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 $60^{\circ}S$ may have been near $+\ 8^{\circ}C$, 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

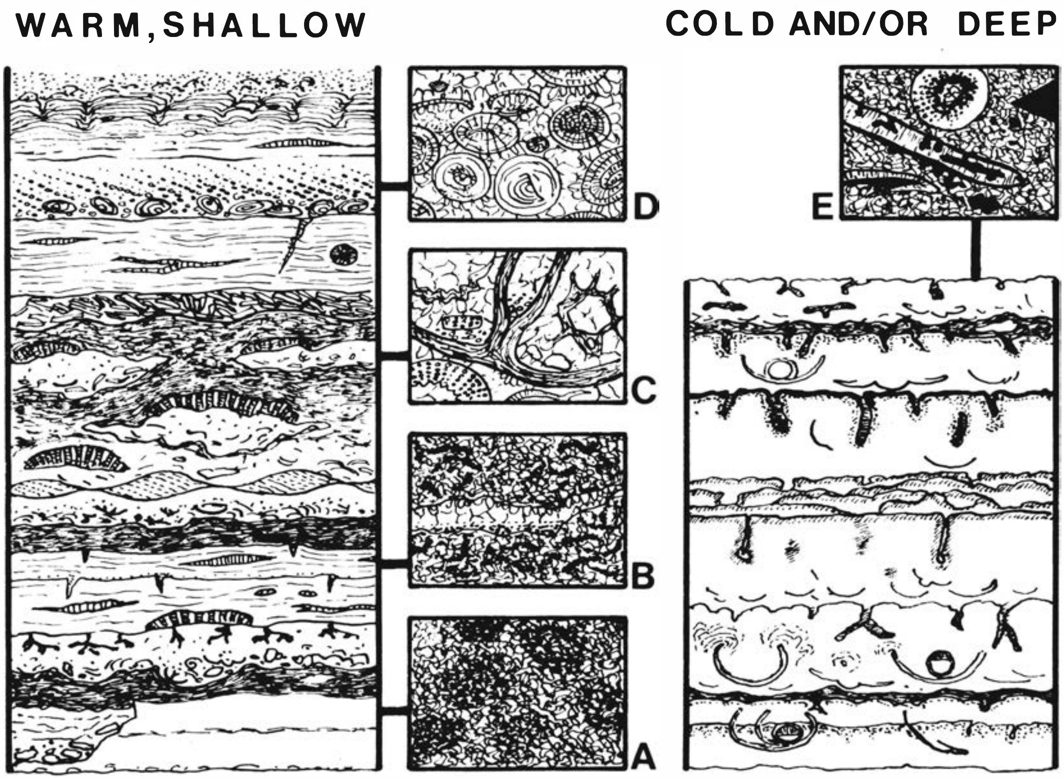

Fig. 1 – Comparison between features of carbonate deposits formed in warm, shallow seas and those formed in cold and/or deep seas. The warm section (left) may represent about 2 m thickness; the cold, deep section (right) corresponds to about 0.5 m. The left section shows, in ascending order, erosional channel with lag deposit; storm grading with load marks; bored hardground with attached skeletal epifauna; laminate with birdseyes, sheetcracks, and desicitation 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,

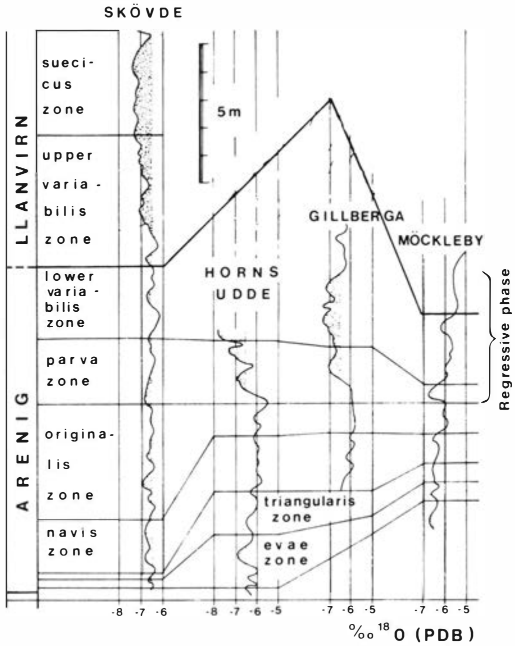

Fig. 2 – ^{18}O variation in four limestone sections in Sweden, with conodont zonation. Stippled zones have values lower than –6.5‰. The curves are based on means for each pair of successive samples. Data from Friedrichsen & Lindström (in prep.).

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 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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