Journal of Structural Geology, Vol. 14, No. 10, pp. 1173 to 1192, 1992
Printed in Great Britain

0191-8141/92 $05.00+0.00

Morphology and geometry of duplexes formed during flexural-slip folding

P. W. GEOFF TANNER

Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, U.K.

(*Received* 15 *May* 1991; *accepted in revised form* 29 *April* 1992)

Abstract—Bedding-parallel slip on the limbs of chevron-style folds during flexural-slip folding results occasionally in the formation of small-scale duplexes. The duplexes described here from Upper Carboniferous turbidites in SW England generally range in thickness from 0.3 to 30 cm, and are up to 2.4 m long. They have smooth nearly flat roofs, and both the internal thrusts and the slickenfibres on them are commonly oblique to the mean transport direction as indicated by slickenfibres on the floor and roof thrusts. The morphology of the duplexes suggests that they developed between active floor and roof thrusts and they show a characteristic lack of hangingwall anticlines on the link thrusts. Their shear sense always corresponds to that required by flexural slip, and the total bedding-parallel displacement across each is commensurate with that predicted by the flexural-slip model. All thrust surfaces in the duplexes are marked by quartz fibre veins (a feature which distinguishes them from soft-sediment and other pre-folding duplexes) and carry slickenfibres whose mean orientation and complex variations in slip direction on different fibre sheets on the same slip surface are identical to those on flexural-slip movement horizons from the same fold limb. The duplexes form either as a result of resistance to thrust propagation by local facies or bed thickness changes, or develop as transfer structures between the tips of movement horizons propagating along adjacent slip surfaces. Late-stage duplexes develop from imbricated fibre veins and also form on slip surfaces oblique to the axial planes of major folds.

INTRODUCTION

DUPLEXES consist of detached slices of rock, called horses, stacked in a systematic manner between a floor fault and a roof fault (Dahlstrom 1970, Boyer & Elliott 1982) (Fig. 1), and can form in both extensional and contractional environments. Contractional duplexes analogous to those reported here, but generally on a much larger scale, have been described from many thrust belts (Boyer & Elliott 1982, Mitra 1986, and references therein) and are also inferred to occur on a crustal scale (Hatcher & Hooper 1992). They are scale-invariant structures whose dimensions are controlled primarily by the thickness of the bed or stratigraphic unit which has been imbricated. The generally accepted model for their formation is that they are forward-propagating structures in which the development of each new thrust slice (horse) is accompanied by the backward rotation and 'piggy-back' transport of the earlier-formed horses. Parameters which possibly determine the final geometry of the duplex include the ramp angle, the initial and final spacing of the thrusts, and the amount of displacement on them (Mitra 1986, Cruikshank *et al.* 1989). However, comparison of the morphologies of well-documented and completely preserved contractional duplexes with that predicted by the classical duplex model (Tanner 1992) (Fig. 1) shows that there are important differences which require explanation.

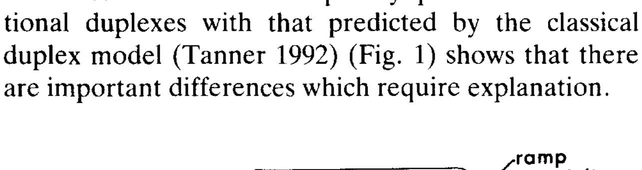

Fig. 1. The basic forward-propagating contractional duplex model, after Boyer & Elliott (1982).

Duplex formation is initiated when the forward propagation of a thrust is impeded by some perturbation or sticking point (Knipe 1985), or by layer-parallel shortening of a bed or horizon in advance of the thrust tip (Cooper *et al.* 1983, Nickelsen 1986). The thrust ramps up to a higher glide horizon and subsequent erosion of the footwall ramp by the formation of a series of imbricated slices creates a duplex structure. Duplexes also develop in systems of stepped faults when displacement is transferred from one fault segment to another by means of a transfer zone (Woodcock & Fischer 1986, Aydin 1988, Cruikshank *et al.* 1991).

Considering how duplexes form, it is entirely predictable that they should develop during flexural-slip, in which packets of welded beds to over a m thick slip over one another during the folding of regularly bedded rock sequences such as turbidites, to form chevron-style folds (Tanner 1989). As individual slip surfaces, or movement horizons, are constrained to a single fold limb (and form between different beds or units on adjoining fold limbs), each surface is only required to be planar and able to sustain bedding-parallel slip without hindrance for the length of one limb, which in the rocks studied here is 10–200 m. In regularly-bedded sequences such as distal turbidites, where beds maintain constant thickness and lithology over large distances, this condition is readily met and duplexes do not appear to develop. For example, such duplexes have not been reported from regularly bedded turbidites found on the northeast coast of South Georgia, South Atlantic (Tanner 1989) and are rare in similar rocks of the Cardigan Bay area, Wales (Tanner 1990). However in more proximal or shallow-water turbidites where beds change thickness and/or facies in relatively short distances, or are affected by large-scale soft-sediment structures, some movement horizons will meet an obstacle to their further propagation even within the length of a single fold limb and be forced to transfer displacement to a higher glide plane. This can lead to the development of a great variety of structures such as ramps, blind thrusts, imbricate structures and duplexes. Also, if adjacent movement horizons initiate with an offset configuration during fold growth and amplification, then the resultant overlap of these bedding-parallel fault segments may give rise to linking structures with a duplex-like geometry such as occur in strike-slip fault systems (Aydin 1988, Cruikshank *et al.* 1991).

The examples described here are from well-documented coastal sections in the Crackington and Bude formations of the Culm Measures in North Devon and North Cornwall (hereafter referred to as 'North Devon') (Fig. 2). These rocks were deposited in a foreland basin in Upper Carboniferous times and deformed during the northward migration of Variscan deformation across the area at the end of the Carboniferous (Seago & Chapman 1988, and references therein). The base of the Bude Formation is taken at the top of the Hartland Quay Shale (Freshney *et al.* 1979) and this formation was deposited in a shallow-water lake-shelf (Higgs 1987) or subsea fan (Melvin 1986, 1987) environment which contrasts with the deeper water, more distal, marine facies of the Crackington Formation.

Both formations are affected by upright, open, gently plunging or horizontal chevron folds which have been progressively modified south of Hartland Quay by southward-directed simple shear (Sanderson 1980), to become tight and recumbent at Millook (Fig. 2). The early folds are also affected by numerous northward-dipping normal faults south of Bude (Freshney *et al.* 1972). The main study area at Hartland Quay was chosen as it is superbly exposed and is in rocks where the upright chevron folds are unaffected by later deformation. This is also largely true for the Maer High Cliff section farther south (Fig. 2), but at Bude and Widemouth Bay the effects of the later deformation become noticeable.

In the Bude area, Whalley & Lloyd (1986) reported evidence for northward-directed thrusting (associated with the development of an antiformal imbricate stack) having occurred *prior* to the development of the upright

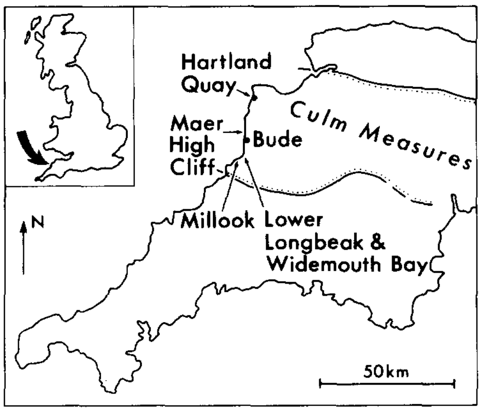

Fig. 2. Location of the main study areas in SW England.

major folds; other pre-folding contractional structures include small-scale imbricate structures found at Maer Cliff which were interpreted by Enfield *et al.* (1985) to have formed during soft-sediment deformation. However, duplexes figured by Price & Cosgrove (1990, fig. 7.25) from Maer Cliff and Lower Longbeak, north of Widemouth Bay were inferred to have resulted from regional thrusting, and Tanner (1989) reported examples of duplexes thought to have formed during flexural-slip folding.

This paper is concerned with carefully documenting the morphology and geometry of small-scale duplexes, mainly of flexural-slip origin, from North Devon. As the coastal sections in this area are subject to very active marine erosion—one duplex illustrated previously (Tanner 1989, fig. 6c) has already been destroyed in this way and others are being actively dissected—such documentation is essential. In view of the conflicting interpretations which have been proposed for some of these structures, emphasis is given to establishing criteria by which duplexes formed as a result of flexural-slip folding may be distinguished from those formed during: (a) soft-sediment deformation; (b) layer-parallel shortening prior to folding; or (c) regional thrusting not directly associated with flexural-slip displacements on the limbs of folds. The terminology used in this paper is shown in Fig. 3; the term 'link thrust' is used to describe an imbricate thrust which links the floor and roof thrusts of a duplex (cf. McClay & Insley 1986).

MORPHOLOGY

Pre-folding duplexes

Relatively few examples of duplexes formed during soft-sediment deformation or layer-parallel shortening prior to folding have been found in the main study areas around Hartland Quay and Maer High Cliff but a number occur elsewhere in the Bude Formation. Structures of soft-sediment origin might be expected considering the abundant evidence of sedimentary slumping and dewatering which has been reported from the Bude Formation (Burne 1970, Enfield *et al.* 1985, Whalley & Lloyd 1986). The early duplexes formed in rocks which had been lithified but may not have been deeply buried at the time. They are highly irregular structures with much fracturing of beds on all scales, may show equally well-developed back- and fore-thrusts and related folds, and, most significantly, lack quartz fibre veins on the thrust planes.

A good example of this type of structure is the 9 m-long duplex found south of Hartland Point lighthouse and just north of the Cow & Calf (SS 228272), which has an irregular, hummocky roof, shows much disruption of beds, and lacks continuous fibre veins along any of the thrust surfaces (Fig. 4). Sedimentary slump folds with wavelengths of up to 0.5 m occur in nearby rocks on the same fold limb.

A well-known imbricate fan structure (not strictly a

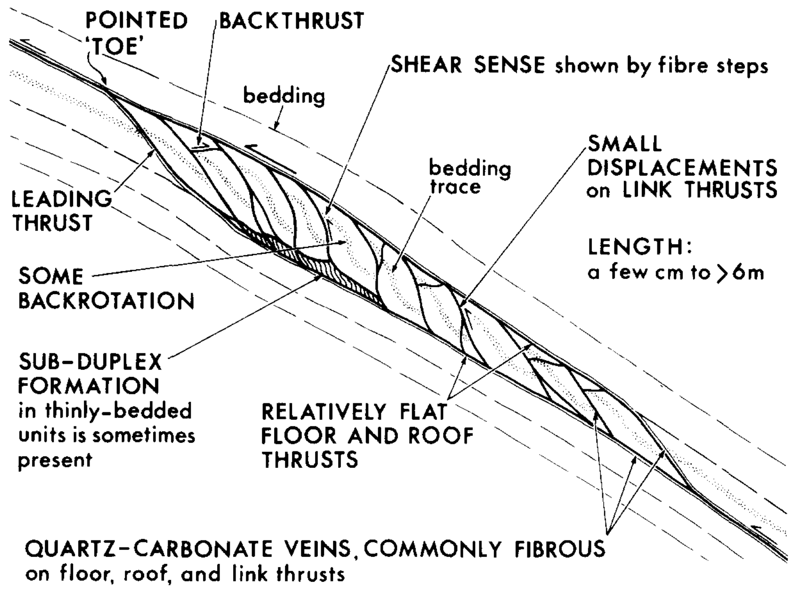

Fig. 3. General features of duplexes of flexural-slip origin. Note the smooth roof and streamlined shape which contrast with those of the classical Boyer & Elliott (1982) model (Fig. 1).

duplex as it lacks a roof thrust) found 20 m north of the steps to the beach at Sandy Mouth (SS 202100, 4 km north of Bude) may have also originated in this way. It can be traced for 3–4 m above beach level where it passes upwards into a disturbed zone (containing both folds and faults) which is oblique to bedding. The imbricate fan (Fig. 5) affects a distinctive 3–4 cm thick unit which is thickened to over 8 cm in the imbricate zone. The unit (X on Fig. 5) consists of grey shale with ripple-drift laminations at the base and a thin pale sandstone bed with regularly-spaced load balls at the top. A pair of earlier extensional faults are preserved in the footwall to the imbricate, and the northerly one (which is now marked by a thin, late cross-fibre quartz vein which also cuts the floor thrust of the imbricate fan) may have behaved as a syn-sedimentary fault during the deposition of the black shale. Figure 5(b) shows the hanging wall cut-off above one of the imbricate thrusts in the fan, and the footwall structure, drawn from a thin section of a loose *in situ* fragment. When the thin competent layers (in black) are traced around the hanging-wall structure it can be seen that different parts of the same layer have undergone extension or contraction, and are displaced by many small fractures of limited extent. The imbricate fan has a typically disturbed internal structure which is

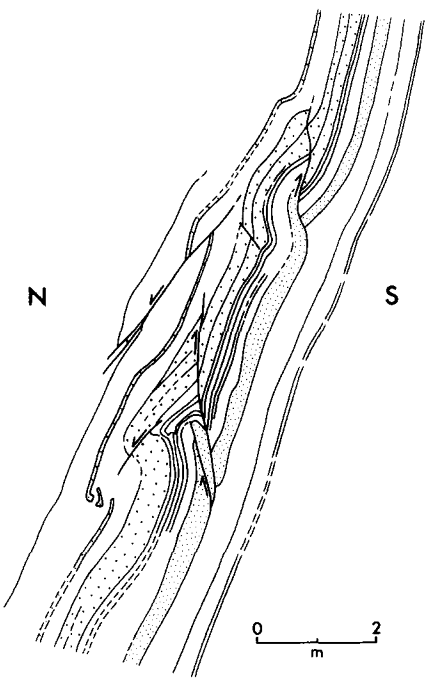

Fig. 4. A duplex of pre-folding age exposed in a steep cliff face 0.6 km south of Hartland Point.

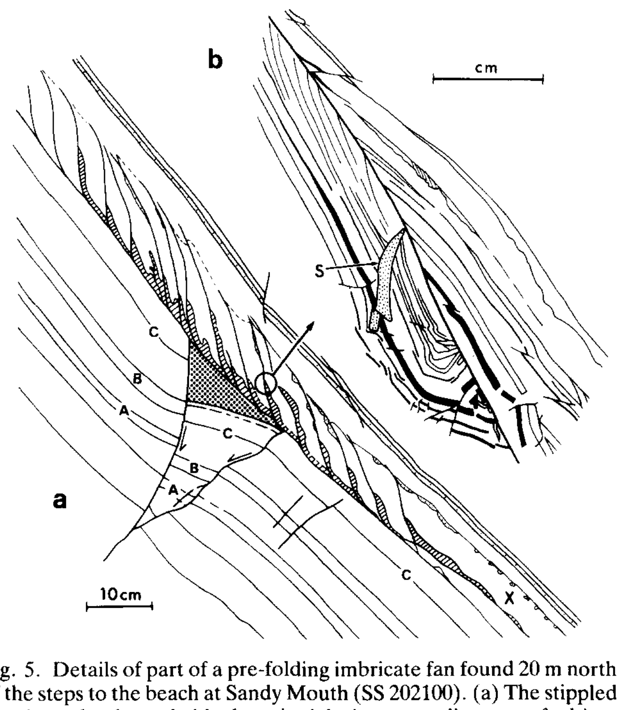

Fig. 5. Details of part of a pre-folding imbricate fan found 20 m north of the steps to the beach at Sandy Mouth (SS 202100). (a) The stippled area is mudrock, probably deposited during syn-sedimentary faulting. Unit X has a cross-laminated bed at the base (ruled ornament), and a thin sandstone bed with load structures at the top (stippled). A–C are beds which can be traced laterally. (b) Details of the circled area in (a) drawn from a thin section. S, sandstone dykelet.

quite unlike the highly ordered and coherent internal structure of the duplexes which formed synchronous with folding (cf. Figs. 11d and 12c) and it shows the type of small-scale brecciation which exemplifies the pre-folding duplexes. No fibrous quartz veins are developed on the link thrusts. The time of injection of the sandstone dykelet in Fig. 5(b) is unknown but it may have formed during the thrusting. During flexural-slip deformation these early duplexes are preserved within welded sedimentary units, with slip taking take place preferentially upon adjoining planar surfaces, but some movement may occur locally on any surfaces within the duplex that are suitably oriented. Thus a mm-thick veneer of quartz fibres seen intermittently on the basal thrust of the imbricate fan probably formed as a result of very limited flexural-slip movement on this planar surface, and a similar fibrous veinlet is seen on a single thrust surface in the lighthouse duplex described earlier (Fig. 4).

Carefully documented studies of minor structures from compressional zones in soft-sediment slumps and slides (Farrell 1984, Martinsen & Bakken 1990) show that the minor thrusts in imbricate structures from these zones are accompanied by soft-sediment deformation, including the injection of sand along minor faults. The thrusts are marked in some cases by sheared mud horizons but no associated quartz fibre growth or veining has been reported.

Duplexes of flexural-slip origin

A total of 27 duplexes of this type have been identified: 20 from the Hartland Quay area (18 to the north of the Quay (Fig. 6) and two to the south), and one imbricate structure; six from Maer High Cliff (Fig. 7); and one each from south of Compass Point, Bude and from Lower Longbeak, north of Widemouth Bay (Fig. 8).

They vary in length from a few centimetres to several decimetres, with the largest (Fig. 8) being 90 cm across and over 6 m long, and all are flat- or smooth-roofed. The duplexes involve either single bands of slate <3 cm thick; interbanded siltstone and slate laminae on a mm- to cm-scale; single sandstone beds (6–25 cm thick); or thicker packets of sandstone beds with intervening slate or siltstone (Fig. 9). Figures 10–13 illustrate both the morphology and geometry of these different types of duplex. Black-and-white photographs were taken of all of the duplexes, as near as possible at right angles to the mean transport direction shown by slickenfibre lineations on the roof and floor thrusts, and large prints were used in the field to record the finer details of each structure. This procedure was especially useful in the case of the Lower Longbeak duplex (Fig. 8). As the major folds have near-horizontal E–W axes, the steep N–S-trending cliff faces and ac joints provide approximate profile sections of most duplexes.

General features of the different types of duplex are summarized first before describing their geometry and mode of development in more detail. The generalized profile of these duplexes seen in a section normal to the mean transport direction is shown in Fig. 3: this profile is common to all duplexes regardless of the rock type in which they have developed and is exemplified by the slate duplex shown in Fig. 10(d). In no case is the unit which has been imbricated present as a continuous layer above the roof thrust as in the Boyer & Elliott (1982) model (Fig. 1).

Most duplexes arise from a simple transfer of slip from one movement horizon to a higher one over a short distance, but in a few cases a duplex has formed locally between two approximately parallel movement horizons which can be traced for a relatively long distance away from it. An example of the latter is shown in Fig. 9(d) where a movement horizon found near beach level splits into two (locally three) slip surfaces which are 4–8 cm apart and can be followed obliquely across the cliff face for some 8 m before a duplex (No. 5 in Fig. 7; illustrated in Tanner 1992, fig. 6a) develops at the tip of the structure, and slip is finally transferred to the upper movement horizon. The duplex is 2.4 m long and has a maximum thickness of 13 cm. Seven widely-spaced ramps occur in the 8 m section, and at one place a small thrust transfers some of the bedding-parallel displacement to the hangingwall packet above the topmost movement horizon.

Slate duplexes are miniature duplexes 0.3–3.0 cm. thick and a few decimetres long which occur in thin bands of mudrock between sandstone or siltstone beds. They consist of an array of closely spaced sigmoidal horses of slate sandwiched between floor and roof thrusts marked by quartz fibre veins (Figs. 9a and 10d) (Tanner 1989, fig. 6c). The thrust surfaces between the horses are generally highly polished, are rarely marked by fibre veins, but occasionally carry a fine striation lineation. Some slate duplexes pass laterally into massive or fibrous quartz veins, or to a set of en échelon quartz veins which indicate the same sense of shear as the duplex (Fig. 9a). Most contain thin ribs of siltstone in the individual horses and the duplexes appear to be small-scale analogues of the sandstone and siltstone duplexes described below, a transitional type being illustrated in Fig. 11(d). The pale sandstone rib seen in the main duplex is locally imbricated to form a sub-duplex up to 2 mm thick (Fig. 11d-1) and a thinner band forms a mm-thick supra-duplex above one of the sandstone horses (d-2). As there is no direct evidence that they formed as a result of layer-parallel shortening of the mudstone layers in advance of the propagating tip of a movement horizon and prior to the development of the link thrusts (a 'cleavage duplex' as defined by Nickelsen 1986), the non-genetic term 'slate duplex' is used to describe these structures. They closely resemble the duplexes described by Bosworth (1984) from black shales of the Appalachian Plateau but are an order of magnitude smaller in size. Some may post-date cleavage formation and represent late-stage adjustments on the limbs of the folds.

Isolated ramps with a veneer of fibrous quartz are found where a movement horizon (Tanner 1989)

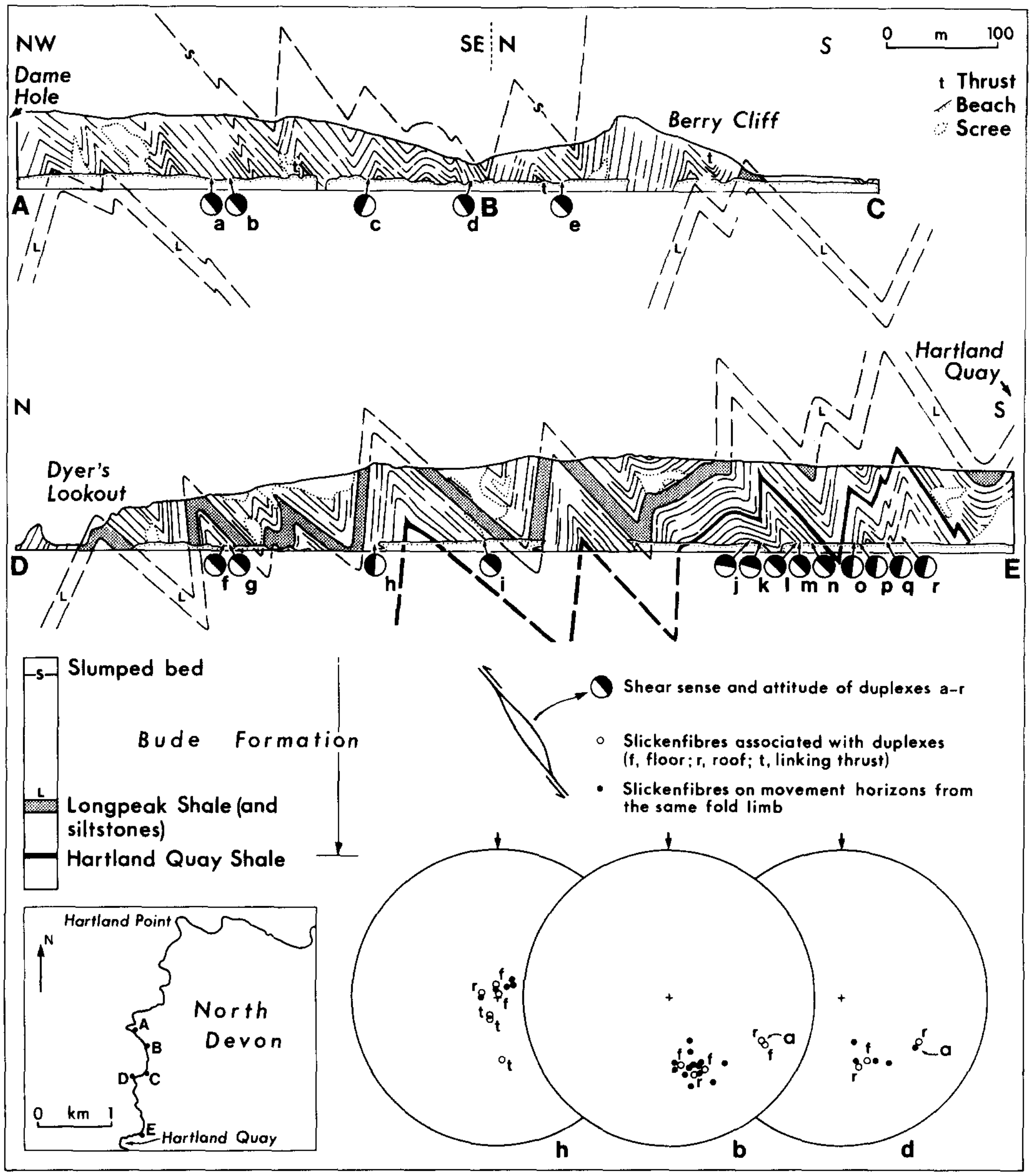

Fig. 6. Location and shear sense of duplexes **a–r** shown on a true-scale, vertical cross-sections drawn normal to chevron fold axes north of Hartland Quay, North Devon (see inset for lines of sections). The orientation of slickenfibres from duplexes **h**, **b** and **d** (open circles) is shown on equal-area, lower-hemisphere projections and compared in each case with that of slickenfibres found on movement horizons from the same fold limb (closed circles). a, slickenfibres oblique to the main set.

changes level from the base to the top of a bed on a fold limb. Sandstone duplexes develop from the imbrication of single sandstone beds, and are the most common form of flexural-slip duplex. Figure 9(c) shows a large sandstone duplex whose floor thrust forms a near vertical, 25 m-long face which has a prominent slickenfibre lineation plunging at >70°. Other examples are given in Figs. 10(a)–(c) and 12(d) & (e) and in Tanner (1992, fig. 4c).

Multilayer duplexes formed by the imbrication of packets of beds are the most suitable for detailed study as bedding traces, and sometimes a stratigraphy, can be followed through the structure (Figs. 11a–c and 12a–c). The best example of this type is the Lower Longbeak duplex in which two beds 0.7–1.0 cm thick and 14–18 cm apart (A and B, Fig. 8), which weather to a chocolate colour, can be correlated from horse to horse and define the internal structure. Imbricate fans (Fig. 9e) are rare. Stacked duplexes are structures in which duplexes on two or more different scales occur together, the size of each being proportional to the thicknesses of the constituent layers. Slate duplexes, or similar structures made up of inter-laminated slate and thin sandstone beds, commonly form a sub-duplex (Figs. 8 and 11d) (Tanner 1992, fig. 6) and in two instances, a supra-duplex (Figs. 11d and 12a–c).

Morphology of the toe region. Knowledge of the precise geometry of the 'toe' of a duplex (Fig. 3) is critical to an understanding of how the whole structure has formed. Unfortunately most duplexes do not show the fine lithological variations within each horse that enable the internal geometry of this part of the structure to be analysed. In slate and sandstone duplexes the leading thrust (Fig. 3) is invariably at a low angle to the

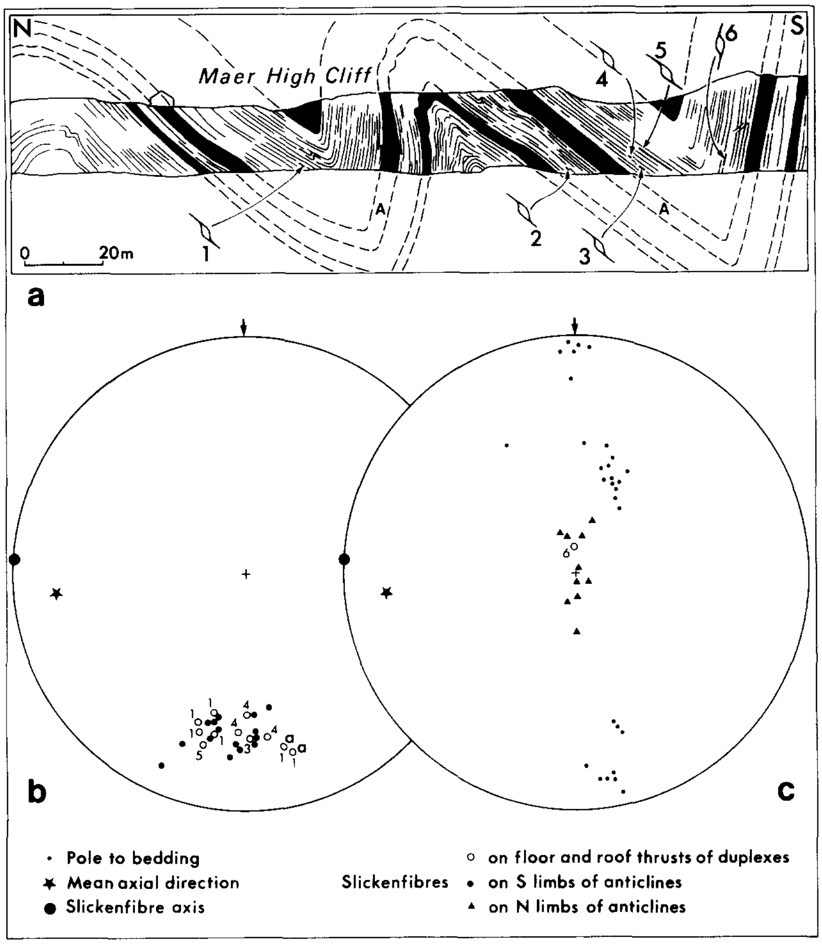

Fig. 7. Geometry and shear sense of duplexes found at Maer High Cliff, 2.5 km north-northwest of Bude (Fig. 2) (SS 201083). (a) True-scale cross-section prepared from a photo-mosaic. The folds face upwards and shale horizons are shown in solid black. (b) Orientations of slickenfibres associated with duplexes **1**, **3**, **4** and **5**, compared with those of slickenfibres from the two S-dipping fold limbs in (a). a, slickenfibres oblique to the main set. (c) Orientations of slickenfibres on the floor thrust of duplex **6** compared with those from movement horizons on the two steep, N-dipping fold limbs in (a). The poles to bedding are from measurements on all four fold limbs and were used to compute the mean axial direction. The slickenfibre axis (Tanner 1989) is the computed normal to the plane containing all movement horizon slickenfibre lineations. (b) & (c) are equal-area, lower-hemisphere projections.

roof thrust, and it appears that the sigmoidal shape of the earlier-formed horses has resulted from their having been affected by rotation and layer-parallel shortening during the progressive accretion of low-angle slices at the toe. There is no evidence that a hangingwall anticline was developed above the leading thrust as in the classical Rich (1934) model.

Critical evidence in this regard is given by duplex **6** at Maer High Cliff (Figs. 7 and 12a–c). Figure 12(b) shows a near-vertical section of the duplex parallel to the transport direction shown by slickenfibres on the floor thrust, photographed in September 1989; Fig. 12(c) shows a drawing of the toe region of the duplex made in August 1990 when the storm beach was 2 m higher and the uppermost part of the duplex had to be excavated in order to examine it. An exposed 70 cm section of the duplex, which affects a 4 cm-thick laminated sandstone–siltstone sequence, contains 24 horses, and correlation of a sandstone doublet between horses shows that there is a thrust displacement of 1–6 cm across each link thrust in the section. Note that the apparent hangingwall fold in Fig. 12(a) is due to a change in level of the rock surface. The most significant features of this duplex are that it has a low angle, slightly listric, leading thrust and that each horse and its bounding thrusts (marked by fibre veins, as are the floor and roof thrusts) was progressively backtilted between the (?active) floor and roof thrusts as new horses were accreted at the toe. This backtilting, accompanied by some bedding-parallel shortening as the horses rotated between the bounding thrusts, gives the misleading impression that shallow hangingwall anticlines had developed above each link thrust. An unusual feature is the presence of a supra-duplex (Figs. 12a & c), 2 cm thick and consisting of >30 horses, above the propagating tip of the main structure.

The roof thrust of duplex **6**, which continues up-dip as a bedding-parallel fibrous quartz vein, occurs 2–4 cm below the extremely lobate base of a 20 cm-thick sandstone bed containing rip-up mudstone clasts (Fig. 12b). This is precisely the position below a thick bed with a welded base at which flexural-slip movement horizons are most commonly found (Tanner 1989, p. 644).

Three-dimensional geometry. Most duplexes are only seen in cross-section but a few more deeply eroded sections show that they are pod-shaped to cylindrical in

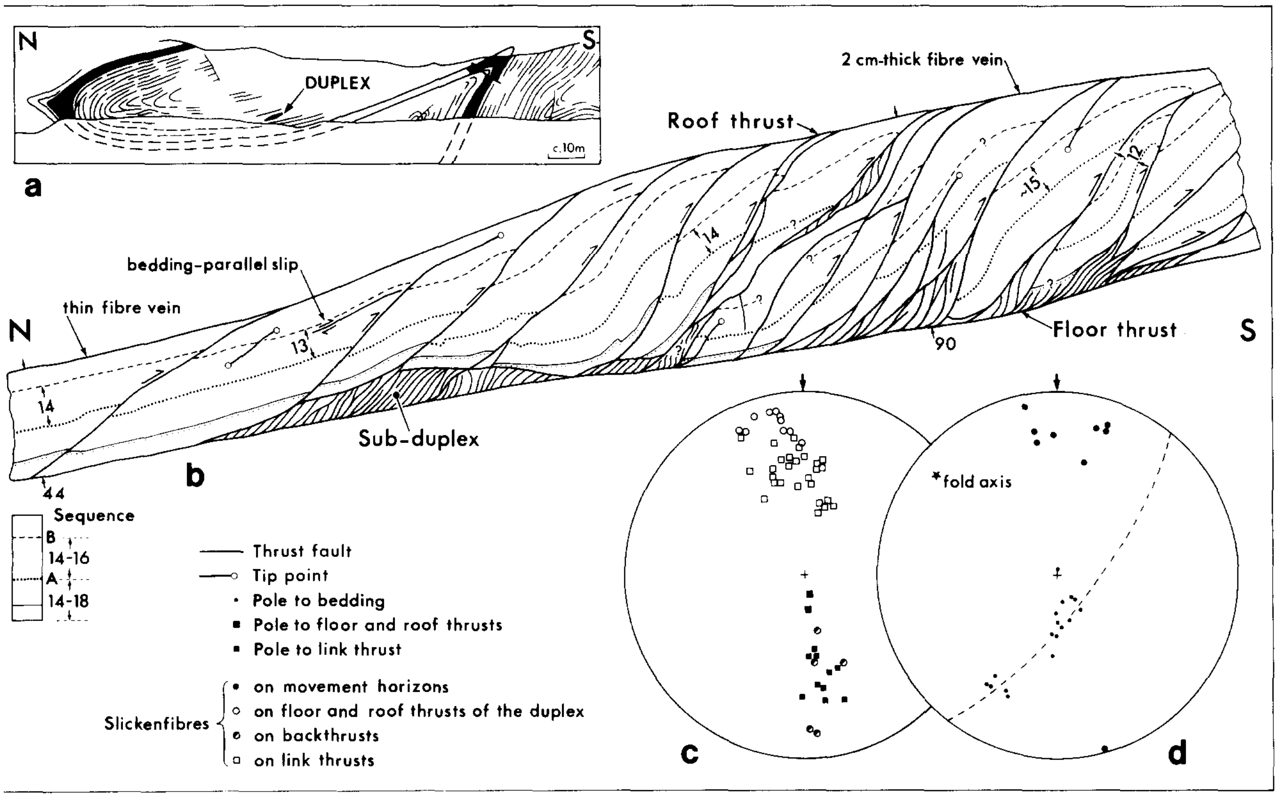

Fig. 8. Morphology and geometry of the Lower Longbeak duplex, 1.3 km north-northwest of Widemouth Bay (SS 198035). (a) True-scale cross-section showing the location of the duplex on the northern, right-way-up limb of an anticline. (b) Internal morphology of the entire exposed portion of the duplex. All measurements in centimetres. (c) Orientations of the bounding and internal thrust surfaces, and of the slickenfibres on them. (d) Orientations of the slickenfibres on movement horizons from the same fold limb as the duplex, and of poles to bedding from both limbs of the anticline. (c) & (d) are equal-area, lower-hemisphere projections.

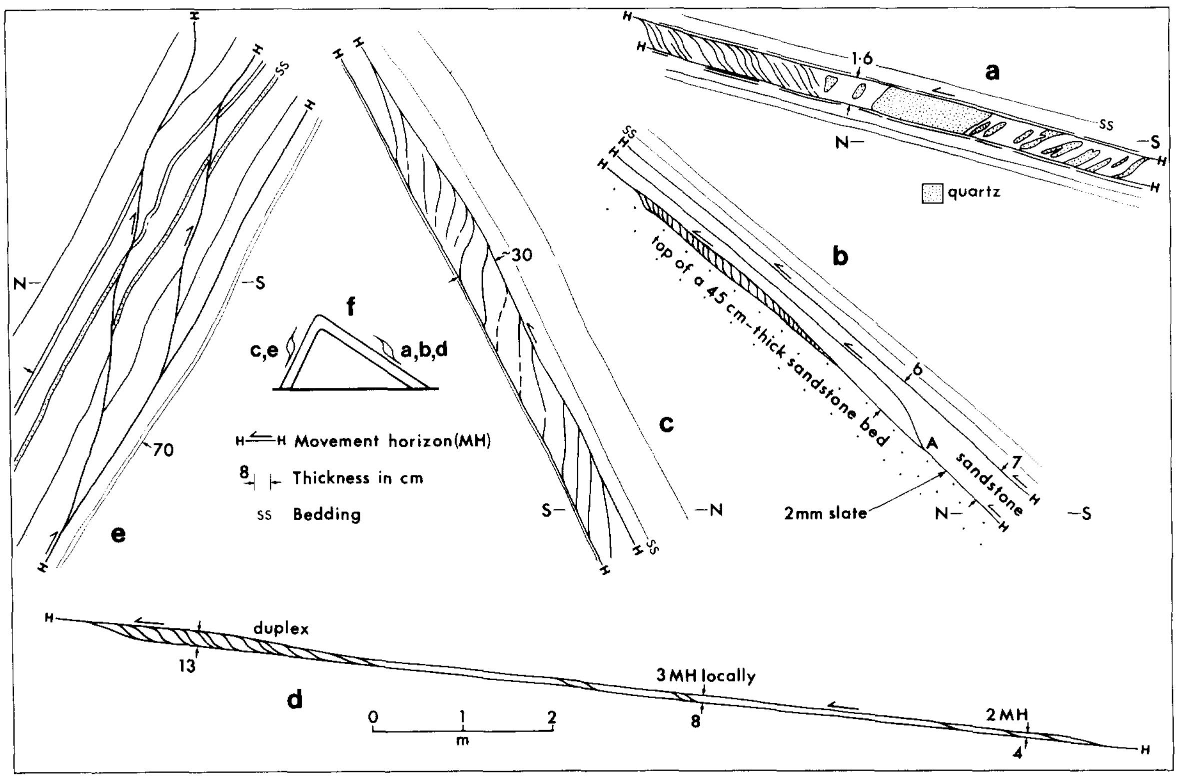

Fig. 9. Some examples of duplexes from the Hartland Quay area, shown in their correct relative orientations. (a) Slate duplex **b** (Fig. 6). (b) Slate duplex **d** (Fig. 6) and bifurcating movement horizon at A. (c) Sandstone duplex south of Hartland Quay at Speke's Mill Mouth (SS 224231). (d) The relationship of duplex **5**, Maer High Cliff, to contiguous movement horizons. (e) Imbricate structure at **c** (Fig. 6). (f) Schematic diagram showing the relative positions of the imbricate structures with reference to an hypothetical anticline.

three dimensions. For example, the sandstone duplex in Figs. 10(a)–(c) is seen in sections both parallel with, and normal to, the movement direction and in three dimensions is a pod-like body whose longest dimension is parallel to the major fold axis.

Two sections at right angles through a duplex or duplexes developed along the same movement horizon are seen at locality **h** (Fig. 6). That exposed in plan view on the rock platform (X in Fig. 10f) represents a section near normal to the transport direction, as the slickenfibres on both floor and roof thrusts plunge at >70°. Figure 10(e) shows an oblique view of this duplex as seen from position Y on Fig. 10(f). Some 3 m east along the same movement horizon the profile of a duplex (see Tanner 1989, fig. 6b) eroded parallel to the movement direction is seen on a near vertical rock face at Z. If X and Z are sections through the same duplex, then it has the form of a cylindrical body whose long axis plunges at a low angle to the west, parallel to adjacent major fold hinges.

Amount of displacement. The amount of bedding-parallel slip represented by the total displacement on the duplexes described here is difficult to quantify due to a general lack of marker horizons which can be correlated from horse to horse within each duplex, and to appreciable variations in displacement vector between the various link thrusts. The Lower Longbeak duplex is the best exposed of these structures but although the profile (Fig. 8b) was prepared in the field using large prints of photographs taken as near as possible at right angles to the mean transport direction there were perspective and scaling problems in drawing the final section. As horizons A and B can be traced with reasonable certainty throughout the structure these were used to make a line-balanced restoration; due to the problem mentioned above and that of correlating beds below A across the structure (Fig. 8b), an area balance was not attempted. From line-balanced restorations the total exposed portion of the Lower Longbeak duplex (Fig. 5) gives an approximate shortening of −41% and a 50 cm portion of duplex **6** (Fig. 4) gives a value of −46%. Other duplexes give comparable though less precise results. The −46% shortening is equivalent to 0.84 m of slip per *present* metre length of duplex and would give displacements of 0.92, 0.84, 0.45, 2.0 and 0.59 m for the Maer High Cliff duplexes **1**, **3**, **4**, **5** and **6**, respectively. The mean spacing of movement horizons from a single measured section near Hartland Quay is 0.78 m and the largest spacing is 1.65 m (Tanner 1989). Using the flexural-slip model of Ramsay (1974), where slip on a surface is equivalent to the thickness of the adjacent layer times the tangent of the angle of dip of the fold limb, a chevron fold with an interlimb angle of 60° (as seen at Maer High Cliff, Fig. 7) would generate 1.35 and 2.85 m total slip for each of these spacings, respectively. Thus the amounts of slip shown by the North Devon duplexes are comparable to those predicted by the flexural-slip model, even allowing for the facts that duplex formation may not have taken place throughout the entire history of fold development and that the shortening on each may vary between 40 and 60%. On South Georgia, South Atlantic, sedimentary dykes displaced by bedding-parallel flexural-slip movements on the limbs of close folds show displacements (measured oblique to the slip direction and therefore minimum values) of up to 2.7 m on individual surfaces (Tanner 1989).

The most difficult case to explain is that of the Lower Longbeak duplex for which the calculated total displacement on the presently exposed portion is 5 m, and for the whole structure would be of the order of 7–8 m. This amount of slip would take place at the boundary of a welded packet of beds about 4.6 m thick on the limb of a chevron fold with an interlimb angle of 60°. The large amount of slip represented by the duplex could be explained by its location above a sequence of massive sandstone beds which show little evidence of inter-bed slip. Release of this accumulated slip, together with the extra slip generated on the limb of a fold with an interlimb angle much less than 60° (Fig. 8a), possibly explains why so much flexural-slip has taken place on a single surface in this case.

Causes of duplex formation. It is seldom evident from an examination of the local rock sequence why a duplex has formed at a particular place. Evidence of increased resistance to slip at some point on a bedding-parallel surface, such as may be given by the localized development of sedimentary structures, is likely to have been destroyed during the development of the duplex and the formation of quartz fibre sheets along the thrust contacts. This is especially so in the case of slate duplexes and other small-scale structures. It is also difficult to distinguish between the local sedimentary thickening of a bed, and thickening which resulted from layer-parallel shortening which occurred in advance of the propagating movement horizon and actually caused the duplex to form (cf. Cooper *et al.* 1983).

Most duplexes, such as that at locality **e** (Fig. 6) show little variation in the thickness of the bed or package which has been imbricated (bed thickness varies from 5 to 6 cm over a distance of 2 m parallel to the transport direction), and significant variations associated with other duplexes are almost certainly of sedimentary origin. An example of the latter is duplex **1** (Fig. 4) shown in Fig. 10(c), in which the imbricated bed increases in thickness from 2–4 cm (at A where it is a cross-laminated sandstone with an eroded top) south of the duplex, to 5–6 cm within the duplex, and thins to 5.5 cm at B to the north of it. These thickness changes are associated with changes in the nature of the bed and are unlikely to have resulted from bedding-parallel shortening immediately preceding duplex formation. At Lower Longbeak, outwith the duplex and north of it, the thickness of the sandstone between the two marker horizons is 14 cm (A and B, Fig. 8b). This thickness varies little within the duplex and even in the most rotated horses it still averages 14 cm (range: 12–16 cm). The beds below horizon A are 18 cm thick north of the duplex and 17 cm

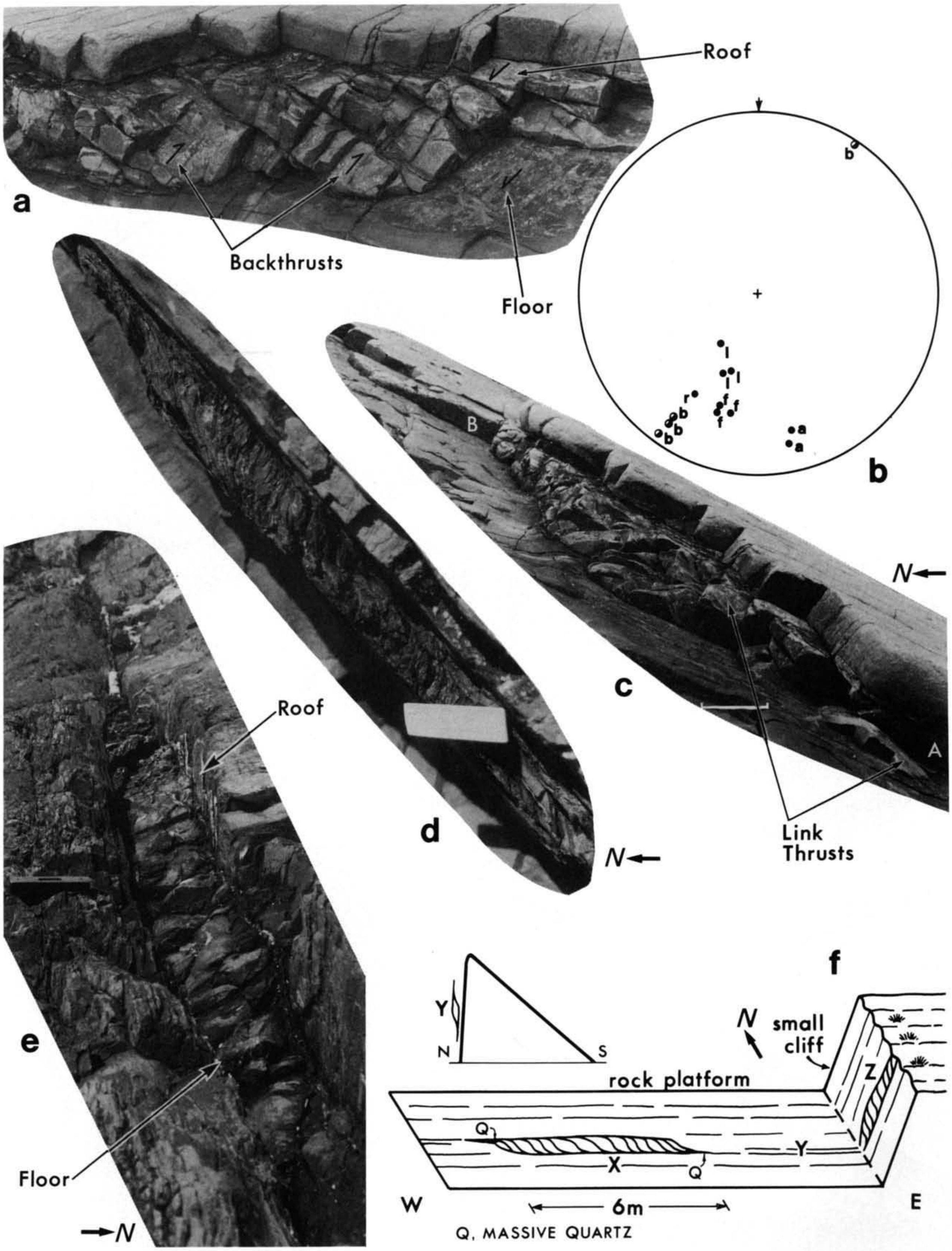

Fig. 10. Three-dimensional geometry and morphology of sandstone and slate duplexes. (a) & (c) Frontal and side views of sandstone duplex 1 from Maer High Cliff (Fig. 7). A and B see text. (b) Orientation of slickenfibres on thrust surfaces in duplex **1**: a, oblique to main set; b, backthrust; f, floor thrust; l, link thrust; r, roof thrust. (d) Slate duplex **d** (Fig. 6). (e) Sandstone duplex **h** (Fig. 6); note the low-angle link thrusts, and near-vertical slickenfibres on the roof thrust. (*f*) Sketch of the three-dimensional relationship of the duplex in (e) to its probable continuation in a low cliff to the east; for *X*–*Z*, see text. The fold profile above shows its local structural setting. The scales in (a)–(d) are 5.5 cm long and that in (e) is 15 cm long.

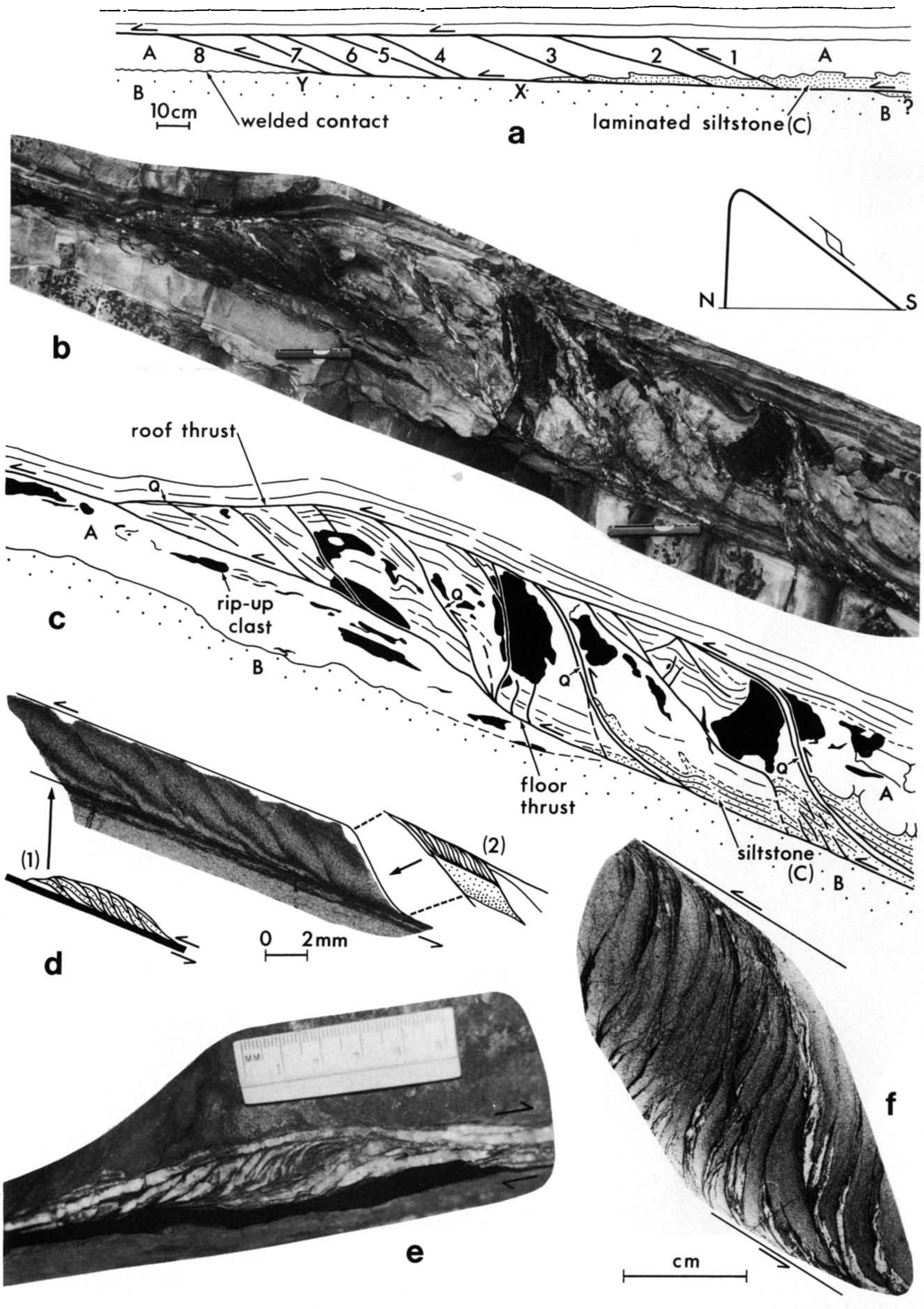

Fig. 11. Internal morphology of duplexes. (a) Diagrammatic reconstruction of the beds affected by duplex **3** (Fig. 7). (b) Photographic mosaic, and (c) line drawing, of duplex **3** from Maer High Cliff. Note the change in scale across each, marked by the spirit levels, each 10 cm long, in (b). Q, thick bedding-parallel quartz veins. (d) Small-scale duplex in siltstone from locality **f** (Fig. 6). (e) Duplex formed from an imbricated movement horizon, locality **q** (Fig. 6). (f) Part of the sub-duplex to duplex **h** (Fig. 6).

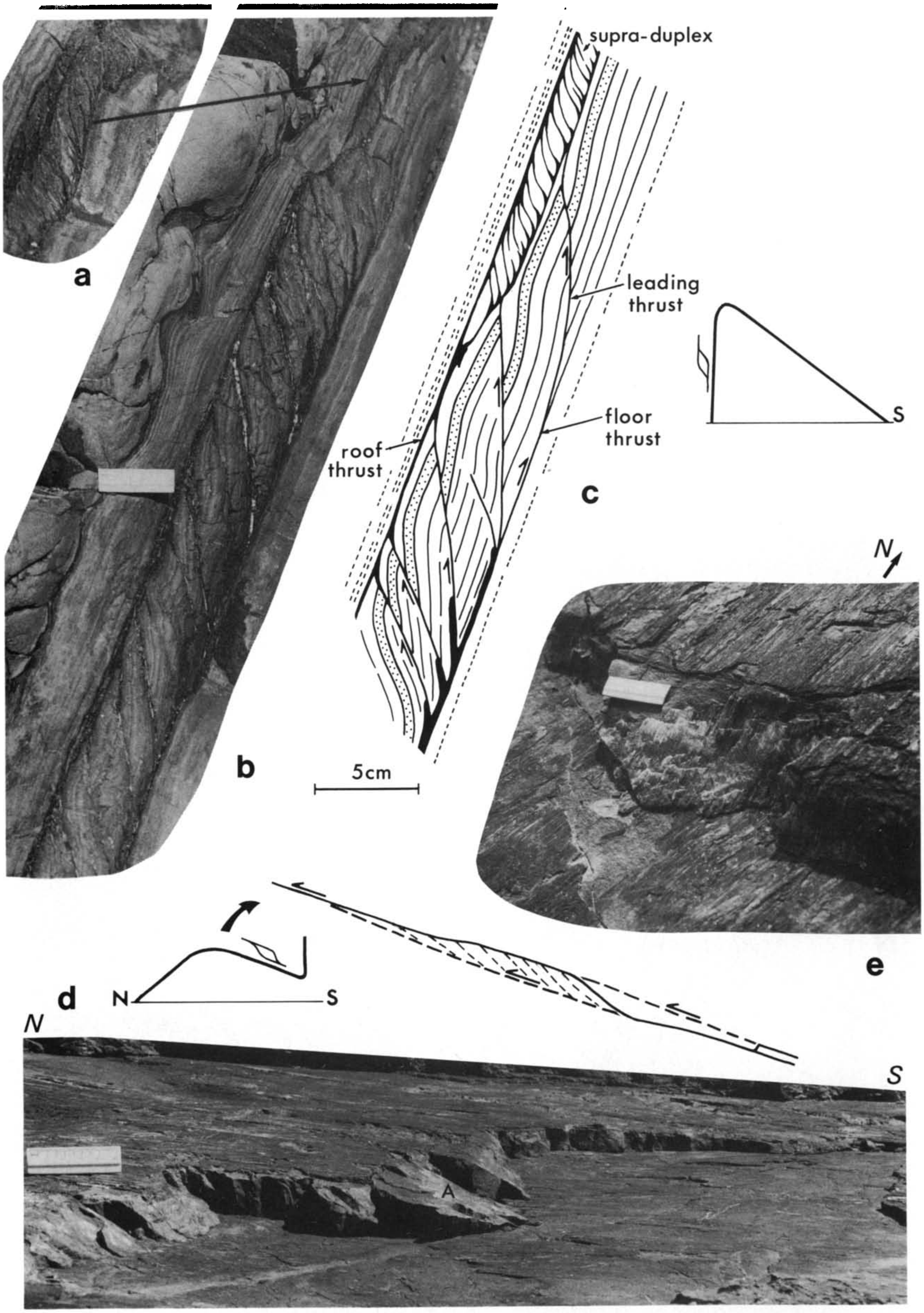

Fig. 12. Details of multilayer duplex **6** from Maer High Cliff (Fig. 7) (a)–(c), and of sandstone duplex **I** from Hartland Quay (Fig. 6) (d) & (e). (a) Detail of the supra-duplex in duplex **6**, the frontal part of which is shown in (b). (c) Drawing of the toe region of this duplex viewed approximately normal to the transport direction as shown by slickenfibres on the floor thrust; the structural setting of the duplex is shown to the left of (c). (e) is the frontal view (looking northwest) of link thrust A shown in photograph (d) of duplex **I**. Note the clear sense of shear given by the fibre steps. The line drawings show the overall profile, correct orientation, and structural setting of this duplex. The scales in (b), (d) and (e) are 5.5 cm long.

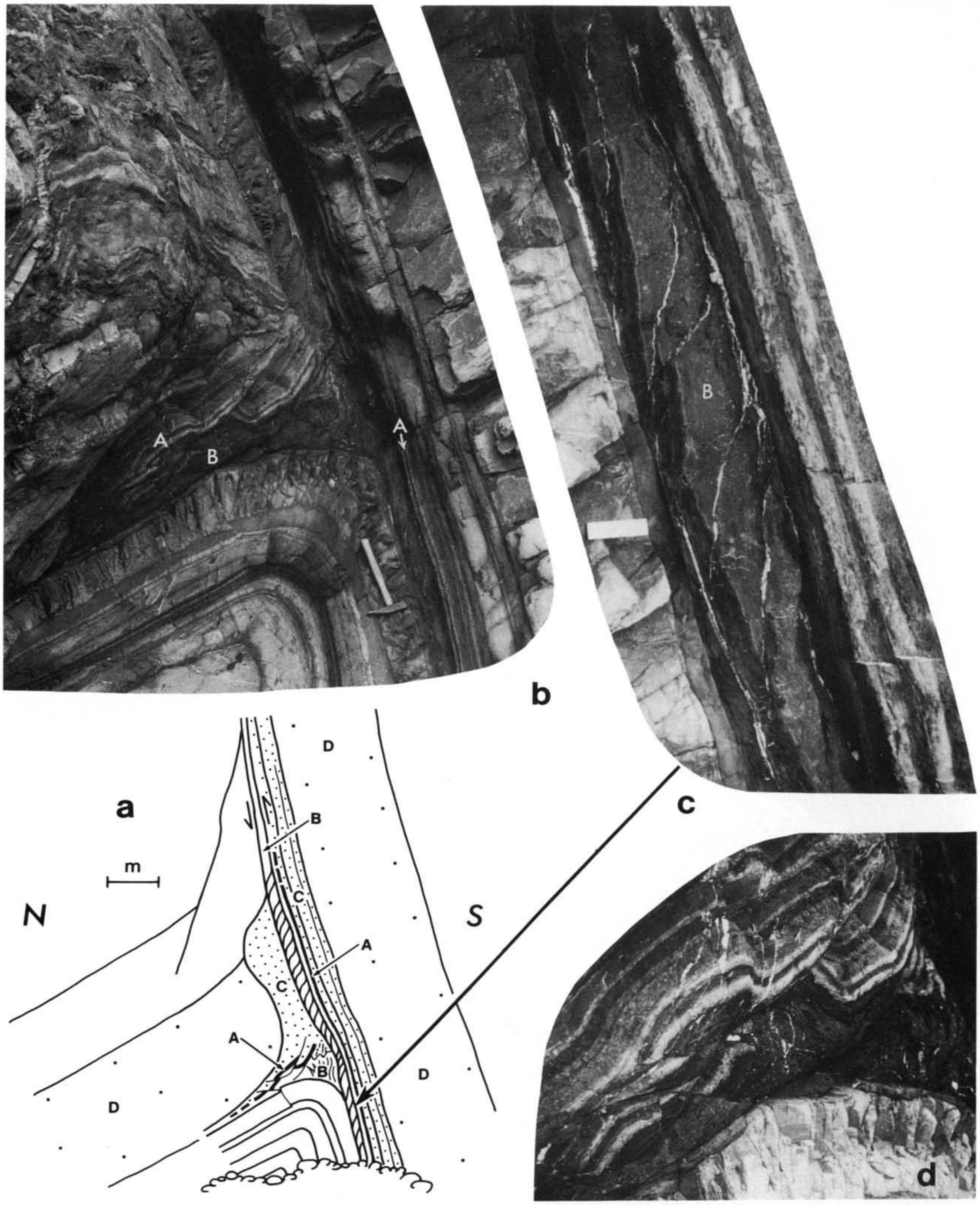

Fig. 13. Location and morphology of a late, fold-related duplex from the south end of Compass Cove, west of Bude (SS 199060). Synoptic sketch (a), and photograph (b), show a bedding-parallel duplex seen near beach level on the south limb of an anticline; it continues upwards along a slip surface which cuts across the axial surface of the fold. A–D, stratigraphical units seen on both limbs of the fold. The hammer is 45 cm long. (c) The flexural-slip duplex developed in mudrock on the south limb of the fold. All thrust surfaces are marked by quartz fibre veins. (d) Details of accommodation structures in the hinge zone of the anticline.

thick where the sequence appears complete within the duplex.

A clear illustration of the effect of facies change in localizing duplex formation is given only by duplex **3** (Fig. 7). North and up-dip from the duplex (Figs. 11a–c) two sandstone beds are seen: an upper one with abundant rip-up clasts of mudrock (A) which lies with an erosional contact upon a lower sandstone bed (B) which shows marked lateral thickness changes and is affected by soft-sediment deformation. Before deformation a schematic reconstruction (Fig. 11a) shows that the upper bed passed south into a package consisting of a sandstone bed with a strongly loaded base overlying a 3–4 cm-thick laminated siltstone sequence (C) which had an undulating contact with the thick sandstone bed (B) below. It can be inferred that slip on the lower movement horizon propagating north along the base of the laminated sequence was impeded between X and Y (Fig. 11a) because farther north the extension of this surface passed laterally into a welded, erosional, contact between two sandstone beds. As a consequence, the movement horizon ramped up to a slip surface above the upper sandstone bed and subsequent collapse of the footwall ramp containing that part of the sequence in which the most marked facies changes occur, led to the development of duplex 3. Thrusts 1–5 (Fig. 11a) appear to be early as they have been strongly rotated during the formation of the structure and are marked by thick quartz fibre veins (Fig. 11c), indicating that considerable movement has taken place on them. The earliest thrust (1) cut through unit A where it was thin, some 1 m in advance of the 'sticking point'. An alternative interpretation is that thrust 4 formed first, followed by 5–8, and 3–1 as 'out-of-sequence' thrusts, but this is less likely because of the features noted above.

Stacked duplexes are common on all scales from 0.3 to 90 cm thick (Fig. 8), with the smaller duplex usually being the lower one (sub-duplex) and lying on a planar floor thrust. In no case does the upper duplex appear to have been deformed by the formation of the sub-duplex and the latter must have formed either earlier than, or at the same time as, the main duplex. A possible sequence of events is that a small-scale duplex formed first from a thin bed above a forward-propagating movement horizon, the duplex impeded further slip on that surface and caused the movement horizon to ramp up through the overlying bed to an easier-slip horizon, and this led to the development of the upper duplex. An example of an early stage in this process is possibly seen in Fig. 9(b) where a movement horizon has ramped up through the sandstone bed at A to avoid the slate duplex; further slip on this surface could have led to the imbrication of the sandstone bed and the formation of an upper duplex.

Alternatively, both structures in a stacked duplex could have developed synchronously to minimize space problems in a duplex forming between active roof and floor thrusts, with beds or packets of beds that differ marked in relative thickness (? and rheology) forming separate imbricate structures. This mechanism seems more appropriate for the development of the supra-duplex shown in Figs. 12(a)–(c). In this case the movement horizon may have ramped up through the multilayer sequence, initiating formation of the main duplex structure, only to encounter a 'sticking-point' on the new roof thrust which then led to the contemporaneous development of both the supra-duplex and of the main duplex beneath it.

The distinction between a sequential and an integrated development of the two parts of a stacked duplex is important as in the first case the total displacement across the structure is the sum of the displacements on each duplex, and in the second case the displacements may be equal and each equivalent to the total displacement.

Late-formed duplexes

These are of two types: duplexes incorporating fibre vein material which had formed previously on flexural-slip movement surfaces, and duplexes associated with late-stage modifications to the geometry of major flexural-slip folds.

When bedding-parallel slip continues after a movement horizon has ceased to accrete new layers of fibrous quartz, the vein itself can become imbricated to form a miniature duplex (Fig. 11e). Alternatively, where asperities have developed on a vein surface during slip in one direction on a fold limb, when the slip vector changes direction during the progressive amplification of the fold, these irregularities can impede further movement and the vein may become imbricated as a result. These types of *fibre vein duplex* are uncommon, develop on the same scale as the slate duplexes, and occasionally form hybrid structures in which a fibre vein and the overlying thin slate band are imbricated together.

As flexural-slip folds amplify, movement horizons pass from concordant (bedding-parallel) on the limbs to cross-cutting in the hinge zone, and continue for a short distance across the opposing limb (Tanner 1989, fig. 22). In some cases the beds break in the hinge zone and a shear zone develops parallel to the beds on one limb. Figure 13 illustrates this situation, where a bedding-parallel duplex (in slate unit B) continues along a slip surface which cuts across the hinge zone of the fold. The roof and floor thrusts of this duplex are marked throughout by fibre veins up to 0.5 cm thick which carry a steeply plunging slickenfibre lineation. Nearby folds south of Efford Ditch (SS 191056) show fractures of this type in the hinge zone which are up to 50 m long in plan view and also carry steeply plunging slickenfibre lineations.

SHEAR SENSE

All of the 27 duplexes described in the second part of the previous section are contractional and *all* show a shear sense consistent with flexural slip on the fold limb on which each duplex is located. The shear sense inferred from the way in which the floor thrust ramps up to a higher stratigraphic level, or from the sigmoidal shape of the horses (see Fig. 9 for some examples), is in several cases confirmed by the shear sense given by fibre steps on the link thrusts between the horses as shown in duplex **l** (Fig. 12e) or, less commonly, by fibre steps on the floor or roof thrusts. Other good examples of fibre steps on the link thrusts are shown by duplex **d** in Fig. 6, where clear fibre steps are preserved on five of the link thrusts (Tanner 1992, fig. 4c); duplex **1** (Figs. 10a–c); and the Lower Longbeak duplex (Fig. 8). Backthrusts which are smaller in scale than the forethrusts and generally restricted to a single horse are found in many duplexes and fibre steps on some of these (i.e. Fig. 10a) confirm the sense of displacement.

Of the duplexes listed above, 18 lie on the south limbs of anticlines and have a N-directed shear sense, and nine lie on the north limbs of anticlines and have a S-directed shear sense. The chance of this occurring as a result of random selection is 1 in 10^8. The unequal partition between N- and S-directed shears may be partly due to a sampling bias, as the accessible part of the cliff section on the more gently dipping south limbs of anticlines is longer than that on the steep to vertical north limbs.

At Hartland Quay and Maer High Cliff, movement horizons are spaced at 10–100 cm apart and the location and shear sense of individual duplexes is shown on Figs. 6 and 7. It is significant that, in both areas, duplexes from approximately the same stratigraphical level on two adjacent fold limbs show a reversal of shear sense across the fold axis. For example, duplexes **h** and **i**; and **j–n** and **o–r** at Hartland Quay (Fig. 6) and duplexes **3–5** and **6** at Maer High Cliff (Fig. 7). Duplex **6** occurs 4.35 m stratigraphically above the top of shale unit A (Fig. 7); on the opposite (north) limb of the syncline duplexes **3**, **4** and **5** occur at 1.16, 2.02 and 5.36 m, respectively, above the same contact. Also on the north limb, a movement horizon which occurs 4.36 m above shale A is marked by ~2 cm of disrupted shaly material which lies 3.5–8 cm below a 20–24 cm-thick sandstone bed with a strongly loaded base. This movement horizon and the overlying bed appear to lie at precisely the same stratigraphic level as those at locality **6** on the south limb.

The above relationships argue against the possibility that packages of N- and S-directed structures occur at different levels and are located by chance on alternate fold limbs. Purely on a basis of shear sense, they may represent populations of forward- and hindward-propagating structures at the same level, developed during layer-parallel shortening prior to, or synchronous with, fold initiation (Price & Cosgrove 1990, p. 392) but their close association with laterally extensive flexural-slip horizons precludes this possibility. Alternatively, they could be compared with soft-sediment duplex-like structures such as those recently described from fan channel sequences in the Ouachita Mountains, Arkansas (Shanmugam *et al.* 1988) which apparently formed with opposing transport directions in adjoining sandstone units. However, even if either of these explanations were possible it would be a remarkable coincidence that the duplex structures in the Culm Measures of SW England of the type classified here as flexural-slip duplexes should divide without exception into two groups on the basis of present-day limb attitudes. The significance of this observation is considered further in the Discussion.

DISPLACEMENT DIRECTION

Flexural-slip surfaces (movement horizons) between beds on the limbs of folds at Hartland Quay are marked by quartz fibre veins which preserve one or more fibre orientations and show a consistent shear sense reversal from limb-to-limb throughout the section (Tanner 1989). Statistical analysis of these data show that the fibres have a mean orientation within 1σ of being at right angles to the related fold axis (Tanner 1989, fig. 16) with slightly oblique lineations (at up to 20° to the mean orientation) representing adjustments in slip vector resulting from the development of slightly curvilinear, doubly plunging major folds. More strongly oblique lineations possibly represent late (? post-folding) adjustments on the fold limbs.

The most significant geometrical features of the duplexes described in this paper are that (a) slickenfibre orientations on the floor and roof thrusts are, in all cases where they are measurable, statistically indistinguishable from those on movement horizons found between packets of rock throughout the same fold limb, and (b) link thrusts are commonly oblique to the mean movement direction and in some cases are lateral ramps.

Slickenfibre orientations on floor and roof thrusts

In the section north of Hartland Quay, slickenfibre orientations for duplexes **b** (slate duplex, Fig. 9a), **d** (sandstone duplex (Tanner 1992, fig. 3d)) and **h** (multilayer duplex (Figs. 10e & f)) are shown on Fig. 6. The geometry of the major folds in this section has been described by Tanner (1989): they plunge at low angles to the ENE or WSW and the slickenfibre lineations lie in a plane which is statistically almost orthogonal to the fold axes. Duplex **h** lies on the steeply dipping north limb of an anticline and the slickenfibres on the floor and roof thrusts are indistinguishable in orientation from those measured on movement horizons from the same limb (Fig. 6, projection h). The same correspondence between slickenfibre orientations on the floor and roof thrusts, and on movement horizons, is seen on the moderately dipping south limbs of anticlines (Fig. 6, projections b and d). In addition, two sets of slickenfibres are found on the floor and roof thrusts of duplex **b**, and three sets on roof thrust of **d**: they correspond to the oblique sets of slickenfibre lineations previously documented from movement horizons on most fold limbs in this area (Tanner 1989, pp. 646–647). Two sets of slickenfibres are also found on the roof thrust of duplex **e** and the presence of these multiple sets of slickenfibre lineations further strengthens the link between the kinematic history deduced for the movement horizons and that for the duplexes.

At Maer High Cliff the major folds plunge at about 17–265° (Fig. 7c) but the slickenfibre lineations lie within a plane whose pole is at 18° to the fold axes. Whether or not the cause for this oblique-slip was a transpressional element in the deformation, as was suspected for the southern end of the Hartland Quay section (Tanner 1989), it is significant that slickenfibre orientations on the floor and roof thrusts of duplexes **1–5** (mean: plunge 34–186°; N = 13) cluster *precisely* in the same field of the equal-area projection as do the movement horizon slickenfibres from the moderately dipping south limbs of the anticlines (mean:plunge 35–188°; N = 14) (Fig. 7b). The steeply plunging slickenfibre lineation on the floor thrust of duplex **6** also lies within the field of steeply plunging movement horizon lineations from the north limbs of the anticlines (Fig. 7c). As in the Hartland Quay section, two sets of slickenfibre lineations are found on the floor thrust of duplex **1**.

The Lower Longbeak duplex is situated on the north limb of an anticline whose axial plane dips at a moderate to low angle northwards (Fig. 8a) and whose axis plunges at about 16–306°. Despite its size, it is similar in morphology to other flexural-slip duplexes in these rocks, has fibre veins on all of the thrust surfaces, and shows the appropriate sense of shear. In addition, the slickenfibre lineations on floor and roof thrusts have a very similar orientation (mean:plunge 16–348°; N = 10) to those on movement horizons on the same fold limb (mean: 21–002°; N = 10) (Figs. 8c & d). Later deformation has warped the major folds in whose common limb the duplex is located, and this makes such a comparison less precise than in the case of duplexes associated with upright folds whose planar limbs are virtually unaffected by subsequent deformation. It cannot however disguise the fact that, although the bedding-parallel slip vector is even less oblique to the fold axes than at Maer High Cliff, the slickenfibre orientations associated with the duplex again closely mirror those found on the movement horizons. As with the other duplexes, the floor thrust of the Lower Longbeak duplex shows several sets of slickenfibre lineations.

Oblique displacements on link thrusts

Link thrusts commonly strike obliquely to the transport direction as defined by slickenfibre lineations on the floor and roof thrusts. The angle between the line of intersection of the link thrust with the floor thrust, and the slickenfibre lineation on the floor thrust, can be as great as 40° (duplex **1**, Figs. 10a–c) but is generally 10–20°. In the Upper Longbeak duplex the total variation in strike shown by 10 link thrusts is 20° (Fig. 8c) and they are arranged almost symmetrically about the transport direction.

The slickenfibre lineations on the link thrusts are commonly oblique to those on the floor and roof thrusts, and vary in trend (as well as in plunge due to the variable dip of the link thrusts) from horse to horse in duplexes of all sizes from slate duplexes to large multilayer duplexes. A typical example is illustrated in Fig. 12(e), where the difference in trend between slickenfibres on the link thrust and those on the floor of the duplex is 22°. In other duplexes such as **b**, **e** and **h** (Fig. 6), where the slickenfibres on the floor and roof are not parallel, those on the link thrust tend to be parallel with, or at a small angle to, the main set of slickenfibres on the floor thrust but are more highly oblique to those on the roof thrust. In duplexes **3** (Fig. 7) and **d** (Fig. 6) all three sets are approximately parallel but in the Upper Longbeak duplex the slickenfibres on the link thrusts range in trend by 53° compared with a variation of 24° for all orientations of slickenfibres found on both roof and floor thrusts (Fig. 8c). In a few cases, multiple sets of slickenfibres are found on a single link thrust, the total variation in pitch being <30°.

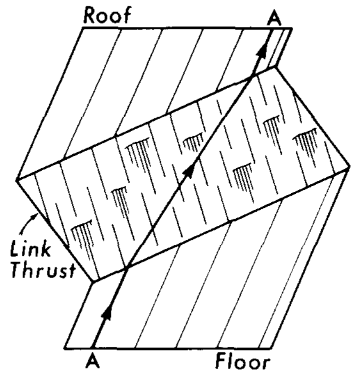

Fig. 14. Typical movement vector shown by slickenfibres on a link thrust in a flexural-slip duplex, compared with the trajectory A–A predicted by Apotria *et al.* (1992) for movement of the hangingwall across an oblique thrust ramp. Fibre steps on the link thrust are shown diagrammatically.

Where the link thrusts strike at an oblique angle to the transport direction, the slickenfibre lineation on the floor thrust changes trend towards a more steeply plunging down-dip orientation as it continues across the link thrust (Figs. 12e and 14). This change in slip direction does not agree with that predicted by Apotria *et al.* (1992) for the passage of the hangingwall over a footwall ramp oblique to the main transport direction (Fig. 14) but instead shows that naturally-formed duplexes have a more complex internal three-dimensional geometry and movement pattern: material is expelled laterally from the structure as it develops (cf. McClay & Insley 1986). A more detailed analysis is not possible as it is not known whether the lineations now seen on the uppermost fibre sheets on the floor, roof, and link thrusts actually formed at the same time, and the presence of two or three sets of lineations on some of these surfaces further complicates the situation.

Backthrusts

These occur in most duplexes but are less common than the forward-propagating thrusts, are smaller in size, and generally restricted to a single horse. In two cases where it was possible to measure the slickenfibre orientations on the backthrusts they were found to lie within the same swathe of orientations on the equal-area projection as slickenfibres on the floor, roof and link thrusts (Figs. 8c and 10b). This suggests that they formed broadly contemporaneous with the other structures and are part of the same overall kinematic picture.

Summary

Slickenfibres on the floor and roof thrusts of duplexes from different limbs of the major folds (which vary in attitude from vertical to gently dipping) are identical in their mean and range in orientation to those found on flexural-slip surfaces on the corresponding fold limb. If the duplexes had formed earlier and were folded by the major folds then some angular divergence between the two sets of linear structures might be expected, especially as the flexural-slip lineations have developed progressively during rotation of the fold limbs, and lineations from the two limbs only become co-linear as a result of 'unfolding' when the major fold axis is horizontal (Tanner 1989). It is also unlikely that pre-folding duplexes would maintain an identical geometrical relationship to the major folds in areas which are separated by a cross-strike distance of over 20 km in the fold belt. Careful examination of the geometry of many natural examples shows that any single section through these duplexes is non-restorable.

TIMING OF DUPLEX DEVELOPMENT

The highly ordered duplex structures of flexural-slip origin described in this paper occur on a variety of different scales, have quartz fibre veins on all of the thrust surfaces, and developed in rocks which had already been lithified. This conclusion is supported by the occurrence of highly polished abraded surfaces on some of the link thrusts, especially in mudrock. These duplexes contrast with the much less common pre-folding duplexes and imbricate structures found in the same rocks, which have a more disordered, even chaotic, appearance, lack quartz veins of any type on their thrust surfaces, and are associated with much small-scale fracturing and brecciation.

The shear sense reversal shown by the duplexes, which occurs from limb to limb across the area (Fig. 6), suggests that these structures formed as a result of bedding-parallel slip after folding began, rather than during the bedding-parallel shortening which preceded it. Layer-parallel shortening which occurs with the minimum principal compression direction (σ_3) vertical and orthogonal to bedding gives rise to small thrusts which develop with opposing dips and shear sense, and often only affect single competent beds in a multilayer sequence. Some of the isolated thrusts and bedding wedges which occur in the rocks described here may have formed in this way (Price & Cosgrove 1990, pp. 389–391). However most of the small thrusts and ramps seen in the sandstone beds formed during the flexural-slip movements and mark the place where a *laterally continuous* bedding-parallel slip surface has changed

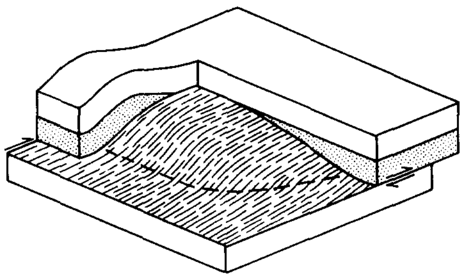

Fig. 15. A lens-shaped wedge block, after Faill (1973). Slickenside lineations are shown by pecked lines.

level within the sequence. The 'wedge faults' described by Faill (1973) from the Valley and Ridge province in Pennsylvania are of the latter type and probably represent the first stage in development of a flexural-slip duplex (Fig. 15). The slickenside lineations on the wedge faults lie in the same plane, at right angles to the average fold axis (Faill 1973, fig. 13), as that defined by the slickenside lineations on the folded bedding planes. Bedding 'slips' and wedges figured by Cloos (1961) are more difficult to interpret due to a lack of information about the nature and location of some of the thrust planes, and of the orientation of the lineations on them, but some (especially those in figs. 11–13) appear to have resulted from flexural-slip and others (i.e. fig. 3) (Price & Cosgrove 1990, fig. 15.6) from layer-parallel shortening prior to folding.

Other features of the duplexes described here which are crucial in linking their development to the flexural-slip process are: (a) the two or more sets of slickenfibre lineations found on floor, roof, and link thrusts in duplexes correspond closely in orientation to the slickenfibre lineations found on the movement horizons on the same fold limb, even where these become increasingly oblique to the major fold axes southward from Maer High Cliff to Lower Longbeak; (b) the laminated quartz-fibre veins found on the roof and floor thrusts in the duplexes reach a thickness of 2 cm, comparable to that of the thicker movement horizon veins, suggesting in a qualitative manner (see Tanner 1989, p. 645) that the amounts of slip on each were of the same order; and (c) the fibre veins in the duplexes and on the movement horizons show identical features in thin section such as slickolites, occasional crack–seal inclusion bands and mylonitic fabrics.

The relationship between cleavage associated with the major folds, and the duplexes, is difficult to see in the field as the sandstones only develop a spaced cleavage in the hinge zones of the folds. However a finely spaced cleavage congruous to the major folds is clearly overprinted on part of duplex **h** at location Z (Fig. 10f) and passes without deflection across the roof thrust and parts of adjoining horses. These relationships merely show, as would be expected for all of the duplex types discussed so far, that movement on the duplex thrusts had ceased before the main penetrative cleavage developed in the rocks. Cleavage in the country rocks also generally abuts against the movement horizon veins and is seen in thin section to continue undeflected in rocks on either side of them. In one instance, cleavage is deflected into the movement horizon over a distance of <2 cm and the latter appears to have been an active shear zone during or after cleavage formation.

DISCUSSION

The external morphology of flexural-slip duplexes was compared to that of well-documented examples from other contractional settings by Tanner (1992) and it was noted that most of the latter have a smooth-roofed, streamlined shape and lack the hangingwall anticlines above each link thrust that are a characteristic feature of the classical duplex model (Boyer & Elliott 1982) (Fig. 3). In the flexural-slip duplexes this is due to the small amount of displacement which occurs on each link thrust, and the low angle of the leading ramp (far less than the 30° dip in the classical model): each horse is not carried sufficiently far up the ramp and on to the roof thrust for a hangingwall anticline to develop. These features are clearly seen in Fig. 12(c); at no stage in the development of a flexural-slip duplex is the roof thrust arched up as it is in the Boyer & Elliott (1982) model. A spectrum of duplex geometries can therefore be recognized from the flat-roofed type, through the Boyer & Elliott type, to the antiformal stack (Figs. 16a–c), the distinguishing factor being the amount that each horse has been transported forward in proportion to its total length. It is a striking feature of these structures that with few exceptions a similar displacement pattern is shown by all of the horses in a single duplex. The gross differences in morphology seen in Fig. 16 could be explained simplistically as being due to differences in confining stress normal to the floor thrust (decreasing from c to a) but it is more likely that these structures originate by different mechanisms.

Previous attempts to explain the origin of flat- and smooth-roofed duplexes (Groshong & Usdansky 1988, Cruikshank *et al.* 1989) have relied upon varying the parameters used in kinematic models of duplex formation to obtain this rather special configuration. Recent detailed mapping of strike-slip fault zones has however provided an alternative, and very different, solution to this problem. Some of the strike-slip faults in the Entrada Sandstone in Arches National Park, Utah, studied by Cruikshank *et al.* (1991) are band faults (narrow zones in which the constituent grains in the rock are crushed and slip is distributed to accommodate cm-scale displacement; Aydin 1978) which consist of a series of overlapping segments. As these segments propagate laterally and eventually overlap, the tips curve inwards and transfer zones develop between them. Within the transfer zones, band faults form at a low angle to the strike-slip fault segments and in some cases develop into smooth-roofed duplex-like structures (Figs. 17a & b) which are from 10 cm to tens of metres long, and closely resemble the duplexes of flexural-slip origin described in this paper. A similar duplex-like structure is described from a normal fault in the same area (Fig. 17c). An important conclusion with respect to the possible mode of formation of flexural-slip duplexes is that the complete pattern of fractures in the transfer zone is thought to have formed *before* any movement takes place on them (Cruikshank *et al.* 1991). These authors also conclude that the present structural relief (S on Fig. 17) is largely a function of the original curvature of the bounding faults (and their separation) and is much greater than would be produced for the small observed displacement by the kinematic (Boyer & Elliott 1982) model.

A complex extensional duplex analysed by Swanson (1990) from York Cliffs, southern Maine, was likewise considered to have formed in a transfer zone in a strike-slip fault system and the overall morphology of the structure, in particular the double-tapered, lozenge-shaped horses (Fig. 17d) is identical to that of some of the sandstone duplexes from North Devon (Fig. 9c) (Tanner 1992, fig. 4c), although the latter are of undoubted contractional origin. Swanson (1990, fig. 12) inferred that new fractures in the York Cliffs duplex formed symmetrically at either end of the transfer structure as it developed. Similar transfer structures were

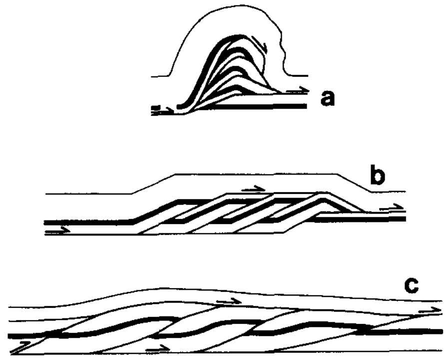

Fig. 16. Contrasting geometries of (a) an antiformal stack, (b) the Boyer & Elliott (1982) duplex model and (c) a typical flexural-slip duplex.

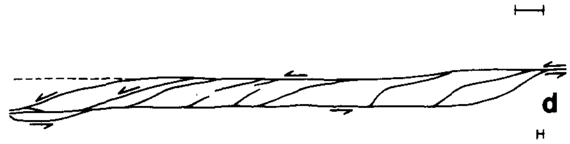

Fig. 17. Duplexes formed as transfer structures between overlapping segments of strike-slip and normal faults, and a sidewall ripout structure developed on a strike-slip fault. The bar scale in all cases is 10 cm long. S, step height. (a) & (b) Duplex-like structures from the Entrada Sandstone, Utah (after Cruikshank *et al.* 1991, figs. 5d & e). (c) The Three Penguins duplex, formed between segments of a normal fault in the Entrada Sandstone (after Cruikshank *et al.* 1991, fig. 5g). (d) An extensional duplex formed between overlapping strike-slip fault segments in quartzite (after Swanson 1990, fig. 7).

modelled theoretically by Aydin (1988) for faults with a thrust displacement and the results used to explain the mode of formation of the cleavage duplexes described by Nickelsen (1986).

In flexural-slip folding, movement horizons (bedding-parallel slip surfaces) develop as the beds rotate on the limbs of a growing fold. As the limb dip increases, new movement horizons nucleate (Tanner 1989) and propagate throughout the structure parallel to bedding. When a fault tip encounters an obstacle, the fault ramps up to a higher 'easy-slip' horizon and an asymmetric duplex results, such as that in Fig. 3. The same morphology would result where local layer-parallel thickening of a bed or packet of beds occurred in the ductile bead in advance of the thrust tip and prior to the development of the link thrusts.

An entirely different scenario results when the tips of two movement horizons propagating in opposing directions overlap, and a transfer structure consisting of fractures at a low angle to bedding develops between them. New link thrusts may form symmetrically at opposing ends of the transfer structure and, depending upon the relative displacement between the two movement horizons, the earlier thrusts may be rotated and steepened. The presence of a 'sticking point' could be important in determining when and where a transfer structure forms, as duplexes of this type only appear to be common where the slip surfaces are not perfectly planar and continuous. As the duplex develops between two essentially parallel thrusts the total displacement will be shared between all of the link thrusts and movement may take place on any one of them throughout the development of the structure. A feature that might be expected is that the floor and roof thrusts would continue beyond the limits of the duplex, as shown by the largest of the duplex-like structures described by Cruikshank *et al.* (1991, p. 1193) (see Fig. 16c). This was a previously unexplained feature of several of the North Devon duplexes, for example the Lower Longbeak duplex (Fig. 8b), and many of the duplexes described here from SW England may have formed as transfer structures, especially the sandstone duplexes and the structure illustrated in Fig. 9(d) associated with one of the Maer Cliff duplexes. This mode of development would explain not only the smooth roof to many of the duplexes but also the low take-off angle of thrusts in the leading and trailing ends of the structure, the lack of hangingwall anticlines on the link thrusts, and the common occurrence of out-of-plane and out-of-sequence movements in naturally-occurring examples (McClay & Insley 1986, Tanner 1992). The wider implication of this conclusion is that many duplexes on all scales associated with both extensional and contractional faults are transfer structures and this would resolve the paradox noted by Tanner (1992) that the morphology of many naturally-occurring duplexes does not conform with that of the Boyer & Elliott (1982) model.

An interesting problem arises when one considers the implications of the above model for flexural-slip folding

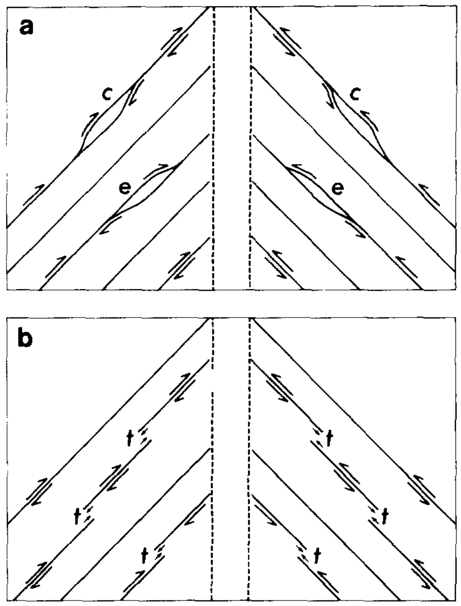

Fig. 18. (a) Hypothetical transfer structures formed on the limbs of flexural-slip folds, with half-arrows showing the sense of slip. c, contractional, and e, extensional, duplexes. (b) The configurations of overlapping segments of movement horizons which appear to have given rise to the transfer structures (t) in North Devon. The directions of propagation of the movement horizons are shown by the small arrows.

as there are two possible configurations for overlapping movement horizon segments on each limb of a fold: right- or left-stepping (Fig. 18a). Left-stepping segments result in transfer structures which have the geometry of contractional duplexes and change in profile from limb to limb of the fold whereas right-stepping segments give rise to extensional duplexes which have mirror-image profiles to the other set (Fig. 18a). As the movement horizons are constrained to being bedding-parallel (except in the transfer zones) there would appear to be an equal chance of either contractional or extensional duplexes developing on a particular fold limb. However, all of the duplexes of flexural-slip type found in the study area are contractional and this suggests that there is some control on the stepwise development of the growing movement horizons such that they always appear to climb up in the transport direction (Fig. 18b). All of the strike-slip-related transfer structures reported by Cruikshank *et al.* (1991) are also contractional but this is explained by the fact that for the conjugate set of strike-slip faults in the area, the dextral set step predominantly to the left and the sinistral set step to the right.

Significance of the fibre veins

Early duplexes and imbricate structures formed either during soft-sediment deformation or during layer-parallel shortening which preceded the folding. The chevron folds were then initiated and the identification of a later population of duplexes as being of flexural-slip origin depends primarily upon the recognition that quartz fibre-coated surfaces found on the fold limbs represent movement horizons active during the flexural-slip process (Tanner 1989). This interpretation has been challenged by Fitches *et al.* (1990) where it was applied to turbidite facies rocks folded by chevron folds in the Cardigan Bay area, Wales, but it can be demonstrated there (Tanner 1990), as in North Devon, that not only are the slickenfibres on the movement horizons statistically orthogonal to the major fold axes, but that the sense of displacement shown by fibre steps on the veins changes systematically from one fold limb to the next and always has the shear sense required by the flexural-slip mechanism. If one therefore accepts that the fibre sheets developed during flexural slip (and are only folded by the chevron folds where the latter have undergone hinge migration) then the recognition that some duplexes are associated with the formation of identical fibre sheets which show the same sense and direction of movement as is seen on adjacent movement horizons, indicates that these particular duplexes also developed during the flexural-slip process. This conclusion is strengthened by the fact that the duplexes also show a consistent shear sense reversal across the folds.

The presence of the fibre veins not only links the development of the duplexes with that of the folds but gives a clue as to the conditions under which both formed. Layer-parallel shortening which occurs under conditions of high fluid pressure prior to folding results in hydraulic jacking and the formation of bedding-parallel veins with growth fibres oriented normal to the vein walls (cf. Henderson *et al.* 1986). Flexural slip not only causes shearing and modification of any early veins, but is accompanied by the development of new fibre veins on movement horizons and on thrust surfaces in duplexes. Individual crystals or fibres in the latter veins are oriented at a very low angle to bedding. Formation of fibre veins synchronous with layer-parallel slip demonstrates that high fluid pressure played an important role in facilitating slip by reducing the normal stress across the movement horizon, and preservation of crack–seal inclusion bands in quartz fibres from such veins (Tanner 1990, fig. 1) suggests that movement took place in a stick–slip manner in response to cyclic variations in fluid pressure. Loss of this fluid during the late stages of folding led to the development of brittle features and polishing on some movement horizons and finally to a 'locking-up' of the structure as layer-parallel shear strain increased. It also resulted in the localized ripping-up of thin veins on these surfaces to form fibre vein duplexes.

The relationships between folds, veins and duplexes in these rocks are very complex. The duplexes have formed at various times from shortly after sedimentation to late in the folding process, and have formed by two, possibly three, different mechanisms. In studying the inter-relationships between, and kinematics of, such structures it is clear that the field relationships need to be ever more carefully examined and documented, and recent studies of faults and duplexes such as those by Martel *et al.* (1988), Swanson (1990) and Cruikshank *et al.* (1991) point the way.

Acknowledgements—I thank Dr R. P. Nickelsen and an anonymous reviewer for their constructive comments, and for posing a number of thought-provoking questions.

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