

The structural relations and regional tectonic implications of metadolerite dykes in the Kongsfjord Formation, North Varanger Region, Finnmark, N. Norway

A. H. N. RICE & W. REIZ

Rice, A. H. N. & Reiz, W.: The structural relations and regional tectonic implications of metadolerite dykes in the Kongsfjord Formation, North Varanger Region, Finnmark, N. Norway. *Norsk Geologisk Tidsskrift*, Vol. 74, pp. 152–165. Oslo 1994. ISSN 0029-196X.

Mafic dykes in the Kongsfjord Formation, Barents Sea Group, North Varanger Region, lie sub-parallel to the steeply dipping axial surfaces of NE–SW to ENE–WSW trending F1 folds formed during dextral strike-slip movement along the Trollfjorden–Komagelva Fault. Previously, it has been asserted that the dykes were broadly syntectonic with the folding. However, several lines of evidence indicate that the dykes pre-dated deformation; sediments within unbroken bridge structures are folded along atypical axial trends; atypical dyke directions (e.g. in linking dykes in bridge structures) have atypical fold axial orientations at their margins; the position of fold axes was affected by bridge structures in the dykes; thin dykes are offset by flexural-slip folding. These folds all have the typical S1 slaty cleavage. In the proposed model, dyke emplacement was caused by pure-shear extension, related to a regional late Sturtian–early Vendian reactivation of the rift systems surrounding Baltica, and predated dextral strike-slip movement. This extension may have caused a minor component of sinistral movement along the Trollfjorden–Komagelva Fault.

A. H. N. Rice* & W. Reiz, *Geologisch-Paläontologisches Institut, Ruprecht-Karls Universität, Im Neuenheimer Feld 234, 69120 Heidelberg, Germany* *Present address: *Institut für Geologie, University of Vienna, Universitätsstraße 7, 1010 Vienna, Austria*.

Introduction

Metadolerite dykes cutting the late Riphean and Sturtian Barents Sea and Løkvikfjellet Groups in northern Varangerhalvøya (Fig. 1) are locally abundant and lie essentially sub-parallel to the axial surfaces of folds (Roberts 1972, 1975; timescale of Harland et al. 1989). Previous publications have stated that these dykes were emplaced *broadly* syntectonically (Roberts 1972, 1975; Siedlecka & Roberts 1992) although no detailed supporting evidence has ever been published, despite the significance of this statement. This article presents preliminary observations recorded to evaluate the relationship of the dykes to the folds and to assess their regional tectonic significance.

Regional geology

The North Varanger Region of East Finnmark, N. Norway (Fig. 1), is an epizone grade fragment of the Caledonian Lower Allochthon (Rice et al. 1989a; previously this area was known as the Barents Sea Region, but this name is thought to be misleading, since the Barents Sea region is, of course, the Barents Sea). The North Varanger Region is juxtaposed against diagenetic to lower anchizone grade rocks of the Gaissa Thrust Belt (also Lower Allochthon) and East Finnmark Autochthon, along the steeply dipping WNW–ESE trend-

ing Trollfjorden–Komagelva Fault (Fig. 1), a fault with a polyphasal movement history (see below).

The rocks on Varangerhalvøya form the most easterly part of the Scandinavian Caledonides. The Gaissa Thrust Belt continues westwards from this area and forms the largest and external part of the Lower Allochthon in Finnmark. The internal part of the Lower Allochthon is represented by the Komagfjord Antiformal Stack and Laksefjord Nappe Complex, both of which contain pre-Caledonian basement lithologies (Føyen et al. 1983; Pharaoh 1985; Gayer et al. 1987). These two units suffered Caledonian epizone metamorphism, comparable to that in the North Varanger Region and this similarity was used to restore the three units to similar along-strike positions in a palinspastic reconstruction (Gayer & Rice 1989; Rice et al. 1989b). Deformation within the Lower Allochthon occurred in the Scandian event between ca. 425 and 380 Ma (Dallmeyer et al. 1988, 1989; Roberts & Sundvoll 1990).

Lithostratigraphically, the rocks north of the Trollfjorden–Komagelva Fault are largely distinct from those lying to the south, although of the same age (cf. Siedlecka & Roberts 1992 for stratigraphic details). To the north of the fault a 15.81 km-thick late Riphean to ?Cambrian sequence has been mapped, comprising the Løkvikfjellet Group unconformably overlying the Barents Sea Group. In the northwest, the Løkvikfjellet Group has been overthrust by turbiditic rocks of the Tanahorn Nappe (Berlevåg Formation; Levell & Roberts 1977). To the

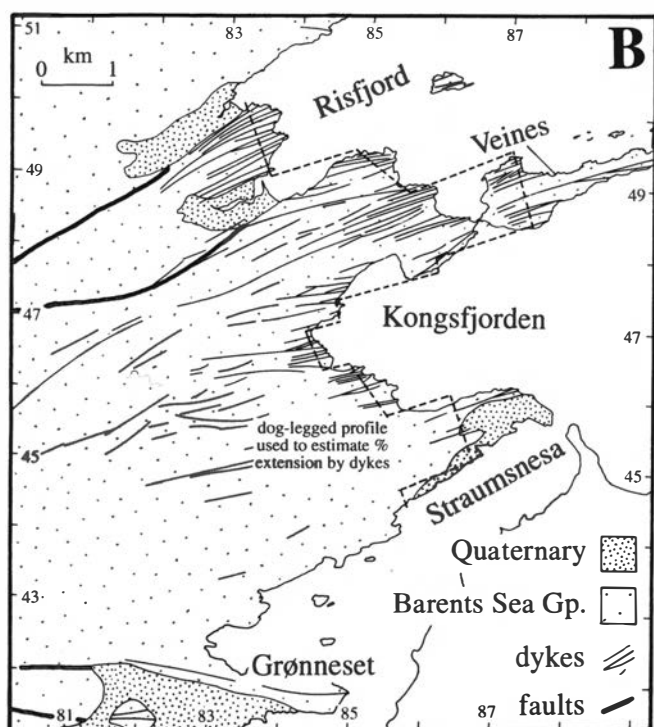
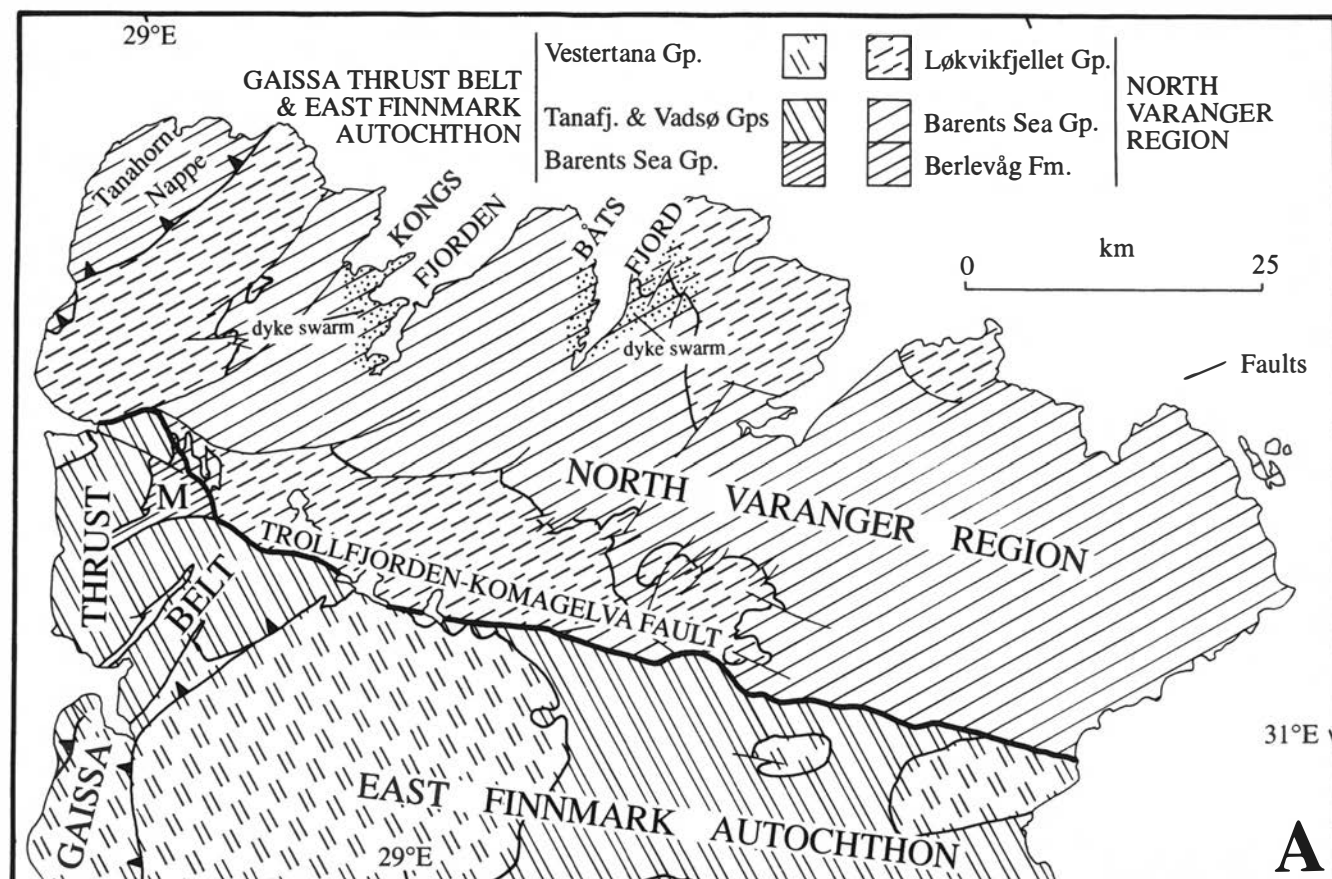


Fig. 1A. Geological map of Varangerhalvøya, showing location of the Båtsfjorden and Kongsfjorden dyke swarms and the main stratigraphic units. M = Manjunnas area. Simplified from Siedlecki (1980) and Siedlecka (1987, 1989). B. Map of the Risfjorden-Kongsfjorden-Straumen area, showing some of the dykes (from Siedlecka 1989).

south of the Trollfjorden-Komagelva Fault a 5.84-km-thick sequence of late Riphean to Ordovician age has been found in the Gaissa Thrust Belt and East Finnmark Autochthon, comprising the Vadsø, Tanafjorden, Vestertana and Digermulen Groups. In the Manjunnas area of the Gaissa Thrust Belt (Fig. 1) the stratigraphy found south of the Trollfjorden-Komagelva Fault unconformably overlies rocks typically found to the north of the fault, providing an important stratigraphic link across the fault (Rice 1994). Both the sequences to the north and to the south of the Trollfjorden-Komagelva Fault contain evidence of a major end Sturtian to early Vendian unconformity.

The stratigraphy reveals a three-fold increase in sedimentary thickness across the Trollfjorden-Komagelva Fault (the Manjunnas area is an exception). The rocks to the north were deposited within the WNW-ESE trending late Riphean Timanian Aulacogen, whilst rocks to the south were deposited on the rift flanks, with the Trollfjorden-Komagelva Fault forming the aulacogen margin (Siedlecka 1975, 1985).

The displacement and sense of lateral movement along the Trollfjorden-Komagelva Fault remain in doubt, but were polyphasal (Karpuz et al. 1993). Torsvik et al. (1994) determined a net strike-slip displacement less than the resolution of the palaeomagnetic method they used (ca. 200–300 km).

Taylor & Pickering (1981) determined a whole-rock age of 520 ± 47 Ma (MSWD 1.35) from cleaved mudstones of the Kongsfjord Formation in eastern Varangerhalvøya. Reuter & Dallmeyer (1988) have shown that whole-rock dating of very low-grade cleaved rocks gives numbers that are a mixture of the age of detrital material and neocrystalline grains. Thus the data of Taylor & Pickering (1981) are probably of a mixed detrital and metamorphic age. However, the date is an indication of the maximum age of cleavage formation.

Age and composition of metadolerite dykes in the North Varanger Region

Two metadolerite dykes suites have been dated in the North Varanger Region (Fig. 1; Beckinsale et al. 1976; all K-Ar data recalculated according to Dalrymple 1979). *Group A* dykes, which lie in the Kongsfjord area, have K-Ar ages of ca. 945–1945 Ma, older than the estimated age of the sediments (Vidal & Siedlecka 1983), suggesting that the dates are spurious. *Group B* dykes, which have ages of between 550 and 650 Ma, come from the Båtsfjord area and the Tanahorn Nappe (Fig. 1). The younger ages come from strongly sheared rocks of the Tanahorn Nappe and probably represent partial or full resetting during Caledonian orogenesis. The dykes in Båtsfjord, which are essentially undeformed, cut across the end Sturtian–early Vendian unconformity (Siedlecki & Levell 1978) and the age determined from them is inferred to be the emplacement age. Note that these ages are all older than the maximum age of cleavage formation, determined by Taylor & Pickering (1981; see above).

The metadolerites in the Tanahorn Nappe and Kongsfjord, which have transitional WPB–MORB compositions, have been related to rifting processes (Roberts 1975), in the development of the Timanian Aulocogen. The extension required to develop such a deep basin suggests a high β factor, with which the presence and composition of the dykes are consistent (McKenzie 1978). These Kongsfjord dykes have not been dated, but it is here presumed that they are broadly contemporary with the Båtsfjord dykes.

The Kongsfjord area

Kongsfjord Formation – lithology and structure

The Kongsfjord Formation is a >3.5 km-thick sequence at the base of the Barents Sea Group and is the oldest stratigraphic unit known within the Lower Allochthon of

this region (cf. Siedlecka & Roberts 1992). The unit consists of well-bedded turbiditic black shales and greywacke sandstones, locally containing spectacular soft sediment slump/thrust horizons (Fig. 3A).

Roberts (1972) presented an overview of the structure of the North Varanger Region. The structural data presented here from the Kongsfjord region have been processed using the program 'Stereo' (© D. B. McEachran), in which the K and C parameters are calculated (Woodcock 1977; Table 1). The K parameter is a measure of the shape fabric of the data; $K = 0$ indicates a perfect girdle whilst $K = \infty$ indicates a perfect cluster, with $K = 1$ defining the cluster/girdle boundary. The C parameter is a measure of the quality of the shape fabric; $C = 0$ indicates a uniform distribution, $C < 2$ indicates a diffuse (tending to random) data set and $C > 4$ indicates a concentrated data set (i.e. a tight cluster or a well-defined girdle).

The most prominent structure is an upright to steeply inclined NE–SW to ENE–WSW striking cleavage (S1;

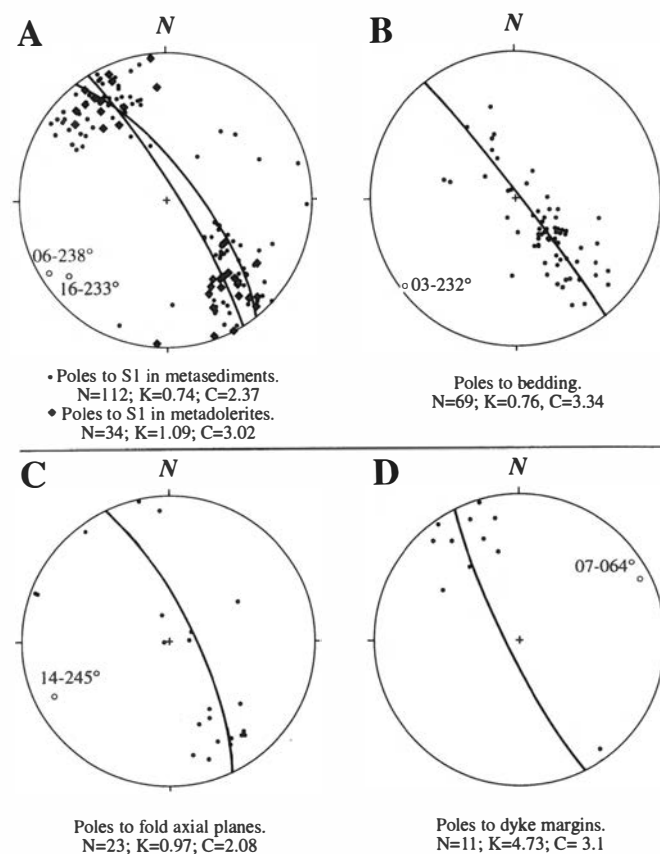


Fig. 2. Equal area lower hemisphere projections of structural data. A. Poles to S1 in the metasediments and metadolerite dykes. B. Poles to bedding. C. Poles to fold axial surfaces. D. Poles to dyke margins. Open circle is the mean lineation vector.

Fig. 3A. Spectacular soft sediment slump thrust fault. Folding and refolding are pervasive and an early (?compactional) cleavage is folded. Although the folding is chaotic, the axes have a consistent direction, plunging 47° towards 283° . West of Straumnesa. B. Thin metadolerite within black mudstones of the Kongsfjord Formation containing two cleavages. One cleavage is parallel to the dyke margin, the other is markedly oblique and is refracted across the dyke. View looking down onto a bedding plane. North of Straumen. C. Two cleavages in a dyke at a bridge; one is sub-parallel to the hammer handle and varies slightly in orientation in different parts of the dyke. The other cleavage is sub-parallel to the margin of the main dyke and is best seen where it is thinning out, to the right of the bridge. Southeast Kongsfjord. D. Well-exposed margin of a metadolerite dyke showing abundant quartz veins regularly spaced in some areas and which in some instances have a sigmoidal *en echelon* form. Cliff is ca. 30 m high. Southeast coast of Veines.



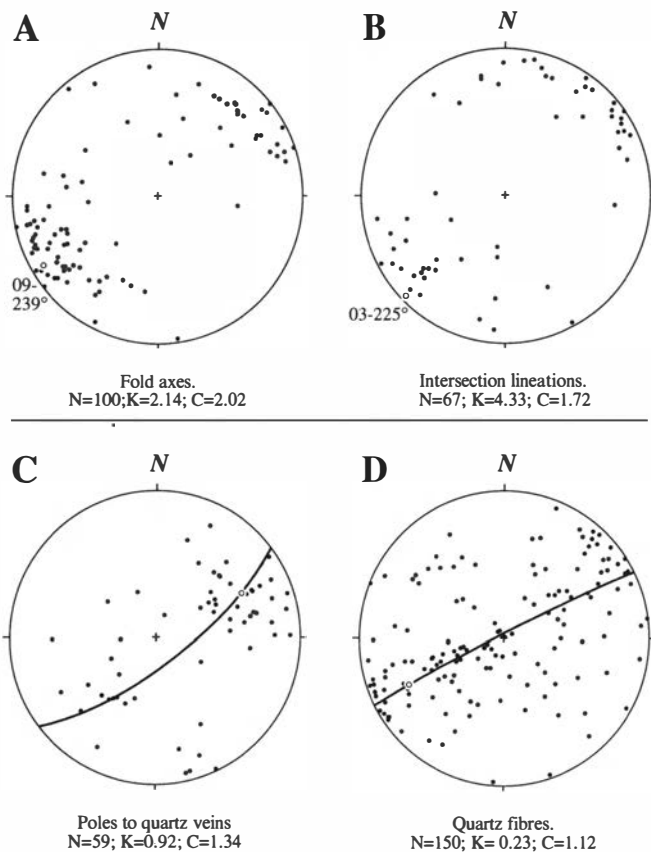


Fig. 4. Equal area lower hemisphere projections of structural data. A. Fold axes. B. Bedding-S1 intersection lineations. C. Poles to quartz veins cutting the metadolerites. D. Quartz fibres within the quartz veins. Open circle is the mean lineation vector.

Fig. 2A), which is penetratively developed in pelitic lithologies and variably developed in sandstones and siltstones. The upright nature and orientation of the cleavage is consistent with it developing during dextral strike-slip movement along the Trollfjorden–Komagelva Fault. In some cases, particularly in the Straumsnesa area, two sub-vertical cleavages have been observed (Fig. 3B; first noted by Dr C. Townsend, pers. comm.). In folds, these both transect the axial surface such that on one limb one cleavage tends to be prominent and on the other limb the other cleavage is prominent. Transecting cleavages are again a feature of strike-slip deformation regimes.

Small- to intermediate-scale folds are locally abundant within the Kongsfjord Formation and are often associated with the metadolerite dykes (see below for these relationships). The mean fold axis is 09° towards 239° (Fig. 4A), at ca. 45° to the trend of the Trollfjorden–Komagelva Fault and consistent with a dextral strike-slip regime. The K parameter indicates a clustered distribution, although the low C parameter indicates a diffuse data set, reflecting both the range of axial trends developed and the periclinal nature of the folds. Fold profiles are essentially parallel in form, with upright to steeply inclined axial surfaces and close to open profiles. Wavelengths vary from <1 m to 20 m, with larger folds being

Table 1. Summary of the structural data illustrated in Figs. 2, 4 and 5. PBFGC = pole to best fit great circle; MLV = mean lineation vector.

Equal area lower hemisphere data					
Structure	N	PBFGC	MLV	K	C
S1 sediments	112	16–233°	01–142°	0.74	2.37
S1 dykes	34	06–238°	07–148°	1.09	3.02
Bedding	69	03–232°	61–138°	0.76	3.34
Axial surfaces	23	14–245°	24–148	0.97	2.08
Dyke margins	11	07–064°	18–332°	4.73	3.10
Fold axes	100		09–239°	2.14	2.02
Inters. Lin.	67		03–225°	4.33	1.72
Quartz veins	59	15–323°	34–064°	0.92	1.34
Quartz fibres	150	02–153°	28–244°	0.23	1.12

Rose diagram data				
Data set	N	Range	Vector mean	Conf. angle
Airphoto	306	0–360°	070°	3°
Field	50	0–360°	060°	9°

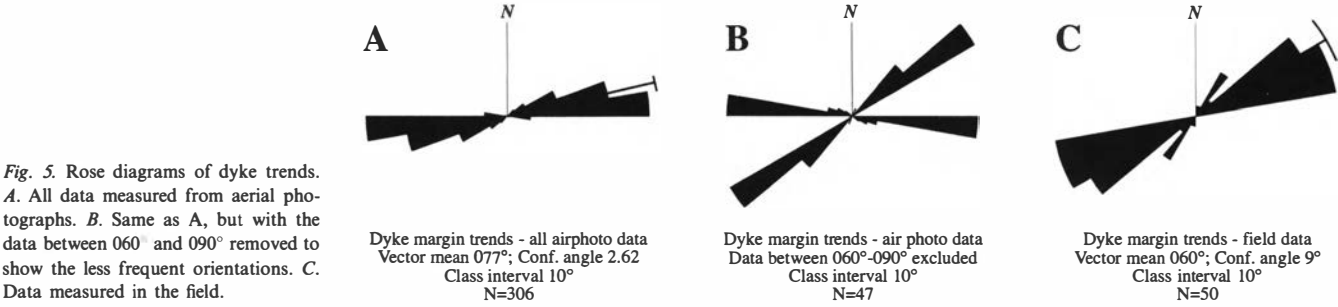
more open. Bedding-plane slickensides and offset markers demonstrate a flexural-slip fold mechanism.

The S1 cleavage is fanned around fold axial surfaces, with a fracture cleavage in sandstones and a slaty cleavage in mud/siltstones. The pole to best fit great circle through cleavage planes gives a regional fold axis plunging 16° towards 233°, similar to that determined from bedding readings (03° towards 232°) and the limited axial plane data (14° towards 245°; Fig. 2A, B, C; Table 1). The transecting nature of the cleavage in some areas will have caused a spread in the data, but since there are two cleavages transecting, with opposite senses of rotation, the transecting geometry should not have substantially affected the mean orientation. Bedding-S1 cleavage intersections give a mean orientation of 03° towards 225° (Fig. 4B); the argument pertaining to the transecting cleavages also applies here.

Metadolerite dykes in the Kongsfjord area

Sub-vertical metadolerite dykes are abundant in the Straumen–Kongsfjord-Risfjord area and, although they do not penetrate far inland (Fig. 1), some dykes can be traced for several kilometres. Most of the dykes are essentially sub-parallel, trending NE–SW to ENE–WSW, mirroring the orientation of normal faults within the region (Roberts 1972; Siedlecka 1989), although cross-cutting relationships occur. Regionally, there is a swing in trend from NE–SW in the Risfjord area to ENE–WSW in the Straumsnesa area and to WNW–ESE further south, on Grønneset (Fig. 1).

The regional mean direction, determined from aerial-photo analysis, is 077° (Fig. 5A), indicating a NNW–SSE extension direction. Fig. 5B shows the orientations recorded with data between 060° and 090° removed, highlighting the less frequent trends. The mean orientation of dyke trends recorded in the field is 060° (Table 1).



The angular difference between the regional mean and the measured mean reflects the deliberate sampling of dykes with anomalous directions in the field, which tend to be short and could not be measured accurately on the aerial photographs, and so were left out of that analysis. Dykes are typically 1–2 m thick, although they may be as thin as a few centimetres (Fig. 3B & C) and thin out entirely, or up to 5 m thick. Using an estimated typical width of the dykes of 1.5 m and a minimum of 115 dykes on a 5-km-long dog-legged section across the Straumen–Risfjord area (Fig. 1) gives a local extension of only

2.3%. The overall extension, accommodated by faults, must have been greater for dyke intrusion to have occurred (cf. McKenzie 1978). Thick dykes have both chilled margins and baked sedimentary margins. Intrusions into sandstones have sharp margins whilst those into mudstones often have irregular profiles (Figs. 7A & B). Rarely, a poorly preserved hexagonal jointing has been discerned. Most dykes are persistent along strike, although they are frequently slightly offset by bridge structures (Figs. 3B, 6, 7A, B, C; Nicholson & Ejiofor 1987). In thin dykes the

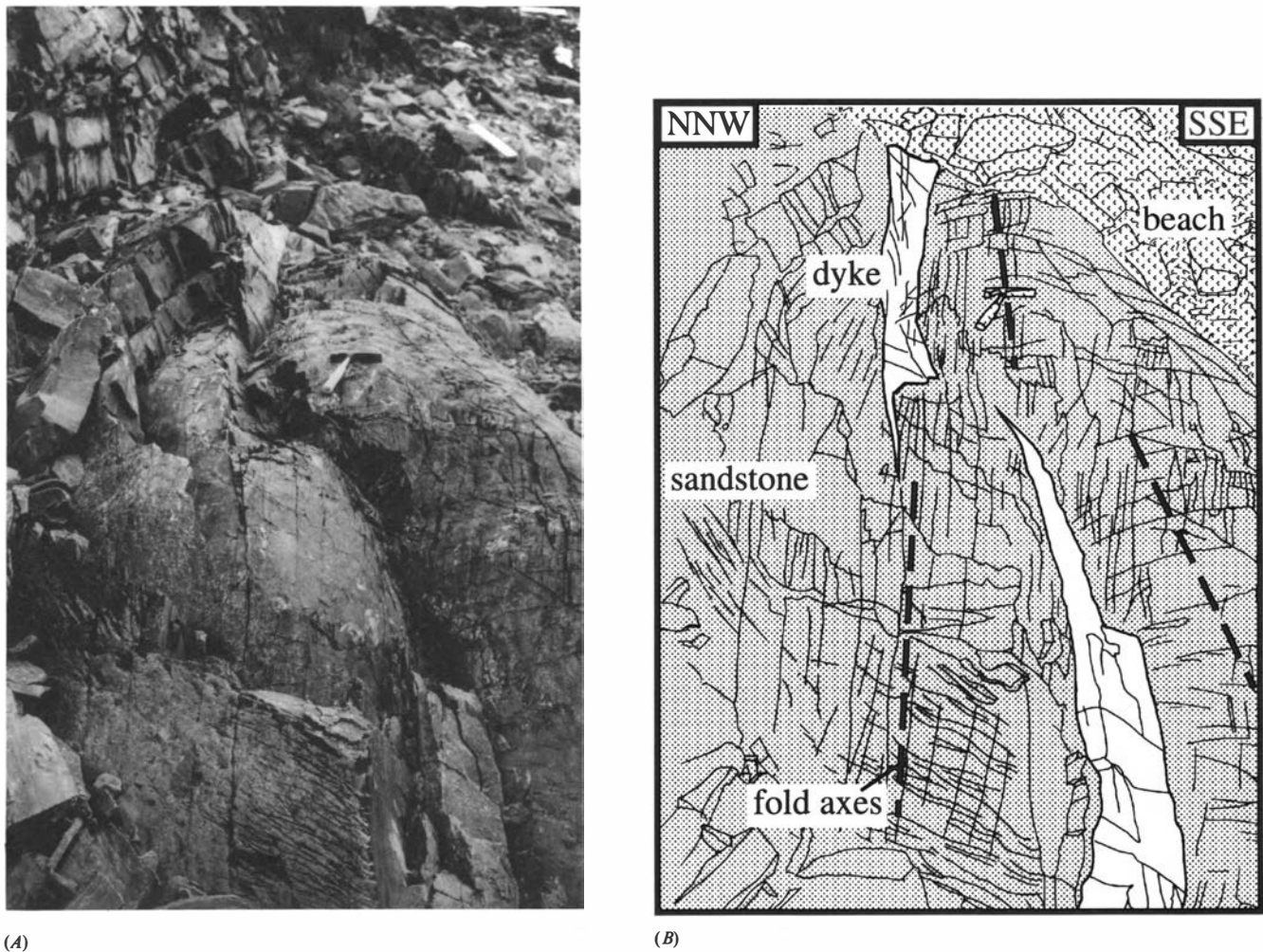


Fig. 6. Photograph (A) and accompanying drawing (B) of a bridge structure in a dyke running along the crest of a fold to the north of Straumen. The structure is exposed essentially as a single folded bedding surface. Anticlinal fold axes shown as heavy dashed lines; these are offset at the bridge structure.

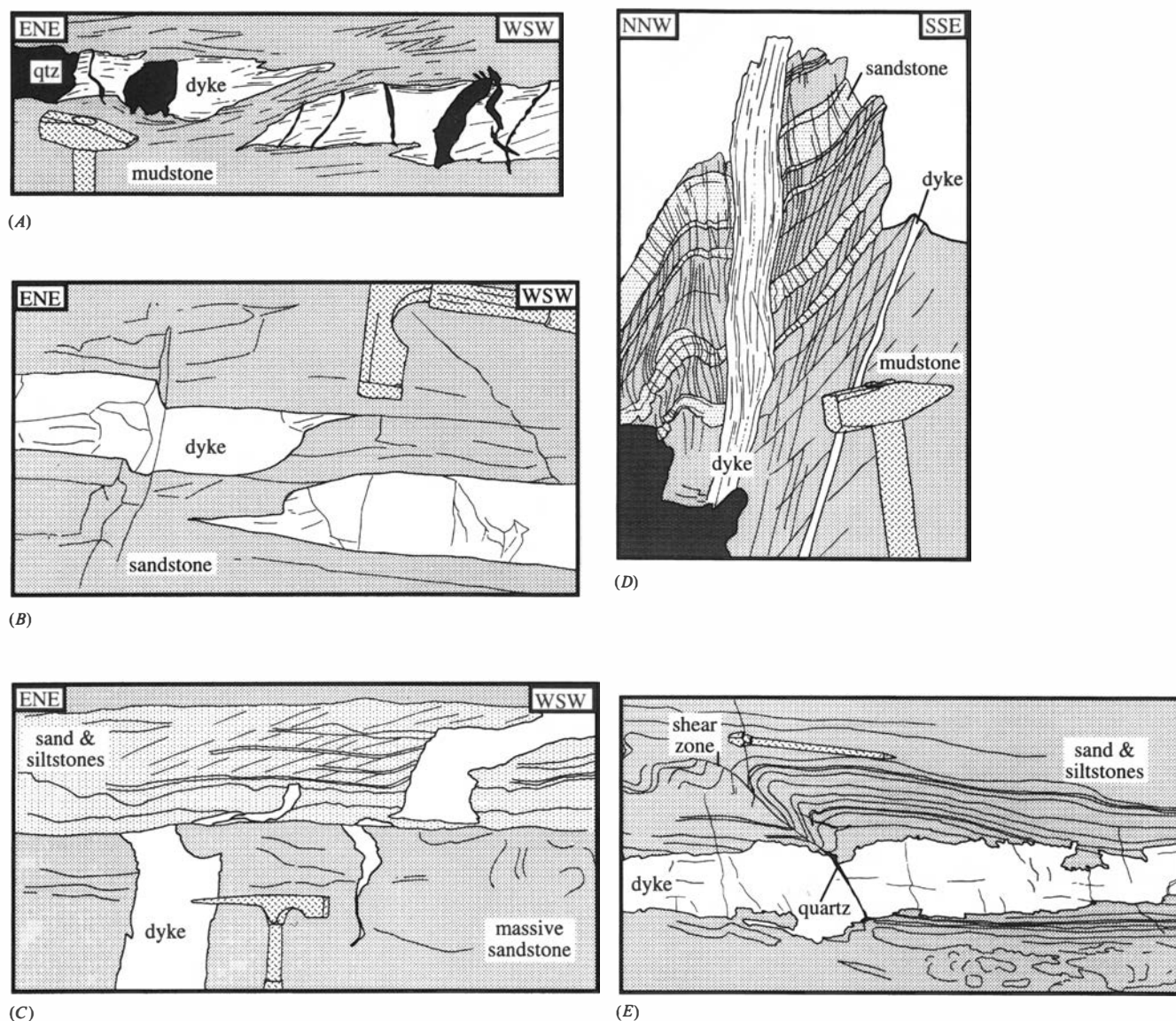


Fig. 7. Accurate drawings from photographs of structures on the north side of Straumen. A. Typical bridge structure in a metadolerite cutting black mudstones, within which two cleavages are present. Note the abundant quartz veins, representing significant extension, and the irregular dyke margin. B. Typical bridge structure in a metadolerite cutting sandstones. Note the planar margins of the dyke. C. Bridge structure in thin dyke offset by flexural-slip folding along an axis oblique to the dyke trend. D. Well-exposed example of a fold developed at the margin of a thin metadolerite dyke. The cleavage within the sandstone beds is somewhat more strongly developed nearer to the dyke, reflecting shear along the dyke-sediment interface. E. Plan view of ambiguous structure at the margin of a narrow dyke cutting thin-bedded mudstones and sandstones. See text for details.

bridges result in lateral offsets of generally <15 cm and the sedimentary bridge may be unbroken, although abundant brittle deformation can be seen in the sediments in the bridge zones. In thick dykes such minor bridges which may have formed during the initial stages of their emplacement are now wholly broken and cannot be observed except as minor offsets of the dyke margins where the strain is low. Large-scale bridges have also been found, either indirectly from an *en echelon* intrusion pattern, or, more directly, when the bridge breaks and a linking dyke segment develops. Field measurement of the trends of these linking dykes accounts in part for the large number of dykes with anomalous orientations in Fig. 5C. Linking dykes may also be segmented by bridge structures.

Within the dykes, polyphasal fibrous quartz-calcite veins, up to a metre thick, are abundant (Fig. 3D), especially in dykes cutting mudstones, where there is a significant ductility contrast between dyke and country rock. In most cases the veins appear to be randomly oriented, although dominantly sub-vertical. In some cases, however, thicker veins form regularly spaced sets, whilst others have a sigmoidal *en echelon* form (Fig. 3D). Some veins have cut into the hard contact-metamorphosed baked margins, although in thin dykes the quartz forms irregularly shaped blocks which have abrupt margins with the adjacent sediments (Fig. 7A). The mean lineation vector for the pole to the vein surfaces plunges 34° towards 063° (Fig. 4C); the K parameter is close to the girdle-cluster transition and the C parameter is very

low (Table 1), indicative of a relatively incoherent data set, although the mean pole direction is sub-parallel to the dyke trend, indicating that, statistically, the veins are normal to the dyke margins. The mean plunge angle is not in itself thought to be particularly significant, as the C parameter is so low. The large variation in vein orientation may reflect the primary hexagonal jointing, although this is not demonstrably so.

Quartz fibres within the quartz veins are ubiquitous, having either a very fine elongate fibrous form or a much coarser blocky habit. The fibres, which show a wide range of orientations and are not necessarily normal to the quartz vein margin, have a mean direction of 28° towards 244° , a similar trend to the mean pole of the veins (Fig. 4D), but again the data set is very incoherent statistically (Table 1) and it seems probable that the plunge angle is not particularly significant. The best fit great circle through the fibre orientations is sub-parallel to the mean dyke margin surface (Table 1). Altogether, the veins and fibres indicate a chocolate-slab boudinage (Harland & Bayly 1958).

Dyke-sediment relationships

Many folds in the area, both anticlines and synclines, lie adjacent to mafic dykes. These folds are frequently also periclinal, so that folds die out and reappear along dyke margins. Conversely, not all dykes have folds developed at their margins.

Detailed measurements by Roberts (1972, fig. 14) indicated that in some cases the dyke margin has a slightly shallower dip to the southeast than the S1 fabric in the adjacent sediments and a more northerly strike, but in many cases they are parallel. Comparison of the mean lineation vector data for poles to S1 and the available poles to dyke margins reveals a more northerly mean direction for the dyke margins (Fig. 2A, D), although the mean of the cleavage dips towards the northwest. More significantly, the available data show that the mean dyke margin strike is parallel to the mean fold axial plane strike (Fig. 2C, D), although the former dips to the southeast and the latter more steeply to the northwest. The angular difference in dip of these two surfaces is in the order of 42° , implying that the dykes must cut across the fold axial surfaces, unless the latter die out up/down the dip of the axial surface.

Since the cleavage is fanned around folds, due in part to the lithology, and the above regional data show that the dip direction for the S1 and dyke margin is different, the detailed observations of Roberts (1972), for which the S1 measurements are presumed to come from immediately adjacent to the margin, imply either that the dykes lie exactly along the axial surfaces, which is not supported by field observations, or that greater shearing has occurred along these dyke margins, a zone of major rheological anisotropy, to produce a cleavage specifically related to a dyke margin high strain zone. This can be

seen both in the sediments in some instances, where a more intense or closely spaced fabric has developed in the quartzo-felspathic lithologies adjacent to the dyke (Fig. 7D), and in the more highly sheared and often weathered out margins of some dykes. Baked sedimentary margins carry a cleavage, but qualitatively this often appears to be less intense than in unbaked margin rocks, reflecting the greater competence of the contact-metamorphosed sediments.

Thick dykes may have an unfoliated or only weakly foliated centre (cf. Roberts 1975 for mineralogy and textures of the dykes). The margins, however, are foliated and thinner dykes are penetratively deformed. Regionally, the orientation of S1 within the dykes is parallel to that in the sediments (Fig. 2A, Table 1). In bridge zones the cleavage may be oblique to the dyke margin, and the dyke itself may have refracted the cleavage orientation. Locally, two cleavages have been found cutting dykes (Fig. 3C), especially in association with bridge structures.

The data show that on a regional scale the minor fold axes are sub-parallel to the trend of the dykes (Table 1). The difference in direction is significantly reduced if the dyke trend data collected in the field are used, rather than the aerial photographic interpretation data, since these dykes come from the same area as the measured folds.

Atypical dyke-sediment structures

Seven examples of dyke-fold relationships which are atypical of the region are described here; these are of particular importance because they give a different insight to the regional structural history. Grid references to the localities are given for 1:50,000 map sheet KONGS-FJORD 2336 II.

In *Example A* (GR 8886 4959) a dyke trending 080° joins or intersects a dyke with an atypical trend of 022° . Bedding beside the dyke is openly to closely folded, with an axis plunging 22° towards 030° , sub-parallel to the dyke. Two cleavages have developed, both of which show a fanned relationship, and are thus related to that fold and with an intensity not less than that in D1 folds seen elsewhere. Intersection lineations, adjacent to a nearby dyke, which trends 259° , plunge 10° towards 246° , sub-parallel to the dyke. In *Example B* (GR 8686 4612) a linking dyke trends 042° , compared with the regional trend of 070° . Minor F1 folds in the vicinity of the linking dyke trend between 018° and 052° , a direction which in some instances is parallel to the linking dyke and elsewhere lies closer to parallel to the regional trend. In these examples the trend of the fold axis is controlled by the orientation of the dyke, which must, therefore, have predated the deformation.

In *Example C* (GR 8861 4928) a dyke with a trend of 032° , which may be a linking dyke between two large dykes trending 252° , is divided by an unbroken sediment bridge structure trending 045° . Sediments within the

bridge are folded, with an axis plunging 18° towards 040° , sub-parallel to the trend of the bridge structure. Again, the axial planar cleavage has an intensity typical of D1 folds. Minor folds away from the linking dyke plunge 33° towards 251° , parallel to the trend of the nearby major dykes. Here, therefore, the fold axis orientation is controlled by the orientation of the bridge structure and consequently the dyke must have preceded the deformation.

In two outcrops on the north side of Straumen, dykes <25 cm thick have been broken across and offset by flexural slip during folding. In *Example D* (GR 8714 4616) the dyke margin trend is sub-parallel to the fold axis, so that in plan view the dyke does not appear to have been affected by the folding. However, in *Example E* (GR 8686 4612) the thin tail of a main dyke is markedly oblique to the fold axis developed adjacent to the linking dyke in *Example B*. In plan view the dyke is exposed as a series of *en echelon* segments, with the dyke thick and normal to bedding in the sandstones and thin and oblique to bedding where it cuts less competent horizons and has been deformed (Fig. 7C). Slickensides on the bedding parallel shear planes cutting the dyke are sub-vertical and indicate a reverse fault displacement along the bedding planes, consistent with a flexural-slip fold mechanism. In *Example D* it is possible that some of the folding occurred prior to dyke intrusion, but this is not the case with *Example E*, where intrusion must have pre-dated folding, since the fold axis is parallel to the oblique trend of the linking dyke.

In *Example F* (GR 8714 4616) a 25-cm-wide dyke lies along the crest of a fold developed at the margin of a ca. 2 m-thick dyke. Where the thin dyke bridges, the axis of the fold has split into two parts, both of which are displaced relative to the axis on the other side of the bridge. The relationships can be seen in Fig. 6, which is a view along the length of the dyke over the bridge and in which the sediments are exposed essentially as a single bedding surface. In this case the dyke must have been present before much or all of the folding.

Example G (GR 8680 4607; Fig. 7E) lies adjacent to a dyke with a very irregular margin, suggesting low strain, although the dyke is cleaved. The dyke is cut by a very thin quartz vein, against which the bedding has been displaced, although the dyke margins are essentially not offset. On the one hand, it could be argued that the fold is cut by the quartz-filled shear and that this was later reactivated to cut through the dyke, so that the dyke post-dated the fold. However, this cannot be conclusively demonstrated, since the exposure is on a wave-smoothed surface and no section normal to the axis is exposed. Alternatively, it could be argued that if the slip vector in the quartz shear was parallel to the dyke margin, as is the general case for the quartz veins (see above), slip along the shear would not significantly offset the margins of the dyke, but would affect any fold within the adjacent sediments, in which case the dyke pre-dates the quartz shear and could pre-date the fold. The observation that

the dyke cuts some beds on both limbs of the fold is not significant; since the fold axis is parallel to the surface of the picture, any fold with a curved hinge line developed at the margin of a dyke will have such an outcrop pattern in a plan view.

Discussion

Previous models

Roberts (1972, 1975) and Siedlecka & Roberts (1992) regarded the dykes as being *broadly* syntectonic with D1. Roberts (1972) wrote that '... after the bulk of the fold development, the dykes were emplaced along the axial surfaces; the subsequent formation of cleavage testifies to a continuation of D1 strain'. In support of this, Siedlecka & Roberts (1992) figured a minor upright open fold with a dyke lying markedly oblique to the axial surface (reproduced here as Fig. 8A). They also figured a moderate to gently plunging fold with axial surface traces oblique to the dyke trend (Siedlecka & Roberts 1992, fig. 38).

In Fig. 8B the fold has been unfolded, using a very simplistic line-length technique and assuming both a flexural-slip fold mechanism and constant bed thickness, although the latter is clearly not strictly valid. The essentially planar and vertical axial surface of the synform on the left has been taken as a marker plane, since in strike-slip deformation the XY plane (and thus the axial plane) remains vertical during progressive strain. Fig. 8B illustrates that, after unfolding, the dyke is still essentially planar, although the acute angle between the main dyke and both the marker axial surface and the apophysis has increased. The curvature of the axial surface of the fold containing the dyke/apophysis is seen in both the deformed and unfolded state and is a result of the relative thinning of the pelitic rocks in beds D–G (Fig. 8A) on that limb.

Fig. 8C & D shows the effects of restoring the structure to a pre-dyke emplacement configuration for the folded and unfolded case. In the case of the dyke post-dating the fold, different closing directions are required for the apophysis and the main dyke, each essentially parallel to the local bedding (Fig. 8C). In the fold pre-dating the dyke situation only a single restoration direction is required, parallel to bedding (Fig. 8D).

These geometrical points show that, on the basis of the crude restorations undertaken, the structure is compatible with models of dyke intrusion both pre-folding and syn- to post-folding. This is itself, however, demonstrates that the original figure is spurious to arguments about the dyke–fold age relationships, and cannot be used to support *either* model.

Folds with an axis oblique to and apparently cut across by a dyke (e.g. Siedlecka & Roberts 1992, fig. 38) are to be expected, since the dykes are not exactly parallel to the fold axes in all cases. Compression oblique

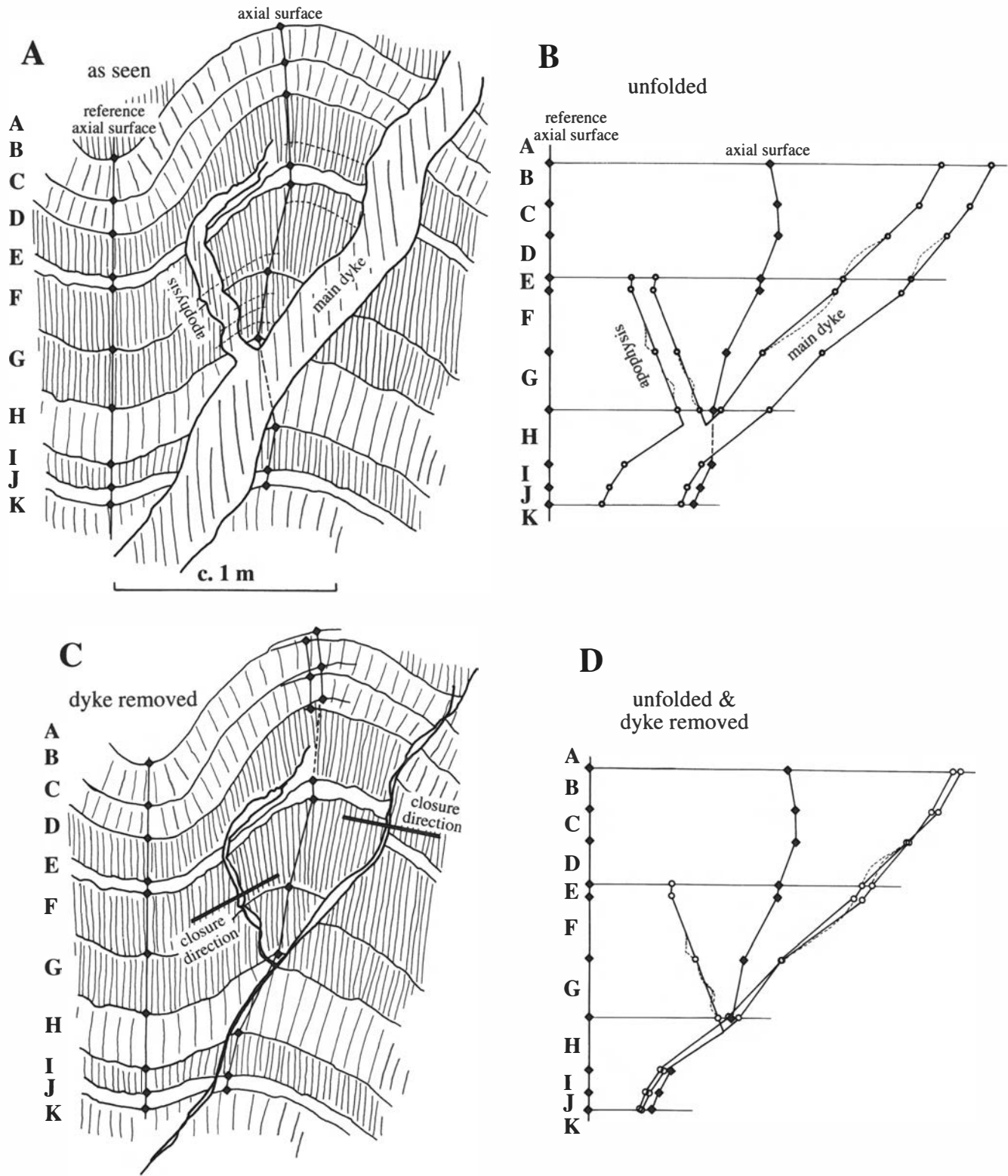


Fig. 8A. Sketch and restoration of a dyke lying markedly oblique to the fold axial surface (from Siedlecka 1990; Siedlecka & Roberts 1992, fig. 38). Dashed lines are inferred bedding surfaces. B. Unfolded form of A, assuming flexural-slip folding and constant bedding thickness (not strictly valid). Dashed lines are the detailed morphology of the dyke using the inferred bedding surfaces in A. C. Closed up version of A. Heavy lines indicate relative movement vectors. D. Closed up version of B. Better fit of bedding across the closed dyke in this case compared to C results from the assumed constant bed thickness used in B, and is therefore spurious.

to the dyke will cause folding with an axis orientation part way between the regional trend and a direction modified by the obliquity of the dyke to that trend. Space problems caused by the sediments shortening

oblique to the dyke margin could have been accommodated by sub-vertical slip along the dyke margin, producing the more intense fabric seen in some examples at the margin, both in the dykes and sediments.

Criteria of dykes cutting folds

Where fold axes are sub-parallel to dykes, structural criteria to assess whether small-scale folding occurred before or after intrusion are limited to cases where the dyke cuts across the same bedding surface on both limbs of a fold in a section effectively normal to the fold axis (assuming no obvious difference in fabric history in the dyke and wall rock or in metamorphic grade). For this to occur, the angle between the dyke margin and axial surface must be more than half the fold interlimb angle. At greater angles the dyke will only cut through a particular bedding surface on one limb, although altogether both limbs of the fold may be cut (e.g. Fig. 8A). Since the dyke will invariably and inevitably lie oblique to the axial surface of all folds, unless either the dyke margin and axial surface are exactly parallel or the fold dies out up/down the axial surface dip, observations that the dyke cuts the axial surface, in axis normal sections, are of no significance.

The data collected indicate that regionally the dyke margins and fold axial planes have parallel strikes, with the axial planes dipping northwestwards at 66° and the dyke margins southeastwards at 72° . This gives an intersection angle of 42° , which, from the definition above, requires a fold interlimb angle of less than 84° . Many folds have open to close interlimb angles and thus fall into the potential range of the criteria. However, wherever smaller interlimb angles have been found the angle between axial surface and dyke margin is also very much reduced, often approaching parallelism (e.g. Fig. 7D, or Roberts 1972, fig. 13). No example of a fold-dyke pair with the required angular relationship for the dyke to cut a single bedding surface on both limbs has been observed.

An alternative criterion would be to examine the orientation of the dyke-sediment contact. If the rocks were cleaved prior to intrusion, then fractures propagating in the rock during extension would split along the existing cleavage planes and a zig-zag intrusion geometry would develop, due to cleavage refraction. In the present case the high strains after emplacement have resulted in strain localization and slip along many margins; to date, no clear example of the dyke following an earlier cleavage across a fold has been found, although there is, perhaps, a hint of this in Fig. 7D. Such relationships, however, could be ambiguous, since the high strains in the softer layers might rotate a thin dyke during folding, giving the appearance of emplacement after cleavage formation; an example of this can be seen in Fig. 7C.

On a regional scale an appraisal of whether the dykes post-dated folding might be obtained from the unconformity at the base of the Løkvikfjellet Group, which is cut by the dykes in the Båtsfjord area. This unconformity has an overall westwards tilt relative to the bedding in the Barents Sea Group, such that the Løkvikfjellet Group oversteps the Barents Sea Group westwards (Siedlecki 1980), implying an essentially eastwards tilt of the North Varanger Region during the development of the uncon-

formity. In principle, detailed examination of the orientation of the bedding in the Barents Sea Group relative to the bedding in the Løkvikfjellet Group immediately above the unconformity should reveal whether the underlying rocks were folded on a large scale prior to erosion, thus implying a pre-unconformity (and thus pre-dyke) deformation, assuming the correlation of the Kongsfjord and Båtsfjord dykes. Such data have not yet been sought and are not determinable from Siedlecki (1980).

Dyke emplacement model

From the examples outlined above, it is clear that the dykes had a significant effect on the orientation of folds developing in their neighbourhood. However, whether the dykes actually *caused* the folds to develop at their margins, due, perhaps, to their differing rheology, is unclear. Folding is not ubiquitous at dyke margins and, where present, does not seem to have been more intense than elsewhere. Since the dyke margins are sub-vertical, and thus sub-parallel to the XY plane of the strain ellipsoid, progressive strike-slip deformation would not cause the dykes to rotate out of the vertical.

The model for the development of the structures is presented in Fig. 9. Fig. 9B is a schematic plan view of part of the dyke swarm, with atypical fold axial trends developed at the margins of the oblique dykes and in the sediments in the bridge structures. These are observed structures and for these there is *no doubt* that the dykes were emplaced prior to folding. In principle, other orientations of oblique and linking dykes and bridge structures would have caused other atypical fold axial trends.

In the schematic cross-section (Fig. 9A), the thin dyke has been offset by flexural-slip folding. In this case some of the folding may have occurred prior to intrusion, and, if a suitable example were found, this could be tested. Offset dykes which lie oblique to the fold axis will develop an *en echelon* pattern in a plan view (Fig. 9B).

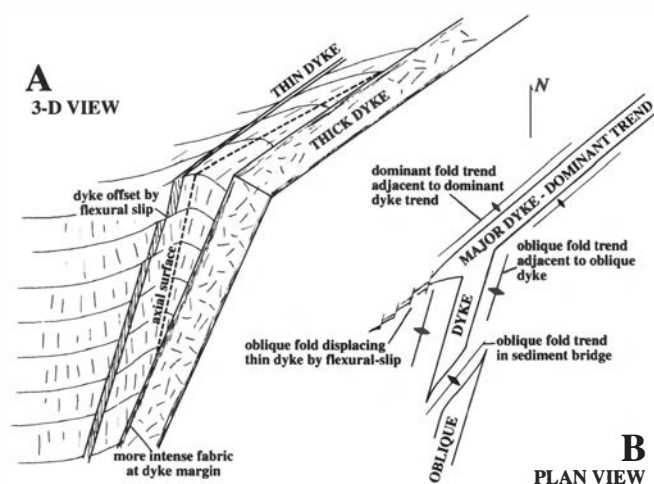


Fig. 9. Summary model for the evolution of the fold-dyke relationships. See text for details.

The thicker dyke in Fig. 9A is slightly oblique to the fold axial surface and crosses it in both the *b* and *a* directions, so that the fold effectively dies out along the dyke; this effect would be enhanced with steeper fold plunges. Such a model could account for the structure shown in Fig. 38 of Siedlecka & Roberts (1992). Roberts (pers. comm. 1993), however, argues that the presence of a fold axis on *both* sides of the dyke, and in line with each other, indicates that the fold predates the dyke. Whilst acknowledging that this is perhaps surprising, it is not proof.

Regional implications of the dykes

The ca. 650 Ma age of the Båtsfjord dykes (Beckinsale et al. 1976) is similar to the age determined for the emplacement for the MORB–WPB Ot fjället dolerites in Central Scandinavia (Claesson & Roddick 1976; Solyom et al. 1979), which suggests a late Sturtian–early Vendian reactivation of the Iapetus Rift and Timanian Aulacogen (see fig. 1 inset in Kumpulainen & Nystuen 1985). For this reason, the MORB–WPB Kongsfjord dykes are inferred

to be essentially contemporary with the Båtsfjord dykes and of ca. 650 Ma age.

Dykes with transitional MORB–WPB compositions similar to those in the Kongsfjord area have been found elsewhere in the Middle and Upper Allochthons in both the Caledonides of northern Norway (Gayer et al. 1985; Rice 1987; Roberts 1990) and in Scandinavia generally (Solyom et al. 1989; Berg & Andresen 1987; Andreasson 1987; Greiling et al. 1989), generally cutting late Riphean–Sturtian rocks, or only the lower part of the unconformable overlying Vendian tillites or younger rocks (cf. Gee 1975).

If the North Varanger Region were restored to a position at the western margin of the Lower Allochthon (Fig. 10A), which was probably the western end of the Timanian Aulacogen (cf. Gayer & Rice 1989), late Sturtian–early Vendian reactivation of the rift system surrounding Baltica is likely to have affected the Timanian Aulacogen. A variety of models can be envisaged for this reactivation, depending on the relationship of the extensional vectors within the three arms of the triple junction. Three examples are given below (Fig. 10B, C, D).

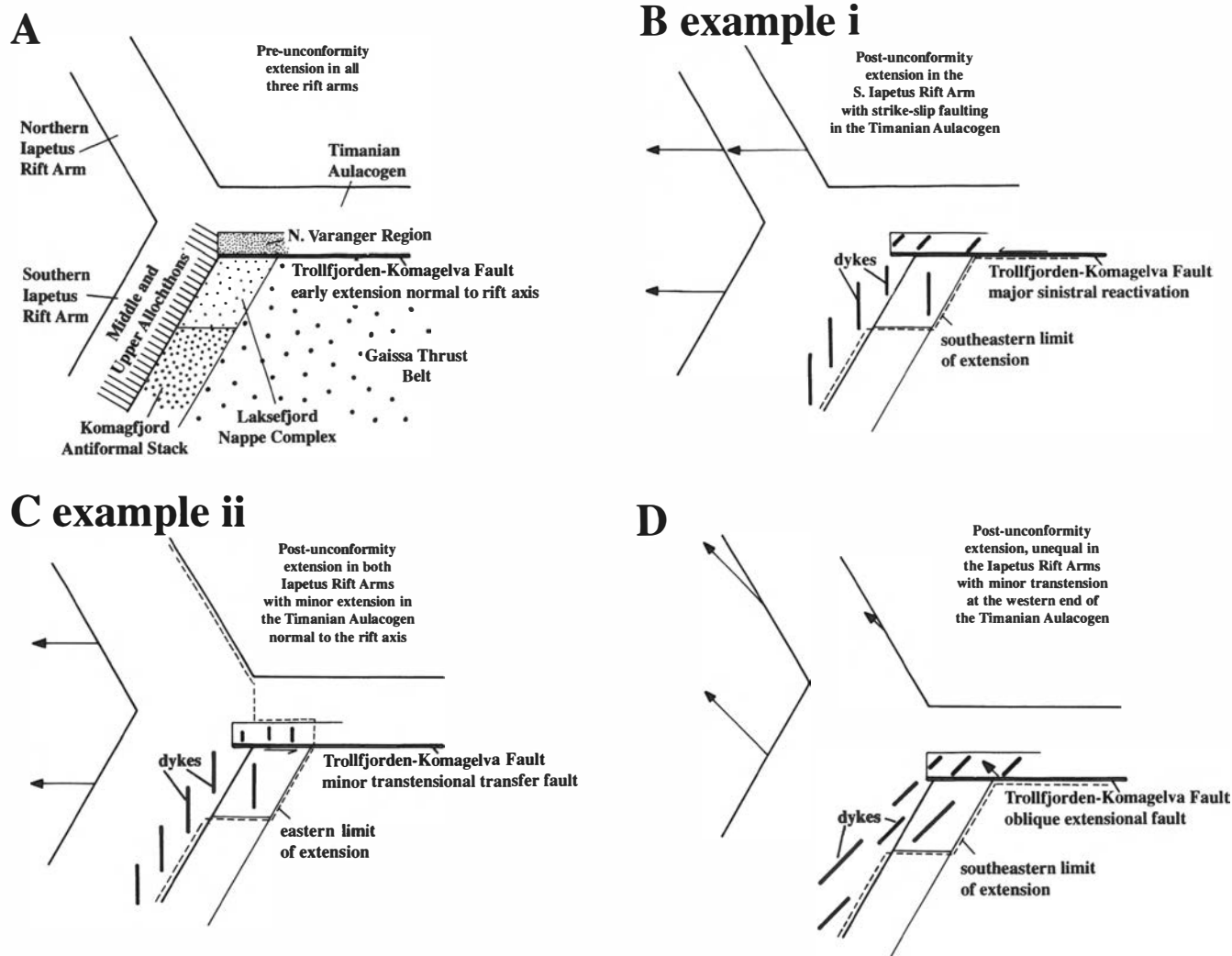


Fig. 10. Schematic model for the regional significance of the dykes. Arrows indicate schematic movement of blocks relative to Fennoscandian Shield. See text for details.

In the first example (Fig. 10B) extension occurred only in the Southern Iapetus Rift, with a vector sub-parallel to the Timanian Aulacogen. Since, in the model, no relative displacement occurs between the western sides of the northern and southern arms of the Iapetus Rift, the extension is accommodated by the development of a sinistral shear zone within the aulacogen and the Trollfjorden–Komagelva Fault is reactivated as the southern margin of this shear. As long as the extension vector is sub-parallel to the trend of the Timanian Aulacogen, sinistral transcurrent with minor transtensional to transpressional deformation occurs, with dyke emplacement in the North Varanger Region at ca. 45° to the Trollfjorden–Komagelva Fault.

The WSW–ENE orientation of the dykes in Kongsfjord and Båtsfjord is consistent with sinistral strike-slip parallel to the Trollfjorden–Komagelva Fault. Subsequent reversal of the strike-slip movement vector would cause extension along the dykes and folding, with axes sub-parallel to the earlier dykes, as long as the relative amounts of transpression and transtension, with consequent rotation of the strain ellipsoid (Sanderson & Marchini 1984), remained the same. Superficially, this argument can be used to account for the observed sub-parallelism of the folds and dykes in the North Varanger Region. However, this model fails to explain why the intense dextral strike-slip deformation did not also lead to intrusion of mafic dykes, since intrusion occurred during an intervening (or later) extension event and, further, why no associated NNW–SSE oriented compressional structures formed during the emplacement of the Kongsfjord dykes.

A simpler explanation is to infer that the dyke emplacement was not related to a major strike-slip deformation, but occurred earlier. This is shown in Fig. 10C, where the same extension vectors are modelled in both of the Iapetus Rifts, again parallel to the trend of the Timanian Aulacogen. Owing to the thinness of the lithosphere within the aulacogen, extension at the eastern margin of the Southern Iapetus Rift propagates into the Timanian Aulacogen, pulling the North Varanger Region out. This results in dyke emplacement in an orientation normal to the aulacogen margin and reactivation of the Trollfjorden–Komagelva Fault as a minor sinistral transfer fault west of the eastern limit of extension into the aulacogen, allowing the central part of the rift to be extended normal to the rift axis, whilst the rift flanks are unaffected.

However, in this model the dyke orientation, relative to the Trollfjorden–Komagelva Fault, is incorrect. Changing the extension direction to lie at 45° to the trend of the Timanian Aulacogen gives the correct dyke orientation and one which is *suggestive* of emplacement during major sinistral strike-slip movement (Fig. 10D). In this model, the Trollfjorden–Komagelva Fault would have been reactivated as an oblique slip transfer fault, with the significant extension causing further subsidence within the aulacogen resulting in the deposition of the

Løkvikfjellet Group. The oblique displacement would have had a minor sinistral movement. This reactivation could have been restricted in the area west of the Båtsfjord dyke swarm, if the western end of the Timanian Aulacogen were extended, or could have extended along its whole length, if a more regional extension across the aulacogen occurred.

Conclusion

There is striking evidence that some fold axial directions in the Kongsfjord Formation were controlled by the orientation of either dykes with atypical trends or bridge structures, so that these dykes were definitely emplaced pre-folding. Other dykes predated all or at least most of the folding, since they are significantly offset by flexural-slip folding. Nevertheless, just as Roberts (1975) stated that some dykes may have been emplaced pre-D1, so it should be noted here, but in the opposite way, that some dykes *may* have been emplaced broadly syn-D1, although there is no unequivocal evidence for this as yet. Finding such proof is likely to be very difficult, given the general parallelism of the fold axial surfaces and dyke margins. Unlike the model of Roberts (1975) and Siedlecka & Roberts (1992), which is based on essentially unsupported statements, the model proposed here is both documented and open to testing.

From correlations with the Båtsfjord dykes, which have been dated with both radiometric methods (Beckinsale et al. 1976) and palaeontological data (Siedlecki & Levell 1978; Vidal & Siedlecka 1983), and from comparisons with the geochemistry of similar dykes elsewhere in Scandinavia, the Kongsfjord dykes are thought to have been emplaced at some time near the Sturtian–Vendian boundary. The extensional stress leading to dyke intrusion was most probably caused by re-activation of the Iapetus–Timanian rift system, as recorded elsewhere throughout the Scandinavian Caledonides, with extension at ca. 45° to the trend of the Timanian Aulacogen. This was followed by dextral strike-slip movement along the Trollfjorden–Komagelva Fault, with associated folding and cleavage development.

Acknowledgements. – We thank Arild and Jorunn Pettersen of Tana Bru and Åge and Solveig Olsen of Veines for hospitality whilst doing fieldwork. We also thank Chris Townsend and Julie Jones for discussions; Charlie Offenbecher, Christa Hofmann and Rafael Herrgård for assistance in the field; and Rod Gayer, David Roberts and Roy Gabrielsen for reviewing the manuscript.

Manuscript received February 1994

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