Sedimentological changes across the Ordovician–Silurian boundary in Hadeland and their implications for regional patterns of deposition in the Oslo Region

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Basin faulting and global eustatic sea-level fluctuations during the latest Ordovician and early Silurian generated new and complex sedimentation patterns in the Oslo Region. In eastern Hadeland a fall in sea level, possibly in the earliest Hirnantian, resulted in the incision of channels which were later filled by carbonate debris flows and finally became emergent during the period of maximum withdrawal. The drowning of these deposits as sea level rose was followed by the deposition of storm-dominated shelf-derived carbonate and siliciclastic sediments. During this period transport was from the north and east. A subsequent gradual shallowing culminated in local emergence. The early Silurian was marked by new shelf-flooding and by the delivery of siliciclastic sediments, again from the east. In contrast with earlier models, an eastern pattern of derivation is thought to have been a general feature of the basin during the late Ordovician and early Silurian.

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Introduction

Although the Oslo Region has long been important for the study of Lower Palaeozoic palaeontology and stratigraphy, relatively little attention has been given to its interpretation as a developing sedimentary basin, forming as it did on the margin of one of the major cratons bordering the Iapetus Ocean. Only in the last two decades have detailed facies analyses been published and few of these have been linked to petrography.

The predominant view of the basin hitherto has been of a series of concentric litho- and biofacies belts forming during the post-Arenig Ordovician, and deepening eastwards from a high referred to as ‘Telemarkland’ (Størmer 1967; Bjørlykke 1974; Brenchley et al. 1979; Worsley et al. 1983). This model also allows for a northeastward spread of sands in the Oslo–Asker District forming a barrier bar migrating obliquely across the palaeoslope during the latest Ordovician and having a northerly directed shoreface (Brenchley et al. 1979; Brenchley & Newall 1980). The Caledonian nappes which were developing to the west and northwest (beyond ‘Telemarkland’) are generally regarded as having been a major source of siliciclastic sediment delivered to the basin (although they are not shown on the cross-section of Brenchley et al. 1979). Little importance has been attached to sediment derived by erosion of the adjacent Precambrian basement to the north and east.

A similar palaeogeography is believed to have persisted into the Silurian (Worsley et al. 1983) with only a brief episode in the late Llandovery when the palaeoslope in part of the central Oslo Region may have reversed behind a southeastwards migrating shoal (Möller 1989) or crustal bulge (Baarli 1990).

In direct contrast to many of these models, we present evidence for the derivation of both bioclastic and siliciclastic sediments from the east during the late Ordovician and early Silurian in at least part of the Oslo Region. This reinterpretation is based partly on our own observations, but would not have been possible without the significant volume of data presented by others. We believe that our alternative views are consistent with their observations.

Setting

The Lower Palaeozoic rocks of the Oslo Region are preserved within the Oslo Graben, a late Carboniferous and early Permian rift basin. The tectonic and magmatic evolution of the Oslo Rift has been reviewed by Neumann et al. (1992). The Carboniferous to Permian movements partially reflect reactivation of earlier structures (Stanistreet 1983; Swensson 1990) and the rift completes a feature linking southwards with elements of the later North Sea Rift system (see Ziegler 1992 for a review). It also extends northwards to include parts of the Hedmark Basin, one of several NNW/SSE-trending ‘Sparagmite Basins’ regarded as aulacogens associated with the open-
ing of the Iapetus Ocean (see Nystuen 1987 and references therein). Significantly, the succession in the Hedmark Basin is believed to have originated 130–230 km northwest of its present position (Nystuen 1981, 1987). The Lower Palaeozoic successions of the Oslo Region are regarded as predominantly parautochthonous (Bockelie & Nystuen 1985; Morley 1986, 1987) and are believed to reflect a shortening of at least 50 km.

During the late Cambrian, the initial stages of the Iapetus closure were marked by westward-directed (in present-day terms) subduction (Ramsay & Sturt 1986). At this time the passive continental margin of Baltica extended at least 400 km west of the present thrust front (Stephens & Gee 1985) and no effects of this subduction are seen in the Oslo Region. However, to the north a second major deformation, beginning early in the Ordovician, resulted in obduction and eastwards (again in the present sense) thrusting of nappe sequences (Sturt et al. 1978; Roberts 1988). The Oslo Basin was then established as a distinct feature, probably by reactivation of Precambrian lineaments, and eventually influenced by nappe loading to the northwest (Ro et al. 1990).

Cambrian and earliest Ordovician successions in Baltoscandia are widespread and uniform (Bergström & Gee 1985), but from the late Arenig onwards platform sediments were deposited in a series of facies belts (Bruton et al. 1985) which remained distinct until the late Ordovician. Such belts are broad in Sweden and the Baltic countries (the facies belts of Jaanusson (1976)) but are narrow in the Oslo Region where some districts show marked lateral variations. In the late Ordovician and early Silurian, sediment associations became increasingly complex as a result of subduction, nappe loading to the west, and the eastwards translation of nappe stacks. In the mid-Silurian, subsidence rates within the basin increased dramatically as a result of crustal loading by the nappes (Bjørlykke 1983).

The lithostratigraphy of the Ordovician of the Oslo Region has been revised by Owen et al. (1990) and that of the Silurian by Worsley et al. (1983). Størmer (1953) subdivided the region into eleven districts and here we focus on Hadeland (Fig. 1) and on deposition taking place in the latest Ordovician and earliest Silurian (late Rawtheyan to early Aeronian). The picture which emerges from this area contrasts with that of the hitherto better documented Oslo–Asker District to the south, which has been taken as the norm for the whole region.

Lithological descriptions

The stratigraphy and biota of the succession around the Ordovician–Silurian boundary in Hadeland have been described by Heath & Owen (1991). Three formations are represented: the Kalvsjøen Formation, dominated by carbonates; the Klinkenberg Formation, a mixed carbonate-siliciclastic unit, and the Sælabonn Formation, predominantly siliciclastic sandstones. The last two together form the Skøyen Group with the Ordovician-Silurian boundary lying at or near the top of the Klinkenberg Formation (Fig. 2). As elsewhere in the Oslo Region, the succession around the Ordovician–Silurian boundary is only weakly constrained chronostratigraphically. The trilobite Acernaspis appears close to the base of the Sælabonn Formation and may coincide with the basal Silurian Parakidograptus acuminatus Biozone, although a slightly higher level in the Silurian cannot be ruled out (Heath & Owen 1991, p. 104).

Kalvsjøen Formation

The Kalvsjøen Formation (Owen 1978; Heath & Owen 1991) comprises two lithologies: nodular limestones and channelized chaotic limestone breccias. The former are

![Fig. 1. The Lower Palaeozoic outcrops and districts of the Oslo Region (after Størmer 1953).](image-url)
Sedimentological changes, Ordovician–Silurian boundary

Fig. 2. Stratigraphy of the uppermost Ordovician and lower Silurian of Hadeland, Ringerike and Oslo–Asker districts (after Worsley et al. 1983; Baarli 1990; Owen et al. 1990; Heath & Owen 1991). Levels containing channel debris flows (asterisks) and millet-seed sands (open circles) are indicated.

typically thin bedded (2–15 cm) and stylonodular, interlayered with shale and siltstone and with limestone forming 50–70% of a succession about 40 m thick in western Hadeland. The limestones vary from mudstones to grainstones but muddy lithologies predominate and beds are typically poorly sorted or graded, with coarse bases and finer tops. Bioclasts are commonly abraded and include fragments of calcareous algae, which are locally abundant, gastropods, echinoderms, brachiopods, orthocones and bryozoans. These occur with 5–40% angular to subangular silt-size quartz grains. Some lower boundaries of beds are burrowed. Burrows were preserved by nodule formation, but in shaley interlayers were generally destroyed by compaction. No cross-lamination or features such as hummocky cross-stratification are present. These rocks are the broad lithostratigraphical equivalents (Fig. 2) of both the nodular limestones of the Langåra Formation in western Oslo–Asker (Brenchley & Newall 1975) and the interbedded limestones and shales of the Bønsnes Formation of Ringerike (Owen 1979; Owen et al. 1990).

At the top of the Kalvsjøen Formation in eastern Hadeland, a series of coarse chaotic breccias occupy channels cut into the nodular limestones (Braithwaite & Heath 1992). The largest and best exposed of these is at Kalvsjø Quarry. The breccia unit is about 15 m thick, consisting of interbedded wackestones and breccias. The breccias are either clast or matrix-supported. Blocks within them are lithoclasts of millimetres to more than 50 cm diameter and vary from mudstones to grainstones. Much of the biota of the blocks is similar to that of the nodular limestones but corals and bryozoans are more abundant (Heath & Owen 1991) and there are in addition colonial filamentous algae, stromatoporoids, brachiopods, echinoderms and molluscs. Different blocks carry different biotas. Importantly, the diagenetic history recorded in blocks also varies. Cement sequences differ from block to block and cements are commonly truncated at the margins of blocks. The blocks are enclosed in a wackestone matrix which may contain up to 3% silt-size quartz, while the blocks contain only about 1%.

The blocks and wackestone matrix form irregular lenses of breccia 1–3 m thick and several metres in lateral extent (as seen in section in the quarry walls). They are interbedded with wackestones which, while they contain a less diverse biota, may themselves consist of 10–30% bioclasts. These laterally discontinuous bedded wackestones resemble the nodular limestones beneath. Limestone beds 2–4 cm thick are separated by 1–3 cm calcareous shales and form units up to 2 m thick extending laterally up to 30 m. A few grainstones 5–10 cm thick and dominated by echinoderm fragments are interbedded.

In some breccia localities (Braithwaite & Heath 1992) the block deposits are cut by irregular cavities and fractures filled with coarse echinoderm-bearing grainstones. These again show a distinctive diagenetic paragenesis, contrasting with those of both the blocks and nodular limestones.

Klinkenberg Formation

The base of the Klinkenberg Formation is marked by the appearance of the first thick sandstone above the nodular limestones of the Kalvsjøen Formation. However, the sandstone is nowhere exposed in contact with the breccia.
deposits thought to occur at the top of the Kalvsjøen Formation in eastern Hadeland. The Klinkenberg Formation varies from more than 45 m thick in the east of Hadeland, where the exposure is incomplete, to about 24 m at Bjertnestangen in the west. Six laterally impersis-
tent lithofacies are represented (Fig. 3) but boundaries between these are often gradational. The heterogeneity of this unit contrasts with the relative uniformity of the laterally equivalent and predominantly siliciclastic Langøyene Formation in Oslo–Asker.

Fig. 3. Lithofacies distribution within the Kalvsjøen, Klinkenberg and Sælabonn formations across a composite west–east section in Hadeland (after Heath & Owen 1991). Note that Lithofacies 5 and 6 of the Klinkenberg Formation, although widely distributed, are too thin to be represented on this illustration.
Lithofacies 1 comprises lenses of coarse bioclastic calcarenite interbedded with fine-grained quartz sandstones, siltstones and shales. In eastern Hadeland, where the calcarenites are most abundant, the lithofacies is 26 m thick, but it decreases to approximately 14 m in the west. Bioclasts include brachiopods, molluscs, echinoderms, corals and algae, (Verniporella and Girvanella), and form lenses up to 10 cm thick which locally make up 50% of the lithofacies. The fine-grained sandstones and siltstones are typically dark grey and calcareous, sometimes dolomitic, forming beds 10–70 cm thick. Up to 50% of angular to subangular grains are quartz, while 25% are feldspars, dominantly microcline. The sandstones are commonly rich in Chondrites, although lamination is locally preserved. Load casts up to 2 m across and several metres long are present locally, as in a railway cutting 2 km north of Sløvika (NM 797855).

Lithofacies 2 consists predominantly of dolostones but locally contains 20–80% siltstones and sandstones and the base is gradational from Lithofacies 1. The rocks are commonly mottled, partly as a result of burrowing, but partly also in response to dolomitization, so that most fossils are unidentifiable. Mottling is common in the east, but in many western sections dolostone and quartz sandstone alternate on a millimetre scale. The lithofacies varies from 2–32 m thick, generally increasing to the west. However, around Gamme (NM 831912) the top of the dolomite is an erosion surface and within this, moulds of gastropods and other presumed formerly aragonitic fossils are filled with millet-seed quartz sands. Irregular fractures several centimetres wide and up to 1 m deep are present in the surface. Some of these are also filled with millet-seed sands (see Lithofacies 6) while others are lined with coarse baroque dolomite. It is not clear how much dolostone may have been lost in eastern areas by late Ordovician erosion.

Lithofacies 3 is about 11 m thick in south-central Hadeland (Klinkenberg), and thins westwards to about 2 m, interfingerig with Lithofacies 2 and 4. At its thickest the unit consists of regularly bedded pale quartz sandstones and fine siltstones, with low-angle cross-lamination, interfingerig with dolostones and forming diffusely laminated beds 5–30 cm thick. Locally, fragments of echinoderms, brachiopods and corals form wedges of grainstone.

Lithofacies 4 comprises about 16 m of cross-bedded grainstones and siliciclastic sandstones in central and eastern Hadeland. At Klinkenberg (NM 811822) fine sandstones, with angular to subangular grains, alternate with calcarenites to form sets of planar and low-angled cross beds, trough cross-bedding and climbing ripples in sets up to 50 cm thick. Small-scale structures commonly fill scoured depressions with coarse shelly lenses at their bases. Some hummocks and depressions resemble the hummicky cross-stratification of Harms et al. (1975) but are typically smaller scale. Bioclast grainstones form discontinuous lenses. Abraded bioclasts are typically of low diversity but include fragments of corals, brachiopods, echinoderms and bryozoans. Well-rounded millet-seed quartz grains (see Lithofacies 6) locally form up to 20% of grains and contrast with the fine-grained angular to subangular quartz and feldspars typical of the lithofacies. Locally, 70% of the rock is dolomite and a number of erosion surfaces have been recognized within this.

Lithofacies 5 comprises only about 1 m of thin-bedded nodular limestones, generally similar to those of the Kalvsjoen Formation. In central and eastern localities these overlie Lithofacies 4, while in western Hadeland they are partly dolomitized and grade laterally into the upper parts of Lithofacies 2. The limestones are wackestones up to 7 cm thick separated by 3 cm thick shales. Bioclasts again include fragments of algae, brachiopods and echinoderms.

Lithofacies 6 is a distinctive sandstone which characteristically consists of well-rounded, near-spherical, 'millet-seed' quartz grains. These are up to 2 mm diameter and are mono- or less commonly, poly-crystalline and may be strained. Inclusions indicate a broadly granitic source although a few resemble fragments of chert. They are uniformly dull in cathodoluminescence. SEM examination of grain surfaces has failed to reveal the frosting characteristic of wind-blown sands, but since they have been etched and pitted by dissolution during diagenesis and many have syntaxial silica overgrowths, any original features have been lost. By contrast, up to 30% of the sharply angular grains in finer-grained sandstones and siltstones are feldspars. A few of these are twinned plagioclase, but the majority are untwinned and are only identified by their bright blue luminescence. Importantly, they are characteristically unweathered. Millet-seed sandstones form distinct beds 30–50 cm thick above Lithofacies 5 but millet-seed grains are also dispersed in finer-grained sandstones and dolostones, and form discontinuous lag aggregates. In all of these occurrences the millet-seed grains form a distinct grain-size population. In NE Hadeland, in Mjosa, and in Ringerike, they overlie and fill fissures in erosion surfaces truncating Or dovician rocks. Similar millet-seed lithologies are found in the Langara and Langeylene formations of Oslo–Asker, (Brenchley & Newall 1975, 1980) (Fig. 2). Millet-seed quartz grains are also present in the lower part of the Sælabonn Formation in Hadeland but as dispersed grains and diffuse discontinuous beds.

Sælabonn Formation

The predominantly siliciclastic Sælabonn Formation was originally defined in Ringerike (Worsley et al. 1983) and is the lateral equivalent of the Solvik Formation in
Kalvsjøen Formation

Debris-flows
Nodular Limestones

Klinkenberg Formation

Lithofacies 4
Lithofacies 3
Lithofacies 2
Lithofacies 1
Sedimentological changes, Ordovician – Silurian boundary

Fig. 4. Schematic illustrations of deposition in Hadeland during the latest Ordovician and early Silurian. □ A. Kalvsjøen Formation. □ B. Klinkenberg Formation. □ C. Sælabonn Formation. Note the exaggeration of bottom topography to emphasize variation.

Sælabonn Formation

Oslo–Asker (Fig. 2). It contrasts markedly with the nodular limestones, dolostones and 'millet-seed' sandstones of the Klinkenberg Formation on which it rests. It is about 140 m thick in the north and east of Hadeland, where it is also generally coarser, thinning to 100 m in western outcrops. In the Mjøsa districts to the NE, and in Ringerike to the SW, it overlies an erosion surface. Almost all (90%) of the sandstones are subarkosic arenites, the remainder have up to 25% matrix. Three lithological subdivisions are recognized in Hadeland (Fig. 3):

Unit 1 consists of 15–25 m of parallel-bedded (1–25 cm) fine-grained sandstones and shales. The basal sandstones contain scattered millet-seed grains which locally form thin discontinuous beds. The thicker sandstones commonly have bioclast lags at their bases, including fragments of brachiopods, trilobites, bryozoans, corals, orthocones, tentaculites and algae. The bases of beds are generally sharp and planar while upper surfaces may be planar or rippled. Many beds show surface-parallel laminations. Clay drapes are common on ripples and are locally marked by Planolites-type feeding traces. In some localities small load casts protrude downwards into underlying sandstones. Decimetre-thick thinning- and thickening-upwards cycles are common, but there is a gradual increase in shale content upwards to about 25%.

Unit 2 consists of regularly bedded sandstones with minor (< 10%) shales. The sandstones are mineralogically mature with a small percentage of carbonate clasts. The unit is 142 m thick at Svea in the east, where it forms the entire Sælabonn Formation, thinning to 30–38 m in central areas. Beds are typically 5–20 cm thick, locally reaching 60 cm. Thickening- and thinning-upwards cycles 50 cm to 1 m thick include some coarsening-up units. The bases of beds are generally sharp and some contain thin lag concentrations of brachiopods and molluscs. A few broad (4 m) low-angle troughs are present. Scattered Skolithos and Planolites burrows are preserved but many outcrop surfaces also show prominent mottling believed to result from burrowing. Bed surfaces are wave-ripped (vectors NNW/SSE in present terms) but laminae are concentrated in the upper centimetre or so of beds.

Unit 3 consists of a lower sandstone/shale interval, a middle thin-bedded sandstone, and an upper sandstone/limestone interval, with limestones increasing upwards into the Rytteråker Formation above. The unit is thickest (56 m) at Bjellum, thinning westwards. Wave ripples and clay drapes are common in the sandstones, accompanied by burrows and surface feeding traces.

The Rytteråker Formation

In Hadeland the Rytteråker Formation is only about 20 m thick, consisting of a basal 9.7 m of bioclastic
grainstones and packstones (with included siliciclastic grains) overlain by 11 m of bioclastic limestone, the basal few metres of which contain abundant pentamerid shells (Møller 1989).

Environmental interpretation

Kalvsjøen Formation

The Kalvsjøen Formation is interpreted as representing basin slope deposits. Much of the biota, particularly the algae, in the thin-bedded nodular limestones reflects shallow-water environments. However, there is no evidence of shallow-water reworking or current activity. Bioclasts are broken and abraded and the beds are usually graded suggesting that they are tempestites or relatively proximal turbidites while the interbedded shales reflect background deposition (Braithwaite & Heath 1992).

The interpretation of the chaotic breccias is more complex (Braithwaite & Heath 1992). The channels in which the breccias occur were cut into the nodular limestones but blocks were not derived from these and the event which formed the channels was probably quite separate from, and perhaps considerably earlier, than that generating the deposits. The varied lithologies and biotas contained within the blocks reflect deposition on a shallow carbonate shelf but do not include any peritidal lithologies. There are no in situ deposits representing this shelf preserved in the Ordovician of Hadeland. Petrography suggests that blocks were lithified before derivation and had a variety of diagenetic histories. They reflect the sampling of a stratigraphy rather than simply erosion of a lithified sea floor. Although deposited contemporaneously with blocks, part of the micrite and bioclasts forming the wackestone matrix evidently settled between them. The bedded wackestones associated with breccias contain a less diverse biota than that within the blocks, which was probably locally derived. The breccias are interpreted as debris-flow deposits (Fig. 4).

The stratigraphical age of the debris flows is not clear. Although their tops are everywhere concealed by superficial deposits they appear to be situated at the very top of the Kalvsjøen Formation. The faunas in the nodular limestones into which the channels were cut are diverse and almost certainly Rawtheyan in age (Heath & Owen 1991). Dr S. K. Donovan of the University of the West Indies informs us (pers. comm. 1994) that crinoid columnals from the Kalvsjøen Formation at Elgsjøen, close to the southeastern margin of Hadeland, resemble a form from the Hirnantian of Kiseley in northern England. The Kalvsjøen Formation has not been examined at Elgsjøen by the present authors and it is not known what lithofacies the crinoids in question came from. However, they provide evidence, albeit slender, that the cutting and filling of the channels might have taken place in the Hirnantian and would therefore accord with such events elsewhere in the Oslo-Asker district and in several other places throughout the world (e.g., see Brenchley 1988).

Following deposition of the breccias, a fall in sea level apparently exposed eastern Hadeland and resulted in karst erosion of the debris-flow deposits. This produced cavities and fissures within the lithified debris flows, although not apparently in the nodular limestones. As sea level rose again, flooding the shelf, these cavities were filled with marine grainstones, presumably derived from the adjacent shallow shelf margin. This may indicate relatively rapid flooding of the shelf and a substantial lag between this and the onset of grainstone deposition. It is arguable whether the cavity-filling grainstones should be regarded as ‘Kalvsjøen’ or ‘Klinkenberg’ Formation.

The debris flows of Hadeland have only been found close to the eastern margin of the district (Fig. 4) and sediments within them were probably only transported 5–10 km westwards (Braithwaite & Heath 1992). This distance accounts for both the small area of the deposits and for the lack of large-scale slump and slide features which would be expected proximally. It also allows water depths to be set at realistic limits to accommodate the subsequent emergence of the deposits. Although the Mjøsa districts to the north were probably emergent in the latest Ordovician, they were too far away to be considered as a source for the debris flows.

Klinkenberg Formation

Following the postulated transgressive shelf-flooding, the Klinkenberg Formation is thought to have been deposited during a regressive phase (Fig. 4). The initial stages of sea-level lowering are indicated by the general increase in siliciclastic material within the formation. The final stages are demonstrated by the emergence of northern Hadeland around Gamme and perhaps larger areas. However, the substantial gap in the sequence and obvious erosion in the Mjøsa districts further north mean that the timing and place of the start of emergence is not known. Lithofacies 1 is thought to have formed partly in response to storm activity. Load structures generated by liquefaction trend NNW–SSE, almost parallel to the presumed gradient as well as to the margin of the basin as indicated by the earlier distribution of the debris flows, by the isopachytes for the Skøyen Group (Heath & Owen 1991, Fig. 10), and by the distribution of lithofacies therein. Bioclast layers, which are more common in the east, are probably storm-lag deposits. To the west, burrowing and poorly preserved lamination point to quiet water deposition below wave base and to a generally more distal environment.

Lithofacies 2 reflects a progressive decrease in the supply of siliciclastic material. The deposition of fine-grained, locally laminated and burrowed sediments points to a continuity of quiet water environments. The decrease in siliciclastic sediment may reflect the development of a rimmed margin which acted as a barrier to sediment movement, with periodic breaches in the rim allowing both carbonate and siliciclastic sediments to be
transported onto the basin slope (Lithofacies 3, 4). The distribution of these sediments in distinct lobes suggests local channelized by-passing of the rim rather than a general overspill. The rim probably consisted of shallow water carbonates resembling those in Lithofacies 1 and 4, and may have been lithified. Within Lithofacies 4 the bioclastic limestones and siltstones were evidently deposited in shallow water, perhaps in the order of tens of metres deep, and reworked by waves generated by onshore winds. Ripple marks suggest wind vectors either NW–SE or NE–SW (in present terms) although they may, of course, indicate wave refraction. The sharp bases and coarse lags resemble those of storm deposits (Aigner 1985; Brenchley et al. 1979).

If winds were westerly, the waves generated on a west-facing shore could have been large and reworking effective in relatively deep water. If, however, winds and waves were from the east, quite shallow sea floors could have been below wave base. Brenchley et al. (1979) suggested that during the Ordovician the Oslo area lay within the SE Trade Wind belts. They presented an elaborate series of models to explain how winds from the south, SE and west impinging on (for them) an eastwards-facing shore were able to generate NE-flowing currents. The recent palaeogeographical maps of Trench & Torsvik (1992) show Baltica lying almost 60°S and aligned roughly E–W during the early Ordovician but by the late Ordovician it lay about 30°S and had rotated towards its present orientation. The SE Trade Wind belt in the present Southern Ocean is centred on 20°S. At higher latitudes in this area winds are more variable and for much of the year may blow either from the NE or SW. However, between 30° and 40°S the ocean circulation generated by such winds and by convection shows a consistent easterly sense of movement, partly because circulation is almost unimpeded in the Antarctic Ocean. This was not so in the Ordovician where currents may have been analogous to those of the present northern Indian Ocean or indeed the Pacific where smaller circulation cells are hemmed in by land masses. Thus, both the general current regime and the progressive relative rotation of Baltica through the Ordovician present formidable barriers to interpretation. The sense of wave movement measured here is consistent with (now) NE–SW pattern (in accordance with Brenchley et al. (1979)) but we must rely on other criteria to determine the net transport direction. Geostrophic flow certainly provides a mechanism, turning an 'eastwards' surface movement into a 'westwards' bottom flow. The strongest argument, however, is in the eastwards thickening and shallowing of siliciclastic deposits in Lithofacies 1 and 4, suggesting derivation from this direction.

The dolostone unit of Lithofacies 2 probably reflects deposition of carbonate muds on a surface starved of siliciclastics. The erosion surfaces within it may reflect pauses in deposition, or storm erosion, possibly following sea-floor lithification. Evidence of this has been lost in dolomitization. However, dolostones show greater petrographic diversity and variety of cathodoluminescent zones in the east, probably reflecting proximity to a draining shelf, while in the west less dolomite and fewer dull zones suggest a lack of penetration of waters into the more poorly oxygenated basin fill. These variations will be discussed elsewhere.

The occurrence of the millet-seed sandstones of Lithofacies 6 within widely different lithologies, and their ubiquitous appearance as a distinct size mode, suggests that they were not transported by the same water masses and current systems as other sediments. The apparent mineralogical contrasts between millet-seed grains and 'host' sandstones seem to imply differing sources. Millet-seed sandstones occur principally in eastern areas, concentrated at the top of the Klinkenberg Formation, although they are not confined to it. Similar millet-seed sandstones are present in the uppermost Ordovician of the Oslo–Asker District (Brenchley & Newall 1975, 1980) and, as in NE Hadeland, they fill fissures in erosion surfaces at or near the top of the Ordovician succession in Ringerike and in the Mjøsa districts. While grain surfaces are not now frosted, their degree of rounding is distinctive. Wind transport is the most effective agent for rounding sand grains (Kuenen 1960) and it is thought that the millet-seed grains of Hadeland could only have been rounded in such a system. Winds travelling at about 17 m/sec⁻¹ could have rounded grains in distances as small as 350 km (Kuenen 1960) and would have been able to carry them for more than 10 km from the landmass, to be deposited with sediments of other origins. A small amount of reworking (which may have removed frosting) was insufficient for grains to become integral parts of the grain-size distribution of the host sediments. It seems, therefore that the millet-seed grains were derived from a significant mature landmass to the north and east of Hadeland, and may have been deposited on (?subaerial) erosion surfaces on the margins of the land before being swept generally south and westwards into the adjacent shallow marine environment.

**Sælabonn Formation**

The Sælabonn Formation is predominantly siliciclastic. It is thicker and coarser in the north and east (Fig. 4), but the variations in shale content in units 1 and 3 appear to be unsystematic. Lateral changes from sandstone dominance to shale dominance are abrupt and localized, suggesting that sand was distributed along distinct pathways rather than uniformly along a slope. Wave ripples indicate a dominant NW–SE oscillation (Heath & Owen 1991, Fig. 10). This may reflect the passage of Baltica to a lower latitude, or its rotation relative to a stable oceanic circulation, or indeed the change to the non-glacial climate of the early Silurian. The planar laminated beds commonly show current lamination and were produced under upperflow regime conditions, perhaps in shore-face environments. However,
such conditions were sporadic and clay drapes and burrows indicate long periods of quiet water.

The absence of large-scale cross-bedding precludes deposition in sand waves or megaripples. Beds within the sequence show similarities with Bouma B, C and D divisions, where parallel-laminated sands are interbedded with a minor shale component. However, the absence of graded bedding, sole marks and load and slump structures argues against a turbidite origin. Nevertheless, sand supply must have been from point sources. The eroded troughs and hummocks are similar to those in hummocky cross-stratification (Harms et al. 1975). The sandstones are similar to those described by Aigner (1985) from the North Sea and are interpreted as resulting from deposition by storms. Coarse shelly bases are common, and similar features have been attributed to storm debris associated with flushing of channels (Brenner & Davies 1973). Parallel lamination is thought to reflect deposition of sand from suspension, with the overlying rippled sequences indicating reworking by waves and currents as storms declined. The bioturbated tops of sandstones reflect the return to quieter conditions and the re-establishment of an infauna.

The fining-up units, which generally reflect a decline in current activity, are intercalated with coarsening and thickening units thought to indicate the progradation of individual sandstone lobes. These were more confined during the early stages of deposition of the Sælabon Formations, allowing the accumulation of finer-grained sediments in interdistributary areas, but became more widely dispersed with time. This may reflect the gradual filling of relief created during the low-stand, but we have no direct evidence of this. There is no conclusive evidence of the direction of sediment transport, but the concentration of coarser-grained rocks and the overall thickening of the formation to the east accords with the transport directions deduced for the Kalvsjøen and Klinkenberg formations. The increase in siliciclastic supply seems to point to a relative fall in sea level. The sandstones are mineralogically mature and thus are likely to have been derived from mature sedimentary or metasedimentary sequences rather than a newly unroofed igneous complex or rising ophiolitic nappes. The 'Sparagmite' sequences, which could have produced such mature sediment, may not have been included within the nappe pile at this time.

Discussion

Hadeland and the Oslo Basin

During the Ordovician and Silurian, much of Baltica formed a broad platform covered by a shallow epicontinental sea (Jaanusson 1976, but see also Torsvik et al. 1991; Trench & Torsvik 1992; Perroud et al. 1992 for palaeogeography). Deposition within this sea was typically slow and took place in a series of broad facies belts.

By contrast, sediments in the Oslo Region occupied a narrow N-S trending depression and facies changes were more rapid. However, the lithofacies distributions of the Early Palaeozoic basin cannot simply be superimposed on the present geography of the Oslo Region. There are major faults and marked differences in deformational style between and within districts which have prevented either palinspastic reconstruction or the construction of balanced cross-sections of the region as a whole.

The Cambrian to Lower Ordovician succession of the northern part of the Mjøsa districts is allochthonous, and overlies the late Precambrian to early Cambrian Hedmark Group (Nystuen 1987). South of this, the parautochthonous Cambro-Silurian succession is deformed by W-E striking folds and is imbricated, probably above a basement detachment. A similar parautochthonous deformation is seen in northern Hadeland where large numbers of hinterland-dipping thrust slices of Cambrian to middle Ordovician rocks rise from a plane of décollement near the base of the Cambrian (Morley 1987, but see also Morley 1994 for an additional mechanism for imbrications). Morley (1987) postulated a second detachment within the poorly exposed middle Ordovician shale succession, above which the middle Ordovician to Lower Silurian sequence shows fairly broad open folds plunging westwards and cut, inter alia, by W-E reverse faults (Owen 1978). The sense of movement on the higher detachment may have been towards the orogen (Morley 1987).

The Lower Palaeozoic succession of Hadeland is overlain in the east by Permian igneous rocks and is truncated in the west by the Randsfjord Fault marking the western edge of the Oslo Graben here. This fault strikes SSE from the southern end of the Randsfjord before it is concealed beneath the Permian nordmarkites of Nordmarka (Ramberg & Larsen 1978, pl. 1). It separates Hadeland from Ringerike to the SW. The Cambrian to middle Ordovician succession of Hadeland is again strongly imbricated by hinterland-dipping thrusts rising from a major basal décollement, but the principal thrust ramps up along the Klekkken Fault to lie above the upper Silurian Ringerike Sandstone Group (Harper & Owen 1983). In contrast to the W-E striking folds of Hadeland, the middle Ordovician to Silurian succession in southern Ringerike, south of the Klekkken Fault, is only gently tilted to the SE.

In further contrast, the Cambro-Silurian succession of Oslo-Asker, which lies south of Nordmarka and SE of Ringerike, is tightly folded with axes striking NE-SW and strike-parallel listric splays rising from a basal décollement (Bockelie & Nystuen 1985; Morley 1986). Further SW in the Oslo Region, the tectonics affecting the Lower Palaeozoic successions become progressively simpler from tight easterly-plunging folds in western Modum (Wandås 1982) to gentle eastward-tilting in Skien-Langesund (Owen et al. 1990). Even here, in the southernmost part of the Oslo Graben, most of the Lower Palaeozoic succession is parautochthonous, al-
though there is debate as to the level of the décollement (Owen et al. 1990, p. 40).

The Oslo Region is thus an incomplete three-dimensional jigsaw, with the relative palaeo-positions of the pieces only approximately known. Moreover, the evidence of early Palaeozoic block-faulting in Oslo–Asker (Brenchley & Newall 1980; Stanistreet 1983) indicates that local topographies and sediment distributions within parts of the basin may not have reflected those of the basin as a whole.

Gross variations in lithology in the Oslo Region have hitherto been regarded as indicating a westwards shallowing (Størmer 1967; Bockelie 1978). No autochthonous Ordovician or Silurian rocks are exposed west of the Oslo Basin but successive authors since Skjeseth (1952) have regarded the western area, 'Telemarkland', as being one of at least shallow water and possibly an emergent source of silicilastic sediment throughout the Ordovician and Silurian (Bjørllykke 1974; Brenchley et al. 1979; Worsley et al. 1983; Baarli 1985).

However, while carbonate rocks are generally thickest in the west, and may point to shallower water here, they also indicate a lack of any silicilastic source. Thus, Bjørllykke (1974, p. 75) noted that, while there may be evidence indicating westward shallowing, the lower Ordovician rocks show no increase in thickness or grain size in this direction. Moreover, chlorite (one of his indicators of derivation from the advancing nappes) appears much earlier in sediments in the north of the region than in those in the west.

There have been a few, generally unsupported, allusions to some silicilastic sediment having entered the basin from the east in the latest Ordovician (Størmer 1967, p. 204; Worsley et al. 1983, p. 45) and during part of the Llandovery (Møller 1989, Fig. 2A). Our detailed study of the uppermost Ordovician and lowest Silurian succession in Hadeland points to silicilastic (and periodically carbonate) sediments being transported from the east. Evidence from other districts outlined below suggests that this may have been a widespread feature of the basin at that time.

Mjøsa

North of Hadeland, in the Mjøsa districts, (Fig. 1) the Middle Ordovician (upper Caradoc? to lowest Ashgill) Mjøsa Limestone (Bjørllykke 1974; Opalinski & Harland 1981) is overlain by rocks of Aeronian age and the uppermost Ordovician and lowest Llandovery are absent. Since the boundary is common an erosion surface the timing of emergence of the Ordovician sequence is not known and may well vary in different localities.

The Mjøsa Limestone is a shallow water carbonate sequence including block deposits and so called 'reefs' (Opalinski & Harland 1981; Harland 1981). At Rud (NN 914281), for example, cross-bedded grainstones are overlain by pelletal limestones containing algally coated grains and fenestrae characteristic of peritidal environments. These contrast with the coeval deeper water sediments in Hadeland and indicate that shallowing and perhaps emergence occurred far earlier in the Mjøsa area than in Hadeland. Some Mjøsa Limestone lithologies resemble those sampled in the debris flows of the Kalvsjøen Formation and may thus represent a shelf similar to but older than that eroded in the Rawtheyan. However, there is more evidence of shoreline deposition than is apparently represented in the blocks of the younger deposits.

At some localities (e.g., Rud, NN 922281; Dølbakken, NN902261) the surface of the Mjøsa Limestone is a karst surface, with coarse sandstones of the Helgøya Quartzite Member of the Sælabonn Formation filling cavities and depressions. Elsewhere, (e.g., Vestby, NN 921379) the boundary is planar but nevertheless reflects a considerable break in deposition (Worsley et al. 1983). At Rud and elsewhere, the Helgøya Member contains millet-seed quartz grains. These are usually larger than those in Hadeland and are most obvious at the base of the Helgøya Member, although they occur throughout the unit. The Helgøya Member is lithologically similar to the upper part of the Sælabonn Formation in Hadeland. This supports the view that a considerable part of the lowest Silurian is missing (Worsley et al. 1983). As in Hadeland, limestones become more common upwards in the transition to the Rytteråker Formation.

Feiring

To the ENE of Hadeland the Lower Palaeozoic rocks of Feiring are thermally metamorphosed, poorly exposed and poorly understood (see Owen et al. 1990, p. 45 and references therein). Major (1946) described late Ordovician or early Silurian sandstones from Skreikampen, 45 km from the upper Ordovician of Hadeland. These are overlain by what is probably the Rytteråker Formation but their precise age range and thickness are not clear. Given the effects of Caledonian and even early Palaeozoic tectonics in the Oslo Region, the palaeogeographical location of the deposition of these sandstones relative to those of Hadeland is unknown.

Ringerike

The Bønsnes Formation of Ringerike, SW of Hadeland, is probably laterally equivalent to the Kalvsjøen Formation (Owen 1979; Heath & Owen 1991) (Fig. 2) and is predominantly of similar nodular limestones and shales. The top of the Bønsnes Formation, on Store Svartøya, consists of silty shales with calcareous nodules. The overlying unit, equivalent to the Klinkenberg Formation, (Owen et al. 1990) was described but not formally named by Hanken (1974; Hanken & Owen 1982) and the name Langøyene Sandstone, that of the broadly equivalent unit in Oslo–Asker, tentatively applied by Owen et al.
(1990). Quartz sandstones and grainstones at the base are overlain by crinoidal grainstones and by what Hanken referred to as a ‘Carbonate Bank’. This in turn is overlain by calcarenaceous sandstones and sandy crinoidal grainstones. The top of each lithofacies is marked by an erosion surface, and in the most distinctive of these, on Store Svartøya, runnels in the top of the limestone are filled with shales of the Sælabonn Formation. This sequence is equivalent to the Klinkenberg Formation of Hadeland but here includes four separate erosion surfaces.

The sandy sequence in Ringerike is thinner than that in Hadeland but the sharp bases to beds, the shelly lags and the low-angle parallel lamination all point to generally similar conditions of deposition, probably dominated by storm-generated waves and currents. The intercalated shales and more common burrows in Ringerike suggest a relatively distal position. As in Hadeland, there was an increase in sandstone deposition at the end of the Ordovician.

The ‘carbonate bank’, well exposed at Ullerntangen, requires further discussion. This was described by Kær (1897) as a ‘reef’. However, Hanken (1974, but see also Hanken & Owen 1982) referred to it as ‘a bank’ and noted that it lacks a framework and contains transported material including randomly orientated stromatoporoids. The outcrop at Ullerntangen covers a large area (350 x 100 m). Like the so-called ‘bank’ at Kalvsjø quarry, it consists of lithologically diverse lithoclasts up to 85 cm diameter. Hanken (1974) described dislodged blocks within the limestone and he noted, but undervalued, the importance of these and of derived stromatoporoids. Clasts include a variety of fossils but these are commonly distinct in adjacent blocks. Brenchley & Cocks (1982) noted also that very few species are in common with those of the open-shelf environments of Oslo–Asker. Fossils are often truncated at block margins and these, together with geopetal structures, indicate the random orientation of blocks which were lithified prior to transport. Hanken examined cathodoluminescent cement sequences within fossils and noted that, as in Hadeland, there are several generations. He also referred to marly intercalations which he regarded as indicating that the mounds had a low relief. In the absence of any autochthonous biogenic structure, we suggest that the Ullerntangen ‘mound’ is a channel-fill sequence of carbonate debris-flow deposits analogous to but slightly younger than those at the top of the Kalvsjøen Formation. As at Kalvsjø quarry, the breccias are truncated by a karst surface, in this case with a relief of up to 3 m. Some crinoidal grainstones drape over the limestone breccias but their position is not clear since they also have been eroded and cavities cut through them into the breccias are filled with millet-seed quartz grains.

The crinoidal grainstones at the top of the Langøyene Formation in Ringerike were eroded before deposition of the Sælabonn Formation. The erosion surface forms irregular runnels and presumed aragonitic fossils, cephalopods and gastropods, within it were selectively dissolved and the moulds subsequently filled with sandstone, as in the Klinkenberg Formation at Gamme in Hadeland. Grahn et al. (1994) stated, on the basis of the chitinozoans, that the uppermost part of the Ordovician is missing in Ringerike and Thomson (1982) suggested that the brachiopod faunas indicate that a significant part of the Rhuddanian is also missing.

During the latest Ordovician, Ringerike seems to have lain further from a siliciclastic source than Hadeland. Carbonates pass eastwards into sandstones (Hanken & Owen 1982, Fig. 4) suggesting that the supply of siliciclastic sediment was from that direction. Emergence was frequent and in each event renewed flooding seems to have been followed by siliciclastic deposition.

The Sælabonn Formation of Ringerike has been divided into three members (Thomsen 1982), the Store Svartøya Member, 20 m of calcareous silty shales with thin limestones, the Djupvær Member, 48 m of calcarenitic sandstones, and the Limovnstangen Member, 40 m of siltstones and shales with increasing numbers of limestones towards the boundary with the Ryttéråker Formation (Fig. 2). These divisions are similar to those in Hadeland but the succession is not as thick and our observations suggest a greater percentage of carbonate, increasing to 60% at the top of the succession. This again implies deposition at a greater distance from the siliciclastic source, but the area was certainly shallow and Thomsen (1982) thought that storm events introduced sediment from the north. Importantly, Möller (1987) concluded that during deposition of the Ryttéråker Formation the mainland area lay to the east of Ringerike. Baarli (1990) recorded a shallow-water benthonic fauna from the Sælabonn Formation in Ringerike.

**Oslo–Asker**

Brenchley & Newall (1975) divided the uppermost Ordovician of Oslo–Asker into three formations (see Fig. 2). The Husbergøya Shale Formation is broadly equivalent to the Kalvsjøen Formation in Hadeland and also to the Bønnes Formation in Ringerike (Owen 1979). To the NW it becomes more calcareous and passes laterally into, and is overlain by, the lower part of the Langåra Limestone-Shale Formation. To the SE, in Oslo, the Husbergøya Shale is overlain by the Langøyene Sandstone Formation, the lower part of which is laterally equivalent to the upper part of the Langåra Limestone-Shale formation of Bærum and Asker.

Brenchley & Newall (1980) and Brenchley et al. (1979) suggested that both the Husbergøya Shale and the Langåra Limestone-Shale Formation were deposited on a SE-directed palaeoslope. They considered that the Langøyene Sandstone was derived from the west and formed a northwards-migrating bar, believing that the increasing siliciclastic supply reflected a shallowing which culminated in the development of oolitic limestones.
Shallowing is certainly indicated by an upwards change in faunas (Brenchley & Cocks, 1982) and fully accords with the model proposed for Hadeland. However, we believe that the sedimentological evidence indicates a different palaeogeography.

Stanistreet (1983, 1989) accepted the existence of the western shoals of Telemarkland but suggested a number of important amendments to the facies interpretation of Brenchley & Newall (1975 et seq.). In 1983 he drew attention to three north–south trending faults, the Nesøya, Bronnøya and Bunnefjord lines. These were believed to have been active during deposition of the uppermost Ordovician and apparently limited the westwards extension of the arenaceous facies. If these faults are real, there is clearly a problem in how the horst limited by the Nesøya and Bronnøya lines could retain sediment derived from the west while the western area retained no trace of its passage. Processes moving sand on top of the horst were not operative to the west and the late Ordovician oolites are virtually confined to this structure. Equally compelling is the evidence from trace fossils (Stanistreet 1989). Three well-defined associations were recognized. An offshore Thalassinoides association, widespread in the earlier stages but later restricted to carbonate-rich offshore environments west of the Nesøya fault, a Diplchnites-Phygod association, reflecting a lower shoreface environment and largely confined to sand-rich lithologies to the east of the Nesøya fault, and a Skolithos-Diplocrater association in plane and cross-bedded sandstones reflecting upper to middle shoreface environments largely between the Nesøya and Bunnefjord faults. The occurrence of the last of these ichnofacies to the east of the Bunnefjord fault is believed to reflect synsedimentary movement of the fault.

The revised interpretation proposed here can be considered in relation to three areas which broadly follow the divisions of Stanistreet (1983), the islands in Bunnefjorden to the east, the SW islands of Oslofjorden (essentially central to Oslo–Asker), and the NW sections of Holmen, Kalvøya and Sandvika. However, we emphasize again that the detailed palinspastic relationships of these areas are not known.

(1) Bunnefjorden (Brenchley & Newall 1975, pp. 246–247, Figs. 4–5, Logs 1–11). The Husbergøya Shale Formation is 17–25 m thick consisting of interbedded shales, thin sandstones and calcareous beds, the last increasing in frequency upwards. These rocks are capped by a thin bioturbated sandstone. The Langøyene Sandstone above is much more varied. It is thickest (53 m) on the southern islands (SE in a palinspastic sense) and thins NW (again palinspastically) to 27 m on Hovedøya. However, the sandstones are everywhere truncated by deep channels so these are minimum figures. Contorted bedding and channels are common (Brenchley & Newall 1977). Millet-seed sandstones are common here also, forming laterally continuous beds and filling channels up to 6 m deep. Well-rounded millet-seed grains are about 1 mm in diameter and, as in Hadeland, form distinct populations in the sediments in which they occur. In addition, and unlike Hadeland, they are associated with ooids and, locally, coarse bioclastic grainstones. A few beds of millet-seed sandstones and oolites appear as storm-graded beds low in the Langøyene Sandstone in what are now southeastern locations. Millet-seed sandstones with some ooids occur in channels in the southern islands in the middle part of the formation, in a channel on Nesøya and on Bjørkøya in the southwest. An oolite sheet caps the Langøyene Sandstone between the Nesøya and Bunnefjord faults.

A number of channels, with fills lithologically similar to those in the Kalvsjøen Formation in Hadeland cut the Langåra Limestone-Shale and the Langøyene Sandstone. These were originally referred to the Langåra Formation (Brenchley & Newall 1975) but have since been included within the Langøyene Sandstone (Brenchley & Cocks 1982). On Hovedøya a channel 6 m deep with an exposed width of 40 m contains blocks, up to 2.5 m maximum diameter, of a variety of limestone lithologies. Six separate incursions of blocks and matrix can be identified, with at least three additional beds in which small lithoclasts and bioclasts were apparently washed down the same path disrupting normal deposition of sandstones and shales. As in Hadeland it seems unlikely that the event which cut the original channel was responsible for depositing the first or any layer of blocks.

(2) Oslofjorden. The Husbergøya Shale Formation is predominantly shales and sandstones but the latter are less common than in the eastern area (Brenchley & Newall 1975), only appearing higher in the succession (Brenchley et al. 1979). In time, the sediments apparently became more calcarenaceous and the top of the formation passes up into the Langåra Limestone-Shale Formation except in the southern sections such as those on Høyervolmen and Skogerholmen which are along-strike correlatives of the Bunnefjord outcrops. In these southern outcrops, the Langøyene Sandstone Formation is thinner than to the east and less variable. The base is formed by 3–4 m of limestones and shales. Above these, thickly bedded sandstones commonly contain convolute bedding. The thickness of these sandstones varies, but may total 18 m. Finally, the sequence is capped by about 2 m of oolitic limestone containing millet-seed quartz grains.

(3) Sandvika, Holmen and Kaltøya. These localities to the NW straddle Stanistreet’s (1983) Nesøya fault and in none of them is the section complete. On Kalvøya and Holmen the Husbergøya Shale Formation is apparently 10–14.5 m thick (Brenchley & Newall 1975) and rests on the nodular limestones of the Skogerholmen Formation. In contrast, at Sandvika (west of the fault) the shale is more than twice as thick and the base is not exposed. Two sandstone units 5–10 m thick in the lower part of the section consist of thin storm sandstones interbedded
with mudstones and are the only significant coarse siliciclastics here.

The boundary between the Husbergøya Shale and the Langåra Limestone-Shale is recognized in all three locations. The Langåra Limestone-Shale is a monotonous sequence of silty shales and calcareous nodules. The silty units are commonly rippled or burrowed. On Kalvøya and Holmen (on Stanistreet's 1983 Oslofjord horst) the unit is cut by shallow channels filled with carbonate lithoclasts which contrast with the lithologies of the host sediments. At Sandvika the Langåra Limestone-Shale is 35 m thick in the section examined by Brenchley & Newall (1975) and consists of silty nodular limestones with nodules centred around clusters of bioclasts including gastropods, Catenipora and solitary corals. Two sandstone intervals, 3 and 6 m thick, are present in the middle of the unit. However, at Kampebrøtten, west of Sandvika, more than 15 m of fine sandstones include incursions of millet-seed sandstones and oolites, probably derived from the horst.

The channels on Kalvøya and Holmen are only about 2 m deep, much smaller than that on Hovedøya. Lithoclasts in that on Kalvøya are up to 30 cm diameter and are of oolitic, pelletal and coral-bearing limestones, contrasting with the host limestones, but they include the alga Palaeoporella, which is common in NW Asker, suggesting that material has been transported from shallower areas nearby. Channels on Holmen, also show lateral failure of relatively weak walls, yielding additional blocks to the fill.

**Interpretation.** — Our observations in Oslo-Asker, with those of Brenchley & Newall (1975 et seq.), suggest that the uppermost Ordovician sediments were deposited on a storm-influenced slope in progressively shallowing waters. Brenchley et al. (1979) suggested that the lower Langøyene Sandstone and Husbergøya Shale were deposited by wind-forced ebb currents which initially flowed offshore and were deflected NE by the Coriolis effect. However, while cross-bedding measurements record current directions they do not necessarily indicate the origin of the sediment, which may have been reworked. It seems unlikely that sands were derived from the west when so little sand is present in that area. Moreover, the markedly cross-cutting facies relationships suggested by Brenchley et al. (1979) seem highly improbable, demanding as they do the synchronous deposition of sands of the putative northward-migrating bar and the sediments of the eastward dipping palaeoslope with limestones in the west and muds in the east. The derivation of the eastern muds and the absence of an interdigitating sandstone-limestone sequence at the northern edge of the supposed sand bar are serious problems with this model as is the apparent confinement of the sands to the east of the Nesøya fault.

Brenchley & Newall (1980) believed that the channels of the Oslo-Asker area were tidal, cutting through offshore bars. There are three reasons why this is unlikely: (1) There are no peritidal deposits amongst the variety of lithoclasts which were derived either from a very wide area or a very thick succession. (2) The large clast sizes suggest that the trigger for transport was a catastrophic event. There is no unequivocal evidence of the direction of transport within channels. Local imbrication is apparently normal to channel margins. (3) As in Hadeland, the clasts imply proximity to a carbonate source area free from siliciclastic influence, although channels in some areas (Langøyene) do contain sands. Borings of *Tripanites* in blocks (Brenchley & Cocks 1982) indicate that limestones exposed were already lithified.

Importantly, as in Hadeland, channel fills do not form part of the sedimentary sequences in which they are now lodged. Stanistreet (1989, Fig. 5) described and illustrated a fill sequence of the channel at the top of the Ordovician on Hovedøya which includes a series of trace fossil assemblages. That on Torbjørnsøya remained open until filled with Silurian shales (Brenchley & Newall 1980) and the relief on the base of the Silurian may be up to 100 m (Bjørlykke 1974). However, while the channels may all be products of a single major event they do not relate to a particular instant of time and on Langøyene three superimposed channel systems are seen.

In much of the Langøyene Sandstone Formation the distribution of lithofacies has been modified by block faulting (Stanistreet 1983, 1989). The thickness variations related to faults seem well established. However, the lack of lithological variation across these lines and their subsequent overlap by units in which faults were no longer apparent argue for erosion and redistribution of the surfaces they created, smoothing the topography. Seismic activity related to faulting may have been responsible for the generation of the contorted bedding within the unit.

Brenchley & Newall (1980) use the presence of thick sandstones low in the formation to the south and higher to the north of the Oslo-Asker district to support the idea of a northward-migrating bar. However, the upper Ordovician sandstones clearly thicken to the east. The generalized isopachyte map for Oslo-Asker showing sandstone thickness published by Brenchley & Newall (1975, Fig. 12b) could be reinterpreted in terms of a westward-thinning sand lobe, sourced from the east as in Hadeland. Stanistreet's description of apparently lower-energy ichnofacies within the Langøyene Formation east of the Bunnefjord fault is a problem in this interpretation in that it is difficult to supply sediment derived from the east to the upthrown side of the fault. However, it is equally difficult to envisage derivation from the west (across the carbonate-dominated belt) or the south where transport would be required to be confined to the axis of a topographical high which cuts across the lithofacies boundaries deduced by Brenchley et al. (1989).
The basal part of the Silurian in Oslo–Asker is the Solvik Formation (Worsley et al. 1983; Baarli 1985). This is up to 240 m thick, dominated by shale, and contrasts with the thinner (150 m) arenaceous sequences in Ringerike and Hadeland. It is not clear whether this thickness variation reflects lateral progradation against an underlying topography, the filling of an irregular topography of the underlying Ordovician (Worsley et al. 1983; Baarli 1985), or the irregular subsidence of local fault blocks (Worsley et al. 1983). Baarli (1985, 1988) referred to the continued effects of a tectonic high (of at least 20–30 m relief) in Asker on the site of the late Ordovician Oslofjord Horst (of Stanistreet 1983, 1989) and Stanistreet (1989) regarded westerly occurrences of oolites and plane- and cross-bedded sandstones as reflecting shallowing over this block. However, while there is convincing evidence of thickness variations, no distinctive lithofacies were formed by erosion of fault scarps.

Worsley et al. (1983) subdivided the Solvik Formation in Oslo into the Myren and Padda members, while Baarli (1985) subdivided the generally more calcareous succession in Asker (and the Holmegrend district to the south) into the Myren, Spirodden and Leangen members with the first two of these temporally equivalent to most of the Myren Member in Oslo.

Baarli (1985) regarded the Solvik Formation as reflecting a prograding system. Deposition was thought to have occurred on a storm-dominated shelf but he believed that western areas were more proximal. The arenaceous content of the Solvik Formation varies vertically and laterally. Thin beds of siltstone are most common in the east (Malmøya) in the lower part of the formation (the Myren Member), but at the top arenaceous beds are more common and coarser to the west in Asker. Worsley et al. (1983, p. 15) considered that this change reflected a reversal in the slope of the basin but Baarli (1985) argued that abundant sole structures (including gutter casts) and striations are consistently orientated NNE–SSW in both Oslo and Asker and indicate transport from the SSW. It is not clear why these linear vectors could not reflect movement from the NNE. Baarli also noted that in Oslo–Asker, individual gutter casts and those associated with very thin beds are more common in the east, while beds to the west are generally thicker and contain climbing ripples, suggesting that they were more proximal. Small-scale cross-lamination indicates currents from the west. She argued that hummocky stratification, small channels, medium-bedded intercalations and more abundant storm-lag deposits in the west also indicate a more seasonal environment. However, all of these features relate to the upper part of the sequence, where regionally extensive shallowing is widely accepted and the possibility remains that, at least for the Rhuddanian, siliciclastic material may have been transported from the east.

Baarli (1988) described a proximality trend analysis of the Solvik Formation in Asker (see also Aigner & Reineck 1982; Aigner 1985). This suggested that the area lay offshore during the deposition of the Myren Member and the basal part of the Leangen Member, but that there was a gradual shallowing towards the transition zone below the shoreface. She explained the apparent deepening at the top of the formation as reflecting a cut-off in siliciclastic supply. However, community analysis undertaken at the same time (Baarli 1988) provided a conflicting picture. It suggested a lengthy period of gentle shallowing (indicated by the Spirodden Member) followed by an abrupt deepening in the lower part of the Leangen Member. Baarli explained this anomaly as arising from deposition of the Spirodden Member above a tectonic high (see also Baarli 1985), making it unsuitable for proximality analysis. Nevertheless, the presence of Clorinda communities in the lower part of the Leangen Member whilst beds of equivalent age in Oslo contain Stricklandia communities (see also Baarli 1985; Johnson et al. 1991) is counter to the general pattern of communities in the rest of the formation, which indicates deeper water to the east.

**Modum**

Reconnaissance work on the upper Ordovician of the Sylling area (SE Modum) by Owen et al. (1990, p. 42) described a limestone-dominated succession similar to that of Ringerike to the north. In the same area Baarli (1988) described a lower Silurian unit which she regarded as intermediate between those of Ringerike and Oslo–Asker and hence termed the Sælabonn–Solvik Formation. The lowest part of this, the Sylling Member, rests on a karst surface of Ordovician limestone (as in Ringerike) which is covered by 8–10 cm of 'millet-seed' sandstone. Most of the rest of the 47 m member consists of shales and thin siltstones similar to those of the Solvik Formation, but silt and sandstone couplets above resemble the Djupvarp and Limovnstangen members of the Sælabonn Formation of Ringerike. Brachiopod faunas indicate that the Sylling Member is equivalent to beds in Asker ranging from the middle of the Spirodden Member up to the lower part of the Leangen Member; hence there is a significant lower Silurian hiatus (Baarli 1988). Proximality trends analysed by Baarli (1988) show that within the 'tempestites' of the Sylling Member, there was apparently a shallowing from offshore muds into the transition zone below the shoreface and back again. Similarly, the lower part of the Djupvarp Member above shows shallowing in the transition zone (perhaps into the lower shoreface) followed by a slight deepening. However, Baarli rejected the proximality analysis of the Limovnstangen Member on the grounds that the sediment supply was cut off during the change to the overlying Ryttæråker Formation. Significantly, the bathymetric curves derived from proximality and community analyses for the Sylling Member in Modum show the same basic pattern as that derived from community analysis of the...
Spirodden Member in Asker but the apparent deepening seen above the lower parts of the Djupvarp Member contrasts with the shallowing seen in the equivalent Leangen Member in Asker. Baarli (1988) suggested that this discrepancy reflected the shelter effect of a migrating barrier in the Sylling area, subsequently accommodating this idea within her model of a peripheral bulge advancing in front of the Caledonian nappes (Baarli 1990).

Skien–Langesund

Rønning (1979) described the uppermost Ordovician sediments in this area at the southern end of the Oslo Region. There is again a mixed carbonate-siliciclastic succession which, as in northern districts, contains features indicating storm deposition. Sandstones are irregularly distributed and this distribution was thought by Rønning to reflect deposition on a partially fault-controlled topography, shallower in the north and deeper in the south. Rønning suggested that the basin had filled by a gradual southwards progradation but concluded that the sediment source lay north or west of the Oslo Region.

Two limestone members are present within the uppermost Ordovician of Skien–Langesund (Rønning 1979). The younger of these contains structures referred to as 'bioherms' but we have not examined them to determine whether, like those in Hadeland and Ullerntangen, other interpretations might be possible. The upper boundary of this limestone member is a karst surface which again reflects sea-level lowering near the end of the Ordovician. The overlying sediments contain millet-seed quartz grains. In some localities coarse sandstones at the base of the Sælabonn Formation, which is generally similar to that in Ringerike and Hadeland, rest directly on this surface (Worsley et al. 1983).

Swed en

The Upper Ordovician rocks of Sweden are exposed in outliers in Jämtland, Västergötland, Siljan and Scania. Most lie within Jaanusson’s (1976) Central Confacies Belt, but Scania forms part of the deeper western belt which includes the Oslo Region. The Lower Allochthon in Jämtland may also have been part of this western belt (Jaanusson 1982). The whole autochthonous Ordovician sequence is thin in Sweden and contrasts markedly with the thick successions of the Oslo Region. This alone suggests that Sweden lay far from the sediment source, only a little fine-grained siliciclastic sediment reached these areas and carbonate production was low. Interpretations of water depth for lower and middle Ordovician are conflicting (see Lindström 1971; Jaanusson 1982) but faunas from mid-Ashgill mudstones indicate deeper waters than those in the Oslo Region (Owen et al. 1991).

At the end of the Ordovician, conditions changed dramatically with an influx of coarse siliciclastic material in the Lower Allochthon in Jämtland (Karis 1982; Chorns & Karis 1995) and also in Västergötland where oolites and peloids are present (Stridsberg 1980). In Siljan, the top of the Boda Limestone is a karst surface and fissures are filled with graptolitic shales formed after the Silurian transgression (Brenchley 1988).

Sedimentology of the Oslo Region around the Ordovician–Silurian boundary

Sedimentation pattern

Towards the end of the Ordovician the Oslo Region Basin contracted as sea level fell. The Mjøsa districts and presumably areas to the north became emergent, although the precise timing of this emergence is unclear. During the Rawtheyan, a shallow shelf was established to the east of Hadeland which generated carbonate sediments and intermittently received siliciclastics during the latest Ordovician (‘Hirnantian) and early Silurian. To the south the basin persisted, deepening through areas now occupied by Hadeland, Ringerike, and Oslo–Asker. However, in Asker and Ringerike shallower areas formed sites for carbonate deposition away from the eastern edge of the basin. These features are best explained in relation to tilted fault blocks (Fig. 5) and were modified at least locally by fault-controlled highs within the basin (e.g. Stanistreet 1983, 1989).

Throughout the Oslo Region, successions show evidence of this late Ordovician regression. Channels incised into the shelf margin provided conduits for the transport of lithified material, forming the debris-flow deposits of Hadeland, and similar boulder deposits in Ringerike, Oslo–Asker and possibly Skien–Langesund. The shedding of boulders from a carbonate platform may have occurred during episodes of relative high stand as discussed by Schlager et al. (1994). As sea level continued to fall some of these deposits were exposed, as in Hadeland and Ringerike, and karst surfaces developed on them. However, the regression was interrupted by a minor transgression recorded in Hadeland and Ringerike in the filling of karst fissures with crinoidal grainstones. The culmination of the regression was marked by emergence and karst development in northern Hadeland (extending that which had already begun in Mjøsa) and along the western margin of the basin immediately to the south. In western Oslo–Asker, shallowing is marked by faunal changes, by the generation of oolites, and by the formation of deep incised channels. Millet-seed sands were transported as far south as Skien–Langesund and filled cavities in the tops of previously emergent limestones. Changes in the diagenetic environment resulting from these sea level changes are reflected in Hadeland in variations in calcite cement and dolomite sequences.

During the latest Ordovician, substantial quantities of siliciclastic material were introduced into the Oslo Region for the first time. In Hadeland, this was demonstra-
bly transported from the east and may have originated from erosion of the Hedmark Group and Precambrian basement ahead of the advancing nappes to the north. In Hadeland at least, sands were deposited in lobes. Lobe distribution was related to bypass gaps in a rim which may have been emergent, and siliciclastic grains were variably mixed with shelf-generated carbonate. Sea level may have fluctuated at this time but waters were always shallow and sediments were predominantly storm and tide-influenced.

Transgression in the early Llandovery, possibly in response to wasting of the Gondwanaland ice cap, reintroduced fine sands from the uplifted source to the north. Fine sands were deposited in Hadeland, Ringerike and Skien–Langesund, forming the Salabon Formation, while shale and silt deposition dominated in Oslo–Asker, with variable amounts of limestone formed in Asker. The maximum flooding eventually resulted in the drowning of areas to the north and deposition of the upper part of the Salabon Formation on the eroded mid-Ordovician Mjøsa Limestone.

As the siliciclastic supply decreased, carbonates were again prominent in the Rytteråker Formation. Möller (1989) suggested that the reappearance of carbonates might reflect a climatic change and Aldridge et al. (1993) included the sedimentological and concomitant faunal changes in their model linking climate and oceanic state. The shelf model proposed here allows such lithological changes to occur rapidly in response to small eustatic changes or fault movements at the basin margin.

**Sedimentation controls**

Part of the increasing complexity of depositional environments towards the end of the Ordovician may be attributed to tectonic activity in the developing Caledonides (Bjerlykke 1983). This activity included the closure of both the Tornquist Ocean to the (present) SW (Cocks & Fortey 1990; Fortey & Cocks 1992; Oliver et al. 1993) and of the Iapetus Ocean to the (present) NW. Roberts (1988) noted the protracted and polyphasic character of the orogeny and Bjerlykke (1974) demonstrated that the Ordovician sedimentary rocks in the region carried an increasingly ‘oceanic’ geochemical signature with time. Bassett (1985) demonstrated a progressive SSE progradation of terrestrial ‘Old Red Sandstone’ facies across Baltica during the Silurian.

Brenchley & Newall (1980), and Brenchley & Cocks (1982) stressed the importance of sea level changes in controlling late Ordovician sedimentation in the Oslo Region. The global regression in the Hirnantian, resulting from the very brief glaciation of Gondwanaland, generated a fall in sea level of 45–60 m (Brenchley et al. 1994 and references therein). A fall of this magnitude must have been significant. It would have exposed some areas, rejuvenated river systems, and brought siliciclastic source regions previously below wave base within reach of transport. In the Oslo Region it resulted in the deposition and reworking of significant amounts of siliciclastic material. Superimposed on the subsequent transgression, uplift of areas adjacent to the basin (probably as a result of the encroaching Caledonide nappes to the northwest) maintained a supply of siliciclastic sediment to parts of the basin for much of the early Llandovery and periodically thereafter.

Faulting has been considered an important influence on late Ordovician sedimentation in the Oslo Basin (Brenchley & Newall 1980; Brenchley & Cocks 1982; Stanistreet 1983, 1989) (Fig. 5). All of these have regarded faulting as the product of basement tectonics and Stanistreet (1983) suggested that the early Palaeozoic history of the area might be compared with that of the Tertiary of the North Sea, in which subsidence of a graben system allowed further basin development. This assumes that the Oslo Region was in a tensional regime during the latest Ordovician and early Silurian, which seems likely, but since, on a larger scale, Baltica was bounded to the NW and possibly also the SW by destructive margins, the detail must have been more complex.

The E–W asymmetry of the Ordovician–Silurian basin of Hadeland, and N/NE source of siliciclastic sediment determined from lithofacies distribution was probably fault controlled. Major N–S faults parallel to the present margins of the Oslo Graben may have been reactivated in the Permian, although not necessarily with the same sense of movement. Strike-slip movement during the late Ordovician could have permitted uplift of the Mjøsa areas simultaneous with continued subsidence of the basin further south. Crowell (1974), discussed basin formation by such strike-slip movement along a non-planar fault.

The Upper Ordovician platform sequences in Sweden are significantly different from the Ordovician deposits in Hadeland and could not have been the source for debris-flow deposits. Such blocks might have formed >100 km to the north but this would require the successions now preserved in Hadeland to have moved considerable distances southwards to their present positions whereas the recognized folding and faulting account for only a few tens of kilometres of displacement. Carbonates may have formed between the Oslo Region and the nearest Ordovician sediments in Sweden. The evidence from the debris flows in Hadeland indicates the existence of a mature and diverse carbonate platform, close to sea level, only 5–10 km to the east (Braithwaite & Heath 1992).

The Mjøsa area, north of Hadeland (Opalski & Harland 1981) was probably emergent throughout deposition of both the Kalvsjeen and Klinkenberg formations. In Hadeland there is evidence of two distinct periods of emergence. The first, interrupting, or more likely terminating deposition of the uppermost Kalvsjeen Formation, was probably the result of a relative sea level fall and resulted in shelf instability and deposition of the debris flows. The second may have been a result of
faulting, with downwarping of the southern and western margins of the basin and uplift of the northern and eastern margins. Bjørlykke (1983) suggested that uplift and exposure of the Mjøsa area was synchronous with subsidence further south.

Petrography shows that the uppermost Ordovician and lower Silurian sandstones in Hadeland are not flysch in character. Their composition provides ambiguous evidence of provenance. The absence of lithic grains or mafic minerals suggests that they were second cycle sediments but the unweathered feldspars indicate either rapid unroofing or arid erosion of a granitic igneous or metamorphic complex. The only older sedimentary or metasedimentary rocks known in this part of Scandinavia are the possible aulacogen sequences in the Hedmark Basin, now at the northern end of the Oslo Graben. Uplift of these and crystalline basement might have been caused by movement of the nappe pile developing to the NW but would have been enhanced by the general lowering of sea level at the end of the Ordovician. The millet-seed sands must have been transported hundreds of kilometres across a mature landscape which probably lay to the north or NE. The sources of siliciclastic sediment in Hadeland may well have also been sources for other districts in the Oslo Region but this requires petrographic confirmation.

In sequence stratigraphical terms, the top of the Ordovician in the Oslo Region could be regarded as a Type 1 sequence boundary associated with the well-established glacio-eustatic global sea-level fall. However, the detailed picture is much more complicated. The combined effects of tectonics on the margins of Baltica, basement tectonics.
within the Oslo Region, and global eustatic fluctuations generated a complexity in sedimentary pattern across the Ordovician–Silurian boundary never before experienced in the Oslo Region.

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