Intra-continental rifting inferred from the major late Carboniferous quartz-dolerite dyke swarm of NW Europe

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Synopsis

The late Carboniferous (Stephanian) quartz-dolerite dyke swarm of northern Britain extends eastwards from the Outer Hebrides on an arcuate trend for up to 200 km from the eastern UK coast, as far as the western margin of the Central Graben. The average trend of the swarm changes by about 45° between its western and eastern extremities. Individual dykes, which are generally up to 30 m wide onshore, attain widths of well over 1 km offshore. Magmatically and spatially, the swarm is closely related to the coeval Oslo Graben volcanic rocks and dykes of southern Norway, which lie on the extrapolation of the arcuate trend. The dyke trends point to a focus of relative tensile stress centred on the West Shetland Shelf region. Finite element modelling of the NW European area corroborates the stress-focusing hypothesis, which we interpret in the context of the major period of intra-continental rifting along the line of the proto North Atlantic (the Rockall Trough, Faeroe–Shetland Trough and eastern Norwegian Sea) at about 300 Ma. The regional distribution of Archaean and Caledonian lithosphere within Pangaea may account for the fact that both the dyke swarm and the Oslo Graben were located in northern Britain and southern Norway, rather than in the Faeroe–Shetland region.

The late Carboniferous quartz-dolerite dyke swarm of northern Britain has hitherto been regarded in global terms as a relatively minor swarm. However, compilation of relevant geological and geophysical data from around Britain demonstrates that the swarm is, in fact, one of the major dyke swarms of NW Europe (Smythe 1994). Individual dykes or dyke concentrations (i.e. multiple dykes) are inferred by geophysical modelling to be over 1 km wide; these may be the widest dykes discovered to date. Figure 1 summarizes the compilation. The Dunbar dyke group, for example (D in Fig. 1), has a cumulative width of 2–3 km across strike (NNW–SSE) over a distance

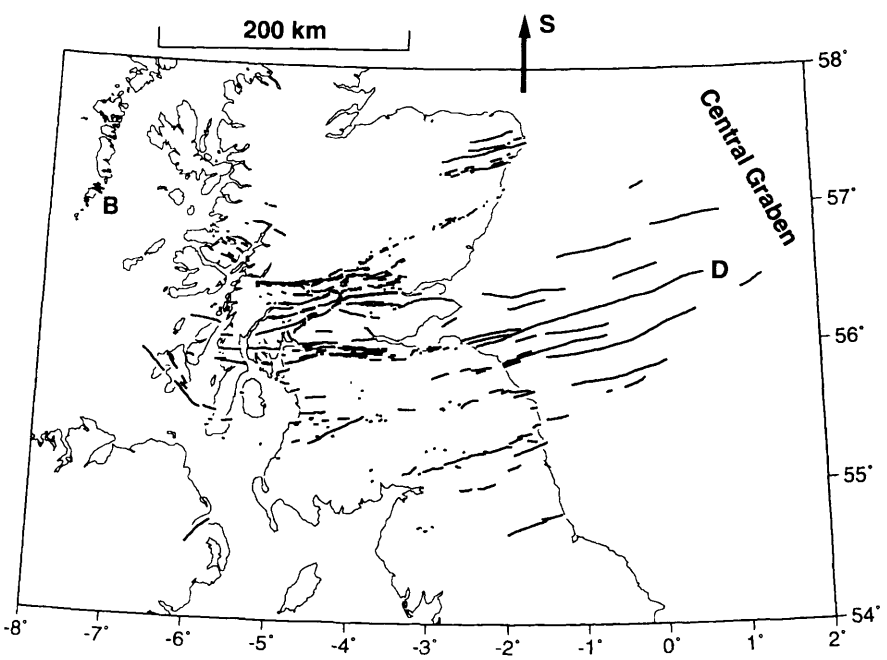

FIG. 1. Stephanian (300 Ma) quartz-dolerite dykes of northern Britain, after Smythe (1994). B, Barra (Fig. 2a); S, towards Shetland (Fig. 2b); D, Dunbar dyke anomalies. UTM projection, central meridian 3°W.

of 200 km, which makes them (or it) as large, volumetrically, as any intra-continental dyke yet modelled. In comparison the Great Abitibi Dyke on the Canadian Shield is 600 km long, but only 250 m wide (Ernst *et al.* 1987). The Great Dike of Nova Scotia (also known as the Shelburne Dyke) is up to 200 km long and 60–180 m wide (Papezik and Barr 1981). Gravity modelling of the Great Dyke of Zimbabwe shows that it is bell-shaped, with the main feeder about 1 km wide (Podmore and Wilson 1987). This dyke is 550 km long.

An intra-continental dyke swarm is an excellent record of regional crustal strain. Firstly, its trend is little affected by local inhomogeneities in the upper crust; secondly, most dyke rock is suitable for radiometric dating; and lastly, a swarm, once emplaced, may subsequently be buried and deformed, but it is practically impossible for the swarm to be eroded away. We can use a swarm to infer contemporary stress fields. Even the occasional dyke swarm without a preferred orientation can tell us something useful, viz. that the difference in the two horizontal principal stresses must have been small (Tokarski 1990).

Precambrian dyke swarms (Halls 1982) imply the existence of a rather simple world-wide stress system (Morris and Tanczyk 1978), whereas Phanerozoic dolerite dyke swarms at passive margins were recognized by Scrutton (1973) to be of a similar age to that of the adjacent lithospheric plate break-up. Fahrig (1987), considering mainly the Precambrian mafic swarms of the Canadian Shield, classifies swarms as either 'early passive margin' or 'failed arm', and therefore links both types to plate break-up processes. The usefulness of a dyke, or better, a dyke swarm, thus lies in its plate tectonic implications.

Many examples documented in the volumes edited by Halls and Fahrig (1987) and Parker *et al.* (1990) demonstrate that major dolerite dyke intrusion can give a good estimate for the age of initiation of a phase of intra-continental rifting prior to sea-floor spreading. The main caveat is that in some examples the dykes signify the start of an abortive rift phase, as in the Labrador Sea and South Australia, and that continental break-up may only occur after a subsequent phase of rifting. On the other hand, there are no well-documented examples of major swarms which are *not* connected with a rift event.

The main aim of this paper is to show that the late Carboniferous quartz-dolerite dyke swarm in northern Britain is a time-marker of a major late Palaeozoic rifting episode in the North Atlantic region, by analogy with dolerite dyke swarms at other passive margins. The Hercynian Orogeny affecting the south of Britain at around the same time is of no direct relevance. We restrict ourselves to consideration of the tholeiitic basalts because they define a single geological event, from which we aim to infer lithospheric stresses; however the olivine dolerites and other more alkaline swarms in similar localities, and with similar trends and slightly younger ages, are probably related to the same general stress regime.

Distribution of dykes and related igneous rocks

Northern Britain

Figure 1 shows a compilation map of all the mapped and inferred dykes. The late Carboniferous quartz-dolerite dykes are sparsely developed in the west of Britain, where they trend at 110° (all azimuths are quoted clockwise from true north). They have not been traced beyond longitude 7° 30′W. The most westerly dykes occur in the Outer Hebrides (Fig. 2a; B in Fig. 1). The dykes trend at about 90° in the Midland Valley of Scotland, where they are generally continuous, and of the order of 30 m wide (Walker 1935; Richey 1939; Cameron and Stephenson 1985). The swarm is regionally arcuate, because the dyke trend swings around to about 70° along the eastern seaboard of Britain, and to about 65° offshore to the east. The arc therefore spans a range of azimuths of about 45°.

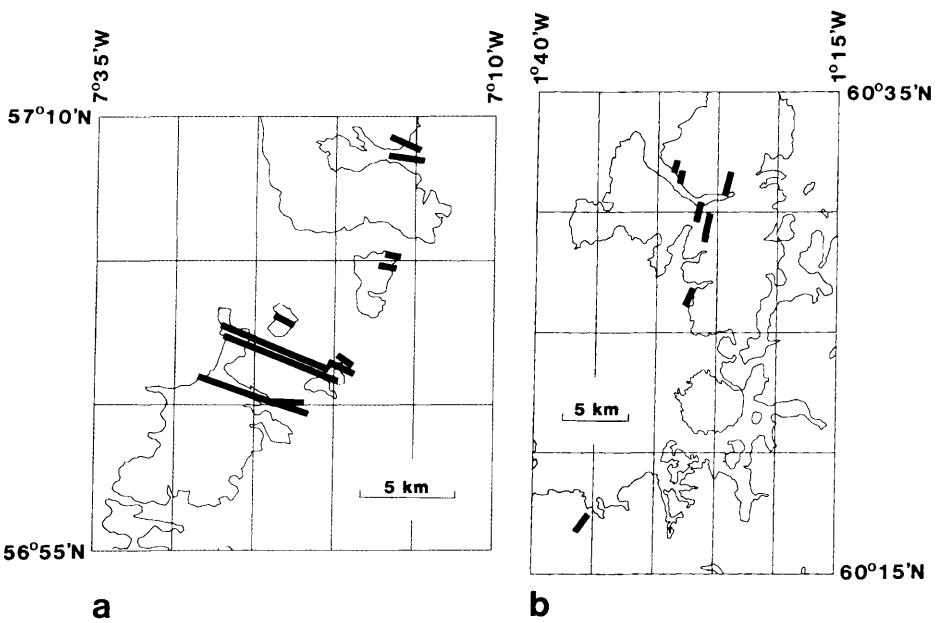

FIG. 2. (a) Detail of WNW–ESE trending dykes of Barra and South Uist. UTM projection, central meridian 3°W. (b). Detail of NNE–SSW-trending dykes in west Shetland. UTM projection, central meridian 3°W.

Local deviations from the regional trend occur in the vicinity of the Great Glen and Highland Boundary faults. The few contemporaneous quartz-dolerite dykes in Shetland, 500 km to the north (Fig. 2b) trend at about 10–45°, sub-parallel to the present continental margin. Although they do not belong geographically to the main arcuate swarm, the stress analysis presented below shows that they could have been emplaced at about the same time.

The age of the quartz-dolerite intrusions is 301 ± 6 Ma, as inferred from a K/Ar study of the Whin Sill (Fitch and Miller 1967; De Souza 1979; Fitch *et al.* 1970). Two of the dykes on the Outer Hebrides (Fig. 2a) give Rb/Sr ages of 284 ± 4 and 300 ± 4 Ma, respectively (Fettes *et al.* 1992). Chemically they are high in Fe and Ti (Macdonald *et al.* 1981), and are analogous to the FETI-basalts of Brooks and Jakobsson (1974) which often occur at spreading centres and over hot spots. They are also chemically comparable to the tholeiitic members of the basic lava suite of the coeval Oslo Graben (Weigand 1975; Macdonald *et al.* 1981).

Scandinavia

Although there is no evidence for the quartz-dolerite dyke swarm extending further eastwards than the western margin of the Central Graben, other relevant features are picked up along the anticlockwise extrapolation of the arcuate trend (Fig. 3a). These include the Oslo Graben, with its gravity high extending SW into the Skaggerak (Ramberg and Smithson 1975), and the distribution of some Permo-Carboniferous igneous intrusive rocks. In the Oslo Graben itself there was a major, but short lived, period of basic volcanism (Oftedahl 1978), which has been dated at 293 ± 10 Ma (Rb/Sr: Sundvoll 1978); this is statistically indistinguishable from the age of the British quartz-dolerite suite. The bulk of the extrusive rocks are alkali basalts and porphyritic trachyandesites, and in the south of the Graben they make a pile up to 1.5 km thick (Segalstad 1975). Northwards the pile thins progressively, until near the centre in the Krokskogen area it is only 30 m thick, but here Weigand (1975) has shown that it includes tholeiitic basalts closely matching the quartz-dolerites in Britain (Macdonald *et al.* 1981). There is a relatively small dolerite dyke swarm of the same age (K/Ar: 300 ± 4 Ma; Klingspor 1976) developed in Scania, southern Sweden (Fig. 3b), which does not appear to continue offshore to the NW (J. M. Bergstrom, pers. com. 1981), nor is it to be found on the island of Bornholm, 40 km to the SE.

Context of emplacement

The dykes, together with related volcanism in southern Scandinavia, have been interpreted previously as a late event in the Hercynian Orogeny (Hjelmqvist 1939; Anderson 1951; Francis 1978*a*, *b*). The east–west average trend of the swarm across northern Britain is far from parallel to the north–south and NE–SW trends of the passive continental margin bounding NW Europe, so to relate the dykes to rifting to the NW of Britain, rather than to the east–west trending Hercynian collision to the south, seems at first glance to be rather implausible. This problem can be resolved. Firstly, we show by graphical methods that the dyke swarm has an arcuate trend, the approximate centre of which is in the West Shetland Shelf area. Secondly, we test by quantitative stress analysis its relationship (along with the Oslo Graben) to our postulated model of intra-continental rifting in the North Atlantic. Lastly we discuss what geological inferences can be drawn from the analysis.

The arcuate pattern of the swarm

Methods used

The dykes and dyke anomalies were compiled in the first instance onto 1:250 000 scale UTM plot sheets, then re-drawn on 1:1 000 000 scale UTM bases and digitized. This yields a file of about 8000 points, the data being stored as latitude, longitude pairs. Each dyke is represented by a series of points which are joined together to make segments for plotting. Each dyke segment represents a section of dyke of the order of a kilometre in length. No attempt has been made to regularise the sampling increment, nor to smooth the series of points describing each dyke.

Assuming that the dyke segment defined by a pair of latitude–longitude coordinates is a segment of a great circle on the sphere of the Earth, the normal great circle through the mid-point of the dyke segment is computed. Our program uses an elegant formulation by Stuart (1984), in which spherical coordinates are transformed by stereographic projection onto the complex number plane. Each dyke segment normal is then represented digitally by a series of geographical coordinate pairs at a regular spacing, at, say 5°, either side of the dyke segment mid-point. This file of dyke normals can then be considered as a representation of the extensional stress trajectory along each dyke segment. The 8000 or so normals thus available create a far too cluttered map, so the file is winnowed for display. At the plotting stage, a polygon made up of connected great circle segments can be defined, to one side of which all dykes are deleted. This is used to remove heading and trailing dykes (see next section).

The angular error in dyke segment trends (and therefore also of their corresponding normals) is in the order of 1° due to rounding error in storage of the coordinates. Digitizer 'jitter' probably leads in addition to rather greater errors, but we surmise that it will average out over a long dyke.

Dyke trailing and heading

Anderson (1951, p. 30) described the phenomenon of faults being deflected by a pre-existing fault. This is called trailing, in the sense of following a trail. The same effect is exerted by large faults on dykes, even though, as stated above, the regional trend of dyke swarms tends to be oblivious to local geological discontinuities. In Scotland, the late Carboniferous quartz-dolerite swarm studied here is affected by the Great Glen and Highland Boundary faults. As predicted by theory, these large faults act as linear

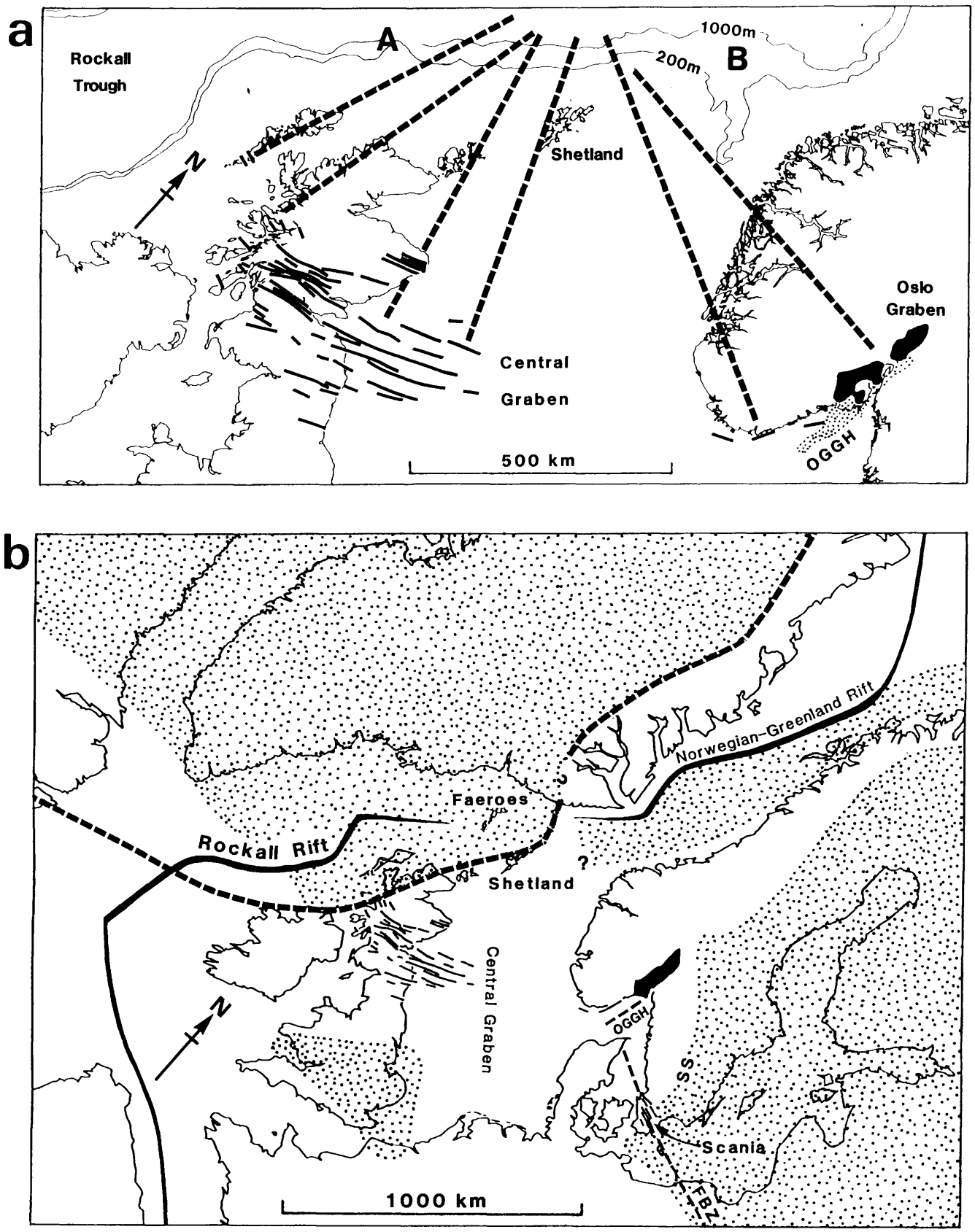

FIG. 3. (a) Late Carboniferous dykes of northern Britain and the North Sea region (solid lines), together with the basalts of the Oslo Graben (black areas) and its SW continuation delineated by the Oslo Graben Gravity High (OGGH; stippled area). Extensional stress trajectories (dashed lines) are sketched normal to the dykes and the Oslo Graben. A–B shows the approximate location of the line of nodes AB of Figure 7. (b) 300 Ma continental reconstruction. The proto-North Atlantic Rift is incomplete in the Faeroe area. Note that the Scania dykes, which do not fit in to the pattern of radial stress trajectories (a), are confined to the intersection of two major crustal lineaments, the Fennoscandian Border Zone (FBZ) and the Småland Suture (SS). Stipple indicates pre-Grenville crust and lithosphere, speculatively interpolated in the Faeroe–Shetland region. Dashed line is the western margin of the Caledonian orogen. Both maps are oblique Mercator projections.

zones across which shear stress is poorly transmitted. The principal stress directions deviate in the vicinity of the fault so that they become orthogonal to the fault plane. Dykes, presumably following the local trend of the maximum principal compressive stress, will then either trail along the fault, or will head into the fault at right-angles. The latter case cannot be described as trailing, although the mechanism is essentially the same as trailing. We propose the analogous term 'heading' to describe this.

Figure 4 shows the trailing and heading dykes. In the case of the Great Glen Fault, dykes of the quartz-dolerite swarm swing round from a regional trend of 090–100°, to head into the fault at about 120°. The Highland Boundary Fault, in contrast, causes the roughly east–west-trending dykes on its NW flank to trail into a trend nearly parallel to the fault. It is interesting to note that the dykes appear to be unaffected south of 56°N, where this important crustal fracture is represented by a set of splays at current erosion levels rather than by a single fault (Paterson *et al.* 1990).

The third big fault in Scotland, the Southern Upland Fault (Floyd 1994), appears to have no effect at all on the swarm, which crosses it at an angle of about 30°. This suggests that the fault is not a crust-penetrating fracture, unlike the Great Glen and Highland Boundary faults.

Results

The dykes deviated by the Great Glen and Highland Boundary faults, shown in Figure 4, have been removed from the digital file used for the regional trend study. The regional map of the great circle normals to the remaining dykes shows a concentration, or focusing, of the normals in the region of northern Scotland, north of the dyke swarm (Fig. 5). A swarm without any preferred curvature would produce a set of sub-parallel normals which would have their greatest concentration in the area of the swarm itself. The actual focus is 400–500 km north of the outcrop of the swarm, and is therefore not an artefact. The focus of the normals is poorly defined, but on the West Shetland Shelf in the area of 60 ± 1°N, 4 ± 2°W. Note that the solitary normal displayed in Figure 5 trending WNW through Shetland suggests that the Shetland quartz-dolerites (Fig. 2b) do indeed belong to the main northern British swarm.

No attempt has been made to include the dykes, faults, or other possible trends, of the contemporaneous Oslo Graben tholeiites in the statistical study. The Graben has a regional NNE trend, but within it there are more local north- or NNW-trending faults and dykes. Nevertheless, it is clear that the Graben does lie approximately on the eastward (anticlockwise) extrapolation of the arcuate swarm of northern Britain (Fig. 3).

The basic alkaline dyke swarms of the Scottish Highlands

We have omitted the Scottish Permo-Carboniferous swarms of basic alkaline dykes (camptonites and monchiquites) from the study. These occur mostly NW of the Great Glen Fault (Rock 1983), and are probably also common in Ayrshire. Volumetrically, the suite is insignificant compared with the tholeiitic suite, because the alkaline dykes are only a metre or so thick on average. Rock pointed out the petrological similarity of these dykes, and the associated volcanic necks and plugs, to the alkaline intrusive phase of volcanism of the Oslo Graben. The suite

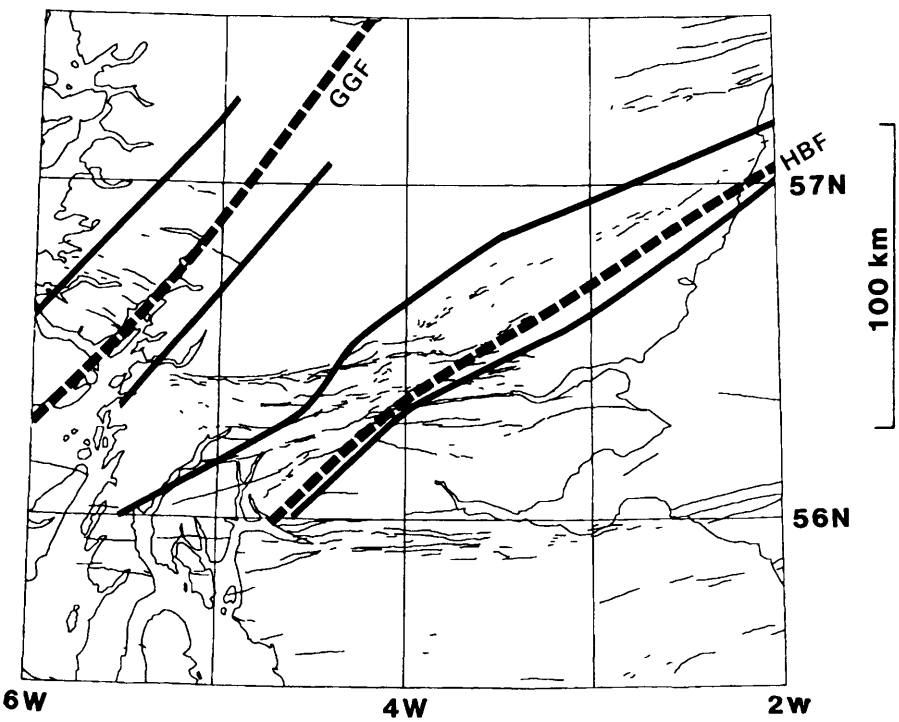

FIG. 4. Dykes deviate from the regional arcuate trend, either heading into the Great Glen Fault (GGF) or trailing along the Highland Boundary Fault (HBF). Solid lines flanking each fault define the limits within which the deviated dykes have been deleted from the digital file for regional trend study. UTM projection, central meridian 4°W.

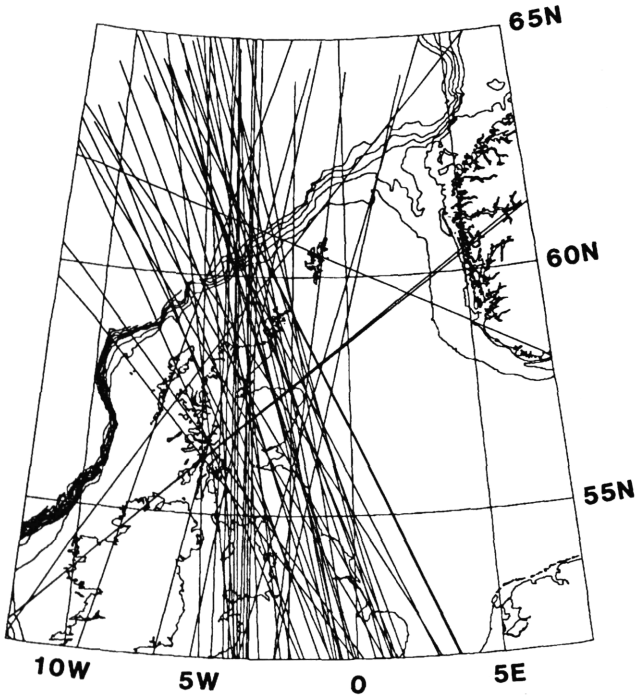

FIG. 5. Normals to the quartz-dolerite dykes, indicating the local relative extensional stress direction, focused on the West Shetland Shelf region, around 59–61°N, 2–6°W. Each normal is a great circle through the mid-point of a digitised linear dyke segment which is also treated as a great circle. Dykes (not shown) are made up of digitised line segments each about 1 km long. The digital file has been winnowed to avoid excessive clutter until only 2% of the great circles are displayed. The NW European continental margin is indicated by bathymetric contours at 200 m intervals. The Faeroe Islands and other microcontinental fragments are omitted.

does not represent one episode of extension, however, as the ages of individual swarms vary from about 330 Ma to 240 Ma (Baxter and Mitchell 1984; Esang and Piper 1984). Figure 6 shows rose diagrams for the various swarms (after Rock 1983) with ages in Ma added where known (Baxter and Mitchell 1984).

The dyke trends in Figure 6 follow a similar arcuate pattern to that inferred for the quartz-dolerites. How close are they in time to the quartz-dolerites? We have modified Baxter and Mitchell's conclusions regarding the ages of some of the swarms. Firstly, the Orkney swarm may well consist of two episodes, the older of which is 288 Ma. Secondly, the east–west Eil–Arkaig swarm, for which Baxter and Mitchell prefer a 326 ± 8 Ma age based on eight samples, may also comprise a younger group of dykes. Two anomalous dykes, omitted by Baxter and Mitchell because they appeared to be 50 Ma younger than the 326 Ma grouping, when added to a further two dated camptonites from the same swarm (Fettes *et al.* 1992), together suggest a consistent age of around 285 Ma. It is therefore possible that one phase of the long-lived basic alkaline dyke suite resulted from the same extensional stress pattern as that which gave rise to the quartz-dolerites, but was intruded about 10–20 Ma later. The implications of this are discussed below.

Quantifying the stress field inferred from the dykes

Regional geological constraints

May (1971) argued, in his analysis of the stresses responsible for rifting of the Central Atlantic, that as the early Jurassic dyke trends were independent of metamorphic and tectonic grain, it was reasonable to draw extensional stress trajectories normal to the dykes. Following May's example, we have sketched extensional stress trajectories normal to the late Carboniferous dykes, to the Oslo Graben and to its gravity high. These sketched trajectories converge on the continental margin west of Shetland (Fig. 3a), and the graphical analysis of normals to the northern British dykes alone (excluding Scandinavian data) supports this convergence (Fig. 5). We previously proposed (Russell and Smythe 1983) an explanation for this concentration of stress, comprising two main elements.

Firstly, the supercontinent of Pangaea, which was just about assembled at 300 Ma, had a neck, or narrow zone, in the Europe–Greenland region, This is the area shown in Figure 3b. Any rifting of Pangaea is likely to have been concentrated in this region, due to the relatively short distance from the Boreal ocean (just beyond the top right-hand corner of Figure 3b) to the Tethys (bottom left of the

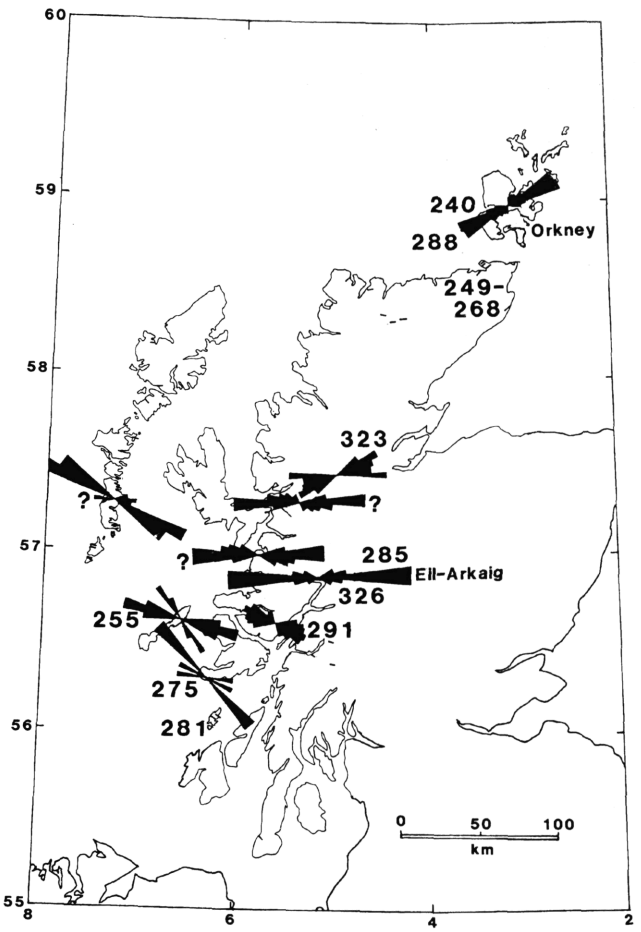

FIG. 6. Rose diagrams of trends of the various swarms of the late Palaeozoic basic alkaline suite of dykes, after Rock (1983). Radiometric age determinations are shown in Ma, where known (modified after Baxter and Mitchell 1984).

same figure). Rifting along what later became the Rockall Trough and the Norwegian Sea, is now generally accepted to have occurred at this time (see, for example, Ziegler 1982, 1988 and Doré 1992; also the discussion in Smythe 1989). This implies a regional tensional stress field oriented NW–SE over the area of Figure 3. A likely way in which the rifting would have occurred could have been by fracture propagation into this continental neck from both oceanic regions simultaneously. The last place to rift would then be the Faeroe–Shetland area (Fig. 3b), implying a concentration of NW–SE-oriented tensile stress trajectories through this area.

Secondly, lithospheric rifting may have been more difficult to achieve between Scotland and Greenland in the late Carboniferous, because the pre-Grenville Laurentian Shield was continuous across this region (Fig. 3b). The thermal lithosphere here would then have been extremely thick (perhaps 300 km or so; Pollack and Chapman 1977) and therefore strong. The western margin of the Caledonian orogen, bounding the relatively young, thin lithosphere, must have taken a sharp bend between Shetland and east Greenland (Fig. 3b; see also Brewer and Smythe 1984), even if we allow for differing levels of erosion at 300 Ma and now. Nevertheless, the rifting west of Britain and Ireland did eventually take a path through what we believe to be pre-Grenville lithosphere; the problem in testing or improving upon the lithospheric strength conjecture is that the relevant basement now lies offshore and is deeply buried.

Finite element models

Next we test the hypothesis that the arcuate trend of the late Carboniferous dyke swarm is due to a concentration of regional horizontal tensile stresses through the Faeroe–Shetland area. A quantitative way to do this is using two dimensional finite element modelling. Many simplifying assumptions must be made, but where possible we can justify these in terms of our inferred global plate tectonic setting for Permo-Carboniferous times, taking into account also the relative importance of various proposed plate tectonic driving forces (e.g. Forsyth and Uyeda 1975; Chapple and Tullis 1977; Richardson *et al.* 1979; Backus *et al.* 1981).

The NW European plate model

The upper, north-western edge of the model (Fig. 7) is based on the shape of the continent–ocean boundary, on a transverse Mercator projection (central meridian 9°W) of its present day position. The other three sides of the model (left, right and bottom in the orientation of Fig. 7) are arbitrarily defined at a large distance from the northern Britain – southern Norway area. The lithosphere is assumed to behave like a flat, isotropic, elastic membrane in plane stress. A Poisson ratio of 0.25 is assumed. The values of Young's modulus and the model thickness used are not important, since absolute stress magnitudes are not considered, and the model is assumed to be laterally homogeneous.

The 2D plane model, rather than a spherical cap plate model, is justifiable since we are only interested in calculating stresses within a small region of about $10^3 \times 10^3$ km^2, where the elements are most concentrated (Fig. 7). All elements are triangular with three Gauss points. Our simple elastic model, with no discontinuities, does not require any more sophistication. Some error in stress computation will arise because we are considering the elements to be coplanar, rather than to tessellate a sphere. However, the angular difference between pairs of elements in these two models is on the order of 1°, which is negligibly small, so there will be no significant error in stress tensors computed using the plane model rather than the spherical model. There is an analogy here with interpreting maps; as long as we have a sensible map projection for a region (Maling 1992) — Europe for example — the geologist does not

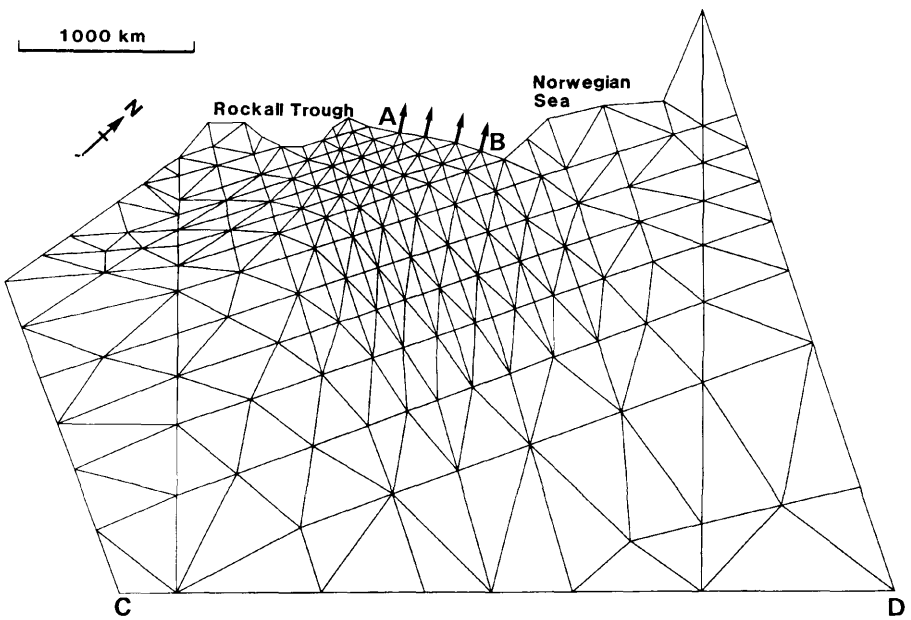

FIG. 7. Finite-element grid consisting of 374 elements and 211 nodes, covering NW Europe. The upper edge of the model represents the NW European plate margin. Lower edge C–D in SE Europe, but for the purposes of the modelling effectively at infinity, is fixed. Arrows on the four nodes along the line A–B (Fig. 3a) represent the extensional force imposed on the edge of the plate in the area of the Faeroes by the Laurentian sub-plate during the phase of rifting shown in Figure 3b. The stresses calculated from this model are shown in Figure 8.

hesitate to compare trends of features on one part of the map with another.

Effects of pre-existing important features such as the Great Glen Fault and the Fennoscandian Border Zone have been ignored for the present. The model is assumed to be in static equilibrium, and all stresses in the model are assumed to arise from imposed boundary forces. Other possible sources of horizontal stress in the lithosphere due to thermal changes, south to north movement on an ellipsoidal earth, surface loading changes (e.g. erosion), stress due to isostatically compensated topography (e.g. young continental margins) and residual tectonic stresses are neglected.

Boundary conditions

The pertinent stress field, as inferred from the Carboniferous geology of Britain, Norway and Greenland, is clearly one of lithospheric extension, so the primary forces on our model should be tensional stresses, whatever their source. What about second-order stresses? The mechanical interaction between the lithosphere and asthenosphere is generally considered to resist plate movement, especially under the continents (Chapple and Tullis 1977; Backus *et al.* 1981). Secondary-scale convection below the lithosphere may contribute significantly to the state of lithospheric stress, but since this effect is poorly understood even in present-day tectonics, we are forced to ignore it in our modelling. We have also assumed that the elastic strength of the asthenosphere is negligible, and that stress diffusion effects occur over time periods too short to be considered in the modelling. All interactions between the lithosphere and asthenosphere have therefore been neglected.

Earthquake source mechanisms show that the East African Rift is an extensional feature, with the predominant trend of the least compressive stress directions perpendicular to the rift. However, similar focal mechanism studies in ocean basins reveal that the state of stress is mostly compressional, resulting from 'ridge push' (Richardson *et al.* 1979). At the time of the formation of the dyke swarm we envisage that major intra-continental rifting was under way in the Rockall Trough and the eastern Norwegian Sea. Therefore we do not apply either compression or tension along these parts of the model boundary, because mature sea floor spreading, complete with elevated mid-ocean ridge, was not occurring; but on the other hand continental rifting would have reached a stage where any regional extension would be quickly accommodated by dilatation and dyke intrusion.

Since all our models are fundamentally extensional perpendicular to the rift (the upper edge of Fig. 7), the parts of the model which bound the Rockall Trough and the eastern Norwegian Sea are left as free boundaries. Only the segment of this boundary in contact with the Faeroe Plateau area is considered as being able to transmit tensional forces between Greenland–North America and NW Europe (A–B in Fig. 7). We presume that the origin of this tension was the attempted relative movement of Greenland–North America away from NW Europe (Russell and Smythe 1983) during the late Carboniferous.

Results

By experimenting with a number of models we found that the length of the line of nodes A–B along which tension is applied, rather than the direction of the applied forces, was critical in determining the stress directions in the northern Britain – southern Norway area. This result can also be appreciated from consideration of the load along A–B as a continuous line load, with an evenly distributed force per unit length, on an infinite two-dimensional half-space. The two ends of the line are then the foci of two sets of stress trajectories, one set being elliptical, the other set hyperbolic. Visual matching of ellipses (i.e. the analytical approximation to the dyke trajectories) to the dykes constrains the location of A–B fairly tightly on the NW European margin. The particular line A–B shown in Fig. 7 is our preferred model. However, it was found by the trial and error method of running many numerical models, rather than by ellipse matching.

Since the model must remain in static equilibrium, it is necessary to apply forces elsewhere in the model to balance those applied at the Faeroes. This was achieved by prescribing zero displacements along a line approximating the southern margin of the European plate (Fig. 7, boundary C–D), effectively assuming that the tension applied at the Faeroes had no effect on the lithosphere to the SE of this line. Forces pulling the overthrust plate at a subduction zone trenchwards (Forsyth and Uyeda 1975; Backus *et al.* 1981) may be the primary component of this balancing force, but as discussed above, the elastic reaction or viscous drag of the asthenosphere may also play a significant role.

In the final version the remaining NE and SW boundaries of the model (i.e. the left and right sides of Fig. 7) were left free to move. An alternative model in which zero displacements were prescribed at these boundaries produced little change in the stress directions, especially in the area near the Faeroes. Horizontal extensions are mainly responsible for rift formation, although vertical forces in the immediate vicinity of the rift are also important. Therefore these vertical forces will not significantly affect the stress trends in the area in which the quartz-dolerite dykes are found.

Interpretation of the stress pattern

The stress pattern produced by the model agrees remarkably well with the observed trends of the dykes (Fig. 8). However, the region in which the magnitudes of the deviatoric tensional stress are greatest, that is, the region in which the model predicts that dykes are most likely to form, lies on a roughly semi-circular arc centred on the Faeroes, passing through the Orkney Islands, whereas the dykes are actually to be found some 200–400 km radially outward from this region. This can be accounted for qualitatively by the assumption in the model of lateral homogeneity of the lithosphere, whereas, as we have pointed out, the contemporary lithosphere was probably much stronger (by virtue of being much thicker) in north-western Scotland and the Faeroes area, than in central and southern Scotland within the still-warm Caledonian fold belt.

The magnitudes of the stresses would thus be higher in the thinner lithosphere to the south. This means that dykes would be intruded further south than predicted by the model, given the empirical evidence for the presence of basaltic magma at depth beneath the Midland Valley and its surrounds throughout the Carboniferous (MacDonald *et al.* 1977). It is likely that models which take account of such lateral variations in the lithosphere would produce slightly different trends in the stress directions, but considering present geological uncertainties (particularly under the North Sea), and other uncertainties about the mechanical properties and thicknesses of lithosphere in shield areas and 100 Ma old orogens, we feel that detailed models which consider such effects quantitatively should not yet be attempted.

The dykes become thicker eastwards from Britain into the North Sea before their trace is lost under the Central Graben (Smythe 1994). This reflects the variations in the effective extensional stress calculated in our model (Fig. 8). However, the occurrence and trend of the Scania dyke swarm is not explained by our hypothesis. This swarm is situated near the intersection of two ancient crustal lineaments, the Fennoscandian Border Zone (the Tornquist Line) and the Småland Suture (the eastern edge of the Grenville Province) (Fig. 3b). It is probable that these pre-existing features would cause an anomalous modification of the stress in such a presumably weak area. This may be why Scania has been subjected to no less than five basic or intermediate magmatic events (including the late Carboniferous dolerites) since 1600 Ma (Klingspor 1976).

Can other models account for the arcuate dyke swarm?

We have suggested that two rifts propagated head-on towards each other, from the ends of dextral transcurrent (incipient transform) fault zones in the north and south, respectively (Fig. 3b). This geometry might seem less plausible than one in which a single rift propagated all the way from one end to the other. In the latter case, both sets

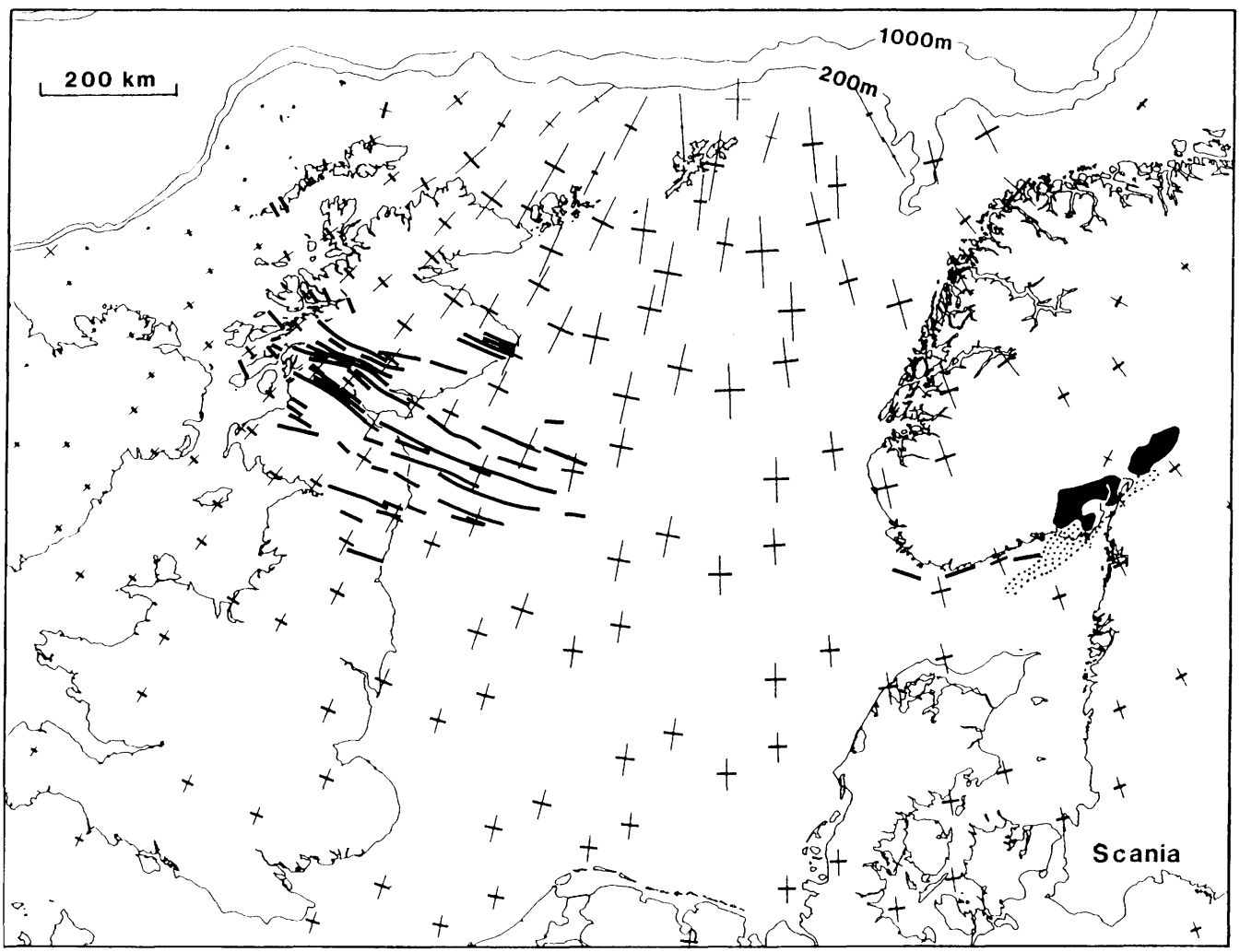

FIG. 8. Results from the finite element model of Figure 7. Calculated orthogonal principal deviatoric stresses are plotted as crosses at the centroid of each of the elements, overlain on the geological map of Figure 3a. Thick bar represents compression, thin bar represents extension. The length of each bar is directly proportional to the magnitude of the stress. Bathymetric contours at 200 m and 1000 m outline the present NW European continental margin, which follows the line of the proto North Atlantic Rift developing at 300 Ma (Fig. 3b).

of stress trajectories (compressive and tensile) would be hyperbolic in plan, and dykes would, to a first approximation, lie on an arcuate trend like the one observed. Although we have not tested this alternative model quantitatively, we suspect that a tolerably good fit of dyke trends with predicted maximum principal deviatoric stress trajectories could be obtained with a southerly rift alone, which is presumed to have propagated about as far north as the Rockall Rift shown in Figure 3b. However, the mismatch of the Oslo Graben – whether of its regional NNE trend or of the local north–south trend within it (Schonwandt and Petersen 1983) – with the predicted stress trajectories would then be very large (40° or more). A two-rift system, giving the line load force along A–B as in Figure 7, resulting in approximately elliptical stress trajectories (Jaeger and Cook 1969, fig. 10.16.3), is required. What occurred in reality may have been the following sequence:

(1) Propagation of one of the rifts as far as the Faeroe region,
(2) Propagation of the other rift to the Faeroes, resulting in the concentration of stresses shown in Figure 3,
(3) Stretching of the Faeroe region, accompanied by lithospheric separation on either side, followed by,
(4) Completion of lithosphere separation and release of stress.

We have no way at present of determining which of the two rifts developed first. However, it is interesting to note that the above sequence suggests that the Faeroe region would have behaved like a locked zone (Courtillot 1982) in the development of the proto North Atlantic Rift.

Another speculative possibility is that a rift propagated across the 'neck' in Pangaea from the Boreal Ocean to the Tethys (or vice-versa), but that the lithosphere in the Faeroes region remained stronger after completion of rifting, whether by virtue of greater friction on normal faults, higher Young's modulus in the elastic upper crust, or higher viscosity in the lower crust. Although this would not account for the essentially transient phenomenon of the concentrated stress trajectories which gave rise to the dyke swarm, it would explain why the more alkaline dykes, intruded some 10–20 Ma later (Fig. 6), seem to be controlled by a similar stress distribution. The source of the tensile stress must then have been varying in amplitude to permit short epochs of dyke intrusion.

Conclusions

The late Carboniferous dolerite dyke swarm of northern Britain is one of the major dyke swarms of NW Europe. Taking the examples of other major dyke swarms worldwide, we have looked to rifting as a cause of the intrusion. Previous models relying solely upon Hercynian subduction and continental collision cannot account either for the arcuate trend of the dyke swarm, or for the related Oslo Graben volcanism. In the region of Britain, this means that we must consider the dykes in the context of events to the NW, rather than to the SE.

Using the dykes as indicators of contemporary tensile stress, we have shown graphically that the tensile stress trajectories are concentrated through the Faeroe–Shetland region, This inference comes directly from the fact that the dykes swing through an azimuth of 45° between the Outer Hebrides and the western North Sea. The unusual distribution of lithospheric stress around 300 Ma has to be explained. Our North Atlantic rifting hypothesis suggests that rifting was occurring in the Rockall Trough and eastern Norwegian Sea regions, but for a possibly short period, there was unrifted lithosphere in the Faeroes area. Stresses were then focused through this lithospheric 'bottleneck' between NW Europe and Greenland, rather than being transmitted uniformly across the rifts.

We have tested the rift hypothesis quantitatively by a simple two-dimensional elastic finite element model of the NW European plate, the essence of which comprises free boundaries along the margins of the rifts, and tensional forces through the edge adjoining the Faeroes block. Our numerical models demonstrate a radial system of minimum principal stress trajectories centred on the Faeroes area, with not only the northern British dyke trends, but also the *overall trend of the Oslo Graben, predicted successfully* (Fig. 8).

We envisage the dykes, together with the Oslo Graben, to be a nascent lithospheric rupture. The formation of dykes requires high rates of strain, consistent with our view that lithospheric rifting had already taken place in the Rockall Trough and eastern Norwegian Sea. We assume that this period of dyke intrusion and volcanism was short-lived, as these two rifts eventually coalesced across the locked zone (Courtillot 1982) of the Faeroe area, to bring about the inception of the Faeroe–Shetland Trough. Had this not occurred, the proto North Atlantic Rift zone might instead have developed in an east–west direction across Scotland and the central North Sea.

The tholeiitic igneous activity was succeeded by a period of major alkaline plutonism of early Permian age in the North Sea region (see for example, Oftedahl 1978; Dixon *et al.* 1981), together with the minor episodes of alkaline basic intrusion in western Scotland. Both sets of later events are closely associated both in space and time with the dyke swarm.

We conclude that this quantitative study of the dyke swarm provides independent evidence for the inception of a proto North Atlantic Rift in late Carboniferous time. However, it does not allow us to estimate when rifting became 'drifting', with the creation of the 250 km-wide zone of quasi-oceanic crust in the Rockall Trough, and the rather narrower width of thinned and subsided continent in the eastern Norwegian Sea.

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