with either trinucleid or an asaphid species. At about this stage a trinucleid is also the dominant trilobite species, not only in the Middle Ordovician (Brestky & Brestky 1975; Hurst 1979; Lockley 1980) but also in the pre-Hirnantian Upper Ordovician (J. Bergström 1973; Sheehan & Lespérance 1978, Table 1). Further stages are difficult to analyse at present owing to the paucity of quantitative information, and because of differences in tabulation conventions between various published data.

The described succession of Middle Ordovician trilobite associations is developed parallel to, but largely independent of that of brachiopod associations along the same environmental gradient. Trilobites and articulate brachiopods frequently occur in association but the correlation along the gradient is not very close, implying that the environmental factors that controlled the distribution of these two groups were partly different. In the Koängen core (Figs. 1–2) the Middle Ordovician sequence is almost devoid of trilobites; conversely, the *Triarthus* association tends to be very poor in articulate brachiopods.

Fortey (1975) analysed the succession of upper Lower Ordovician trilobite associations along the same gradient from Ny Friesland, Spitsbergen. There the olenid-dominated association has a far higher taxonomic diversity than in the Middle Ordovician, and is followed by assemblages in which nileids form an important constituent. In the contemporaneous rocks of the central Baltoscandian belt *Nileus* is also common, and here it appears to be characteristic of mud bottoms, now represented by calcilutites and mudstones; on skeletal sand bottoms the genus is rare to absent, as are other nileids. In Ny Friesland the nileid-rich beds are succeeded by high energy, shallow water limestones in which *Illaenus* is the dominant component in a fauna of North American type. In the central Baltoscandian belt, the dominant forms in the roughly contemporaneous calcarenites are the endemic genera *Megistaspis* (lower part) and *Asaphus* (upper part; Jaanusson & Mutvei 1982, Fig. 6).

Middle Ordovician successions of trilobite associations have been described from Canada by Chatterton & Ludvigsen (1978 a) and Shaw (1980). Many taxa in the associations of the carbonate platform are restricted to the North American Midcontinent Province. Particularly characteristic is the association with *Bathyurus* and related bathyurids (see also Ludvigsen 1978 b) which is widely distributed in North America (Jaanusson 1979 a, Fig. 7). A succession of trilobite associations along an inferred environmental gradient in Kazakhstan was outlined by Apollonov (1975).

Pelecypod associations

Pelecypods form an important accessory component in many Ordovician assemblages, but they are particularly common in terrigenous siltstones and sandstones where a particular species is often dominant (e.g., Bretsky 1969, 1970a, b; Bretsky & Bretsky 1975; Spjeldnæs 1979). Many of these common Ordovician clams were byssaly attached epifaunal suspension feeders. In some Middle and Upper Ordovician limestones of the North American Midcontinent region pelecypods are also fairly common (Bayer 1967; Copper & Grawbarger 1978) or even dominant in places (Bretsky *et al.* 1977).

Along the environmental gradient from graptolitic shales pelecypods occasionally form an important constituent of the macrofauna, especially in the *Sericoidea*/*Chonetoidea*–dalmanellid association (Fig. 2; Hurst 1979). The dominant forms there, such as *Similodonta,* were infaunal deposit feeders.

Some other associations

Ordovician leperditiid ostracodes are confined almost exclusively to the North American Midcontinent Province. In many places they form the dominant macrofaunal component in assemblages from very shallow water carbonate deposits (Hatfield 1968; Walker 1972). The association commonly has a very low taxonomic diversity and it obviously inhabited a somewhat extreme environment.

Cephalopod-dominated assemblages appear to be mostly allochthonous, transported and sorted. Assemblages in which cephalopods are dominant characterise, for example, certain beds in the "Orthoceratite Limestone" of the central Baltoscandian belt, especially the Middle Ordovician portion of the limestone (Fig. 7). In these beds abundant orthocone cephalopod conchs are associated with large disarticulated asaphid or illaenid trilobites, the asaphids at several horizons being almost exclusively large pygidia. In places the accumulation of cephalopod conchs occurs in association with probable desiccation cracks and stromatolitic algal mats (Larsson 1973), suggesting supratidal conditions. The only modern analogue known to me in which cephalopod conchs are the dominant skeleton-bearing constituents, is on the sandy beaches just east of the southernmost point of Gran Canaria, north of Punta de Maspalomas. Here, above high-water level, two separate counts of skeleton-bearing organisms showed conchs of *Spirula* (71% and 76%) to be the dominant component of the assemblage, together with *Sepia* (4% and 7%), clams, the cirripede *Lepas,* crabs, and a few fronds of *Halimeda.*

Discussion

The environmental gradient reflected in the lithological succession from various bedded limestones through argillaceous, nodular limestones ("marls") and mudstones to graptolitic shales is mostly regarded as being related to increasing depth of the sea. The biofacies succession along this gradient is therefore fairly generally assumed to have been controlled by depth. However, the tendency for hydrodynamic water energy to decrease with increasing water depth has many exceptions, and for this reason this depositional model should not be applied indiscriminately. In particular, in a lithological succession which involves both carbonate and terrigenous sediments, the effect of competitive sedimentation should be considered. In such cases the sediment type, whether predominantly carbonate or terrigenous, is determined by a competition in the sediment supply from two different sources, i.e. from areas producing carbonate particles and those supplying terrigenous material. Thus if little or no terrigenous material were available for deposition in the mudstone belt, the resulting rock would very probably have been limestone. The effect of competitive sedimentation should be carefully considered particularly in basins or troughs such as the Scandinavian Caledonides or Appalachians, in which deposition was pronouncedly assymetrical with the deposition of carbonates on one side and predominantly terrigenous material on the other. In such areas much of the fine carbonate material was supplied from the cratonic side of the basin and the terrigenous material from the opposite side. In such conditions graptolitic clay could well have been deposited within the same basin at the same depth as carbonate mud or even in shallower waters.

The important ecological point in this context is that below a particular, fairly inconsiderable depth, the depth-dependence of the benthic fauna is relatively slight and is normally overshadowed by the dependence on the substrate. In the general dependence of benthic animals on depth conditions within the archibenthal region, the effects of temperature frequently overshadow those of hydrostatic pressure. In northern Europe, for example, the deposition of modern fine terrigenous mud (of a type that might become lithified into a rock comparable to graptolitic shale) begins mainly at a depth of about 200–250 m, but in some fjords the same sediment is produced at depths as low as 30 m, and eurythermal elements of the fauna adapted to this type of substrate follow the sediment type. In other geographical settings a comparable mud could be deposited in even shallower water. For this reason, in the distribution of various benthic associations of fossils more attention should be focused on the physical properties of the substrate than on assumed depth conditions.

The associations of fossils outlined diagrammatically in Figs. 3 and 6 appear to be basically substrate-dependent, although the distribution of the constituent species was also controlled by temperature, hydrostatic pressure, oxygen content of the sea water, and quantity as well as type of nutrients. A problem in assessing the available data on Ordovician faunal associations is the frequent lack of basic information on the properties of the substrate, and particularly the sediment grain size, in carbonate rocks.

The taxonomic diversity of the skeleton-bearing fauna in fine grained subtidal rocks, either carbonate or terrigenous, tends to be much lower than in rocks that were deposited at higher water energies. However, account should be taken of the fact that such fine grained bottoms were inhabited mainly by soft bottom infaunas, and that in comparable modern environments the importance of skeleton-bearing animals in the total fauna is much lower than in many environments associated with a higher water energy, particularly in those that are inhabited also by hard-bottom organisms (see also Craig & Jones 1966). Thus the taxonomic diversity of skeleton-bearing animals does not necessarily reflect the overall diversity of the fauna, soft-bodied organisms included.

The distributional patterns outlined above indicate that shallow water associations on the platforms tend to be geographically restricted, whereas faunas along the margins of the cratons, mainly in fine grained, argillaceous or terrigenous rocks, have a wider spatial distribution (see also Fortey 1975 and Ludvigsen 1978 a). Because the fine grained bottoms extend into deep water there is a wider spatial continuity of the environment, and this decreases the biogeographical differentiation, even when the environment extends into shallow water where it can be inhabited by eurythermal representatives of the fauna adapted to such a substrate.

Acknowledgements

I am particularly indebted to Ragnar Nilsson for putting the extensive set of original data from the Koängen core at my disposal. Some other original data presented here were obtained jointly with Ralph Männil, Harry Mutvei and Hinrek Neuhaus. Michael G. Bassett critically read the manuscript and suggested valuable improvements. The figures were prepared by Lennart Andersson and Bo Bergman at the Section of Palaeozoology, Swedish Museum of Natural History.

Appendix

Quantitative estimates of the composition of large macrofaunas from the Ordovician of Estonia, referred to in the text. Surface sampling, 1–3 by V. Jaanusson in 1940, 4–5 by V. Jaanusson, R. Männil and H. Neuhaus in 1940.

1. Volkhov Stage, Vääna Beds ($B_{II}\beta$). Mäekula. N = 28. Brachiopods: Articulata, 86%, Inarticulata, 2%, Bryozoa: 12%, Trilobita: 1%. Commonest species: *Paurorthis parva* (Pander) 60%, *Productorthis obtusa* (Pander) 21%.
2. Keila Stage, Saue Beds. Jälgimägi quarry (Rõõmusoks 1970: 257). Surface A is about 0.2 m above surface B. N = 54. Brachiopoda Articulata: 41 specimens, Bryozoa: 9, Pelecypoda: 1, Cephalopoda: 2, Gastropoda: 1. Commonest species: *Sowerbyella forumi* Rõõmusoks 28%, *Horderleyella kegelensis* (Alichova) 26%, *Clinambon anomalus* (Schloth.) 9%, *Strophomena ? occidens* (Oraspõld) 8%. B (N = 33): Brach. Art. 26 specimens, Bryozoa 5, Hyolithida 1. Commonest species *Sowerbyella forumi* 58%, *Horderleyella ? kegelensis* 6%, *Clinambon anomalus* 6%.
3. Keila Stage, Saue Beds. Saue quarry (Rõõmusoks 1970: 256), uppermost exposed beds. N = 66. Brachiopoda: Articulata 56 specimens, Gastropoda: 4, Trilobita: 2. Commonest species: *Horderleyella ? kegelensis* 62%, *Clinambon anomalus* 18%.
4. Vormsi Stage, Vormsi, Saxby N. N = 51. Brachiopoda: Articulata 21 specimens (Orthacea 15, Enteletacea 2, *Eoplectodonta schmidti* 3, *Strophomenacea* 1), Inarticulata (*Pseudolingula quadrata*) 1, Bryozoa: 7, Rugosa: 2, Heliolithida 3, Stromatoporoidea 1, Gastropoda: 5, Trilobita: 5, Dendroidea: 6, Conulariida: 1. The only relatively common species: *Nicolella* n. sp. 16%.

6. Pirgu Stage, Cormsi, Hosholm. N = 37. Brachiopoda Articulata, 11 specimens (Orthacea 2, Enteletacea 1, Plectambonitacea 1, Strophomenacea 3, Porambonitacea 1, Atrypacea (*Eospirigerina*) 3, Bryozoa: 4, Rugosa: 3, *Palaeofavosites* 5, other Tabulata 4, Heliolithida: 2, Gastropoda: 2, Trilobita: 3, Dendroidea: 1, Cephalopoda: 2. No dominant species.

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Palaeoecology of Ordovician Bryozoa

By JUNE R. P. ROSS

Most Ordovician ectoproct bryozoans are found in marine carbonate deposits. Evolution in the Early and Middle Ordovician of not only bryozoans but also other calcium carbonate secreting organisms led to marked changes in the organization and species composition of benthic assemblages. In the early stages of establishment of these calcareous assemblages, bryozoans were sediment stabilizers, sediment binders, and framebuilders of structures, which at times formed mounds on the sea floor. Later in the Middle and Late Ordovician increased species diversity and more extensive areas of hard substrate on which benthic assemblages could become established resulted in widely distributed bryozoan assemblages in depositional areas of lagoons, near shore margins, shoals, shelf margins and shallow narrow shelves, more open widespread shelves, and the edges of shallow cratonic basins. Buildups and accumulations of calcareous assemblages, including bryozoans, at some sites formed bioherms. Diversity of bryozoan assemblages in these different facies varied depending on such factors as the initial organisms settling on the substrate (particularly which bryozoan species first became established), organization of tropic levels, rate of influx of clastic material, and other fluctuations, particularly in the physical environment. Distribution patterns of some of the bryozoan species assemblages show migration of species with transgressions and regressions.

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Representatives of ectoproct bryozoans have a long geological history from early in the Ordovician (Tremadoc, about 500 million years ago) to Recent. These colonial benthic animals made up a large part of many epifaunal assemblages in the Ordovician. They were filter feeders.. Many positioned themselves several centimeters or more above the substrate to which they were attached. Some bryozoans, particularly Early Ordovician species, encrusted the substrate and played an important role in stabilizing sediment accumulating on the sea floor. These particular colonies lay close to the water-substrate interface. Most of the bryozoans were passive filter feeders waiting for food to drift to the colonies. The tentacular movements created water currents and the food was caught in these currents and collected for feeding. It appears that the arrangement of zooecial openings across colony surfaces was related to the geometry of tentacle feeding patterns.

Colony forms in Ordovician bryozoans were varied in both shape and size (Fig. 1) and, in addition, encrusting and laminate forms were common. Under different environmental conditions, the colony form of certain species was plastic and varied, e.g., *Batostoma campensis* in Chazyan strata of New York State was encrusting and laminate where it stabilized and bound lime mud and ramose where it was a framebuilder or where it became established on a firm substrate.

Almost all known Ordovician bryozoans belong to the class Stenolaemata which includes the orders Cryptostomata, Trepostomata, Cystoporata, and Cyclostomata. The Cryptostomata, a dominant and highly diverse group in the Ordovician, commonly had bifoliate colonies (Fig. 1A–D) with zooecia opening only on two opposing surfaces. The zooecia diverged from a median lamina. Other Cryptostomata, grouped as Rhabdomesonata (rhabdomesoids), had small stick-like or twig-like colonies. Their tubular zooecia were arranged in distinctive geometric patterns, e.g., *Ulrichostylus, Arthrostylus,* and *Sceptropora.* Another cryptostome

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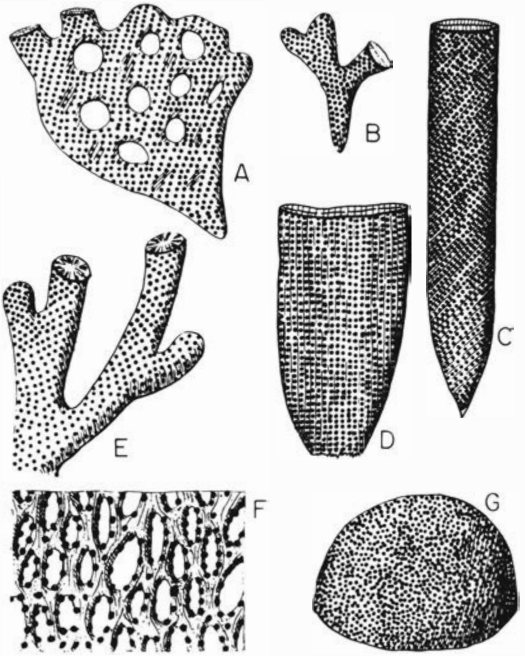


Fig. 1. Forms of bryozoan colonies. An Ordovician genus with the particular colony form is listed. A. Cribrose, Oanduella, X2. B. Strap-shaped, Strictopora, X2. C, Lanceolate, Escharopora, X3. D. Explanate, Phaenopora, X3. E. Ramose, Champlainopora, X2. F. Reticulate, Phylloporina, X10. G. Hemispheric, Prasopora, X2. Colonies in Figs. A to D show bifoliate arrangement of zooecia in cross sectional view. Black dots on colonies represent individual zooecia.

group, the phylloporinids (Fig. 1F) had reticulate colonies with zooecia opening on only one surface. The Cystoporata commonly had dendroid, ramose, laminate, or hemispheric colonies with recrystallized walls forming tubular zooecia. The Trepostomata also had tubular zooecia, many times were polymorphic, and the colonies were hemispherical (Fig. 1G), laminate, ramose (Fig. 1E), or dendroid. The zooecia were commonly crossed by curved or flat platforms. The Cyclostomata, exceedingly rare in the Ordovician, had simple long tubes that were loosely clustered together.

Bryozoans of the order Ctenostomata of the class Gymnolaemata have no skeletons and, therefore, the record of this group is represented by impressions and borings and is very sparse in all parts of the geological record including the Ordovician.

The earliest known bryozoans, represented by rare colonies of Ctenostomata and Cystoporata, are of Early Ordovician (Tremadoc and early Arenig) age. A few fossils from the Cambrian have been erroneously assigned to the Bryozoa. By the middle Arenig, diverse groups of cryptostomes, trepostomes, and cystoporates had broad geographical distributions in carbonate facies on platforms, near shore margins, and open shelves. These groups continued to evolve rapidly in these carbonate facies throughout the Ordovician and they were the dominant group in many assemblages throughout the world, e.g., Middle and Upper Ordovician of the central states of the United States (Ulrich 1890, 1893), of Estonia (Männil 1959), and of the Siberian Platform (Nekhoroshev 1961; Astrova 1965).

Sediment stabilizers, binders, and framebuilders of mounds and bioherms

The evolution in the Early and Middle Ordovician of not only bryozoans but also other calcium carbonate secreting organisms led to significant changes in the structure and species composition of benthic assemblages. Bryozoans were abundant to common in many of these assemblages in the carbonate facies that gradually dveloped. The calcium carbonate secreting organisms, such as pelmatozoans and bryozoans, stabilized, bound, and trapped lime mud, sometimes forming small mounds only a few meters in diameter. Chazyan faunas from upper New York State represent an example of these pioneer assemblages in which bryozoans were widespread in mound and intermound areas (Pitcher 1964; Ross 1981). Other mound deposits with bryozoans as a dominant part of the fauna have been described from Tennessee (Alberstadt *et al.* 1974).

In upper New York State in Chazyan time (Day Point Limestone) (Fig. 2) on the western part of a broad platform near the cratonic margin (Fig. 3), bryozoans, such as ramose colonies of the trepostome *Champlainopora chazyensis*, reticulate colonies of *Phylloporina* sp., and bifoliate strap-shaped blades of the cryptostome *Chazydictya chazyensis*, colonized the shifting sand substrates of the near shore environment. The colonization and establishment of bryozoan assemblages in this unpredictable environment was a chance event and depended on the availability of suitable fragmentary substrate, sometimes pelmatozoan fragments. The bryozoan colonies, fragmented by storm and wave action, made up from 10–20% to 50% by volume of the micritic sediment of mound rock. Some colonies of species such as *Champlainopora chazyensis* also were established in intermound areas where they comprised about 50% by volume of the biosparite. Farther to the east and farther out on the platform, small mounds of 1.5 to 2 m diameter were commonly stabilized and bound by encrustations of the trepostomes *Champlainopora chazyensis* and *Batostoma campensis* and the cystoporate *Ceramoporella* sp. Here the bryozoans made up 20–30% by volume of the micritic rock. Where the mounds were in a nondisrupted environment, *B. campensis* tended to dominate the faunal assemblages.

Higher in the stratigraphic sequence toward the top of the Day Point Limestone, species diversity increased and biohermal buildups developed. The term bioherm is used to describe a carbonate buildup that has topographic relief and the internal composition and framework of which is largely derived from *in situ* production of organisms (Wilson 1975). The same mix of species was present as noted above but here *Batostoma campensis* lacked encrusting growth and was only ramose. Both *B. campensis* and *Champlainopora chazyensis* functioned as framebuilders in the carbonate buildups. Reticulate colonies of *Phylloporina* extended as horizontal growths among the *Batostoma* branches.

The succeeding Crown Point Limestone was also deposited in a platform facies but in quieter, less agitated water. In the western region, closer to the shoreline, isolated thickets of ramose colonies of *Champlainopora chazyensis* and small patches with colonies of *Chazydictya chazyensis, Stictopora fenestrata, Eopachydictya gregaria, Phylloporina* sp., *Nicholsonella* sp. and *Jordanopora heroensis* became established. These were framebuilders and sediment trappers. The bryozoans comprised 20–40% of the biomicritic sediments.

Stromatoporoids commonly dominated bioherms on the eastern part of the platform which was farther from shore. Several cryptostome bryozoans and a few trepostomes were present in bioherms that reached 8–9 m in height. In different bioherms, the bryozoan species were present in different proportions.

Bryozoan carbonate facies of epeiric seas

Later in the Middle Ordovician and subsequently in the Late Ordovician, calcium carbonate secreting organisms dominated the extraordinarily widespread seas of cratonic shelves and shallow basins. Upper New York State and the vast Midcontinent area of North America were two regions in which carbonate facies were widespread and across which successive marine transgressions and regressions occurred.

The carbonate facies and their associated faunal assemblages in upper New York State provide a framework for analyzing Middle and early Late Ordovician bryozoan paleoecology. Bryozoan assemblages were established in a number of depositional sites such as lagoon, near shore, shoal, shallow shelf, open shelf, shelf margin and shallow cratonic basin (Figs. 3, 4) (Ross 1970, 1972). These depositional areas lay to the west of the Adirondack Arch.

In Black Riveran time (Lowville Formation) in upper New York State, bryozoans were restricted in diversity and abundance as a result of the inhospitable lagoonal facies. Where present, they were established in channels and other areas of the platform where water flow was continuous and where colonies were not overcome by lime mud from stagnant turbid waters. Numerous skeletal fragments that had accumulated on the sea floor provided extensive sites for attachment of benthic organisms. The lack of a hard, stabilized substrate was generally no longer a significant limiting factor as it had been in Early Ordovician times.

The accumulating lime muds were not suitable for most bryozoan settlement and growth. There were few species. The strap-shaped cryptostome *Stictopora labyrinthica* was the most abundant and widespread species and large complete colonies were embedded in the lime mud in growth position. This species commonly attached itself to colonies of the tabulate coral *Tetradium.* Variants of this species found at different sites on the platform represent adaptions to particular environmental conditions. Other bryozoan species became established in less silty and less muddy parts of the

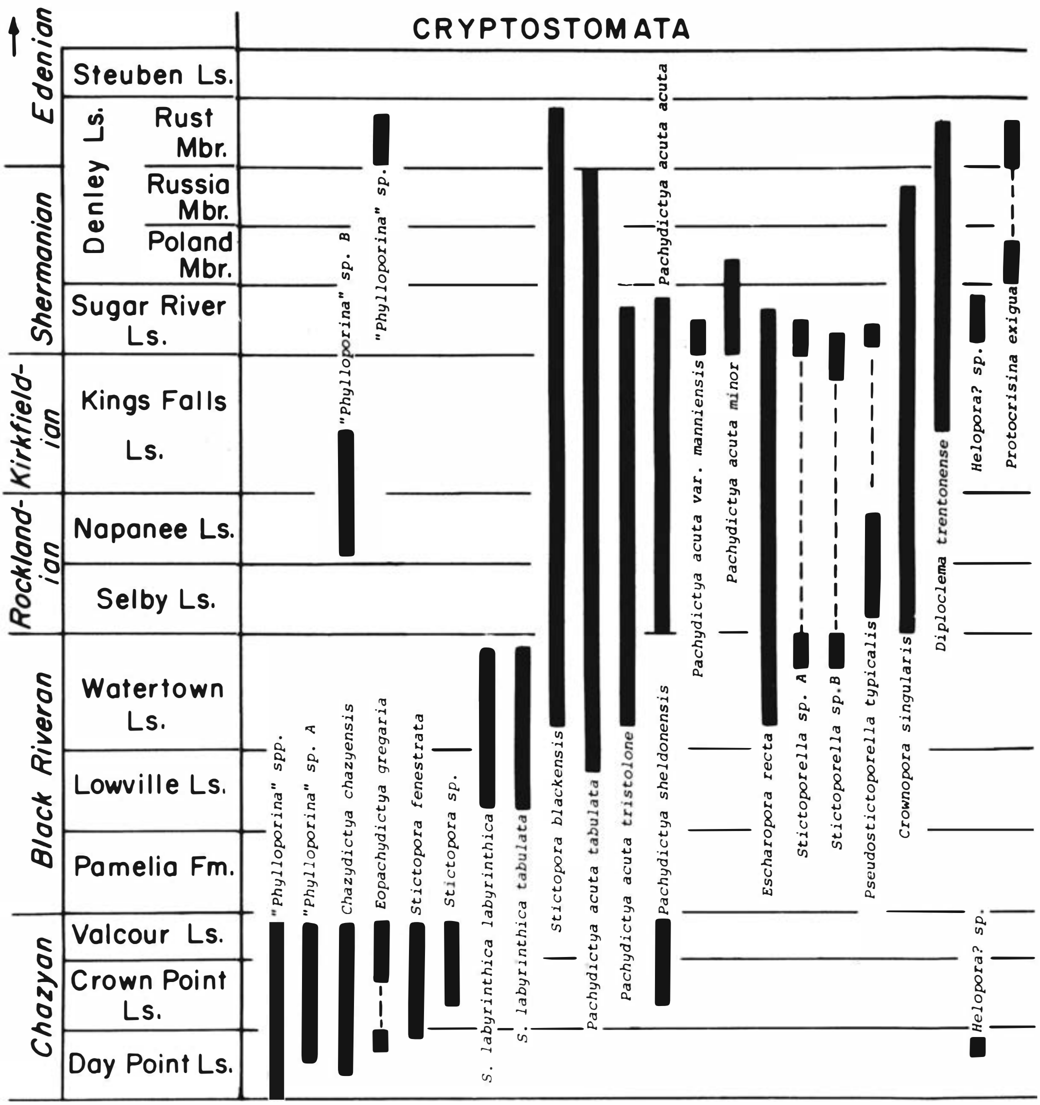
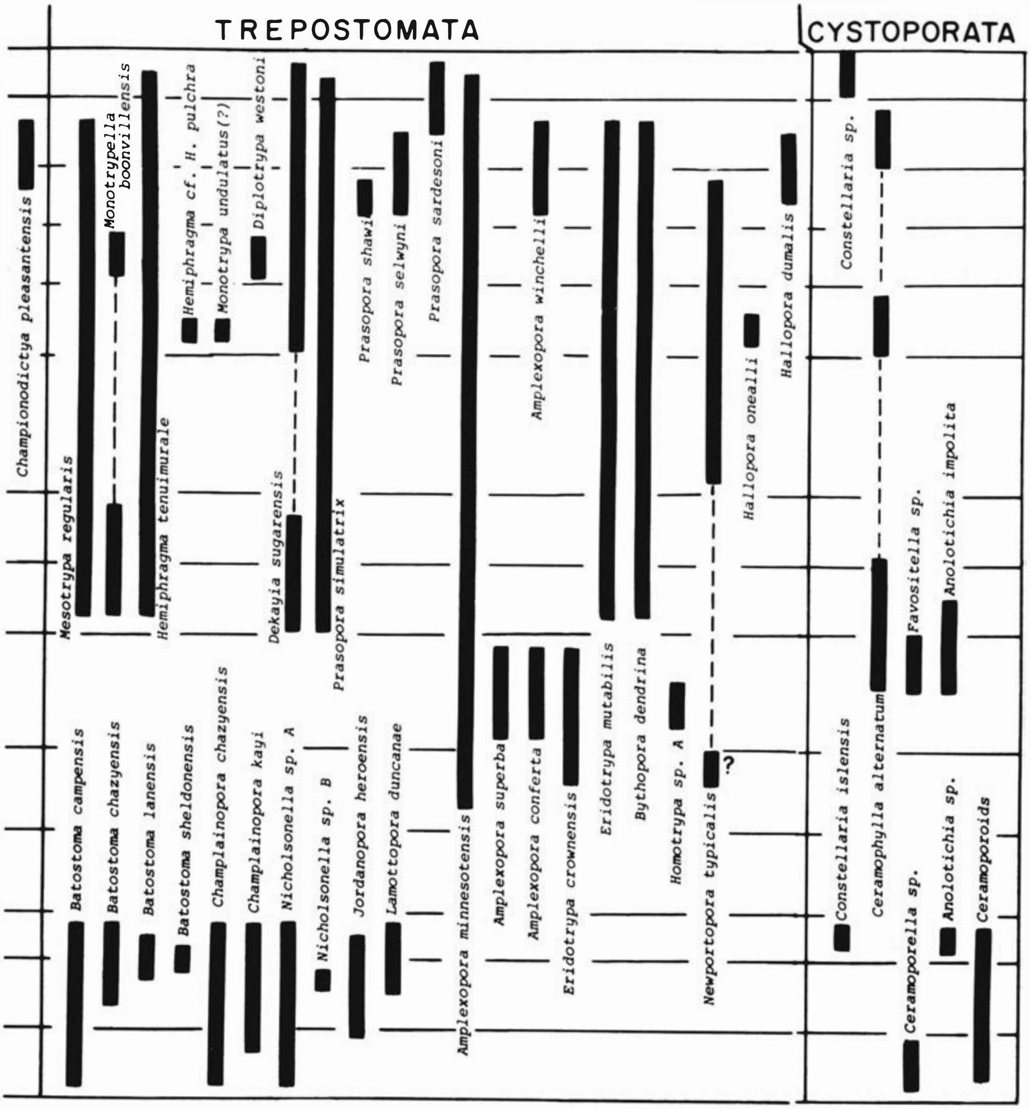


Fig. 2. Stratigraphic range of ectoproct bryozoans in Chazyan through to early Edenian from New York State.

near shore lagoonal area and included the ramose trepostomes *Amplexopora minnesotensis* and *Eridotrypa crownensis* and cryptostomes *Pachydicta acuta* and *Escharopora* sp. Later in Black Riveran time (Watertown Limestone), the lagoonal lime muds had many skeletal fragments and bryozoans had a greater diversity than those in the Lowville Formation. Several cryptostomes and trepostomes were abundant. Cystosporates were sparse.

In succeeding Rocklandian, Kirkfieldian, and early Shermanian times, a major marine transgression from west to east enroached on to the Adirondack Arch. During Shermanian time the Adirondack Arch was submerged and the rising Taconic Highlands to the east of the arch created an influx of clastic sediments onto the shelf and basin to the west, as well as changing the



water circulation patterns. In Rocklandian time (Selby and Napanee Limestones) bryozoan assemblages were established in quiet, near shore environments of the transgressing sea. They formed extensive beds on the sea floor of large unfragmented colonies with a number of species including ramose and dendroid colonies of trepostomes *Monotrypella boonvillensis* and *Bythopora dendrina*, the cryptostomes *Escharopora recta* and *Pachydictya acuta tabulata*, and cystoporates *Ceramophylla alternatum* and the massive ramose *Anolotichia impolita*. A consistent and recurring assemblage included *Bythopora dendrina*, *Eridotrypa mutabilis*, *Dekayia sugarensis*, and *Crownopora singularis*.

In succeeding Kirkfieldian time (King Falls Limestone), bryozoans continued to be dominant in the near shore environment but they

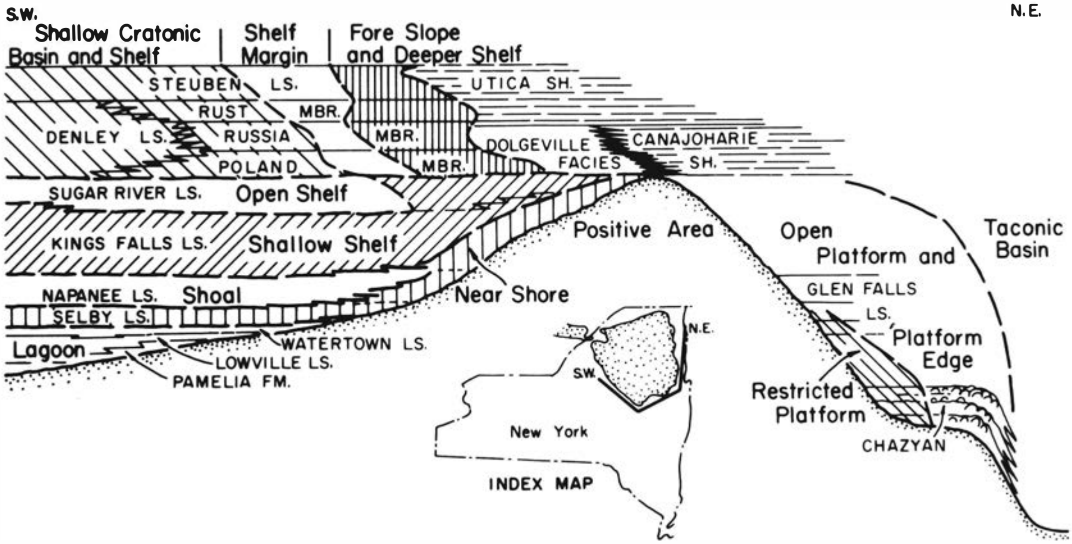


Fig. 3. Stratigraphic cross section along line indicated on index map showing facies relations in Middle Ordovician and early Upper Ordovician in New York State. The index map shows line of section between points marked SW and NE which passes across Adirondack Arch (positive area). Stippled area is Precambrian outcrop of Adirondack Arch.

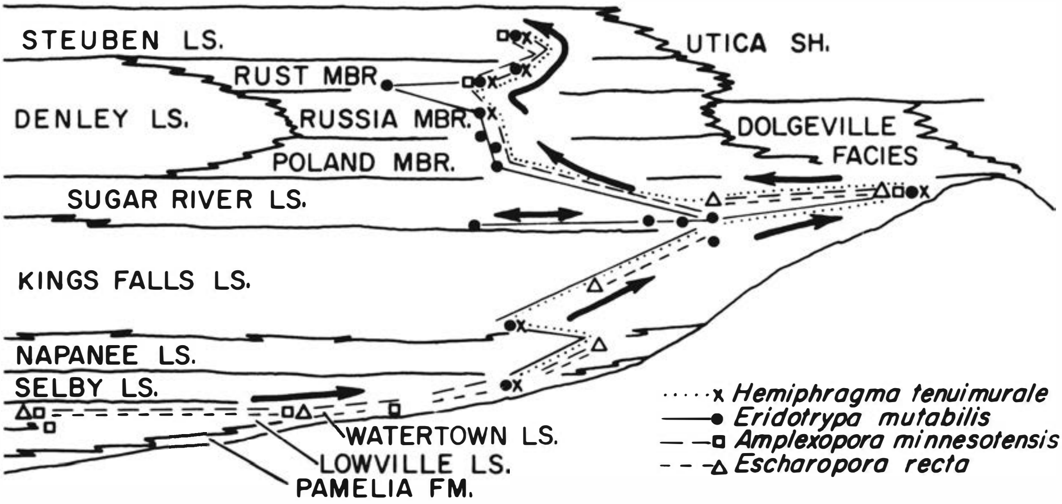


Fig. 4. Stratigraphic cross section showing distribution and migration with facies of four bryozoan species in Middle Ordovician and early Upper Ordovician New York State.

also extended westward across part of the shallow shelf. The benthic faunas continued to form extensive beds on the sea floor.

In early Shermanian time (Sugar River Limestone) as the seas continued to transgress eastward on to the Adirondack Arch, the bryozoan species also continued to migrate eastward (Fig. 4) becoming established in habitats in the near shore, the shallow shelf, and extending out to the open shelf (Fig. 4). However, the continuing influx of clastic material from the rising Taconic Highlands to the east and accompanying modifications in water circulation patterns resulted in a marked change in the bryozoan assemblages. The lanceolate cryptostome *Escharopora recta* and several other cryptostomes disappeared from the assemblages.

Later in Shermanian time (Denley Limestone), other bryozoan species show adaptations to the changing sedimentary patterns. *Prasopora simulatrix* showed marked variability in size and in morphological structures, such as degree of development of acanthopores. In addition, several other species of this genus, namely *P. shawi, P. selwynii,* and *P. sardesoni* appeared. In the later part of Shermanian time (Denley Limestone) a regression took place. This change in the depositional pattern led to a much reduced number of bryozoan species as well as a reduction in the actual number of bryozoans. Another effect on some bryozoans was a reduction in colony size, e.g., *Prasopora shawi* is only a small hemispheric button-like colony.

By the Edenian at the beginning of Late Ordovician time, bryozoans had established themselves as important, and commonly the dominant, substrate colonizers of most carbonate near shore shelf ecosystems. Bryozoans were able to adapt to the wide variety of shelf facies. These evolutionary radiations of bryozoans which are recorded in upper New York State during the Early, Middle, and early Late Ordovician continued into the Late Ordovician epeiric seas. The general community organization that had evolved by the end of the Middle Ordovician remained throughout the rest of the Ordovician. This organization involved a limited number of trophic levels, guild assemblages, niche partitions, and higher taxomonic groups in the ecosystem. Bryozoans are extremely abundant fossils in these strata and many species were of relatively large size so that their proportion of the biomass in many assemblages was relatively high (see Jaanusson 1979).

Summary

During the Ordovician, bryozoans evolved rapidly. They diversified from small, loose or open aggregates of individuals in colonies that encrusted the surfaces of other organisms or fine fragments in lime mud and lime sand into a wide variety of colony forms. The diverse colony forms were adapted to ecological niches that included colonizing of both soft and hard substrate, binding sediment, and forming a framework in and around which other organisms lived. As successful filter feeders, bryozoans adapted by increasing their colony surface and by increasing the kinds of communities in which they were participants.

These invasions into different communities were first strongly evident in the early Middle Ordovician (Chazyan). Several of these adaptions were extremely successful if the abundance of bryozoan fossils is used as an index. Some of the Chazyan communities did not survive into younger parts of the Middle Ordovician. However, new additional communities appeared and into these additional bryozoan radiations gradually became established. The bryozoans were important and dominant participants in most shallow marine carbonate shelves and platforms. In addition to playing an important role as the principal filter feeders in many of these communities, bryozoans became strong competitors for substrate space and some evolved so as to modify substrates by effectively stabilizing and binding sediment.

By the end of Middle Ordovician time, most of the trophic positions for bryozoans had become established. The taxonomic lineages within the bryozoans and the geometry of their zooecial arrangements in colonies had reached the broad limits of their main adaptive diversifications which persisted for the remainder of the Paleozoic.

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Middle-Upper Ordovician shallow platform to deep basin transect, southern Tasmania, Australia

By CLIVE BURRETT, BRYAN STAIT, CHRIS SHARPLES and JOHN LAURIE

There is a profound change in sedimentary environment and in fauna going from the Middle-Upper Ordovician of Ida Bay to Surprise Bay over a present distance of only 25 km. The transition is from peritidal carbonates at Ida Bay, to subtidal carbonates at Precipitous Bluff to deep subtidal calcareous shales at Pt. Cecil to deep water micrites, graptolitic shales and carbonate turbidites at Surprise Bay. The blind (*Nanshanaspis, Bulbaspis*) or large eyed (*Telephina*) trilobites at Surprise Bay suggest sub-photic or twilight depositional conditions and a phosphatic, ironstone hardground enriched in metals and with imploded nautiloids suggests a zone of nutrient-rich upwelling currents at about 300 ± 50 m water depth. Macrofauna from Ida Bay is mainly endemic and is associated with Mid-continent province type or endemic conodonts. The shelf edge sections at Precipitous Bluff and Pt. Cecil contain more widespread macrofossils and Mid-continent conodonts whereas the macrofauna from the deep-water deposits is widespread or cosmopolitan and is associated with North Atlantic province conodonts.

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The Ordovician in Tasmania is chronostratigraphically the most complete sequence in Australia (Banks 1962; Banks & Burrett 1980; Webby *et al.* 1981) and its regional palaeogeographic setting has been reviewed by Webby (1976, 1978). The fossiliferous sequence in the Florentine Valley (Figs. 1, 2) ranges from the Lower Tremadoc through to the Lower Silurian in a conformable succession 2300 m thick (Corbett & Banks 1974). Other important sections occur in southern, northwestern and western Tasmania (Figs. 1, 2). Most sections exhibit a conformable fining-up sequence from siliciclastic conglomerates (often fanglomerates) to quartz arenites (intertidal–high subtidal) to siltstone (mainly subtidal) to limestone (mainly peritidal) up to siliciclastics (subtidal–peritidal).

The tripartite lithostratigraphy is probably best classified at the group level and we recommend that the Denison (dominantly siliciclastic) and Gordon (dominantly carbonate) Limestone Subgroups (of Corbett & Banks 1974 *not* Corbett & Banks 1975) be given the same status as the Eldon Group (dominantly siliciclastic). The Tiger Range Group of Baillie (1979) applied to the siliciclastics overlying the Gordon Group in the Florentine Valley may best be regarded as a subgroup of the Eldon (Fig. 2).

Based on conodont studies it is clear that the base of the Gordon Group is strongly diachronous; being oldest in the east (Middle Arenig in the Florentine Valley), Chazyan at Mole Creek and Blackriveran in the northwest at Queenstown and the Vale of Belvoir (Figs. 1, 2).

The Florentine Valley Formation contains a sequence of trilobite, brachiopod and graptolite faunas ranging from the Lower Tremadoc to the Middle Arenig (Quilty 1971; Laurie 1980; Stait & Laurie 1980; Jell & Stait in prep; Rickards & Stait in prep).

The overlying Gordon Group contains sequences of nautiloids, conodonts, brachiopods, trilobites, stromatoporoids, corals, gastropods, pelecypods, rostroconchs and bryozoa all of which have been the subject of recent intensive

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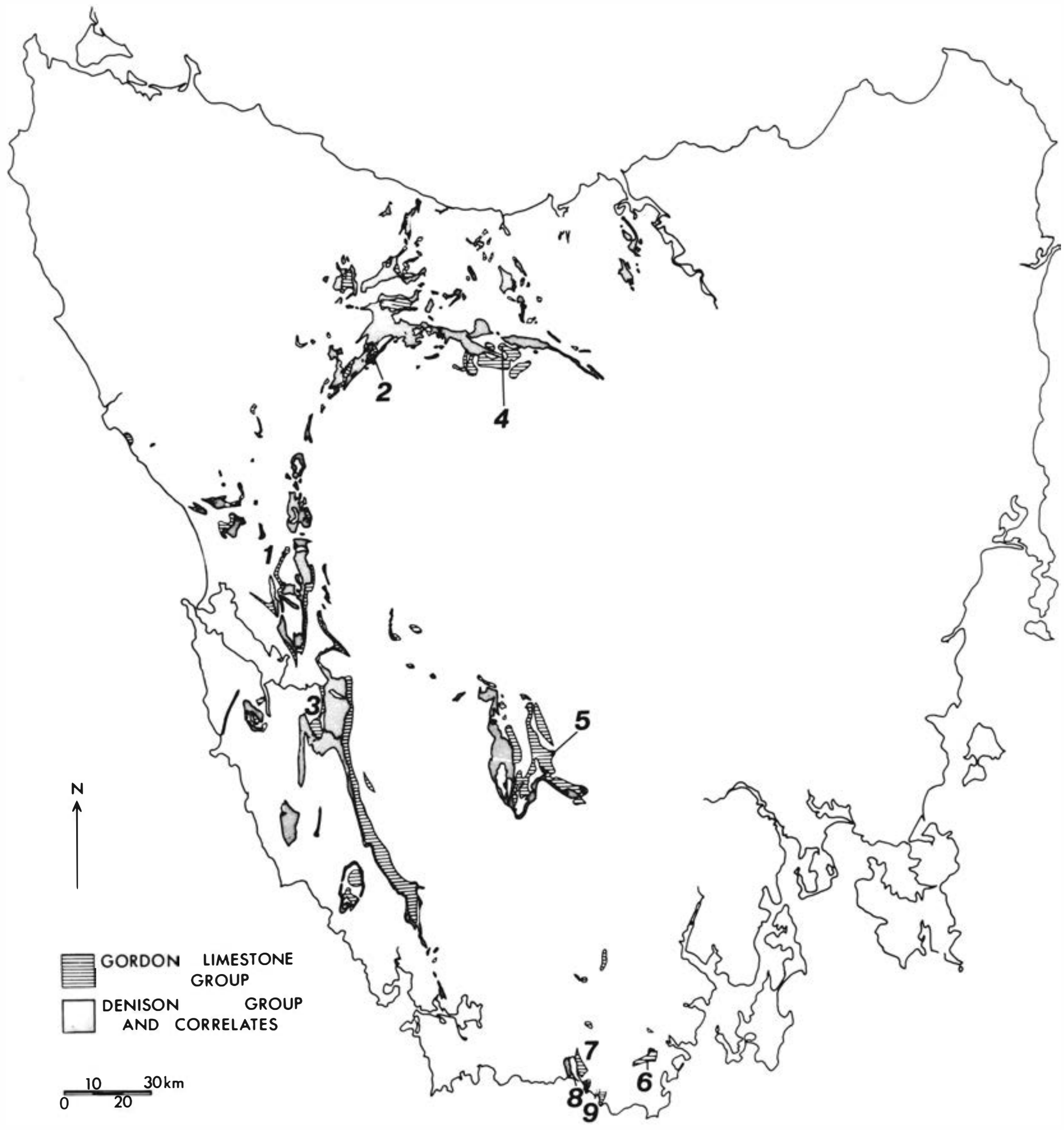

Fig. 1. Locality map of Tasmania showing major sections mentioned in text. 1: Queenstown, 2: Vale of Belvoir, 3: Lower Gordon River, 4: Mole Creek, 5: Florentine Valley, 6: Ida Bay, 7: Precipitous Bluff, 8: Point Cecil, 9: Surprise Bay.

studies. These faunas range in age from Middle Arenig through to Edenian or Maysvillian (Banks & Burrett 1980). The overlying Westfield Beds (the basal formation of the Eldon Group in the Florentine Valley) contains an *Hirnantia* fauna (Laurie 1982) occurring below, with and above a graptolite fauna of the *G. persculptus* or lower *A. acuminatus* zones of the Llandovery (Baillie *et al.* 1978).

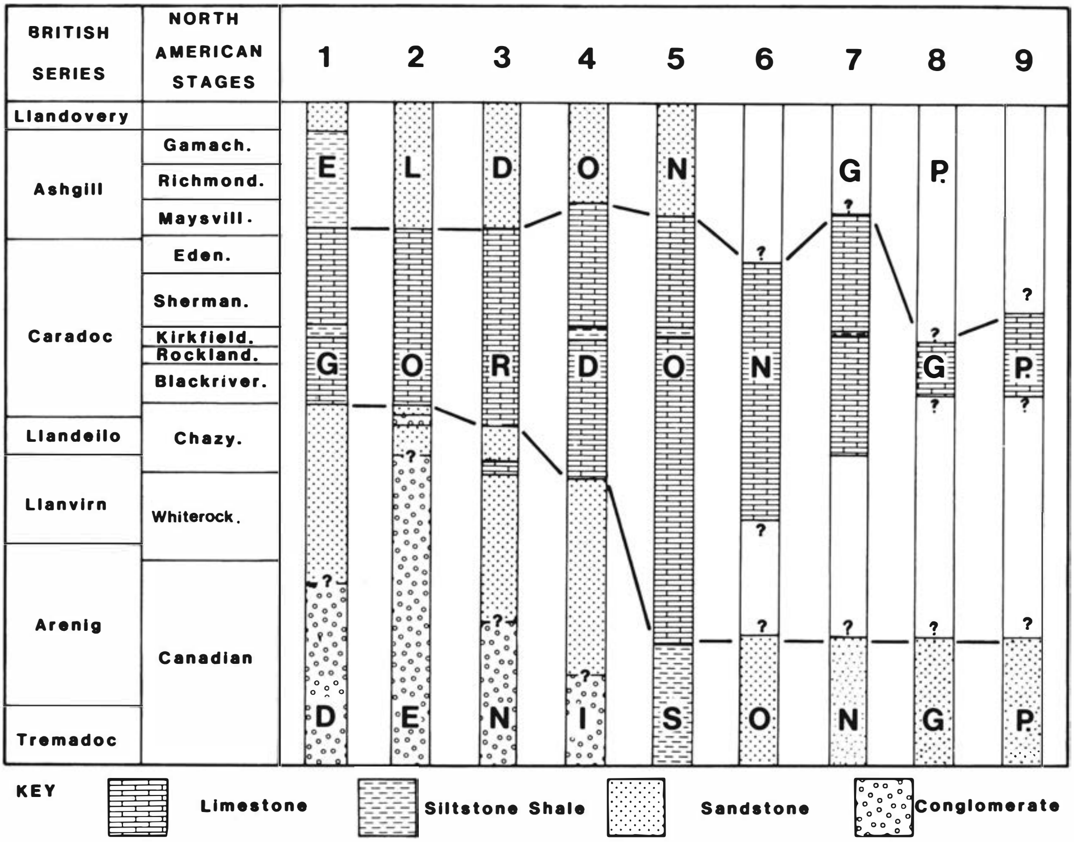

Fig. 2. Simplified stratigraphic columns showing diachronous base of the Gordon Group. Numbers refer to locations marked on Fig. 1.

Ordovician sequences in southern Tasmania

Ida Bay. – Although the limestone section at Ida Bay (Figs. 1, 2) is thrust faulted and incomplete being overlain unconformably by Upper Carboniferous or Lower Permian tillites, it is typical of the mainly peritidal Gordon Group sequences studied elsewhere in Tasmania. Ten lithofacies have been recognised (Fig. 5) ranging from supratidal/or high intertidal to high subtidal. The stratigraphic succession of the lithofacies shown in Fig. 3, indicates that most of the depth changes involve cycles of deposition followed by erosional transgressions in an overall environment of a wide, prograding tidal flat.

Precipitous Bluff and Pt. Cecil. – In contrast to the dominantly peritidal sequences at Ida Bay, the Gordon Group at Precipitous Bluff was deposited in much more open and deeper waters. Burrett *et al.* (1981) have named two conformable lithostratigraphic units. The lower New River Beds consist of 130 m of bryozoan/algal/coralline biosparenites deposited in a high energy subtidal environment. Corals and *Calathium* are sufficiently abundant on many horizons that the designation reef is appropriate. However, thick bush precludes a detailed examination of these beds away from a narrow walking track. The New River Beds are overlain by 230 m of siltstones, calcareous shales and minor biosparites and biomicrites containing a bryozoan/brachiopod/trilobite fauna (the Precipitous Bluff Beds) deposited in a deep subtidal environment.

The Prion Beach Beds at Pt. Cecil, 5 km

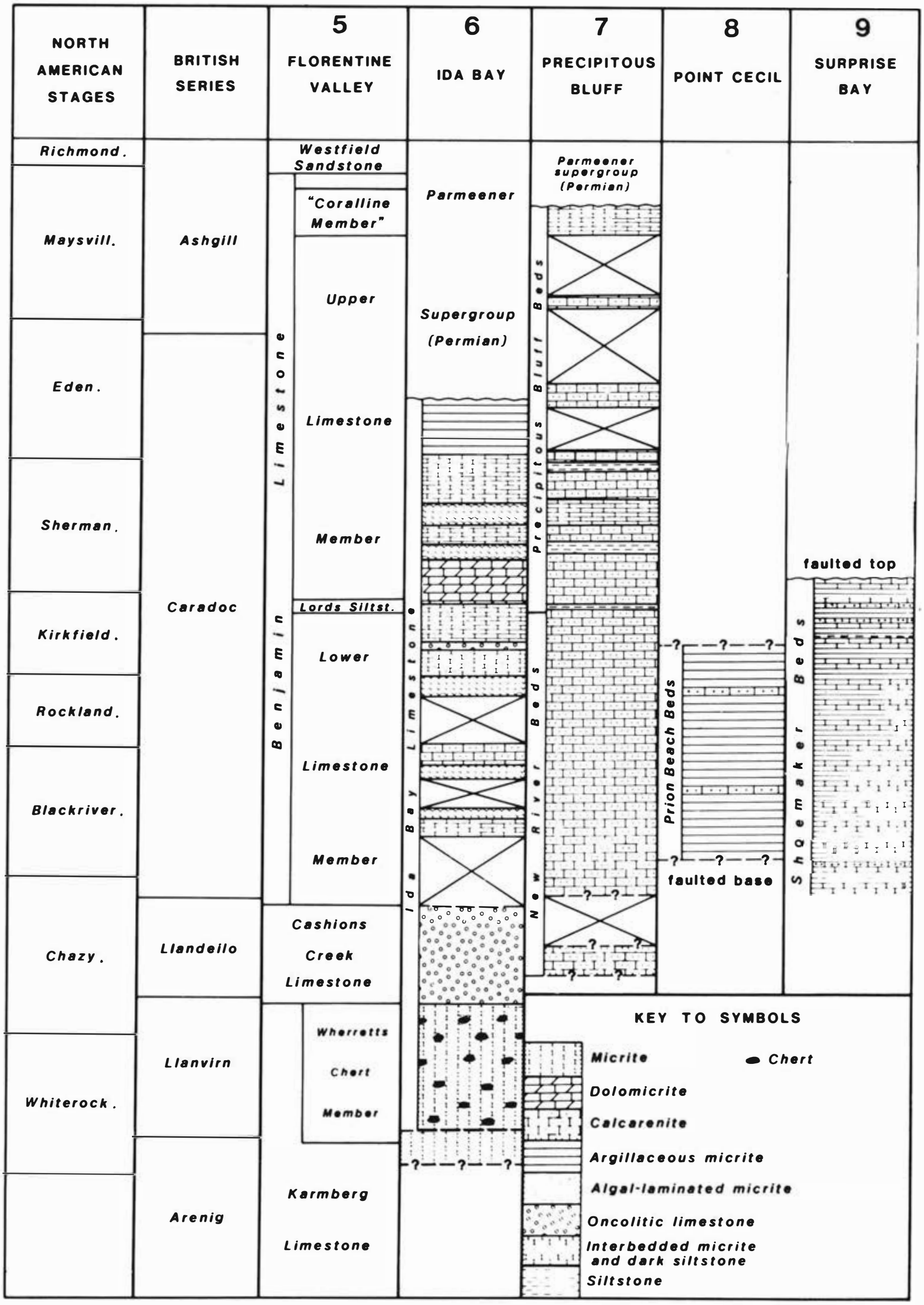

Fig. 3. Simplified stratigraphic columns showing range of carbonate lithologies in southern Tasmanian sections of the Gordon Group. Siltstone (dashed) symbol on Surprise Bay column indicates position of phosphatic hardground. Numbers refer to locations on Fig. 1.

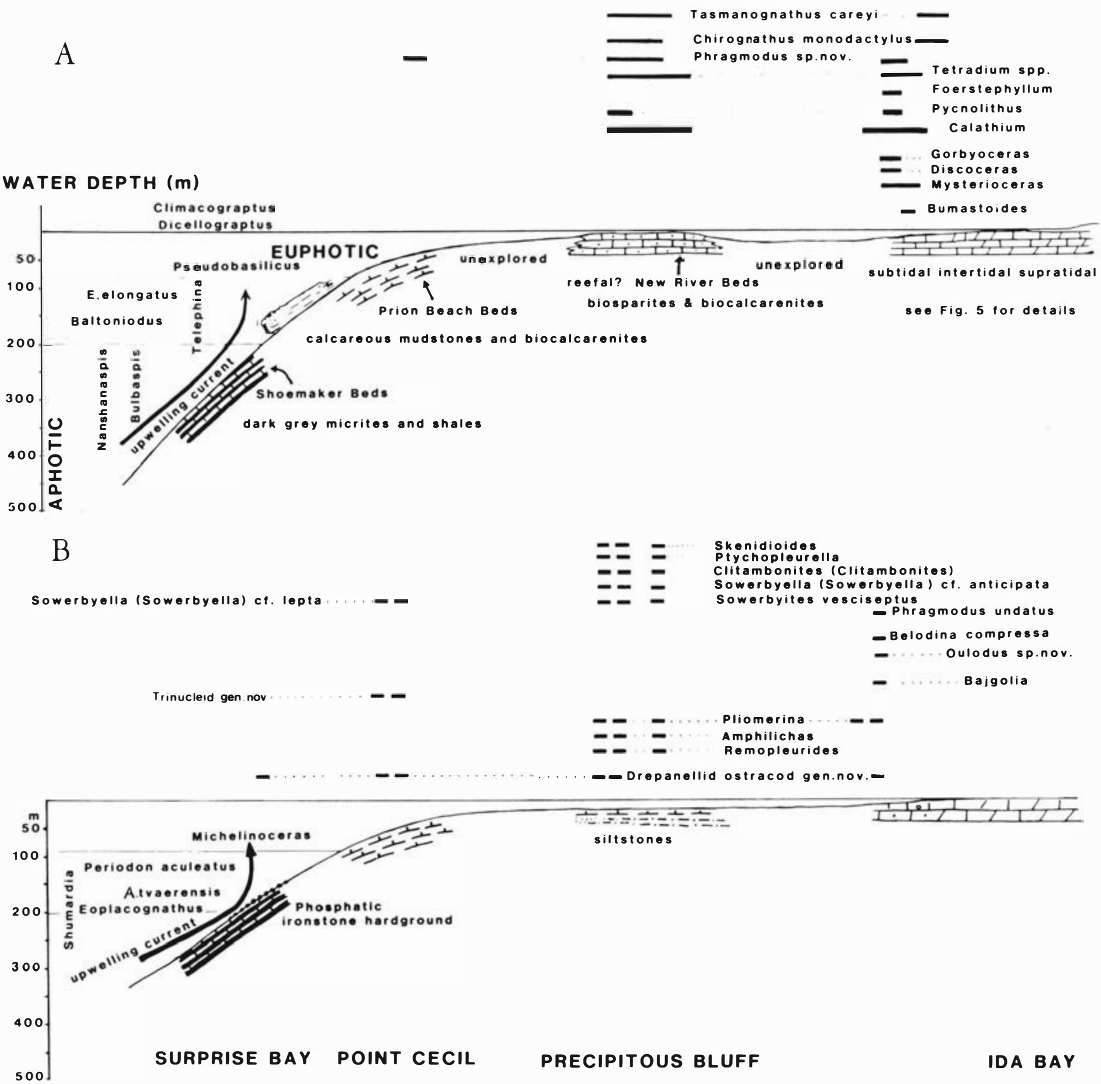

Fig. 4. Generalised environmental model showing transect from Ida Bay to Surprise Bay. Present distance is 25 km. A: Diagrammatic reconstruction for Blackriveran, B: reconstruction for Rocklandian/Kirkfieldian time.

south of Precipitous Bluff, are argillaceous micrites with minor biosparites containing a raphiophorid, a new genus of cryptolithid (Banks 1962, p. 170) and brachiopods (Fig. 4).

Surprise Bay. – The Shoemaker Beds at Surprise Bay are distinctly different from any of the other Ordovician limestones in Tasmania (Burrett *et al.* 1983). They consist of thinly bedded dark grey micrites interbedded with shales, a few lensoidal bodies of biocalcarenite and at least two phosphatic ironstone beds (Figs. 3, 4). About 5% of the micrite beds contain trilobite "swarms". The lower 100 m of section contain *Nanshanaspis murrayi, Bulbaspis* sp., *Telephina (Telephia) twelvetreesi, Pseudobasilicus* sp. and *Nileus* sp. The upper

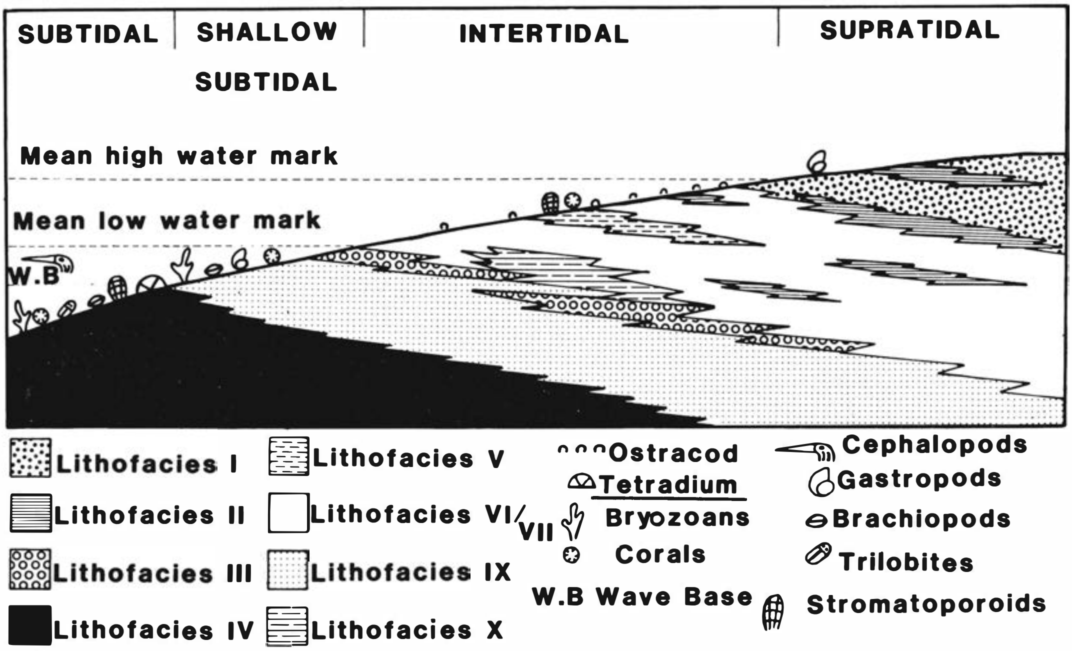

Fig. 5. Environmental distribution of the lithofacies recignised at Ida Bay. Lithofacies I: dimicrites with abundant, irregular spar birdseyes. Fossils rare except Hormotoma*-supratidal or upper intertidal. Lithofacies II: pelmicrites and rarely micrites with abundant laminae (probably algal) of dark bituminous material, abundant dolomite and minor aeolian quartz grains. Well developed mudcracks and rare gastropods. Protected, dessicating environment – peritidal. Lithofacies III: oncolitic limestones between subtidal and intertidal horizons. Oncolites decrease in size towards the subtidal going from 30–40 mm diameter in a dolomitic microspar down to 5–8 mm in a poorly washed intrabiospelsparite matrix. Wave agitated, low intertidal or upper subtidal environment. Lithofacies IV: Biomicrites with dolomitised worm burrows and diverse biota of corals, bryozoa, gastropods, nautiloids, calcareous algae, crinoids and stromatoporoids.* Tetradium *dominated band may represent "wave baffle community" of Walker (1972). Lithofacies V: horizontally burrowed intrapelsparites with minor oosparites. Fauna dominated by large stromatoporoids,* Foerstephyllum *and* Hecatoceras. *This may be analogous to the coeval* Stromatocerium/Foerstephyllum/Antinoceras *community of Walker (1972). Low intertidal, subject to intermittent wave action and crossed by tidal channels. Lithofacies VI: micrites and biomicrites with thin, irregular, dolomitic layers, interbedded with intraclastic breccias and graded calcarenites resting on scoured surfaces. Vertical burrows, probable evaporite pseudomorphs, rare birdseyes and mudcracks are present. Ostracods abundant. Dessicated upper-intertidal environment. Lithofacies VII: is similar to VI but lacks dessication features and contains greater abundance of unbroken fossils. Lower intertidal environment. Lithofacies VIII: is volumetrically minor, rests above Lithofacies II and below I. Micrites with tabular, 1–2 mm diameter birdseyes and abundant vertical worm burrows. Upper intertidal/lower supratidal environment. Lithofacies IX: intrasparites and intrabiosparites with micrite intraclasts several centimetres long. Brachiopods, trilobites, gastropods, bryozoa,* Solenopora, *solitary corals and crinoid debris. Energetic subtidal environment. Lithofacies X: biosparites and biomicrites with silt layers. Micrite beds are sparsely fossiliferous but may contain worm burrows. Most fossils are confined to graded, densely packed coquinae in a spar matrix. Silt bands consist of angular quartz in an argillaceous and/or bituminous groundmass. The silt bands are thought to represent changes in terrigenous input from the source area. Low intertidal environment.*

100 m contains abundant specimens of *Shumardia forbesi*. Of these trilobites *Nanshanaspis*, *Bulbaspis* and *Shumardia* are blind and *Telephina* has extremely large eyes. By analogy with modern marine isopods (Menzies *et al.* 1973; Taylor & Forester 1979) it is probable that the blind forms inhabited depths below the photic zone. The very large eyes of *Telephina* are most reasonably considered adaptations to the disphotic (twilight) zone (i.e. about 150–200 m

in the tropics). In the early–Middle Ordovician, *Nileus* is characteristic of the Nileid Community of Fortey (1975) which is thought to have inhabited the upper slope of the North American Block (Shaw & Fortey 1977).

Several 0.5–1 m thick, graded, lensoidal biocalcarenites with minor cross-bedding at their tops, are probably carbonate turbidites and are similar to modern examples described by Van Tassell (1980).

The lowest of these turbidites contains a North Atlantic Province conodont fauna (*Periodon aculeatus, Protopanderodus varicostatus, Eoplacognathus* and *Baltoniodus gerdae*) whereas higher calcarenites yield *Amorphognatus tvaerensis* and *Protopanderodus liripipus.* None of these genera or coprovincial forms are found in Gordon Group sections of Middle-Ordovician age which are dominated by either Midcontinent Province species or endemic species morphologically similar to Midcontinent genera and species.

A 4 cm thick phosphatic, ironstone bed marks a lithological and biological change in the Shoemaker Beds. Below this bed there is generally no sessile benthos except to within 10 m of the bed where sponges are present. Above the ironstone, bryozoans and articulate brachiopods occur and nautiloids and gastropods are common within the ironstone. The overlying micrites are much lighter than those below which, with the faunal evidence, suggests a transition into a more oxygenated zone. By analogy with modern basins this transition may have occurred at about 250 m water depth (Rhoads & Morse 1971). The ironstone consists of pyrite and collophane with a nodular, oxidised (limonitised), bored crust suggesting that it was a submarine hardground. The ironstone is considerably enriched in Cu, Zn, Pb and Ba (but not Mn, Os and Ir) relative to the bulk of the Shoemaker Beds suggesting that the bed formed in a zone of nutrient rich, plankton rich (and hence trace element rich, Berry 1981) upwelling near a basin margin.

The large numbers of *Michelinoceras* sp. in the hardground perhaps fed on the profuse plankton in this upwelling zone. Most of the *Michelinoceras* specimens are completely broken but a few have their distal camerae crushed in. One specimen is proximally embedded in the ironstone but the crushed distal camerae do not contain any of the overlying shale indicating that the crushing is more likely due to implosion than to overburden pressure or predation. Using the formula of Westermann (1973) which relates septal thickness and curvature to implosion depth, a depositional depth of 300 ± 50 m is suggested.

Nodular phosphates are currently forming at similar depths in areas of upwelling along the East Australian shelf margin (Kress & Veeh 1980) and similar sulphide/ironstone beds have been reported from basinal carbonates of the Ordovician of the Appalachians (Read 1980).

The hardground thus forms a useful, absolute palaeodepth datum within the Shoemaker Beds of 300 ± 50 m. The change to oxygenated conditions above this level may have been caused by a basinal uplift, or by a shift in the zone of upwelling or by a downward expansion of the oxygen-rich zone in the manner suggested by Fischer & Arthur (1977).

Conclusions

During the Blackriveran-Kirkfieldian in southern Tasmania there is a transition from tidal flats, to outer platform reefs, to platform marginal calcareous shales to deep pelagic carbonates. This transition fits the general model of Wilson (1975). Similar transects have recently been documented from the Upper Cambrian of Nevada (Cook & Taylor 1976), from the Lower Ordovician of Spitzbergen (Fortey 1975; Fortey & Barnes 1977), and the Middle Ordovician of Texas (Bergström 1978; King 1977), Virginia (Read 1980), and the Mackenzie Mountains (Ludvigsen 1978). Tasmania appears to have been a mini-platform compared with the North American, Siberian and Australian blocks and the profound sedimentological contrasts occur over a relatively short distance (< 25 km) compared with several hundred on the major platforms. The faunal contrasts are also profound. The peritidal sections supported a strongly endemic macrofauna with a Midcontinent province type conodont fauna. A similar fauna but with elements found elsewhere in Australia occurs in the platform margin whilst the deep water facies contain a totally different macrofauna with cosmopolitan or "Chinese" affinities and a North Atlantic Province conodont fauna. Only one faunal element (a new drepanellid ostracod genus) is found across the whole transect but this is stratigraphically long-ranging and it remains impossible to correlate directly between Surprise Bay and the rest of Tasmania. However, it seems possible that the shallowing exhibited in the top part of the Shoemaker Beds at Surprise Bay and the deepening exhibited at Precipitous Bluff (and indeed in the Florentine Valley) happened isochronously at about the Kirkfieldian suggesting a substantial decrease in depositional slope over a wide area of southern and central Tasmania. Using Read's (1982) terminology this is a transition from a rimmed shelf to a ramp which caused or coincided with a profound change in lithology and fauna across the whole of the Tasmanian platform.

Acknowledgements

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Palaeoecology of graptolitic black shales

By S. HENRY WILLIAMS and R. BARRIE RICKARDS

Lower Palaeozoic black shales were deposited in a variety of environments, including open oceans and shallow, near-shore areas. Some resulted from relatively rapid turbidite deposition, while others were formed slowly as a soft, anoxic ooze. Bioturbation is occasionally present, while associated paler lithologies sometimes contain dark "flakes" which were flocculated organic material or "rip-up clasts" of unconsolidated sediment. Detailed study of deep-water sequences reveals rapid lateral variation in thickness, implying an undulating sea floor in abyssal environments. Currents during the deposition of black shale are often indicated by aligned graptolites. These may also have winnowed unconsolidated graptolitic sediment to form laminae with closely packed, uniformly sized rhabdosomes. This offers an alternative to the commonly preferred explanation of "mass mortalities" which would result in a variably sized assemblage, although this can also often be established. While changes in sea level were responsible for major lithological changes, small-scale lithological alternations were probably related to fluctuations in oxygen levels controlled by current strength and density of organic material.

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A substantial literature exists on black shale of the Upper Paleozoic and of the Mesozoic to Recent, which has been greatly boosted by the economic potential of such sediments as hydrocarbon source rocks (see Dunham 1961; Schlanger & Jenkins 1976). The palaeoecology of these has been studied in great detail, most of them being considered to represent sedimentation in enclosed, oxygen starved basins up to several hundred kilometers wide (Hallam & Bradshaw 1979; Deggens & Stoffer 1980; Schlanger & Jenkins 1976). Although some Lower Palaeozoic black shales were no doubt formed in such environments, many are integrally related to shallow water and sub-aerial deposits, such as the Upper Ordovician of the Oslo Region (Brenchley & Newall 1980) and the Snowdon Volcanic Group of North Wales (Fitch 1967), while others are of great lateral extent and must have been deposited in open oceans (Fig. 1). Perhaps the best known example of this type is the Moffat Shale Group of southern Scotland which is now considered to have been deposited in the Lower Palaeozoic Iapetus Ocean and subsequently formed into a series of imbricate thrust slices above a subduction zone (McKerrow *et al.* 1977). No open ocean black shale environments exist at the present time and it has been suggested (Berry & Wilde 1978) that such conditions may only exist when ocean circulation is reduced. This was probably the case in the early Palaeozoic when the climate was relatively uniform and there was no ice cap in the northern Polar area. During the late Ordovician, glacial conditions clearly existed in Gondwanaland, which have been most frequently documented from North Africa (e.g. Beuf *et al.* 1966; Bennacef *et al.* 1971). This led to widespread regression and probably increased oceanic circulation. Destombes (1976, and following discussion) considered the glaciation to have begun in the Hirnantian and possibly to have extended into "lowest Silurian", depending on the position relative to the South Pole.

If present day sequences are compared, black shales are seen to accumulate under anoxic conditions with a slow "rain" of fine sedi-

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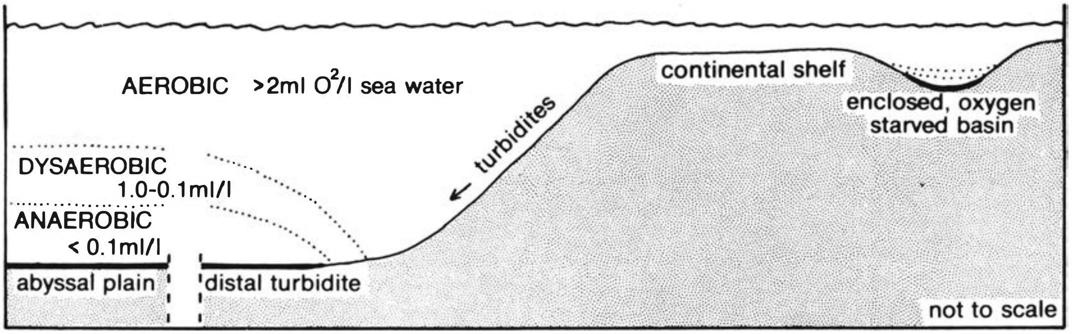

Fig. 1. Idealised profile showing possible sites of formation of graptolitic shale. Names and definitions of oxygen levels from Rhoads & Morse (1971).

ment. The black colour is often, but not always, due partly to the presence of pyrite rather than carbon, which rarely exceeds a few percent (Hallam 1980). Determination of the carbon origin by geochemical means is difficult or impossible in Lower Palaeozoic British rocks due to subsequent alteration (A. Thickpenny, pers. comm.). Byres (1979) described general and detailed sections from two black shale sequences in the Upper Devonian and Upper Cretaceous of the United States, both of which were apparently formed in enclosed basins up to 150 m deep. He summarises ideas on the formation of laminated muds in Recent sediments, concluding that although most laminated fine silts and shales are products of low energy environments, some may have been deposited by relatively rapid turbidity currents, such as the Phyllopod Bed of the Burgess Shale which has clearly graded units (Piper 1972). The lamination of black shales, caused by variation in sediment supply, is destroyed or partially destroyed (Rickards 1964) when bioturbation is present. Normal oxygenation in the ocean surface water is about 7 ml O^2 /l sea water; this can be lowered to about 2 ml/l with little effect on marine communities (Byres 1979). The terms ”dysaerobic” and ”anaerobic” were defined by Rhoads & Morse (1971); dysaerobic conditions occur with 0.1–1.0 ml/l which can support a simple infauna such as small polychaetes and aschelminths but no shelly fauna, while anaerobic conditions, below 0.1 ml/l, are toxic to all life. Although sediments formed under dysaerobic conditions are partially bioturbated the burrows are only up to 2 mm diameter and lamination is commonly preserved.

Rickards (1964) described and illustrated sectioned slabs from the Silurian of the Howgill Fells, northern England, which he considered to have been deposited by distal turbidites, the evidence being graded bedding on a microscopic scale and lateral passage into proximal turbidites. The lithologies were later studied briefly by Piper (1975); he gave no conclusive suggestions on the formation of black graptolite mudstone but suggested that graptolite preservation was favoured by rapid burial. Rickards (1964, text-fig. 2) clearly showed that the graptolitic mudstone was soft at time of deposition and that current strength was low, due to the existence of rare upright orthocone cephalopods. These occur rarely in graptolitic shales of both Wenlock and Ludlow age in northern England and of Ashgill age in the Oslo Region. Rickards also showed that the change from black to pale lithology was a simple one with few sedimentary changes and almost no change in total mineralogy, save for extra S, Fe and As in the black shales (see also Spencer 1966). He concluded that increase in current strength caused aeration of bottom waters, decreasing the preservation of carbon and primary pyrite and permitting the existence of benthic organisms. He also noted, both in 1964 and later (Rickards 1978; Ingham & Rickards 1974), that the presence of narrow carbonaceous

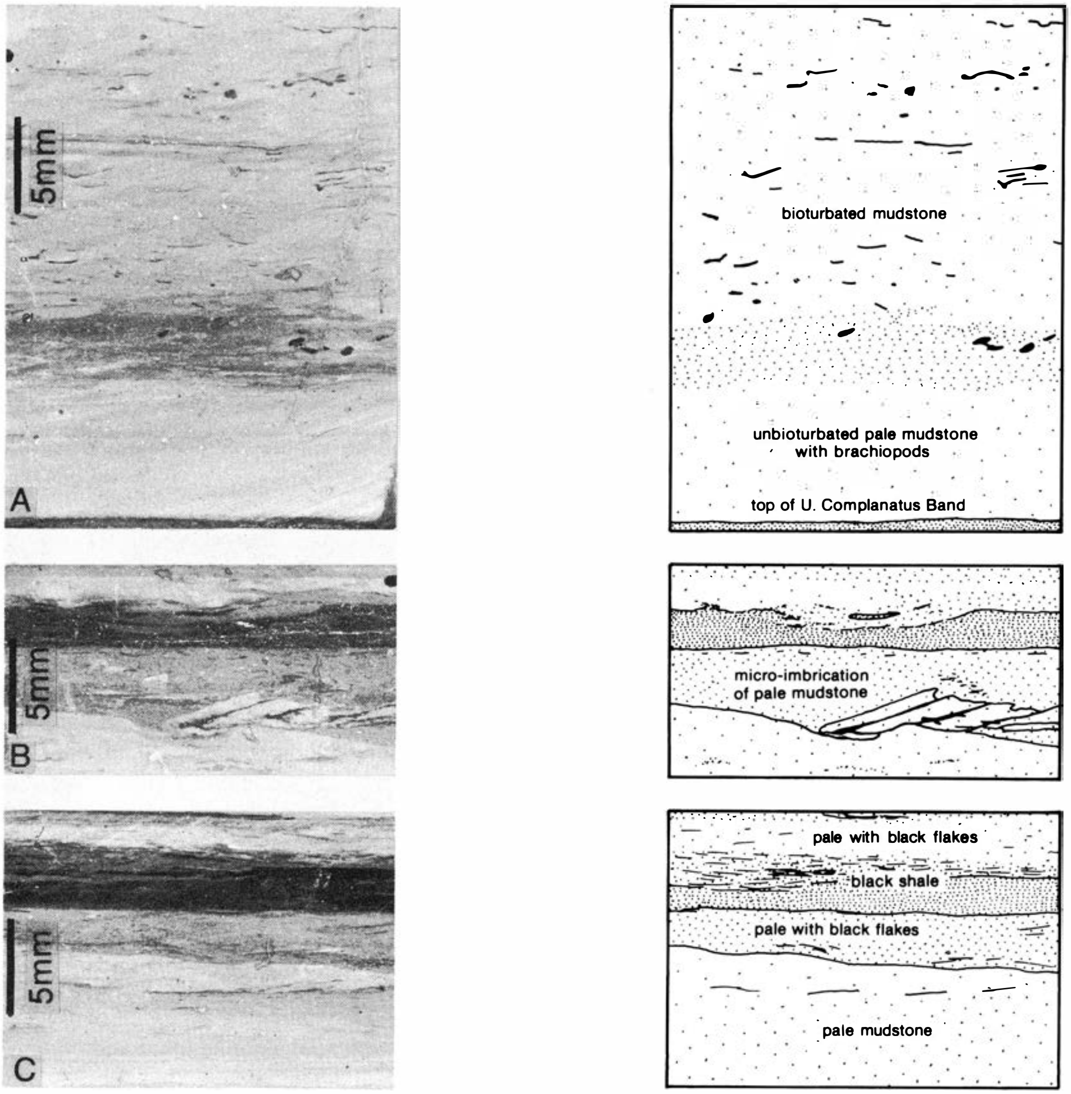

Fig. 2. Sectioned slabs from the Upper Hartfell Shale at Dob's Linn with explanations. A. Strata above the upper Complanatus Band. B. Micro-imbrication just below the lower Complanatus Band. C. Strata at the same level as B, showing black flakes above and below a thin black shale band.

bands in an otherwise pale sequence was due to increased influxes of carbonaceous material (also see Wilson 1954) from a planktonic source. This caused anoxic bottom conditions with a corresponding increase in pyrite formation and preservation of graptolites.

Detailed work at Dob's Linn, southern Scotland, shows that the lithological reversals in the Upper Hartfell and low Birkhill Shale Formations are to some extent characterised by different features from those described by Rickards (1964), although there are certainly some similarities in the Llandovery black shales in the *atavus* to *argenteus* levels. The pale grey mudstone is occasionally bioturbated by horizontal burrows a few millimetres in diameter (Fig. 2A), but these are never present in the black shale. Rickards (1964) considered that black shale lamination was due to variation in density of carbonaceous material and that bioturbation

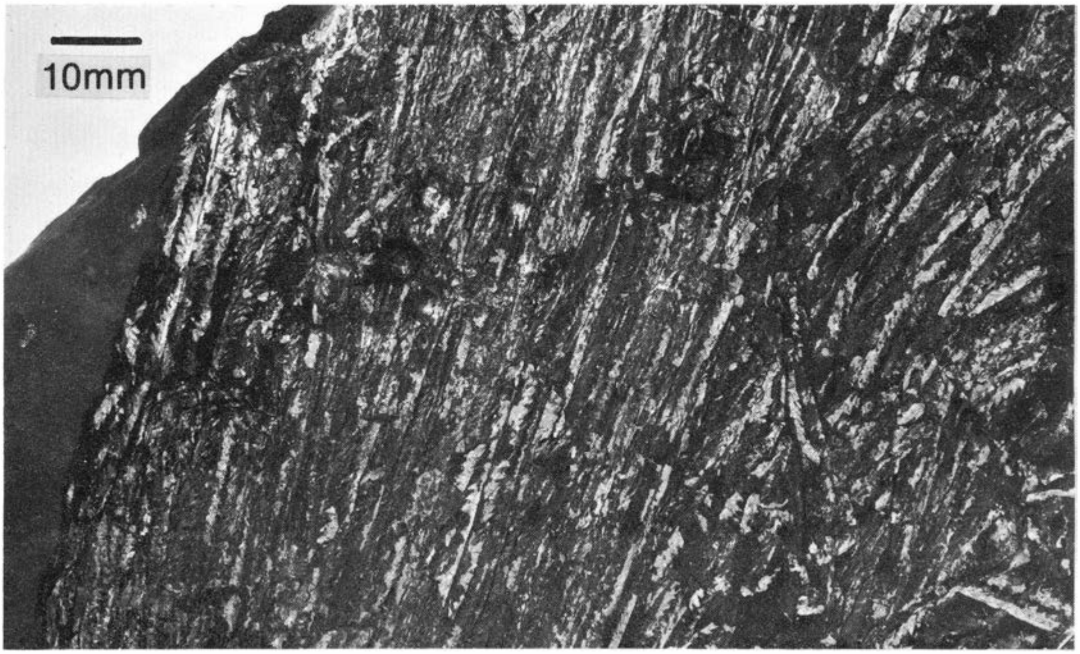

Fig. 3. Lamina of winnowed and aligned extensiform didymograptids with almost no sediment. Lower Didymograptus Shale (Lower Ordovician), Tøyen underground station excavations, Oslo. Coll. B.-D. Erdtmann. Pal. Mus. Oslo, PMO 109.141.

destroyed banding, rather than creating it by the compression of faecal pellets as suggested by Jones (1954). This conclusion was borne out by Byres (1979). It therefore seems likely that the black graptolitic bands of the Upper Hartfell Shale represent anoxic conditions, while the paler mudstones with occasional small, horizontal burrows were deposited under dysaerobic conditions. In addition, the lithological boundaries are commonly complex with scattered flakes of black material at both the upper and lower boundaries with the pale mudstone (Fig. 2B); these also occur in the *sedgwickii* Zone in the Lake District. Micro-tectonic, syn-sedimentary structures also occasionally occur at these boundaries (Fig. 2C). The micro-faulting and micro-imbrication probably indicate unstable deposition on a gentle slope, although some syn-sedimentary micro-faulting may be due to dewatering effects. There are two possible explanations for the black flakes present in pale mudstone; either they are remnants of scattered carbonaceous material which drifted over the area but was insufficiently dense to cause totally anoxic conditions, or they represent syn-depositional erosion of black material which has been reburied soon after to prevent decomposition. In the Howgill Fells the latter certainly applies in the *maximus* Subzone where black flakes up to several centimetres across themselves yield *R. maximus*, yet occur with thin beds of black shale also yelding *R. maximus.*

The Upper Ordovician Lower Tretaspis Shale of the Oslo Region, Norway demonstrates lithologies intermediate between true black shale and dark grey mudstones and silts (Williams & Bruton, in press). The base of the Lower Tretaspis Shale, which is marked by a phosphorite, represents a deepening of the sedimentary basin and gives a sharp contrast with the underlying limestone-shale alternations of the Upper Chasmops Limestone. Bioturbation is present throughout much of the unit and is locally coarse with burrows up to 10 mm in diameter. The finer bioturbation is similar to that recorded by Rickards (1964) from black shales overlain by non-carbonaceous material with rare trilobites and brachiopods. The bioturbation figured by him (Rickards 1964, pl. 16) appears close to that illustrated by Piper & Brisco (1975, fig. 12a) from a Tertiary abyssal plain mud core taken during DSDP research. The presence of bioturbation need not, however, indicate oxygenerated conditions in the black shale as present day organisms may burrow for some depth into anoxic sediment. Occasional darker bands occur throughout the paler sequence in the upper part of the Lower Tretaspis Shale, several possessing carbonaceous flakes at their bases. These clearly represent the situation where carbonaceous material was insufficiently dense to cause anoxic conditions or deposition of true black shale. The presence of a relatively abundant shelly fauna in both the dark shale and nodular limestones, together with increased bioturbation and limestone deposition towards the top, clearly indicates formation of a dark, carbonaceous shale in a relatively shallow and shallowing, near shore environment. Rickards (1964: 422–3) considered that the presence of pyritised graptolites preserved in relief indicated more fully anoxic conditions than those which are preserved in the flattened state. Work in the Oslo Region clearly indicates, however, that this hypothesis is not entirely correct as the most highly pyritised and best preserved three-dimensional specimens occur in the higher, paler lithologies of the Lower Tretaspis Shale, often associated with dense bioturbation.

Evidence such as graptolites cutting bedding (Briggs & Williams 1981) support earlier conclusions that the black shale was very soft during deposition. It is possible that there was no true sediment-water interface but merely a transition zone, in which case syn-depositional erosion seems unlikely in some cases. However, evidence of current activity during black shale deposition is given by aligned graptolites at many localities (Fig. 3); at Dob's Linn these are especially common in the Anceps Bands of the Upper Hartfell Shale. Further indications of current activity and a possible irregular sea bed are shown by rapid lateral variation in metabentonites and thicknesses of pale shale units. Contrary to what would be expected, the thin metabentonites at Dob's Linn are rarely laterally continuous for any distance. This is well illustrated by the lower Complanatus Band which contains a metabentonite near the top in the Main Cliff section, although this is absent while the thickness of intervening pale grey mudstone increases in the Linn Branch section some 70 cm along strike. The thickest Anceps Band, B, is also affected by such lateral change; it is only 17 m thick on the Main Cliff but increases to 19 cm some 120 m along strike in the Linn Branch section (Williams 1982). In [illegible] thin metabentonite is present towards the top of Band B in the Linn Branch which is absent on the Main Cliff, while the top boundary of the black shale in the Linn Branch is gradational although on the Main Cliff it is sharp with a thin orange lamina directly below. It is possible that this orange band and the overlying black lamina are greatly thinned (eroded?) representatives of the metabentonite and overlying black shale in the Linn Branch. In the Lake District Hemsley (pers. comm.), has shown that bentonites can be reworked and even cut out altogether despite the generally low energy environment.

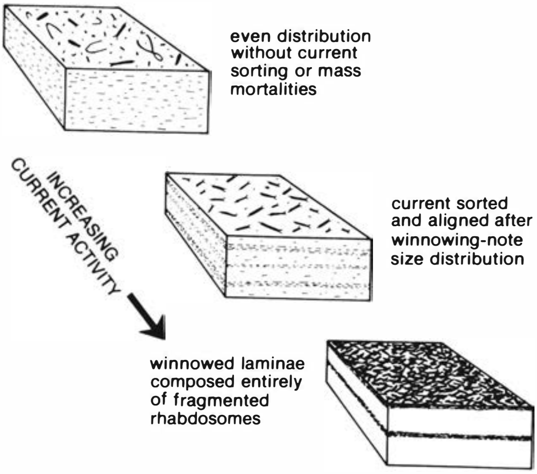

FIg. 4. Postulated effects of current activity of graptolitic shale with syn-depositional erosion and redeposition following winnowing.

There is no doubt that further detailed study of the black bands in the Upper Hartfell Shale would reveal additional changes similar to the great lateral variation in thickness of the Anceps Bands recorded by Williams (1982, text-fig. 3). This variation is original and represents either non-deposition or syn-sedimentary ero-

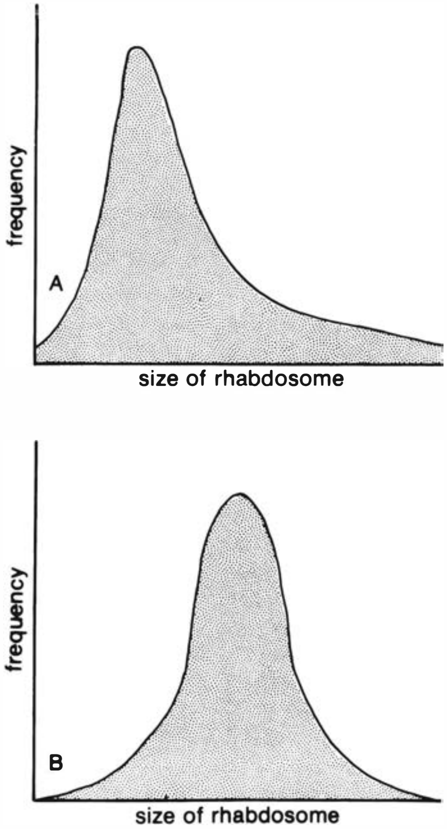

Fig. 5. Qualitative graphs showing: A. Expected size distribution of graptolite rhabdosomes (cf. Curry 1982 for Recent brachiopods). B. Distribution commonly found on bedding laminae, due to current sorting of rhabdosomes.

sion. In the Howgill Fells (Rickards 1970, Ingham & Rickards 1974) syn-sedimentary erosion may well have winnowed thicker black shale sequences, distinctly more pyritous, into hollows on the sea floor, and the positions of these have been accurately plotted. The second hypothesis is therefore preferred as this would also explain the distribution of evenly-sized, current orientated rhabdosomes of two or three species on certain laminae and other bedding planes consisting entirely of fragmented rhabdosomes with no sediment (Figs. 3, 4). Both features are here considered to be products of winnowing; the uniform size distribution could have been produced either as the dead graptolites sank to the bottom in a gentle current, or by syn-depositional erosion of the soft sediment and subsequent sorting during redeposition in hollows. The latter is preferred as a uniform presence of graptolites near the surface would have produced a uniform faunal composition on the bedding planes if sorting occurred during sinking. The winnowing of soft sediment also appears to explain layers consisting solely of fragmented graptolites; this was presumably the product of more intense current activity during which all the enclosing sediment was removed. Occasionally such beds can be 4 mm thick and consist of graptolites and a small number of remnant sand grains and (entrapped?) mica flakes.

If graptolite preservation was not biased in favour of mature colonies one would expect far more juveniles than other growth stages (Fig. 5), as found in Recent marine invertebrates (e.g. brachiopods, see Curry 1982). The presence of bedding planes covered with siculae and juveniles in the low Birkhill Shale and many horizons in the Silurian of the United Kingdom indicates that preservation of juveniles is perfectly feasible and that their absence is likely to be due to sorting. Both this and other horizons with densely packed graptolites are unlikely to be due to mass catastrophies caused by such factors as upwelling of anoxic water (e.g. Berry & Wilde 1978) as this would more often result in a complete range of growth stages. It should be pointed out that if a continuous sequence of black shale is studied in detail, although graptolites occur most commonly at discrete horizons they occur sporadically throughout. However, because of the ease of splitting along planes with abundant specimens, most are collected from such horizons and give the impression that they are almost restricted to such laminae. This often gives the incorrect impression of "mass mortalities" separated by non-fossiliferous intervals.

Many conodonts and one scolecodont have been found in Upper Ordovician black shale at Dob's Linn but only one conodont from the pale mudstone; this is either collection failure, which seems unlikely as the black conodonts are more easily seen in the pale lithology, or an original distribution or preservational feature. In the last case, the conditions affecting preservations of graptolites presumably also applied to conodonts. Inarticulate brachiopods are common in many black shale sequences. At Dob's Linn they also occur at two horizons in the pale mudstone. Specimens in and just above the upper Complanatus Band (Williams & Lockley 1983) are well preserved and in uniformly laminated strata (i.e. not *in situ*). Others between Anceps Bands B and C occur as "nests" of broken fragments in a pyritic, bioturbated interval. Trilobites occur together with fragments of bivalves and nautiloids in an interval just below the Extraordinarius Band (J. K. Ingman and N. H. Trewin, pers. comm.); the Dob's Linn trilobite represents a new blind dalmanitid genus and is clearly a deep water form but not *in situ* (J. K. Ingham, pers. comm.). Algal remains (?) occur in the black shale; these are to be expected if the black shale has a partially algal origin and it is surprising that so few examples of algae in black shales have been recorded. The enigmatic form *Dawsonia,* tentatively assigned to the Ordovician crustacean *Caryocaris* Salter by Rolfe (1969: R316) without comment may be of algal origin due to its similar preservation and original "crocus" flower shape.

At the base of Birkhill Shale is an unfossiliferous interval some 12 cm thick which is mottled and heavily weathered, especially in the lowest part. The different weathering properties indicate a different composition but lithological sections have failed to reveal any clear structures. It is not apparent whether the mottling is an original feature such as bioturbation (although the laminae appear intact) or an irregular distribution of organic material, but the interval does appear to be browner than the overlying black, graptolite-yielding strata. Similar strata occur in Spengill in the Howgill Fells. The Extraordinarius Band is also dark brown rather than black, with a graptolitic unit of only 2 or 3 mm thick. Such brown material may prove important in elucidating the exact formation of black shale as when preparing etchings of graptolites the brown shale/limestone often results in a brown, algal (?) debris, rather than a black one.

McKerrow (1979) and Leggett (1980) have both summarised large-scale processes in British Lower Palaeozoic successions and conclude that change in lithology is directly related to sea level. Although it is clear that certain major events (such as the general regression in the top Ordovician) are related to sea level changes it is improbable that this caused the small-scale reversals described here; such changes are more likely to be related to oxygen content caused by current density and/or density of carbonaceous material. Whichever processes are responsible for the onset of black shale deposition in the Lower Palaeozoic it is clear that there are several different explanations, exemplified by the assortment of structures seen at Dob's Linn, northern England and the Oslo Region. It is therefore hoped that this brief discussion will stimulate further critical study of similar argillaceous sequences elsewhere.

Acknowledgements

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Biogeography of Ordovician graptolites in the southern Appalachians

By STANLEY C. FINNEY

The graptolite fauna of the Middle Ordovician Athens Shale of Alabama shows conspicuous variation between an eastern, deep water, black shale facies and a nearby western, shallower water, calcareous facies. Although this variation involves only seven species out of a large fauna, it is distinctive. These species have very noticeable rhabdosomes and are abundant in the facies in which they occur. They characterize their assemblages and as a result the assemblages of the black shale and calcareous facies are readily distinguished. The two facies are separated by only a few tens of kilometers, and the rapid faunal variation is recognized not only in Alabama but also in Tennessee. The depth stratification model does not explain the faunal variation in which several species are restricted to the shallower water calcareous facies. Lateral fauna differentation produced by water-mass specificity is favored as a more likely explanation.

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One hundred years ago, Lapworth (1879–1880) demonstrated the stratigraphic value of graptolites. Since then graptolite biostratigraphy has been used as the more reliable base for subdivision and inter-continental correlation of Ordovician to Lower Devonian strata. Dependence on, and confidence in, graptolite biostratigraphy was supported by the early realization (Hall 1865; Lapworth 1897) that graptolites were planktonic organisms carried around the world by oceanic currents. This engendered the belief that, once defined, the sequence of graptolite zonses and the assemblages characterizing them could be recognized worldwide (Elles 1922). This assumption, however, became untenable as graptolite research outside northwest Europe intensified.

The reality of marked graptolite provincialism and its constraints on inter-continental correlation was first recignized in the Lower Ordovician graptolite sequence of Australia (Harris & Keble 1932; Harris & Thomas 1938). Subsequent research led to the recognition of: 1. Atlantic and Pacific provinces in the Lower Ordovician (Berry 1960; Skevington 1969, 1973; Bulman 1971). 2. subprovinces within both of these provinces (Jaanusson 1960; Erdtmann 1965; Skevington 1973). 3. an Upper Ordovician fauna endemic to the northern Appalachans in contrast to a coeval cosmopolitan fauna in the rest of North America (Riva 1969; Berry 1977), and 4. gradual changes in faunal composition along paleogeographic gradients (Skevington 1974; Watkins & Berry 1977). Concomintant with these discoveries attempts were made to reconcile the observed geographic distributions with such hypothetical controls as tectonics, depth stratification, vertical and/or lateral temperature differences, water-mass specificity, depth and position of oxygen minimum zone and combination thereof. Most of these hypotheses were offered to explain large scale graptolite provincialism, but in general no studies focused on faunal variation over short distances within a single basin of deposition.

Research in the Middle Ordovician of the southern Appalachians shows that significant variations in graptolite faunas can occur over short geographic distances. Although realization of this phenomenon is not new (e.g. Watkins & Berry 1977; Kaljo 1978), some specialists might discount it because it has not previously been clearly demonstrated. The purpose of the present paper is to document this phenomenon by means of collections from the

In Bruton, D. L. (ed.), 1984. *Aspects of the Ordovician System*. 167–176. Palaeontological Contributions from the University of Oslo, No. 295, Universitetsforlaget.

Athens Shale in Alabama and to evaluate various hypothoses that might be use to explain it.

Geological Setting

Graptolite-rich shales of Middle Ordovician age extend almost the entire length of the Appalachians. The paleogeographic and sedimentologic history of the depositional basin of these shales, which can in general be extended through the length of the southern Appalachians, is summarized by Shanmugam & Walker (1978, fig. 5); and Neuman (1976, text-fig. 2) as follows:

1) In the lowest Middle Ordovician, shallow water carbonates were deposited on the continental shelf, near the eastern edge.

2) This deposition ceased when the shelf rapidly subsided to form a NE–SW trending basin bounded to the west by a carbonate platform and to the east by the uplifted Taconic Highlands, the source area for basinal clastic sediments. The subsidence is reflected lithologically by the abrupt upward replacement of shallow water carbonates (i.e. Lenoir Formation) by graptolitic shales (i.e. Blackhouse Formation) and by the westward transgression of associated facies.

3) After an interval of pelagic sedimentation (graptolitic shales), the influx of terrigenous clastic sediments from the east increased. These sediments (i.e. Sevier Formation) comprised initially of silt and sand and later conglomerate were transported by turbidity currents. Accumulation of graptolite remains within the sediment continued.

4) By the upper Middle Ordovician, the influx of coarse sediments (i.e. Ottosee Formation) filled the basin and spilled over onto the western carbonate platform. Graptolites no longer lived within the basin.

The Athens Shale of Alabama represents the southernmost extension of this geological setting in the Appalachians. It crops out in a narrow band bounded to the west by the Helena thrust fault and to the east by the metamorphic front of the Piedmont Province (Fig. 1). In its eastern outcrops near Calera, it is predominantly black shale, but to the west near Centerville it is replaced by light gray to tan calcareous shale and argillaceous micrite. The black shale facies, resting directly on shallow water carbonates (i.e. Lenoir Formation), represents pelagic sedimentation and possibly a turbiditic influx of clastics from the east (Fig. 2). The western, calcareous facies of the Athens Shale (Fig. 3), which gradationally overlies a slope facies (i.e. the Pratt Ferry Beds), accumulated at a position intermediate between the deep water black shale basin to the east and a contemporaneous shallow subtidal carbonate platform to the west (i.e. the Chickamauga Group). It is a mixture of fine clastic sediment derived from the east and carbonate mud washed eastward from the carbonate platform.

The black shale facies was extensively sampled for graptolites near Calera (locality C in Fig. 1), and graptolites were collected at many horizons in the calcareous facies (localities PF and PS in Fig. 1). Although the two facies are largely contemporaneous and have many graptolite species in common, conspicuous faunal differences do exist between them.

Graptolitic biostratigraphy

Calera section (black shale facies)

Near Calera, Alabama, the Athens Shale is represented by 75 meters of predominantly black shale that is completely exposed, is continuously and abundantly graptolitic, and was extensively sampled, in some parts at 0.2 meters intervals (Fig. 2). Most of the Athens Shale is correlated to the lower half of the *Nemagraptus gracilis* Zone with a lower boundary placed at the first appearance of *N. gracilis* at 10.5 meters above the base of the unit (Finney 1977; Finney & Bergström, in press). Below this level, the Athens Shale is assignable to the *Glyptograptus* sp. cf. *G. teretiusculus* Zone.

The Calera collections are characterized by several long-ranging, cosmopolitan species, while others are rare (e.g. *Pterograptus* sp., *Amphigraptus* n.sp. B), relatively endemic (e.g. *Apoglossograptus lyra, Climacograptus meridionalis*) and/or stratigraphically restricted (e.g. *D. geniculatus, P.* sp. cf. *P. eurystoma,* and *Leptograptus trentonensis*).

The Calera range chart (Fig. 2) consists of data combined from two measured sections (Fig. 1; see Finney 1977 for full description). In the following discussions, both sections are referred to as one by the designation Calera section.

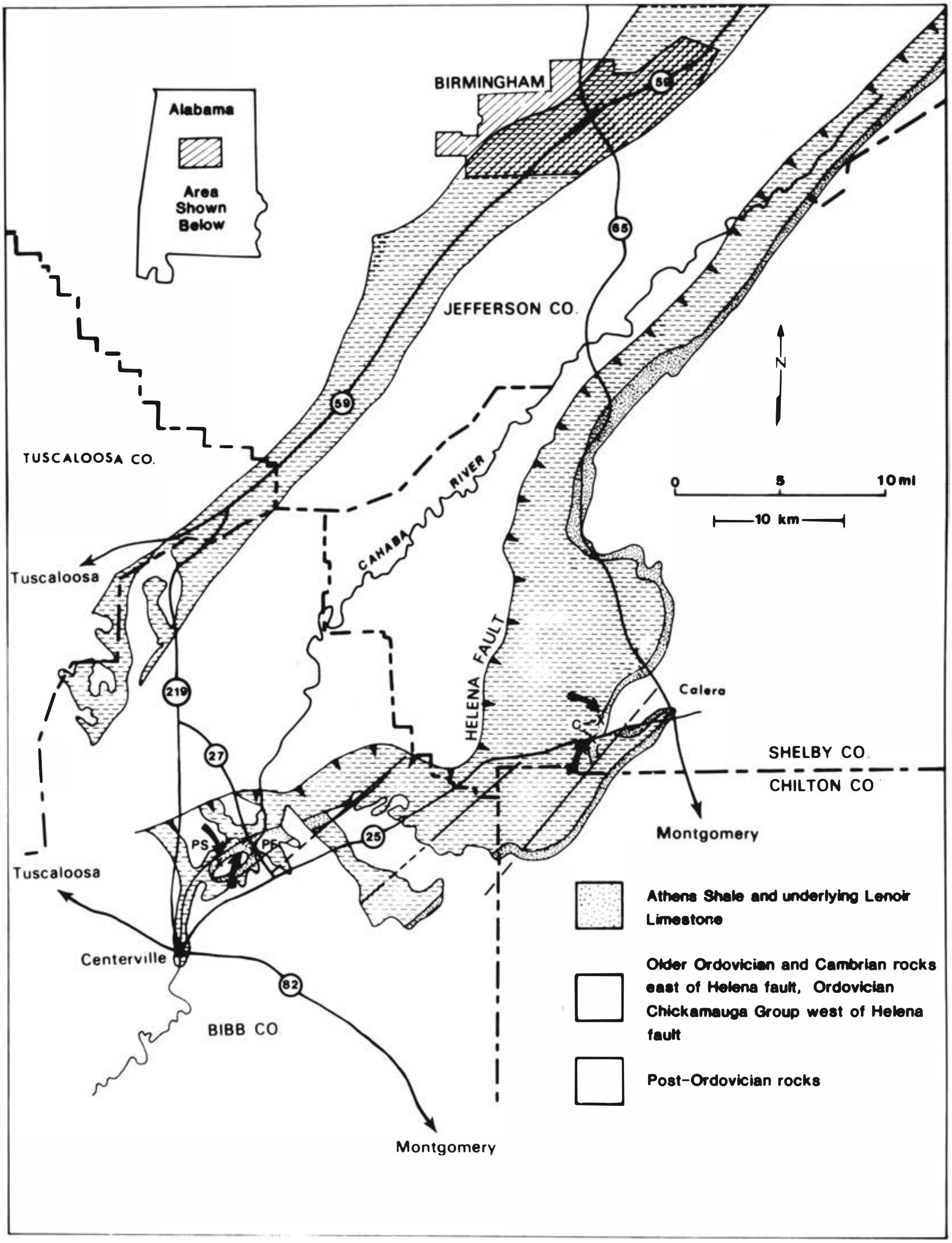

Fig. 1. Index map of central Alabama showing distribution of Ordovician rocks and location of sections at Pratt's Ferry (PR), Pratt's Syncline (PS) and Calera (C).

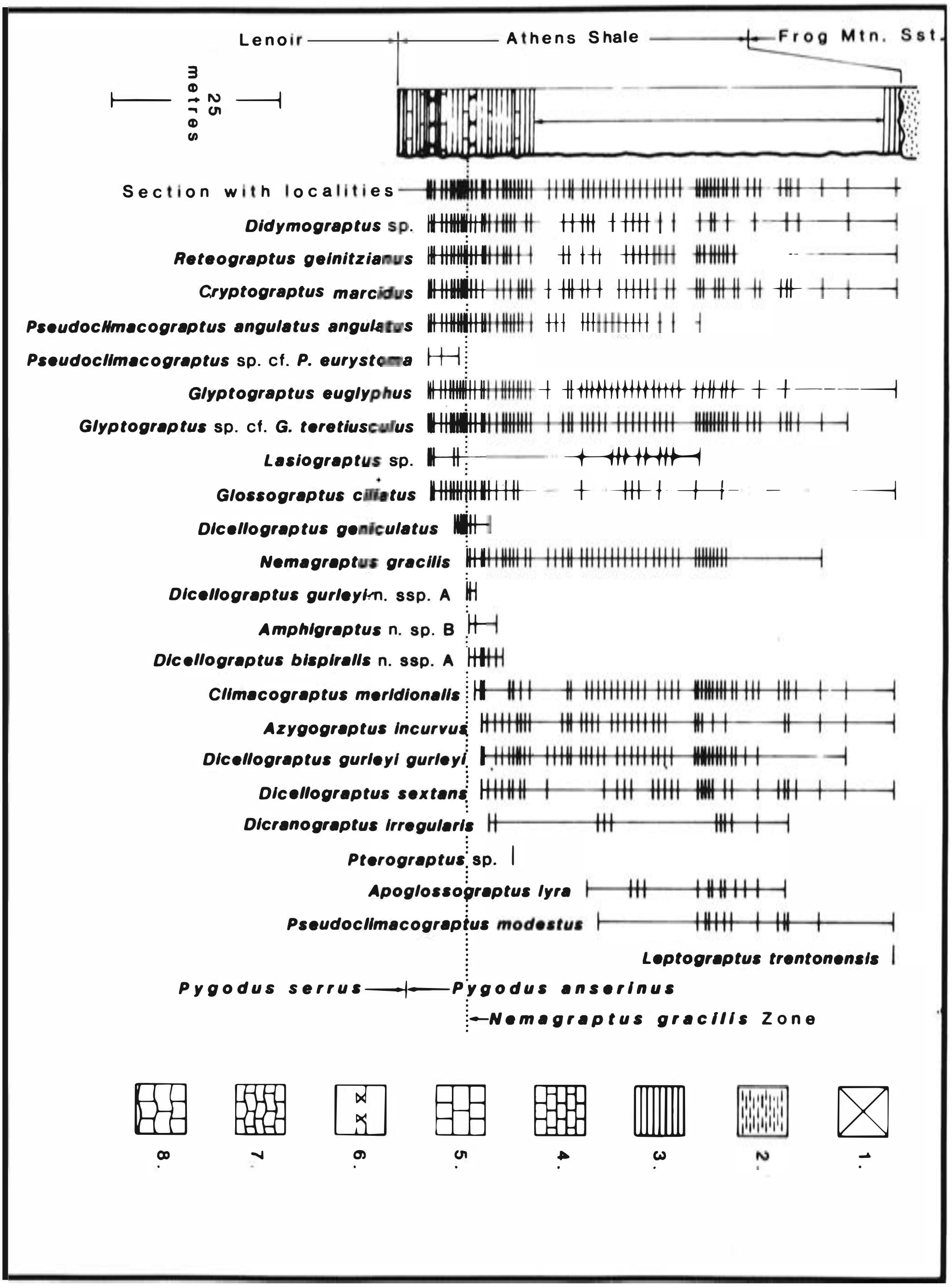

Fig. 2. Range chart and stratigraphic column of Calera section. Taxonomy is after Finney (1977, 1978, 1980). Symbols: 1. Covered interval; 2. Mudstone, generally calcareous; 3. Shale, black shale and calcareous shale; 4. Thin to medium bedded micrite; 5. Thick bedded micrite; 6. Bentonite; 7. Medium bedded fossiliferous limestone with wavy bedding surfaces, i.e. Pratt Ferry Beds; 8. Massively bedded fossiliferous limestone with wavy bedding surfaces, i.e. Lenoir Limestone.

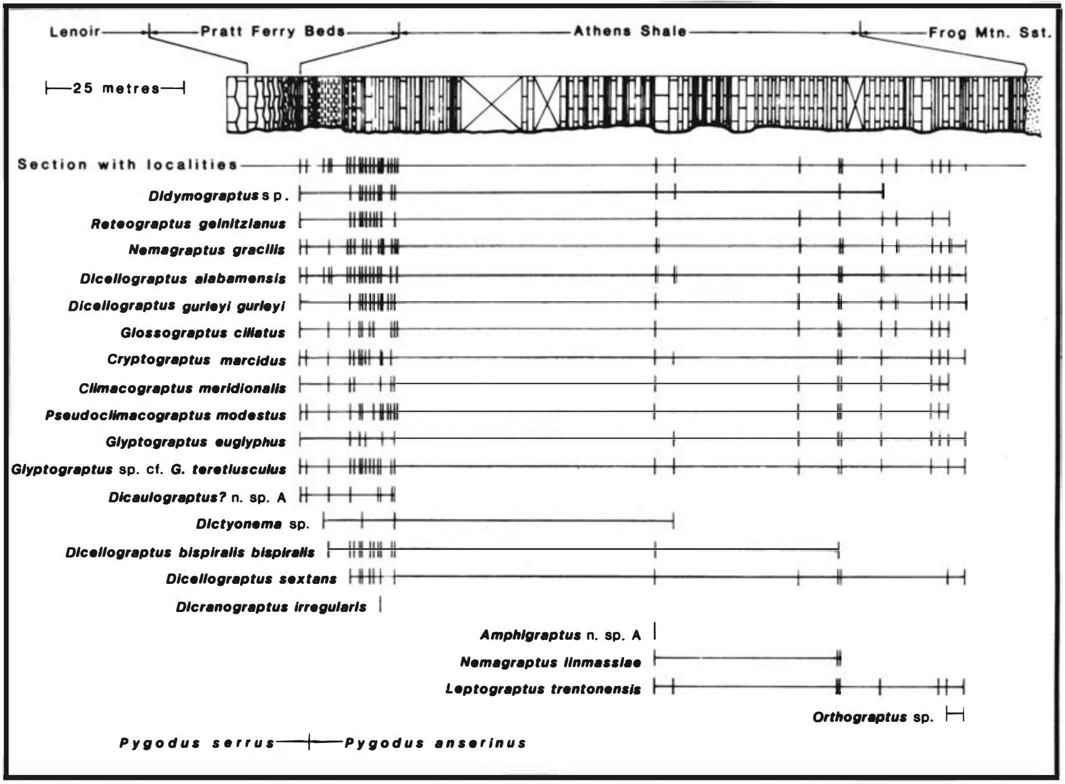

Fig. 3. Range chart and stratigraphic column of Pratt's Ferry – Pratt's Syncline section. Taxonomy is after Finney (1977, 1978, 1980). Symbols as in Fig. 2.

Pratt's Ferry and Pratt's Syncline Sections (calcareous facies)

The range chart for the calcareous facies (Fig. 3) is also a compilation of data from two measured sections (see Finney 1977). Only the lowest 25.5 meters of the Athens Shale is exposed at the Pratt's Ferry section. It is rich in graptolites and was sampled at 22 levels. The nearby Pratt's Syncline section includes the entire 134-meter thickness of the Athens Shale. Its lowest 25 meters is lithologically identical to the Pratt's Ferry section. Its middle and upper parts are dominated by micrites. Shale interbeds are few and thin. Graptolites were found at only twelve horizons that are stratigraphically above the Pratt's Ferry section. However, at these horizons they are abundant and well-preserved. In the following discussions the Pratt's Ferry and Pratt's Syncline sections are referred to collectively as the PF–PS section.

As at Calera, the graptolite collections are characterized by several long-ranging, cosmopolitan species, while others are rare (e.g. *Dicranograptus irregularis, Amphigraptus* n.sp. A, *Nemagraptus linmassiae*), relatively endemic (e.g. *Dicaulograptus?* n.sp., A, *Dicellograptus alabamensis, D. bispiralis*), and/or stratigraphically restricted (*Leptograptus trentonensis* and *Orthograptus* sp.).

The presence of *Nemagraptus gracilis* in the lowest graptolite collection and *Orthograptus* sp., which helps define the base of the *Climacograptus bicornis* Zone (Finney & Bergström in press), in the highest collection indicates that the calcareous facies is entirely within the *N. gracilis* Zone.

Correlation of sections

Comparison of species composition, graptolite ranges, and conodont ranges between the Calera and PF–PS sections shows that, while the black shale and calcareous facies differ somewhat in age, they are largely equivalent. The base of the Athens Shale is diachronous reflecting the basinal subsidence and westward transgression exhibited in the Middle Ordovician throughout the Appalachians. The levels of the first appearance of *gracilis* Zone graptolites and of the *Pygodus serrus* – *P. anserinus* conodont zone boundary demonstrate that graptolite shales of the eastern black shale facies began accumulating before those of the western calcareous facies. The top of the Athens Shale is an unconformity and was eroded to different levels in the two areas before deposition of the Devonian Frog Mountain Sandstone. The level of first appearance of *Leptograptus* in the two sections indicates that the top of the Athens Shale at Calera is correlative with a level within the lower half of the PF–PS section. Thus, although the PF–PS and Calera sections are largely correlative, the Calera section extends to a lower biostratigraphic level than the PF–PS section, and the PF–PS section extends to a higher level than the Calera section.

Geographic faunal variation between Calera and PF–PS sections

Graptolite collections from the Calera and PF–PS sections may have many common species. However, several species are restricted to either the Calera or the PF–PS section, and this makes the assemblages of the two sections distinctive.

Some of the faunal restriction is due to age differences. For example, the biostratigraphic ranges of *Pseudoclimacograptus* sp. cf. *P. eurystoma, Dicellograptus geniculatus, D. gurleyi* n. ssp. A, and *D. bispiralis* n. ssp. A are too low for them to appear in the PF–PS section, and the range of *Orthograptus* sp. is too hight for it to appear at Calera. Differences involving *Amphigraptus* n. sp. A, *Amphigraptus* n. sp. B, *Nemagraptus linmassiae, Lasiograptus* sp., and *Pterograptus* sp. are no doubt due to the scarcity of available specimens. Conspicuous differences between the sections still remain, and these must be due to ecological factors restricting the geographic ranges of seven other species. These species and their geographic and stratigraphic distributions are as follows (all have been or will be described by Finney 1978, 1980, in preparation):

1) *Pseudoclimatograptus angulatus angulatus* Bulman is abundant through most of the Calera section. Its long rhabdosome and distinctive apertural excavations make it easy to recognize. Its complete absence from the PF–PS section is noticeable especially because it is a cosmopolitan species.

2) *Azygograptus incurvus* Ekström is common in 40 collections from the Calera section where distinctive rhabdosomes, each consisting of a single strongly curved stripe, cover bedding planes. Although cosmopolitan, the species is absent in the PF–PS section.

3) *Dicranograptus irregularis* Hadding is a common species in the middle of the Calera section distinguished by a short biserial proximal end. A single specimen was found in the PF–PS section. Its scarcity there is surprising in light of the many specimens collected at Calera and its worldwide distribution.

4) *Apoglossograptus lyra* (Ruedemann) is common in the upper part of the Calera section but completely absent in the PF–PS section. It has a conspicuous rhabdosome consisting of two proximally monopleural but distally diverging stipes.

5) *Dicellograptus alabamensis* Ruedemann has a large distinctive rhabdosome with a spinose proximal end often thickened with cortical tissue and appearing to be biserial (Ruedemann 1908, pl. 20, figs. 1–2). It is abundant throughout the PF–PS section. Its complete absence at Calera is conspicuous especially because it occurs in Virginia and Texas.

6) *Dicellograptus bispiralis bispiralis* (Ruedemann) has a distinctive rhabdosome composed of two helical stipes with extremely introverted thecae. It is abundant throughout the PF–PS section. Yet, it is absent at Calera even though it has been recorded in Tennessee (Decker 1952).

7) *Dicaulograptus?* n. sp. A is peculiar because its biserial rhabdosome displays torsion and it is easily recognized by its highly spinose and introverted theacae. It occurs only in the lower part of the PF–PS section. But because

it is so abundant there and so distinctive, its absence from an equivalent interval at Calera is readily noticed. The species is known from Tennessee (S. Finney personal collections).

The differences between the assemblages of the Calera and PF–PS sections, involving only seven species, is not great. However, it is conspicuous because these seven are so morphologically distinct and so abundant that they readily distinguish assemblages from the sections in which they occur.

The faunal differences occur over a geographic distance of 43 kilometers. If adjusted for structural shortening, the distance is 70 kilometers. The faunal variation is thus surprising and significant because the seven species that do not co-occur in well sampled, correlative sections only a few tens of kilometers apart in Alabama have been reported from localities hundreds and thousands of kilometers away from Alabama.

Other localities in the southern Appalachians

Decker (1952) recorded graptolites from many localities in eastern Tennessee and southwestern Virginia. In addition, USGS field parties, University of Tennessee graduate students, S. Bergström and S. Finney have also made many collections from the Middle Ordovician shale basin in this region. The seven species listed above occur in several of these collections, in most instances in the same associations as the Calera and PF–PS sections. This demonstrates that there are other sections in close proximity (tens of kilometers) that appear to display the same faunal variation as in Alabama. These sections have not been systematically collected so that detailed correlations and comparisons cannot yet be made. They do however show that the geographic differentiation of the graptolite fauna extends throughout the southern Appalachians.

Cause of faunal variation

The graptolite variation so clearly demonstrated in Alabama must be a reflection of ecological controls. The extensive systematic collecting and precise correlations rule out collecting failures and age differences as the cause. Tectonic controls, such as the juxtaposition of once widely separated basins of deposition by plate tectonics that was proposed to explain some graptolite provincialism (Skevington 1973) can also be disregarded because the Alabama localities and those in Tennessee and Virginia were deposited in a single basin of deposition.

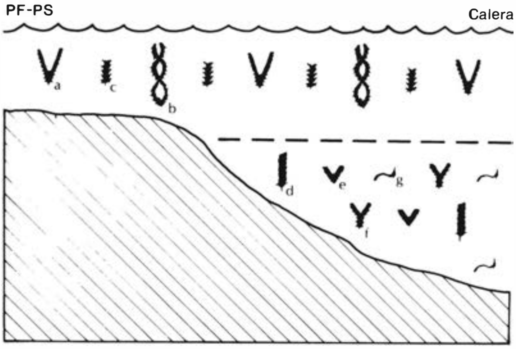

Fig. 4. *Depth stratification model in which* Dicellograptus alabamensis *(a)*, D. bispiralis bispiralis *(b), and* Dicaulograptus ? *n. sp. A (c) inhabit upper layers of water column and can thus occur in the PF–PS region and* Pseudoclimacograptus angulatus angulatus *(d)*, Apoglossograptus lyra *(e)*, Dicranograptus irregularis *(f), and* Azygograptus incurvus *(g) are restricted to deeper water and thus cannot occur in the PF–PS region. This model is invalidated by absence of fossils of* D. alabamensis, D. bispiralis bispiralis, *and* Dicaulograptus ? *n. sp. A at Calera.*

Many ecological factors have been proposed to explain graptolite provincialism. Large-scale controls, in particular the latitudinal distribution of climatic belts favored by Bulman (1964, 1971), Boucek (1972), and Skevington (1974) as the cause of global variation, are not appropriate because the variation in Alabama is sharp, occurs over a small distance, and is repeated several hundred kilometers away in Tennessee. Other possible ecological controls can be grouped into two categories that involve either a vertical differentiation of the graptolitic fauna or a lateral differentiation.

Depth stratification or vertical zonation of graptolites was first proposed by Berry (1962) and later used by Skevington (1969), Berry & Boucot (1972), Erdtmann (1976), and Kaljo (1978) to explain geographic variation. Cisne & Chandlee (1982) recently invoked it in order to relate lateral and vertical variations in abundance of various taxa to changing water depths, i.e. transgressions, regressions, and topography. Depth stratification has been attributed to vertical changes in water temperature (Berry & Boucot 1972), light intensity, nutrient supply, water turbulence, and the position of the oxygen minimum zone (Erdtmann 1976; Cisne & Chandlee, 1982).

Berry (1974, 1977; Watkins & Berry 1977) later attributed geographic variation to a lateral differentiation of the graptolite fauna. This variation occurring across and within a basin of deposition was due to water-mass specificity, i.e. the differentation of surface water masses and the restriction of graptolite species to them.

The Alabama data are best explained by a lateral differentiation of the fauna. The Calera section was probably a deeper water site of sedimentation than the PF–PS section because of the timing of sedimentation and its lithology. The depth stratification model (Fig. 4) can, therefore, account for the restriction of *Pseudoclimatograptus angulatus angulatus, Apoglossograptus lyra, Dicranograptus irregularis,* and *Azygograptus incurvus* to Calera if they are restricted to a deeper water habitat. However, it cannot explain the restriction of *Dicellograptus alabamensis, D. bispiralis bispiralis,* and *Dicaulograptus?* n. sp. A to the PF–PS section. These species must have lived at shallow depths in the sea to occur at the PF–PS section, but this would not have prevented them from inhabiting waters overlying those with deep water species. In fact, they should also occur at Calera. Proponents of the depth-stratification model (Erdtmann 1976; Cisne & Chandlee 1982) explain the absence or scarcity of shallow-water species in deep-water strata by the selective destruction of the rhabdosomes of shallow-water species during their long drift to the deep burial site. However, the large durable rhabdosomes of the shallow-water species in Alabama and their easy preservation and abundance in the oxygenated and biologically-rich (and thus predator and scavenger rich) environment of the shallow-water carbonate facies, strongly suggest that at least some of these rhabdosomes should survive the passage to the deep burial site. Their complete absence from the deep-water facies, which was so extensively collected at Calera, indicates that they were not present in the waters overlying the area. The abundance of *Cryptograptus marcidus* throughout both the PF–PS and Calera sections (Figs. 2, 3) also argues against selective destruction. Among the more than 30 species for which I have studied isolated material, this species has the thinnest, weakest periderm. It is usually preserved on shale surfaces only as a faint film while the periderm of other species on the same bedding surfaces stands up in relief. If any species should show selective destruction in either a shallow, oxygenated environment or during sinking to a deep burial site, then it should be *C. marcidus.* Yet, it does not in the Athens Shale. Finally, if the Athens graptolites were depth stratified, then the manner of initial appearance of graptolites in each section should reflect the gradual deepening of the sea at these localities as it transgressed westward. While the gradual incoming of species in the Calera section might be taken as evidence of this phenomenon, the sudden appearance of many species in the PF–PS section argues against it. The fact that several species are restricted to the shallow water PF–PS section and absent at Calera can be instead easily explained by a lateral differentiation of the fauna (Fig. 5).

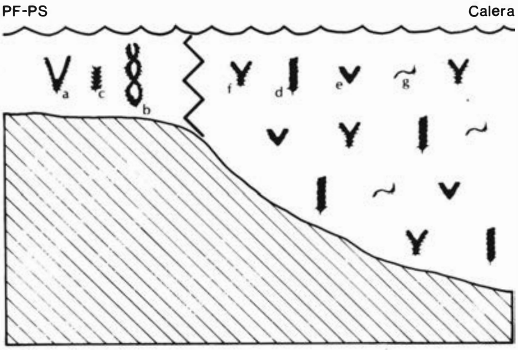

Fig. 5. Lateral differentation in which characteristic species are confined to water masses overlying either the Calera region or the PF–PS region. Letters denoting species as in Fig. 4.

The waters overlying the Calera and PF–PS sections would no doubt have differed in turbidity and salinity as reflected by their lithology and inferred from their paleogeographic positions. Barry (1974, 1977) has pointed out that studies of modern faunas (e.g. Fager & McGowan 1963) show that plankton are often restricted to oceanic water masses with specific hydrographic characteristics. Although it appears so in Alabama, these water masses and their restricted faunas need not be precisely reflected by the character of the underlying sediments. In Tennessee, assemblages such as those at Calera occur in both black shale and calcareous facies.

Conclusions

Graptolites are difficult to analyse paleoecologically. They have no closely related extant analogue, either taxonomically or morphologically. Furthermore, because they were pelagic, rocks containing their remains provide few clues to the nature of their habitat. The Alabama example is not to show that biogeographic variation is consistently associated with calcareous and black shale facies or due exclusively to water-mass specificity, but just that it can indeed exist over short geographic distances and in this case is easily explained by water-mass specificity. In spite of this demonstrated variation, graptolites are still valuable biostratigraphic tools. For example, the base of the *gracilis* Zone at Calera has been utilized by Finney & Bergström (in press) as a standard against which graptolite and conodont data have been used to evaluate the base of the zone in New York, Texas, Scotland, Wales, Scania, and Australia.

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Black shale geochemistry: A possible guide to Ordovician oceanic water masses

By WILLIAM B. N. BERRY

Most modern oceanic plankton faunas are circumscribed in their occurrence by the distribution of the water mass in which they live. Even in shelf seas, water mass properties exert such a control on plankton distribution that certain taxa may be used to recognize the presence of a particular water mass. Distribution of different but seemgly coeval Ordovician graptolite faunas suggests that water mass characteristics played the same role in plankton distribution in the past as they do today. Geochemical analyses of selected Ordovician graptolite-bearing shales indicate that each shale bears trace metals in combinations and abundances unique to each. Trace metals in coeval Late Ordovician shales in New York and the western United States differ in a manner consistent with variations in graptolitic fauna. Modern ocean water masses also have trace metal contents unique to each. Plankton tissues take up these metals which may be transported relatively rapidly to the sea floor by zooplankton faecal pellets. If the pellets fall into anoxic environments, the sediment will reflect the trace metal composition of the waters in which the plankton lived because the metals will remain essentially in place as sulphides. The distribution of water masses may therefore be traced by using the trace metal geochemical signatures of black shales.

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Distribution patterns exhibited by Ordovician graptolites have attracted considerable attention (e.g. Berry 1979; Bulman 1971; Skevington 1974). The Ordovician graptolite faunal regions and provinces are characterized by unique, primarily endemic taxa. For example, isograptids and pseudisograptids typify the Pacific faunal region. These taxa are not found in the Atlantic faunal region. Coeval Atlantic region faunas typically include pendent (tuning fork shape) didymograptids of the *D. murchisoni* group. Skevington (1974) suggested that temperature difference in Ordovician oceanic surface waters may have been a major factor in development of the two Ordovician graptolite faunal regions.

Another, subtle pattern in graptolite faunal associations may be recognized within the context of the faunas that characterize the regions and provinces. Dover *et al.* (1980) described such a pattern in an analysis of coeval sets of Middle Ordovician graptolite collections from stratigraphic sections in thrust belts in central Idaho. The faunas from the Trail Creek section are typified by the presence of distinctive pseudisograptids whilst apparently coeval faunas from the Little Fall Creek section, are characterized by the presence of glyptograptids of the *G. austrodentatus* group. Pseudisograptids are rare in the latter section. Certain isograptid taxa are common to both areas. The stratal sequences in both areas have been sampled extensively over a number of years by several different collectors and sampling bias does not seem to be a significant factor in the faunal difference.

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An explanation for the small scale differences in Ordivician graptolite associations

The small scale differences among seemingly coeval Ordovician graptolite faunas in a number of localities have remained despite intensive collecting and are clearly real rather than due to preservational or sample bias.

One approach to understanding the reasons for these differences among plankton faunas from ancient oceans is to seek comparable or analogous patterns among plankton faunas in modern oceans. Controls on distribution of modern plankton faunas may be recognized. Potentially, distribution of plankton faunas in ancient seas may have been controlled by similar environmental factors.

Russell (1952), Bradshaw (1959) and Fager & McGowan (1963) documented the relationship that major modern oceanic plankton faunas are limited to specific oceanic water masses. Whereas these reflect plankton fauna-oceanic water mass relationships at a relatively broad scale, more detailed analysis of plankton faunas has revealed similar relationships but at a smaller scale. Fraser (1965) recognised eight plankton associations in the North Sea and found that each of these was limited in its distribution by the limits of the water mass in which it existed and the unique physical and chemical properties of the water. Temperature and salinity are the primary properties involved in the differences between water masses.

Johnson & Nigrini (1980, 1982) described radiolarian associations in surface sediments of the western and eastern parts of the Indian Ocean and found that the distribution of the tests of these planktonic organisms in the surface sediments closely reflects water mass distribution. Thus, not only may the plankton living in a water mass, but also their remains in surface sediments reflect control on their distribution by the water mass in which they live.

Those faunal associations found in the present North Sea and especially those in the Indian Ocean surface sediments that reflect waer mass control over their distribution, suggest a possible explanation for the small scale differences cited among coeval Ordovician graptolite faunas. The differences described may indicate that specific graptolite associations were restricted to particular oceanic water masses.

Patterns similar to those described for Ordovician graptolites have been recorded among Palaeogene and Neogene plankton faunal associations (Sancetta 1978, 1979). These faunas, obtained from cores in the Pacific Deep Sea drilling programme (Keller 1978) have proved useful in documenting palaeooceanographic and palaeoclimatic changes through time.

Trace metal geochemistry of Ordovician graptolitic shales

As an aid in tracing rock units that bear a unique graptolite association, and to more fully explore the possibility that water mass properties did influence the distribution of Ordovician graptolite associations, trace metal compositions of selected Ordovician graptolite-bearing strata have been examined. This approach to an understanding of small scale Ordovician graptolite biogeography has developed from results of studies of trace metal associations in modern oceanic water masses. Bewers *et al.* (1976) and Campbell & Yeats (1982) noted that "biological uptake of metals" takes place near the surface in present-day oceanic water masses and that some have unique trace metal associations. Certain plankton appear to have trace metals in their tissues that reflect the trace metal species present in the water. When these plankton are preserved in ocean floor sediment such that decay of their tissues is retarded and release of contained metals inhibited, then the chemistry of the sediments and rocks formed from them, will reflect the trace metal composition of the surface waters at the time the plankton lived.

The unique chemical characteristics of the *Dictyonema flabelliforme sociale*-bearing phyllites at Nordaunevoll, eastern Trøndelag, Norway enabled Gee (1981) to trace these phyllites in the field. Gee (1981: 93) stated that "trace element (V, Mo, U) analysis of the Nordaunevoll phyllites has demonstrated a chemistry that is strikingly similar to that of sediments of similar type and age on the Baltoscandian Platform in Skåne and Östergötland". He suggested that it might be "possible to identify similar tectono-stratigraphic levels in the Swedish Caledonides of northern Jämtland and Västerbotten, and elsewhere in the mountain belt".

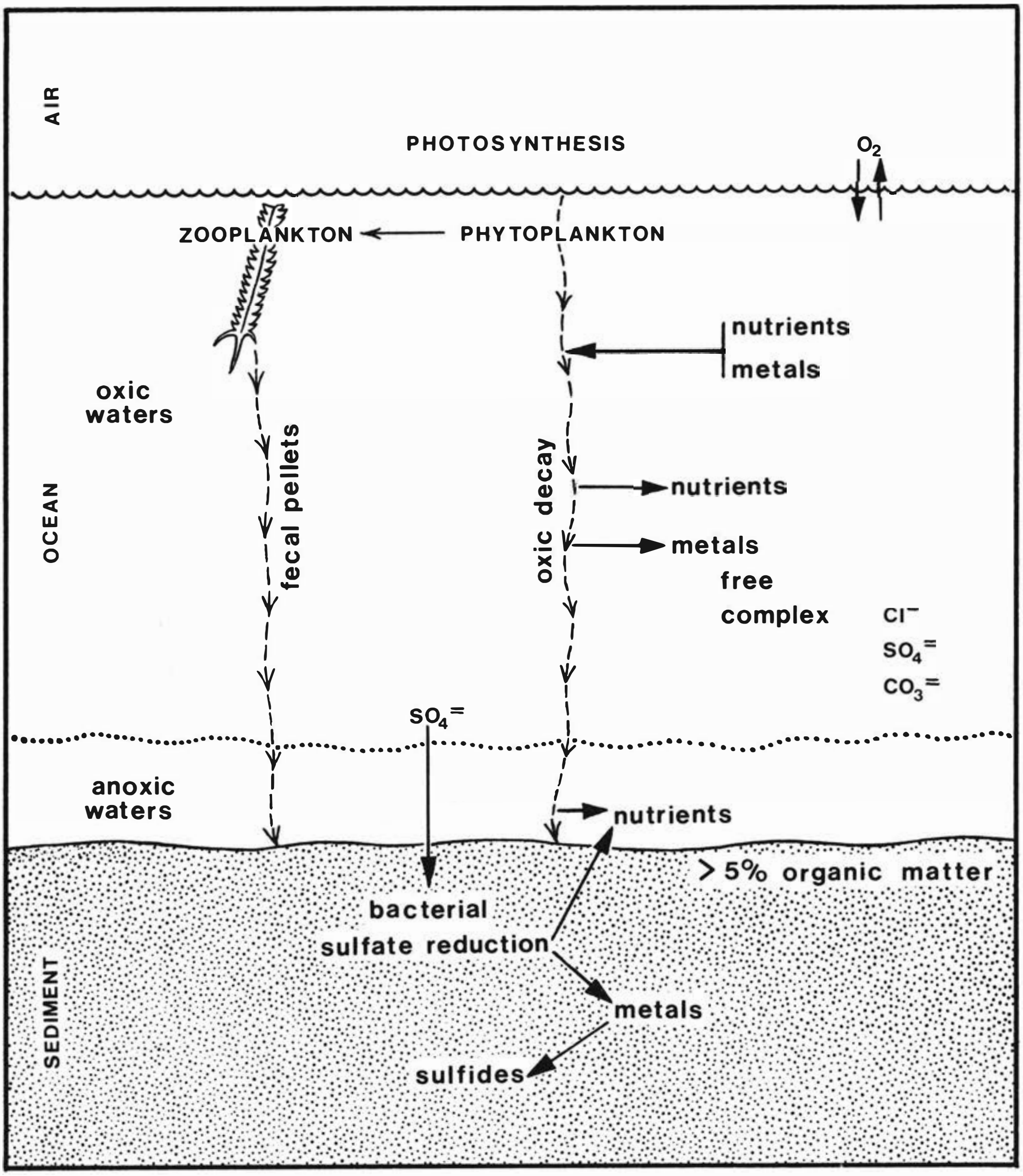

Fig. 1. Schematic diagram of an Ordovician oceanic environment in which graptolites may have lived. The diagram suggests that phytoplankton tissues may be "packaged" into zooplankton faecal pellets, thus providing a mechanism for rapid deposition of metal-bearing organic material. Anoxic bottom waters will result in sulphate reduction which may lead to metals being preserved essentially "in place" as sulphides.

Gee also noted that not only may trace metal geochemistry of the *Dictyonema*-bearing shales play a role in tracing them from area to area, but also it may be useful in biogeographic and palaeogeographic reconstructions as a consequence of identifying these shales in remote areas by their geochemical signature.

Poole & Desborough (1981) and Gee (1981) drew attention to the relatively high (more than 5 percent) organic carbon content of (Lower?) Palaeozoic black shales that bear significant quantities of metal sulphides. Anoxic depositional environments in which organic matter is not fully decomposed by oxidation is important to the preservation of trace metals contained in plankton that fall partly decayed to the sea floor. Under anoxic conditions, many metals tend to form sulphide complexes and to remain in place.

Plankton tissues bearing trace metals may accumulate on the sea floor relatively rapidly by incorporation in faecal pellets (Fig. 1). Smayda (1971) and Porter & Robbins (1981) noted that zooplankton faecal pellets may preserve planktonic organisms in a partially decayed condition both by "packaging" them in pellets that drop relatively quickly to the sea floor and by protecting the organic matter with an organic (probably mucous) rind on the outside of the pellets. The rind retards decomposition by oxidation in oxygenerated waters. Porter & Robbins (1981) examined samples of dark, organic-rich shales of different ages from a number of localities and suggested that faecal pellets were the most abundant organic particles in the shales that they studied and that such pellets appear to be a major source of organic matter.

The trace metal composition of late Ordovician (approximate equivalent of the *Dicranograptus clingani* Zone of Britain), Snake Hill Shale in eastern New York state and approximately coeval black, graptolite-bearing Vinini Shale in Nevada has been determined using neutron activation (see Perlman & Asaro 1969; Alvarez *et al.* 1982, for technique). Poole & Desborough (1981) also recorded geochemical analyses of samples from the Vinini Shale. Analyses of these coeval late Ordovician shales from eastern and western North America that have somewhat different graptolite faunas, show the Vinini Shale, to have relatively greater quantities of chromium, molybdenum, uranium, and vanadium than does the Snake Hill Shale whilst the latter bears significant quantities of cobalt, iron, manganese, and titanium. The difference in trace metals between the shales is consistent with differences in graptolitic faunas of each. Potentially, the trace metal composition of the Snake Hill Shale reflects the influence of waters from the deltaic environments in eastern New York and adjacent areas described by Bretsky (1970) and Dennison (1976).

Conclusion

Trace metal composition of Ordovician graptolite-bearing shales may provide a useful tool in correlations (Gee 1981), and it may reflect differences in water mass chemistry in ancient oceans as is indicated by the analyses of trace metals in the Snake Hull and Vinini Shales. Poole & Desborough (1981) and Trudinger (1981) noted the possibility that Lower Palaeozoic organic rich shales may bear metals in sufficient quantities to make them potentially valuable ore deposits. Trace metal analyses of Ordovician graptolite-bearing strata, though seldom carried out to date, may therefore prove to be economically important as well as significant in correlation and in understanding graptolite biogeography.

Acknowledgement

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Ordovician volcanicity

By CHRISTOPHER J. STILLMAN

Reconstructions of World palaeogeography through the Lower Palaeozoic period suggest extensive movement of long-lived continental or micro-continental masses which produced relatively shortlived ocean basins. These basins opened by rifting, spread with the formation of new ocean crust and then closed by subduction of that crust with the attendant development of island arcs, marginal basins and active continental margins. It was in these situations that the great majority of Ordovician volcanic rocks were erupted. Volcanics are found sporadically throughout Lower Palaeozoic strata and evidence of world-wide sustained climaxes of volcanic activity are preserved commonly in late Cambrian to early Ordovician and in middle to late Ordovician sequences.

There is undoubtedly a connection between geotectonic situation and the type and intensity of volcanism. Products of initial crustal rifting are widely seen in the Cambrian and post-orogenic continental volcanism was common in the Devonian, but Ordovician activity was dominantly subduction-related, pre- or syn-orogenic, occurring in submarine, island, coastal or cordilleran environments, often situated on crust whose instability was responsible for an abundance of distinctive features in the volcano-sedimentary record. The characteristics of this volcanism and the nature of its products is described, with reference to present day analogues.

The chemistry of volcanic rocks is largely controlled by their tectono-magmatic environment and geochemical descriminant analysis has been widely used as a means of distinguishing between the environments. By such means Ordovician volcanic rocks within, for instance, the Caledonide orogen have been recognised as the products of eruption above subduction zones bordering Iapetus; on continental plate margins that came together in collisions that largely destroyed the oceanic lithosphere, preserving it only in small remnants in obducted slices. The substantial replacement of volcanism by plutonic activity in late Silurian to Devonian times is believed to be due to the suturing of the plates which commonly terminated subduction.

Whilst adverting to the world-wide distribution of Ordovician volcanism, emphasis in this paper is placed on the Iapetus region. In particular the volcanic rocks of the British and Irish Caledonides are described in some detail as an illustration of features which might be expected anywhere in similar tectono-magmatic situations.

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Introduction

Many of the problems which face those who seek to reconstruct the palaeogeography of the World in Ordovician times derive from the difficulty of positioning accurately the remarkably mobile continental masses of that time. Palaeomagnetics and terrain analysis suggest the movement of relatively long-lived continental or microcontinental masses producing relatively short-lived ocean basins which opened and closed with extensive subduction and attendant development of island arcs and marginal basins. The "docking" of continental masses with a variety of relative motions ultimately created compressive or transpressive orogenic events, deforming and dislocating the sequences of rocks which were formed in the Lower Palaeozoic at the margins of these continents; rocks which include the great majority of Ordovician volcanics. There is undoubtedly a connection between the geotectonic activity

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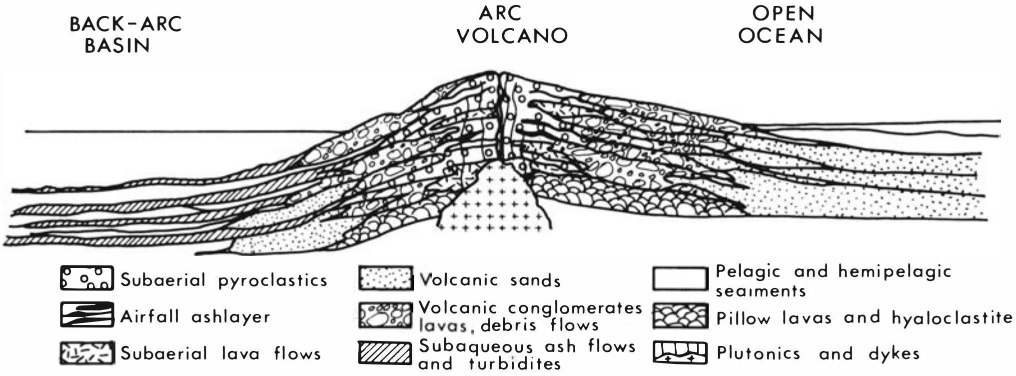
and the intensity, extent and type of volcanicity; volcanic rocks are found sporadically throughout the Lower Palaeozoic, but widespread and sustained periods of volcanic activity are common only in late Cambrian to early Ordovician times and again in the middle to upper Ordovician, in both of which periods worldwide climaxes of activity have been recorded.

Almost everywhere these volcanics erupted in island arcs, marginal basins or volcanically active continental margins. Their magmas belonged to either or both of two magma "series"; 1) the "Orogenic Magma Series", a basalt-andesite-dacite association with chemistry ranging from sub-alkaline island-arc tholeiite through calc-alkaline suites to potash-rich shoshonites; or: 2) a characteristically bimodal association of tholeiitic basalt and calc-alkaline to alkaline rhyolite, often with a "within plate" chemistry. By analogy with modern examples, these Ordovician volcanoes are believed to have erupted above subduction zones, and the dramatic reduction of volcanicity and its replacement by essentially plutonic activity in late Silurian to Devonian times is believed to be due to suturing resulting from continental plate collisions which, in most cases, terminated the active subduction and initiated the variety of cratonising events which make up the end-Caledonian orogeny.

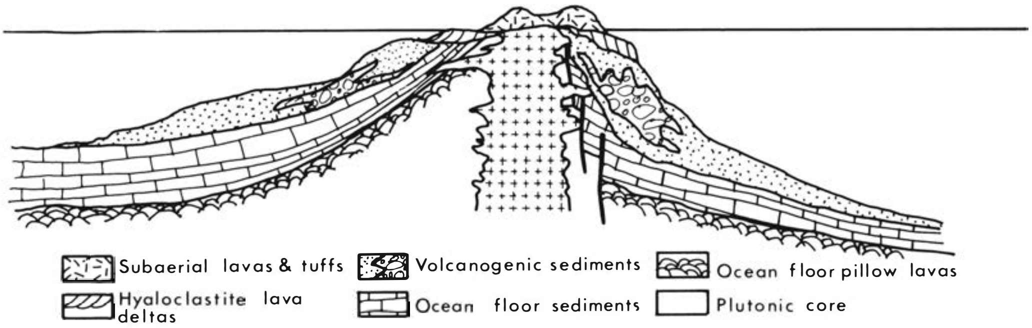
Characteristics of subduction-related volcanism

Present-day analogues indicate two principal environments; firstly the volcanic island arcs, often separated from the continental land masses by marginal back-arc basins, such as are seen in the western Pacific and South East Asia, and secondly the active continental margins such as the American Northwest, or the Andean region. In the first of these, the volcanism is of two distinct types: that of central volcanoes which build up arcs of oceanic islands, and that of sea-floor rifts within extensional basins. The products of this latter type of activity closely resemble the new ocean floor produced by mid-ocean rifting except that, in many cases, the crust is sialic and the rifting is not complete. Thus the result is a suite of tholeiitic basalt dykes intruding a thinned continental basement, on the surface of which submarine basaltic lava flows and hyaloclastites are deposited together with pelagic and hemipelagic sediment. The island-arc volcanoes, on the other hand, are generated by calc-alkaline magma which is largely intermediate in composition, containing higher concentrations of volatiles and producing predominantly explosive eruptions which generate pyroclastic ejecta. This, in a marine environment devoid of terrigenous detritus, provides the volcanic sediment which is the principal infill to the adjacent back-arc basins. An example is the Granada Basin west of the Lesser Antilles arc of the Carribean, where volcanic sediment input has added 7 km of sediment in the 47Ma since the inception of volcanism (Siggurdson *et al.* 1980). The bed-forms, nature of grading and sorting depend both on the character of the source and the transport mechanism. Three of the major forms of transport described by Fisher (1971), are all important in volcanic sediments; these are slides, sediment gravity flows and suspension fall-out. Individual volcanoes provide point sources which may build up unstable piles of volcaniclastic material. These may shed sediment gravity flows which initiate as debris or grain flows and extend down-slope and across the basin floor as turbidites. The latter may demonstrate diagnostic features such as the doubly graded sequence recognised by Fiske & Matsuda (1964) in the submarine ash flows of the Tokiwa Formation in Japan. In some cases major deposits can be correlated directly with individual ignimbrite eruptions which commonly provide the largest volume of sediment; a fine example is the Minoan eruption of Thera, in the Aegean Sea (Bond & Sparks 1976). Ignimbrites with their zones of welding and reworked mudflows were formerly regarded as diagnostic of subaerial eruption but subaqueous examples have now been recognised. It is believed that these flows moved within a carapace of steam which insulated them and permitted a more complete and uniform welding throughout the full thickness of the sheet (Howells *et al.* 1979; Francis & Howells 1973). Accretionary lapilli remain one of the few unique indicators of subaerial eruption, but even these can commonly be reworked in aquagene deposits.

As the volcanoes build up from the sea bed, contemporaneous intrusion into the wet sediment is not unusual; magma bodies reach a



A. Island arc volcano on subsiding crust (back-arc basin subsiding faster than open ocean)



B. Oceanic island volcano on rising ocean crust with rapid uprise of plutonic core

Fig. 1. Schematic cross section of island volcanoes, to portray the distribution of volcanic sediments. A *represents an island arc volcano, based on the diagrammatic model given by Siggurdson* et al. *(1980). The cross section also illustrates the asymmetry produced by the greater oversteepening of the slopes descending into the back arc basin.* B *represents an oceanic island volcano, based on a diagrammatic model given by Stillman* et al. *(1982). The cross section illustrates the effect of the rise at the ocean floor which is particularly rapid immediately adjacent to the plutonic core of the volcano.*

hydrostatic compensation level within the pile and spread laterally into the incoherent sediment producing a range of effects both in the igneous material which pillows or disintegrates, and on the sediment which becomes fluidised thus destroying the normal bed forms (Kokelaar 1982). At the foot of the volcanic slopes the build up of small fans is achieved, often with contributions from more than one volcano, and a multi-stage evolution of the sedimentary pile is common. Further from the vents, on the floor of the basin, the products of several volcanoes may interfinger and the volcaniclastic flows become intercalated with hemipelagic sediment fed from essentially airfall volcanic dust, which provides the main source of sediment at distances beyond the distal limits of the turbidites. Even more complex interrelationships of volcanic and non-volcanic sediment have been recorded in the western Pacific back-arc basins. Klein (1982) reports the recognition of nine depositional systems in the cores recovered by the Deep Sea Drilling Project, of which debris flow, subma-

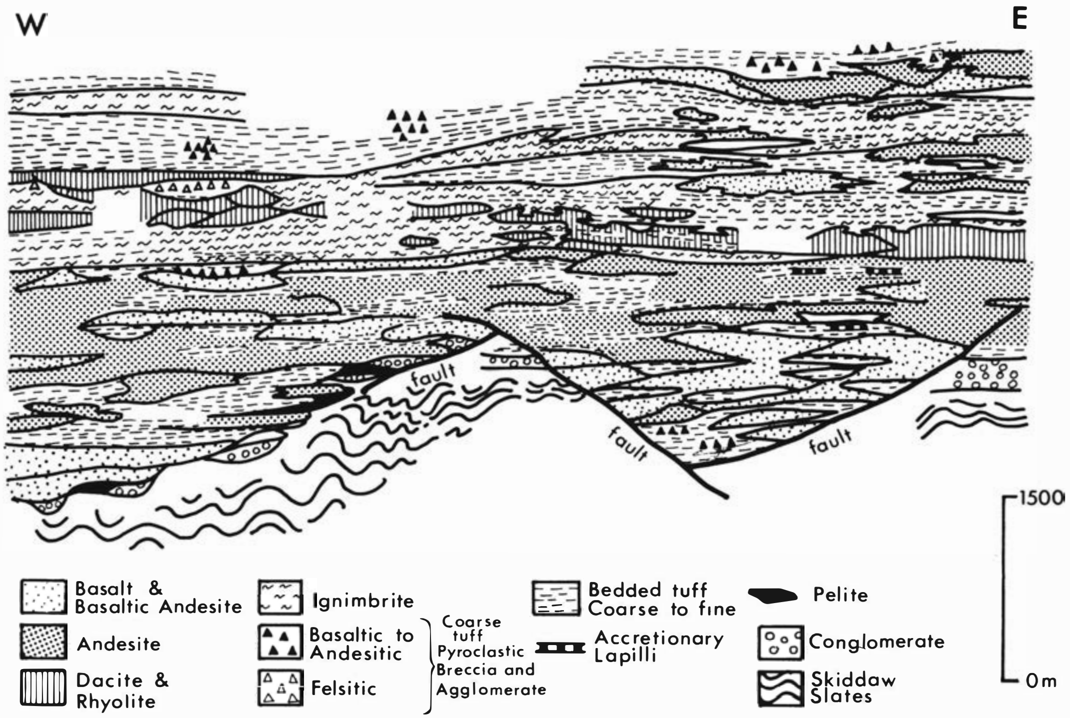


Fig. 2. Schematic representation of a segment of the lower part of the Borrowdale volcanic succession; after Millward et al. *(1978, fig. 37), to illustrate rapid facies changes and intraformational unconformities produced by the derivation of products from a number of separate and overlapping volcanic edifices.*

rine fans, resedimented carbonates and turbid-layer clay deposits occur during times of arc volcanism, of basin spreading and tectonism. Others such as biogenic pelagic systems accumulate all the time, but are dominant only in quiet periods. In these, pelagic clays dominate when the basin floor sinks below the carbonate-compensation-depth; otherwise pelagic carbonates are common. These extensive deposits are distinguished from the restricted island-fringing biogenic reef deposits which build up round the volcanic islands and contain essentially shallow water faunas and floras. Such reefs are often broken up into limestone breccias or conglomerates by seismic activity or by volcanic debris flows into which they may become incorporated. To some extent the bed-forms and the degree of continuity, or lack of it in the sedimentary record depends on the speed with which the submarine volcano builds up, and this in turn seems to relate to whether the volcano has a plutonic core which acts as an upward-moving piston producing extensive intraformational faulting, doming and uplift of the ocean floor (Stillman *et al.* 1982). A significantly different picture may emerge when there is subsidence, which, in arc situations may occur in the back-arc basin but not on the open ocean side of the arc (see Fig. 1).

The depositional environment of a volcanically active continental margin is different in that it is, at least in part, subaerial, and consists of shield volcanoes with abundant parasitic vents. These build up extensive plateaux with intermontane lakes providing scattered lacustrine and deltaic environments, and near the coast, as in Chile in historic times, occasional subsidence provides limited marine incursions. A characteristic of such an environment is extremely rapid lateral facies change with frequent intraformational unconformities and the reworking of volcanic sediments in re-

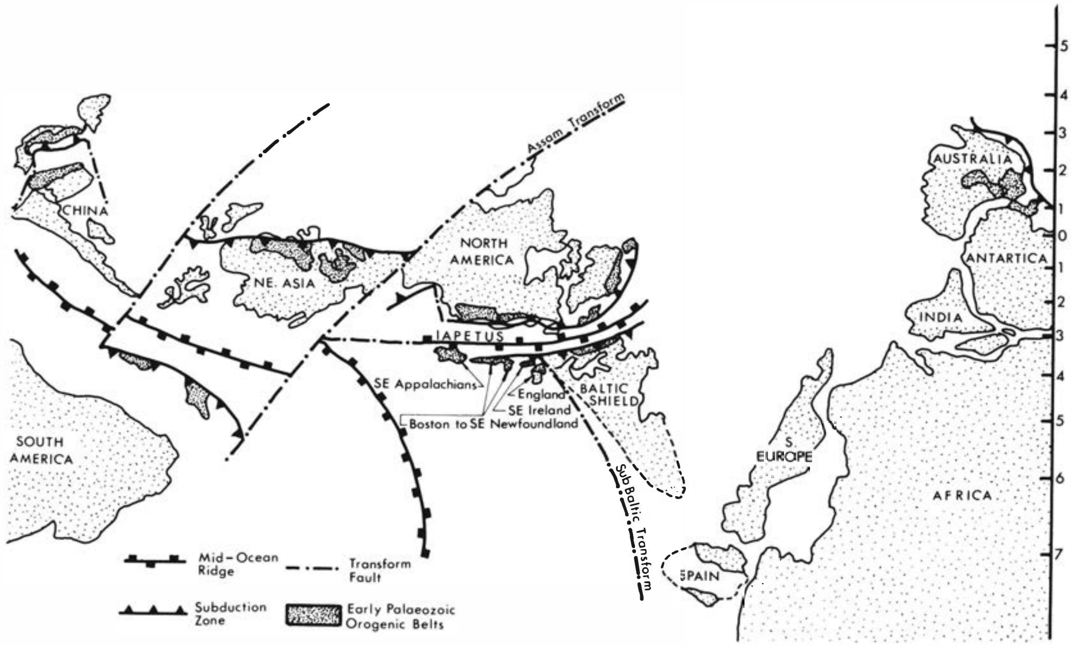


Fig. 3. Early Ordovician world map (on Mercator projection) to show distribution of continental masses, early Palaeozoic orogenic belts, subduction zones transforms and mid-ocean ridges (after Keppie 1977). (The map is centred upon the Iapetus region and part of the Pacific region is omitted.) Ordovician volcanism is largely confined to the orogenic belts, and is subduction-related.

peated cycles of construction and destruction, often overlapping from one centre to another. Such features are well displayed in the upper part of the Ordovician Borrowdale Volcanic Group of the English Lake District which demonstrates close similarities with present day volcanogenic sediments in northwestern USA (Moseley 1982), where, in a complex sequence of events, substantial eruptions of pyroclastics are reworked. However in general the radius of distribution of all but the air-fall ashes is much less than for the equivalent submarine flows and thus the contributions from different volcanic sources provide more rapid intercalation and facies change.

The global setting of Ordovician volcanism

In order to understand the timing and distribution of the Ordovician volcanicity it is necessary to examine briefly the crustal evolution of the world immediately prior to our period of interest. Towards the end of the Precambrian it appears that the continents were amalgamated into a Proterozoic super-continent (Piper 1976). From 1000 Ma to c. 850 Ma this was in a general state of tension with typical intraplate tensional igneous centres: layered intrusions, alkaline complexes, lavas and dykes. Around 850 Ma the super-continent began to crack up (Windley 1977), and by 600 Ma the Baltic and Laurentian plates had separated from the main mass and started to move apart, as had the Antarctic, Australian and Asian plates. The process of extension and rupturing was accompanied by the appropriate volcanic activity with dyke swarms and lavas of tholeiitic basalt; of such are the Tayvallich volcanics of the Scottish Upper Dalradian which Anderton (1982) attributes to the opening of the Iapetus Ocean.

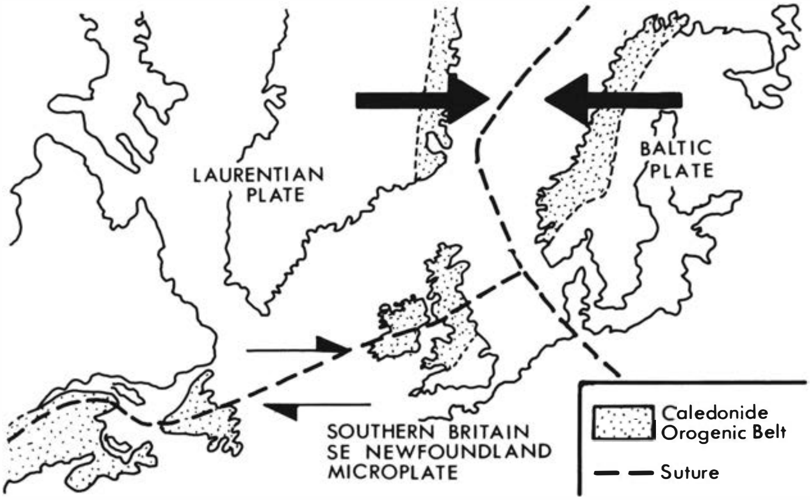


Fig. 4. A Caledonian plate reconstruction of the Iapetus region after closure of the ocean. The plate boundaries are taken from Anderton (1982), but the arrows indicating sense of movement have been added by the author.

New ocean crust to floor the Iapetus and other opening ocean basins was presumably being provided by basaltic mid-ocean rift volcanism, but almost all of this has vanished; traces only are found in scattered ophiolite complexes preserved in obducted slices stacked on the continental plate margins, sometimes in association with accretionary wedges of sediment. This is because by the early Ordovician, the sense of plate motion had changed and the oceans were beginning to close with the consumption of most of the newly formed ocean crust in subduction systems which started in Tremadoc times. Well documented examples are the Ballantrae Complex of the Scottish Southern Uplands (Bluck 1978, Bluck *et al.* 1980) and the ophiolite fragments of western Norway (Furnes *et al.* 1980; Sturt *et al.* 1980).

The processes of closure operating throughout the Ordovician and early Silurian eventually brought together the continental plates to produce a new super-continent, Pangaea. Though the process was complex and the nature and timing of approach of each plate and the tectonic situations on the leading edges was variable, a consistent pattern nevertheless emerges of volcanic activity related to the subduction systems.

Clearly it is not within the scope of this paper to describe all known areas of Ordovician volcanism, thus attention will be focussed on Iapetus, and in particular the British and Irish section of the Caledonides, which contains some of the best known localities (see Stillman & Francis 1979).

Iapetus and Caledonide volcanism

Two aspects of the tectonism of the British Isles sector of the Caledonide orogen are of major significance when discussing the apparent zonation of volcanism as it is preserved today. Firstly, when the continental masses came together to close Iapetus, they did not always approach in a direction normal to the plate boundaries; a great deal of lateral translation must have taken place with one margin sliding past another. Collision and subsequent deformation was not everywhere synchronous nor always compressive; the irregular outlines of the passing continents would have provided opportunity for the generation of pull-apart marginal basins and at a later stage the potential for high level pluton emplacement where thickened crust was wrench-faulted (see Hutton 1982). Secondly, the history of closure and collision of the Laurentian and Baltic plates (see Fig. 4) did not at first involve the crust which underlies southern Britain. This apparently belongs to a micro-plate which was detached from some other part of the African continent and did not enter the "Caledonide" orogenic belt until quite late in the process of closure (Anderton 1982), and then "docked" largely by strike-slip or transpression.

Closure of the Baltic and Laurentian plates however was compressional, and resulted in enormous crustal thickening and the peeling off of nappes which, driven by gravity, slid away from the suture across the Greenland and Scandinavian crust, preserving on the latter dismembered sequences of volcanic and plutonic rocks within the tectonically-bounded stratigraphic packages. A recent compilation by Stephens *et al.* (in press) shows that in the various nappes of the Upper and Middle Allochthon there are preserved relics of pre-Iapetus rifting, Iapetus opening with oceanic crust now seen as ophiolite fragments, and Iapetus closing with subduction-related arc volcanics and even late extensional magmatism developed on newly sutured crust. When reconstructing the ocean margins in which the igneous activity presumably took place, it is necessary to account for the curious asymmetry which resulted in the virtual absence of volcanic rocks in the East Greenland Caledonides until Devonian times, when acid lavas and tuffs were erupted through crust already remelted to produce orogenic granite magmas.

The Caledonide igneous activity in Britain and Ireland marks the approach and "docking" of the southern British microcontinent with the southern flank of Laurentia, south-west of its compressive collision with Baltica, and west of the debatable "Törnquist" line. North of the Iapetus suture, igneous activity relates to the margins of the ancient Laurentian crust, whilst to the south of it, magmatism developed on the edge of a less homogeneous and much younger crustal unit which may well have formed by volcanic arc accretion and cratonisation only a little earlier, in the late Proterozoic (Thorpe 1979, Piasecki *et al.* 1981).

Ordovician volcanicity south of the Iapetus suture in Britain and Ireland

It appears that all Ordovician magmas south of the suture may have been generated on or above a relatively simple southward-dipping subduction system which may have changed its position during Ordovician times. Geochemical and petrographic zonation indicates volcanism on a continental crust which thickened to the southeast (Stillman & Williams 1979). The Eycott volcanics of Arenig-Llanvirn age, rest on the Skiddaw Group on the northern flank of the Skiddaw anticline in the northern Lake District. These submarine basalt and basaltic andesite lava flows are intercalated with occasional more acid pyroclastics, suggesting periodic violent acid eruptions punctuating the quiet effusion of basaltic lavas which built up oceanic shield volcanoes (Millward *et al.* 1978). Chemically the basic rocks are tholeiites characteristic of early stage arc deveopment. They are similar in many ways to the basalts and basaltic andesites of the Ordovician inliers seen north of Dublin in eastern Ireland, which were erupted somewhat later to build up submarine sea mounts and oceanic islands (Stillman & Williams 1979). Earliest eruptives are Llanvirn but the vast majority are Caradoc to Ashgill, and are associated with local shallow water limestones and more widespread black pelagic muds. A Llanvirn plinian pyroclastic eruption at Bellewstown represents the only acid activity here.

Immediately south of the Eycott volcanics are the much better-known and extensive Borrowdale Volcanic Group (Llandeilo to Caradoc) eruptives. These are dominantly andesitic, calc-alkaline, and of an evolved island-arc type, associated with some more acid dacites and rhyolites. The abundant volcanic sediment indicates an initial environment of deposition around a chain of volcanic islands in shallow

Fig. 5. Locality map of the Ordovician volcanic rocks in Britain and Ireland.

seas underlain by continental crust. The islands were dominantly strato-volcanoes composed of interbedded lavas and pyroclastics, together with acid shields of ignimbrite and acid lava. These, despite being continuously subjected to destructional erosion and redistribution in the waters around the islands, some of which, to judge from microfossils (Millward *et al.* 1978) were marine, built up to subaerial plateaux.

In south-east Ireland at this time volcanic activity dominated the infilling of a north-east trending linear basin which had been accumulating the terrigenous clastic sediments of the marine Bray and Ribband Groups. Amongst these are "coticule" beds now thought to indicate ocean-floor metalliferous sediment related to some form of volcanogenic-hydrothermal activity (Kennan pers. comm.). In Llanvirn to Llandeilo times, in what is now the Waterford coastal region, a number of submarine basaltic volcanoes began to erupt island arc transitional tholeiites, then calc-alkaline basalts and basaltic andesites. These built up shield volcanoes with much associated shallow intrusive activity emplacing high-level sills into unconsolidated volcanic muds and hyaloclastites. A pause in the upper Llandeilo allowed the widespread deposition of a calcareous sediment, then renewed igneous activity ejected large volumes of acid pyroclastics as ash-flows, both submarine and subaerial, with occasional andesite and basalt sheets. Quantities of rhyolite were injected as sills, dykes, and domes which sometimes broke surface to provide the setting for "Kuroko" type volcano-exhalitive copper-iron deposits, as at Avoca (Sheppard 1980). Here there is a bimodal association of basic rocks derived from subduction-related calc-alkaline magmas and acid rocks possibly derived from the partial melting of continental crust.

Whilst it seems likely that the south-east Ireland and Lake District volcanoes were parts of arcs situated on the margin of the southern Britain microcontinental plate, a recent very thorough re-examination of the evidence by Kokelaar *et al.* (in press) leaves little doubt that the Lower Palaeozoic Welsh Basin was actually founded on this continental crust and developed by extensional mechanisms, behind the arcs. At the very beginning of the Ordovician, in late Tremadoc times, a major graben controlled by north-east trending faulting and filled with Cambrian marine sediments was subjected to tectonism and associated subduction-related calc-alkaline volcanism, as seen at Rhobell Fawr. The subduction system would appear to have changed position in Lower Ordovician times, as the focus of arc volcanism moved to the southeast Ireland-Lake District centres and was succeeded in Wales by mainly tholeiitic magmatism emplaced in an environment of back-arc extension. This, with the exception of the Llanvirn Fishguard Volcanics (Bevins 1982) seldom fractionated to produce significant volumes of intermediate or acid rocks. The distinct bimodal character of Welsh volcamism is, in fact, provided by the addition of voluminous eruptions of rhyolite which again is interpreted as resulting from crustal fusion. Centres of volcanism migrated both in time and space from the Arenig to Llanvirn activity of southern Wales to the pre-Caradoc volcanism of southern Snowdonia and then the Caradoc episode of central and northern Snowdonia. In the Welsh borderlands and the Lleyn, activity of both episodes is represented. In all areas contemporaneous faults often controlled sites of eruption. Intrusive activity is represented by high-level basic sills and small granitic bosses, some of which are clearly coeval with the volcanics, but others may be later.

Ordovician volcanity north of the Iapetus Suture, in Britain and Ireland

No such simple zonation is apparent here. The Laurentian plate had a longer more complex history, which involved, during the Lower Palaeozoic, both the approach and interaction with the Baltic plate and the southern Britain microplate. The results were first a Cambro-Ordovician (Grampian) compressive event producing pronounced regional metamorphism and deformation, and a second, end-Silurian (Caledonian) event, in which oblique transpression resulted in strike-slip shuffling of segments of crust with relatively mild internal deformation, but with very extensive batholithic granitic plutonism. The sediments deposited in the Dalradian basins on this plate were affected by both, but early Ordovician to mid Silurian rocks deposited on its southern margin show only the Caledonian events.

There is evidence that the Dalradian supracrustal basins which were being stretched and ruptured in mid-Cambrian were, by Arenig times, fringed by ocean basins to the south. Remnants are seen at Ballantrae, where Bluck *et al.* (1980) have described an island-arc and marginal basin assemblage which evolved in a pre-Arenig marginal basin and was obducted northwards onto a pre-Caledonian continental margin between 470 and 490 Ma ago. The Highland Border Complex which is now known to be, at least in part, of Lower Arenig age (Curry *et al.* 1982) also contains southerly derived debris from a back-arc basin. In Northern Ireland the Tyrone Igneous Complex comprises a suite of high-level basic and intermediate intrusions closely related to volcanic rocks of similar composition in a setting which could well be a marginal basin and island arc. The complex is in tectonic contact with Dalradian metasediments in a manner analogous to that of the Highland Border Complex.

There is evidence not only of the existence of oceanic crust but also of the beginnings of subduction, with its related arc magmatism, at the same time as the Grampian orogenic event was taking place on the continental plate. In Connemara the interaction of the two may perhaps be seen; the Connemara migmatites contain an intrusive magmatic component which Yardley *et al.* (1982) interpret as the roots of an early Ordovician volcanic arc emplaced in Dalradian supracrustals during the Grampian event.

Whilst these deep-seated roots to not have direct volcanic associates, compositionally very similar coeval Tremadoc submarine lavas are found at Lough Nafooey, some 15 km to the north. These are island-arc tholeiites which in th Arenig, give way to more calc-alkaline evolved arc types (Ryan *et al.* 1980). It is not likely that these little-deformed arc volcanics with their pelagic black mud intercalations were erupted in close proximity to the Connemara orogenic zone; they must be situated on a block of crust which was tectonically juxtaposed with Connemara only from Llanvirn times onwards, when clasts of Dalradian lithology were transported from the south into the Maumtrasna Group conglomerates. The volcanics apparently initiated as an arc on the southern border of the South Mayo Trough, which evolved throughout Ordovician and Silurian times, during which the volcanicity was, like the Welsh Basin, bimodal acid and basic. Here again it seems that the acid magma may be the product of crustal melting, induced as the crust was thickened.

Similar Arenig volcanics and marine sediments are seen at Charlestown, 45 km to the northeast, where a mineral deposit, apparently of porphyry copper type, is associated with high-level quartz feldspar porphyry intrusions. Such deposits are characteristic of continental margin volcanic areas and again suggest some crustal involvement in the magmatism.

All these early Ordovician volcanic arcs and marginal basins fringe the Dalradian continent and it appears that an active continental margin with subduction-related arc magmatism and extensional marginal basins was established on this segment of the Laurentian plate boundary as early as the beginning of the Ordovician.

Immediately to the south, there is now to be found an accretionary prism of stacked, often inverted, slices of sediments, volcaniclastics and lavas which make up the Southern Uplands. This sequence, lying immediately to the south of the Ballantrae ophiolite, has been interpreted as comprising ocean-floor assemblages obducted northward onto the continental margin (Leggett *et al.* 1979). In the northern belt are basaltic lavas overlain by metalliferous sediments, cherts, graptolitic mudstones and greywackes, largely of Arenig age. Geochemical studies (Lambert *et al.* 1981), indicate a variation from alkaline to tholeiitic in chemical character from north to south through a series of tectonic slices, leading these authors to suggest that the Ordovician sequences were deposited on oceanic crust, alkaline in the north and tholeiitic in the south. They also suggest that this oceanic crust continued to be the basement throughout Ordovician and into Silurian times.

Along the northern edge of the Longford-Down extension of the accretionary prism into Ireland, Llandeilo to Caradoc pillow lavas, found on the northern shores of Belfast Lough at Helens Bay (Sharpe 1970, Craig 1982) and at Stokestown (Morris 1981), appear to have erupted in spreading marginal basins, and in South Connemara Ordovician basaltic pillow lavas are of apparently similar origin (Ryan *et al.*, in press). Although no actual fragments of oceanic crust have been recorded, all the Ordovician lavas are of extensional ocean floor origin.

However, also within the sediments, are volcaniclastic components which become dominant in the younger Ordovician sequences, and which appear to have been derived from the north. The succeeding Silurian beds also contain traces of volcanic material (Phillips & Skevington 1968, O'Connor 1975, Cameron & Anderson 1980), which is largely pyroclastic and may best be explained as the product of subduction-related arc volcanism, the source region of which is not clear but may well be to the north. This rather tentative evidence of a volcanic arc north of the accretionary prism which continued its activity into the Silurian is significant when the situation in Scotland is examined. Here though the Ballantrae conglomerates contain northerly derived clasts of volcanic arc rocks (Bluck 1978), the supply of these had apparently ceased by the end of the Llanvirn. The Southern Uplands Fault must have been active by this time as the Ordovician arc which supplied the clasts can scarcely have been in the region now occupied by the Midland Valley. The dislocations and juxtapositions such as are suggested for the Southern Uplands and for Connemara, which took place during the period of Iapetus closure, illustrate the complexity of the Laurentian plate margin that evolved during that closure. They may go far to explain the absence of a clearly defined geochemical zonation in the volcanic rocks, such as is seen in the subduction-related Ordovician volcanics of England, Wales and south-east Ireland, south of the Iapetus suture.

This detailed account of one small sector of just one plate collision zone should serve to indicate that Ordovician volcanity demonstrates a period of active plate motion. It also suggests that whilst geochemical descriminants provide perhaps the most powerful tool for the recognition of tectono-magmatic setting, palaeogeographic reconstructions may be better analysed with the help of the eruptive and depositional character of the volcanics which can effectively demonstrate the physical environment.

A final and cautionary comment must be that the widespread distribution of Ordovician volcanism virtually precludes its use in long range correlation. Similar tectono-magmatic settings on different plate boundaries can, and do, produce identical volcanic sequences in provinces which are geographically quite distinct and perhaps formed at different times.

Acknowledgements

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Faunas in a volcaniclastic debris flow from the Welsh Basin: A synthesis of palaeoecological and volcanological observations

By MARTIN G. LOCKLEY

Investigations of a diverse, Llanvirn-age faunal assemblage from peri-insular volcaniclastic sedimentary rocks in Central Wales, provide evidence of palaeocommunity mixing by submarine debris flows. The assemblage also contains rare elements, hitherto unknown in this area or at this stratigraphic horizon, which tend to blur distinctions between previously defined faunal provinces.

Together these observations suggest that insular faunas may not be as inherently diverse or endemic as previously supposed. The special sedimentological conditions associated with volcanic islands evidently play a significant role in the mixing and preserveation of faunas. Palaeontological and volcanological observations must therefore be combined in order to evaluate the factors contributing to the marginal rock and fossil accumulates of volcanic islands.

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Recent analyses of palaeocommunity and facies distributions in the Ordovician of Wales and North America (Lockley 1983, Sepkoski & Sheehan 1983) provide useful models for generalised shoreface to open shelf and ocean basin settings. However rocks of the Appalachian-Caledonide Orogen indicate the existence of volcanic islands in the Iapetus Ocean throughout the Ordovician. Such facies contain unusual faunal assemblages variously described as "curious" (Neuman 1976: 13) or "peculiar" (Horne 1976: 1, Bruton & Harper 1981a: 37, 1981b; in press). These faunas are often diverse with many new forms (Neuman 1976) and therefore are frequently regarded as endemic because they cannot be readily compared with those from standard platform successions. Since the general models of Lockley (1983) and Sepkoski & Sheehan (1983) do not embrace island settings, these unusual peri-insular faunas need to be explained in the context of the special environmental and sedimentological conditions which prevail around volcanic islands.

Investigations of a diverse, well-preserved faunal assemblage from peri-insular volcaniclastic sandstones of the Builth-Llandrindod inlier (Jones & Pugh 1949, Furnes 1978) provide evidence that submarine debris flows played a significant role in:

1) mixing and redistribution of well-documented faunas, including most elements of the *Hesperorthis* and *Dalmanella* palaeocommunity spectrum,
2) preservation of valuable "census" samples through rapid burial.

This latter phenomenon appears to have enhanced the quality of preservation. Rarer discoveries, including forms like the brachiopods *Porambonites, Parastrophinella, Mcewanella* and *Christiania* (Lockley & Williams 1981), *Kullervo, Rhynchorthis* and the trilobite *Atractopyge* (this paper) add significantly to our knowledge of faunal distributions. Moreover many of the above-listed forms, now known from the Anglo-Welsh province (*sensu* Williams 1969, 1973), have traditionally been ascribed Celto-Baltic affinities. This blurring of provincial differences by new finds suggest that current concepts of endemism may in some extent reflect an incomplete knowledge or record of the distribution of taxa. Perhaps the endemism is a little more apparent than real.

In Bruton, D. L. (ed.), 1984. *Aspects of the Ordovician System.* 195–201. Palaeontological Contributions from the University of Oslo, No. 295, Universitetsforlaget.

The relationship between faunal assemblages and volcanogenic sediments in the Builth-Llandrindod area

Ordovician rocks of the Builth-Llandrindod inlier have long been the subject of attention and are generally considered to represent a classic example of a volcanic island complex (Murchison 1833, 1839; Geike 1897; Elles 1939; Jones & Pugh 1941, 1948, 1949; Hughes 1969, 1971, 1979; Williams 1969; Williams *et al.* 1972). However, recent studies (Lockley & Williams 1981, Williams *et al.* 1981, Sutheren & Furnes 1980, Furnes 1978 and pers. comm.) have shown that much is still unknown about the faunas and volcanogenic sediments in which they are entombed. Although palaeontology and volcanology are traditionally separate geological subdisciplines, integration of these areas of investigation considerably enhances our understanding of Welsh Basin palaeonenvironments (see Fig. 1 for location of sampled succession and volcanic rocks).

Faunas recovered from various stratigraphical units in the Howey Brook section (Murchison 1839, Elles 1939, Lockley & Williams 1981) include a diverse assemblage (Table 1 herein and Williams *et al.* 1981: Table 4) which Furnes (1978 and pers. comm.) interprets as occurring in a submarine debris flow deposit. According to Furnes the Howey Brook succession consists of a fining upwards sequence of massive and graded sandstones and black shales, respectively representing proximal, mid and distal fan deposits. The sandstones generally lack sedimentary structures but contain "rip up" clasts of black shale indicative of submarine erosion. Such deposits are essentially similar to those described by Carey & Sigurdsson (1980) and Sparks *et al.* (1980); see Figs 2 and 3 herein. Continued investigation of this locality and a consideration of palaeocommunity interrelationships (Lockley 1983) supports the debris flow interpretation and provides further valuable palaeontological evidence with a variety of implications.

The new evidence includes the discovery of a linguilid brachiopod and species of *Petro-*

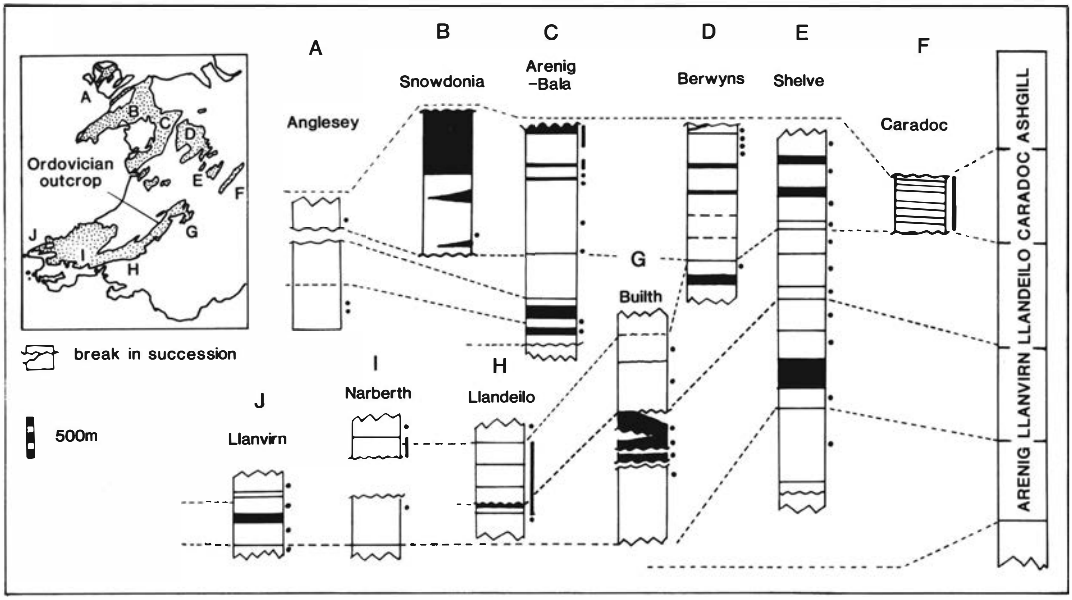

Fig. 1. Type and classic Ordovician (Arenig-Caradoc) sections in Wales and the Welsh borderland (modified after Lockley 1983, Text-Fig. 1) showing the widespread distribution of volcanic rocks and volcanogenic sediments (in black). Vertical bars indicate continuously sampled sections, dots indicate other known faunas.

crania, Rhynchorthis, Kullervo and ? *Tritoechia* as well as the trilobite *Atractopyge* and some hitherto unidentified molluscan and bryozoan remains which differ from these previously listed by Williams *et al.* (1981 Table 4). The diversity of this enlarged collection (see Table 1) is now known to be at least 40 species, a value which exceeds those of all other known Welsh Basin assemblages and diverse shelf faunas of Caradoc age (Hurst 1979).

A convincing explanation of this high diversity assemblage can be derived from a knowledge of palaeocommunity and volcanogenic sediment distribution as shown in Fig. 3. Modern studies of late Quaternary volcanogenic debris flows, "the submarine counterparts ... of subaerial pyroclastic flow deposits" in the Lesser Antilles Arc (Sparks *et al.* 1980; Sigurdsson *et al.* 1980) have convincingly shown that flows will travel for "over 13 km offshore". Such movement perpendicular to shoreline (or arc axis) will cut across any shore-parallell zones thereby mixing faunas whose distribution is controlled by depth or contour-parallel facies belts.

The Howey Brook faunal assemblage contains an anomalous mixture of elements from the "shoreface" *Hesperorthis* palaeocommunity and the "open shelf" mixed *Dalmanella* palaeocommunity as well as a few nektic and planktic elements like cephalopods and graptolites (Fig. 3). The sediments themselves also contain rip-up clasts of fine grained dark mudstone similar to that in which inarticulate-dominated, offshore faunas typically occur (cf. Lockley 1983).

Constitutents of each fauna are readily identified and can be separated out (Fig. 3) to show that the assemblage represents a relatively comprehensive census of at least two palaeocommunities. *Hesperorthis* palaeocommunity constituents identified by Williams *et al.* (1981) and Lockley (1983) as Bryozoan, *Hesperorthis* and *Salopia* associations (A_{1-3} respectively of Fig.

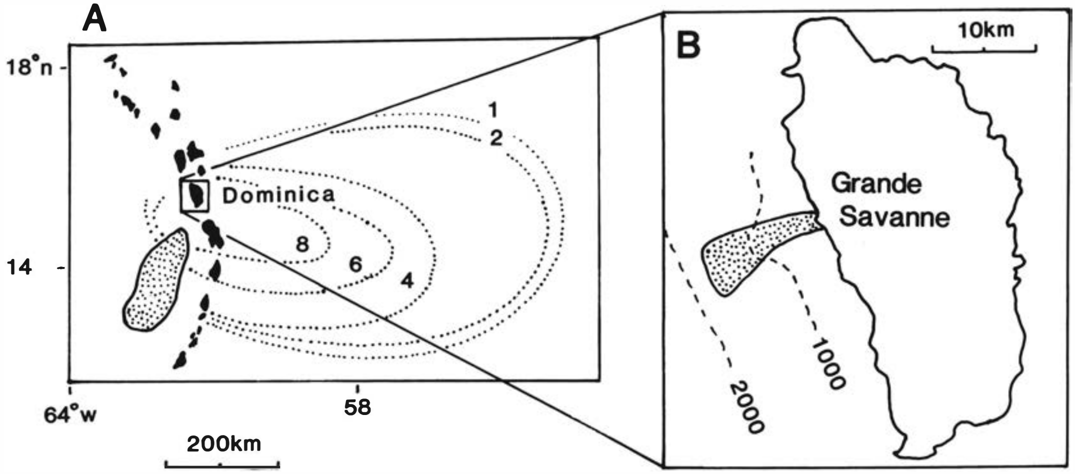

Fig. 2. Distributions of some volcanogenic sediments in the Lesser Antilles Arc lincluding. A. subaqueous pyroclastic debris flows (stippled) and an airfall tephra layer (with isopachs in cm), modified after Carey & Sigurdsson (1980, Fig. 5); B. detail of bathymetry and a submarine debris flow associated with the late Quaternary Grande Savanne pyroclastic apron on the Western flanks of Dominica, modified after Sparks et al. *(1980 Fig. 4).*

3) are all recorded in the Howey Brook assemblage and collectively comprise an estimated 43% of the fauna. The remainder of the fauna is comprised largely of elements like *Glyptorthis, Dalmanella* and *Macrocoelia* which are typical of the mixed *Dalmanella* palaeocommunity (Lockley 1983). The absence of offshore palaeocommunity elements such as *Schizocrania* may be attributed to two factors. Firstly they show relatively low abundance and diversity, and secondly they appear to be comprised largely of nektic and planktic elements (cf. Lockley & Antia 1980) which would not be severely affected by down-slope debris flow. The rarity of elements from this palaeocommunity is therefore predictable.

Although transported faunas are often regarded as inappropriate for palaeoecological analysis, it appears that the high diversity of the Howey Brook assemblage, and the excellent preservation of juveniles and delicate forms, may reflect the potential of such debris flows for preserving valuable census samples (cf. Cisne 1973). Although the higher diversity estimates obtained from such assemblages are only partly attributable to mixing, which does not necessarily explain the preservation of rare forms, unknown elsewhere, there is a close correspon-

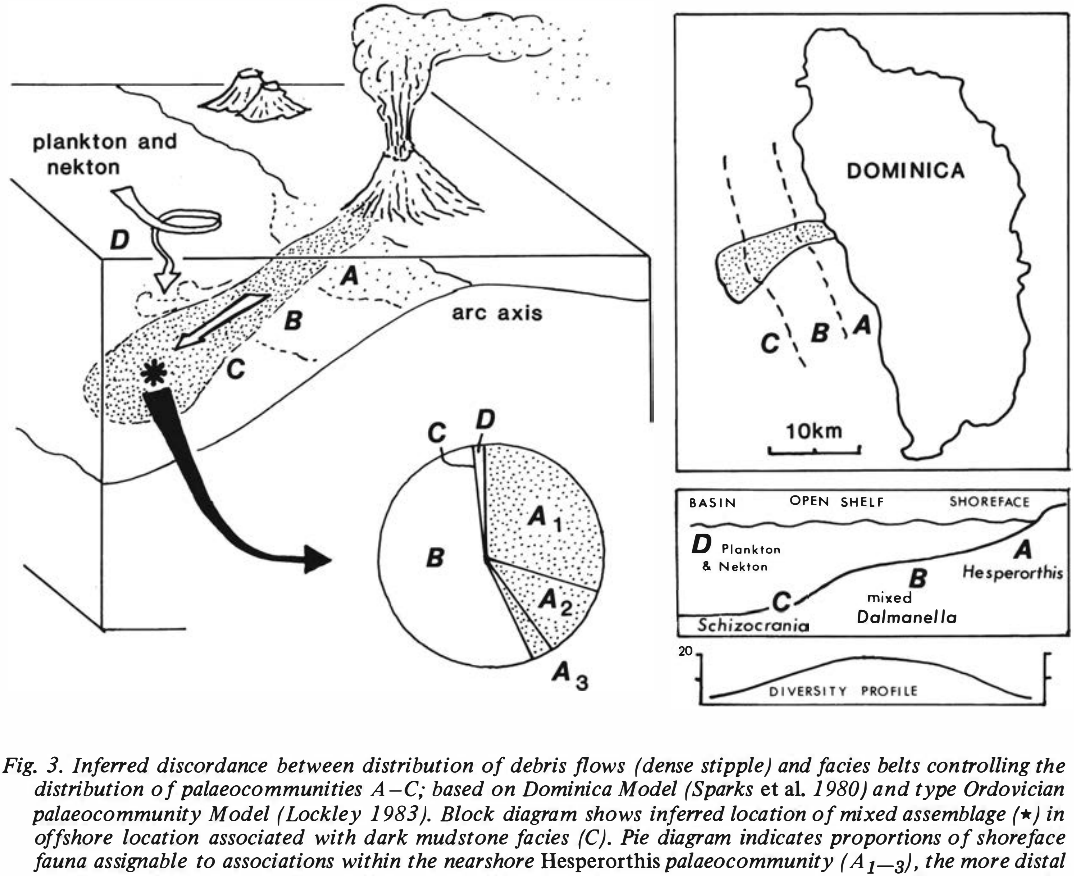

Fig. 3. Inferred discordance between distribution of debris flows (dense stipple) and facies belts controlling the distribution of palaeocommunities A–C; based on Dominica Model (Sparks et al. *1980) and type Ordovician palaeocommunity Model (Lockley 1983). Block diagram shows inferred location of mixed assemblage (⋆) in offshore location associated with dark mudstone facies (C). Pie diagram indicates proportions of shoreface fauna assignable to associations within the nearshore* Hesperorthis *palaeocommunity* (A_{1-3})*, the more distal mixed* Dalmanella *palaeocommunity (B) and nektonic and planktonic palaeocommunities (D); see text for details.*

dence between the total diversity of the Builth sample and the sum of diversities for component palaeocommunities.

Other interpretations

In addition to the taphonomic interpretations given above, reassessment of the Howey Brook fauna permits other conclusions pertaining to faunal province concepts. In particular it is noteworthy that the recently discovered brachiopod genera *Petrocrania, Rhynchorthis* and *? Tritoechia* all occur in Arenig rocks of Anglesey, which also apparently represent peri-insular settings (Neuman & Bates 1978). Consequently the coefficient of association (0.13) between Anglesey and Builth is significantly greater than tha 0.03 value previously estimated from available brachiopod data (Lockley 1983). *Kullervo* and *Christiania* which also occur at Builth are traditionally regarded as Baltic or Celtic province elements rather than of Anglo-Welsh affinity (Williams 1969, 1973).

It is therefore becoming apparent that faunas from volcanic island settings are not necessarily as endemic as hitherto supposed (Williams 1969, 1973; Neuman & Bates 1978). The extent to which some of the recent discoveries in Wales have tended to blur the concept of provincial boundaries can be ascertained from the following list which indicates some of the genera (with age in brackets) recently recorded from Wales, their traditional provincial affinities (Scoto-American, Celto-Baltic or Anglo-Welsh *sensu* Williams 1969, 1973) and the source of information (in brackets):

Protozyga and *Bimuria,* (Caradoc), Scoto-American, (Lockley 1980).
Mesotaphraspis, (Llanvirn), Scoto-American, (Williams *et al.* 1981, Owen & Bruton 1980).
Christiania, (Llanvirn), Baltic, (Lockley & Williams 1981).
Kullervo and Rhynchorthis, (Llanvirn), Celto-Baltic, (This paper).

There is no reason to expect all provincial boundaries to be clearly defined; thus as more data has accumulated, more examples of interprovincial exchange have been recognised. Although he defined provinces on the basis of their distinct faunal composition, Williams (1969, 1973) cited several examples of pandemic genera as well as others which could not be considered strictly endemic. Neuman (1972, 1976) and Neuman & Bates (1978) have already commented on the similarities between American and Celtic faunas, while a view from the Welsh side of the Proto-Atlantic highlights Celtic–Welsh affinities. Current investigation of Arenig rocks in Wales suggest that American-Welsh affinities may also prove to be much greater than hitherto supposed (M. G. Bassett, pers. comm.).

Conclusions

Type and classic Ordovician successions of the Welsh Basin must be viewed in the context of their probable palaeogeographic setting which is generally interpreted as a back arc basin (or marginal sea) dominated by island arc volcamism (Woods 1974, Windley 1977). In such distinctive geological settings local sedimentation patterns are strongly influenced by specific factors including volcanism itself, slope and prevailing wind. It is therefore desirable to view faunas entombed in volcaniclastic sediments from an appropriate depositional environment perspective rather than attempt to reconcile them with shoreface-open shelf models which are appropriate elsewhere (Hurst 1979, Lockley 1983). Furthermore, palaeoecological and volcanological investigateion should not necessarily be treated as mutually exclusive topics; data can be combined into a coherent model (as in Fig. 3) in order to provide a plausible explanation of coeval phenomena. When this was done for the Howey Brook fauna, hitherto unrelated volcanological and palaeoecological investigations appeared to corrobate each other, and demonstrate that peri-insular volcanic islands may have their own special faunal distribution patterns.

Acknowledgements

I should like to thank Drs. H. Furnes and C. J. Stillman for their advice. New brachiopods in the Howey Brook collection have been designated British Museum Numbers BB 65620–24.

Table 1. The composition of a diverse faunal assemblage from Upper Llanvirn volcaniclastic debris flow deposits exposed near Howey Brook (SO 092591), Mid Wales. Brachiopod, mollusc and trilobite diversities are estimated at 15, 10 and 5 respectively. (In other taxonomic categories, where it is hard to estimate number of individuals, figures are given in brackets, or simply as N.)

Taxon	Estimated number of individuals
BRACHIOPODA	
Lingulida indet.	1
Petrocrania sp.	1
Hesperorthis dynevorensis Williams	62
Glyptorthis cf. *viriosa* Williams	25
Mcewanella berwynensis MacGregor	9
Rhynchorthis sp.	1
Dalmanella parva Williams	95
Tissintia prototypa (Williams)	2
Salopia turgida (M'Coy)	15
? Tritoechia sp.	1
Kullervo sp.	2
Macrocoelia llandeiloensis elongata Lockley & Williams	98
Christiania elusa Lockley & Williams	25
Porambonites sp.	2
Parastrophinella parva MacGregor	1
Total Brachiopoda	340
MOLLUSCA	
Lyrodesma sp.	11
Similodonta sp.	3
Pterneid indet.	1
Modiomorphid indet.	1
Matherella acuticostata Bates	1
Helical gastropod indet.	2
Sinuites sp.	1
Ornamented bellerophontid	1
Orthoceras vagens Salter	2
Cyrtoceras macrum Blake	1
Total Mollusca	24

Taxon of individuals	Estimated numbers
TRILOBITA	
Bettonia chamberlaini (Elles)	2
Atractopyge sp.	1
Calymenid indet.	1
Odontopleurid indet.	1
Total Trilobita	7
OTHER TAXA	
Fenestrate bryozoan	(6)
Ramose bryozoan (Morphotype 1)	(157)
Ramose bryozoan (Morphotype 2)	(3)
Boring organism	(1)
? Conularida indet.	(2)
Hyalostelia fasicula (M'Coy)	(N = 1)
Crinoidea	(N = 1)
? Cystoidea	(N = 1)
Spicules (? Porifera)	(4)
Didymograptus sp.	6
Estimated total number of individuals	553
Estimated minimum diversity	40

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Lower Ordovician volcanism in North West Argentina

By FLORENCIO GILBERTO ACEÑOLAZA and ALEJANDRO TOSELLI

Outcrops of Lower Ordovician volcanic rocks occur over large areas of north west Argentina and are best studied in the regions of Famitina and Puna. In the former region the rocks vary from andesites to rhyolites in composition and form tuffs, lava flows and volcanic breccias. The main outcrops are found in the El Chuschin–Cuchilla Negra, El Cachiyuyo, Narvaez-Chaschuil and Les Planchadas. Volcanics occur in the Suri Fm. (*s.l.*), called locally the Morado and Las Planchadas formations. Lower levels contain *Tetragraptus approximatus* and upper levels *Hoekaspis (= Merlinia) megacantha.* The volcanics are therefore of Arenig age. In the Puna region the volcanics are mesosilicic and acid to alkaline forming lavas, often with pillows, tuffs, ignimbrites and spilites. In the Quichagua area the volcanic sequence is 600 m thick. Similar outcrops to the above occur in southern Bolivia and Peru.

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Volcanic rocks and associated Ordovician sediments in north west Argentina were first known through the work of Stelzner (1885) who noted the occurrence of quartz porphyry and intercalated tuff in the Sistema de Famatina rocks then though to be of Silurian age. Bodenbender (1916) recognised "interbedded diorites" in similar rocks and Penck (1920) described vulcanites of "Devonian" age from the northern part of Famatina. Recent studies by Schwab (1973), Coira (1973), Koukarsky & Mirre (1974) at La Puna, and by Lavndaio (1973) and Maisonave (1973) at Famatina, now consider the volcanic rocks and associated sediments to be of Ordovician age.

These studies, together with those of Aceñolaza & Toselli (1977), allow the recognition of major Ordovician volcanic events in north west Argentina and other parts of South America including Bolivia and Peru.

Ordovician Palaeogeography

In order to interpret the palaeogeographical situation of the Ordovician volcanic events it is convenient to take into account the structural framework in which the Ordovician basin in north west Argentina developed. The principal emergent areas in the west-central region of South America was the craton of Arequipa, made up of Precambrian crystalline rocks giving radiometric ages of between 600 and 1900 million years. This probably extended between the regions of Paracas in Peru and Antofagasta in Chile. In the east the continent was defined by the Dorsal or Cratogeno Central Argentino (Bracaccini 1960), consisting of igneous and low grade metamorphic rocks of Upper Precambrian to Lower Cambrian age (Aceñolaza & Toselli 1981), and by the Dorsal Charata-Asuncion of similar age and composition. The Dorsal Central Argentino formed a long wedge which divided the marine basin into the Cuenca de Pampasia on the east and the Cuenca de la Puna, Famatina and Precordillera (Cuyo) to the west.

In the Cuenca de Pampasia the sedimentological and palaeontological evidence suggests shelf conditions with deposition of fine and medium clastic sediments. In the Puna region the combination of turbidite facies and ophiolites indicates a deeper unstable basin with the formation of rhyolite-keratophyre-type volcanics adjacent to the shelf. In Famatina a shelf-slope

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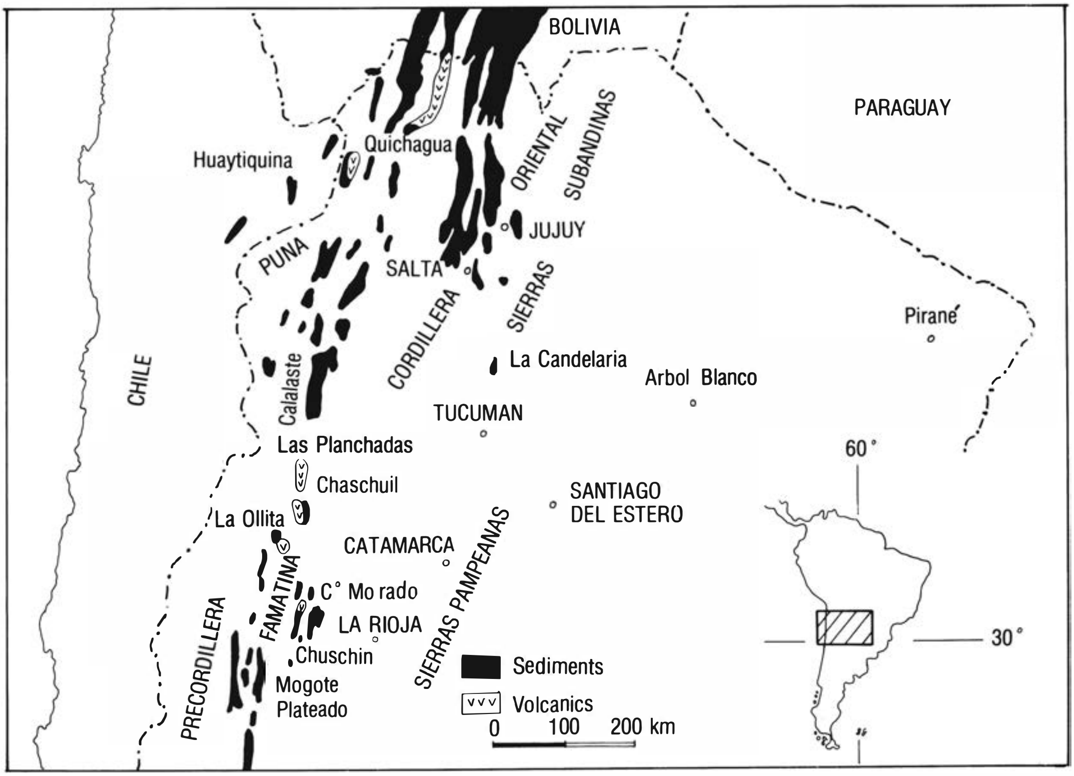

Fig. 1. General map of north west Argentina showing distribution of outcrops of sedimentary and volcanic rocks and position of localities mentioned in the text.

sequence with volcanics is present while in the Precordillera a shelf carbonate sequence grades into a flysch ophiolite facies along the western border (Baldis 1978).

Outcrops of volcanic rock

Outcrops of Ordovician volcanic rocks occur in several sections in north west Argentina extending from the Precordillera (33°S) to the border with Bolivia (22°S).

Puna of Salta and Jujuy areas

The studies of Schwab (1973), Coira (1973, 1975, 1979), Koukarsky & Mirré (1974) and Arganaraz *et al.* 1973) include detailed descriptions of the petrography of the Ordovician volcanic rocks in these areas. Spilite-keratophyre sequences are known in the Cochinoca-Escaya, Quichagua-Queta, Cauchari-Huaitiquina ranges and San Antonio de los Cobres, but one ophiolite is known from the Salar de Pocitos.

Quichagua-Queta and Cochinoca-Escaya ranges (66 00W–22 40 S)

Outcrops of Ordovician volcanic rocks in these ranges are almost continuous down their length. Although known for several decades, it is only from the recent studies of Coira (1973, 1975, 1979), that their composition and genetic aspects have been evaluated. These rocks are associated with a sequence of greywackes and shales (Formacion Acoite) of Arenig age and contain *Didymograptus nitidus* (see Steinmann & Hoek 1912), *D. protobifidus* and *Tetragraptus fruticosus* (see Aceñolaza 1981). Coira estimated the thickness of the Ordovician sedimentary sequence to be some 200 m consisting of grey-green greywackes interbedded with lavas, tuffs of spilitic, dacitic and rhyolitic composition.

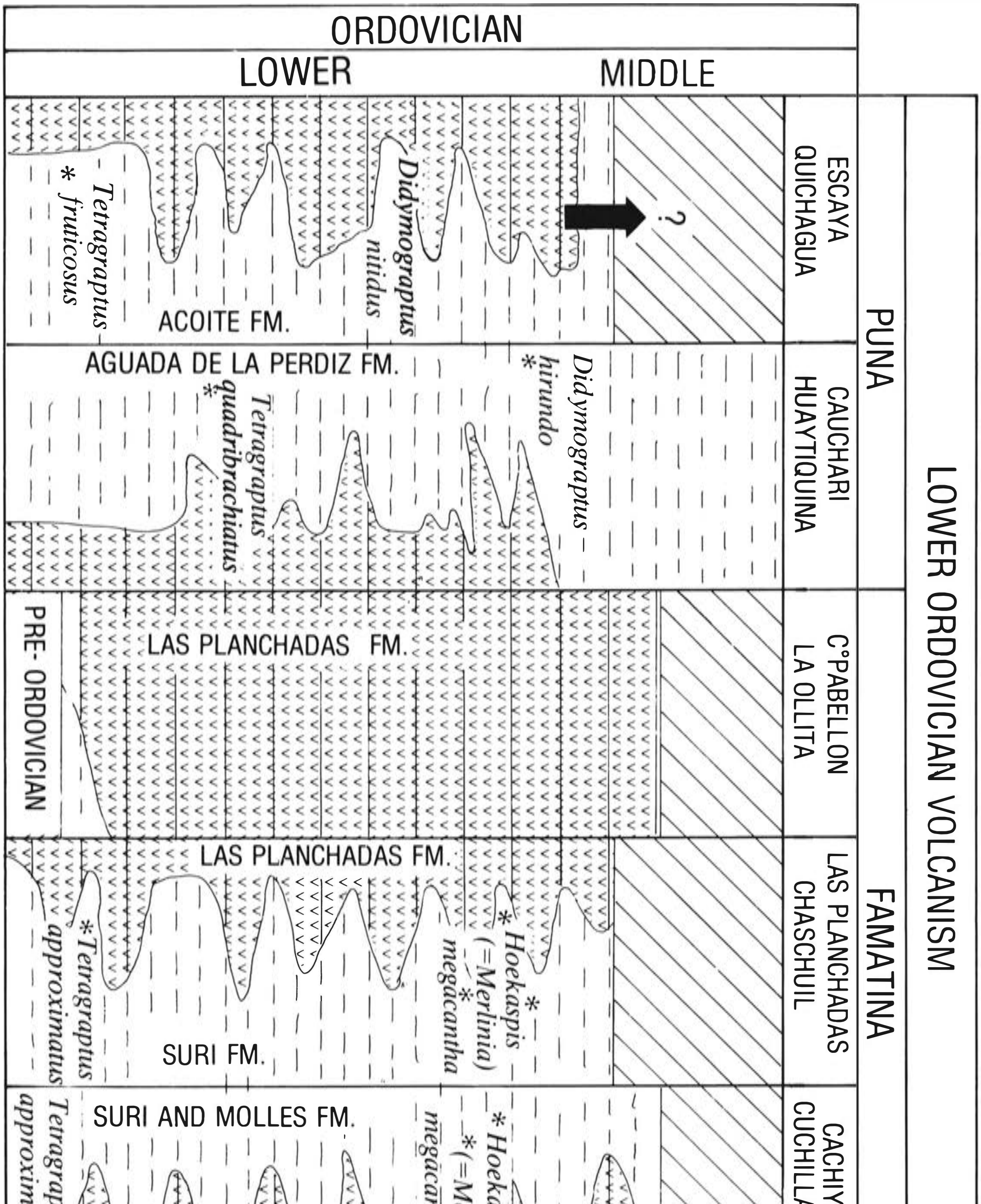

Fig. 2. Summary of Lower Ordovician stratigraphy and vulcanology and occurrence of Arenig graptolites and trilobites.

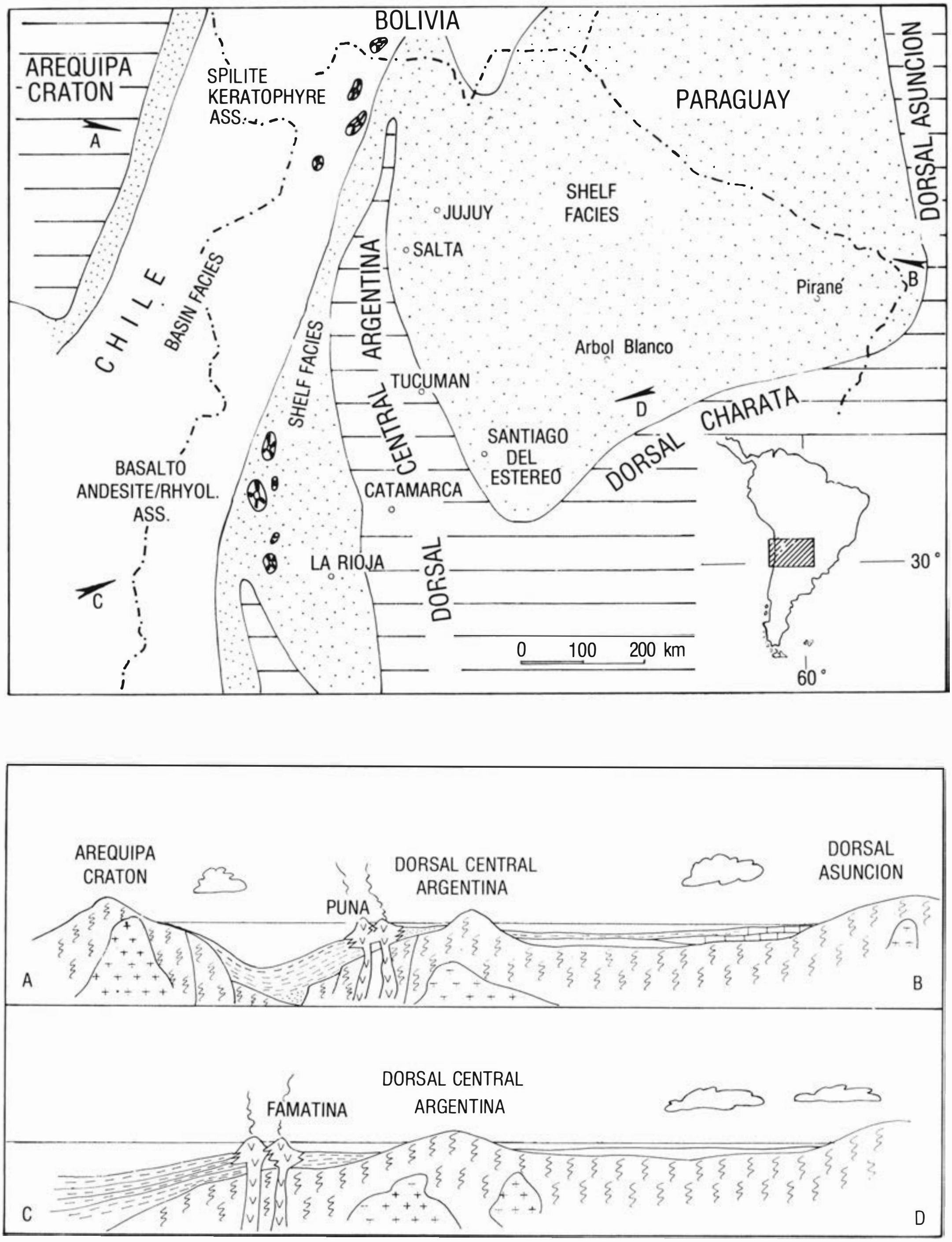

Fig. 3. A, Map of north west Argentina and adjacent areas to show general Lower Ordovician palaeogeography. B, Cross section profiles along line A–B and C–D.

Coira (1979) calculated the volcanic rocks to be about 600 m thick in the Sierra Quichagua. The outcrops of Sierra de Queta and Cochinoca are similar with massive green coloured spilite horizons up to 18 m thick and ash flow tuffs and lava flows with a total thickness of 550 m.

In general terms the spilites are described as porphyry rocks. Coira (1979) distinguished two types of tuff, fine to pumice and thick tuffs rhyolitic to dacitic in composition. The ash flow tuffs and lava flows constitute well differentiated massive bodies.

Cauchari-Huaytiquina Ranges
(67°00'W – 21°20' S)

These mountains are found to the west of Salar de Cauchari and extend to the Argentine–Chilean border. Schwab (1973) estimated the thickness of the Ordovician succession to be in the order of 3000 m of volcanics, cherts greywackes and shales, the latter containing the Arenig graptolites *Tetragraptus quadribrachiatus* and *Dichograptus octobrachiatus* in the zone of Catua (Acenolaza & Durand 1975) and *Didymograptus* cf. *D. hincksii* in the Filo Tropapete. The volcanic rocks include diabases, diabase tuffs, porphyritic andesites and spilites.

San Antonio de Los Cobres
(66°20' W – 24°20' S)

To the west and south west of San Antonio de los Cobres, Koukharsky & Mirre (1974) reported the occurrence of keratophyre tuffs near where Rolleri & Mingramm (1968) discovered Upper Arenig graptolites including *Didymograptus hirundo.*

Sistema de Famatina

The Sistema de Famatina includes a series of mountains extending from Llanos in La Rioja south to the Cordillera of San Buenaventura (27°25' S). Ordovician volcanic rocks occur in the zone of Las Planchadas-Chaschuil, Cerros Pabellon-La Ollita, Cachiyuyo-Cerro Morado-Cuchilla Negra, el Chuschin-Miranda and el Mogote Plateado, and have been interpreted in a series of papers by Maisonave (1973, 1979) and Acenzola & Toselli (1977). The Las Planchadas-Chaschuil Range outcrops extended in a north-south belt approximately 200 kms long by 20 km wide. The oldest rocks in Las Angosturas contain *Clonograptus* and consist of shales and sandstones interbedded with andesites (Acenolaza 1978). In the Chaschuil area the Suri Formation consists of shales, cherts and tuffs, the latter containing a shelly fauna including *Actinodonta, Proterocameroceras, Orthambonites, Annamitella* and *Merlinia* (see Aceñolaza & Toselli 1977).

Cerros Pabellon – La Ollita
(68° 15'W – 28° 05'S)

Extensive outcrops of volcanic rocks of a dacitic-rhyodacitic composition occur in the Catamarca–La Rioja area between the Cerro La Ollita and Las Lajitas. These extend for more than 15 km in a SE–NW direction with a width of 6 km. Maisonave (1979) referred them to the Planchadas Formation noting their similarity to the Suncho Formation in the region of El Suncho gully.

Cachiyuyo–Cerro Morado–Cuchilla Negra (67°45'W – 28°45'S)

Outcrops occur in the Sierra de Famatina, in a region extending from La Cuchilla Negra to the west slope of Portezuelo de las Minitas, and from Cerro del Inca–El Tocino–Morado–El Pelado on the eastern flank of the range. The volcanic rocks consist of some 400 m of rhyodacites to trachytes assigned to the Morado Formation by Turner (1964). According to Turner these rocks in the region of Rio Cachiyuyo are younger than those contining *Merlinia megacantha* but in the southern part they extend down to include levels with *Tetragraptus approximatus.*

El Chuschin–Miranda–Mogote Plateado

The sections here occur on the west flank and middle portion of the Sistema de Famatina and in the Chuschin area consist of a sequence of sandstones and black shales interbedded with tuffs and dacitic lavas (Schalamuk *et al.* 1981).

Interpretation

We agree with Coira (1979) that the Puna spilite-keratophyre sequence and associated sediments with graptolites, indicates a continental slope to basin facies belonging to an early orogenic suite. The presence of serpentinite bodies and gabbroic rocks in Salar de Pocitos, Vega del Cajero, Vega de Tamberias, Antofalla, Sierra de Calalaste and Filo de Copayo (Arganaraz *et al.* 1973; Allmendinger *et al.* 1982), are however associated with a later orogenic stage. Observations by Van Bemmelen (1963) and Souther (1967) on ash flow tuffs, indicate that these eruptions accompany periods of general elevation of the geanticline and related periods of strain and faulting. This stage coincides with the Famatina event. In the Sistema de Famatina, the sediments indicate a shallow shelf environment associated with a synorogenic rhyolite suite of volcanic rocks.

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<tr><td rowspan="3">*Dictyonema spp.*</td></tr>
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Correlation of the British, Baltic and North American series with the Atlantic graptolite and conodont zones. Modified from Bergström 1977 and Jaanusson 1979.

aspects of the Ordovician System

On present estimates, the Ordovician System lasted approximately 70 million years. During this time epicontinental seas were widespread and land areas small. Continental relief was subdued and low run-off led to a limited supply of terrigenous sediment into seas where carbonate sedimentation was extensive. Among the marine faunas a number of major animal groups appeared or became common for the first time and biogeographical distribution was pronounced.

The Ordovician north pole lacked an ice-cap but one was present at the south pole. This caused climatic zonation in the southern hemisphere, global fluctuations in sea-level and an extensive glaciation at the end of the Period.

Movement along continental plate margins was accompanied by sustained periods of volcanic activity more intense than in almost any other geological epoch.

These topics and others were discussed at the IV International Symposium on the Ordovician System held in Norway in August 1982. This book contains 20 of the invited thematic lectures presented on two days of the Symposium and it is hoped that their content and style of presentation will be of general interest to the specialist and non-specialist alike.

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