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Models for the Ellesmerian mountain front in North Greenland: a basin margin inverted by basement uplift

N. J. Soper

Department of Earth Sciences, The University, Leeds LS2 9JT, U.K.

and

A. K. Higgins

Geological Survey of Greenland, Ostervoldgade 10, DK 1350, Copenhagen, Denmark

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Abstract—In North Greenland an early Palaeozoic trough sequence was compressed against a carbonate shelf which flanked it to the south during the mid-Palaeozoic Ellesmerian orogeny. Excellent N–S fiord sections reveal the frontal structures and good stratigraphic control permits their interpretation to depth. In the east, the structures are thin-skinned, with S-vergent folds and thrusts. Traced westwards these structures pass into a mountain front monocline which attains a maximum amplitude of 7 km. Attempts to interpret the structure in terms of familiar thin-skinned mountain front models do not lead to credible deep sections. It is necessary to invoke basement uplift, with an early Cambrian basin margin extensional basement ramp becoming reactivated in compression during the Ellesmerian orogeny. Four restorable deep sections are presented to illustrate this new mountain front model. If realistic, these show that horizontal displacements were modest, of the order of tens of km.

The thin-skinned zone to the east has a hitherto unexplained southward tilt towards the foreland; this may be due to weak reactivation of a gently inclined segment of the basement ramp. The ramp geometry at depth along the four sections is modelled from the shape of the deformed basement–cover interface; the amplitude of the monocline is seen to be controlled by the inclination of the ramp and thrust displacement on it.

There are also implications for the amount of Cenozoic displacement on the Nares Strait lineament between Greenland and Ellesmere Island, a topic of long-standing controversy. The monocline ends at the strait and the Ellesmerian structures revert to thin-skinned on the Canadian side. It is inferred that a transfer fault existed there during the early Cambrian extensional phase, terminating the basement ramp, and which on Ellesmerian reactivation produced the swing in strike and apparent offset of geological markers which some workers believe to be due to post-Ellesmerian strike-slip. If correct, this means that Cenozoic displacement along the strait (Wegener Fault) was limited to a few tens of km.

INTRODUCTION

In the early Palaeozoic a marine basin extended E–W across the northern part of Laurentia, through the area now occupied by the Queen Elizabeth Islands of Arctic Canada and northern Greenland (Fig. 1). The deep-water part of this Franklinian Basin was flanked to the south by a stable platform, and deposition was terminated in Devonian time by the Ellesmerian orogeny (Trettin & Balkwill 1979, Surlyk & Hurst 1984). In Greenland the Ellesmerian orogen is traditionally known as the North Greenland fold belt (Fig. 2). In the northern 'orthotectonic' part of the fold belt, polyphase N-vergent structures are developed in low amphibolite facies metasediments on the north coast of Johannes V. Jensen Land, and both the deformation and metamorphism decrease southwards (Dawes & Soper 1973, Higgins *et al.* 1982). South of a divergence zone (Fig. 2) structures verge south, towards the platform, and take the form of a thin-skinned fold and thrust zone in which

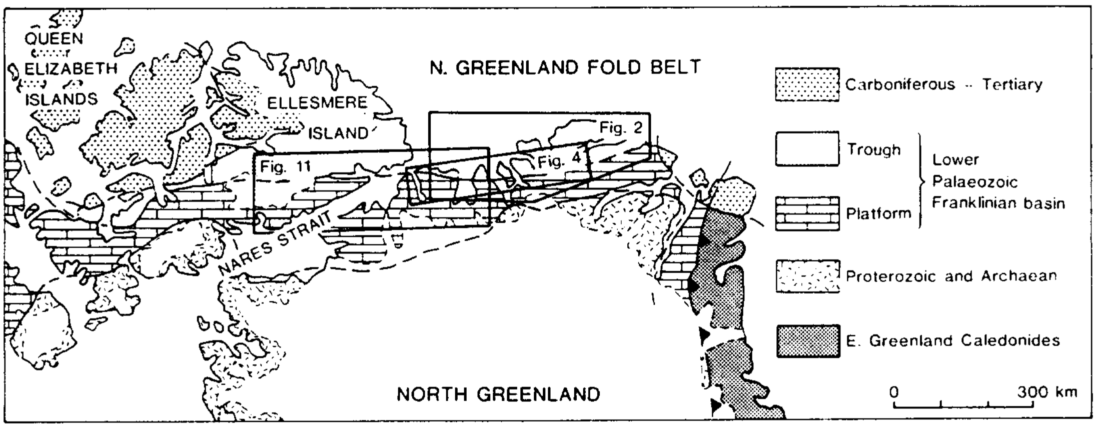

Fig. 1. Location map of Greenland and the adjacent Canadian Arctic showing the Ellesmerian orogen.

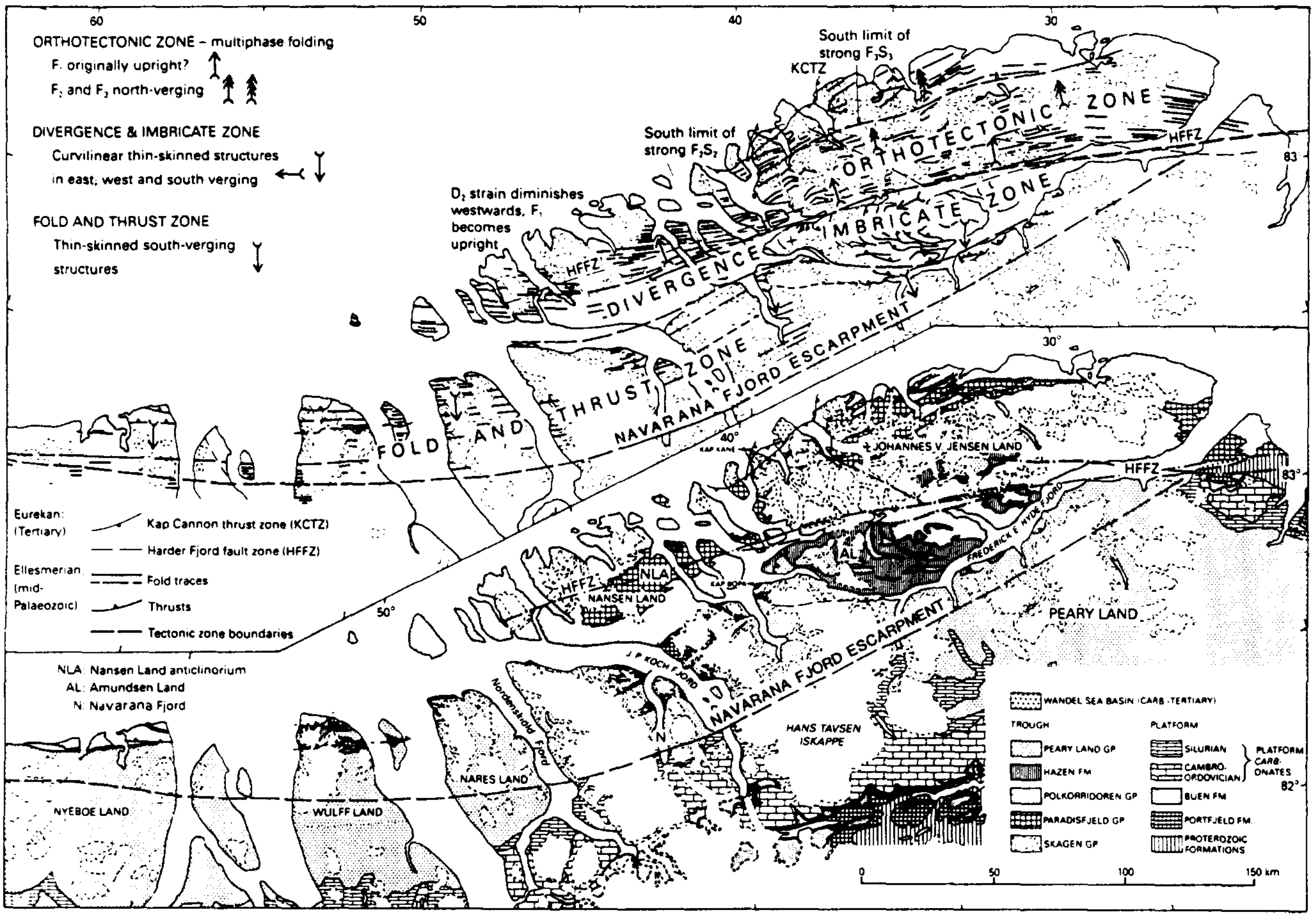

Fig. 2. Structural and stratigraphic map of North Greenland, after Soper & Higgins 1987.

the basinal sediments were compressed against the platform margin (Soper & Higgins 1985, 1987). Traced westwards from the central part of North Greenland, the Ellesmerian margin changes character to become a major mountain front monocline which attains an amplitude of some 7 km at its greatest development in Wulff Land (Figs. 2 and 7).

This paper describes the variation in structural style of the Ellesmerian front by reference to four cross-sections based on field observations made in 1979–1980 and 1984–1985 during a regional mapping programme of the Geological Survey of Greenland. Excellent exposure of the structures is provided by deeply incised N–S fiords and stratigraphic control is good. We have attempted to model the deep structure of the mountain front in terms of restorable sections with minimum displacement. It proves necessary to invoke the involvement of crystalline basement in the formation of the monocline. The structure is thought to have been controlled by basement faults which were active in extension during southward expansion of the basin, and were subsequently reactivated in compression during its inversion.

STRATIGRAPHIC SETTING

From earliest Cambrian to early Silurian time a distinction existed in the Franklinian Basin between the trough, in which more than 8 km of turbiditic and hemipelagic sediments were deposited, and the shelf to the south, on which accumulated a thinner, carbonate-dominated sequence. An outline of the trough stratigraphy in North Greenland is provided by Friderichsen *et al.* (1982). Models for the evolution of the Franklinian Basin in North Greenland have been presented by Surlyk *et al.* (1980), Surlyk & Hurst (1983, 1984) and Higgins *et al.* (in press), to which the reader is referred for details. A simplification of these models (Soper & Higgins 1987) envisages three stages: a period of rapid fault-controlled extension in the early Cambrian, during which up to 4 km of turbidites were deposited; a long period dominated by thermal subsidence in which a thin 'starved basin' sequence accumulated; and a second period of turbidite deposition in the Silurian when a cumulative thickness of some 5 km was laid down, eventually swamping the shelf. The source of the Silurian turbidites is thought to have been the rising East Greenland Caledonides (Hurst *et al.* 1983), that of the early Cambrian turbidites is unknown. During the early Palaeozoic the trough expanded southwards in several stages by foundering of the platform margin along E–W lineaments which are presumed to have been fault controlled (Surlyk & Hurst 1983, 1984). In the present context the most important of these is the Navarana Fjord lineament or escarpment (Figs. 2, 3 and 4) which formed the platform margin in early Silurian time and is discussed below.

Distribution of the main stratigraphic units is shown in

Fig. 3. Note that no attempt has been made to restore the facies boundaries to their pre-Ellesmerian positions. For a discussion of this problem in relation to the early Cambrian platform margin, and of the influence of that margin on subsequent Eurekan (Tertiary) structures, see Soper & Higgins (1987).

Skagen Group (Fig. 3a)

The oldest exposed sedimentary rocks in the Franklinian Basin of North Greenland belong to the Skagen Group (Friderichsen *et al.* 1982) and are loosely dated as latest Proterozoic to earliest Cambrian, equivalent to the Kennedy Channel Formation of Arctic Canada (Kerr 1967). The best known Greenland exposures, in northern Wulff Land, comprise at least 600 m of siliciclastic and carbonate sediments of marine shelf origin (Surlyk & Ineson 1987). The base is not seen and a considerably greater thickness may be present at depth in the northern peninsulas of Greenland, since the Kennedy Channel Formation attains at least 1200 m in Ellesmere Island.

The Skagen Group thins southwards and is absent in southern Wulff Land where the overlying Portfjeld Formation rests directly on crystalline basement. As explained below, our model for the mountain front monocline requires the reactivation of a basement ramp which was active in extension during Skagen deposition.

Portfjeld Formation and Paradisfjeld Group (Fig. 3b)

Carbonate deposition followed the Skagen clastics in the Lower Cambrian and it is possible to define a facies boundary across North Greenland between shelf carbonates of the Portfjeld Formation to the south (generally 200–325 m thick) and the equivalent slope sequence of carbonate and siliciclastic muds of the Paradisfjeld Group to the north which is more than 1 km thick. The Portfjeld thickens to more than 500 m towards the shelf edge, but the transition into the trough facies is both foreshortened and obscured by Ellesmerian thrusting. It has been speculated that this hidden boundary was fault controlled, the deep fractures becoming reactivated in strike-slip mode to produce the Tertiary Harder Fjord fault zone (Surlyk & Hurst 1984, Soper & Higgins 1987). Syndepositional extensional faulting was perhaps responsible for the regressive episode recorded at the top of the Portfjeld Formation near the shelf margin (Surlyk & Ineson 1987).

The Portfjeld carbonates are a competent horizon compared to the more shaly units above and below, so that they tend to control buckling and, when involved in thrusting, contain ramps rather than flat detachments.

Buen Formation and Polkorridoren Group (Fig. 3c)

The regression above the Portfjeld Formation heralded a major change in sedimentation. A thin sequence

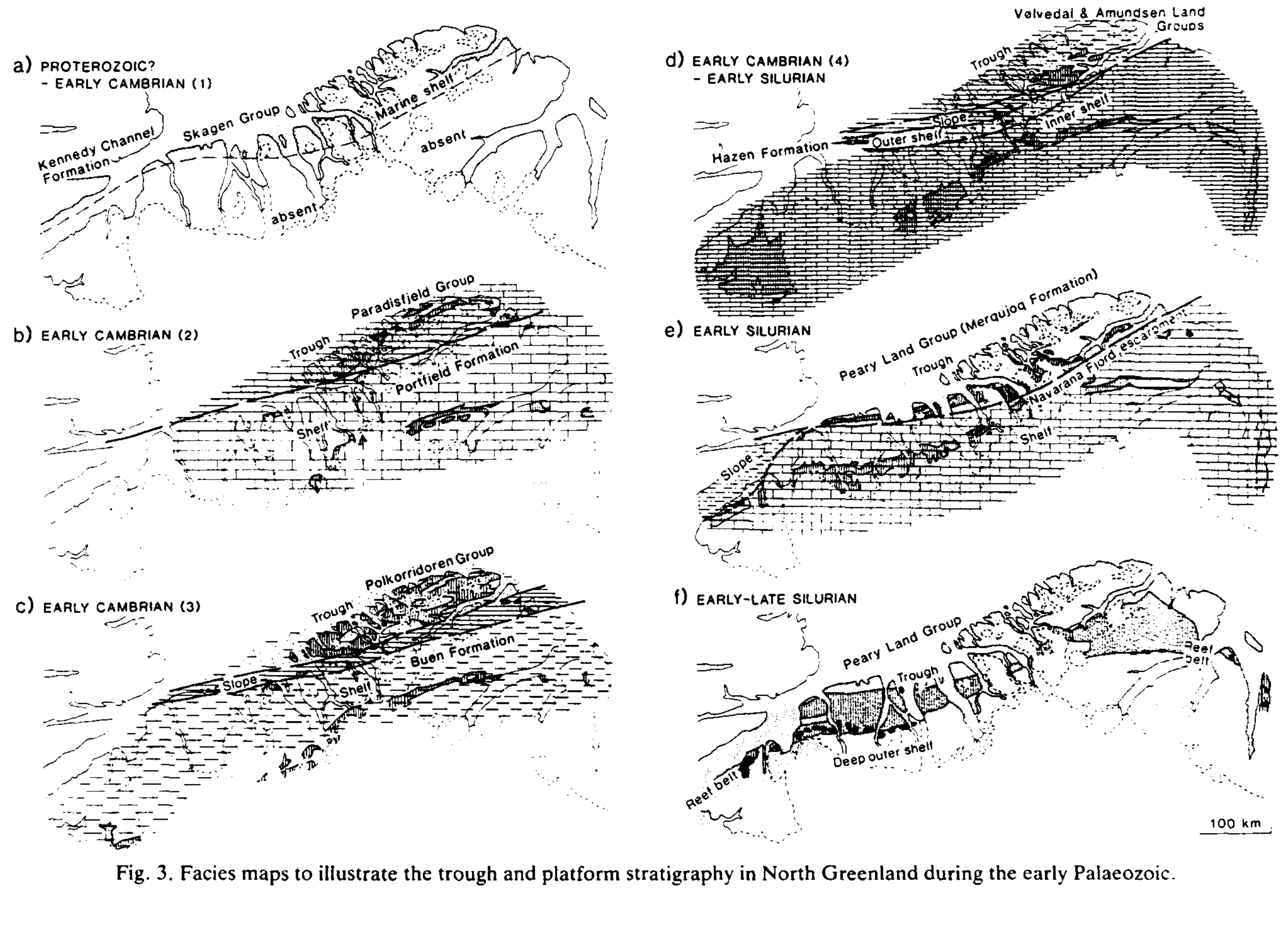

Fig. 3. Facies maps to illustrate the trough and platform stratigraphy in North Greenland during the early Palaeozoic.

of shales and deltaic–marine sandstones of the Buen Formation (250–375 m) was deposited on the shelf while 2–3 km of arkosic turbidites of the Polkorridoren Group accumulated rapidly in the trough (Davis & Higgins 1987, Surlyk & Ineson 1987). The trough turbidites extend westwards into Ellesmere Island as the Grant Land Formation (Trettin 1971) and represent a huge accumulation of first cycle sediment derived from an unknown granitic gneiss source.

Transitional sequences 500–700 m thick are referred to the slope (Fig. 3c); these were displaced southwards over normal Buen Formation by Ellesmerian thrusts. The trough sequence was likewise displaced over the slope, so that the transition is not exposed at the present erosion level. Fault control is again inferred, with the development of active scarps, as shown by the presence of olistoliths of Portfjeld carbonates within the Polkorridoren sequence (Friderichsen & Bengaard 1985, Soper & Higgins 1987).

The shale-dominated Buen Formation on the outer shelf is thought to have provided a detachment horizon during Ellesmerian deformation, mainly below the present level of exposure, while the interbedded sand turbidites and shales of the slope deformed in typical fold and thrust mode. At the top of the trough sequence is a thick purple and green mudstone unit, the Frigg Fjord Mudstone. Detachments in this horizon produced the spectacular arcuate imbricate structures of Amundsen Land (Fig. 2) which were described by Pedersen (1980, 1986) and are not considered in the present paper.

Cambro-Ordovician shelf and slope sequences (Fig. 3d)

Carbonate deposition resumed on the shelf late in the Lower Cambrian and some 1250 m had accumulated by the end of Ordovician time. On the outer shelf carbonates and carbonate conglomerates were followed by thin chert and black shale sequences; normal thicknesses are 300–450 m, but occasionally attain 600 m where thick conglomerates are present. On the slope and trough margin a condensed sequence was deposited, often consisting of less than 100 m of cherty shales, but with many Ordovician graptolite zones preserved, and apparently representing continuous sedimentation from late in the Lower Cambrian to the Llandovery (Soper & Higgins 1985).

These variegated outer shelf and slope sequences are exposed on the northern parts of the peninsulas between Nyeboe Land and J. P. Koch Fjord (Fig. 2) where they form a useful mapping division between the Buen Formation below and Merqujoq Formation of Silurian turbidites above. This division was recognized by Soper & Higgins (1985) as equivalent to the Hazen Formation of Ellesmere Island (Trettin 1971, Trettin *et al.* 1979). Pending formalization of this part of the stratigraphy in North Greenland we apply the name Hazen Formation informally.

The response of this heterogeneous sequence to compressional deformation is variable. In areas where the lower part of the overlying Silurian turbidites crop out extensively, anticlines cored by outer shelf facies Hazen Formation are sometimes present, of both tip-line and ramp type, and low angle detachments are also exposed within this formation. Farther north, where the Hazen is thin and dominantly cherty, it forms trains of tight cusp-like anticlines and more lobate synclines, presumably reflecting the contrast between competent Silurian turbidites above and less competent transitional Buen Formation below.

Early Silurian facies and Navarana Fjord escarpment (Fig. 3e)

On the inner shelf or platform, carbonate deposition continued during the Llandovery, while a marked change to turbidite deposition took place in the trough. The boundary is a 30–40° escarpment in reefal platform edge carbonates, exposed in Navarana Fjord and J. P. Koch Fjord, against which the turbidites are banked with no tectonic disturbance (Hurst & Surlyk 1984, Surlyk & Ineson 1987). The early Silurian carbonates are about 750 m thick while the equivalent late Llandovery turbidites (Merqujoq Formation) are much thicker—at least 1.8 km in the vicinity of the escarpment and probably more to the north and west (Hurst & Surlyk 1982, Larsen & Escher 1985, Surlyk & Ineson 1987). Differences in sediment thickness across the escarpment are thought to have been controlled primarily by extensional faulting. Hurst & Surlyk (1984) inferred the existence of a deep-seated growth fault, on whose footwall the carbonates accreted and which limited the northward progradation of the platform facies in the Ordovician. We have incorporated this interpretation into our cross-sections (Figs. 6–9). No significant facies change across the Navarana Fjord escarpment can be discerned in the Portfjeld or Buen Formations, so the fault is likely to have been active later, probably at about the Cambrian–Ordovician boundary, controlling the regressive episode recorded on the platform at that time. The escarpment was overlapped by the unit above the Merqujoq Formation (Thors Fjord Member) after which turbidite deposition spread across the whole area in the Wenlock and the Navarana Fjord escarpment ceased to control sedimentation.

The location of the westward, subsurface extension of the Navarana Fjord escarpment is important to our model for the monocline and is discussed below.

Silurian turbidites (Fig. 3f)

The outer platform subsided in the latest Llandovery and turbidite deposition expanded southwards as far as a reef belt, which in the west marked the new northern limit of the outer shelf. A cumulative thickness of some 5 km was deposited, including the Merqujoq Formation which, as explained, is confined to north of the Navarana Fjord escarpment. The preserved sequence extends into the Ludlow, possibly into the Lower Devonian in Hall Land to the west of the study area (Hurst & Surlyk 1982). The Silurian turbidites are less feldspathic and more calcareous than those of the Polkorridoren Group, consistent with derivation from the siliciclastic and calcareous Caledonian nappes to the east (Hurst *et al.* 1983). However, the stratigraphic equivalent in Ellesmere Island, the Imina Formation, is lithologically similar but largely derived from the northwest (Trettin 1971, 1979), implying a different source. Several finer grained intervals are present, providing detachment horizons for shallow accommodation structures, as well as chert conglomerate units.

SOUTHERN MARGIN OF THE NORTH GREENLAND FOLD BELT

As outlined above, the trough–platform transition in North Greenland migrated southwards during early Palaeozoic time. In the early Silurian the platform margin was located at the Navarana Fjord escarpment, as explained above. The scarp is now exposed in the eastern part of the area under consideration (J. P. Koch Fjord and Navarana Fjord, Fig. 4) and its position can be closely inferred in the west, in NW Nyeboe Land (Escher & Larsen 1987). Several attempts have been made to interpolate its subsurface position through the intervening central area. Hurst & Surlyk (1984, fig. 4) implied that it continues en échelon as the Nyeboe Land fault zone of Dawes (1982), a structure which is now known to be simply the steep limb of the mountain front monocline (Soper & Higgins 1985). Escher & Larsen (1987, fig. 5) located it at the northern limit of shallow-rooted box folds to the south of the monocline, on the assumption that those folds were controlled by a detachment at the top of the platform carbonates. Our reconstruction of the deep structure gives a northern limit for the subsurface trace of the escarpment west of Navarana Fjord which is somewhat to the south of that adopted by earlier workers (Fig. 4).

Fold traces in the orthotectonic part of the fold belt (Nansen Land, Johannes V. Jensen Land and the intervening islands, Fig. 2) trend parallel to the Navarana escarpment. In the thin-skinned fold and thrust zone immediately to the north of the lineament, major fold traces and linear steep belts show a similar parallelism to it (Figs. 2 and 4). Clearly, the early Silurian facies boundary did influence the pattern of deformation which developed in the trough sediments as they were compressed against the platform margin. In detail however, the southern limit of Ellesmerian deformation as seen at the surface does not coincide with the Navarana escarpment, and diverges some 10–15 km to the south of it in the west (Nyeboe Land, Fig. 4).

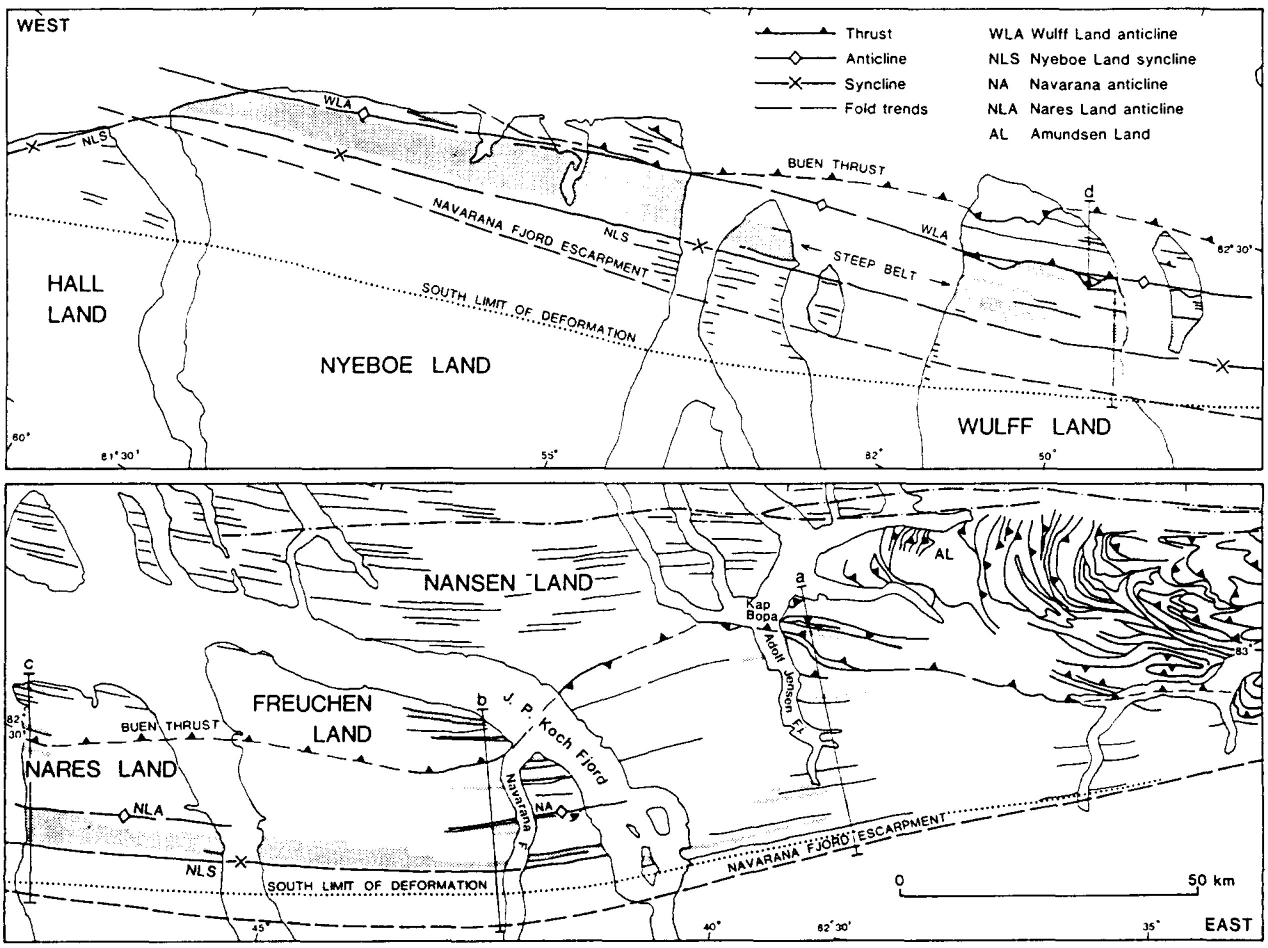

Fig. 4. Structural map of the southern margin of the North Greenland fold belt between Hall Land and Amundsen Land, showing location of cross-sections. Subvertical beds stippled.

Changes in structural style along the Ellesmerian mountain front

In the eastern part of the region, in the vicinity of Adolf Jensen Fjord (Fig. 4), the structure at the southern margin of the fold belt is of typical thin-skinned fold and thrust type. It has been described by Soper & Higgins (1987) and the published cross-section is reproduced here with some changes and additions as Fig. 9. These modifications arise from insights gained from our study of the thicker-skinned structures farther west and are described in a subsequent section. Stratigraphic relationships at the Lower Silurian platform margin escarpment have also been changed to conform with the analysis of Surlyk & Ineson (1987), and the inference that the Navarana escarpment is located above a steep growth fault, as argued above.

At Adolf Jensen Fjord, the southern limit of deformation is located a few kilometres north of the Lower Silurian platform margin at a synclinal up-turn of the gently N-dipping Silurian turbidites, associated with a pair of anticlines to the north. A thrust ramp beneath the northern anticline is exposed some 5 km east of the section line and, on the assumption that shortening across these structures is accommodated at a detachment, an excess area calculation places it at about 2 km depth, in the Buen Formation. The buried tip-line of this detachment must lie between the northern anticline and the Lower Silurian platform margin. Farther north the thin-skinned deformation style is indicated by S-vergent thrusts, tip-line anticlines cored by Hazen Formation and by the complex refolded backthrusts at Kap Bopa, whose restoration has already been presented (Soper & Higgins 1985, fig. 3). Note that in the southern part of this section the stratigraphic level rises to the north; the Wulff Land Formation is below sea level just north of Navarana escarpment but rises through the synclinal up-turn and beyond until it is elevated above the local summit levels. This southward tilt to the whole structure cannot be entirely explained by thin-skinned minor thrusts rooting in an essentially horizontal detachment within the Buen Formation; the detachment itself must be tilted towards the foreland. We address this problem subsequently.

Navarana Fjord, 60 km to the west (Fig. 4), provides the next cross-section (Fig. 8). The northward synclinal up-turn from the gently N-dipping platform is more pronounced, producing a 2.0 km elevation of the Silurian turbidites above their level in the core of the syncline, with an additional 1.5 km due to the Navarana Anticline. This is a large symmetrical box-fold to the north of which medium scale folds verge south and are associated with gently N-dipping thrusts. The up-turn presents problems. The elevation associated with it is maintained for at least 25 km to the north, with the base of the Silurian turbidites remaining close to sea level along the outer part of Navarana Fjord. In principle, this elevation could be accommodated by a major duplex at depth, but this duplex would have to be developed entirely in sub-Portfjeld strata (Skagen Group) and it would have to be at least 25 km long, implying improbably large displacements.

The problem is more acute in the Nares Land section, a further 75 km to the west (Fig. 4). This section (Fig. 6) shows that the synclinal up-turn has now developed into a mountain front monocline giving an elevation of the Silurian strata 3.0 km above platform regional. Westwards from Nares Land, the monocline is defined by a steep, foreland-dipping or vertical panel of strata, the Nyeboe Land steep belt or linear belt of Dawes (1982). Its amplitude increases to 6.5–7.0 km in Wulff Land, 30 km from the Nares Land section (Fig. 7), and then decreases to 5.5 km in eastern Nyeboe Land, a further 75 km to the west. In both areas Lower Silurian strata are exposed overlying the platform to the south of the monocline, Lower Cambrian strata to the north. Since in Wulff Land and Nyeboe Land the upper limb of the monocline is deformed by later folds and thrusts, we return to the Nares Land section to define the problem posed by this major example of a mountain front monocline.

Figure 5(a) shows the monocline as developed in Nares Land, in simplified form. The two component fold hinges have been named the Nares Land Anticline (NLA) and Nyeboe Land Syncline (NLS—these fold traces are located on Fig. 4) and the steep, foreland-dipping panel between them is not homoclinal in this section but contains an additional fold pair. South of the monocline the exposed Silurian strata are affected by box folds, presumably associated with a shallow detachment not far below fiord level. The monocline, of much larger scale, must be related to a deep detachment. North of the monocline, asymmetric S-vergent folds and thrusts are developed in Lower Cambrian strata, presumably above a detachment at intermediate depth. The problem is simply stated: what fills the space below the elevated tract? Larsen & Escher (1985) have illustrated the problem in their fig. 8, which relates to Nyeboe Land.

Models for the mountain front monocline

Vann *et al.* (1986) have offered four possible solutions to the general mountain front problem (see their fig. 2). Two involve emergent forethrusts ahead of the monocline. These can be eliminated immediately since no major emergent thrusts are observed, nor is any post-Ellesmerian molasse present which might mask them. The third possibility involves an antiformal stack associated with a buried tip line, as illustrated in Fig. 5(b). The main objection to this has already been raised in relation to the Navarana Fjord up-turn: the structural level to the north of the monocline declines very gently, so to fill the space the antiformal stack would have to take the form of a long duplex, which implies excessively large displacements and probably requires a passive roof backthrust as described by Banks & Warburton (1986). The hidden duplex could not be composed of the Portfjeld, Buen or Hazen Formations since these are exposed in

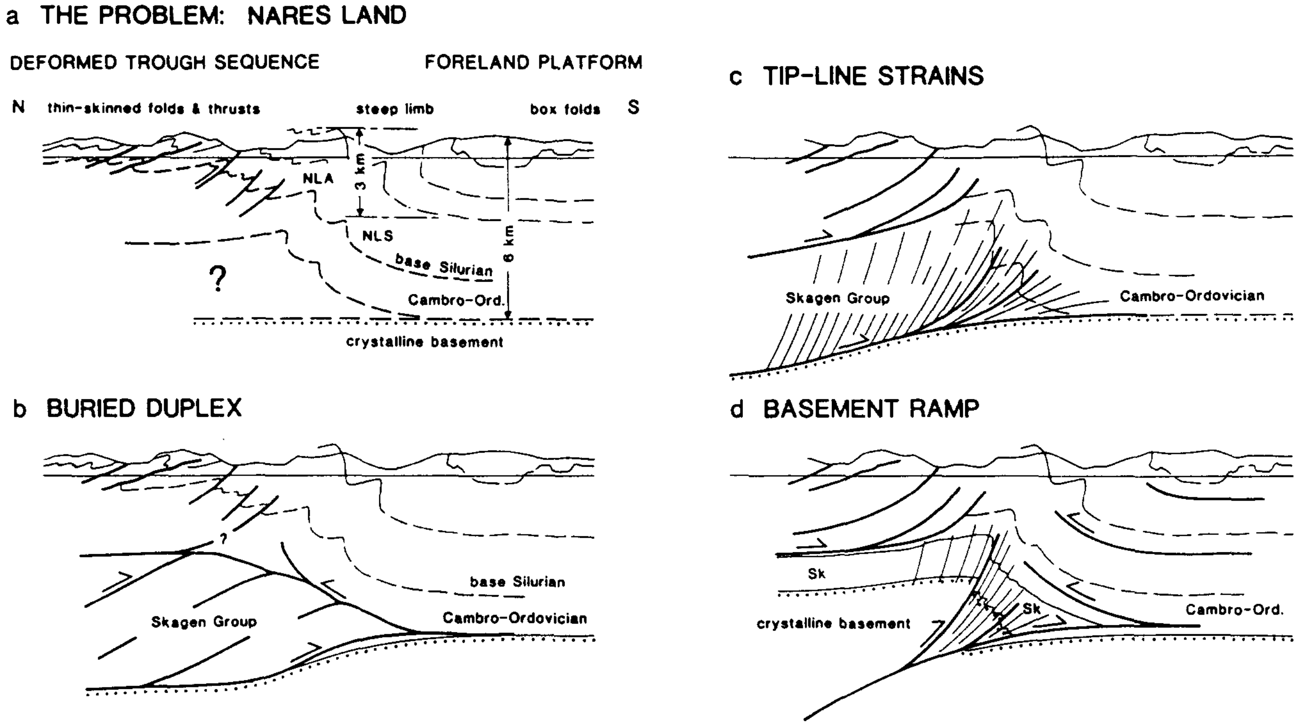

Fig. 5. The mountain front monocline in Nares Land: (a) basic geometry and the problem; (b) a thin-skinned interpretation involving a passive-roof duplex; (c) an interpretation involving loss of displacement to a buried tip-line; (d) the preferred solution: basement uplift. For discussion see text.

several places (Navarana Anticline, Wulff Land Anticline) and show no evidence of backthrusting or involvement in a major duplex. It would therefore be necessary to infer a major duplex development entirely in sub-Portfjeld (Skagen Group) strata.

A fourth possible solution involves rapid loss of displacement associated with the buried tip-line of a major forethrust (Fig. 5c). The thickness of the known pre-Silurian stratigraphic succession (about 1.5 km) would need to be doubled to achieve an uplift of 3.0 km, but to maintain this for an indefinite distance to the north seems unrealistic; to produce almost 7 km of uplift in Wulff Land by this mechanism is impossible. Thickening of the sub-Portfjeld sequence into the trough is again a possibility (Fig. 5c), but does not ease the problem, since before the Ellesmerian deformation the basement surface would have descended northward to accommodate the extra sediment, thus increasing the area to be filled.

The only realistic solution to the space problem is to invoke basement uplift to the north of the monocline. Basement uplift mechanisms have been extensively studied in the Rocky Mountain foreland region under the stimulus of petroleum exploration and a range of geometries have been described, from block uplifts on steep faults of possible strike-slip origin to fold-thrust uplifts on inclined basement ramps (see review by Lowell 1985, Chapter 3 and references therein). There is no evidence of major strike-slip or block faulting in this part of the Ellesmerian orogen, so a thick-skinned mechanism is not appropriate, but there is ample evidence of thin-skinned thrusting. As a hypothesis it is therefore proposed that a shallow basement thrust of Wind River type (Bally 1981) propagated into the cover as a monoclinal uplift. We have no drill-hole or seismic evidence to support this interpretation, but we have established that it is geometrically feasible, by developing a restorable model for the deep structure, using iterative line and area balancing.

The hypothetical ramp would have been located parallel to, and in the vicinity of, the early Palaeozoic basin margin, so it may have been active in extension during sedimentation. Figure 5(d) illustrates the basic geometry of this solution, as applied to Nares Land. As explained in more detail below, the model involves thrust reactivation of the basement ramp, with loss of displacement to a buried tip line producing considerable ductile strain accommodated within a triangle zone, uplift on steep splays from the ramp, and also shallow out-of-syncline backthrusting to accommodate flexural strains ahead of the main monocline. This combination of mechanisms is necessary to account for the observed structures, within the constraints provided by good stratigraphic control. The solution is not unique, mainly because the depth of the basement–cover interface north of the monocline is unknown and therefore we do not know the thickness of the sub-Portfjeld (Skagen Group) sediments. We have produced a conservative solution in terms of depth to basement and displacement on the ramp, but one which requires considerable distortion of the leading part of the basement wedge.

We now apply this model to the four cross-strike sections mentioned above, discussing them and their pre-Ellesmerian restorations in turn, starting with the Nares Land section, moving westwards to Wulff Land where the mountain front monocline reaches its greatest development and then returning to consider Navarana Fjord and Adolf Jensen Fjord where the deformation is essentially thin-skinned.

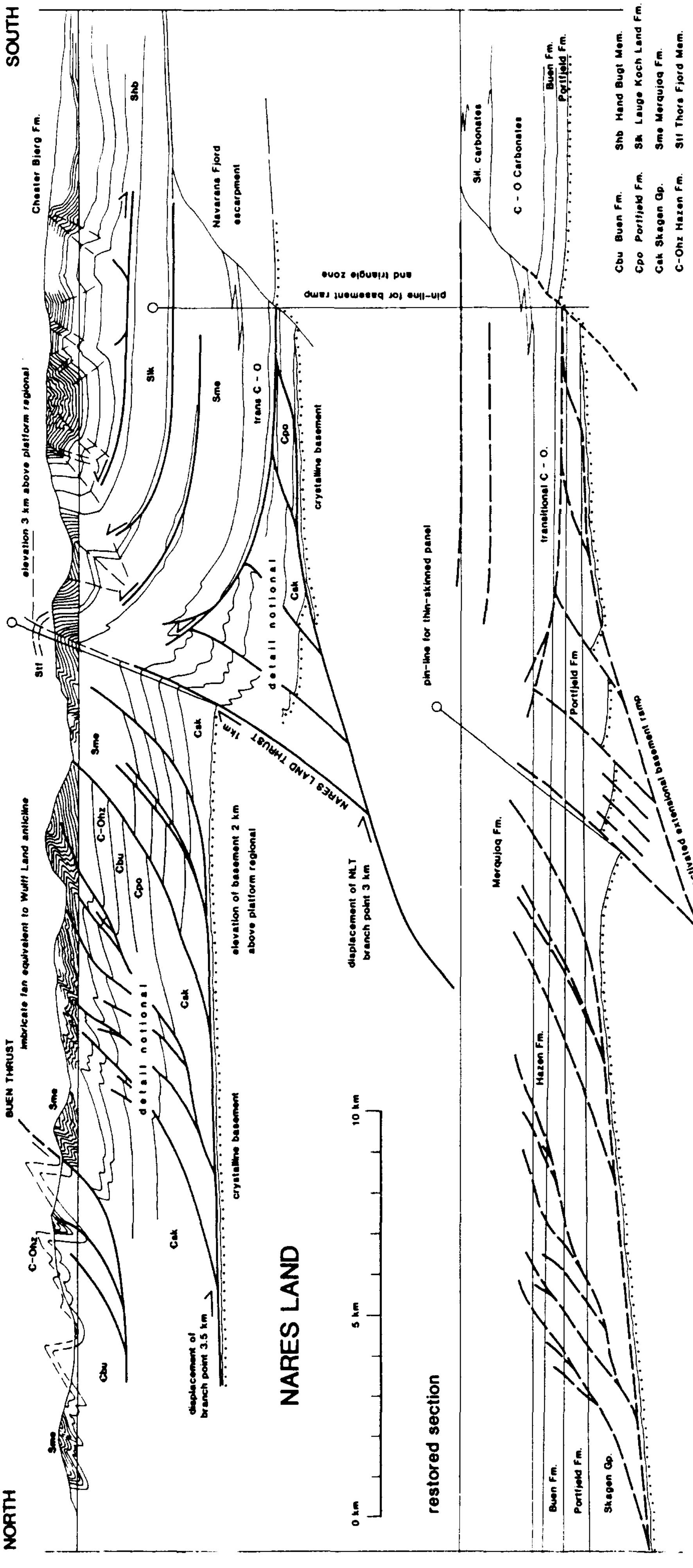

Fig. 6. Interpretive deep section through the Ellesmerian mountain front in Nares Land and (below) restored section.

STRUCTURAL CROSS-SECTIONS

The sections have been constructed from field observations made by helicopter and foot traverses, greatly aided by aerial photography of the fiord walls undertaken during Twin Otter flights by Niels Henriksen (GGU, North Greenland Project Leader) and Jakob Lautrup (GGU, photographer).

Nares Land (Fig. 6)

The west coast of Nares Land provides a N–S section through the Ellesmerian front which is almost continuously exposed for more than 35 km. Five main structural zones can be recognized. From south to north these comprise sub-horizontal platform strata at the top of the Silurian turbidite sequence (Chester Bjerg Formation); a zone of box folds on several scales; the mountain front monocline with a 4 km wide steep zone which produces an elevation of 3.0 km; a long section of gently N-dipping Lower Silurian turbidites (Merqujoq Formation), deformed by asymmetric, S-vergent folds; and in the north a zone of spectacular S-vergent folds and thrusts in which the anticlines are cored by the pre-Silurian Hazen and Buen Formations. The latter is much thicker than on the platform and is transitional to the basinal turbiditic Polkorridoren Formation.

The monocline is sufficiently well exposed to enable its geometry at depth to be reconstructed in terms of the basement ramp model described above, using the standard technique of iteration between the deformed and restored sections. The axial surface of the main anticline (Nares Land Anticline) dips north at 65°. Displacement on the ramp is clearly less than its length. The kink-fold construction appropriate to this situation (Suppe 1983, fig. 3), whereby the position of the ramp top is located by projecting the axial surface of the main anticline to depth, produces insufficient cross-sectional area in the triangle zone beneath the up-turn. The axial surface, passing into an upthrust at depth (the Nares Land Thrust or NLT, Fig. 6), must therefore root on the ramp some distance to the north of the ramp top. Some indication of the restored position of the NLT can be gained from an area balance of the Hazen, Buen and Portfjeld Formations (Fig. 6). Unless its location is random, the NLT must have originated as a syndepositional extensional splay from the ramp. Extensional faults of this type could accommodate a northward-thickening wedge of Skagen Group sediments beneath the Portfjeld.

The subsurface geometry of the mountain front monocline has been reconstructed (Fig. 6). It is inferred that the main detachment on the basement ramp loses displacement southward to a tip located at the buried Lower Silurian platform margin, climbing into Buen shales where the Skagen pinches out. Strains associated with loss of displacement on this thrust must provide sufficient stratal thickening in the wedge beneath the up-turn to give the observed monocline geometry. A backthrust is inferred to separate strongly deformed rocks in the wedge from less deformed rocks above in the up-turn, making the wedge a triangle zone. The details shown within the triangle zone are simply to complete the section and to illustrate that a line balance can be achieved. The position of the tip-line, and hence of the Lower Silurian platform margin in the subsurface, is constrained to the north by the area balance of the triangle zone. If this reasoning is correct, the platform margin is located some 5 km further south than the southern limit of shallow rooted box-folds.

Area balance of the triangle zone also allows the ramp angle to be determined. In the Nares Land section a dip of about 17° is adopted for the upper part of the ramp, the deep trajectory of which is considered in a later section. The area balance also permits restoration of the NLT branch point on the basement ramp. A 3 km displacement of the branch point and 1 km on the upthrust itself combine to give a basement elevation of about 2.5 km. The remaining uplift to give 3.0 km elevation can be accommodated by compressional strains in the cover associated with steepening of the NLT during formation of the triangle zone.

The panel of S-dipping Silurian strata which forms the steep limb of the monocline is interrupted by an anticline which we infer to be associated with a bedding-subparallel backthrust within the Merqujoq Formation. Box-folds in the Nyeboe Land Formation south of the main syncline are similarly developed above a shallower detachment, in the shaly Hand Bugt Member. These detachments are viewed as typical out-of-syncline thrusts which accommodate inner-arc compressional strains in the main synclinal up-turn, not forethrusts at the top of the buried shelf as suggested by Larsen & Escher (1985, fig. 8).

Turning now to the northern end of the Nares Land section, it can be inferred that the spectacular train of S-vergent folds exposed on the west coast is associated with a detachment about 1 km below sea level. The anticlines are cored by transitional Buen Formation at least 1 km thick in this area. The detachment is therefore probably located in shales near the base of this formation and may be called the Buen Thrust. It is thought to emerge at the southern end of this fold train as a steep thrust which juxtaposes Hazen Formation to the north in the hangingwall with Merqujoq Formation in the south. The Buen Thrust is clearly an important dislocation, traceable across the whole area and truncating the slope–outer shelf transition in Cambro-Ordovician strata (Fig. 4). In the fold train exposed in northern Nares Land this thrust accommodates about 5.5 km shortening over a horizontal distance of 8 km.

Between the emergent Buen Thrust and the main monocline, Lower Silurian strata of the Merqujoq Formation dip generally north at about 20° on the north limb of the Nares Land Anticline, as described above. This homoclinal region is affected by S-verging fold pairs with several exposed minor thrusts, with about 2.5 km shortening. Area balance considerations require these structures to involve thick Skagen Formation at depth, implying that their controlling detachment is at the top of the crystalline basement. The buried tip of this inferred basement–cover detachment defines the southern thin-skinned thrust front in Nares Land; all structures to the south of it are associated with the mountain front monocline. The folds and thrusts associated with this buried detachment in Nares Land are along strike from the Wulff Land Anticline to the west (Fig. 7), in which, as explained below, the basement–cover detachment is emergent and is inferred to post-date the formation of the mountain front monocline.

Wulff Land (Fig. 7)

This section differs from that of Nares Land in two main respects. The elevation produced by the mountain front monocline is much greater, some 7 km, and the upper part of the monocline is modified by a major anticline, the Wulff Land Anticline (WLA), which is developed in part as a hangingwall ramp anticline above an emergent thrust, the Wulff Land Thrust (WLT, Fig. 7). The north limb of the Wulff Land Anticline dips north at about 45°. This implies that the controlling thrust ramp has a similar inclination. Since at least half a kilometre of Skagen Group is exposed in the core of the WLA, the thrust must lie deeper than this, presumably near the basement–cover interface. Because the WLT is not folded round the monocline, but truncates vertical strata associated with it, it must be interpreted as later, displacing the NLT. An additional complication is that the upper section of the latter thrust has been reactivated in extension.

Reconstruction of the subsurface structure follows the Nares Land interpretation, with purely illustrative detail added. Again, a major backthrust is inferred to exist beneath the up-turn, mainly in Buen shales and not quite emergent at the present erosion level, on which root tight chevron folds in the Merqujoq Formation. As in Nares Land, the subsurface position of the Lower Silurian platform margin inferred from the triangle zone area balance cannot lie as far north as the onset of shallow box-folding.

The restored section (Fig. 7) allows an estimate to be made of the displacements involved in the formation of the monocline at its maximum development. The NLT has a displacement of 0.5 km and its branch point has been displaced 7 km up a steep section of the ramp, giving an elevation of the basement of about 5.5 km above its original level and about 4 km above its reference level at the top of the ramp. Additional uplift to achieve the 6.5–7.0 km amplitude of the monocline is achieved by displacement on the WLT and ductile strains associated with steepening of the NLT splay.

Nyeboe Land

In Nyeboe Land (Fig. 4) an along-strike equivalent of the Wulff Land Anticline is not seen at the present erosion level. Instead, an out-of-sequence thrust, which appears to be the westward continuation of the Buen Thrust of Nares Land, overrides the upper limb of the main monocline and precludes analysis of its geometry.

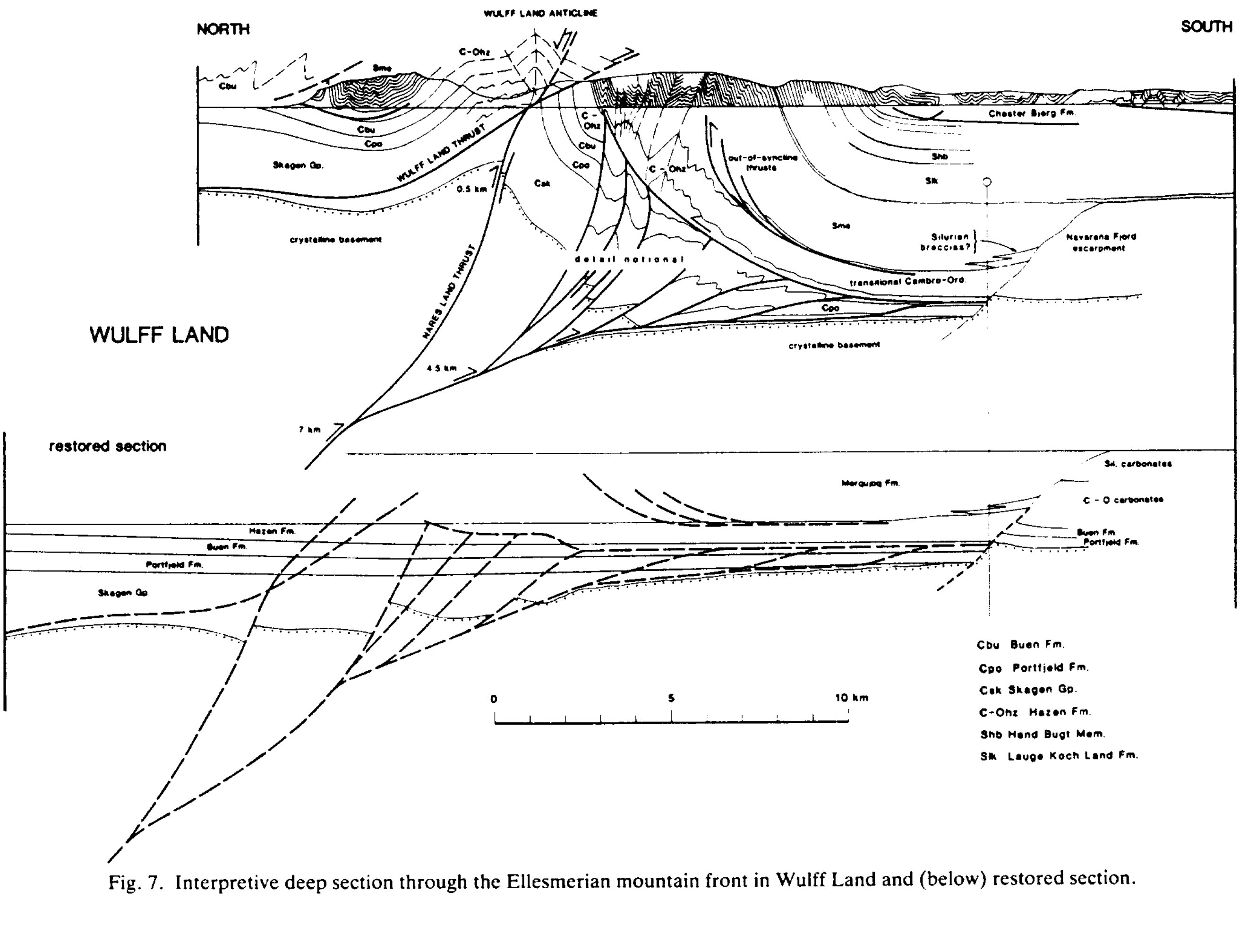

Fig. 7. Interpretive deep section through the Ellesmerian mountain front in Wulff Land and (below) restored section.

This thrust post-dates the basement thrusting which produced the monocline, as does the WLT, so it seems clear that the monocline is the earliest Ellesmerian structure in the region, and all the thin-skinned, south vergent structures post-date it.

We now return to the eastern, thin-skinned part of the Ellesmerian thrust front, to show that the model developed for the monocline can help to resolve the remaining problems posed by the Navarana Fjord and Adolf Jensen Fjord sections.

Navarana Fjord (Fig. 8)

This section illustrates the problem alluded to above—the 2 km elevation associated with the synclinal up-turn near the southern end of the section. If the up-turn is interpreted as an incipient development of the main mountain front monocline, it is possible to similarly reconstruct the deep structure in terms of a reactivated basement ramp. The partial restoration (Fig. 8) shows that the required displacements are modest: about 1 km on the upthrust and 3.75 km movement of its branch point up the ramp.

One of the models rejected during development of the basic Nares Land model was that the basement ramp and the deep-seated fault beneath the Navarana Fjord escarpment were the same structure. The Navarana Fjord section shows that this cannot be so; the two structures are separated by a 10 km tract occupied by the broad synclinal up-turn.

The Navarana Anticline itself can be interpreted as a large box-fold associated with loss of displacement on a thrust at the top of the basement below thick Skagen Group; by analogy with the Wulff Land section, this thrust may post-date the up-turn. Farther north, tight S-vergent folds bring up Hazen and Buen Formations. Again by analogy with Wulff Land, these folds may be associated with a continuation of the Buen Thrust; a comparable thrust has been mapped in Freuchen Land just west of the entrance to Navarana Fjord (Fig. 4).

Adolf Jensen Fjord (Fig. 9)

Along this section the southward tilt to the whole structure mentioned above comprises about 1 km elevation at the main up-turn and a further 1 km over a distance of 20 km to the north of that point. By analogy with the more westerly sections this could have been generated by a modest displacement, 4 km or so, on a gently inclined section of the basement ramp. The backthrust connection in Buen shales between the ramp and main detachment shown in Fig. 9 is speculative.

DISCUSSION

Geometry of the basement ramp

The position and shape of the uppermost part of the basement ramp is quite tightly constrained by area and line balance considerations in the five sections presented above. Some indication of the deeper trajectory can be gained from a study of regional elevations north of the main monocline, using well-known principles which, for example, link a uniformly inclined ramp to uniform uplift, a convex-up ramp to a forward tilt and a concave-up ramp to a backward tilt. Many methods of construction are available, just as there are for determining extensional fault trajectories from roll-over geometry, and all are artificial in that they involve particular displacement criteria. The method used to construct Fig. 10, a combination of area and horizontal line balance, with constant displacement, does model the inferred basement elevation reasonably well, and leads to an interesting insight into the origin of the monocline.

In Fig. 10, the top of the Skagen Group (base of the Portfjeld Formation) between the ramp and the Navarana escarpment was chosen as the regional reference level, and the elevation of either the basement (Figs. 10b & c) or of the Skagen Group itself (Figs. 10a & d) related to it. From the elevation above regional of an arbitrary point P at the north end of the section, and the displacement on the ramp, point P′ and the horizontal displacement P′Q can be determined. Angle PP′Q gives the ramp inclination beneath P′ and its depth can be found from an excess area calculation relating areas A and A′. It proved possible to generate the required shape of the elevated tract by supposing the basement wedge to consist of horizontal layers and applying a 'flexural-slip' construction, maintaining layer thickness and length. The basement did not of course deform in this way, but the ease with which it proved possible to model the forward and backward tilts and the folds shown in the Navarana and Wulff Land sections lends credibility to the inferences drawn.

Controls on development of the monocline

The main conclusion to be drawn from Fig. 10 is that the westward increase in amplitude of the monocline is due to two factors, increasing dip of the ramp and increasing displacement on it. The actual location of the monocline was of course controlled by the basement ramp, on which displacement was concentrated as the sole thrust locked at the Navarana escarpment.

It is not possible to model the deep trajectory of the ramp any farther north than shown in Fig. 11, but it may be presumed that it reverts to a constant dip, commensurate with displacement, to maintain Lower Cambrian rocks at the surface throughout Nansen Land. This may perhaps be seen in the Navarana Fjord section (Fig. 10b).

Implications for Nares Strait

The amount of early Tertiary displacement on the Nares Strait, as Greenland rotated away from North America, continues to be controversial (see the symposium volume edited by Dawes & Kerr 1982). An analysis of the problem in the light of new work on

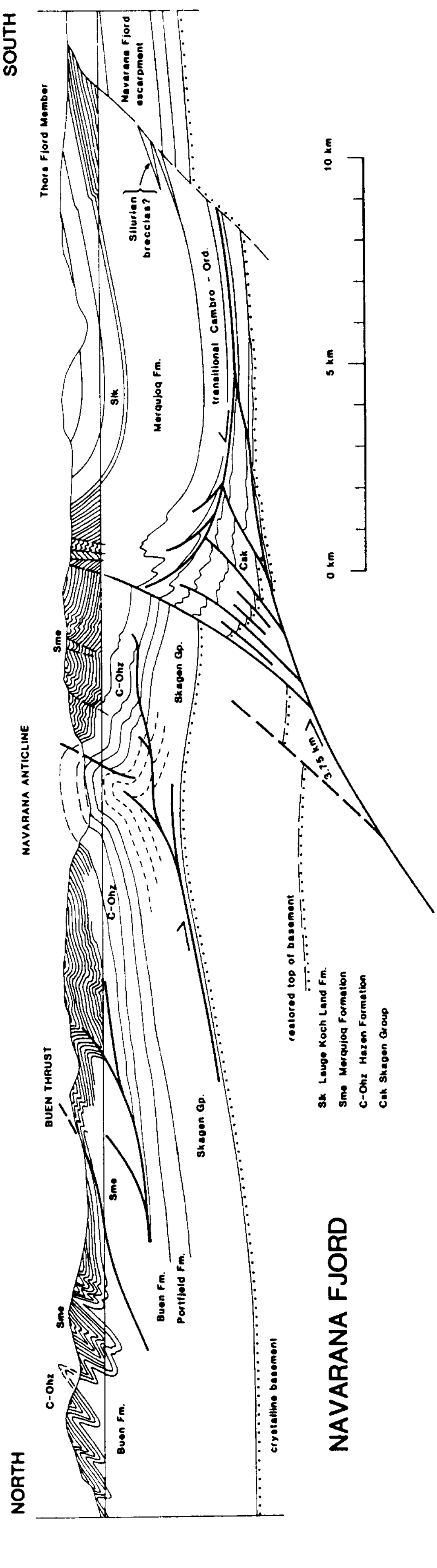

Fig. 8. Interpretive deep section along Navarana Fjord.

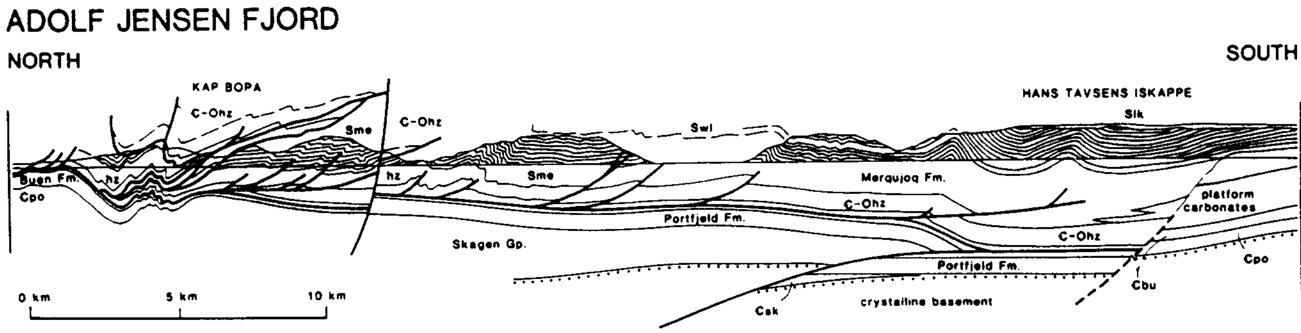

Fig. 9. Interpretive deep section along Adolf Jensen Fjord, modified after Soper & Higgins 1985.

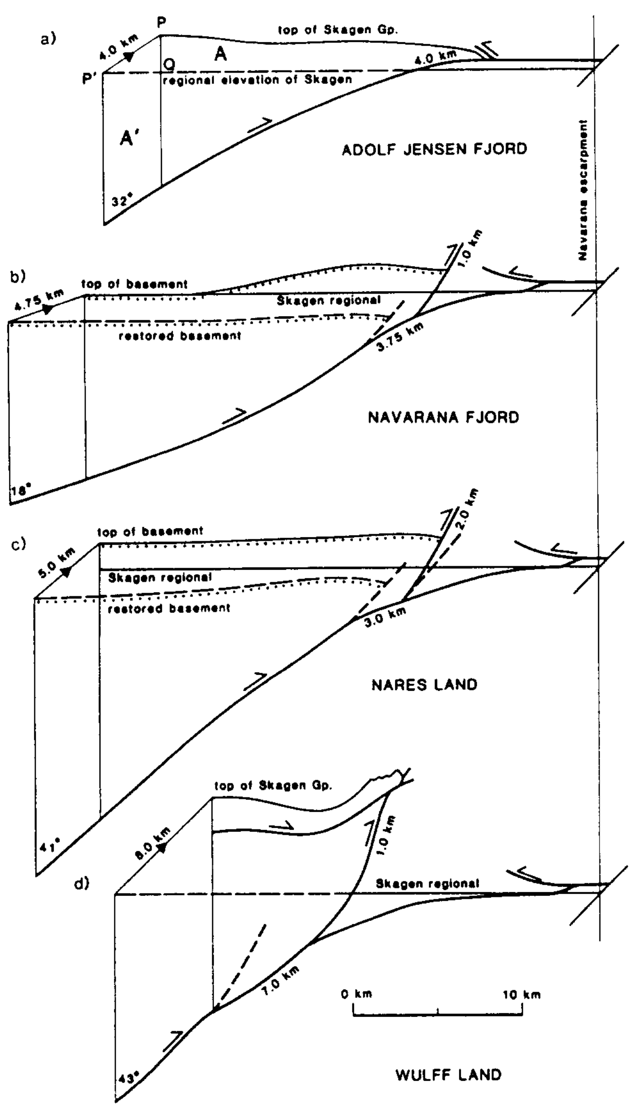

Fig. 10. Reconstruction of the ramp trajectory at depth along the four N–S cross-sections. For explanation see text.

Ellesmerian structures on both sides of the strait is in preparation with Canadian colleagues, and very brief mention will be made here. The main problem is that Palaeozoic structural trends curve parallel to the strait and facies boundaries cross it at a small angle (Christie *et al.* 1981, Higgins *et al.* 1982), so that offset depends on how the lines are interpolated across the seaway (Dawes & Kerr 1982).

In Greenland the monocline passes offshore NW of Nyeboe Land, but the synclinal up-turn (Nyeboe Land Syncline—NLS) curves to the SW sub-parallel to Nares Strait (Figs. 4 and 11), just cutting the coast of Hall Land, the next peninsula to the west (Dawes 1987). Across the strait in Ellesmere Island, on Judge Daly peninsula, the monocline is not developed and Ellesmerian folds are concentric in style and almost parallel to the strait (Trettin & Balkwill 1979). This rapid along-strike termination of the monocline implies the existence of an Ellesmerian accommodation structure along Nares Strait.

Our analysis suggests that the strait was the site of an important transfer fault or zone during the early Palaeozoic at the margin of the Franklinian Basin, which terminated the basement ramp and shifted the early Cambrian basin margin farther south on the Canadian side. We suppose that this deep fracture system was utilized as a tear fault during the Ellesmerian orogeny, confining westward propagation of the monocline, and that it suffered limited sinistral reactivation during the early Tertiary rotation of Greenland away from North America. If correct, this implies that the curvature of Palaeozoic lineaments in the vicinity of the strait is an original feature, and that Tertiary displacement is no more than a few tens of km. The most prominent facies marker which crosses the strait, the Ordovician platform margin (marked by the Navarana escarpment on the Greenland side), shows a combined displacement due to Ellesmerian shortening and Cenozoic strike-slip of about 25–40 km dependent on the line of projection west of Nyeboe Land (Fig. 11).

SUMMARY

The Ellesmerian front in North Greenland is characterized by E–W-trending thin-skinned structures in the east, but west of Adolf Jensen Fjord changes into a mountain front monocline which attains an amplitude of almost 7 km in Wulff Land.

It is not possible to interpret the monocline as entirely thin-skinned; crystalline basement must be involved.

The deep structure of the monocline can be modelled as an early Palaeozoic extensional basement ramp which was reactivated during the Ellesmerian orogeny. The

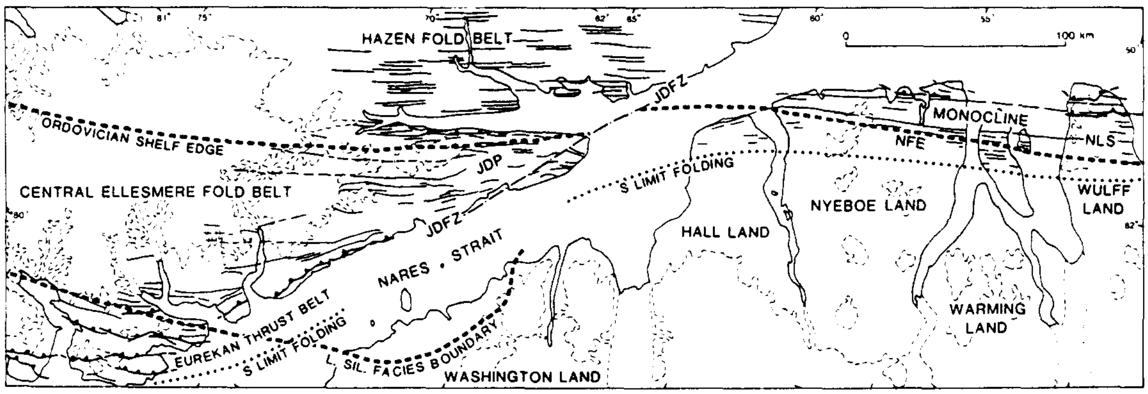

Fig. 11. Map of the Nares Strait region showing some early Palaeozoic facies boundaries and Ellesmerian structural trends. The Lower Silurian facies boundary is a steep contact between shelf carbonates to the south and shales to the north (Christie *et al.* 1981). Other markers are discussed in the text. JDP, Judge Daly Promontory; JDFZ, Judge Daly fault zone; NFE, Navarana Fjord Escarpment.

resulting thrust is thought to have terminated at a tip-line located at another early extensional fault beneath the Navarana escarpment, which was too steep to be reactivated, and a backthrust developed, producing a triangle zone at depth and the monocline above it.

The model has been generated by iterative balancing and is internally consistent, but geophysical evidence of the depth to basement is needed to test it.

After the structure locked, shortening continued by the development of thin-skinned thrusts on the upper limb of the monocline. The effect was to displace Lower Cambrian trough sediments southwards over the shelf margin and slope sequences, and to elevate them to the present erosion level.

The scale of the monocline appears to have been controlled by the inclination of the basement ramp and the displacement on it which reached a maximum of about 7 km in Wulff Land.

The abrupt reversion to a thin-skinned deformation style in Ellesmere Island implies the existence of a transfer fault in the vicinity of Nares Strait. This may have suffered limited strike-slip reactivation in Tertiary time.

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